1　The temperature of the deep ocean is a robust proxy for global mean surface temperature
2　during the Cenozoic

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8　Reconstructing past changes in global mean surface temperature (GMST) is one of the key
9　contributions that palaeoclimate science can make in addressing societally relevant questions and
10　is required to determine equilibrium climate sensitivity (ECS). GMST may be derived from proxy
11　data compilations and/or via simple metrics that can be directly related to global temperature. One
12　such metric is the temperature of the deep ocean (T_d), which has the advantage of being
13　reasonably well-constrained for much of the Cenozoic. Previous work has suggested a simple T_d-
14　GMST scaling factor of 1 prior to the Pliocene. If this were to be accurate, it would enable GMST to
15　be precisely determined for any interval of interest. However, this metric lacks a robust mechanistic
16　basis, and indeed, such a relationship is intuitively difficult to envisage given that polar
17　amplification is a ubiquitous feature of past warm climate states and deep water overwhelmingly
18　forms at high latitudes. Here, we interrogate whether and crucially, why, this relationship exists
19　using a suite of curated data compilations generated for key deep-time climate intervals (Pliocene,
20　early Palaeogene) as well as two independent sets of palaeoclimate model simulations. We show
21　that models and data are in full agreement that a 1:1 relationship is a good approximation.
22　Mechanistically, both sets of climate models suggest that i) increasingly seasonally biased deep
23　water formation, and ii) a faster rate of land versus ocean surface warming are the two processes
24　that act to counterbalance a possible polar amplification-derived bias on T_d-derived GMST. Using
25　this knowledge, we interrogate the quality of the existing deep ocean temperature datasets and
26　provide a new Cenozoic record of GMST. Our estimates are substantially warmer than similar
27　previous efforts for much of the Paleogene and are thus consistent with a substantially higher-
28　than-modern ECS during deep-time high CO_2 climate states.

29

30　Key Points:

31　• Deep ocean temperature changes are used to constrain global mean surface temperature
32　　yet the underlying assumptions have not been scrutinised in detail
33　• Both curated data compilations and climate model simulations demonstrate that deep
34　　ocean-derived GMST estimates are robust
35　• We update the transformation equations and provide a revised estimate of GMST through
36　　the Cenozoic

1. Introduction

One of the most important contributions that quantitative reconstructions of Earth's climate can make to society is as an empirical method of constraining key aspects of Earth's climate system (e.g. Gulev et al., 2021; Tierney et al., 2020). Perhaps the most fundamental parameter of interest within this context is Earth's equilibrium climate sensitivity (ECS), which broadly describes the change in global near surface temperature per CO_2 doubling (Sherwood et al., 2020), knowledge of which is required to determine the degree to which our planet will warm over the coming centuries and millennia.

The precise definition of ECS depends on a number of factors such as which long-term feedbacks are taken into account (Rohling et al., 2012; Sherwood et al., 2020), but from a past climate perspective, constraining ECS at the very least requires a reconstruction of radiative forcing (CO_2) and global temperature, and the timescale of interest to be defined. The late Pleistocene, and the last glacial maximum (LGM) in particular have received much attention in terms of deriving ECS from the palaeoclimate record (Osman et al., 2021; Rohling et al., 2012; Schmittner et al., 2011; Sherwood et al., 2020), because direct measurements of atmospheric CO_2 are available from the ice core record, and because an enormous wealth of proxy information exists from both the terrestrial and marine realm.

The use of deep-time palaeoclimate records (pre-Pleistocene) to determine the value of important parameters like ECS has received an increasing amount of attention over the course of the last few decades (Anagnostou et al., 2016; Hansen et al., 2013; Inglis et al., 2020; Martínez-Botí et al., 2015; Zhu et al., 2019). One advantage of this approach is that CO_2 was greater than at present, providing insights into nonlinear features in Earth's climate system such as ice sheet dynamics (Foster and Rohling, 2013; von der Heydt et al., 2014) which cannot be fully determined from the study of cooler-than-modern climate states (Anagnostou et al., 2016; Inglis et al., 2020; Martínez-Botí et al., 2015; Pagani et al., 2010).

The development of precise and accurate methods of reconstructing past changes in CO_2 throughout the Cenozoic (Anagnostou et al., 2016; Foster and Rae, 2016; Hönisch et al., 2012; Pagani, 2002; Pearson and Palmer, 1999) opened up the study of geologic intervals prior to the ice core CO_2 record for this purpose, which was previously challenging in part because of the large uncertainties associated with the CO_2 data (Covey et al., 1996; Hoffert and Covey, 1992). With the production of high quality CO_2 data for much of the Cenozoic (Rae et al., 2021), the accuracy and precision with which global mean surface temperature (GMST) is known has become an increasingly important source of uncertainty in the derivation of ECS from pre-Pleistocene warm intervals (Inglis et al., 2020; McClymont et al., 2020).

Datasets designed to reconstruct past changes in GMST can be broadly placed into two categories: i) the synthesis of large amounts of spatially-distributed data sufficient to constrain the global climate at the time, or ii) the parameterisation of GMST in terms of a single, well-constrained aspect of Earth's climate. In an ideal world, the first of these is preferable as it avoids any

75 assumption that goes into indirect approaches, but it requires sufficient palaeotemperature
76 reconstructions to be available within a sufficiently narrow time interval to be able to constrain
77 GMST unbiased by (e.g.) latitudinal and zonal heterogeneities in surface temperature. However,
78 few Cenozoic warm intervals exist with sufficient data density to facilitate the calculation of GMST
79 with sufficient certainty, with possible exceptions being the Pliocene Warm Period (PWP, ~3.2 Ma
80 (Dowsett et al., 2016; Haywood et al., 2013; McClymont et al., 2020)), Miocene climatic optimum
81 (MCO, ~16-14 Ma (Burls et al., 2021)), and perhaps also intervals within the early Palaeogene (Hollis
82 et al., 2019). In this latter case, the DeepMIP project recently produced a curated data compilation
83 for the early Eocene climatic optimum (EECO; ~49.1-53.3 Ma), Paleocene-Eocene thermal
84 maximum (PETM; ~56 Ma), and the latest Paleocene (LP, the interval immediately preceding the
85 PETM; ~57-56 Ma), compiling over 1500 'high confidence' quantitative estimates of terrestrial and
86 ocean surface temperature and CO_2 for these intervals (Hollis et al., 2019), see
87 https://www.deepmip.org/data/. Using multiple methodologies, Inglis et al. (2020) derived GMST
88 for these intervals, constraining ECS to 3.1 to 4.5°C albeit with large uncertainties, while Tierney et
89 al. (2022) constrain ECS to 5.7-7.4°C using LP and PETM data coupled with a paleoclimate data
90 assimilation approach.

91 The difficulty in compiling and quality checking datasets that are large enough to constrain GMST
92 given the spatial heterogeneity in Earth's surface climate has led to the development of techniques
93 based on a simple, (relatively) easily determinable parameter. The key feature of Earth's climate
94 system that has formed the basis of a transformation relationship to GMST is the temperature of
95 the deep ocean (or rather, some closely related parameter such as the oxygen isotopic
96 composition of foraminifera), because a continuous, high temporal resolution record exists for the
97 entirety of the Cenozoic (Billups and Schrag, 2003; Cramer et al., 2011; Lear et al., 2000; Lisiecki and
98 Raymo, 2005; Westerhold et al., 2020; Zachos et al., 2001, 2008). If records such as these can be
99 reliably related to GMST, then many of the issues with generating and compiling large surface
100 temperature datasets can be avoided. This approach, pioneered by Hansen et al. (2013, 2008),
101 requires that the temperature of the deep ocean be coupled to that of the global surface climate.
102 This is likely a reasonable assumption given that the temperature of the deep ocean is broadly
103 similar to the temperature of the surface ocean in the regions of deep water formation, and the
104 temperature of these regions may, in turn, be expected to relate predictably to GMST. The
105 attraction of this approach is that GMST can be immediately calculated for any time interval of
106 interest, and it is for this reason that the approach of Hansen et al. (2013) has been widely cited and
107 reproduced (Lunt et al., 2016; Mills et al., 2019; Rae et al., 2021; Tierney et al., 2020; Westerhold et
108 al., 2020). However, the underlying rationale for the details of the methodology are complex and
109 have largely not been empirically tested, especially in deep time.

110 Here, we focus on addressing the question of whether GMST and the temperature of the deep
111 ocean are linearly related with a scaling factor of 1, as suggested by Hansen et al. (2013), given that
112 this may seem intuitively problematic. Deep water formation occurs at high latitudes in the modern
113 ocean with both fully coupled climate models and proxy data suggesting that this was similarly the
114 case for most, if not all, of the Cenozoic, although the locus of deep water formation likely shifted

115 (Ferreira et al., 2018; Ford et al., 2022; Valdes et al., 2021; Zhang et al., 2022). While much remains
116 to be understood about deep-time climatic variation, a ubiquitous and well-constrained feature of
117 multiple past warm climate states is that these are characterised by polar amplification (Burls et al.,
118 2021; Cramwinckel et al., 2018; Evans et al., 2018; Gaskell et al., 2022; Lunt et al., 2012) and overall
119 increased ocean stratification (Green and Huber, 2013; Winguth et al., 2012). Given that deep ocean
120 temperature records are therefore effectively a record of high latitude surface temperature, we
121 may expect a temperature record derived from the deep ocean to include a polar amplification
122 component. As a result, surface temperature estimated from that of the deep ocean using a linear
123 one-to-one relationship would, to a first-order approximation, be expected to result in a GMST
124 overestimate, with the severity of the inaccuracy directly related to the degree of polar
125 amplification in a given time interval.

126 In the following sections, we: identify the potential issues with this simple transformation of deep
127 ocean temperature (inferred via benthic foraminifera $\delta^{18}O$) into GMST in Section 2, critically
128 evaluate the quality of the Cenozoic deep ocean temperature datasets (Sec. 3.1), and then test
129 whether the methodology can usefully approximate GMST by comparing to a combination of
130 curated proxy data compilations and two sets of climate model simulations, including the DeepMIP
131 model database (Sec. 3.2 and 3.3). In comparison to previous work, especially that of Goudsmit-
132 Harzevoort et al. (2023) and Valdes et al. (2021), our analysis differs in several key respects in that i)
133 we do not limit our analysis to climate model output, ii) we explicitly set out to test what
134 relationship between GMST and deep ocean temperature is expected, for the entirety of the
135 Cenozoic, and iii) present several advances in the way in which climate model data are evaluated
136 for these purposes, detailed below (Sec. 3.3).

137

138   2. Relating global surface temperature to the deep ocean

139 The geochemistry of deep-dwelling benthic foraminifera forms the basis of our long term records
140 of deep ocean temperature change, because of the near-continuous nature of the fossil record of
141 these organisms in sediments of Cenozoic age and beyond (e.g. Westerhold et al., 2020; Zachos et
142 al., 2008). At least three proxy methods exist for reconstructing the temperature of the deep ocean
143 all of which are based on the geochemistry of benthic foraminiferal (Evans, 2021), namely, the
144 stable isotope (the oxygen isotopic ($\delta^{18}O_b$) and clumped isotope composition (Δ_{47}) of their shells
145 (e.g. Leutert et al., 2019; Marchitto et al., 2014)), and trace incorporation of metal impurities (Mg/Ca
146 (Rosenthal et al., 1997)), each with their own advantages and disadvantages, discussed in
147 supplementary text S2. For the purposes of this introductory discussion, we focus on benthic
148 foraminifera $\delta^{18}O$ and the transformation equations of Hansen et al. (2013), hereafter abbreviated
149 H13, because this approach is by far the most widely utilised, and because the benthic oxygen
150 isotope stack has a far higher temporal resolution than any other deep ocean proxy.

151 H13 first calculate the temperature of the deep ocean (T_d) from $\delta^{18}O$, and then transform this into
152 GMST. In the first step, the Cenozoic benthic $\delta^{18}O$ stack is divided into three portions on the basis

153 that i) all change in $\delta^{18}O_b$ can be ascribed to temperature prior to the growth of a major ice sheet
154 on Antarctica at ~34 Ma, ii) that ice volume changes increasingly contribute to $\delta^{18}O_b$ when Earth is
155 cooler, because there is a lower limit of the temperature of seawater. Specifically:

156 $\qquad T_d = -4\,\delta^{18}O + 12$ (Eq. 1.1)

157 prior to 35 Ma (equation 3.1 of H13), m = 4°C ‰$^{-1}$

158 $\qquad T_d = 5 - 8\,(\delta^{18}O - 1.75)/3$ (Eq. 1.2)

159 since 35 Ma and when $\delta^{18}O_b$ <3.25‰ (equation 3.5 of H13), m = 2.67°C ‰$^{-1}$, and

160 $\qquad T_d = 1 - 4.4\,(\delta^{18}O - 3.25)/3$ (Eq. 1.3)

161 when $\delta^{18}O_b$ >3.25‰ (equation 3.6 of H13), m = 1.47°C ‰$^{-1}$. This latter equation in effect covers all
162 of the Pleistocene and the glacial intervals of the late Miocene and Pliocene, and a slope of 1.47 is
163 broadly similar to the canonical view that two thirds of the Pleistocene $\delta^{18}O_b$ signal can be ascribed
164 to ice growth and decay (Raymo et al., 1989 and references therein). The result of applying Eqs 1.1-
165 1.3 to the most recent version of the benthic foraminiferal $\delta^{18}O$ stack (Westerhold et al., 2020) is
166 shown in Fig. 1A, alongside independent temperature proxies based on the Mg/Ca and clumped
167 isotopic composition of benthic foraminifera.

168 With a deep ocean temperature record, H13 separate the Cenozoic into the Pleistocene, Pliocene,
169 and pre-Pliocene. For the Pleistocene, H13 relate surface temperature (T_s) to the deep ocean by the
170 following relationship:

171 $\qquad T_s = 2\,T_d + 12.25$ (Eq. 2.1)

172 This is equation 4.1 of H13, which was rooted in a Holocene-LGM ΔT of 4.5°C, and a Holocene
173 GMST of 14.15°C, requiring that GMST changed twice as quickly as the deep ocean given a deep
174 ocean LGM-Holocene $\Delta\delta^{18}O_b$ of ~1.7‰ of which approximately two thirds is assigned to ice sheet
175 decay. Then, in the Pliocene:

176 $\qquad T_s = 2.5\,T_d + 12.15$ (Eq. 2.2)

177 Equation 4.2 of H13, which is the sensitivity required to match an "Early Pliocene mean temperature
178 3°C warmer than the Holocene" (Hansen et al., 2013). Beyond the Pliocene, H13 relate T_s to T_d in a
179 1:1 relationship, i.e.

180 $\qquad T_s^{t=t} - T_s^{t=5.3} = T_d^{t=t} - T_d^{t=5.3}$ (Eq. 3)

181 Where t=t is the time interval of interest, and t=5.3 is the base of the Pliocene. The result of
182 applying the transformation Eqs 2.1, 2.2, and 3 to both the $\delta^{18}O$ and Mg/Ca-derived deep ocean
183 temperature records (Fig. 1A) is displayed in Fig. 1B. The focus of the analysis presented here is on
184 the warm intervals of the early Cenozoic, and therefore it is Eq. 3 which we most closely scrutinise.

185 A simple illustration of whether we might expect Eq. 3 to be accurate is shown in Fig. 2A, wherein
186 surface area-weighted global ocean temperature is calculated for an arbitrary degree of tropical

187 and high latitude SST increase for an ocean planet. Specifically, a value for SST is assigned to the
188 equator and poles, varying linearly as a function of latitude, from which GMST is computed (see
189 supplementary text S1 for a comprehensive description). The contours in Fig. 2A show the ratio of
190 the increase in GMST compared to the temperature of the deep ocean, assuming that deep ocean
191 temperature is equal to surface area-weighted high latitude SST between 65-80°. Note that the
192 position of these contour lines is independent of absolute temperature, depending only on the
193 choice of latitude representative of deep water formation. The theoretical case of exactly no polar
194 amplification is represented by the line with m = 1 (equal warming at 0° and 90°), which naturally
195 coincides with the $\Delta GMST/\Delta T_{deep}$ contour equal to 1 as the high latitudes warm at an identical rate
196 to the rest of the planet. All scenarios in which the high latitudes warm at a faster rate than the
197 tropics are characterised by $\Delta GMST/\Delta T_{deep}$ relationships <1.

198 In order to determine the extent to which GMST could be overestimated by Eq. 3 within the
199 context of this simple calculation, surface ocean proxy data from three key intervals within the early
200 Palaeogene are overlain on Fig. 2A (the latest Paleocene (LP), Paleocene-Eocene Thermal
201 Maximum (PETM) and the early Eocene climatic optimum (EECO)). These average values for each
202 interval represent the mean of all data in the DeepMIP database (Hollis et al., 2019), available at
203 https://www.deepmip.org/data, conservatively excluding planktonic foraminifera $\delta^{18}O$ data points
204 impacted by diagenesis following Inglis et al. (2020). The change in tropical and high latitude
205 temperature is calculated relative to modern (Locarnini et al., 2018), averaged over 0-30° and
206 >60°N/S respectively. Two estimates of high latitude warming are shown, based on the high
207 latitude SST proxy data and the temperature of the deep ocean, the latter calculated using the
208 benthic foraminiferal oxygen isotope stack (Westerhold et al., 2020) and Eqs 1.1-1.3. We include this
209 alternative indirect assessment of high latitude SST to determine the extent to which the analysis
210 could be impacted by a potential summer bias in the high latitude SST proxy data (Hollis et al.,
211 2012), which can be avoided via the assumption that the temperature of the deep ocean may more
212 closely represent mean annual high latitude SST than the direct surface ocean proxy data (Evans et
213 al., 2018). The results of this simple analysis confirm that, whichever dataset is used to constrain
214 high latitude SST, all three intervals are characterised by high latitude warming ~2-5 times greater
215 than in the tropics, as previously described (e.g. Hollis et al., 2019; Tierney et al., 2022). While we
216 stress the simplicity of the approach used to calculate the position of the contours in Fig. 2A (e.g.
217 ignoring for the moment that this is a water world), the data suggest that, in the absence of other
218 processes driving the system in the opposite direction, the temperature of the deep ocean would
219 be expected to rise at a rate of ~0.6-0.8 times that of the global surface ocean for this degree of
220 polar amplification. If this simple analysis is applicable to Earth, then the corollary of this is that H13
221 likely overestimate Palaeogene GMST by approximately ~15-30% (less than ~0.2-0.4 times (1 – 0.6-
222 0.8) as the 1:1 relationship (Eq. 3) is anchored to the base of the Pliocene).

223 There are of course several possible reasons that this analysis may be incorrect, and several
224 processes that may mean that the key assumption of H13 (Eq. 3) is not biased by polar
225 amplification/stratification; exploring whether or not such processes exist is the focus of the
226 remainder of this contribution, and indeed, weakened polar amplification mechanisms in warm

227 climate states have been posited (Cramwinckel et al., 2018). An immediate indication that the
228 approach of H13 may perform better than might be expected based on the simple reasoning
229 outlined above is that GMST estimates based on the approach of H13 are in overall good
230 agreement with several fully independent approaches based only on surface proxy datasets (Inglis
231 et al., 2020). For example, EECO GMST estimates based on a climate model-derived transfer
232 function (Farnsworth et al., 2019) or Gaussian processes regression (Inglis et al., 2020) agree with
233 H13 to within 1.5°C, suggesting that no large bias results from applying Eq. 3 to the early Cenozoic,
234 although we note that a weakness of comparison derived (in part) from climate model simulations
235 is that most underestimate polar amplification (e.g. Burls et al., 2021; Evans et al., 2018; Lunt et al.,
236 2012).

237 For H13 to accurately reconstruct GMST requires process(es) that either i) drive GMST increase at a
238 faster rate than the average global ocean, and/or ii) bias the temperature of the deep ocean below
239 that of mean annual high latitude SST. For these processes to result in a world that is well
240 characterised by the assumption of Eq. 3, the magnitude of the effect must be exactly equal to the
241 surface area-weighted difference between the change in temperature of the global relative to the
242 high latitude ocean. This requirement was of course clear to H13, and indeed, alternative
243 assumptions were tested in that study requiring an even greater counteracting force; namely that
244 GMST increases at 1.5 times the rate of the deep ocean.

245 There are, to our knowledge, two key mechanisms that have the potential to drag GMST and deep
246 ocean temperature onto a 1:1 relationship within the context of polar amplification and increased
247 ocean stratification under past warm climate states:

248   1) While changes in global mean (near) surface temperature (e.g. Morice et al., 2012) and
249   global mean sea surface temperature (GMSST) may be approximately equated over the
250   instrumental period (Hansen et al., 2010), this is not the case when considering more
251   extreme (Cenozoic) climatic change given the greater specific heat capacity of water
252   compared to most other surface materials and a strongly different land-ocean
253   evaporative flux and different response of that flux to warming (Henry and Vallis, 2022;
254   Roderick et al., 2014). For these reasons, it is important to bear in mind that ΔGMSST
255   and ΔGMST are not equivalent throughout much of the Cenozoic (see e.g. Fig. 1B;
256   Gaskell et al., (2022)).
257   2) A bias in the season of deep water formation towards the winter (or possibly, an
258   increased seasonal bias as GMST increases), could counteract the effect of amplified
259   high latitude SST increase relative to the global mean. In the modern ocean, there is no
260   strong seasonal trend in meridional overturning circulation (MOC) in the North Atlantic
261   west of Greenland and a strong seasonal variation in MOC between Greenland and
262   Scotland, although with a seasonal timing that varies between years (Wang et al., 2021).
263   In contrast, the mixed layer depth (MLD) close to Antarctica is characterised by a strong
264   seasonal variation (Pellichero et al., 2017), with sea ice-driven densification thought to
265   play an important role in deep overturning (Pellichero et al., 2018). Given a likely

266   increase in high latitude SST seasonality under past greenhouse climate states (Hollis et
267   al., 2012), it is at least possible that overturning was biased towards winter to a greater
268   degree than at present during these times, although we note that a sea-ice related
269   seasonality in deep water formation is unlikely to have been a relevant process for
270   much of the Cenozoic.

271   A simple illustration of this from a theoretical point of view is shown in Fig. 2B. Here, the same
272   calculation of the relationship between polar amplification and the ratio of GMST/deep ocean
273   temperature as in Fig. 1A is shown, except with a GMST-dependent bias of SST during the season
274   of deep water formation of 1/5°C (i.e., for each 5°C increase in GMST, the temperature of the
275   surface ocean in the regions of deep water formation is biased by 1°C below that of mean annual
276   high latitude SST). Comparing the recomputed contours to the same estimates of GMST for three
277   key early Cenozoic intervals demonstrates that a relatively modest increasing seasonal bias in deep
278   water formation is sufficient to bring these observations of tropical versus high latitude warming
279   almost exactly in line with a GMST/deep ocean temperature ratio of 1. The implication of this is
280   that, in absolute terms, the necessary seasonal bias in the temperature of the surface ocean in the
281   regions of deep water formation is minor. For example, the EECO was characterised by a GMST
282   ~15°C warmer than pre-industrial, thus requiring only a ~3°C difference between the temperature
283   of subducting deep water and that of mean annual high latitude SST.

284   Determining the real-world applicability and magnitude of the two mechanisms listed above
285   requires robust observational evidence for the relationship between deep ocean and GMST
286   throughout the Cenozoic and a physical mechanistic basis. In order to provide this, we tackle the
287   following four questions in the remainder of this contribution: 1) Do we know the Cenozoic
288   evolution of deep ocean temperature sufficiently well for it to find utility as a proxy for GMST? (Sec.
289   3.1), 2) What is the empirical slope of the relationship between deep ocean temperature and
290   GMST? (Sec. 3.2), 3) Are fully coupled climate models characterised by a similar relationship, and
291   what can model data tell us about the mechanistic basis for the deep ocean-GMST relationship?
292   (Sec. 3.3), and 4) Should the approach of H13 be revised in light of this analysis, and if so, how?
293   (Sec. 4). In each case, we introduce the relevant methodology and datasets at the beginning of the
294   section.

295

296   3. Results and discussion
297   3.1 How well do we know the Cenozoic evolution of deep ocean temperature?

298   Three key proxy methodologies exist for the temperature of the deep ocean (Evans, 2021), all of
299   which are based on the geochemistry of the shells of calcifying benthic foraminifera: i) the
300   magnesium to calcium ratio (Mg/Ca), ii) the oxygen isotopic composition, and clumped isotopic
301   composition of the shell. Each of these has benefits/limitations summarised briefly in the
302   supplementary materials (Text S2). It is important to note that all three proxies have nonthermal
303   controls or present analytical challenges, particularly the extent to which secular changes in the

304   seawater Mg/Ca ratio impact the Mg/Ca proxy (e.g. Evans and Müller, 2012; Lear et al., 2015), the
305   portioning of measured $\delta^{18}O$ changes between temperature, ice volume, and possible seawater pH
306   (e.g. Meckler et al., 2022; Raymo et al., 2018; Rohling et al., 2021), and the relatively large analytical
307   uncertainty and sample size requirements of Δ_{47} (e.g. Affek, 2012; Meckler et al., 2022).

308   A compilation of deep ocean temperature data (T_d) derived from these three independent
309   techniques is shown in Fig. 1A. Overall, the datasets are characterised by a remarkable degree of
310   consonance, with all three delineating a long-term cooling trend through the Cenozoic of ~15°C
311   since the EECO and being characterised by several structural similarities, such as the relatively rapid
312   cooling across the Eocene-Oligocene Transition (EOT) and after the Miocene Climactic Optimum
313   (MCO) visible in all or most of the records. On the other hand, major discrepancies exist, most
314   notably that the clumped isotope-derived temperature record is substantially warmer than either
315   of the other proxies throughout much of the Cenozoic (by up to 7°C) and has structure in the early
316   Eocene that is virtually entirely absent in the other proxy data (Meckler et al., 2022). While the
317   $\delta^{18}O$-derived deep ocean temperature record is based on multiple assumptions (Eq. 1.1-1.3), the
318   presence of a previously unidentified deep ocean cooling of ~10°C in the earliest Eocene would
319   have major implications for our interpretation of traditional $\delta^{18}O$ datasets and our understanding
320   of the Cenozoic evolution of $\delta^{18}O_{sw}$ if it is globally representative. Meckler et al. (2022) argue that
321   $\delta^{18}O_b$ may be driven by coincident temperature variation and density-driven changes in $\delta^{18}O_{sw}$,
322   while the overall much warmer Δ_{47} temperatures may imply previously unidentified long-term
323   $\delta^{18}O_{sw}$ shifts related to (e.g.) climactically-driven changes in groundwater storage such that the
324   canonical assumption of bulk ocean $\delta^{18}O_{sw}$ = -1‰ in an ice-free world (Zachos et al., 1994) may
325   often not apply. In contrast, the Mg/Ca-derived record is not systematically offset from either of
326   the other proxies across the Cenozoic, showing good agreement with the clumped isotope data in
327   the Oligocene and Neogene (Fig. 1A) but suggesting substantially cooler temperatures than the
328   clumped isotope record during the mid-late Eocene. Mg/Ca-derived temperature is additionally
329   substantially warmer than that based on $\delta^{18}O$ in the Paleocene, which could, for example, be driven
330   by the lack of an accurate seawater Mg/Ca record for this interval. It is also important to note that
331   the Mg/Ca and $\delta^{18}O$ datasets are characterised by very different resolutions, such that the apparent
332   agreement between the two proxies during the PETM is an artefact of the way the Mg/Ca record
333   has been smoothed.

334   While it is beyond the scope of this contribution to reconcile all the aforementioned proxy-proxy
335   offsets, reconstructing GMST from deep ocean proxy data with a useful accuracy is contingent on
336   understanding these discrepancies, such that this issue deserves urgent attention. In Sec. 4 we
337   show, as a starting point, how the majority of the Cenozoic Δ_{47} and $\delta^{18}O$ data may be reconciled
338   with each other, therefore also bringing $\delta^{18}O$ and Mg/Ca into agreement during the Neogene. We
339   also note that, notwithstanding the importance of the Mg/Ca data compilation and analysis by
340   Cramer et al. (2011), several aspects of that data analysis require revision, particularly in light of new
341   information regarding the Cenozoic evolution of seawater Mg/Ca and updated benthic
342   foraminiferal Mg/Ca temperature calibrations (Evans et al., 2018; Lear et al., 2015); revisiting the
343   Paleogene portion of the Mg/Ca dataset with these advances in mind may help to resolve the

344   Eocene Mg/Ca-Δ_{47} offset. Irrespective, the central hypothesis to be tested here is that GMST and T_d
345   are characterised by a linear 1:1 relationship. Fortunately, doing so is reasonably insensitive to the
346   deep ocean temperature proxy discrepancies (Fig. 1) because the method of H13 anchors this 1:1
347   relationship to the base of the Pliocene (Eq. 3). Given that the clumped isotope record is warmer
348   than the $\delta^{18}O$ transformation (Eq. 1.1-1.3) throughout most of the Cenozoic, the outcome of
349   assessing this relationship empirically via the combination of deep ocean and surface temperature
350   datasets (Sec. 3.2) does not greatly depend on the choice of deep ocean proxy data.

351

352   3.2 Empirical evidence for the relationship between deep ocean and global mean surface
353   temperature

354   In order to assess whether quantitative, independent proxy data support the notion of a linear, 1:1
355   relationship between T_d and GMST prior to the Pliocene, we combine the deep ocean temperature
356   records described in Sec. 3.1 (Fig. 1A) with curated data compilations from five well-studied
357   intervals: the last glacial maximum (LGM), Pliocene (mid-Piacenzian) Warm Period (PWP), and three
358   early Cenozoic warm intervals described in Sec. 2 and Fig. 2 (the LP, PETM, and EECO).

359   Global mean surface temperature for each interval is based on the following datasets: the LGM
360   GMST used here is that of the data-model assimilation exercise of Osman et al. (2021), who
361   estimated a LGM-pre industrial (PI) ΔT of 7.0±1.0°C, i.e. LGM GMST = 6.9°C based on a PI GMST =
362   13.9°C. The GMSST of the PWP (mid-Piacenzian; 3.264-3.025 Ma) was estimated using the surface
363   area-weighted mean of the PRISM3/4 SST dataset (Dowsett et al., 2016, 2013), which is 18.7°C.
364   Alternatively using the alkenone-only GMSST reconstruction of McClymont et al. (2020) yields
365   17.2°C for a narrower interglacial (KM5c, 3.2 Ma) and would therefore result in pre-Pliocene
366   reconstructions 1.5°C lower when using this as an anchor. The early Cenozoic GMS(S)T estimates
367   were taken from the DeepMIP data compilation and associated GMST analysis (Hollis et al., 2019;
368   Inglis et al., 2020), with GMST/GMSST estimated in five different ways using surface ocean and
369   terrestrial temperature estimates. In addition, we provide a new estimate based on a combination
370   of SST data from the mid/low latitudes and deep ocean temperature as a proxy for high latitude
371   SST avoiding potential seasonal bias (Evans et al., 2018; Hollis et al., 2012; Inglis et al., 2020), with
372   GMST equal to the surface area-weighted mean in each latitudinal band (0-30, 30-65, >65°). In this
373   latter case, the estimates of deep ocean and GMSST are not fully independent of each other,
374   although we note that they agree well with four other GMS(S)T calculation techniques that are
375   independent of the deep ocean temperature data (Inglis et al., 2020). These GMS(S)T estimates are
376   compared to $\delta^{18}O$ and Mg/Ca-derived T_d in Fig. 3, calculated using the mean of all data within the
377   sampled interval in all cases, using the H13 transformation equations in the case of $\delta^{18}O$ (Fig. 1B)
378   and Eq. 7a of Cramer et al. (2011) in the case of Mg/Ca. The only exception to this is the LGM, for
379   which we use the deep ocean temperature estimate of Adkins et al. (2002). An estimate of the
380   relationship between T_d and GMST based on clumped isotope deep ocean temperatures is not
381   given here as a result of the sparsity of data in certain key intervals, but note that a comparison
382   between Δ_{47}-derived T_d and GMST is given by Goudsmit-Harzevoort et al. (2023).

383 The data compilation shown in Fig. 3 unavoidably combines GMST (PI, LGM, some early Cenozoic
384 estimates) and GMSST reconstructions (PWP, some early Cenozoic estimates), such that caution is
385 required in extrapolating between them. For this reason, the data compilations described above
386 are compared to the T_d-GMST relationship of H13 (Eqs 1.1-3) as well as a data-derived T_d-GMSST
387 relationship (Gaskell et al., 2022). Indeed, coupled climate models consistently predict that GMST
388 and GMSST diverge at GMST $<\sim$20-25°C (Haywood et al., 2020; Lunt et al., 2021; Valdes et al.,
389 2021) but are broadly similar above this, discussed in more detail in Sec. 3.3. This is in agreement
390 with a comparison of the proxy-based GMSST analysis of Gaskell et al. (2022) with H13, which
391 suggests coincident GMST/GMSST at very high GMST and a divergence of the two below $\sim$20°C
392 (Fig. 1A; 3A). If correct, this suggests that the Pliocene-Eocene portion of the analysis shown in Fig.
393 3 should be limited to the relationship between T_d and GMSST, given that the PWP dataset
394 contains only SST estimates, or that the 1:1 line anchored to the Pliocene should be translated
395 down the y axis by several °C. In the absence, to our knowledge, of a true PWP GMST estimate
396 based on a comprehensive terrestrial and marine proxy data compilation, and to avoid
397 complications and uncertainties associated with correcting GMSST to GMST (see Sec. 4) we initially
398 approach the proxy data analysis with both possibilities in mind.

399 Oxygen isotope-based deep ocean temperature (Fig. 3A): Anchoring a 1:1 GMST-T_d relationship to
400 the mid-Piacenzian δ^{18}O-derived deep ocean and PRISM SST datasets defines a 17.2°C offset
401 between the two and thus a LP, PETM, and EECO GMST of 28.4, 30.8, and 35.4°C respectively
402 (given by the y axis location of the black dashed line at the respective T_d for these intervals). These
403 estimates are substantially warmer than the majority of the independent GMST estimates for the
404 EECO and PETM with the exception of the surface-area weighted SST-derived estimate of this
405 study (solid green EECO datapoint in Fig. 3A), and the majority of the LP estimates, which fall within
406 2°C of the 1:1 line. In contrast, the transformation equations of H13 result in early Paleogene GMST
407 estimates $\sim$3°C cooler than the 1:1 line anchored to the PWP, and overall excellent agreement
408 between the DeepMIP database GMST and T_d-derived estimates (red line in Fig. 3A). As discussed
409 above, a likely reason for this is that the Pliocene anchor represents PWP GMSST rather than
410 GMST, such that earlier Cenozoic GMST derived from this may be overestimates (black dashed line
411 in Fig. 3A). The agreement between the T_d-GMSST relationship of Gaskell et al. (2022) and the PWP
412 data point (blue line and red data point in Fig. 3A) adds support to this caveat. Either way, minor
413 discrepancies exist, for example the EECO GMSST estimate of this study (solid green symbol; Fig. 3)
414 is $\sim$3°C warmer than the T_d-GMSST relationship of Gaskell et al. (2022). Understanding whether or
415 not these offsets imply (e.g.) a state-dependent GMST-T_d relationship remains challenging given
416 the certainty with which deep-time GMST can be independently estimated from surface proxy
417 datasets (Anagnostou et al., 2020; Inglis et al., 2020) and should be the subject of future research
418 and data compilation efforts. Nonetheless, the proxy data analysis shown in Fig. 3A is consistent
419 with both the GMSST-T_d and GMST-T_d relationships (Gaskell et al. 2022; H13), especially in the latter
420 case if the PWP data point is considered to represent GMSST and thus overestimates GMST.

421 As an aside, we note that the LGM GMST estimate of Osman et al. (2021) is cooler than that
422 predicted by the δ^{18}O transformation of H13, requiring a steeper Pleistocene GMST-T_d slope (Eq.

423   2.1), which directly follows from the greater Holocene-LGM ΔT than that utilised by H13 (7 cf.
424   4.5°C).

425   Mg/Ca-based deep ocean temperature (Fig. 3B): Compared to using the δ^{18}O transformation
426   equations of H13, the Mg/Ca-derived T_d-GMST relationship differs principally in that the Mg/Ca
427   PWP T_d is ~3°C warmer, whereas the early Paleogene deep ocean temperatures are broadly not
428   (Fig. 1A). This has the effect of shifting the 1:1 GMST-T_d relationship to higher T_d for a given GMST,
429   bringing the EECO and PETM GMST estimates, especially those of this study, into excellent
430   agreement with the deep ocean-based estimate (Fig. 3B). In contrast, the LP is offset from the 1:1
431   line by >5°C, which is driven by the very high Mg/Ca deep ocean temperatures in the late
432   Paleocene compared to those derived from δ^{18}O, in contrast to the EECO, where the two proxies
433   are in good overall agreement. The unexpected nature of the structure of the Mg/Ca temperatures
434   in the earliest Cenozoic suggests that this is probably an artefact of the Mg/Ca data or
435   transformation, given that in this analysis the LP and PETM have similar T_d but were clearly
436   characterised by very different climate states (Dunkley Jones et al., 2013; Penman et al., 2014). The
437   reason for this is likely routed in either the seawater Mg/Ca correction (very little data exist for the
438   Paleocene, none of which was available at the time these Mg/Ca deep ocean temperatures were
439   calculated (see Cramer et al., 2011; Evans et al., 2018b; Gothmann et al., 2015)), or suggests a
440   diagenetic issue with the Paleocene deep ocean Mg/Ca data. Assuming the PWP GMST is an
441   overestimate (see discussion above) would alternatively suggest that T_d underestimates PETM
442   (±EECO) GMST as the red data point and black dashed line anchored to it in Fig. 3B would shift
443   down the y axis, possibly arguing for a relationship between GMST and T_d with a slope >1.
444   Alternatively viewing the PWP datapoint as GMSST and comparing to the early Paleogene GMSST
445   estimates (this study; solid symbols in Fig. 3B) would constrain a pre-Pliocene T_d-GMSST slope of
446   ~1, substantially steeper than that derived by Gaskell et al. (2022). That is, the Mg/Ca-derived T_d
447   analysis cannot be fully reconciled with both the transformation equations of H13 and Gaskell et al.
448   (2022), although the δ^{18}O-derived relationship of that latter study crosses the 1:1 line at a GMSST
449   approximately equidistant between the PWP and EECO, such that the datasets may nonetheless fall
450   within uncertainty of each other.

451   In conclusion, irrespective of which deep ocean temperature dataset is used, and whether the
452   surface temperature estimates based on the data compilations utilised here are considered to
453   represent GMST or GMSST (or both, in warmer climate states), the above data analysis is consistent
454   with the notion of an approximate 1:1 relationship between T_d and GMST, as proposed by Hansen
455   et al. (2013). In addition, reframing parts of this analysis in terms of GMSST provides support for a
456   T_d-GMSST slope of 0.73 (Gaskell et al., 2022; see the coloured solid data points in Fig. 3A).

457

458   3.3 Constraints from fully coupled climate models

459   In section 3.2 we show that there is good empirical evidence that the central assumption of the
460   pre-Pliocene GMST estimate of H13 is a reasonable approximation. To mechanistically understand

461 why this is the case, we interrogate the output of two sets of Paleogene coupled climate model
462 datasets: 1) the DeepMIP set of model simulations (Lunt et al., 2021, 2017), which incorporates eight
463 climate models run under different pCO_2 but otherwise similar (early Paleogene) boundary
464 conditions between models, and identical boundary conditions within a set of model simulations,
465 plus associated PI controls, and 2) the Cenozoic portion of the Phanerozoic HadCM3 simulations
466 (Valdes et al., 2021), which includes 12 simulations (1 per Stage) with varying palaeogeography and
467 other boundary conditions, run under two (broadly similar) pCO_2 within each time slice. These
468 latter simulations were not part of DeepMIP. Both sets of model output are described in detail
469 elsewhere, including: the experimental design and rationale (Lunt et al., 2021; Valdes et al., 2021,
470 and references therein), the degree to which the deep ocean has reached equilibrium (Zhang et al.,
471 2022), the location(s) of deep water formation (Zhang et al., 2022), and the spatial heterogeneity in
472 modelled deep ocean temperature (Goudsmit-Harzevoort et al., 2023; typically <1°C). Those
473 analyses are not repeated here, with the exception of the key regions of deep-water formation,
474 which we interrogate when comparing high latitude SST to T_d, and the relationship between GMST
475 and T_d, which was explored in detail by Goudsmit-Harzevoort et al. (2023), but is expanded upon
476 here and compared in detail to the simulations of Valdes et al. (2021). The majority of the
477 simulations in both sets are considered to have a reached a reasonable degree of equilibrium with
478 respect to the deep ocean (<1°C drift in the volume-integrated mean ocean temperature; Valdes et
479 al. (2021)), with the exception of the 9× CO_2 CESM simulation (~1.5°C in the final kyr; Fig. S2 of
480 Zhang et al. (2022)). This latter data point is clearly an outlier (see below), and while we include it in
481 all relevant figures, we exclude it from any regression analysis on the basis that deep ocean
482 temperature is likely underestimated in this simulation. In all cases in the discussion below we
483 define T_d as equal to the mean of all temperature data below 3000 m.

484 The two sets of model simulations have their own advantages and limitations. The DeepMIP output
485 allows the role of CO_2 to be more readily separated from other factors, as all other boundary
486 conditions were held constant with the exception of the PI controls. Conversely, key boundary
487 conditions such as paleogeography were modified for each time slice in the HadCM3L simulations
488 of Valdes et al. (2021), making it more challenging to pinpoint the factors driving a given output,
489 but arguably makes them a better test of the relationship between T_d and GMST, given that (e.g.)
490 the Cenozoic paleogeographic changes are incorporated.

491 3.3.1   Model deep ocean temperature versus GMS(S)T

492 As shown by Goudsmit-Harzevoort et al. (2023), the DeepMIP simulations are characterised by a
493 T_d-GMST slope close to 1 based on the entire ensemble (excluding the 9× CO_2 CESM simulation
494 and PI controls), with m = 1.026 (Fig. 4A). Anchoring a 1:1 line to the mean of the lowest CO_2
495 simulation for each model with Eocene paleogeography (in order to broadly follow the assumption
496 of Hansen et al., 2013) demonstrates that all simulations fall within ±2°C, with the majority falling
497 within ±1°C (mean average error = 0.82°C). Thus, a 1:1 relationship between T_d and GMST appears
498 to be a robust assumption based on both proxy data (Sec. 3.2) and climate models. The HadCM3L
499 simulations with variable boundary conditions (hereafter HadCM3L$_{V21}$ to distinguish them from

500   those performed as part of DeepMIP) paint a similar picture, with most falling with ±2°C of a 1:1
501   line anchored to the mean of the two simulations at 3 Ma (Fig. 4C). The exception to this are the
502   Miocene simulations, which are characterised by GMST 0-3°C warmer than PI but mean deep
503   ocean temperatures up to 2°C cooler, irrespective of which CO_2 scenario is used. This yields an
504   overall T_d-GMST slope much lower than the DeepMIP simulations (m = 0.715), although excluding
505   these from the analysis results in a slope much closer to unity (m = 0.858). The HadCM3L$_{V21}$
506   Miocene anomaly appears to be driven by a shift in the dominant region of deep-water formation
507   from the North Atlantic to the Southern Ocean compared to the PI control, resulting in cooler deep
508   ocean temperatures. This is likely caused by salinity-driven changes in density driving a slowdown
509   in N. Atlantic overturning such that deep water formation shifts to the fresher but colder Southern
510   Ocean, yet CO_2 is insufficiently high to drive increases in GMST. Conversely, T_d is higher in the
511   HadCM3L$_{V21}$ Pliocene simulations because the N. Atlantic remains the dominant region of deep
512   water formation. Of the 86 simulations that we consider in our analysis, it is only the six HadCM3$_{V21}$
513   from the Miocene that show a substantial deviation from a 1:1 T_d-GMST relationship, such that
514   climate models with both constant and temporally varying boundary conditions run at different
515   pCO_2 overall provide strong evidence in favour of the 1:1 T_d-GMST hypothesis. Nonetheless, given
516   that the HadCM3L$_{V21}$ simulations are arguably a better test of whether T_d and GMST are directly
517   related throughout the Cenozoic (as opposed to when pCO_2 changes within a given interval), as
518   paleogeography related phenomena are capable of driving large changes in GMST independent of
519   pCO_2 (Caballero and Huber, 2013), the possible breakdown of this relationship in the Miocene is a
520   key target for future research. That there is limited proxy evidence for a similar Miocene and PI
521   deep ocean temperature (Fig. 1A), possibly suggests an issue with the Miocene model data, and
522   highlights the need for further Miocene pCO_2 estimates (Valdes et al., 2021).

523   Both sets of model simulations are characterised by a GMSST-T_d and GMSST-GMST relationship
524   with a slope substantially lower than 1, with m = 0.870 and 0.456 in the DeepMIP and HadCM3L$_{V21}$
525   simulations respectively (GMSST-T_d; Fig. 4) and m increasing to 0.587 in the latter case if the
526   Miocene simulations are excluded. These slopes bracket the empirical ($\delta^{18}O$-derived) relationship of
527   Gaskell et al. (2022), which has m = 0.73, with the DeepMIP suite of models conspicuous in being
528   consistently characterised by a steeper relationship and GMSST consistently (~2-3°C) warmer than
529   the data suggest for a given T_d at pCO_2 >2× PI (Fig. 4B). While the HadCM3L$_{V21}$ dataset is
530   characterised by a lower slope than the proxy data-based estimate, almost all simulations remain
531   similarly offset to higher GMSST. In general, the model-data GMSST-T_d disagreement is likely
532   driven by the model difficulty in capturing the magnitude of polar amplification implied by the
533   proxy data (e.g. Evans et al., 2018b; Kiehl and Shields, 2013; Lunt et al., 2012; Sagoo et al., 2013;
534   Hollis et al., 2019; Lunt et al., 2021), which results in a lower degree of high latitude and therefore
535   deep ocean warming for a given global GMSST increase, rather than an issue with the analysis of
536   Gaskell et al. (2022). However, we note that this cannot explain the good agreement between the
537   early Cenozoic HadCM3L$_{V21}$ simulations with Gaskell et al. (2022) (Fig. 4D).

538
539

540   3.3.2   Mechanistic basis for a ~1:1 deep ocean-GMST relationship

541   Both model and proxy data are in overall agreement that GMST and the temperature of the deep
542   ocean are linearly related with a slope close to 1, supporting one of the central assumptions of
543   Hansen et al. (2013). We next revisit the two key reasons that this relationship might emerge,
544   outlined in Sec. 2, despite polar amplification and the stratification of the ocean in warm climate
545   states, using both the DeepMIP and HadCM3L$_{V21}$ simulations. That is: 1) whether deep water
546   formation becomes increasingly seasonally biased in warm climate states, and 2) if a faster rate of
547   land versus ocean warming counteracts the effect of polar amplification to result in a 1:1 T$_d$-GMST
548   relationship.

549   Is deep water formation seasonally biased in warm climate states? In order to address this
550   question, we examine the relationship between T$_d$ and high latitude SST (SST$_{HL}$) in the regions of
551   deep-water formation. The sensitivity of this analysis to three different methodologies was
552   explored, in which SST$_{HL}$ was variously calculated as:

553   1. Mean annual SST in all regions of the ocean likely to be important for deep-water
554   formation, that is, the N. Atlantic, N. Pacific, and Atlantic, Pacific, and Indian sectors of the
555   Southern Ocean. See Zhang et al. (2022) and Fig. 5 for the definition of these boxes.
556   2. Winter SST in only the boxes relevant for deep-water formation, on a simulation-specific
557   basis. We define winter as June, July, and August in all southern hemisphere boxes, January
558   and February in the N. Pacific, and March, April, and May in the N. Atlantic to reflect that
559   maximum overturning occurs shortly after winter in that region (Wang et al., 2021). The
560   choice of relevant box(es) for a given simulation was based on the region of maximum
561   mixed layer depth, detailed in Tab. S1. In cases where this is ambiguous, such as the 9× CO$_2$
562   CESM simulation (Fig. 5), we use the boxes from the closest CO$_2$ simulation which clearly
563   shows deep water formation as evidenced by the mixed layer depth (mld).
564   3. The SST in the grid cells that have a simulation-specific mld at least 90% as deep as the
565   global seasonal maximum, only considering grid cells >50°N/S (Valdes et al., 2021). T$_d$ was
566   compared to both mean annual and winter SST in the mean of all grid cells meeting these
567   criteria.

568   In all cases, SST is based on the mean of all relevant grid cells averaged over 0-100 m. A sensitivity
569   analysis was performed to understand the impact of in/excluding the Arctic, which may at times
570   have been disproportionately fresher and warmer than other high latitude regions (Brinkhuis et al.,
571   2006). Doing so has no significant impact on any aspect of the data analysis presented below.

572   The results of the first two of these analyses is shown in Figs 6 and 7 for the DeepMIP and
573   HadCM3L$_{V21}$ simulations respectively, in all cases anchored to i) the simulation with the lowest pCO$_2$
574   but Eocene paleogeography in the case of the DeepMIP simulations, as the closest possible
575   representation of Pliocene-like conditions, and ii) to the Pliocene in the case of HadCM3$_{V21}$, again,
576   because the principal aim of this study is to test the pre-Pliocene T$_d$-GMST assumption of Hansen
577   et al. (2013). Note that while a similar interrogation of the relationship between deep ocean

578   temperature and GMST was performed by Goudsmit-Harzevoort et al. (2023), the analysis

579   presented here differs in that we interrogate all simulations by determining the relative change

580   from the Pliocene or that with the lowest CO_2 and paleo boundary conditions, although we replot

581   the data in absolute terms for comparison (e.g. Fig. 6C,F). This is an important distinction, given

582   that it is the *relative* change in T_d and GMST prior to the base of the Pliocene that we are

583   principally interested in, and doing so avoids potential bias derived from model-specific skill in

584   capturing past climate states.

585   In the case of the DeepMIP models, mean annual SST (MASST) in all high latitude boxes rises at a

586   faster rate than GMSST in response to CO_2 (ensemble m = 0.88; Fig. 6A), as expected given

587   enhanced poleward heat transport in warm climate states (Kelemen et al., 2023). Limiting the

588   analysis to include only winter SST in the relevant deep water boxes (Fig. 5; Tab. S1) is a simple

589   method of removing the summer bias in SST_{HL} in the comparison. Indeed, doing so results in an

590   increase in the whole ensemble GMSST-SST_{HL} slope (m = 0.96; Fig. 6B), albeit with a greater degree

591   of variance around the least squares regression (the uncertainty in the slope increases from 0.018

592   to 0.038). A similar picture emerges when assessing the change in T_d as a function of SST_{HL} (Fig.

593   6D,E), which is overall characterised by a slope close to 1 (m = 1.05) such that $\Delta SST_{HL} \approx \Delta T_d$,

594   implying that changes in deep ocean temperature are directly coupled to the high latitude surface

595   ocean in all simulations. As before, an increased gradient (to m = 1.13) and variance is observed

596   when winter SST in the filtered high latitude boxes is used (0.034 versus 0.074), implying slightly

597   worse agreement between SST_{HL} and T_d in absolute terms across the ensemble (Fig. 6F). A faster

598   rate of temperature increase in the deep ocean than high latitude SST (Fig. 6E) is physically

599   implausible and implies a limitation of this simple analysis (e.g. the inclusion of grid cells in the high

600   latitude boxes (Fig. 5) that do not contribute to overturning), or that a portion of deep water is

601   formed outside of the months and/or boxes considered here, explored in more detail below.

602   Overall, however, this analysis must mean that seasonally biased deep water formation is important

603   in offsetting T_d from mean annual SST_{HL}, strongly arguing for this as a mechanistic cause of a 1:1 T_d-

604   GMST relationship.

605   In the case of the HadCM3L$_{V21}$ simulations, the results of the analysis is broadly similar in that

606   accounting for the seasonality and relevant location of deep water formation results in a ΔGMST-

607   ΔSST_{HL} slope marginally close to 1 compared to a similar analysis based on MASST in all high

608   latitude boxes (Fig. 7A,B). While this adds further model-based support for a 1:1 relationship

609   between GMST and SST_{HL}, these simulations differ from the DeepMIP set in that the unfiltered

610   analysis (Fig. 7A) is characterised by a slope marginally >1, i.e. GMSST broadly increases at a faster

611   rate than SST_{HL}. This is not the case for the HadCM3 DeepMIP simulations, highlighting the

612   potential impact of palaeogeography over CO_2 alone, although we also note that the version of

613   HadCM3L utilised to produce the DeepMIP simulations is one of several models less able to

614   capture the proxy data-derived degree of Eocene polar amplification in several deep-time warm

615   intervals (Burls et al., 2021; Evans et al., 2018; Lunt et al., 2021). The HadCM3L$_{V21}$ simulations are also

616   characterised by a T_d-SST_{HL} slope >1, irrespective of whether the relevant boxes and season of

617   deep water formation are taken into account (Fig. 7D,E), and absolute deep ocean temperature

618   substantially offset from winter SST in the boxes relevant for deep water formation (Fig. 7F). This is
619   also not the case for the DeepMIP HadCM3 simulations and is mostly easily explicable a limitation
620   of our analysis, likely suggesting the inclusion of cooler high latitude grid cells that are not relevant
621   for deep water formation and/or that there is an important contribution to deep water formation
622   beyond the core winter months.

623   Both the DeepMIP and HadCM3L$_{V21}$ simulations overall provide strong evidence that deep ocean
624   temperature, GMSST, and high latitude SST are tightly, linearly linked to each other, especially
625   when a winter season bias in deep water formation is accounted for. While the degree of
626   divergence from 1:1 high latitude seasonal SST-T_d and GMSST-T_d relationship is small in both sets
627   of simulations (within ±5°C in almost all cases, e.g. Fig. 6B,E; Fig. 7B, E), we explore how much of
628   the remaining variance in the data analysis is a result of the approach of averaging data across
629   large high latitude ocean boxes (Fig. 5), which is the simplifying approach of both our analysis thus
630   far and previous studies (Goudsmit-Harzevoort et al., 2023; Valdes et al., 2021; Zhang et al., 2022).
631   To assess this, Fig. 8 shows the same analysis of the DeepMIP simulations as in Fig. 6, except that
632   only grid cells with a mixed layer depth (mld) within 90% of the global seasonal maxima were
633   considered (approach 3 above), thus avoiding possible bias from the inclusion of grid cells within
634   the broad boxes shown in Fig. 5 that are not relevant for deep water formation. For example, in the
635   3×CO_2 CESM simulation, this approach excludes coastal and more northerly grid cells in the
636   Weddell Sea, as well as the eastern portion of the Indian Ocean sector of the Southern Ocean box
637   (Fig. 5).

638   This more nuanced analysis highlights a stronger degree of polar amplification in both mean
639   annual and seasonal SST in grid squares with the deepest seasonal mld relative to the all grid cells
640   within the high latitude boxes discussed previously. For example, there is a shallower slope
641   between ΔSST_{HL} and $\Delta GMST$ in the model ensemble relationship between GMSST and mean
642   annual SST_{HL}, which is characterised by a slope that is reduced to 0.81±0.05 in the mld analysis
643   compared to m = 0.88±0.02 when using all grid cells in the high latitude boxes (cf. Fig. 8A, Fig. 6A).
644   A similar reduction in slope is observed when comparing the seasonal SST in the grid cells
645   characterised by the deepest mld and limiting the analysis to include only grid cells in the
646   hemisphere in which deep water formation dominantly occurred, with a reduction in slope from
647   0.96±0.04 to 0.89±0.04 (Fig. 8B, Fig. 6B; with CESM being a notable exception). Therefore, winter
648   SST in these grid cells warms faster than global average MASST, which means that overall, while an
649   increasing seasonal bias in deep water formation is a key process that results in an approximate 1:1
650   T_d-GMSST relationship (all DeepMIP simulations fall within ±5% of the 1:1 line; Fig. 8B), a polar
651   amplification signal is present in subducting water, such that a seasonal bias in deep water
652   formation cannot be the only process resulting in a 1:1 GMST-T_d relationship (Fig. 4A).
653   Unsurprisingly, the mld analysis fully resolves the discrepancy between SST_{HL} and T_d observed
654   previously (Fig. 6E). Considering only the relevant grid squares results in a slope of 0.99±0.07 (Fig.
655   8D), i.e., the two parameters have the same value.

656 The above analysis demonstrates that an increasing seasonal bias in deep water formation offsets
657 polar amplification in the regions of deep water formation to a substantial degree, but is, alone,
658 insufficient to mechanistically explain a 1:1 T_d-GMST relationship. The other key factor, as noted by
659 Hansen et al. (2013) and Goudsmit-Harzevoort et al. (2023) is that land surface air temperature are
660 more sensitive to CO_2 than SST. This is the case to a greater degree in the HadCM3L$_{V21}$ simulations
661 than the DeepMIP ensemble (GMSST versus GMST m = 0.84 and 0.72 respectively, Fig. 4E,F),
662 implying that non-CO_2 boundary conditions can be important in modulating this slope given that
663 isolating the DeepMIP HadCM3 and HadCM3L simulations demonstrates that these are
664 characterised by a steeper slope than HadCM3L$_{V21}$ (m = 0.795 versus 0.716). Nonetheless, in the
665 case of the DeepMIP ensemble, the slopes between ΔGMSST and ΔSST$_{HL}$ (accounting for
666 seasonality in the relevant high latitude boxes) and GMSST-GMST are almost identical (m = 0.89
667 and 0.84, respectively; Fig. 8B, Fig. 4E), which is the explanation for a 1:1 T_d-GMST relationship as
668 hypothesised by H13 and the reason that this is an emergent model property (Fig. 4A).

669 We stress that while previous analyses have reached the same or similar conclusions (Goudsmit-
670 Harzevoort et al., 2023), the key assumption of H13 was that this is the case in the earlier Cenozoic
671 when GMST and T_d are *anchored to the base of the Pliocene*, i.e. prior to the strong modulation of
672 this relationship by ice sheet growth and ice-sheet climate feedbacks. Thus, while it is strongly
673 encouraging that this conclusion has now independently been reached several times, it is only by
674 performing the analysis in the way presented here, and assessing whether it is the case in model
675 simulations with both variable palaeogeography and constant palaeogeography but variable $p\mathrm{CO}_2$
676 that we can mechanistically understand whether or not this relationship is likely to have been the
677 case for the entirety of Cenozoic prior to ~5 Ma.

678

679 4   Reformulation of the deep ocean-GMST relationship

680 Climate model simulations and proxy data are in remarkable full agreement (within uncertainty)
681 that the temperature of the deep ocean and GMST are characterised by a 1:1 relationship prior to
682 the Pliocene. Crucially, given that this is empirically the case, this lends strong support to the notion
683 that a reconstruction of deep ocean temperature is a reliable proxy of GMST irrespective of
684 whether the model-derived mechanistic basis for this relationship discussed in Sec. 3.3.2 is correct.
685 Nonetheless, the GMST approach of H13 requires revision, particularly in light of our greatly
686 improved understanding of i) the Cenozoic evolution of T_d, including a more thorough grasp of the
687 nonthermal controls on some key temperature proxies, and ii) the Cenozoic evolution of
688 continental ice volume, i.e., that the assumption of ice-free conditions before the base of the
689 Pliocene is no longer tenable (Lear et al., 2015; Leutert et al., 2021; Rohling et al., 2022). Focusing on
690 the benthic oxygen isotope stack ($\delta^{18}O_b$) because the temporal resolution of the dataset is
691 unparalleled by the other proxy datasets (Fig. 1), we explore whether $\delta^{18}O_b$-derived temperatures
692 can be reconciled with constraints from clumped isotope analysis of benthic foraminifera, and how
693 our improved understanding of sea level variation impacts GMST estimates based on these data. In
694 order to do so, we i) revisit the ice volume/sea level component of $\delta^{18}O_b$ using the analysis of

695   Rohling et al. (2022), ii) explore the impact of a pH/[CO_3^{2-}] correction on $\delta^{18}O_b$, and iii) rescale the
696   resulting deep sea temperature record using three intervals with reasonable constraints on both
697   deep ocean and GMST (the LGM, present-day, and PWP).

698   *Sea level*: An extremely comprehensive analysis of the sea level/ice volume contribution to $\delta^{18}O_b$ is
699   available (Rohling et al., 2022), which uses a process-based model to determine the nonlinear
700   relationship between $\delta^{18}O_b$ and sea level. This nonlinearity largely results from the relationship
701   between GMST and ice volume (e.g., the absence of ice above a certain GMST) and the change in
702   mean ice sheet $\delta^{18}O$ as a function of total ice volume (Rohling et al., 2022, 2021; Spratt and Lisiecki,
703   2016). Here, we use the median of the boot-strapped Monte Carlo results of the preferred process-
704   based model of Rohling et al. (2022), from which we derive a sea level-free $\delta^{18}O_b$ record by
705   converting the deep sea temperature of that study back to $\delta^{18}O_b$ simply by dividing by -0.25‰°C^{-1}.
706   As Rohling et al. (2022) studied the interval 0-41 Ma, we extend the record to the entirety of the
707   Cenozoic by appending the remainder of the $\delta^{18}O_b$ record of Westerhold et al. (2020), assuming no
708   ice volume contribution to $\delta^{18}O_b$ before 41 Ma (e.g. Scotese et al., 2021).

709   *pH effect on $\delta^{18}O_b$*: A seawater carbonate chemistry effect on $\delta^{18}O_b$ has been found for both species
710   of foraminifera (*Orbulina universa* and *Globigerina bulloides*) for which sufficient data are available
711   to make an assessment (Bijma et al., 1999; Spero et al., 1997), as well as coccolithophores and
712   calcareous dinoflagellates (Ziveri et al., 2012), and inorganic calcite (McCrea, 1950). While there is
713   no direct evidence for a similar impact on the oxygen isotopic composition of the shells of benthic
714   foraminifera, we advocate for a correction because it is a ubiquitous feature of all calcitic plankton
715   studied so far, and has a strong basis in theory, being rooted in the pH-dependent speciation of
716   dissolved inorganic carbon (Zeebe, 1999). We nonetheless stress that correcting $\delta^{18}O_b$ data for past
717   changes in seawater carbonate chemistry remains fraught with uncertainty because the slope of
718   the relationship strongly differs between both foraminifera species studied thus far (by a factor of
719   ~2), which has a large impact on the resulting correction when considering large whole-ocean
720   changes in pH (Evans et al., 2016). Here, we explore a correction using the theoretical slope
721   between pH and $\delta^{18}O$ across the pH range 7-9 (Eq. 2 of Zeebe (1999)), which covers the possible
722   range of past ocean pH variation, across which a linear approximation suffices. Doing so yields a
723   pH-$\delta^{18}O$ slope of -1.50‰ per pH unit (see supplementary Text S3), which is intermediate between
724   the two planktonic foraminifera species for which data are available (-0.89 and -2.51‰ per pH
725   unit). While this control on $\delta^{18}O$ is often thought of as a carbonate ion effect (e.g. Gaskell et al.,
726   2022; Spero et al., 1997), it is more appropriately mechanistically ascribed to pH because this is, in
727   effect, the dominant control on seawater [HCO_3^-]/[CO_3^{2-}], and the effect of seawater carbonate
728   chemistry on carbonate $\delta^{18}O$ occurs via the differential fractionation factor between water and
729   these DIC species (Zeebe, 1999). Parameterising the seawater carbonate chemistry effect on $\delta^{18}O$ as
730   a function of pH additionally has the advantage that a direct proxy for seawater pH is available
731   from measurements of the boron isotopic composition of foraminifera (Anagnostou et al., 2020;
732   Foster and Rae, 2016; Hönisch et al., 2012; Penman et al., 2014). In order to apply a pH correction,
733   we fit a smoothing spline to the benthic foraminifera $\delta^{11}B$-derived pH record (Greenop et al., 2014;
734   Meckler et al., 2022). The available benthic foraminifera-derived pH dataset is low-resolution (31

735 datapoints spanning the last ~60 Ma) and contains no data between the mid-Eocene and Miocene.
736 As such, our T_d and derived GMST record will require revision as more data become available, and
737 it is possible or likely that the details of the records presented here contain artefacts related to the
738 long-term smooth applied to the pH data.

739 The resulting Cenozoic T_d reconstruction is shown in Fig. 9A, with that of Hansen et al. (2013) and
740 independent estimates of deep ocean temperature for comparison (Fig. 1A). Notwithstanding the
741 uncertainties in the approach driven by the sparsity of deep ocean pH data, pH correcting $\delta^{18}O_b$
742 following the ice volume deconvolution of Rohling et al. (2022) results in a Cenozoic $\delta^{18}O_b$-derived
743 T_d record which agrees well with the majority of the clumped isotope data, such that this revised
744 analysis of the $\delta^{18}O$ data resolves much of the pre-existing discrepancy between the $\delta^{18}O$ and Δ_{47}
745 records (cf. Meckler et al. (2022); Westerhold et al. (2020)), especially when the $\Delta_{47}/\delta^{18}O$ reanalysis
746 of Daëron and Gray (2023) is considered (Fig. 9A). The magnitude of the Cenozoic T_d decrease
747 between the EECO and late Pleistocene (~17°C) is indistinguishable between the proxies, with
748 major discrepancies remaining only in the early Eocene, with a transient cooling event constrained
749 by Δ_{47} but not $\delta^{18}O$, and in the Miocene, wherein the majority of the Δ_{47} data are ~5°C warmer
750 than the $\delta^{18}O$-derived record presented here. This latter discrepancy either implies that the
751 clumped data record regionally warmer-than-global temperature at ODP Site 761 (where the
752 majority of the Δ_{47} data come from in this interval, NW Australian margin, see the discussion in
753 Evans (2021) and Modestou et al. (2020)), or that mid-Miocene deep ocean pH and/or sea level are
754 substantially overestimated. Further work is of course required to understand whether the Δ_{47} data
755 from this interval is a truly global signal and to determine the cause of the remaining discrepancies
756 within the Eocene portion of the datasets .

757 We convert the $\delta^{18}O$-derived T_d record into GMST using a similar approach to H13, by splitting the
758 Cenozoic into three intervals characterised by overall different relationships between T_d and GMST,
759 namely, the Plio-Pleistocene intervals with T_d cooler than modern, the Plio-Pleistocene with T_d
760 warmer than modern, and the remainder of the Cenozoic before, in this case, the PWP. Doing so
761 requires three tie points at which GMST is well characterised and the assumption that the
762 relationship between T_d and GMST remains constant within each of these portions of the dataset.
763 As tie points, we use the 20th Century GMST (13.9°C, e.g. Trenberth and Fasullo (2013)), the LGM
764 GMST analysis of Osman et al. (2021), and the mid-Piacenzian PRISM4 GMSST (Dowsett et al., 2016,
765 2013), coupled with the minimum T_d of the last 25 ka and mean T_d of the interval 3.00-3.05 Ma.
766 Here, the switch between LGM and Pliocene scaling occurs at a $\Delta T_d = 0$ based on the 1 ka T_d
767 datapoint. The accuracy of this analysis obviously depends on the quality of the GMS(S)T data, all
768 of which are based on large, independent proxy data compilations or observations. In the case of
769 the mid-Piacenzian, we unavoidably use a GMSST rather than a GMST estimate to anchor the
770 Cenozoic GMST reconstruction, because there is, to our knowledge, currently no curated data
771 compilation on which such an estimate could be based. However, we note that both the
772 HadCM3L$_{V21}$ and DeepMIP suite of climate models (Fig. 4E,F) fall onto a single emergent GMST-
773 GMSST relationship, which we use to convert Pliocene GMSST to GMST. Performing this calculation
774 using HadCM3L$_{V21}$ yields a mid-Piacenzian GMST = 15.0°C, preferred here because these

775   simulations have variable palaeogeography. However, the exercise is overall insensitive to this

776   choice; alternatively using the DeepMIP set of simulations would result in pre-Pliocene GMST

777   0.45°C cooler. Together, this exercise defines $\Delta GMST/\Delta T_d$ scaling factors of 2.2, 1.3, and 1.0 for the

778   Plio-Pleistocene cooler and warmer than present, and pre-3.025 Ma, respectively.

779   The resulting GMST estimates, both with and without a pH correction on $\delta^{18}O$, are shown in Fig. 9B.

780   This updated Cenozoic GMST reconstruction constrains the magnitude of cooling from the EECO

781   (53-50 Ma) to the 21st Century to 17.3°C, of which 45% occurs during the Eocene, 20% across the

782   Eocene-Oligocene Transition, 20% during the Miocene, and 15% during the Plio-Pleistocene. The

783   pH correction on $\delta^{18}O_b$ exerts a strong control on Palaeogene-Miocene GMST reconstructed in this

784   way, e.g. elevating EECO GMST by ~5°C. Understanding whether or not this correction should be

785   applied is clearly an urgent priority, and if so, as is the production of a high-resolution deep ocean

786   pH record. More broadly, placing an uncertainty estimate on T_d-derived GMST is challenging

787   because the pre-Pliocene scaling factor is an assumption that has not been derived from any

788   specific dataset. While we show that it appears to be a good assumption, the independent proxy

789   data and associated transformations are currently insufficiently error-free to place further

790   constraint on the scaling factor (Fig. 3) while the climate model simulations interrogated here

791   suggest that it lies between 1.03 (DeepMIP) and 0.86 (HadCM3L$_{V21}$), see Fig. 4. Notwithstanding the

792   potential pitfalls in using the model derived slopes to constrain uncertainty in the approach overall,

793   we apply a ±2°C uncertainty to the T_d and GMST reconstructions shown in Fig. 9, which is the

794   approximate difference that would result in the Eocene between a scaling factor of 0.86 and 1.03,

795   but again highlight that further systematic bias is possible, particularly related to the pH correction

796   outlined above.

797   The main differences between our GMST reconstruction (Fig. 9) and that of H13 derive from i) the

798   revision of the LGM-modern $\Delta GMST$ from 4.5 to 7°C by Osman et al. (2021), resulting in a

799   substantially greater glacial-interglacial GMST change in the late Pleistocene, and ii) from the pH

800   correction on $\delta^{18}O$, resulting in substantially warmer Palaeogene GMST as a result (average EECO

801   GMST of ~27°C in H13 compared to ~31.°C in this study (Fig. 9B)). The revised Cenozoic GMST

802   reconstruction agrees well with several independent lines of evidence from both models and proxy

803   data (Fig. 10). For example, the Pliocene Model Intercomparison Project range ($\Delta GMST$ = 1.8-5.2°C

804   relative to the pre-industrial era; Haywood et al. (2020)) covers the deep ocean-derived maxima for

805   this interval (2.5°C). Our results are also in reasonable agreement with the model-informed

806   Oligocene GMST estimates of O'Brien et al. (2020), with the latter offset to values ~0-3°C higher

807   (Fig. 10). In addition, we observe excellent agreement between the data compilation-derived

808   GMSST estimates of Ring et al. (2022) coupled with our assessment of deep ocean temperature

809   and the T_d-GMSST relationship of Gaskell et al. (2022), see Fig. 10. However, our GMST estimates

810   are broadly substantially cooler for much of the Neogene, and warmer during the early Paleogene

811   than those of Ring et al. (2022). We note that the GMST and GMSST estimates of that study are

812   broadly similar, which is at odds with modern observations and climate model simulations which

813   require substantially warmer GMSST than GMST in cooler climate states (Fig. 4E,F), with the two

814   converging only under early Paleogene-like global warmth. This discrepancy potentially points to a

815   systematic bias in the terrestrial proxy records (also discussed in Ring et al. (2022)), which warrants

816   further investigation, rather than a failure of climate models to capture this feature of Earth's

817   climate.

818   In the early Paleogene, the PETM and pre-PETM GMST estimates of Tierney et al. (2022) of 34.1°C

819   (33.1-35.5) and 28.5°C (27.5-30.1°C) are within uncertainty of this study (35.2 and 28.1°C,

820   respectively; note that the resolution of the core-PETM $\delta^{18}O$ data in the stack utilised here is

821   insufficient to place a precise estimate on the PETM using this approach). Our analysis constrains

822   EECO GMST to 31.3±1.3°C, slightly higher than the upper range of the estimates provided in Inglis

823   et al. (2020) but within uncertainty of both the data analysis and 6× CO_2 CESM simulation of Zhu et

824   al. (2019). While there are important aspects of the deep ocean-derived estimates that require

825   further research (see above), if correct, this would also constrain 'bulk' equilibrium climate

826   sensitivity at the upper end of the range reported in Inglis et al. (2020), i.e. ~5°C, in agreement with

827   the LP-PETM-derived ECS estimate of Tierney et al. (2022). The offset of this study as well as that of

828   Tierney et al. (2022) and Zhu et al. (2019) compared to Inglis et al. (2020) is likely driven in large

829   part by the inclusion of a substantial amount of terrestrial temperature data which may be cool

830   biased in several of the approaches included in that latter study (also compare our early Eocene

831   GMST estimate to that of Ring et al. (2022)). Support for this is provided by the fact that our T_d-

832   derived GMST is in excellent agreement with the DeepMIP database-derived estimate of coeval

833   GMSST of this study and the T_d-GMSST analysis of Gaskell et al. (2022), see Fig. 10, potentially

834   implying an issue with some of the terrestrial but not the ocean datasets in that database. This is

835   encouraging because both all model simulations considered here and our T_d-GMST analysis

836   coupled with the T_d-GMSST analysis of Gaskell et al. (2022) suggest a convergence, within ~1-3°C,

837   of GMST and GMSST under early Palaeogene-like climate states (Fig. 4E,F; 10). As such, cooler

838   EECO GMST estimates (Fig. 10) are difficult to reconcile with our data-derived GMSST, and/or

839   require this aspect of the climate model output to be inaccurate. We alternatively argue that the

840   consistency between this study, Gaskell et al. (2022), and this emergent property of climate models

841   provides strong support for the very warm EECO GMST presented here.

842

843   5   Conclusions

844   Here, we interrogate the use of a deep ocean temperature (T_d) record to infer global mean surface

845   temperature (GMST) in detail, using both curated data compilations and two sets of climate model

846   simulations. In particular, we address the question of whether T_d and GMST are linearly related with

847   a slope of 1, as previously hypothesised (Hansen et al., 2013). We show that no such relationship

848   would be expected in a world characterised by polar amplification, because i) the high latitude

849   regions warm faster than the global mean, and ii) deep water is thought to have formed at high

850   latitudes throughout most, if not all, of the Cenozoic. However, proxy data compilations of the

851   Pliocene and early Palaeogene fall within uncertainty of a 1:1 T_d-GMST relationship, suggesting that

852   (some) other process(es) act to balance polar amplification. Using both the DeepMIP set of

853   simulations (Lunt et al., 2021) with varying CO_2 and fixed palaeogeography and a set of Cenozoic

854   HadCM3L simulations with covarying palaeogeography and CO_2 (Valdes et al., 2021), we show that

855   these processes are: 1) an increasing seasonal bias in deep water formation as GMST increases, and

856   2) the fact that the land surface warms at a faster rate than the ocean surface (see also Goudsmit-

857   Harzevoort et al. (2023)). While this provides a mechanistic basis for a 1:1 T_d-GMST relationship

858   prior to the (mid)Pliocene, we note that some HadCM3 simulations do not adhere to this, with

859   GMST underestimated by up to 3°C during the Miocene (Fig. 4C). This occurs when a relatively

860   small CO_2 change is sufficient to shift the principal locus of deep-water formation without a large

861   associated change in GMST, resulting, in these simulations, in a cooler deep ocean as North

862   Atlantic overturning ceases to be an important source of deep water. Although there is, to our

863   knowledge, no direct evidence for this scenario in the Cenozoic, it highlights that there is at least

864   the potential for substantial deviations from a 1:1 T_d-GMST relationship, particularly in deeper time.

865   More broadly, we stress that our key finding – that T_d-GMST is characterised by a 1:1 relationship

866   prior to the Pliocene within the certainty of the proxy data records – is robust irrespective of the

867   mechanism, and further work will of course be required to empirically determine whether the

868   above causal processes inferred from climate model simulations were indeed responsible.

869   Our contribution substantially strengthens the notion that GMST may be simply calculated from

870   that of the deep ocean with a useful degree of precision. However, we highlight that recent

871   advances in proxy methodologies for deep ocean temperature have arguably increased the

872   uncertainty in our knowledge of T_d itself, particularly in the early Cenozoic. Specifically, clumped

873   isotope-derived palaeotemperatures are substantially warmer and more variable than our

874   canonical understanding of benthic foraminiferal $\delta^{18}O$ and/or the Cenozoic evolution of $\delta^{18}O_{sw}$.

875   Solving this issue is clearly an urgent priority, although we show that the majority of the

876   discrepancy can be explained by a seawater carbonate chemistry (pH) effect on $\delta^{18}O$ (Fig. 9A), as

877   also suggested by Meckler et al. (2022).

878   Using the pH-corrected $\delta^{18}O$ data, we construct a revised Cenozoic GMST record, broadly

879   following the methodology of Hansen et al. (2013), but incorporating advances in our

880   understanding of LGM and Pliocene GMST as well as the evolution of sea level throughout the past

881   40 Ma (Dowsett et al., 2016; Osman et al., 2021; Rohling et al., 2022). Our GMST record is warmer

882   throughout much of the Cenozoic, with large (up to ~5°C) differences present prior to the

883   Miocene. Overall, these estimates are in excellent agreement with several independent early

884   Cenozoic GMST reconstructions, adding confidence to all of these various lines of evidence. Finally,

885   we constrain the magnitude of the Cenozoic GMST decrease to 17.3°C (EECO to 20th Century), and

886   EECO GMST to 31.3±1.3°C, slightly above the upper end of previous reconstructions. If correct, this

887   would support the notion that 'bulk' equilibrium climate sensitivity of was higher than modern in

888   this past warm climate state (Tierney et al., 2022).

889

Data availability statement

No new datasets are presented here. The climate model output interrogated here is available from https://www.deepmip.org/data/ and https://www.paleo.bristol.ac.uk/ummodel/scripts/papers/Valdes_et_al_2021.html (Lunt et al., 2021; Valdes et al., 2021). They key datasets used in the data analysis performed here are the WOA2018 ocean temperature dataset (https://www.ncei.noaa.gov/products/world-ocean-atlas (Locarnini et al., 2018)), the DeepMIP proxy database (Hollis et al., 2019), Cenozoic benthic foraminifera $\delta^{18}O$, Mg/Ca, Δ_{47}, $\delta^{11}B$, and sea level/temperature deconvolution (Cramer et al., 2011; Leutert et al., 2021; Meckler et al., 2022; Modestou et al., 2020; Rohling et al., 2022; Westerhold et al., 2020), and mid-Piacenzian SST synthesis (Dowsett et al., 2016). The python and Matlab scripts used to perform the data analysis presented here, and to produce all figures, can be found at https://zenodo.org/xxxxxxxxx [provided as a supplementary zip during review].
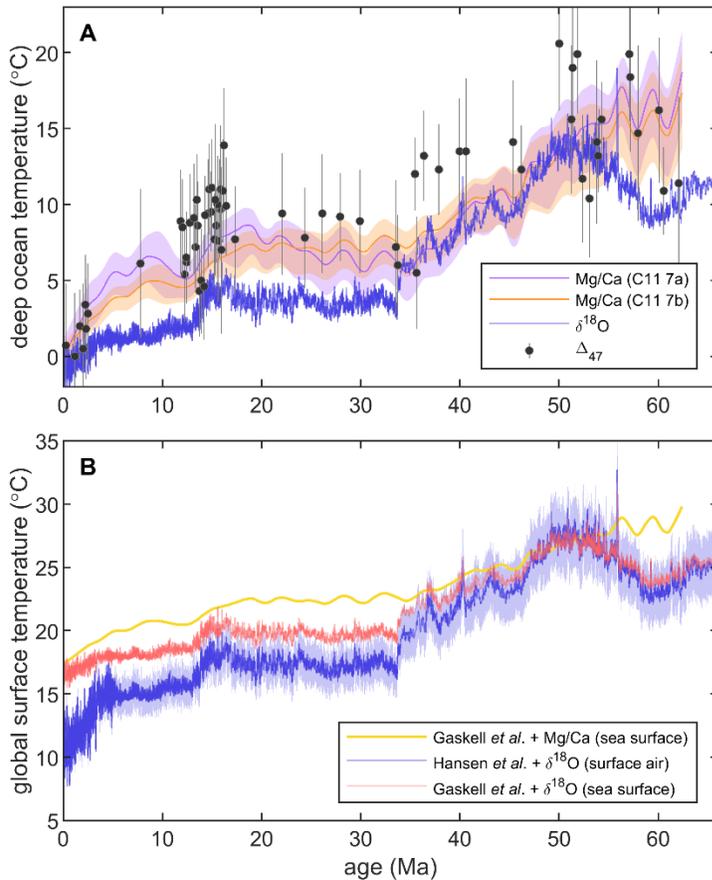
Fig. 1. (A) Deep ocean temperature based on the oxygen isotope composition of foraminifera (Westerhold et al., 2020) converted to temperature following the approach of Hansen et al. (2013), the Mg to Ca ratio of foraminifera (Cramer et al., 2011), and the clumped isotopic composition of foraminifera (Δ_{47}; Leutert et al., 2021; Meckler et al., 2022; Modestou et al., 2020). In the case of Mg/Ca, the two transformation equations refer to the two Mg/Ca-temperature calibrations explored by (Cramer et al., 2011). (B) Global mean surface air temperature (Hansen et al., 2013) and global mean sea surface temperature (Gaskell et al., 2022) calculated according to the methodologies outlined in the original studies except using the revised deep ocean benthic foraminiferal oxygen isotope stack (Westerhold et al., 2020). An arbitrary ±10% uncertainty has been added to the global mean surface air temperature estimate. Note that the deep ocean to surface temperature transformation of (Hansen et al., 2013) is parameterised according to some specific features of the benthic oxygen isotope stack, such that calculating GMST from benthic foraminifera Mg/Ca in a directly comparable way is not possible and was not attempted.
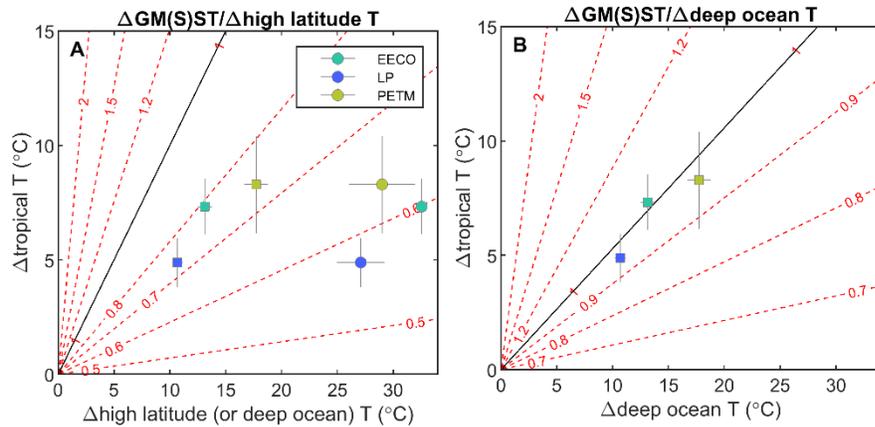
927
928
929 Fig. 2. (A) The idealised relationship between GM(S)ST and high latitude/deep ocean temperature
930 for a given degree of high latitude/deep ocean and tropical warming (contour lines, see text and
931 supplementary Text S1 for details) for a zonally homogeneous waterworld. Assuming deep ocean
932 and high latitude SST are exactly coupled and in the absence of other processes, only in a world
933 characterised by exactly zero polar amplification is a 1:1 relationship between GM(S)ST and high
934 latitude SST expected (black line). Estimates of high latitude and tropical SST change for three
935 DeepMIP target intervals (Inglis et al., 2020) are shown (circles). Given a likely seasonal bias in high
936 latitude proxy SST, high latitude temperature is alternatively calculated for the same intervals using
937 the benthic foraminiferal oxygen isotope stack (squares; see Fig. 1 and the text for methodological
938 details). (B) The same analysis assuming that deep water formation becomes increasingly biased
939 towards winter as high latitude SST increases (specifically, for every 5°C high latitude SST increase,
940 deep ocean temperature is biased by 1°C below the mean annual high latitude average). This has
941 the effect of stretching the contour lines shown in panel A towards higher deep ocean
942 temperatures. Note that the prescribed change in the seasonal bias in deep water formation is
943 entirely without basis; the exercise is intended as a simple illustration of a process that could result
944 in a 1:1 relationship between deep ocean temperature and GM(S)ST.
945
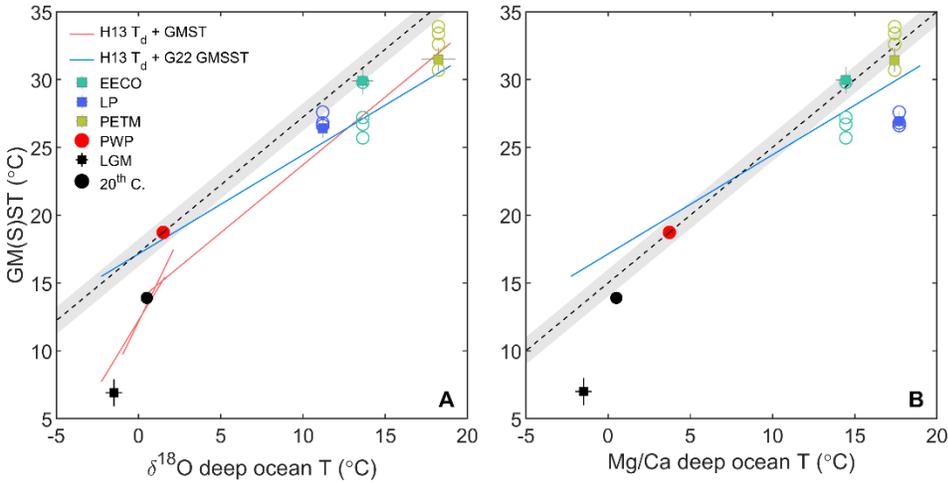
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947
Fig. 3. Empirical estimates of GMS(S)T as a function of deep ocean temperature for key Cenozoic
intervals for which curated data compilation efforts exist (Dowsett et al., 2016; Hollis et al., 2019;
Osman et al., 2021). Estimates of GMST for three DeepMIP target intervals (Inglis et al., 2020) are
shown (open circles, see that study for uncertainties) as well as GMSST (filled squares; this study)
(A) Deep ocean temperature calculated from the benthic foraminifera oxygen isotope stack
following Hansen et al. (2013). (B) Deep ocean temperature calculated from benthic foraminiferal
Mg/Ca following Cramer et al. (2011). LGM and pre-industrial deep ocean temperature is not based
on foraminiferal $\delta^{18}O$/Mg/Ca, see text for details. The grey shaded region depicts a 1:1 increase in
deep ocean and GMST anchored to the Pliocene Warm Period (red circles), with an arbitrary ±1°C
uncertainty. Previously suggested relationships between deep ocean temperature and GMST
(Hansen et al., 2013) and GMSST (Gaskell et al., 2022) are shown with red and blue lines
respectively. Note that: 1) the Paleogene GMST estimates of this study are not fully independent
from estimated deep ocean temperature, because deep ocean temperature was used to estimate
high latitude SST in order to avoid a seasonal bias in the surface proxy data, and 2) the PWP data
point is a GMSST estimate that may overestimate GMST. This may explain the discrepancy between
the Palaeogene estimates from the 1:1 line anchored to the PWP, see text for details.
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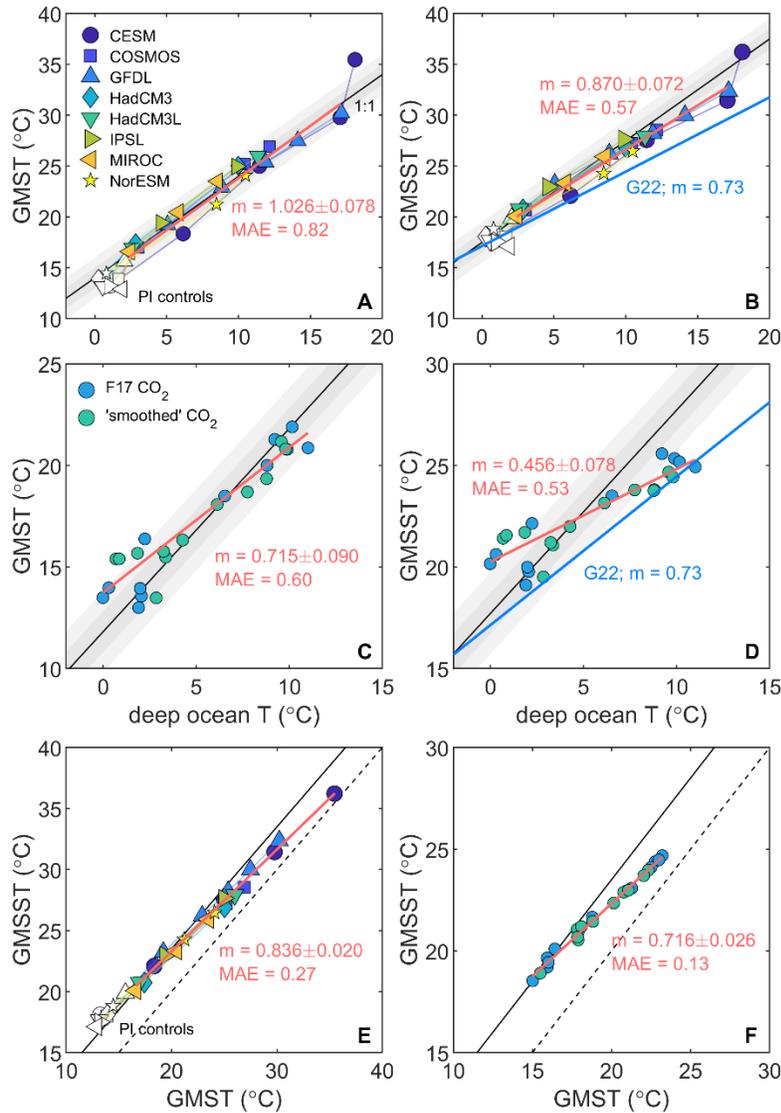
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967 Fig. 4. (A) Global mean surface temperature and (B) Global mean sea surface temperature as a
968 function of deep ocean temperature (>3000 m) in the DeepMIP set of model simulations, see
969 panels C and D for x axis labels. One-to-one lines are anchored to the mean of the simulations
970 conducted at 1×CO_2 and Eocene palaeogeography plus the IPSL simulation at 1.5×CO_2; shaded
971 regions depict ±1 & 2°C from this line. The least squares linear regressions (red lines) include all
972 model simulations with Eocene palaeogeography, except for the 9×CO_2 CESM simulation (see
973 text). (C,D) A similar analysis performed for the Cenozoic HadCM3 simulations of Valdes et al.
974 (2021) with 1:1 lines anchored to the mean of the two 3 Ma simulations. A foraminiferal δ^{18}O-
975 derived relationship between GMSST and deep ocean temperature is shown in panels B and D,
976 calculated following Gaskell et al. (2022). (E,F) The relationship between global mean surface
977 temperature and global mean sea surface temperature in the DeepMIP and HadCM3 simulations
978 of Valdes et al. (2021), respectively. Lines with a slope of 1 are shown anchored as described for the

979   other panels (solid) as well as 1:1 lines (dashed). All model data were interpolated to a 1×1° grid
980   before further calculations were performed.
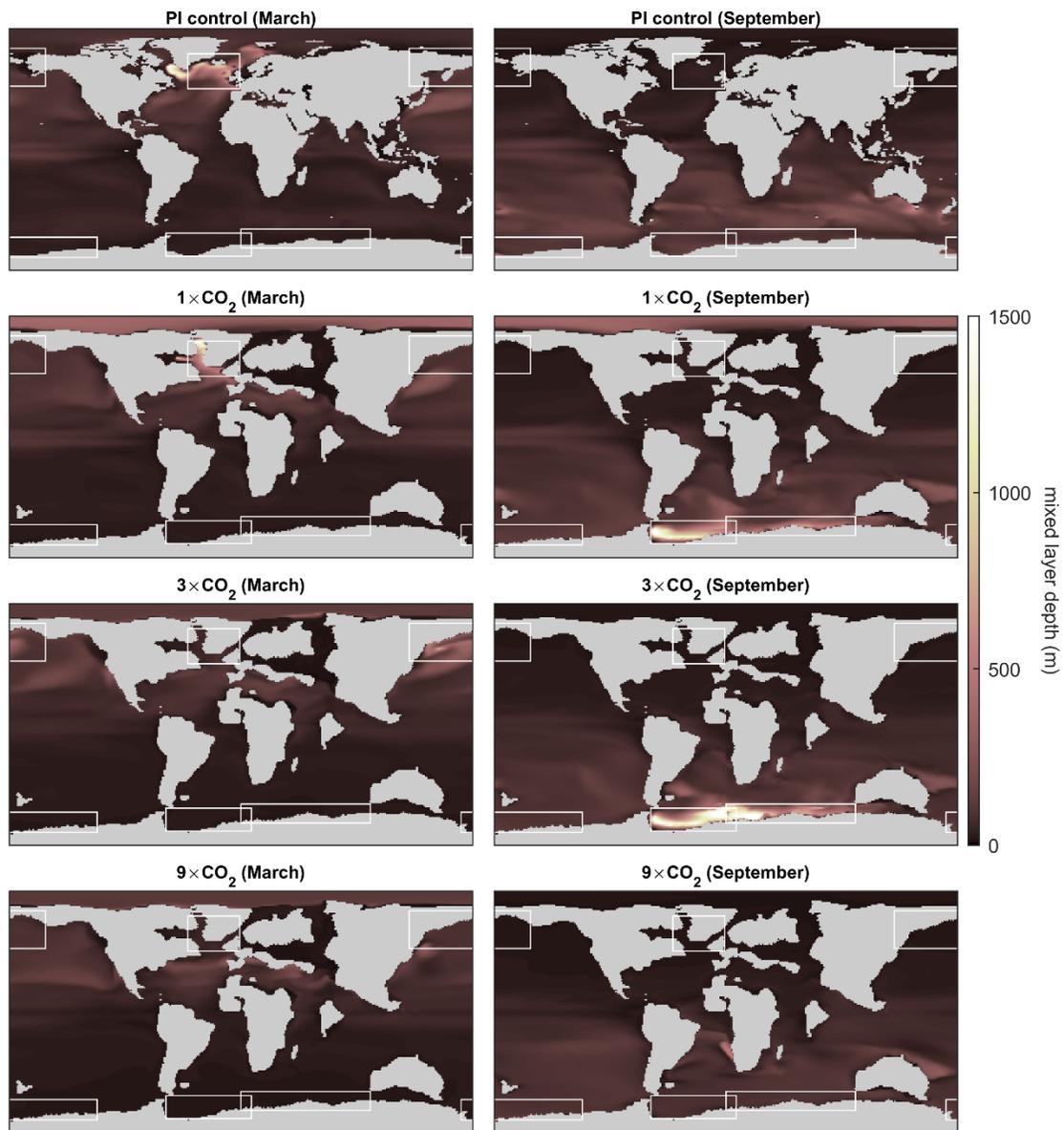
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983   Fig. 5. March and September mixed layer depth in the four DeepMIP CESM simulations (PI control,
984   as well as 1,3, and 9× CO_2). Boxes showing the key regions of deep water formation in the full suite
985   of models contributing to the DeepMIP set of simulations are overlain (Zhang et al., 2022).
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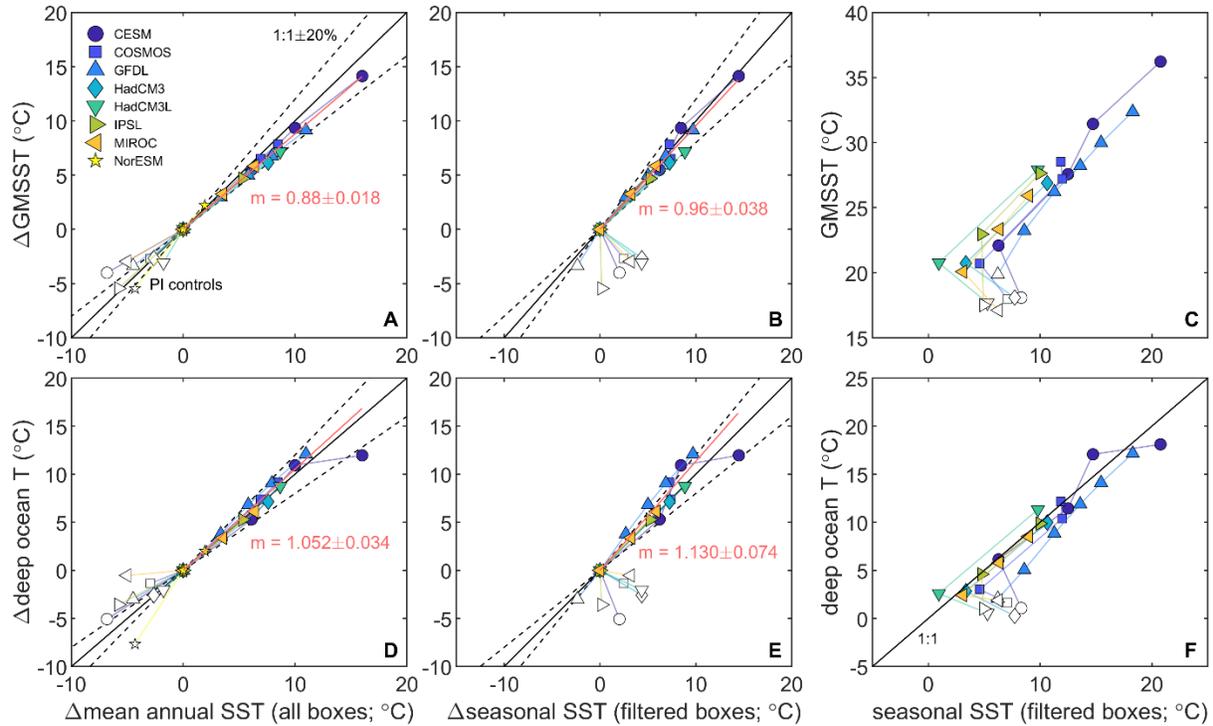
Fig. 6. The relationship between sea surface temperature (0-100 m) in the broad regions of deep-water formation and global mean sea surface temperature (GMSST) or deep ocean temperature in the DeepMIP set of model simulations. (A,D) Mean annual SST in all boxes shown in Fig. 5 plotted relative to the 1×CO_2 simulation with Eocene palaeogeography (except IPSL; 1.5×CO_2). (B,E) Winter SST in the model-specific box(es) relevant for deep water formation. Note that seasonal SST data for NorESM was not available. (C,F) As panel B/E, except in absolute temperature space. In all cases, the least squares linear regressions are forced through the origin and fit to the ensemble, excluding the pre-industrial controls. The 9×CO_2 CESM simulation was excluded from the fit.
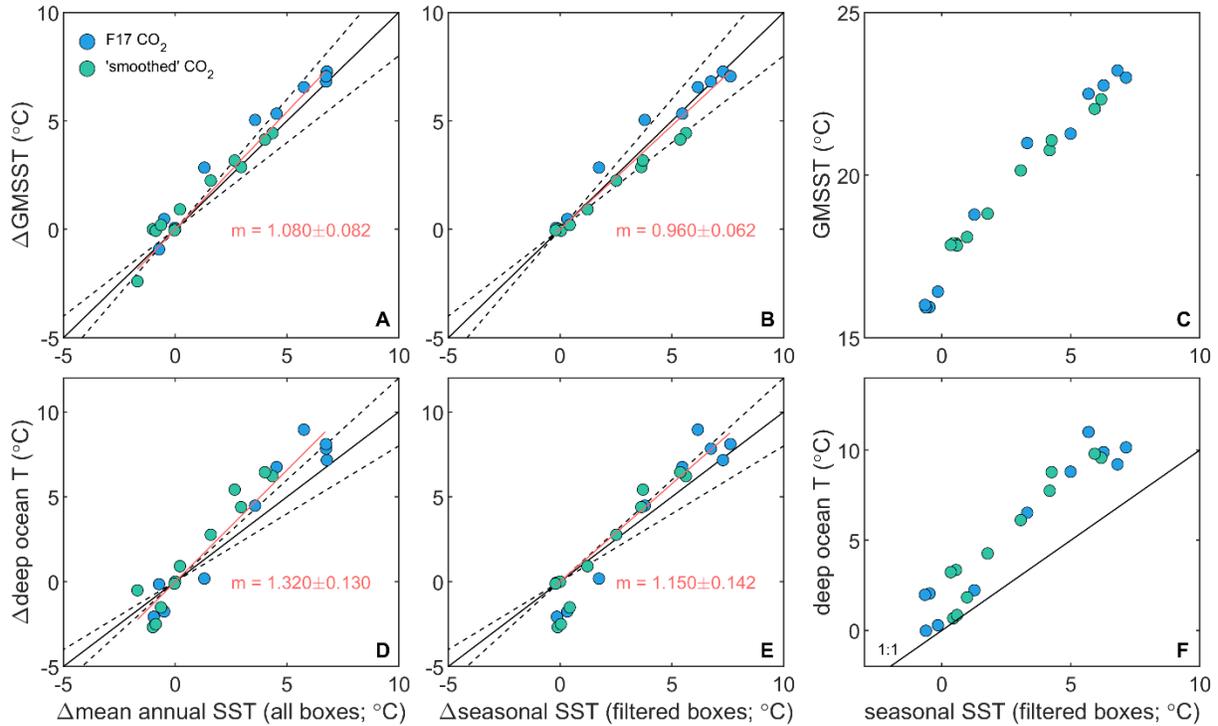
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1000 Fig. 7. The relationship between sea surface temperature (0-100 m) in the broad regions of deep-
1001 water formation and global mean sea surface temperature (GMSST) or deep ocean temperature in
1002 the HadCM3L simulations of Valdes et al. (2021). Two simulations were performed for each time
1003 slice, at two different CO_2. (A,D) Mean annual SST in all boxes shown in Fig. 5 plotted relative to the
1004 3 Ma simulation. (B,E) Winter SST in the model-specific box(es) relevant for deep water formation.
1005 Data from the 0 Ma simulation are not shown as these fall off of the trend as a result of being
1006 characterised by N. Atlantic deep water formation. (C,F) As panel B/E, except in absolute
1007 temperature space. In all cases, the least squares linear regressions are forced through the origin
1008 and fit to all simulations.
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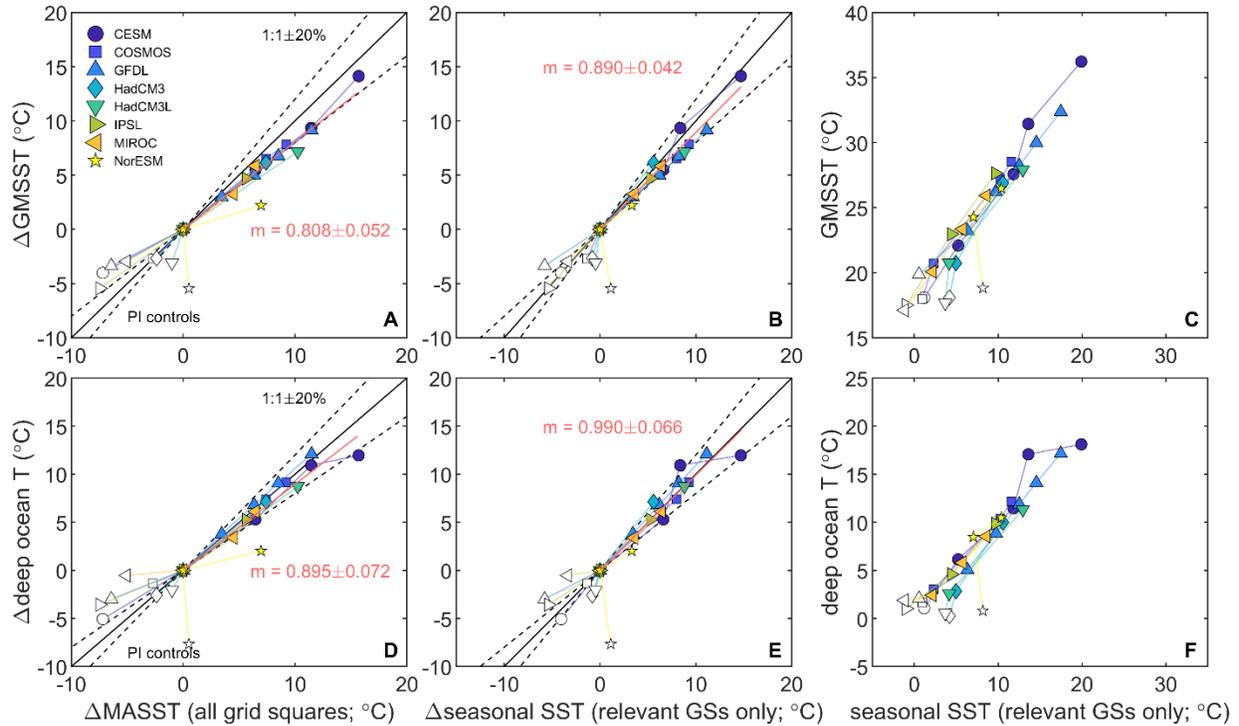
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Fig. 8. The relationship between SST in the model grid cells with a mixed layer depth (mld) at least
90% of the global seasonal maximum and global mean sea surface temperature (GMSST) or deep
ocean temperature in the DeepMIP set of model simulations. (A,D) Mean annual SST in all grid cells
meeting the mld criteria plotted relative to the 1×CO_2 simulation with Eocene palaeogeography
(except IPSL; 1.5×CO_2). (B,E) SST in all grid cells (GSs) meeting the mld criteria during the season of
maximum mixed layer depth, and limited to the model-specific hemisphere(s) relevant for deep
water formation. Note that seasonal SST data for NorESM was not available. (C,F) As panel B/E,
except in absolute temperature space. In all cases, the least squares linear regressions are fit to the
ensemble, excluding the pre-industrial controls. The 6× and 9×CO_2 CESM simulations were
excluded from the fit.

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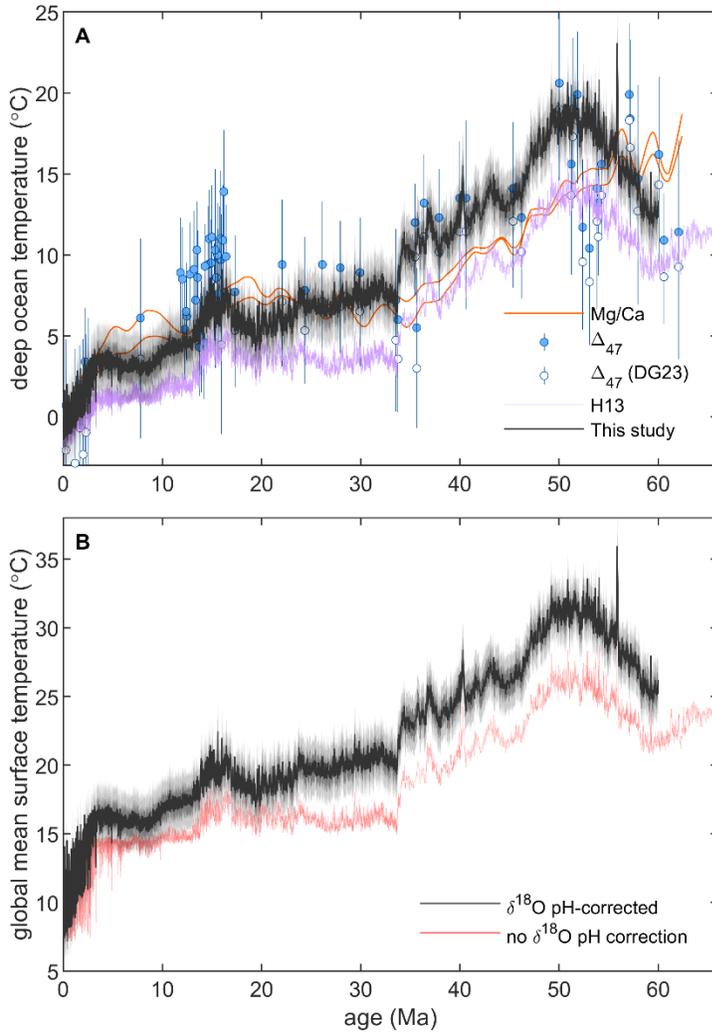
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1025 Figure 9. (A) A revised estimate of the Cenozoic evolution of deep ocean temperature based on
1026 the sea-level and pH-corrected benthic foraminifera oxygen isotope stack (black line with an
1027 arbitrary ±2°C uncertainty, see text for details) in the context of other proxy estimates (see Fig. 1),
1028 including the Δ_{47} reanalysis of Daëron and Gray (2023). (B) GMST based on the deep ocean
1029 temperature record from this study (see text) and the sensitivity of this reconstruction to whether
1030 or not the benthic foraminiferal $\delta^{18}O$ data are pH corrected.
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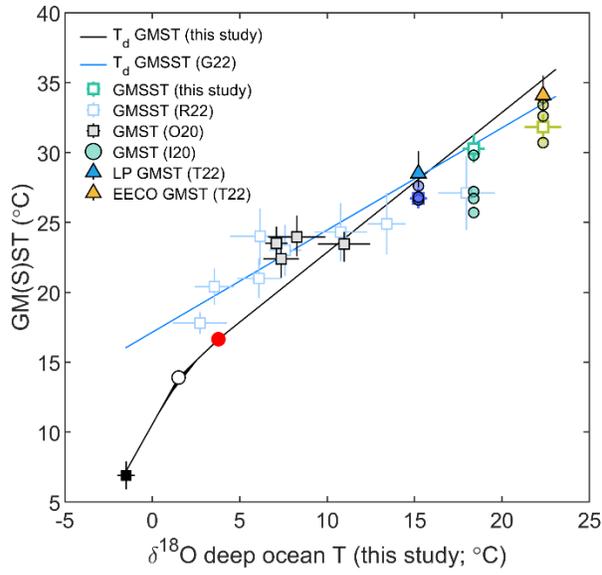
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1034 Figure 10. The revised relationship between deep ocean temperature (T_d) and GMST (black line, this
1035 study), showing the three anchor points used here (LGM, 20th Century, and PWP, black square,
1036 white circle, and red circle respectively; see text for details). Following our analysis and H13, the
1037 relationship between T_d and GMST has a slope of 1 for all climate states warmer than the PWP.
1038 Independent estimates of GMST (Inglis et al., 2020; O'Brien et al., 2020; Tierney et al., 2022 (I20,
1039 O20, T22, respectively) and GMSST (open squares; this study, based on the DeepMIP database
1040 (Hollis et al. 2019) and the revised assessment of T_d (Fig. 9), as well as those of Ring et al. (2022);
1041 R22) are shown. Note that the blue line is the T_d-GMSST relationship of Gaskell et al. (2022) and
1042 not the best fit regression of the estimates from Ring et al. (2022).

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