



1 **Maastrichtian-Rupelian paleoclimates in the southwest Pacific – a critical**
2 **evaluation of biomarker paleothermometry and dinoflagellate cyst**
3 **paleoecology at Ocean Drilling Program Site 1172**

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5 Peter K. Bijl^{1*}, Joost Frieling^{1,2}, Marlow J. Cramwinckel^{1,3}, Christine Boschman¹, Appy
6 Sluijs¹, Francien Peterse¹

7

8 ¹Department of Earth Sciences, Utrecht University, Utrecht, the Netherlands.

9 ²now at: Department of earth Sciences, University of Oxford, South Parks Road OX1
10 3AN, UK

11 ³now at: School of Ocean and Earth Science, National Oceanography Centre
12 Southampton, University of Southampton, Southampton, UK

13

14 * corresponding author. email: p.k.bijl@uu.nl

15

16 **Abstract**

17 Sea surface temperature (SST) reconstructions based on isoprenoid glycerol dialkyl
18 glycerol tetraether (isoGDGT) distributions from the Eocene southwest (sw) Pacific
19 Ocean are unequivocally warmer than can be reconciled with state-of-the-art fully
20 coupled climate models. However, the SST signal preserved in sedimentary archives
21 can be affected by contributions of additional isoGDGT sources. Methods now exist to
22 identify and possibly correct for overprinting effects on the isoGDGT distribution in
23 marine sediments. We here use the current proxy insights to assess the reliability of
24 the isoGDGT-based SST signal in 69 newly analysed and 242 re-analysed sediments
25 ODP Site 1172 (East Tasman Plateau, Australia) following state-of-the-art
26 chromatographic techniques, in context of paleo-environmental and
27 paleoclimatologic reconstructions based on dinoflagellate cysts. The resulting ~130
28 kyr-resolution Maastrichtian-Oligocene TEX₈₆-based SST record confirms previous
29 conclusions of anomalous warmth in the early Eocene sw Pacific and remarkably cool
30 conditions during the mid-Paleocene. Dinocyst diversity and assemblages show a
31 strong response to the local SST evolution, supporting the robustness of the TEX₈₆
32 record.



33 Soil-derived branched GDGTs stored in the same sediments are used to reconstruct
34 mean annual air temperature (MAAT) of the nearby land using the MBT'_{5me} proxy.
35 MAAT is consistently lower than SST during the early Eocene, independent of the
36 calibration chosen. General trends in SST and MAAT are similar, except for: 1) an
37 enigmatic absence of MAAT rise during the Paleocene-Eocene Thermal Maximum and
38 Middle Eocene Climatic Optimum, and 2) a subdued middle-late Eocene MAAT
39 cooling relative to SST. Both dinocysts and GDGT signals suggest a mid-shelf
40 depositional environment with strong river-runoff during the Paleocene-early
41 Eocene, progressively becoming more marine thereafter. This trend reflects gradual
42 drying and more pronounced wet/dry seasons in the northward drifting Australian
43 hinterland, which may also explain the subdued middle Eocene MAAT cooling relative
44 to that of SST. The overall correlation between dinocyst assemblages, marine
45 biodiversity and SST changes suggests that temperature exerted a strong influence on
46 the surface-water ecosystem, probably in part through sea level changes caused by
47 steric effects. Finally, we find support for a potential temperature control on
48 compositional changes of branched glycerol monoalkyl glycerol tetraethers
49 (brGMGTs) in marine sediments. It is encouraging that a critical evaluation of the
50 GDGT signals confirms the vast majority of the generated data is reliable. However,
51 this also implies the high TEX_{86} -based SSTs for the Eocene sw Pacific, and the
52 systematic offset between absolute TEX_{86} -based SST and MBT'_{5me} -based MAAT
53 estimates remain unexplained.

54



55 1. Introduction

56 1.1 The Paleogene Southwest Pacific Ocean

57 Reconstructions of deep-sea (Westerhold et al., 2020) and sea surface
 58 temperature (Bijl et al., 2009; 2013a; Cramwinckel et al., 2018; Frieling et al., 2014;
 59 Hollis et al., 2014; 2019; Inglis et al., 2015; 2020; O'Brien et al., 2017; Evans et al.,
 60 2018; O'Connor et al., 2019; Sluijs et al., 2020) have revealed overall cool climate in
 61 the Maastrichtian and Paleocene, long-term warming towards the early Eocene
 62 Climatic Optimum (EECO; 53.4 – 49.2 Ma), and subsequent cooling during the middle
 63 and late Eocene (48.6 – 33.6 Ma). The EECO stands out as particularly warm, with ice-
 64 free polar regions (Bijl et al., 2013a; Hines et al., 2017; Pross et al., 2012; Frieling et
 65 al., 2014). Certain southern high-latitude regions retain warm-temperate conditions
 66 into the late Eocene (Bijl et al., 2009; Houben et al., 2019) and, despite ample
 67 evidence for pronounced Antarctic glacial expansion across the Eocene-Oligocene
 68 transition (Salamy and Zachos, 1999; Bohaty et al., 2012), even into the early
 69 Oligocene (Hartman et al., 2018; Passchier et al., 2013; 2017; O'Brien et al., 2020).
 70 Variations in atmospheric CO₂ concentrations (Beerling and Royer, 2011; Anagnostou
 71 et al., 2016; Foster et al., 2017) are likely the primary driver of these multi-million-
 72 year climatic trends (Cramwinckel et al., 2018). While equatorial proxy-based sea
 73 surface temperatures (SSTs) and deep-sea temperatures, assumed to reflect high-
 74 latitude SSTs, show good correspondence with numerical model simulations under
 75 Eocene boundary conditions and with varying CO₂ forcing (Cramwinckel et al., 2018),
 76 proxy-based SST reconstructions of the southwest (sw) Pacific remain warmer than
 77 those from model simulations (Bijl et al., 2009; Cramwinckel et al., 2018; Hollis et al.,
 78 2012; Lunt et al., 2021), despite proposed zonal heterogeneity (Douglas et al., 2014).
 79 Specifically, numerical models are currently unable to simulate a paleoclimate in
 80 which the annual SST difference between the equatorial Atlantic Ocean (Cramwinckel
 81 et al. 2018) and the sw Pacific Ocean (Bijl et al., 2013a; Hollis et al., 2012) is as small
 82 as the proxy data suggests. Without this model-data mismatch resolved, it remains
 83 unclear to what extent numerical models properly simulate polar amplification of
 84 Paleogene climates and if the current proxies properly reflect high latitude
 85 temperatures under greenhouse conditions (Lunt et al., 2012).



86 Proxy evidence for warmth in the Eocene sw Pacific region derives from a
87 suite of organic and calcite-based proxies. In terms of the latter, oxygen isotope ratios
88 ($\delta^{18}\text{O}$) and trace element ratios (Mg/Ca) from well-preserved planktonic foraminifera
89 indicate warm temperatures from Eocene sections in New Zealand (Creech et al.,
90 2010; Hines et al., 2017; Hollis et al., 2009; 2012). These proxies require assumptions
91 regarding seawater chemistry (Mg/Ca, $\delta^{18}\text{O}$ seawater, pH) that carry significant
92 uncertainty (e.g., Kozdon et al., 2020, Evans et al. 2018). The application of clumped
93 isotope paleothermometry has great potential to partially alleviate such concerns, as
94 evident from work on Seymour Island (Douglas et al., 2014). Pollen-based vegetation
95 reconstructions from New Zealand, the Tasman region and Wilkes Land (Carpenter et
96 al., 2012; Contreras et al., 2013; 2014; Huurdeman et al., 2020; Pross et al., 2012),
97 however, confirm warm conditions, and arguably deliver the best constraints on
98 winter temperatures. This is because of fundamental physiological restrictions in
99 their individual tolerances (e.g., Reichgelt et al., 2018), whereas mean annual air
100 temperature (MAAT) reconstructions from pollen assemblages are complicated
101 because MAAT exerts much less control on the standing vegetation than seasonal
102 temperature and hydrological extremes. Reconstructions of the warm Eocene
103 primarily relied on organic geochemical proxies, notably TEX_{86} (Bijl et al., 2009;
104 2013a; Cramwinckel et al., 2018; 2020; Crouch et al., 2020; Hollis et al., 2009; 2012;
105 Sluijs et al., 2006; 2009; 2011). These absolute SST estimates for the sw Pacific are
106 closer to those from the equatorial Atlantic as they are to the deep-sea (Cramwinckel
107 et al., 2018), which is surprising given that the South Pacific was presumably the
108 dominant region of deep-water formation during the Eocene (Huber and Thomas,
109 2010; Thomas et al., 2003; 2014).

110

111 1.2 GDGT paleothermometry

112 TEX_{86} utilizes the correspondence of higher abundances of cyclopentane rings
113 in sedimentary archaeal membrane lipids termed isoprenoid glycerol dialkyl glycerol
114 tetraethers (isoGDGTs) with higher SST of the overlying surface water (Schouten et
115 al., 2002). This relation is attributed to a viscoelastic adaptation of the membrane of
116 pelagic Thaumarchaeota, the dominant source organisms of isoGDGTs, to temperature
117 (Schouten et al., 2002; 2013). For some periods in geological deep time, including the



118 Paleocene and Eocene, TEX₈₆ calibrations based on GDGTs in core top sediments need
 119 to be extrapolated above the modern SST range (~30°C) to estimate SST. The
 120 linearity of the relation at and beyond the high-end of the core-top calibration is
 121 poorly known, leading to very high uncertainty in SST estimates at the warm end of
 122 the calibration (Hollis et al., 2019). However, as the absolute TEX₈₆ values of many
 123 Eocene sediments exceed those observed for modern core-tops, even the most
 124 conservative calibrations yield SSTs >30°C in the warmest intervals. Increasingly,
 125 such temperatures are corroborated by estimates from other SST proxies (e.g., Zachos
 126 et al, 2006; Frieling et al., 2017; Evans et al., 2018).

127 Along with calibration uncertainties, a number of confounding factors have
 128 been identified since first publication of the isoGDGT-based SST records from the sw
 129 Pacific (Hollis et al., 2009; 2012; Bijl et al., 2009; 2013), relating to a suite of pre-, syn-
 130 , and post-depositional processes that might alter the pelagic isoGDGT signal in
 131 marine sediments. IsoGDGT contributions from methanogenic (Blaga et al., 2009) and
 132 methanotrophic (Weijers et al., 2011; Zhang et al., 2011) archaea, deep-dwelling
 133 archaea (Taylor et al., 2013) and terrestrial sources (Hopmans et al., 2004; Weijers et
 134 al., 2006) to the sedimentary isoGDGT pool have been determined. These factors can
 135 be recognized in GDGT distributions, leading to better interpretation of TEX₈₆-based
 136 SST reconstructions. However, the influence of growth phase (Elling et al., 2014), and
 137 environmental ammonium and oxygen concentrations (Qin et al., 2015; Hurley et al.,
 138 2016) on sedimentary isoGDGT distributions are as yet poorly constrained.

139 Branched GDGTs (brGDGTs) produced by soil bacteria provide
 140 reconstructions of mean annual air temperature (MAAT) using the MBT'_{5me} index (De
 141 Jonge et al., 2014a; Naafs et al., 2019; Peterse et al., 2012; Weijers et al., 2007; Dearing
 142 Crampton-Flood et al., 2020). Based on brGDGTs supplied to marine sediments, the
 143 MAAT evolution of Australian-New Zealand continents followed the trends in deep-
 144 and surface ocean temperature remarkably well, but with much lower absolute
 145 values than SST (Bijl et al., 2013a; Pancost et al., 2013). Albeit to a lesser extent than
 146 SST, southern high-latitude MAAT reconstructions remain warmer than model
 147 simulations (Huber and Caballero, 2011) as well, even when simulations replicate
 148 equatorial surface and global deep ocean temperatures (Cramwinckel et al., 2018).
 149 With improved analytical techniques (Hopmans et al., 2016), brGDGT isomers with a



150 methylation on the 5- or the 6-position of the alkyl chain can now be separated and
 151 quantified (De Jonge et al., 2013). With this separation, the pH co-dependence of the
 152 brGDGT signal can be removed, isolating the temperature relation (De Jonge et al.,
 153 2014a; Naafs et al., 2017a). The separation of brGDGT isomers also allows to
 154 recognize and correct for potential contributions of aquatic brGDGTs to the soil-
 155 derived brGDGT signal stored in marine sediments that complicate the use of
 156 brGDGTs as continental paleothermometer in continental margin sediments (De
 157 Jonge et al., 2014b; 2015; Dearing Crampton-Flood et al., 2018; Peterse et al., 2009;
 158 Sinninghe Damsté, 2016; Tierney and Russell, 2009; Zell et al., 2013).

159 Lipids from a related biomarker family, the branched glycerol monoalkyl
 160 glycerol tetraethers (brGMGTs), were identified in the marine realm in core-top
 161 sediments (Liu et al., 2012), oxygen minimum zones (Xie et al. 2014), and later in peat
 162 (Naafs et al 2018a) and East African lake sediments (Baxter et al 2019), the latter
 163 identifying 7 individual brGMGTs. The brGMGTs contain a covalent bond connecting
 164 the two alkyl chains. The abundance of brGMGTs relative to that of brGDGTs, as well
 165 as the relative distribution of brGMGT isomers seems to vary with temperature in
 166 some degree (Baxter et al., 2019; Naafs et al., 2018a; Tang et al., 2021), although this
 167 is thus far only based on empirical relationships. The exact sources of these
 168 compounds, and consistency of such signals in various terrestrial, lacustrine and
 169 marine realms, are as yet not fully understood. The degree of methylation of a specific
 170 subset of brGMGT compounds (Naafs et al., 2018a, Sluijs et al. 2020) is however
 171 similar to those in brGDGTs for which membrane stability regulation is proposed as
 172 underlying mechanism (Weijers et al., 2007). Paleogene marine sediments from the
 173 Arctic Ocean do contain abundant brGMGTs (Sluijs et al., 2020), and these seem to be
 174 produced in the marine system with a strong oxygen minimum zone, and substantial
 175 terrestrial input. While the degree of methylation of acyclic brGMGTs (HMBT_{acyclic})
 176 does show similar trends to TEX₈₆ in the Paleogene Arctic record, brGMGTI, which
 177 was tentatively calibrated to temperature using a suite of tropical lakes (Baxter et al.,
 178 2019) does not (Sluijs et al., 2020). The response of brGMGTs to environmental and
 179 climatic changes, as well as their sourcing in the marine realm, is clearly diverse,
 180 complex and, as yet, poorly understood.

181

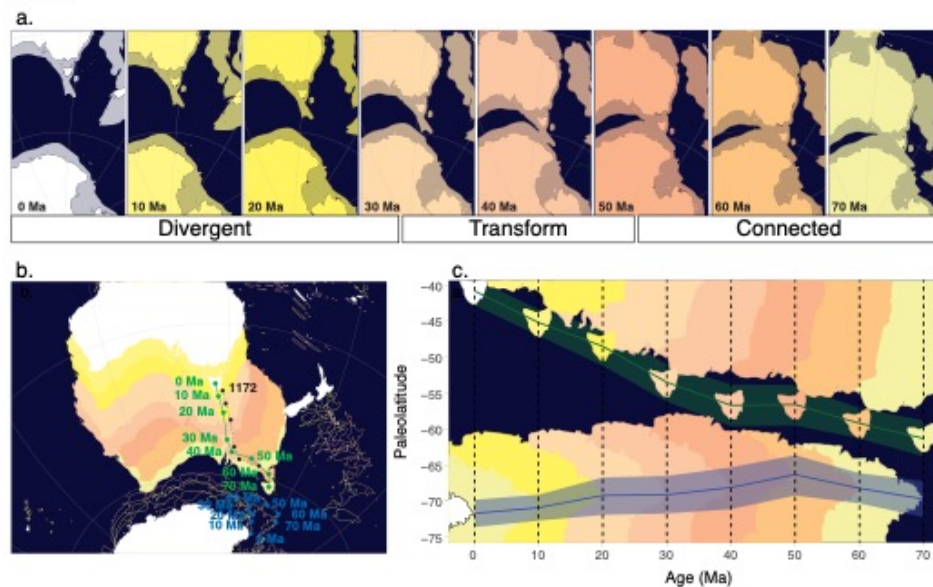


182 1.3 Revisiting GDGTs, and investigating GMGTs at Site 1172

183 From the new insights about the GDGT-based proxies, it is clear that assessing
184 the reliability of previously published GDGT-based temperature reconstructions
185 requires a revisit and a detailed constraints on past environmental, climatological and
186 depositional conditions. Moreover, the presence and proxy potential of brGMGTs the
187 early Paleogene sw Pacific has not yet been assessed. To this end, we have revisited a
188 sedimentary record from the sw Pacific Ocean: ODP Site 1172 on the East Tasman
189 Plateau (ETP), which contains an almost complete succession of late Cretaceous to
190 early Oligocene sediments (Bijl et al., 2013b; 2014; Brinkhuis et al., 2003;
191 Schellenberg et al., 2004; Stickley et al., 2004a). We have newly analysed isoGDGTs,
192 brGDGTs, and brGMGTs from the Maastrichtian and Paleocene section of the record,
193 and re-analysed previously published sample sets using the chromatography method
194 with improved compound separation (Hopmans et al., 2016). We critically evaluate
195 the biomarker results using established indicators for pre-, syn- or post-depositional
196 overprint of the primary sea surface and air temperature signals. In addition, we use
197 dinoflagellate cyst assemblages and terrestrial palynomorph abundance as recorders
198 of paleoenvironmental change on the continental shelf, for constraints on
199 depositional, environmental and hydrological changes, in order to aid interpretations
200 of the GDGT indices. We also evaluate the nature, source and possible temperature
201 affinity of the brGMGTs in our record. After this critical (re-)evaluation, we interpret
202 the Maastrichtian to early Oligocene sea surface and air temperature, and
203 paleoenvironmental evolution of the sw Pacific region.



Fig. 1



204

205 Figure 1. Absolute plate tectonic changes around Tasmania during the Maastrichtian
206 to recent. a. TG opening in 8 time slices, from 70 Ma to present. Relative tectonic
207 motion between Australia and Antarctica is limited until about 53 Ma, transfer until
208 34 Ma and divergent from 34 Ma onwards. b. Like a, but on an orthographic
209 projection (compiled with Gplates, using paleomagnetic reference frame (Torsvik et
210 al., 2012) from Seton et al. (2012)). Green and blue lines and dots represent the
211 pathways of Tasmania and Cape Adare, respectively, black dashed line and dots
212 represent the paleo-position of Site 1172. c. The absolute paleolatitude of Cape Adare,
213 Antarctica (blue line, blue shading = uncertainty), as indicative of the plate tectonic
214 motion around the pathway of the Tasman Current, and of Tasmania (green line,
215 green shading = uncertainty), as indicative of the plate tectonic motion of the source
216 area of the terrestrial organic matter, in 10 Myr time steps from 70–0 Ma. obtained
217 from paleolatitude.org (Van Hinsbergen et al., 2015). Plate contours represent
218 paleolatitude of present-day shorelines, for orientation (obtained from Gplates). Note
219 that in visualizations b and c, submerged continental crust is not shown, but does
220 limit TG opening.

221

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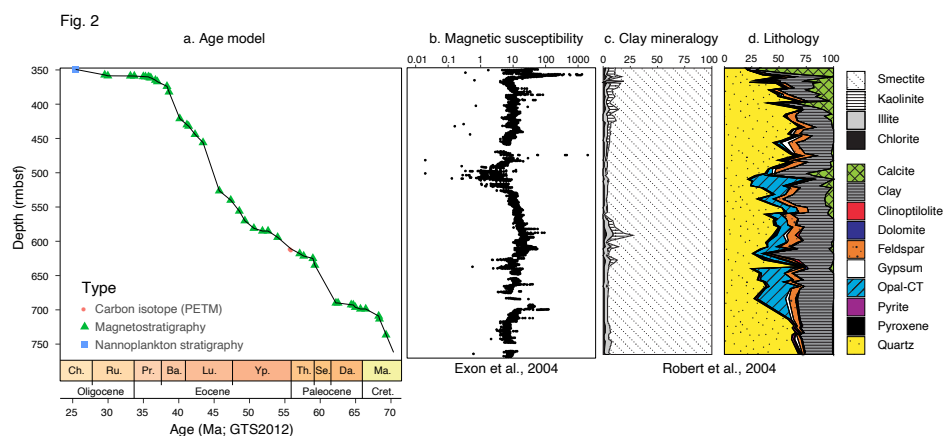


2. Material

2.1 Site locality and tectonic evolution

Since the Cenomanian, the continental complex including Australia, Tasmania, ETP and the South Tasman Rise (STR) tectonically moved as one continental plate (Müller et al., 2000) – here referred to as Australia. True polar wander, although relatively poorly constrained, has caused absolute plate motions relative to the spin axis of the Earth (Torsvik et al., 2012; Van Hinsbergen et al., 2015). On a paleomagnetic reference frame, Antarctica and Australia rotated northwestward by 3° of latitude and over 3° longitude between 70 and 50 Ma (Fig. 1), with only little transform displacement between them. From 50 Ma onwards, the tectonic drift orientation of Antarctica shifted slightly more southwards than that of Australia, causing left transform motion between notably west of Cape Adare and west Tasmania (Fig. 1; Williams et al., 2019). Although this divergence effectively opened the Tasmanian Gateway (TG) to surface flow of the Antarctic Counter Current close to the early-middle Eocene boundary (Bijl et al., 2013a), probably through subsidence of the Antarctic continental margin and STR, rapid northward movement of Australia and southward movement of East Antarctica did not start before about 40 Ma (Fig. 1; Cande and Stock, 2004; Seton et al., 2012). Yet, a connection between Australian and Antarctic continental blocks persisted until 34Ma when transform motion between STR and Wilkes Land changed into spreading and ocean crust formation (Cande and Stock, 2004; Seton et al., 2012).

244



245



246 Figure 2. a. Age tie points used to construct the age model for ODP Site 1172, based on
 247 carbon isotope (Bijl et al., 2010; Sluijs et al., 2011), magnetostratigraphic (Bijl et al.,
 248 2010; Houben et al., 2019; Stickley et al., 2004a), and nannofossil (Houben et al.,
 249 2019) age constraints. b. Shipboard magnetic susceptibility from Exon et al. (2001). c.
 250 Shore-based clay mineralogy and d. smear slide-based lithological observations from
 251 Robert (2004).

252

253

254 2.2 Lithology

255 A 760 meter thick sediment sequence was recovered at ODP Site 1172, on the
 256 western part of the ETP, east of Tasmania, Australia (Exon et al., 2001). We studied
 257 the interval from 760 meters below sea floor (mbsf) to about 350 mbsf in a composite
 258 (Röhl et al., 2004a) of Hole A and D. The carbonate ooze from 350 mbsf upwards did
 259 not yield any GDGTs. The studied succession consists broadly of green-grey silty
 260 claystones from 760 to ~505 mbsf, grading into dark grey to black silty claystones
 261 from 636 – 625 mbsf, then a return of the greenish grey silty claystones, which is
 262 interrupted by a more lithified unit at 610 mbsf (Exon et al., 2001; Röhl et al., 2004a,
 263 b; Schellenberg et al., 2004; Sluijs et al., 2011). The silty claystones turn gradually and
 264 cyclically lighter from 570 mbsf upwards until ~361 mbsf, with gradually more
 265 calcium carbonate and diatoms preserved (Röhl et al., 2004a). The clay mineralogy is
 266 dominated by smectite but contains progressively more kaolinite above ~500 mbsf
 267 (Fig. 2; Exon et al., 2001; Robert, 2004). Grey silty claystones give way to a green
 268 glauconitic horizon between 360.1 and 357.3 mbfs (Exon et al., 2001). Above this
 269 greensand, carbonate oozes continue further upwards. The record is quasi-
 270 continuously bioturbated by zoophycos (Exon et al., 2001). We follow the depth
 271 corrections published in (Sluijs et al., 2011) for Cores 12R–31R, based on the
 272 correlation of core-log and downhole log magnetic susceptibility.

273

274 2.3 Age model

275 For age tie points (Table S1; Fig. 2), we use the identification of the PETM
 276 (Sluijs et al., 2011) and Middle Eocene Climatic Optimum (MECO; Bijl et al., 2010) and
 277 robust biostratigraphical constraints from the Eocene–Oligocene Boundary interval



278 (Houben et al., 2019; Sluijs et al., 2003). In the intervals in between, the age model
 279 relies on magnetostratigraphy, which in some intervals suffers from a strong normal
 280 overprint (Dallanave et al., 2016), calibrated using nannofossil- diatom- and dinocyst
 281 biostratigraphy (Bijl et al., 2013b; Stickley et al., 2004). Despite the overprint, there
 282 does seem to be a paleomagnetic signal preserved in the inclination data (Fuller and
 283 Touchard, 2004), and biostratigraphic constraints are broadly consistent with nearby
 284 sites (Bijl et al., 2013b; Dallanave et al., 2016).

285 The nature of the sediments and depositional setting (see below) implies that
 286 small hiatuses may exist between the age tie points (Röhl et al., 2004a). Indeed,
 287 hiatuses were already identified across the Cretaceous–Paleogene Boundary
 288 (Schellenberg et al., 2004), in the mid-Paleocene (Bijl et al., 2013b; Hollis et al., 2014),
 289 early Eocene (Bijl et al., 2013b) and in the middle Eocene (Röhl et al., 2004a),
 290 corresponding to increases in magnetic susceptibility (Fig. 2). The section across the
 291 Eocene–Oligocene transition is strongly condensed (Houben et al., 2019; Stickley et
 292 al., 2004a, b).

293

294 2.4 Depositional setting

295 From the lithology (Robert, 2004) and palynological content (Brinkhuis et al.,
 296 2003), the Maastrichtian–Eocene part of the record was interpreted to represent a
 297 shallow-marine, mid-continental shelf depositional setting, with gradual deepening in
 298 the middle Eocene based on an increase in calcium carbonate content (Fig. 2; Röhl et
 299 al., 2004a). The late Eocene greensands have been initially interpreted as evidencing
 300 strong deepening and current inception as a result of widening of the TG (Stickley et
 301 al., 2004b). Later studies have related the greensands to invigorated ocean circulation
 302 and winnowing (Houben et al., 2019), and not necessarily deepening of the site.
 303 Volcanic activity of the Balleny plume in the late Eocene (Hill and Moore, 2001) might
 304 have played a profound role in the bathymetric changes of the ETP, in addition to the
 305 tectonic stresses that act on diverging plates in the TG area.

306 The regional tectonic evolution has implications for interpreting the marine
 307 and terrestrial temperature record at this site, because it influenced regional
 308 oceanography and climatic conditions in the hinterland catchment area. Field data
 309 and model simulations indicate that with a closed TG, the Tasman Current, a strong



310 western boundary current of the proto-Ross gyre, bathed the plateau with Antarctic-
 311 derived surface waters (Bijl et al., 2011; 2013b; Huber et al., 2004; Sijp et al., 2014;
 312 2016). Palynological evidence confirms that the Proto-Ross Gyre influence persisted
 313 at the ETP until the late Eocene (Bijl et al., 2011; Warnaar et al., 2009). This means
 314 that despite northward tectonic drift, the same strong western boundary current
 315 bathed the site during the Maastrichtian to early Eocene (Sijp et al., 2016), with
 316 perhaps some intermittent influence of East Australian Current waters from the north
 317 (Bijl et al., 2010; Cramwinckel et al., 2020). This ended when the proto-Leeuwin
 318 Current started to flow through the progressively widening TG (Fig. 1), bringing the
 319 ETP under the influence of more northerly sourced surface waters (Houben et al.,
 320 2019; Stickley et al., 2004b).

321 The source area for the terrestrial organic matter (OM) and detrital input was
 322 likely Tasmania. Persistent terrigenous input (Brinkhuis et al., 2003) arguably
 323 requires a large terrestrial catchment area, and the ETP seems too small (~50,000
 324 km²) to have had vast areas above sea level. Moreover, Paleocene–Eocene terrestrial
 325 palynomorph assemblages contain common Permian–Triassic elements (Contreras et
 326 al., 2014); the Permian–Triassic upper Parmeener group contains thick terrestrial
 327 (coal) deposits and comprises the surface lithology of most of eastern Tasmania
 328 today. Although that same formation might be present in the ETP subsurface as well,
 329 it was probably covered with sediment throughout the Cenozoic (Hill and Moore,
 330 2001). Rivers flowing from southeast Australia drained into the Gippsland and Bass
 331 Basins, and that terrigenous material is unlikely to have reached the ETP. Seismic
 332 information from the East Tasman Saddle, connecting the Tasmanian Margin to the
 333 ETP, does not suggest there was a deep basin in between (Hill and Exon, 2004).
 334 Therefore, Tasmanian-sourced detrital material could reach the ETP. The ETP was
 335 close enough to the Antarctic margin during the Maastrichtian–early Eocene to have
 336 received perhaps a minor component of Antarctic-sourced terrestrial OM input, in
 337 addition to the dominant Tasmanian source. The regional palynology (Macphail,
 338 2000; 2002; Carpenter et al., 2012; Pross et al., 2012; Contreras et al., 2013; 2014;
 339 Truswell, 1997), the abundance of peatlands (Holdgate et al., 2009), and the felsic
 340 lithology (Moore, Betts, and Hall, 2013) suggests the hinterland catchment contained
 341 acidic, wet soils and peats.



342

343 2.5 Samples

344 For this study, we used lipid extracts that have been analysed for GDGTs
 345 previously (Bijl et al., 2009; Hollis et al., 2014; Houben et al., 2019; Sluijs et al., 2011).
 346 We augmented these with 69 new extracts of sediments from the Maastrichtian and
 347 the Paleocene, to extend and improve the temporal resolution of the record.
 348 Unfortunately, not all archived samples from the PETM interval (Sluijs et al., 2011)
 349 could be located, and the interval was reanalysed in a lower resolution (6 samples
 350 over the PETM interval). For the remaining PETM, we used the published peak areas
 351 (Sluijs et al., 2011) to calculate TEX₈₆, which is warranted given the new analytical
 352 technique does not affect isoGDGT peak area ratios (Hopmans et al., 2016). For
 353 palynology, we collated and revisited data presented in (Bijl et al., 2010; 2013b;
 354 Brinkhuis et al., 2003; Houben et al., 2019; Sluijs et al., 2011) and generated higher-
 355 resolution data for the Maastrichtian and Paleocene.

356

357 3. Methods

358

359 3.1 Organic geochemistry

360 3.1.1 Extraction, column separation and analysis

361 Earlier work (Bijl et al., 2009; 2010; 2013a; Houben et al., 2019; Sluijs et al.,
 362 2011) presented in detail the extraction, Al₂O₃ column separation and filtering
 363 techniques used for the samples. We followed the same procedures for the processing
 364 of the new samples. We reanalysed all available polar fractions using the double
 365 column UHPLC-MS approach as described in (Hopmans et al., 2016). In short,
 366 processing involved extraction with a Dionex accelerated solvent extractor using
 367 dichloromethane:methanol (DCM:MeOH) 9:1 (v/v), column separation of the total
 368 lipid extract using solvent mixtures hexane:DCM 9:1 (v/v), hexane:DCM 1:1 (v/v) and
 369 DCM:MeOH 1:1 (v/v) for apolar, ketone and polar fractions, respectively. Polar
 370 fractions were filtered using a 0.45 µm polytetrafluorethylene filter, and analysed
 371 using an Agilent 1260 Infinity series HPLC system coupled to an Agilent 6130 single-
 372 quadrupole mass spectrometer.

373 3.1.2 Data analysis: indices, overprints



374 Since the discovery of isoGDGTs as proxy for SST (see Schouten et al. (2013)
 375 for a review), several non-SST effects have been identified that may affect the
 376 distribution of isoGDGTs in the sediment. Several indices have been developed to
 377 identify most of the known sources of overprints (Table 1). Next to signalling SST-
 378 unrelated influences on the isoGDGT pool, these indices also provide information on
 379 the prevailing marine and paleoenvironmental and depositional conditions.

380

381 Table 1: Indices from iso- and brGDGTs and brGMGTs. For the chemical structure of
 382 these components see Fig. 3.

Index name	Equation	Proxy for	Cut-off value	Source
TEX ₈₆	$\frac{GDGT2 + GDGT3 + Cren'}{GDGT1 + GDGT2 + GDGT3 + Cren'}$	Sea surface temperature	-	(Schouten et al., 2002)
BIT index	$\frac{IIIa + IIIa' + IIa + IIa' + Ia}{Cren + IIIa + IIIa' + IIa + IIa' + Ia}$	Terrestrial input	>0.4? Site-dependent	(Hopmans et al., 2004)
fcren'	$\frac{\%Cren'}{\%Cren' + \%Cren}$	Non-thermal contribution of crenarchaeol isomer	0.25	(O'Brien et al., 2017)
Methane Index	$\frac{GDGT1 + GDGT2 + GDGT3}{GDGT1 + GDGT2 + GDGT3 + Cren + Cren'}$	Contribution by methane-metabolising archaea	>0.3	(Zhang et al., 2011)
AOM ratio	$GDGT2/Cren$	Contribution by anaerobic methane oxydizers	>0.2	(Weijers et al., 2011)
GDGT2/3 ratio	$GDGT2/GDGT3$	Contribution by deep-dwelling archaea	>5	(Taylor et al., 2013)
Methanogenesis	$GDGT0/Cren$	Contribution by methanogenic archaea	>2.0	(Blaga et al., 2009)
Ring index (RI)	$0\%GDGT0 + 1\%GDGT1 + 2\%GDGT2 + 3\%GDGT3 + 4\%Cren + 4\%Cren'$	Non-pelagic GDGT composition	$\Delta RI > 0.3^*$	(Zhang et al., 2016)
MBT _{5me}	$\frac{IIIa + IIIb + IIIc}{IIIa + IIIb + IIIc + IIa + IIb + IIc + Ia}$	Mean annual air temperature	-	(De Jonge et al., 2014a)
CBT	$\frac{IIIb + IIb' + IIc' + Ia'}{IIIa + IIa + Ia}$	(soil-)pH	-	(De Jonge et al., 2014a)
#rings _{tetra}	$\frac{Ib + 2 * Ic}{Ia + Ib + Ic}$		>0.7	

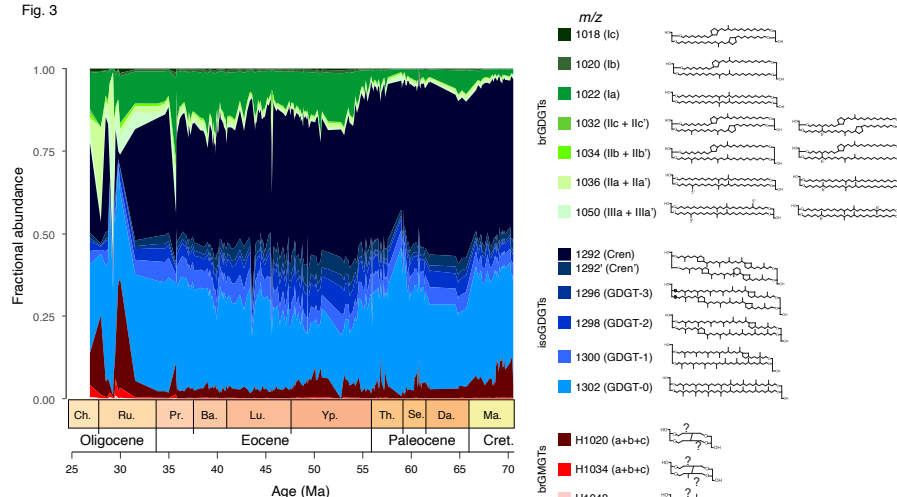


#rings _{penta}	$\frac{I Ib + I Ib' + 2 * I Ic + 2 * I Ic'}{I Ia + I Ia' + I Ib + I Ib' + I Ic + I Ic'}$	Marine in situ brGDGT production	-	(Sinninghe Damsté, 2016)
#rings _{penta5}	$\frac{I Ib + 2 * I Ic}{I Ia + I Ib + I Ic}$		-	
#rings _{penta6}	$\frac{I Ib' + 2 * I Ic'}{I Ia' + I Ib' + I Ic'}$		-	
IR	$\frac{I Ia' + I Ib' + I Ic' + I a'}{I Ia + I Ia' + I Ib + I Ib' + I Ic + I Ic' + I a + I a'}$	River in situ brGDGT production	Depends on soil value	(De Jonge et al., 2014b)
IR _{hexa}	$\frac{I a'}{I a + I a'}$		-	(Sinninghe Damsté, 2016)
IR _{penta}	$\frac{I Ia' + I Ib' + I Ic'}{I Ia + I Ia' + I Ib + I Ib' + I Ic + I Ic'}$		-	
brGMGTI	$\frac{H1020c + H1034a + H1034c}{H1020b + H1020c + H1034a + H1034b + H1034c}$	Temperature, oxygenation?	In lakes	(Baxter et al., 2019)
%brGMGT, %brGMGT _a	$\frac{[brGMGT]}{[brGMGT] + [brGDGT]} \times 100$		In peats, lakes	(Baxter et al., 2019) (Naafs et al., 2018a)
cyclic	$\frac{[brGMGT]}{[Ia + I Ia + I Ia + brGMGT]} \times 100$		In peats	(Naafs et al., 2018a)
HMBT _{acyclic} , all	$\frac{H1020a + H1020b + H1020c + H1034a + H1034b + H1034c}{H1020a + H1020b + H1020c + H1034a + H1034b + H1034c}$		In Arctic Ocean sediments	Sluijs et al., 2020, sensu Naafs et al., 2018a)
HMBT _{acyclic}	$\frac{H1020c}{H1020c + 1034b + H1048}$			

383 * Cutoff depends on TEX₈₆ value. ΔRI = RI_{TEX} - RI, where RI_{TEX} = -0.77*TEX₈₆ + 3.32*TEX₈₆² + 1.59

384

Fig. 3



385



386 Figure 3. Fractional abundances of the various branched (green) and isoprenoidal
 387 (blue) GDGTs, and branched GMGTs (red) at ODP Site 1172.

388
 389

390 All of the ratios that signal biases express the influence of this bias relative to
 391 modern “normal” or idealized compositions. The abundance of overprinted
 392 components is divided over either a total sum of overprinted and non-overprinted
 393 components (closed sum; e.g., methane index, IR) or over only the non-overprinted
 394 (ratio; GDGT2/3 ratio, AOM ratio). The complication is that a change in index value
 395 can be caused by changes in the denominator or numerator. This is particularly
 396 evident in the interpretation of the branched and isoprenoid tetraether (BIT) index. It
 397 was proposed that the closed sum ratio between brGDGTs, which were thought to
 398 derive exclusively from soils, and the isoGDGT crenarchaeol, which was thought to be
 399 produced exclusively by marine archaea, could be an indicator for the relative
 400 contribution of soil organic matter (OM) into marine sediments (Hopmans et al.,
 401 2004). However, it was subsequently shown that crenarchaeol, but critically also
 402 other isoGDGTs are produced on land (Weijers et al. (2006), and that a large
 403 contribution of soil-OM to marine sediments can cause a bias in TEX₈₆. Weijers et al.
 404 (2006) used an end-member model to calculate the potential bias in TEX₈₆-based SST
 405 reconstructions in the Congo Fan and this shows significant bias (>2°C) for BIT above
 406 0.3. However, it is rather arbitrary to assign a cut-off for BIT based on the magnitude
 407 of the SST bias alone, as this bias depends primarily on the difference between the
 408 isoGDGT composition of the soil and marine endmembers. As a consequence, there is
 409 no uniform cut-off value for BIT index above which TEX₈₆ should be discarded.
 410 Secondly, because the BIT index is a closed sum, the index is equally affected by
 411 increased production of crenarchaeol as it is by the input of brGDGTs. Although this
 412 could be verified with absolute concentrations of GDGTs, these are not always
 413 presented. Third, brGDGTs are produced in situ in the marine realm (Dearing
 414 Crampton-Flood et al., 2019; Peterse et al., 2009; Sinninghe Damsté, 2016). The
 415 corollary of this is that the cut-off value for BIT to infer an overprint in the isoGDGT-
 416 based SST signal should be assumed to differ per environmental setting, and perhaps
 417 also through time if the setting changes majorly (e.g., Sluijs et al., 2020). As a result,



cut-offs were set higher (e.g., 0.4; Bijl et al., 2013a) when no relationship between
 TEX₈₆ and BIT index values was detected. This approach is also not without
 complications, because the abundance and composition of isoGDGTs (including
 crenarchaeol) in soil material varies (Weijers et al., 2006; De Jonge et al., 2015; De
 Jonge et al., 2016; Naafs et al., 2019; Sluijs et al., 2020). This in turn implies that in
 samples with high BIT (Weijers et al., 2006), the terrestrially-derived isoGDGT
 contribution will not be uniform, and might become impossible to detect in a TEX₈₆-
 BIT index cross plot. In some paleo-environmental settings that evidently lack high
 input of soil-OM, BIT index values are high (Leutert et al., 2020). For reasons given
 above, this could either mean very low crenarchaeol production in the marine realm,
 or in situ production of brGDGTs. These two scenarios make for a completely
 different paleoenvironmental interpretation, and fail to indicate whether the
 brGDGTs can be used to reconstruct MAAT of the hinterland or whether isoGDGTs
 can be used for reliable paleothermometry. For that, it is important to assess the
 sources of brGDGTs, as brGDGTs produced in rivers (De Jonge et al., 2014b; Zell et al.,
 2013) or the (coastal) marine environment (Dearing Crampton-Flood et al., 2019;
 Peterse et al., 2009; Sinninghe Damsté, 2016) may contribute and thereby bias the
 initial soil-derived brGDGT signal, affecting the reliability of MAAT reconstructions.
 Fortunately, production of brGDGTs in the marine realm can be recognized based on
 the weighed number of rings of the tetramethylated brGDGTs, quantified in the
 #rings_{tetra}, where values >0.7 indicate a purely marine source of the brGDGTs
 (Sinninghe Damsté, 2016). Similarly, brGDGTs that are produced in rivers are
 characterized by a relatively high abundance of the 6-methyl brGDGTs relative to the
 5-methyl brGDGTs, expressed as the Isomerization Ratio (IR; De Jonge et al., 2014b).

For the other overprinting indices, cut-off values may not be as uniformly
 applicable to all depositional settings either. Leutert et al. (2020) showed that
 modern samples exceeding cut-off values for the GDGT2/3 ratio and methanogenesis
 index do not have anomalous TEX₈₆ index values in the modern core-top dataset
 based on the Ring index. This suggests that the TEX₈₆ index value is not equally
 influenced by non-thermal contributions in all depositional settings. As a result,
 evaluating non-thermal contributions on TEX₈₆ index values should consider the
 depositional and environmental setting along with the cut-off values.



450

451 3.1.3 Calibrations for TEX_{86} and $\text{MBT}'_{5\text{me}}$

452 An extensive number of calibrations has been proposed to convert TEX_{86} index
 453 values to sea surface temperatures. At the heart of the calibration discussion is the
 454 uncertainty whether the TEX_{86} -to-SST relationship continues in a linear way beyond
 455 the modern SST range, or in an exponential way (Hollis et al. 2019). The answer to
 456 that is beyond the scope of this paper, and we refer to (Cramwinckel et al., 2018;
 457 Hollis et al., 2019; O'Brien et al., 2017; Tierney et al., 2017) for detailed discussions,
 458 specifically about using TEX_{86} in regions with SSTs warmer than modern. Following
 459 recent recommendations (Hollis et al. 2019), we here apply several calibrations to
 460 convert our TEX_{86} values to SSTs. We apply an exponential calibration (Kim et al.,
 461 2010) for which we acknowledge that this calibration suffers from a regression
 462 dilution bias (Tierney and Tingley, 2015), and two linear calibrations; one on the
 463 warm part ($>15^\circ\text{C}$) of the core-top dataset (O'Brien et al., 2017) and one using
 464 Bayesian statistics (BAYSPAR; Tierney and Tingley, 2015) to convert our TEX_{86} values
 465 to SSTs. BAYSPAR produces a linear calibration based on a subset of the core top data
 466 with similar GDGT assemblages as the measured TEX_{86} , with a user-defined tolerance.
 467 The larger the subset of core tops (i.e., when the number of identified core-top
 468 analogues to the sample TEX_{86} values is large), the closer the Bayesian calibration will
 469 approach a global linear regression. For MAAT reconstructions, we use the Deming
 470 regression of the soil-specific calibration of Naafs et al. (2017b; $\text{MAAT}_{\text{soil}}$) and the
 471 Bayesian BayMBT₀ of Dearing Crampton-Flood et al. (2020). The latter follows the
 472 approach of BAYSPAR, but then for $\text{MBT}'_{5\text{me}}$. We will compare the calibrations for
 473 both proxies to middle Eocene U^K_{37} -based SSTs (from Bijl et al., 2010, but using
 474 BAYSPLINE calibration of Tierney and Tingley, 2018) and Paleocene–early Eocene
 475 sporomorph-based MAAT estimates (Contreras et al., 2014) from the same site.

476

477 Table 2. Calibration equations for TEX_{86} and MAAT

Calibration	Equation	Type	Proxy for	Source
SST_{exp} ($\text{TEX}_{86}^{\text{H}}$)	$68.4 * \log_{10}(\text{TEX}_{86}) + 38.6$	Exponential (regression dilution)	Mean annual SST (0– 20m)	(Kim et al., 2010)



SST _{lin}	58.8 * TEX ₈₆ - 11.18	Linear	Mean annual SST (0–20m)	(O'Brien et al., 2017)
BAYSPAR	Prior mean = 20, prior std = 10, search tolerance = 0.15	Bayesian linear	Mean annual SST (0–20m)	(Tierney and Tingley, 2015)
MAAT _{soil}	40.01 * MBT' _{5me} - 15.25	Linear, Deming regression	Mean annual air temperature (for days above freezing)	(Naafs et al., 2017b)
BayMBT ₀	Prior mean = 20, prior std = 15	Bayesian linear	Mean annual air temperature	(Dearing Crampton-Flood et al., 2020)

478

479

3.1.4 R-script for data analysis and evaluation

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To facilitate systematic calculation of GDGT ratios, data analysis, visualization, and evaluation, we constructed a set of R markdowns

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(<https://github.com/bijlpeter83/RGDGT.git>) that can be applied to any time- or

482

depth series of isoGDGT, brGDGT and/or brGMGT data. The R script loads peak areas

483

of GDGTs/GMGTS from Microsoft excel spreadsheets, calculates and plots fractional

484

abundances, overprinting indices and paleotemperature time or depth series.

485

486

3.2 Palynology

487

3.2.1 Sample processing

488

Palynological sample processing techniques were published in the original papers describing these datasets (Bijl et al., 2010; 2013b; Brinkhuis et al., 2003; Houben et al., 2019; Sluijs et al., 2011). We followed the same procedure for the new Maastrichtian–Paleocene samples. The different publications do indicate differences in the employed sieve mesh sizes (10 or 15 µm) but given all counted dinocysts are larger than 15 µm, this has not led to differences in the dinocyst results.

489

3.2.2 Taxonomy

490

We used the taxonomic framework cited in Williams et al., (2017) in our counts down to the species level, with one exception. For the Wetzelielloidae subfamily, we follow Bijl et al. (2017), for reasons stated therein. We also follow the supra-generic classification based on Fensome et al. (1993); genera described post-1993 were added into that classification accordingly.

491

3.2.3 Ecological affinities of dinocyst ecogroups



502 In this paper, we present our results plotted in eco-groups and complexes,
 503 which groups dinocyst species with similar ecologic affinities based on modern (for
 504 extant taxa) and empirical (for extant and extinct taxa) data (Frieling and Sluijs, 2018;
 505 Sluijs et al., 2005). The species of which the ecologic affinity was not assessed
 506 previously were included into larger groups based on shared morphological
 507 characteristics; primarily tabulation, archaeopyle and cyst- and process shapes. For
 508 example, the “*Apectodinium* complex” is equated to “Wetzellioids” following Frieling
 509 et al. (2014; see Table S2 for taxonomic grouping). We directly compare dinocyst- and
 510 terrestrial palynomorph indices with GDGT-based indices (following the approach of
 511 Frieling and Sluijs, 2018) to arrive at multi-proxy reconstructions of SST, river runoff
 512 and distance to shore (Table 3). For this, we resampled and binned GDGT-based
 513 indices to the sample resolution of the dinocyst data.

514

515 Table 3. Environmental parameters, and their corresponding GDGT indices and
 516 dinocyst eco-groups, based on Sluijs et al. (2005), Frieling and Sluijs (2018) and Sluijs
 517 and Brinkhuis (2009).

Environmental parameter	GDGT index	Dinocyst eco-group
Sea surface temperature	TEX ₈₆	%Thermophiles (<i>Wetzellioideae</i> , <i>Hafniasphaera</i> spp., <i>Florentinia reichartii</i> , <i>Polysphaeridium</i> spp, <i>Homotryblium</i> spp., <i>Heteraulacacysta</i> spp., <i>Eocladopyxis</i> spp., <i>Dinopterygium</i> spp.)
Runoff, fresh water, salinity	IR, BIT	% <i>Senegalinium</i> cpx, % <i>Phthanoperidinium</i> spp., %Terrestrial palynomorphs
Distance to shore	BIT, #ringstetra	%Open marine (<i>Apectodinium</i> spp., <i>Impagidinium</i> spp., <i>Operculodinium</i> spp., <i>Spiniferites</i> spp.), % <i>Glaphyrocysta</i> cpx., %Epicystal Goniodomids (<i>Polysphaeridium</i> spp, <i>Homotryblium</i> spp., <i>Heteraulacacysta</i> spp., <i>Eocladopyxis</i> spp.,



		<i>Dinopterygium</i> spp.), %Terrestrial palynomorphs
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518

519

520 3.2.4 Diversity and variability indices

521 To gain insight into the diversity and variability of the dinocyst assemblage
 522 through time, we employ several indices (Table 4), and compare their results at
 523 species and ecogroup level. We note that diversity in dinocyst taxonomy is
 524 complicated since dinocysts are dominantly produced during the hypnozygotic phase
 525 of the dinoflagellate life cycle following sexual reproduction. Only ~15% of modern
 526 dinoflagellates features this cyst stage (Fensome et al., 1993), which leaves a large
 527 proportion of the biological group unrepresented in the fossil record, including for
 528 example coral and foraminifer symbionts, but also more closely related free-
 529 swimming taxa. Dinocyst diversity can therefore only be used as an indicator for
 530 dinocyst-producing dinoflagellates. As a further complication, taxonomic divisions of
 531 cysts, although strongly related to dinoflagellate morphology and taxonomy, is fully
 532 based on the morphology of dinocysts. Even modern dinoflagellate-dinocyst
 533 relationships are often not unambiguous. Some dinoflagellate species produce
 534 various cyst morphologies (Rochon et al., 2008) that represent multiple cyst genera
 535 or species. Some of these cyst morphological variations are the result of ecology
 536 (Mertens et al., 2011). This complicates comparing cyst datasets on the species level
 537 and affects assessing biological diversity based on cysts. On the genus level,
 538 taxonomic division of cysts is in most cases based on plate tabulation, which is the
 539 morphologic feature that has the closest relationship to dinoflagellate biological
 540 diversity (Fensome et al., 1993). However, ecologic or biogeographic affinities have
 541 been established for some dinocysts on a species level (e.g., Frieling and Sluijs, 2018),
 542 which does suggest subtle morphological features may have biologic and ecologic
 543 significance. The eco-groups we use here are the result of extensive reviews of
 544 empirical data. These groups combine dinocyst genera and species with
 545 fundamentally similar plate tabulations, and thus probably group cysts of closely
 546 affiliated biological dinoflagellate species, and, as such also ecologic affinities. For our



diversity calculations we use both the ungrouped data on species level and the diversity in dinocyst ecogroups.

As the simplest approximation of biological diversity, the richness R in terms of dinocyst taxa was summed. Furthermore, several diversity measures were calculated using the R package Vegan (Oksanen et al., 2015). Of these, Fisher's alpha (α) is based on the count data, whereas the Shannon index (H') and Simpson index (D) derive from the relative abundance data. Finally, we employed the Σcv metric (Gibbs et al., 2012), which we here dub the "Gibbs index", and which provides a measure for assemblage variability. Together, these metrics can give insight into changing stability and diversity of these regional dinocyst assemblages over the Maastrichtian to early Oligocene.

Table 4. Diversity and variability indices for (fossil) assemblages. In the Shannon and Simpson indices, p_i represents the proportional abundance of the i^{th} taxon of the total amount of taxa R . The Gibbs index summates the coefficients of variation (SD/mean) of all taxa (i to R) over a certain rolling window $t_1 \rightarrow t_2$. In Fisher's alpha, the α parameter is estimated from the dataset in which S is the expected number of species with an abundance of n . x represents a nuisance parameter estimated from the dataset, generally between 0.9 and 1. While the Gibbs index is calculated over a rolling window, the other indices are calculated per sample.

Index	Equation	Source
Shannon index (H')	$H' = \sum_{i=1}^R p_i * \ln p_i$	(Shannon, 1948)
Simpson index (D)	$D = \sum_{i=1}^R p_i^2$	(Simpson, 1949)
Gibbs index (Σcv)	$\Sigma cv = \sum_{i=1}^R \left(\frac{SD_{i,t_1 \rightarrow t_2}}{\text{mean}_{i,t_1 \rightarrow t_2}} \right)$	(Gibbs et al., 2012)
Fisher's alpha (α)	$S_n = \frac{\alpha x^n}{n}$	(Fisher et al., 1943)

3.3 Comparison of GDGT and dinocyst assemblage data



Both dinocyst ecogroups and GDGT indices bear information on SST, runoff (~salinity), nutrients and marine primary productivity, and relative distance to shore (Table 4). As dinocyst and GDGT analyses were performed on partly separate datasets, the highest resolution dataset (dinocysts) was linearly resampled to the depth intervals of the GDGT data to facilitate inter-comparison.

4. Results

4.1 GDGTs

The resulting dataset has an average temporal resolution of ~130 kyrs for the time interval between 70 and 30 Ma. IsoGDGTs dominate the pool of GDGTs (particularly GDGT-0 and crenarchaeol), with a gradual increasing relative abundance of brGDGTs (particularly Ia) throughout the record (Fig. 3). BrGMGTs are in low (<10%) relative abundance, except in the Oligocene, where they account for up to 30% of the total GDGT/GMGT pool.

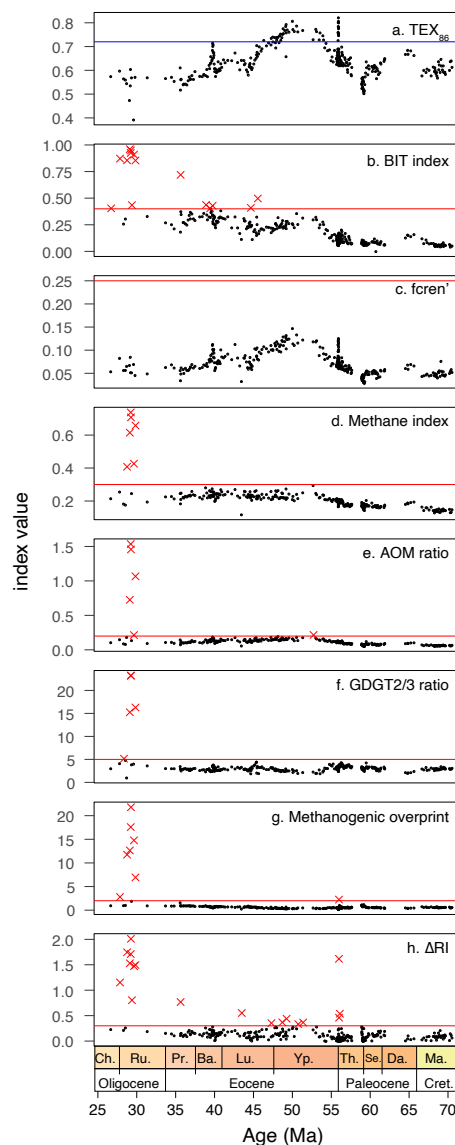
4.1.1 Isoprenoidal GDGTs

Before the TEX₈₆ index results (Fig. 4a) can be interpreted in terms of SST, we assess whether the isoGDGTs are primarily derived from the sea surface, or have received contributions from non-pelagic sources. Across the Ypresian, the BIT index (Fig. 4b) increases from values around 0.1 in the underlying sediments to values at times exceeding 0.4. Most EECO and middle Eocene values, however, have BIT index values between 0.2 and 0.35. The remaining sediments have BIT index values around 0.3, with the exception of some upper Eocene and Oligocene sediments that occasionally have values around 0.75. There is no correlation between TEX₈₆ and BIT in our entire dataset ($p=0.15$). None of the samples have $f_{\text{cren'}}$ (Fig. 4c) values above the cut-off of 0.25, suggesting no non-temperature related contribution of the crenarchaeol isomer to the isoGDGTs. In fact, the consistent trend in ($f_{\text{cren'}}$) and the TEX₈₆ index (Fig. 4) (and TEX₈₆^L, not shown) demonstrates the temperature sensitivity of $f_{\text{cren'}}$ and confirms that discarding it from the index as was done for the cold-temperature index TEX₈₆^L (Kim et al., 2010) is not justified here. With the exception of 6 samples in the Oligocene, all samples have methane index values below the conservative cut-off value of 0.3 (Fig. 4d), although in the early and middle



601 Eocene, methane index values do approach the cut-off. AOM ratio values suggest an
602 overprint from anaerobic methane oxidisers in one sample in the early Eocene that
603 also has a high BIT index, and 5 samples from the Oligocene where isoGDGTs
604 supposedly received a contribution of anaerobic methane oxidisers (Fig. 4e). As a
605 result, these samples were discarded for TEX_{86} .

Fig. 4



606



607 Figure 4. TEX₈₆ index and overprinting indices. For the equations of these indices see
 608 Table 1. Red line indicates the cut-off value, red crosses indicate samples exceed the
 609 cutoff a. TEX₈₆ (Schouten et al., 2002), with blue line indicating the maximum modern
 610 core-top value (~0.72). b. BIT index (Hopmans et al., 2004). c. fcren' (O'Brien et al.,
 611 2017), no samples discarded; d. Methane index (Zhang et al., 2011), n_{discarded} = 6. e.
 612 AOM ratio (Weijers et al., 2011), n_{discarded} = 6. f. GDGT2/3 (Taylor et al., 2013),
 613 conservative cut-off of 5, n_{discarded} = 4, one sample retained for absence of anomalous
 614 Ring index values (see text). g. Methanogenesis (Blaga et al., 2009), cut-off of 2.0,
 615 n_{discarded} = 8. h. ΔRI n_{discarded} = 18.

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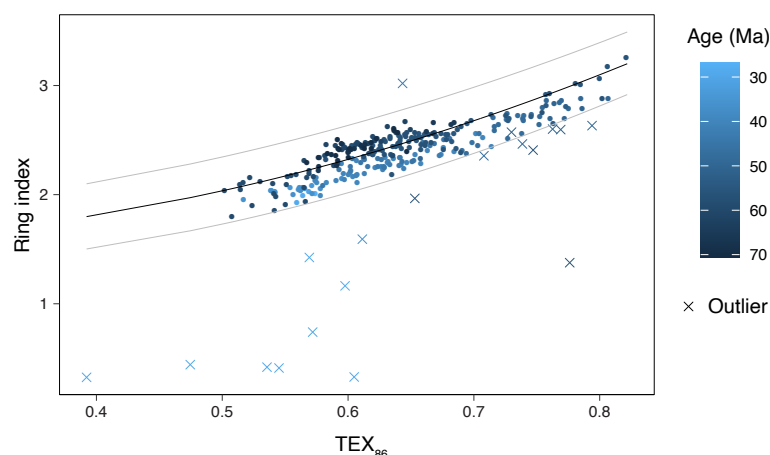
619 Values signalling a deeper water column overprint (GDGT2/3 ratio) fall below
 620 5 for most of the data (Fig. 4f), which was expected given the depositional setting on a
 621 continental shelf. A total of 5 samples has water GDGT2/3 values exceeding 5, 4 of
 622 which also had other indications of overprinted isoGDGT signals. We discard eight
 623 TEX₈₆ datapoints because GDGT-0/crenarchaeol ratios are over 2 (Fig. 4g). One of
 624 these is from the PETM, with a value just above the cut-off. The others represent
 625 Oligocene samples.

626 In total, 18 samples fall outside the range of RI values that characterize
 627 modern pelagic-derived isoGDGT compositions (ΔRI >0.3; Fig. 4h, or 95% confidence
 628 interval; Fig. 5). Nine of these 18 come from the interval >43 Ma and have no other
 629 indications of overprints. These samples fall just outside the 95% confidence interval
 630 limits (Fig. 5) and do follow the overall RI/TEX₈₆ trend, but we here discard them to
 631 be conservative. The other 9 samples with too high ΔRI come from the late Eocene-
 632 Oligocene and have BIT index values over 0.4. Two of these samples have no other
 633 overprint indication, the other 7 have multiple other overprints. One sample has RI
 634 values above the range of modern samples (with no other overprinting indications),
 635 the others have too low RI values. Eight samples with normal RI values do have
 636 overprints based on the other indices: BIT index values slightly over 0.4 (6 samples),
 637 GDGT2/3 ratio values over 5 (1 sample), and AOM ratio over 0.2 (1 sample).
 638 Following Leutert et al. (2020) we retained the samples with normal RI values and



639 high BIT or GDGT2/3 ratio values. After considering all these potential biases, and
 640 retaining those with normal isoGDGT distributions, we discard a total of 19 samples.
 641 Most discarded samples are from the Oligocene and meet multiple overprint criteria.

Fig.5



642
 643 Figure 5. Ring index (RI) versus TEX_{86} . Dots indicate RI/ TEX_{86} values for samples that
 644 are retained, crosses indicate samples that are discarded (see Fig. 4). Color of all
 645 datapoints indicates age of the sample, from Maastrichtian (dark blue) to Oligocene
 646 (light blue). Black and grey curves represent the RI/ TEX_{86} relationship of modern
 647 core top samples, and 95% confidence interval, respectively.

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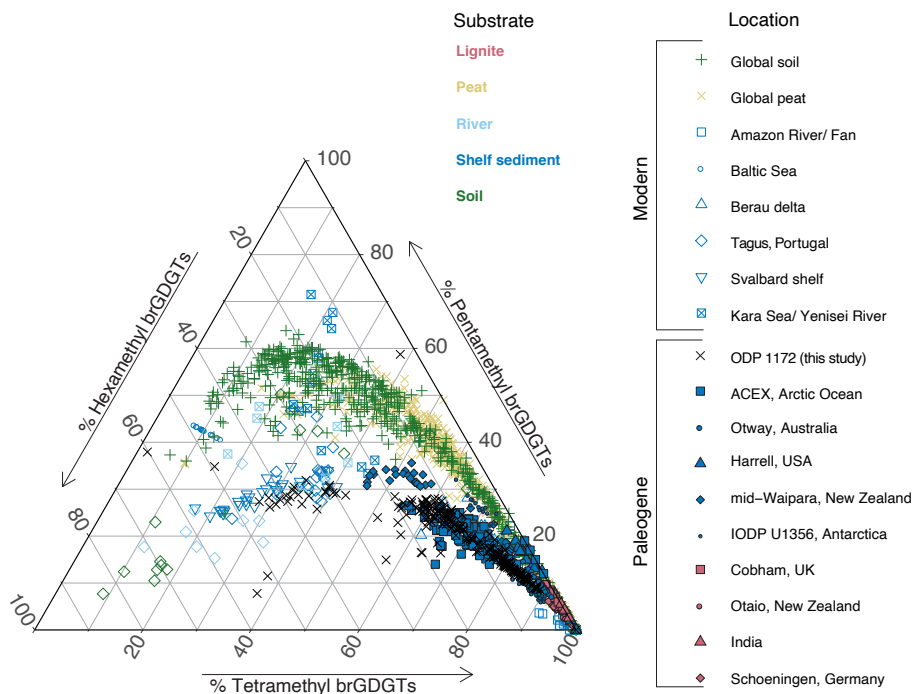
4.1.2 Branched GDGTs

650
 651 Before brGDGT distributions can be interpreted in terms of air temperature,
 652 we assess whether they are primarily derived from soils, or have received
 653 contributions from river-, or marine in situ production. The relative abundance of
 654 tetra-, penta- and hexamethylated brGDGTs in the samples from 1172 deviates from
 655 that in modern and Paleogene soils and peats (Fig. 6): it has either comparatively less
 656 tetra-, less penta-, or more hexamethylated brGDGTs, and the data split into 2
 657 clusters. The largest cluster follows the distribution observed in modern soils and
 658 peats with an offset (Fig. 6). This generally indicates brGDGT contributions from
 659 sources other than soils. The smaller second cluster, containing exclusively mid-to-
 660 late Paleocene samples, is indeed close to that of modern continental shelf sediments



661 (Svalbard and Berau delta; Dearing Crampton-Flood et al., 2019; Sinninghe Damsté,
662 2016) for which in situ brGDGT production substantially exceeds soil contributions.
663 This would generally suggest that the smaller cluster, and perhaps also the larger
664 cluster of samples, have contributions from marine in-situ brGDGT production.
665 However, unlike those modern marine sediments, our entire record does not show
666 elevated #rings_{tetra} values that are associated to marine *in situ*-produced brGDGTs
667 (Fig. 7). The low #rings_{tetra} values on our record (<0.3) suggest brGDGTs have a
668 terrestrial source.
669

Fig. 6

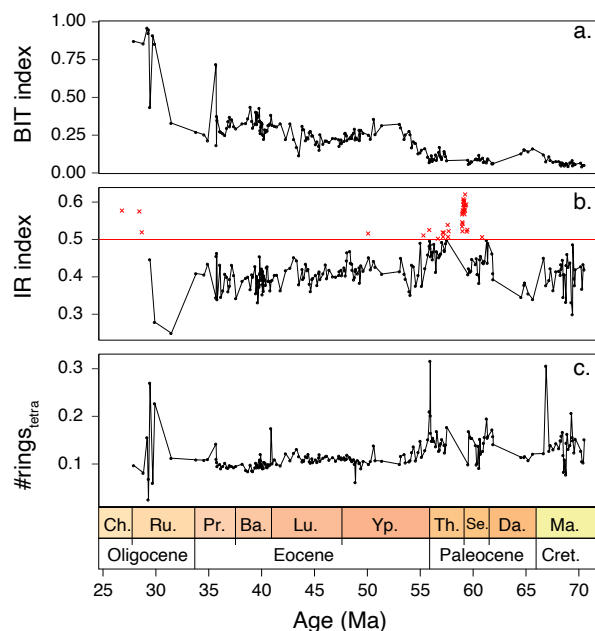


670
671 Figure 6. Ternary diagram showing fractional abundance of tetra-, penta- and
672 hexamethylated brGDGTs from modern soils (Dearing Crampton-Flood et al., 2020),
673 peats (Naafs et al., 2017b), rivers (Tagus River, Portugal (Warden et al., 2016) and
674 Yenisei River (De Jonge et al., 2015)) and marine sediments (Kara Sea (De Jonge et al.,
675 2015), Svalbard (Dearing Crampton-Flood et al., 2019), Berau delta (Sinninghe
676 Damsté, 2016), Portugal (Warden et al., 2016), Baltic Sea (Warden et al., 2018), and



677 Amazon Fan (Zell et al., 2014)), Paleogene lignites (Cobham, Schoeningen, India and
678 Otaio (Naafs et al., 2018b)) and Paleogene marine sediments (Arctic Ocean, IODP
679 Arctic Coring Expedition ACEX (Willard et al., 2019), Otway Basin, Australia
680 (Huurdeeman et al., 2020), Gulf of Mexico coastal plain, Harrel core (Sluijs et al., 2014),
681 New Zealand (Pancost et al., 2013) and Antarctic Margin IODP Site U1356 (Bijl et al.,
682 2013a)). Data from ODP Site 1172 in black crosses.

Fig.7



683
684
685 Figure 7. a. BIT index, b. IR (with cut-off value of the Australian soil IR of 0.5 and
686 discarded MAAT samples in red crosses) and c. #rings_{tetra} from Site 1172.

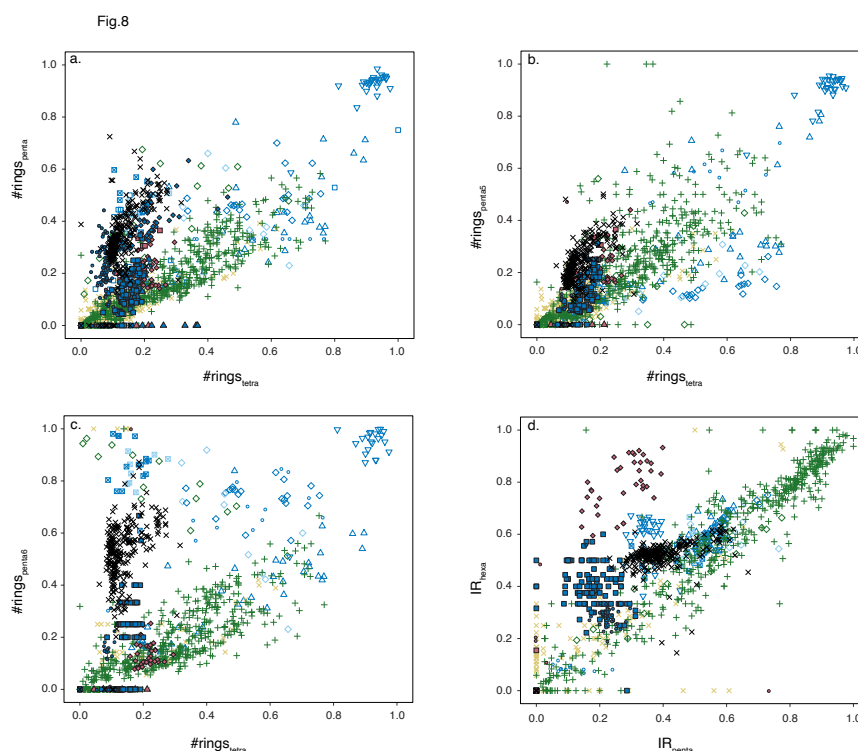
687
688 In the modern soil- and peat brGDGT dataset #rings_{penta} and #rings_{tetra} are
689 below 0.7 and follow a coherent 1:1 correlation (Fig. 8a). Values >0.7 suggest
690 brGDGTs cannot come purely from soils (Sinninghe Damsté 2016). All samples from
691 1172 are below 0.7 for both these indices, which suggests a terrestrial origin is likely,
692 and a strong contribution by marine in situ produced brGDGT is unlikely. We do note
693 that while #rings_{tetra} is below 0.3, #rings_{penta} ranges up to 0.6. This diverts from the



694 1:1 line but brings the brGDGT dataset from Site 1172 within the area of other
 695 Paleogene marine brGDGT datasets (Fig. 8a). This is consistent with the relative
 696 abundances of tetra-, penta- and hexamethylated brGDGTs plotted in the ternary
 697 diagram, where all Paleogene data plotted offset to that in modern soils. #rings_{tetra}
 698 and #rings_{penta} for the 5- (Fig. 8b) and 6-methyl brGDGTs (Fig. 8c) separately shows
 699 that the diversion of the 1172 data from modern soils and peats is primarily caused
 700 by a larger degree of cyclisation of the 6-methyl brGDGTs, suggesting an influence by
 701 river brGDGT production.

702 The samples within the Paleocene cluster do have relatively high IR. The IR in
 703 our record averages ~0.4, but is elevated between 62 Ma and 54 Ma, with peak values
 704 in the mid-Paleocene up to 0.6 (Fig. 7). Although based on only two datapoints,
 705 modern southeast Australian soils have IR values of maximum 0.5 (De Jonge et al.,
 706 2014a). Hence, any values exceeding that of the corresponding hinterland soils could
 707 indicate a contribution by non-soil sources (be it marine or river in situ production;
 708 De Jonge et al., 2014a). Given that #rings indices values argue against a large
 709 contribution from marine in-situ production, we suggest that river-produced
 710 brGDGTs contribute to the GDGT pool in the Paleocene samples with high IR.
 711 Therefore, for MBT'_{5me} calculation, we eliminate all brGDGT samples in our record
 712 that have IR values above that of Australian soils. The IR_{penta} and IR_{hexa} are well within
 713 the range of the modern soil and peat data (Fig. 8d) and settings affected by marine or
 714 river in situ production (Svalbard, Kara / Yenisei), and plot far away from lignite
 715 deposits.

716



717
 718 Figure 8. Crossplots of various brGDGT indices. a. $\#rings_{tetra}$ vs. $\#rings_{penta}$, $\#rings_{tetra}$
 719 vs. $\#rings_{penta5}$, $\#rings_{tetra}$ vs. $\#rings_{penta6}$, IR_{penta} vs. IR_{hexa} . For color and symbol legend
 720 see Figure 6.

721

722 The consistent offset in brGDGT composition in Paleogene marine sediments
 723 relative to modern soils can potentially be explained by non-soil contributions even
 724 though many (most) indices suggest soil-dominated sourcing. Notably, not only
 725 Paleogene marine sediments from Site 1172 are offset from present-day soils, but this
 726 extends to other sites as well, including Paleogene lignites that are not influenced by
 727 marine contributions. We thus argue that there may be an additional evolutionary- or
 728 bacterial community factor that is at least in part responsible for the observed offset
 729 brGDGT distributions during the Paleogene, and argue that, with some exceptions
 730 where indices do suggest non-soil contributions, Paleogene marine sediments contain
 731 a dominant soil-derived brGDGT signal in spite of their offset composition.

732



733

734 4.2 SST and MAAT reconstructions

735 4.2.1 SST

736 By removing all samples with potential overprint from the record we can now
 737 assess the trends in TEX_{86} -based SST (Fig. 9a). Reconstructed SSTs differ to up to 4 °C
 738 between the exponential SST_{exp} and the linear SST_{lin} calibration in the warmest
 739 intervals, and progressively less with lower SSTs (Fig. 9a). The BAYSPAR method,
 740 which uses a linear relation between SST and TEX_{86} generates the lowest SSTs for
 741 the low TEX_{86} values in the Maastrichtian, Paleocene and middle–late Eocene, and
 742 SSTs in between those based on the linear and exponential calibrations for the
 743 highest TEX_{86} index values (Fig. 9a). U^{K}_{37} -based SSTs (Bijl et al., 2010, with
 744 BAYSPLINE calibration) around the MECO are most consistent with the high SSTs
 745 based on the linear TEX_{OBL} calibration, although in this TEX_{86} range all calibrations
 746 yield SSTs within error (Fig. 9a).

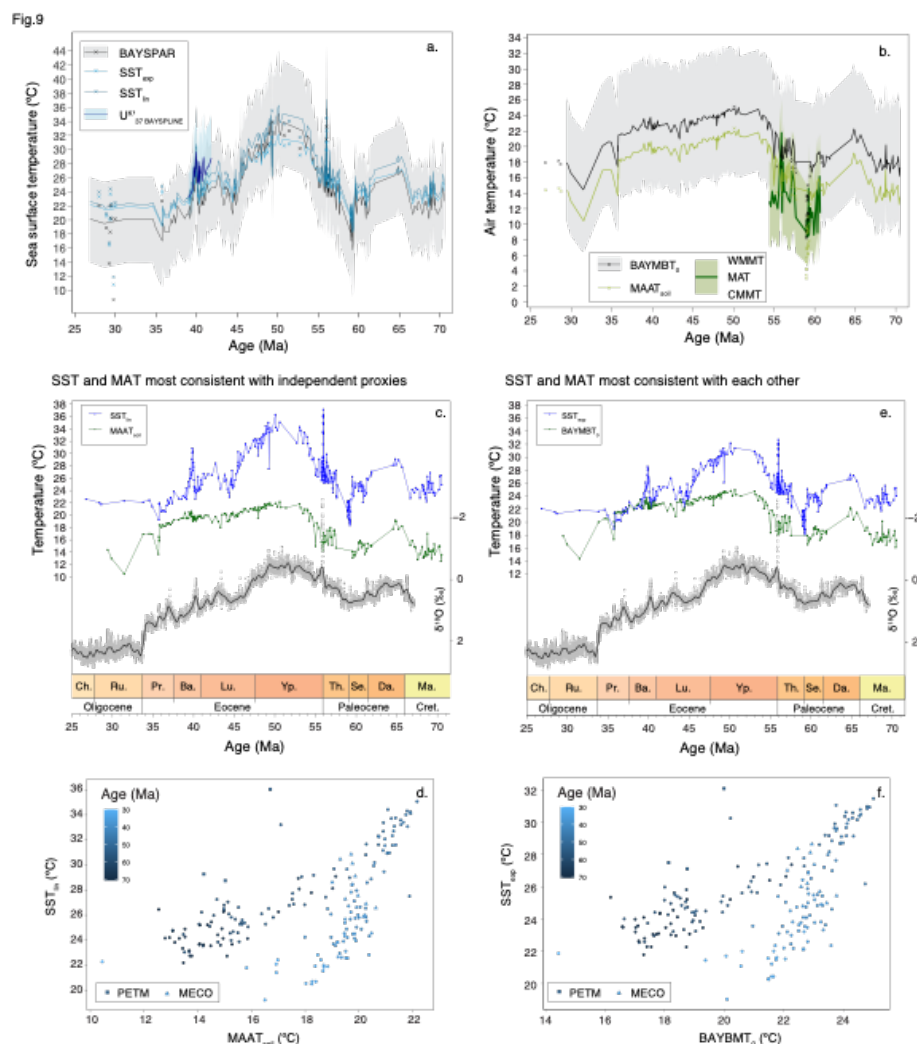
747 In general, Maastrichtian (70.5–66.7 Ma) SSTs show a gradual cooling trend
 748 towards the late Maastrichtian–early Danian hiatus of ~800 kyrs. The early Danian
 749 (~65 Ma) is roughly 3 °C warmer than the late Maastrichtian (~67 Ma). The mid-
 750 Paleocene (62–59 Ma) shows two cooling steps: a ~2 °C cooling at 62 Ma, followed by
 751 a return to higher SSTs roughly at 60 Ma, and a ~5 °C cooling to minimum values of
 752 around 18°C around 59.5 Ma, where the record is truncated by a hiatus.

753 A hiatus straddling the early Thanetian (59.0–57.7 Ma) likely obscures the
 754 onset of late Paleocene warming out of the mid-Paleocene temperature minimum.
 755 Background latest Paleocene SSTs fluctuate considerably, but the PETM warming
 756 clearly stands out (Sluijs et al., 2011), with a magnitude (5–7 °C depending on the
 757 calibration) comparable to that in other mid- to high latitude sections and similar to
 758 the global average (Frieling et al., 2017). Post-PETM SSTs drop back to pre-PETM
 759 values completely, followed by renewed warming towards the EECO. The magnitude
 760 of warming is 6 to 11 °C depending on the choice of calibration. Smaller early Eocene
 761 hyperthermal events, if represented in our record, do not clearly stand out at Site
 762 1172.

763 Highest SSTs of the EECO are slightly lower than peak PETM values in all
 764 calibrations, consistent with other records (Inglis et al., 2020). However, a hiatus



765 between 52.7 and 51.4 Ma may conceal peak EECO temperatures (Hollis et al., 2019;
766 Lauretano et al., 2018). A gradual SST decline commences around 49.2 Ma and
767 terminates the EECO, which is broadly coeval with cooling inferred from benthic
768 foraminiferal isotope records (Lauretano et al., 2018) and proxy records from New
769 Zealand (Dallanave et al., 2016; Crouch et al., 2020).
770



771
772 Figure 9. Paleothermometers. a. TEX₈₆ based SST reconstructions, using a Bayesian
773 (BAYSPAR; black, with 90% confidence interval in grey envelope; Tierney and



774 Tingley, 2015) an exponential (SST_{exp} ; light blue; Kim et al., 2010) and a linear (SST_{lin} ;
 775 dark blue; O'Brien et al., 2017) calibration. Also plotted are $U^{K'_{37}}$ -based SSTs (thick
 776 blue line; from Bijl et al., 2010, recalculated using BAYSPLINE of Tierney and Tingley,
 777 2018). Discarded samples are indicated by the crosses. b. brGDGT-based MAAT
 778 reconstructions using the Bayesian (BayMBT₀; black, with 90% confidence interval in
 779 grey envelope; Dearing Crampton-Flood et al., 2020) and soil (MAAT_{soil}; light green;
 780 Naafs et al., 2017b) calibrations. Discarded samples are indicated in crosses. Also
 781 plotted are pollen-based NLR-approach MAT reconstructions (thick dark green, with
 782 light green envelope representing coldest and warmest month mean temperatures;
 783 from Contreras et al., 2014). c. SST and MAAT reconstructions using the calibrations
 784 that are most consistent with independent proxies from this site (Bijl et al., 2010;
 785 Contreras et al., 2014), with the CENOGRID benthic foraminifer oxygen isotope splice
 786 of Westerhold et al., 2020 (10pnt loess smooth in grey, 500 pnt loess smooth in
 787 black). d. cross-correlation between SST_{lin} and MAAT_{soil}. e. SST and MAAT
 788 reconstructions using the calibrations that are, in absolute temperature, most
 789 consistent with each other, with the CENOGRID benthic foraminifer oxygen isotope
 790 splice of Westerhold et al., 2020 (10pnt loess smooth in grey, 500 pnt loess smooth in
 791 black). f. cross-correlation between SST_{exp} and BAYMBT₀.

792

793

794 Post-EECO cooling is gradual and of small magnitude (2 °C from peak EECO
 795 values) until about 46.4 Ma, after which it accelerates (5 °C) between 46.2 Ma and
 796 44.4 Ma. The age model in this particular part of the sequence is complicated, due to a
 797 problematic assignment of chron C21n (Bijl et al., 2010; 2013b). Following the
 798 minimum SSTs at 44.4 Ma, SSTs rise again towards a plateau at 42.7 Ma. An SST
 799 minimum of ~22 °C is observed just prior to the MECO, at 40.2 Ma, followed by MECO
 800 warming of about 5–7 °C reached at 39.9 Ma. Post-MECO cooling seems more gradual
 801 than MECO warming, although this might be the result of sedimentation rate changes
 802 (Bijl et al., 2010), which are not accounted for in our age model. SSTs are almost
 803 identical to those for the MECO at nearby Site 1170 on the South Tasman Rise
 804 (Cramwinckel et al., 2020). Post-MECO cooling transitions into gradual late Eocene
 805 cooling, down to a minimum of ~19 °C at 35.7 Ma. The following latest Eocene to



806 Oligocene TEX₈₆-based SSTs (35–27 Ma) are 3 °C warmer, with surprisingly no
 807 cooling associated to the Eocene–Oligocene transition (Houben et al., 2019).

808

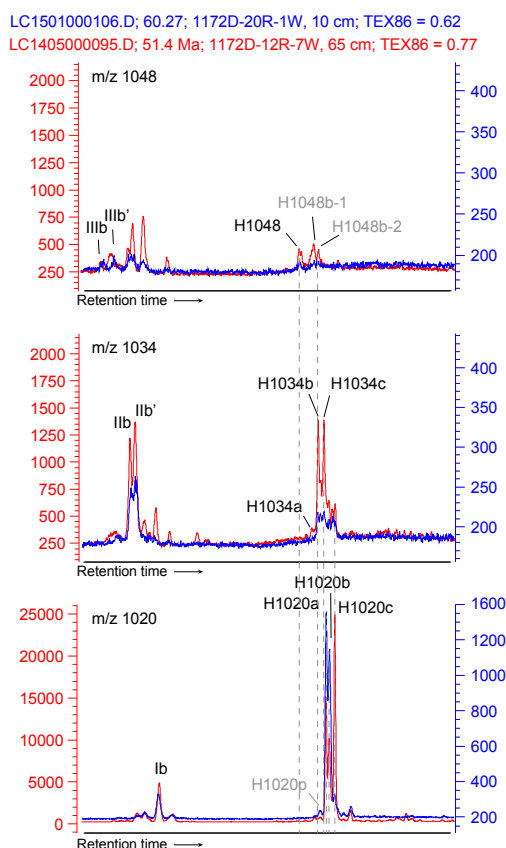
809 4.2.2 MAAT

810 After excluding samples with IR values above those of local soils, we present
 811 MBT'_{5me}-based MAAT reconstructions. These differ substantially (up to 5°C) per
 812 calibration chosen (Fig. 9b). The MAAT_{soil} calibration yield consistently cooler MAATs
 813 (2–3°C) compared to the BAYMBT₀ calibration (Fig. 9b), which may be because the
 814 MAAT_{soil} is calibrated against average temperature of the days above freezing, while
 815 BAYMBT₀ is calibrated against mean annual temperature. Both MAAT calibrations fall
 816 between the MAT and the warmest month mean temperature reconstruction of the
 817 sporomorph nearest living relative (NLR)-based temperature reconstruction
 818 (Contreras et al., 2014), and calibration errors largely overlap.

819 Independent of the calibration chosen, MAAT evolution reveals late
 820 Maastrichtian-early Danian warming of 3°C, gradual 4°C cooling towards the mid-
 821 Paleocene, gradual 8°C warming towards the EECO, with a conspicuous absence of
 822 warming at the PETM. Gradual cooling of 4°C starts at 49.5 Ma and continues into the
 823 Late Eocene. MAAT reaches a minimum at 43 Ma, followed by a plateau-phase. There
 824 is little warming that can be associated to the MECO. MAAT shows a conspicuous
 825 transient 3 °C cooling in the Late Eocene, at the onset of glauconite deposition. The
 826 Eocene-Oligocene Transition is characterized by another 5°C MAAT cooling.



Fig. 10



827

828

829 Figure 10. Stack of 2 UHPLC-MS chromatograms (1172D-12R-7W, 65cm in red and
 830 1172D-20R-1W, 10cm in blue) of m/z 1048, 1034 and 1020 traces, between ~40 and
 831 ~60 minutes retention time, showing the pattern of peaks corresponding to brGDGTs
 832 and brGMGTs. Peak labels refer to the molecular structures in Fig. 3.

833

834

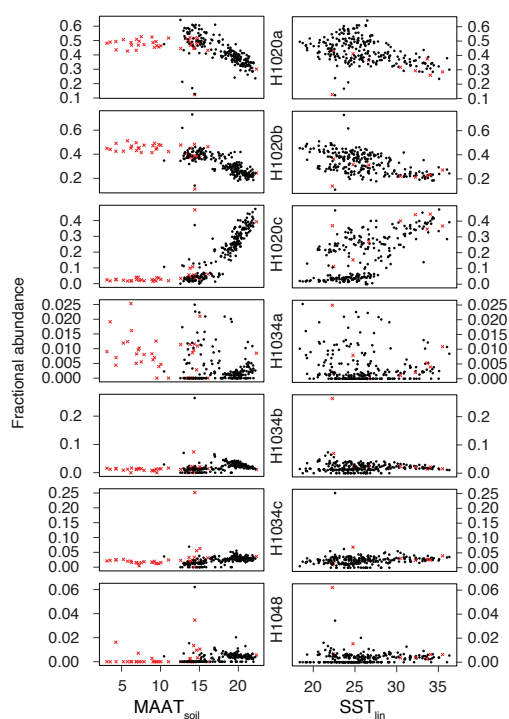
835 4.3 brGMGTs

836 We detected brGMGTs throughout the record (see Fig. 10 for typical
 837 chromatograms). The exact molecular structure and position of the C-C bridge is still
 838 unknown but based on visual comparison with chromatograms of brGMGTs in
 839 African lake sediments (Baxter et al., 2019), all 7 known brGMGTs can be identified.



840 Next to H1020a, b, and c as most abundant brGMGTs, also H1034a, b and c are
 841 detected, although in some samples in the Maastrichtian, individual peaks could not
 842 be separated. Traces of H1048 can be seen at times, suggesting presence of this
 843 compound as well, albeit in low relative abundance. Next to these previously
 844 described peaks, we here recognized several other peaks that consistently occur at
 845 Site 1172: one compound that elutes just prior to H1020a, which we term H1020p
 846 (Fig. 10). ~1 minute after H1048 elutes, two more peaks appear which we here term
 847 H1048b-1 and H1048b-2 (Fig. 10). Because these have so far unknown affinities or
 848 molecular structure, we do not include them here further in our calculations.

Fig. 11



849
 850 Figure 11. Fractional abundances of brGMGTs plotted against BAYMBT₀ and SST_{exp}.
 851

852
 853 The fractional abundances of the H1020 isomers show qualitatively similar
 854 relations to temperature as in the East African lake dataset of Baxter et al. (2019):
 855 H1020b abundance has a negative relationship with MAAT, while H1020c has a

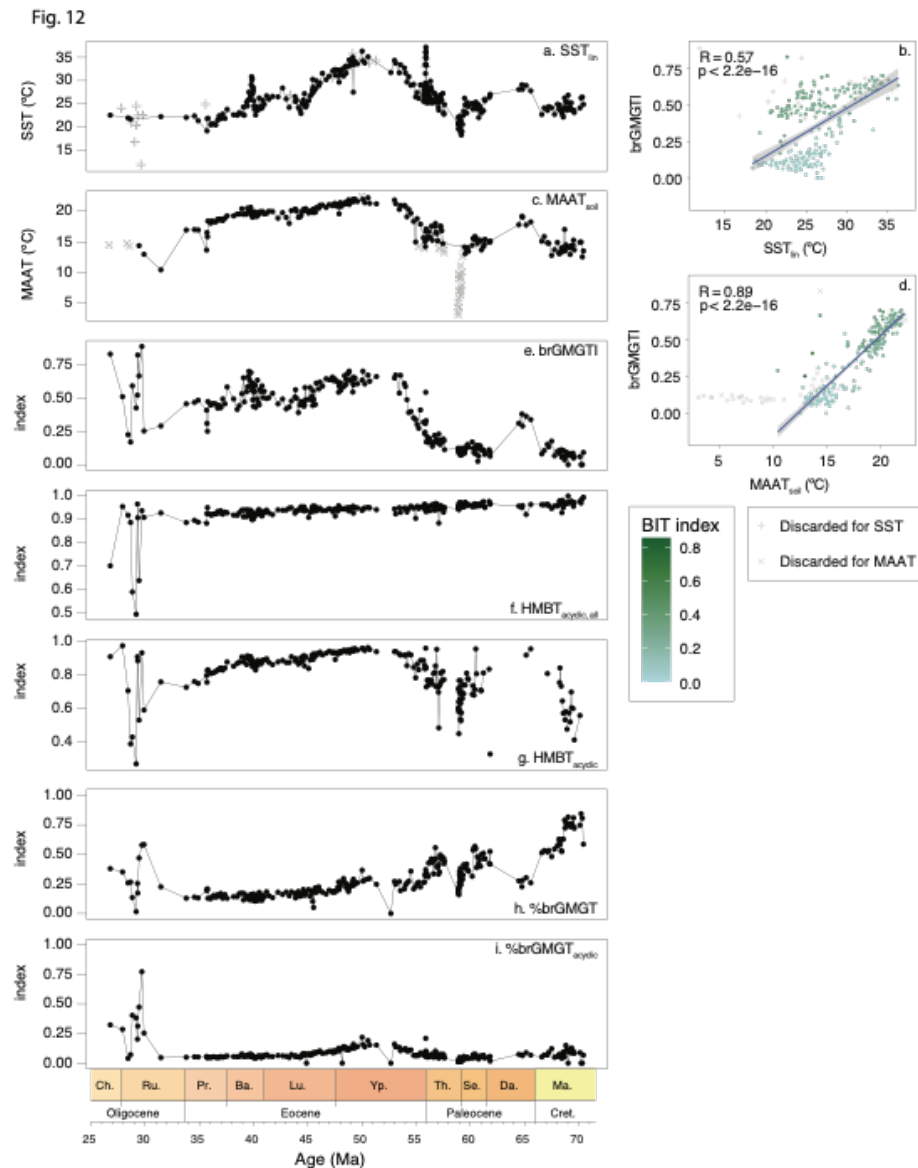


positive relationship (Fig. 11). In addition, H1020a seems negatively correlated with MAAT in our data, although this compound does not show any relation to temperature in lakes (Baxter et al., 2019). The relative abundances of the H1034 isomers do not show a clear relationship with MAAT (Fig. 11), similar as in the tropical lake dataset, where the scatter increases at lower (H1034b) or higher (H1034a, c) temperatures (Baxter et al., 2019).

Downcore trends in brGMGTI are primarily controlled by the relative abundance of H1020c and closely follow those in both SST and MAAT. Upon cross-correlation, we note that brGMGTI has a higher correlation with MAAT ($R = 0.89$) than with SST ($R = 0.57$; Fig. 12). Interestingly, the correlations between brGMGTI and temperature indices only hold for part of the dataset. Samples with high IR values, for which brGDGT-based MAAT could not be reliably interpreted, fall outside the correlation (grey crosses in Fig. 12). This is probably due to the supposed contribution of river-derived brGDGTs that may bias MAAT towards lower temperatures. In addition, we also note that brGMGTI approaches ‘saturation’ (brGMGTI = 0) at reconstructed MAAT of 15°C. The correlation with temperature (MAAT and SST) is divided into 2 clusters. BrGMGTI values for samples from the younger part of the record, with BIT index $< \sim 0.3$, seem to correlate better to temperature than those from the older part of the record, with BIT index values $> \sim 0.3$ (Fig. 12). In the Oligocene, where many isoGDGT signals were evidently overprinted, the %brGMGTs increases (Fig. 3, 12d), and the brGMGT composition is characterized by a relatively higher abundance of H1020c, H1034b and H1048 compared to the rest of the record (Fig. 3; 12).

There is no clear relationship between %brGMGTs and MBT_{5me}-based MAAT (Fig. 12h, i). This slightly differs from the Paleocene-Eocene Arctic Ocean record (Sluijs et al., 2020), where the brGMGTI has no correlation to temperature, while %brGMGTs do. This may indicate differences in sources of brGMGTs between the sites, differences in oceanographic settings, or a non-temperature control on their distribution. HMBT_{acyclic} does seem to show similar trends to the MBT_{5me}-based MAAT (Fig. 12g), like in the Arctic Ocean (Sluijs et al., 2020).

886



887
888 Figure 12. Time series and crossplots of brGMGT-based indices with MAAT and SST
889 reconstructions. (a) SST_{lin} (b) crossplot of SST_{lin} and brGMGTI, (c) $MAAT_{soil}$, (d)
890 crossplot of $MAAT_{soil}$ and brGMGTI, (e) brGMGTI, (f) $HMBT_{acyclic, all}$, (g) $HMBT_{acyclic}$, (h)
891 $\%brGMGT$ (Baxter et al., 2019), (i) $\%brGMGT_{acyclic}$ (Naafs et al., 2018). For equations
892 see Table 1). Colour variable in the crossplots indicate the BIT index value.
893



894 4.4 Palynology

895 4.4.1 Assemblages

896 Dinocyst assemblages are dominated by *Manumiella* spp. throughout the
 897 Maastrichtian, together with Goniodomideae (notably *Dinopterygium* spp.), and
 898 *Cerebrocysta* cpx. (Fig. 13). From the late Maastrichtian onwards, *Senegalinium* cpx.
 899 increases gradually in relative abundance, interrupted by acmes of *Palaeoperidinium*
 900 *pyrophorum* in the Danian and *Glaphyrocysta* cpx. in the lower Selandian. Relative
 901 abundances of *Senegalinium* cpx. reach maximum values during the mid-Paleocene
 902 and during the Paleocene–Eocene transition, while the PETM itself is characterised by
 903 abundant *Apectodinium* (~30%) and common *Senegalinium* cpx. and Goniodomideae
 904 (Sluijs et al., 2011). After the PETM, *Senegalinium* cpx. abundances temporarily
 905 increase, and are then replaced by *Deflandrea* cpx, *Spinidinium* cpx and *Elytrocysta*
 906 spp. and low abundances of various other genera, e.g., *Wetzeliella*, *Hystriocholpoma*,
 907 Goniodomideae (Fig. 13). From 45 Ma to ~37 Ma, *Enneadocysta* spp. alternates
 908 dominance with *Deflandrea* cpx. and *Spinidinium/Vozzhennikovia* cpx, with
 909 contributions of *Phthanoperidinium* spp. (Röhl et al., 2004a). Protoperidinioid
 910 dinocysts appear commonly in the late Eocene, around 35.5 Ma (Houben et al., 2019;
 911 Sluijs et al., 2003; Fig. 13).

912

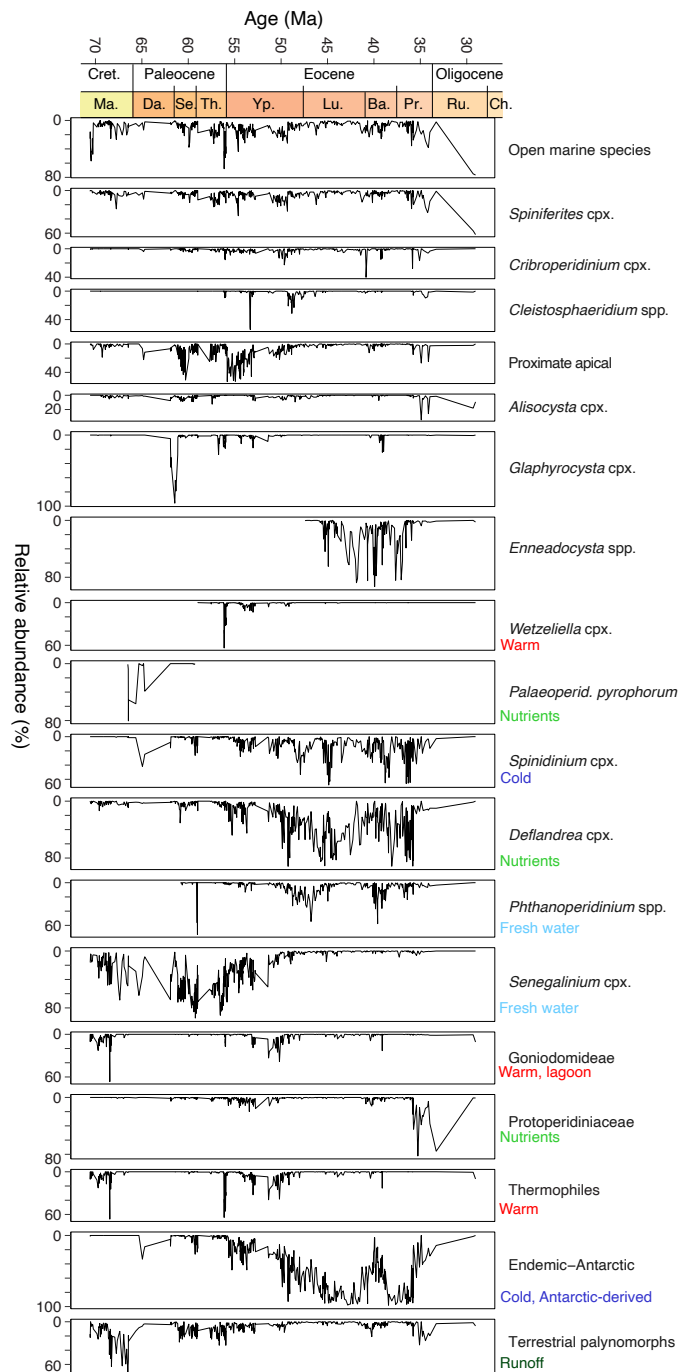
913 4.4.2 Diversity and variability

914 Together with the changing assemblage composition are some strong and
 915 coherent changes in all diversity and variability indices (Fig. 14). First, results are
 916 highly similar for the full species-level dataset and the grouped assemblages,
 917 indicating robustness of both the grouping and diversity analysis. The Maastrichtian–
 918 Paleocene assemblages are characterized by relatively low diversity, which increases
 919 towards the EECO, as variability indicated by the Gibbs index decreases. After this,
 920 diversity drops and variability increases during middle–late Eocene cooling ~49–38
 921 Ma, with an interruption around 41–39 Ma. Towards the top of the record, both
 922 diversity and variability increase.

923



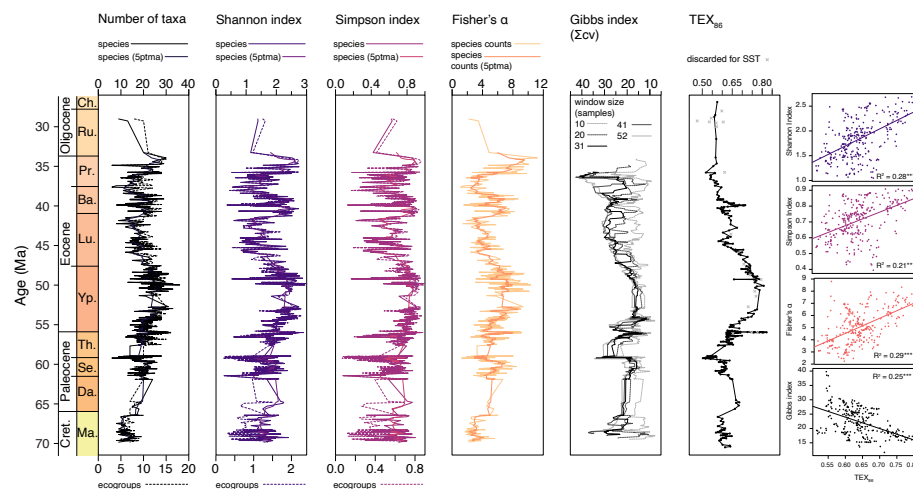
Fig. 13





925 Figure 13. Dinocyst ecogroups and %terrestrial palynomorphs from Site 1172. From
926 left to right, ecogroups are ordered in an in-shore-to-offshore transect. Relative
927 abundance ticks represent 20%.
928

Fig. 14



929
930 Figure 14. Dinocyst taxonomic diversity and variability through time. From left to
931 right are plotted number of taxa (black), Shannon index (purple), Simpson index
932 (pink), Fisher's alpha (orange) (species-based dataset in solid lines; ecogrouped
933 dataset in dashed lines) and Gibbs index (different window sizes in gray-black). TEX₈₆
934 plotted for comparison. Crossplots show correlation of these indices with TEX₈₆.
935

936 5. Discussion

937 5.1 Air and sea surface temperature evolution compared

938 The near-shore depositional setting of our record should have captured a
939 coastal terrestrial environment with similar MAATs similar to the local SSTs. Indeed,
940 the brGDGT-based MAAT record shows strong similarities to SST particularly in
941 multi-million year trends (Fig. 9). Specifically, the early Danian is warmer than the
942 Maastrichtian, and both records show a 2-step Paleocene cooling, late Paleocene
943 warming towards peak values in the EECO, and middle-late Eocene cooling. On
944 shorter time scales both records, occasionally even on sample-to-sample level, show
945 synchronous variability (Fig. 9c, d). The independent SST (based on alkenones; Bijl et



al., 2010) and air temperature reconstructions (based on pollen assemblages; Contreras et al., 2014) show the best fit with the absolute SST values derived from the linear calibration for TEX_{86} (BAYSPAR or SST_{lin} ; generating the highest SST estimates), and MAATs based on the calibration using Deeming regression ($MAAT_{soil}$), respectively. These are the two calibrations with the largest absolute temperature offset, particularly in the warm EECO (Fig. 9c). If the local terrestrial climate in the nearby river catchment is largely controlled by ocean temperature, the absolute mean annual SST and MAAT should be broadly similar in the integrated signal of our samples. This could be an argument to choose transfer functions for SST and MAAT whereby absolute temperature estimates overlap most: SST_{exp} and the Bayesian $BAYMBT_0$. These generate overlapping absolute temperatures for the colder middle-late Eocene and Paleocene, but in the warmer Eocene time intervals SSTs remain consistently warmer (by $\sim 8^\circ C$). The lower MAATs cannot be explained by a saturation of the MBT'_{5me} index that is underlying the paleothermometer, as maximum recorded MBT'_{5me} values are around 0.85 in the EECO. Accounting for the calibration errors of SST ($\pm 2.5^\circ C$) and the large calibration error bar of $BAYMBT_0$ in principle could resolve the offset but applying the extremes in calibration error to TEX_{86} and MBT'_{5me} to make them consistent would make both profoundly inconsistent to the other independent SST and MAAT proxies from the same samples. More importantly, the colder MAAT compared to SST seems to be a consistent feature in many regions where TEX_{86} and MBT'_{5me} were applied together (see e.g., Hollis et al., 2019 and compare Willard et al., 2019 and Sluijs et al., 2020). The offset between MAAT and SST would have been more variable between sites if it were only the result of calibration errors. TEX_{86} -based SSTs in the southwest Pacific realm have been on the high end of many multi-proxy-based temperature reconstructions for the Eocene (Hollis et al., 2012), and therefore may be assumed to have an as yet poorly understood warm bias. However, the consistency of TEX_{86} -based SSTs with other SST proxies for specific time intervals and locations (Bijl et al., 2010; Hines et al., 2017) suggests the SST-MAAT offset cannot be easily reconciled by only invoking a warm bias in TEX_{86} . MAAT reconstructions on the other hand have been broadly consistent with nearest living relative based temperature reconstructions on pollen assemblages in many applications in the Eocene (Contreras et al., 2013; Contreras et al., 2014;



978 Pancost et al., 2013; Pross et al., 2012; Willard et al., 2019), but both brGDGT and
 979 plant-based temperature estimates are arguably prone to cold biases at the high-
 980 temperature end (e.g. Naafs et al. 2018, Van Dijk et al., 2020). In addition, it may be
 981 that the same evolutionary or bacterial community factors that make Paleogene
 982 brGDGT assemblages deviate from that in modern soils, may also cause a deviation in
 983 the calibration to MAAT. This may resolve some of the offset between MAAT and SST
 984 reconstructions. Although this implies that quantitative MAAT estimates based on
 985 MBT'_{5me} in non-analogue settings such as the present one should be taken with care,
 986 the strong temperature dependence in Paleogene brGDGTs cannot be ignored.

987 Another surprising result is that the brGDGT-based MAAT record does not
 988 capture PETM and MECO warming (Fig. 9c), independent of the calibration chosen.
 989 This is remarkable for several reasons: (1) brGDGTs were measured on the same
 990 samples as isoGDGTs, which precludes a sampling bias; (2) SST changes of longer
 991 duration (Maastrichtian–Danian warming, mid-Paleocene cooling and early Eocene
 992 warming towards the EECO) are represented in the MAAT record; (3) The duration of
 993 PETM (~150-250 kyrs; Röhl et al., 2007, Zeebe & Lourens, 2019) and MECO (~400
 994 kyrs; Bohaty et al, 2009) is too long to explain the absence in MAAT warming with a
 995 lag in soil-derived OM delivery to the ocean (up to several kyr, see, e.g., Feng et al.,
 996 2013; Schefuß et al., 2016; Huurdeman et al., 2020); (4) seems incompatible with the
 997 fact that other low-amplitude, shorter-term SST changes are reflected in the MAAT
 998 record (Fig. 9c, d), which suggests that soil-derived OM did capture short-term
 999 climate variability on the time resolution of the samples; (5) Other PETM records in
 1000 the region do show a temperature response in the PETM in various proxies including
 1001 brGDGTs (Hollis et al., 2012; Pancost et al., 2013, Huurdeman et al. 2020). A lack of
 1002 MAAT rise during the PETM and the MECO could be explained by a switch from
 1003 brGDGT sourcing from soils to (peaty) lakes, which are notoriously cold-biased
 1004 (Tierney et al., 2010). Paleocene–Eocene peats are abundant in southeast Australia
 1005 (Holdgate et al., 2009), and it is possible that a contribution from peat lakes at times
 1006 of global warming, contributed to a dampening of the proxy-response. This may also
 1007 explain the lack of response at the EECO and explain why MBT'_{5me} does not reach
 1008 saturation in that interval. The final option, which, although unlikely, cannot be



1009 excluded, is that a dominant fraction of the brGDGT was produced in situ and subduces
 1010 the temperature response during these warming phases.

1011 The subdued middle-late Eocene cooling in the air temperature record
 1012 compared to the strong decrease in SST is particularly clear when cross-plotting SST
 1013 *versus* MAAT (Fig. 9c, d): the relationship between the two proxies is different in the
 1014 Maastrichtian–early Eocene compared to the middle–late Eocene. This may be related
 1015 to the start of a progressive northward tectonic drift of the Tasmanian hinterland,
 1016 which occurred around the same time (Fig. 1). This puts the hinterland of the soil-
 1017 derived brGDGTs into warmer climate zones throughout the middle–late Eocene,
 1018 while the ETP remained under influence of the Antarctic-derived Tasman Current
 1019 (TC) through that time (Huber et al., 2004). The TC cools and likely strengthens in the
 1020 middle and late Eocene. The terrestrial climate cooled as well, but this signal will be
 1021 subdued if the soil material is sourced from a progressively lower-latitude
 1022 environment as Australia drifted northward. Yet, one would expect that the close
 1023 coupling between land- and seawater temperature in near-shore environments
 1024 would not capture such tectonic effects.

1025 The correlation between both temperature proxies changes once more during
 1026 the late Eocene–Oligocene, when they show a sharp cooling of 2–3 °C at the onset of
 1027 greensand deposition, at ~35.5 Ma. Whereas this is merely a continuation of
 1028 progressive late Eocene cooling in the SST record, the decrease in MAAT is strongly
 1029 accelerated compared to middle–late Eocene trends. Another 3–4 °C MAAT drop
 1030 occurs sometime between the late Eocene and the early Oligocene (likely associated
 1031 with the Eocene–Oligocene transition), while SSTs returns to warmer, middle Eocene
 1032 values in this interval. The continued sea surface warmth across the EOT has been
 1033 related to persistent influence of the proto-Leeuwin Current (Houben et al., 2019),
 1034 which apparently retained a constant temperature across the EOT. Since MAAT
 1035 decreased while SSTs remained high, the influence of regional oceanographic changes
 1036 did not impact climate changes in the source region of brGDGTs during this time
 1037 interval. This is surprising given the proximity of the sediment record to the coastline.

1038 Minimum mid-Paleocene SSTs are lower than those for the early Oligocene.
 1039 However, the site migrated northward by ~7 ° of latitude between the mid-Paleocene
 1040 and the early Oligocene (Fig. 1) and the oceanographic regime changed with



1041 throughflow of the proto-Leeuwin Current through the TG in the Late Eocene
 1042 (Houben et al., 2019). More remarkably, the mid-Paleocene SSTs approach those of
 1043 the warmest interglacial intervals of Oligocene Wilkes Land Margin, east Antarctica
 1044 (Hartman et al., 2018), at a similar paleolatitude as the Paleocene ETP. This suggests
 1045 that mid-Paleocene Antarctic-proximal temperatures were similar to those at times of
 1046 major Antarctic glaciation during EOT. Previous work has indeed presented
 1047 widespread evidence for low sea level during this time interval (Frieling et al., 2014;
 1048 2018; Guasti et al., 2006; Hollis et al., 2014), which combined with low SSTs suggests
 1049 the presence of some continental ice on the Antarctic continent during this mid-
 1050 Paleocene interval.

1051
 1052

1053 5.2 BrGMGTs

1054 The continuous presence of brGMGTs in our record appears unrelated to
 1055 depositional conditions or varying relative contributions of terrestrial material,
 1056 suggesting that at least part of the brGMGTs have a marine origin. This confirms
 1057 previous observations from the modern sediments and water column (Liu et al. 2012,
 1058 Xie et al. 2014) and Paleogene Arctic Ocean sediments (Sluijs et al., 2020). Their
 1059 sparse presence in modern soils (Peterse et al., pers comm based on metadata from
 1060 De Jonge et al., 2019; Kirkels et al., 2020) is also in line with this, although abundant
 1061 occurrence in peats and lakes (Naafs et al., 2018, Baxter et al. 2019, Tang et al., 2021)
 1062 shows clear potential for terrestrial input. Despite the uncertainties in sourcing of
 1063 brGMGTs, we find a strong resemblance with the signals from tropical lakes. The
 1064 brGMGTI, which was derived to quantify the temperature relation of brGMGTs in
 1065 surface sediments of East African lakes (Baxter et al., 2019), seems to correlate to
 1066 temperature in our record as well (Fig. 12a–d), although the better correlation of
 1067 brGMGTI with MAAT than with SST (Fig. 12a–d) seems to be at odds with the
 1068 presumed marine source of brGMGTs in our record. Despite all potential
 1069 complications, our study provides additional evidence that a temperature signal may
 1070 be governing the distribution of the different brGMGT isomers, as proposed by
 1071 (Baxter et al., 2019), but there are differences as well. For instance, H1020c is
 1072 altogether absent in the cold mid-Paleocene part of our record. There are numerous



ways in which this can be explained, e.g., the biosynthesis of H1020c occurs only above a certain temperature, or the microbes that biosynthesize H1020c do not live at low temperatures, or the site receives brGMGTs from a different catchment in colder climates. Indeed, H1020c was almost absent in the cold ($<12^{\circ}\text{C}$), high-altitude tropical lake samples (Baxter et al. 2019). However, those cold lakes did have high relative abundance of H1048 (Baxter et al., 2019), which our record lacks in the cold mid-Paleocene, illustrating the differences between modern lakes and our marine depositional setting.

The HMBT record does not reflect a temperature signal when using all brGMGT isomers in the calculation (Fig. 12f), as H1020 isomers are by far the most abundant brGMGTs at Site 1172, and any change in the abundance of H1020c is compensated by both H1020b and H1020a in our record. In contrast, only using the H1020c and H1034b isomers, which following the chromatographic peak identification of Sluijs et al. (2020) match the compounds used to derive the HMBT index based on brGMGTs in peats (Naafs et al., 2018), does show a correlation to temperature in our record (Fig. 12g).

%brGMGT as calculated in Naafs et al., (2018) or as in Baxter et al., (2019) does not show a clear relation with temperature (Fig. 12h, i), in contrast to observations in modern peats (Naafs et al., 2018a) and lake sediments (Baxter et al., 2019). We do note the extremely high %brGMGT ($\sim 80\%$) in the Maastrichtian and Oligocene, much higher than seen anywhere so far. We also do not see a clear inverse relationship between %brGMGT and BIT index, as in Sluijs et al. (2020). This may be because of the complexity in the relative abundance of brGDGTs in the record, which does not necessarily reflect the variation in terrestrial versus marine sourcing.

The higher relative abundance and change in composition of brGMGT composition in the Oligocene that coincides with abundant signs of overprints in all br- and isoGDGT indices may indicate a source change of brGMGTs in this interval, e.g., related to the same changes in contribution that cause the overprints in isoGDGTs. The differences in brGMGT signals between our record, the Eocene Arctic Ocean record and modern sediments demonstrate that more research is needed on their sources and environmental dependencies to further assess their use and value in paleoreconstructions.



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1106

1107 5.3 Integrated paleoenvironmental and -climatological reconstruction

1108 5.3.1 Sea level and ecosystem response to SST changes

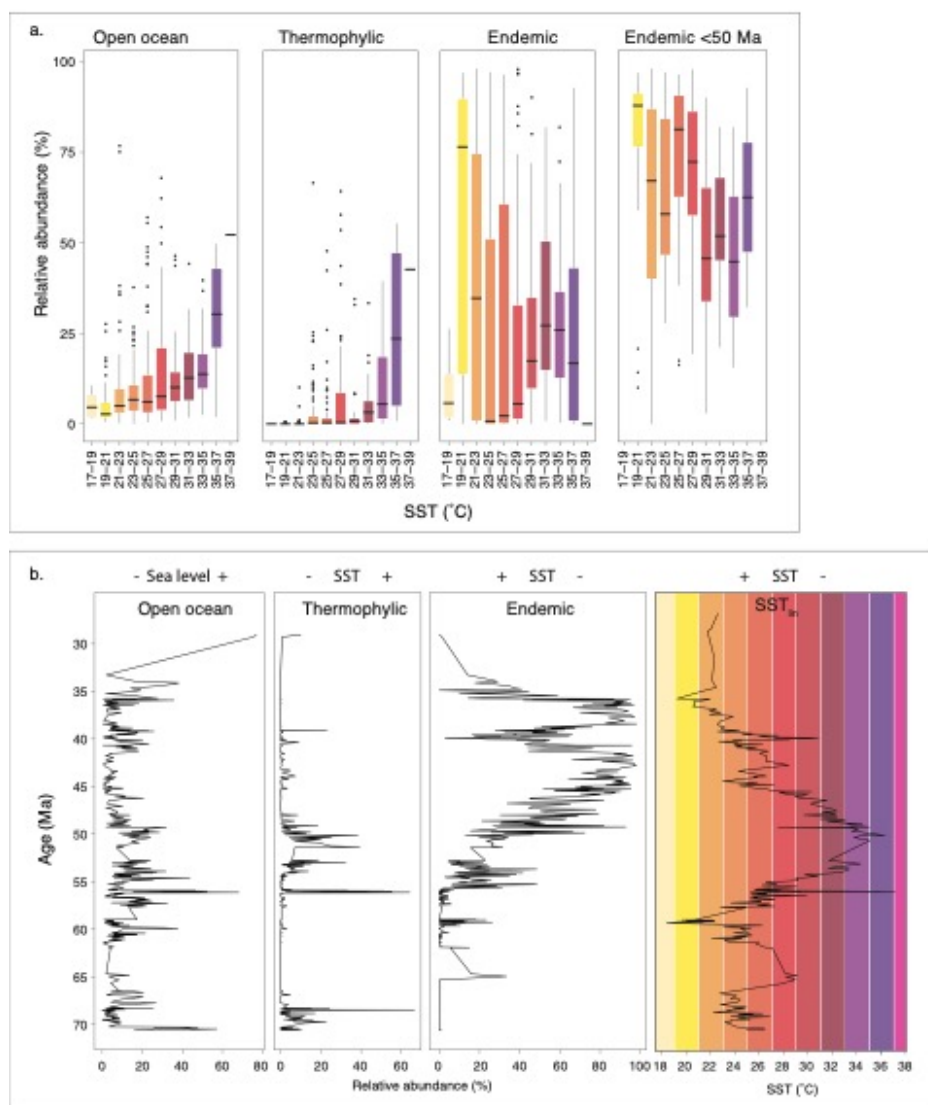
1109 The abundance of temperature-responsive dinocyst eco-groups qualitatively
 1110 confirm our SST trends. The thermophile dinocyst eco-group shows higher relative
 1111 abundance at higher SSTs (Fig. 15). This eco-group has been empirically associated to
 1112 SST based on its strong affinity to high SST in the global compilation of late
 1113 Paleocene-early Eocene records (Frieling and Sluijs, 2018), and we here confirm this
 1114 relationship for the longer early Paleogene. The relationship between isoGDGT-based
 1115 SSTs and relative abundance of the supposedly cold-affiliated endemic dinocysts (Bijl
 1116 et al., 2011) is not as clear. This might be because the majority of dinocyst species
 1117 within the Endemic-Antarctic dinocyst ecogroup are restricted to the Eocene (Bijl et
 1118 al., 2011).

1119 Indeed, from 50 Ma onwards, SST does have a correlation with the %endemic
 1120 Antarctic dinocyst abundance in the record (Fig. 15; Bijl et al., 2011), which may
 1121 indicate a more distinct biogeographical separation through the development of
 1122 stronger meridional gradients or evolutive adaptation of the endemic dinocyst
 1123 community to more polar environments following the extreme warmth of the EECO.
 1124 We particularly note the strong correlation between TEX₈₆-based SST and relative
 1125 abundance of more open ocean-affiliated dinocysts (Fig. 13), which suggests that SST
 1126 correlates with regional sea level. SST-induced sea level changes hint at steric effects
 1127 as driver. Indeed, because of the flatness of continental shelf areas during long-term
 1128 greenhouse climates (Somme et al., 2009) small changes in regional sea level will
 1129 cause large changes in distance to shore and associated characteristics, including e.g.
 1130 salinity, nutrients and suspended sediment loads, which subsequently shapes the
 1131 dinocyst assemblages.

1132



Fig.15
 GDGT- and palynology-based proxies for SST



1133
 1134 Figure 15. isoGDGT- and palynology-based proxies for SST. a. relative abundance of
 1135 open ocean, thermophylic and endemic-Antarctic dinocysts (total dataset and
 1136 samples <50Ma only) plotted against linearly interpolated (resampled) TEX₈₆-based
 1137 SST, in 2 degree bins. b. Time series of the same dinocyst ecogroups and SST.
 1138



1139 The SST control on dinocyst assemblages is further demonstrated in diversity
 1140 and assemblage variability indices. Throughout the long time interval covered in our
 1141 record, dinocyst richness and diversity show a remarkably similar pattern to TEX₈₆-
 1142 based SST. Species richness, as well as diversity expressed in the Shannon and
 1143 Simpson indices and Fisher's alpha reach a maximum during the prolonged warmth
 1144 of the EECO and drop during middle-late Eocene cooling. These relationships are
 1145 demonstrated by modest correlation coefficient between diversity and TEX₈₆ (Fig.
 1146 13). This pattern holds for the long-term trends, as well as for the short-term PETM,
 1147 but not for the MECO. Dinocyst diversity is known to vary in an inshore-to-offshore
 1148 transect in the modern and Eocene (Brinkhuis, 1994; Pross and Brinkhuis, 2005). As
 1149 our assemblages indeed consist of many mid-shelf species, the strong correlation
 1150 between diversity and SST might thus be indirectly related through habitat size, with
 1151 expansion of the flat shelf area and thereby increase in shelf niches during the EECO
 1152 (Somme et al., 2009). Variability as indicated by the Gibbs index records the opposite
 1153 pattern to diversity for the long-term trends into and out of the EECO (Fig 14). That
 1154 is, the lower-diversity assemblages in general have a higher variability, or sample-to-
 1155 sample fluctuations within the relative abundances. This indicates the higher
 1156 diversity dinocyst assemblages might be more ecologically resilient.

1157 5.3.2 Marine environmental response to runoff changes

1158 *Senegalinium* cpx. abundance broadly correlates with the IR (Fig. 16), which
 1159 signals input of river-produced brGDGTs. *Senegalinium* cpx. is generally thought to
 1160 have tolerated low salinity environments (Barke et al., 2011; Sluijs et al., 2005; Sluijs
 1161 and Brinkhuis, 2009). These two proxies together confirm a large input of fresh water
 1162 at this site during the mid-to-late Paleocene in line with high relative abundance of
 1163 terrestrial palynomorphs, low abundances of open ocean dinocysts and high IR values
 1164 (Fig. 16). Interestingly, *Phthanoperidinium* spp., which is also generally associated
 1165 with low-salinity to near fresh-water conditions (Barke et al., 2011; Frieling and
 1166 Sluijs, 2018; Sluijs and Brinkhuis, 2009) does not correlate to IR (Fig. 16). Because we
 1167 find different species of *Phthanoperidinium* spp. than those previously associated to
 1168 fresh-water conditions (e.g., Barke et al., 2011), it could be that the fresh-water
 1169 tolerance of *Phthanoperidinium* spp. was not shared among all species. It could also
 1170 be that *Phthanoperidinium* spp. is slightly less euryhaline than *Senegalinium*.



1171 Tasmania was located at latitudes between 55° and 60° S, in the middle of the
1172 region of strong westerly winds, within the range of low atmospheric pressure and
1173 received a large amount of precipitation during the Paleocene (Huber and Caballero,
1174 2011). River input from Tasmania and perhaps also Australia could have reached the
1175 site. Interestingly, while Tasmania migrated northwards only a few degrees latitude
1176 between 60 and 40 Ma (Fig. 1), the freshwater input decreased, based on both
1177 *Senegalinium* cpx., terrestrial palynomorph abundance (decimated abundance from
1178 50 Ma onwards) and the IR (decrease to normal values around 54 Ma; Fig. 16). This
1179 suggests either a rerouting of river input or a drying of the hinterland. Support for the
1180 latter comes from clay mineralogical data, showing an increase in kaolinite starting at
1181 50 Ma (Fig. 2; Robert, 2004). Kaolinite forms abundantly at the base of acidic peats
1182 (Staub and Cohen, 1978; Korasidis et al., 2019) and in old, leached soils, which were
1183 ubiquitous in the Eocene Australian hinterland (Holdgate et al., 2009). While the
1184 hinterland could well retain the kaolinite-rich clays during the Maastrichtian and
1185 Paleocene, because the site was under a year-round high precipitation regime, it was
1186 less efficient in doing so when the hinterland experienced drier, more variable
1187 climatic conditions as Australia migrated northward. The increase in kaolinite
1188 delivery to Site 1172 is hence interpreted as a signal of enhanced soil (or peat)
1189 erosion from the catchment areas in Tasmania and SW, as a result of a more variable
1190 climate regime.

1191

1192 5.4 Environmental and climatological changes in the catchment

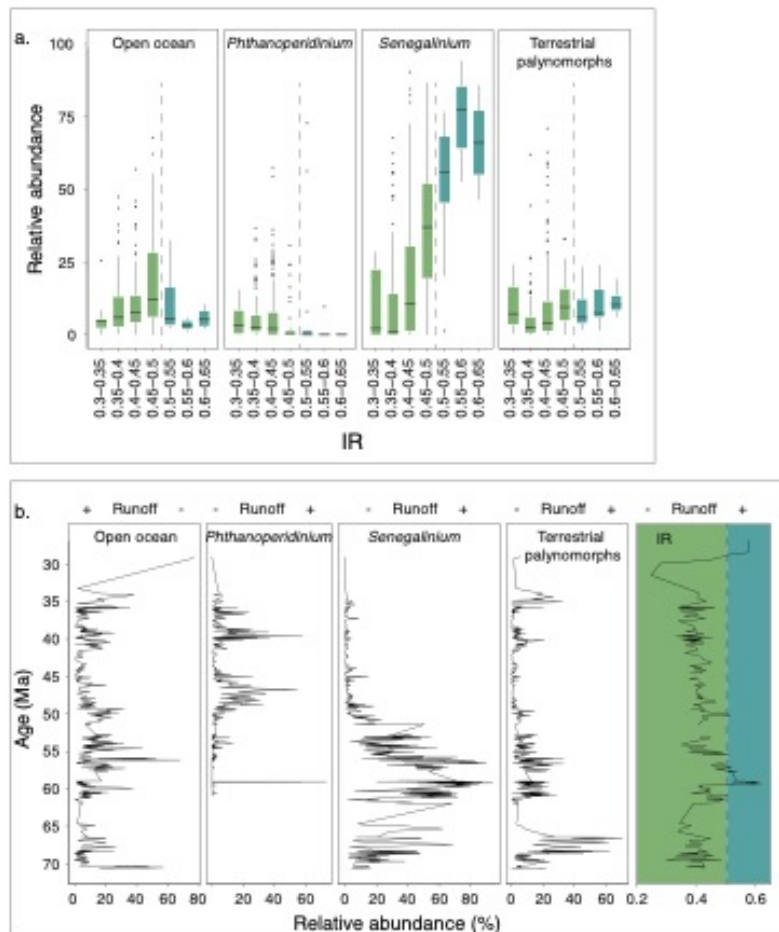
1193 The long-term trends in the BIT index are at odds with all the river runoff
1194 indicators in our data. BIT remains low during the presumed interval with increased
1195 river water discharge, as indicated by the high IR and abundance of *Senegalinium* and
1196 increases when this river-input signal ceases. This is unexpected, because the BIT
1197 index should reflect high input of soil-derived brGDGTs into the marine sediments,
1198 associated with increased discharge (Hopmans et al., 2004). There are two ways to
1199 explain this signal. The BIT index could be responding to marine crenarchaeol
1200 production, rather than to the terrestrial brGDGT flux. Indeed, marine productivity
1201 could have been spurred by runoff and associated nutrient delivery to the coastal
1202 zone during the Paleocene, and decreased in the Eocene, affecting BIT index values.



1203 Another explanation is that the BIT index has been influenced by crenarchaeol
1204 production in the river (Zell et al., 2013), although the river-produced crenarchaeol
1205 concentrations are normally negligible compared to that produced in the marine
1206 realm. Decreasing BIT index values with increasing river-and soil input has been
1207 related to catchment dynamics in the late Quaternary Zambezi river (Lattaud et al.,
1208 2017), involving vegetation dynamics controlling soil-OM input. Studies on modern
1209 soil-river-ocean pathways in the Amazon River system show that whereas river-
1210 suspended brGDGTs in its upper tributaries reflect catchment soils (Kirkels et al.,
1211 2020), brGDGTs in the lower Amazon receive increasing contributions of in situ
1212 produced brGDGTs (Zell et al., 2013), and marine surface sediments of the Amazon
1213 fan contain a mix of soil, riverine, and marine brGDGTs (Zell et al., 2014). In a high-
1214 precipitation, soil-rich environment like the Amazon basin, this can only occur if the
1215 river-produced brGDGT production progressively exceeds the soil-derived brGDGTs
1216 input in river water along the trajectory of the river, or when the soil-derived
1217 brGDGTs are preferentially remineralized during river transport. We here propose
1218 the following scenario to explain the BIT index trends: In the Paleocene–early Eocene,
1219 excessive, year-round precipitation-fed river runoff caused strong fresh-water
1220 delivery into the sw Pacific. Soil-derived brGDGTs were partly remineralized in the
1221 river catchment, and the strongly diluted soil-derived OM was dominated by river-
1222 produced brGDGTs and crenarchaeol, both in river production and on the continental
1223 shelf, which suppressed the BIT index. This explains the unexpectedly cold MAATs
1224 derived from the mid-Paleocene samples with high IR. The high riverine input did
1225 promote pelagic isoGDGT production: crenarchaeol which decreases the BIT index
1226 and isoGDGTs in general. In the early Eocene, precipitation in the hinterland
1227 decreased as the hinterland gradually drifted out of the zone of intense precipitation.
1228 This reduced river discharge, but at the same time the more dynamic climate regime
1229 caused more seasonal precipitation which increased destabilization of Australian and
1230 Tasmanian soils, leading to a higher abundance of soil-derived brGDGTs and kaolinite
1231 from destabilized soils in the river discharge, and relatively little river-produced
1232 brGDGTs (because enhanced soil-OM content increased turbidity) and river-produced
1233 crenarchaeol, which increased BIT index at the ETP.



Fig. 16
GDGT- and palynology-based proxies for runoff



1234
1235 Figure 16. BrGDGT- and palynology-based proxies for river runoff. a. relative
1236 abundance of Open Ocean, *Phthanoperidinium* spp., *Senegalinium* cpx. and terrestrial
1237 palynomorphs in 0.05 index unit bins of the resampled IR. Dashed line separates
1238 elevated IR bins from non-elevated IR bins. b. Time series of the same palynology
1239 ecogroups and the IR.
1240
1241
1242
1243



1244 6. Conclusions

1245 We have critically reviewed the GDGT- and dinoflagellate cyst data in samples from
 1246 the Maastrichtian–lower Oligocene sediments at ODP Site 1172. IsoGDGTs represent
 1247 a pelagic signal throughout the Maastrichtian–Eocene and are influenced by
 1248 sedimentary-produced isoGDGTs in the Oligocene. BrGDGTs at our site are likely
 1249 primarily soil- or peat-derived, albeit evolutionary changes in brGDGT production
 1250 may have altered the Paleogene soil brGDGT composition. Exceptions are the mid-
 1251 Paleocene, where river-produced brGDGTs influence the record, and the Oligocene,
 1252 where marine in situ produced brGDGTs dominate. TEX₈₆ and MBT'_{5me} records reveal
 1253 the SST and MAAT evolution of the region, respectively, consistent with independent
 1254 existing paleotemperature reconstructions from the same samples. A temperature
 1255 offset between the SST and MAAT proxies remains poorly reconciled with the coastal
 1256 proximity of the site but is likely a combined effect of evolutionary differences in
 1257 MBT'_{5me}-MAAT relationships and a disparate integration of climate signals by both
 1258 proxies. The subdued air temperature cooling in the middle Eocene compared to the
 1259 SST cooling could in part reflect progressive northward tectonic drift of the
 1260 hinterland, while SSTs remained continuously influenced by the Antarctic-derived
 1261 Tasman Current. Strong MAAT cooling occurred step-wise around the Eocene–
 1262 Oligocene transition. The absence of SST cooling around the Eocene–Oligocene
 1263 transition suggests some disconnection between oceanographic reorganisations and
 1264 regional MAAT changes. The absence of a MAAT response at the PETM and MECO
 1265 remains unresolved but might be related to shifting sources of brGDGTs during these
 1266 warm phases.

1267 The relatively high IR, and low #rings_{tetra} confirm dinocyst assemblages characteristic
 1268 of an inner mid-shelf depositional setting. We reconstruct strong river-runoff in the
 1269 Paleocene–early Eocene at this site, and normal open-marine conditions thereafter.
 1270 The latter reflects a drying and increased seasonality in precipitation in the
 1271 catchment as the continent tectonically migrates northwards, out of the region of
 1272 strong precipitation during the middle and late Eocene. The correlation between SST
 1273 and changes dinoflagellate cyst assemblages and biodiversity metrics suggest
 1274 temperature-controlled sea level changes influenced the site, probably through steric
 1275 effects. We find corroborating evidence for a temperature signal in brGMGT



1276 assemblages, further demonstrating the potential to develop novel proxies with more
1277 extensive studies on modern affinities.

1278

1279 **7. Supplements**

- 1280 • Table S1: Age tie points
- 1281 • Table S2: grouping of dinocysts in ecogroups

1282

1283 **8. Data availability**

1284 Raw GDGT/GMGT and palynological data presented in this paper, as well as R
1285 markdown code for data analysis and visualization can be found on Github
1286 (<https://github.com/bijlpeter83/RGDGT.git>). Data was deposited at zenodo under
1287 DOI:10.5281/zenodo.4471204.

1288

1289 **9. Author contributions**

1290 AS (PETM) and PKB (part) prepared samples for GDGT analyses. PKB (part), AS
1291 (PETM) and CB (part) integrated UHPLC-MS results. AS (PETM) and PKB (part)
1292 analysed palynological samples. All authors contributed to the interpretation of the
1293 OG results, and PKB, AS, JF and MC interpreted dinocyst results. PKB wrote the R
1294 script for data analyses and visualization, drafted the figures and wrote the paper
1295 with input from all authors.

1296

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1308



1309 11. References

- 1310 Anagnostou, E., John, E. H., Edgar, K. M., Foster, G. L., Ridgwell, A., Inglis, G. N., et al.:
 1311 Changing atmospheric CO₂ concentration was the primary driver of early
 1312 Cenozoic climate, *Nature*, 533, 380-384, doi:10.1038/nature17423, 2016.
- 1313 Barke, J., Abels, H. A., Sangiorgi, F., Greenwood, D. R., Sweet, A. R., Donders, T., et al.:
 1314 Orbitally forced Azolla blooms and middle Eocene Arctic hydrology: Clues from
 1315 palynology, *Geology*, 39, 427-430, 2011.
- 1316 Baxter, A. J., Hopmans, E. C., Russell, J. M., and Sinninghe Damsté, J. S.: Bacterial GMGTs
 1317 in east african lake sediments: Their potential as palaeotemperature indicators,
 1318 *Geochimica et Cosmochimica acta* 259: 155-169,
 1319 doi:<https://doi.org/10.1016/j.gca.2019.05.039>, 2019.
- 1320 Beerling, D. J., and Royer, D. L.: Convergent Cenozoic CO₂ history, *Nature Geoscience*,
 1321 4, 418-420, 2019.
- 1322 Bijl, P. K., Bendle, A. P. J., Bohaty, S. M., Pross, J., Schouten, S., Tauxe, L., et al.: Eocene
 1323 cooling linked to early flow across the Tasmanian Gateway, *Proceedings of the*
 1324 *National Academy of Sciences of the United States of America* 110(24), 9645-
 1325 9650, 2013a.
- 1326 Bijl, P. K., Brinkhuis, H., Egger, L. M., Eldrett, J. S., Frieling, J., Grothe, A., et al.: Comment
 1327 on 'Wetzelilla and its allies-the 'hole' story: A taxonomic revision of the
 1328 Paleogene dinoflagellate subfamily Wetzelilloideae' by Williams et al. (2015).
 1329 *Palynology*, 41(3), 423-429. doi:10.1080/01916122.2016.1235056, 2017.



- 1330 Bijl, P. K., Houben, A. J. P., Schouten, S., Bohaty, S. M., Sluijs, A., Reichart, G.J., et al.:
 1331 Transient middle Eocene atmospheric carbon dioxide and temperature
 1332 variations, *Science* 330, 819-821, 2010.
- 1333 Bijl, P. K., Pross, J., Warnaar, J., Stickley, C. E., Huber, M., Guerin, R., et al.:
 1334 Environmental forcings of Paleogene Southern Ocean dinoflagellate
 1335 biogeography, *Paleoceanography* 26, PA1202, 2011.
- 1336 Bijl, P. K., Schouten, S., Brinkhuis, H., Sluijs, A., Reichart, G.J., and Zachos, J. C.: Early
 1337 Palaeogene temperature evolution of the southwest Pacific Ocean, *Nature* 461,
 1338 776-779, 2009.
- 1339 Bijl, P. K., Sluijs, A., and Brinkhuis, H.: A magneto- chemo- stratigraphically calibrated
 1340 dinoflagellate cyst zonation of the early Paleogene South Pacific Ocean, *Earth-*
 1341 *Science Reviews* 124, 1-31, 2013b.
- 1342 Bijl, P. K., Sluijs, A., and Brinkhuis, H.: Erratum to "A magneto- and
 1343 chemostratigraphically calibrated dinoflagellate cyst zonation of the early
 1344 Paleogene South Pacific Ocean" [*Earth sci. rev.* 124 (2013) 1-31], *Earth-Science*
 1345 *Reviews* 134, 160-163. doi:10.1016/j.earscirev.2014.03.010, 2014.
- 1346 Blaga, C. I., Reichart, G.-J., Heiri, O., and Sinninghe Damsté, J. S.: Tetraether membrane
 1347 lipid distributions in water-column particulate matter and sediments: A study of
 1348 47 European lakes along a north-south transect, *Journal of Paleolimnology* 41(3),
 1349 523-540, doi:10.1007/s10933-008-9242-2, 2009



- 1350 Bohaty, S. M., Zachos, J. C., and Delaney, M. L.: Foraminiferal Mg/Ca evidence for
1351 Southern Ocean cooling across the Eocene–Oligocene transition, EPSL,
1352 doi:<https://doi.org/10.1016/j.epsl.2011.11.037>, 2012
- 1353 Bohaty, S. M., Zachos, J. C., Florindo, F., and Delaney, M. L.: Coupled greenhouse
1354 warming and deep-sea acidification in the middle Eocene, *Paleoceanography* 24,
1355 doi:10.1029/2008PA001676, 2009
- 1356 Brinkhuis, H.: Late Eocene to early Oligocene dinoflagellate cysts from the Priabonian
1357 type-area (northeast Italy); biostratigraphy and palaeoenvironmental
1358 interpretation, *Palaeogeography, Palaeoclimatology, Palaeoecology* 107, 121-
1359 163, 1994.
- 1360 Brinkhuis, H., Sengers, S., Sluijs, A., Warnaar, J., and Williams, G. L.: Latest Cretaceous
1361 to earliest Oligocene, and Quaternary dinoflagellates from ODP site 1172, East
1362 Tasman Plateau, In N. Exon, and J. P. Kennett (Eds.), *Proceedings of the Ocean*
1363 *Drilling Program, scientific results, volume 189*. College Station, Texas: U.S.
1364 Government Printing Office, 2003.
- 1365 Cande, S. C., and Stock, J. M.: Cenozoic reconstructions of the Australia-new Zealand-
1366 south Pacific sector of antarctica, In N. F. Exon, J. P. Kennett and M. Malone (Eds.),
1367 *The Cenozoic Southern Ocean: Tectonics, sedimentation and climate change*
1368 *between australia and Antarctica* (pp. 5-18) *Geophysical Monograph Series*,
1369 American Geophysical Union, 2004.



- 1370 Carpenter, R. J., Jordan, G. J., Macphail, M. K., and Hill, R. S.: Near-tropical early Eocene
 1371 terrestrial temperatures at the Australo-Antarctic margin, western Tasmania,
 1372 Geology, 40(3), 267-270, doi:10.1130/G32584.1, 2012
- 1373 Contreras, L., Pross, J., Bijl, P. K., Koutsodendris, A., Raine, J. I., van de Schootbrugge, B.,
 1374 et al.: Early to middle Eocene vegetation dynamics at the Wilkes Land Margin
 1375 (Antarctica), Review of Palaeobotany and Palynology 197, 119-142, 2013.
- 1376 Contreras, L., Pross, J., Bijl, P. K., O'Hara, R. B., Raine, J. I., Sluijs, A., et al.: Southern
 1377 high-latitude terrestrial climate change during the Palaeocene-Eocene derived
 1378 from a marine pollen record (ODP site 1172, East Tasman Plateau), Climate of the
 1379 Past, 10(4), 1401-1420, doi:10.5194/cp-10-1401-2014, 2014.
- 1380 Cramwinckel, M. J., Huber, M., Kocken, I. J., Agnini, C., Bijl, P. K., Bohaty, S. M., et al.:
 1381 Synchronous tropical and deep ocean temperature evolution in the Eocene,
 1382 Nature 559, 382-386, 2018.
- 1383 Cramwinckel, M. J., Woelders, L., Huurdeman, E. P., Peterse, F., Gallagher, S. J., Pross, J.,
 1384 et al.: Surface-circulation change in the Southern Ocean across the Middle Eocene
 1385 Climatic Optimum: Inferences from dinoflagellate cysts and biomarker
 1386 paleothermometry, Climate of the Past, , 1-34, 2020
- 1387 Creech, J. B., Baker, J. A., Hollis, C. J., Morgans, H. E. G., and Smith, E. G. C.: Eocene sea
 1388 temperatures for the mid-latitude southwest Pacific from Mg/Ca ratios in
 1389 planktonic and benthic foraminifera, Earth and Planetary Science Letters, 299,
 1390 483-495, 2010.



- 1391 Crouch, E. M., Shepherd, C. L., Morgans, H. E. G., Naafs, B. D. A., Dallanave, E., Phillips,
 1392 A., et al.: Climatic and environmental changes across the Early Eocene Climatic
 1393 Optimum at mid-Waipara River, Canterbury Basin, New Zealand, Earth-Science
 1394 Reviews, 200, doi:10.1016/j.earscirev.2019.102961, 2020.
- 1395 Dallanave, E., Bachtadse, V., Crouch, E. M., Tauxe, L., Shepherd, C. L., Morgans, H. E. G.,
 1396 et al.: Constraining early to middle Eocene climate evolution of the southwest
 1397 Pacific and Southern Ocean. Earth and Planetary Science Letters, 433, 380-392.
 1398 doi:10.1016/j.epsl.2015.11.010, 2016
- 1399 De Jonge, C., Hopmans, E. C., Stadnitskaia, A., Rijpstra, W. I. C., Hofland, R., Tegelaar, E.,
 1400 et al.: Identification of novel penta- and hexamethylated branched glycerol dialkyl
 1401 glycerol tetraethers in peat using HPLC-MS2, GC-MS and GC-SMB-MS, Organic
 1402 Geochemistry 54, 78-82, doi:10.1016/j.orggeochem.2012.10.004, 2013.
- 1403 De Jonge, C., Hopmans, E. C., Zell, C. I., Kim, J. -, Schouten, S., and Sinninghe Damsté, J.
 1404 S.: Occurrence and abundance of 6-methyl branched glycerol dialkyl glycerol
 1405 tetraethers in soils: Implications for palaeoclimate reconstruction, Geochimica et
 1406 Cosmochimica Acta, 141, 97-112. doi:10.1016/j.gca.2014.06.013, 2014a.
- 1407 De Jonge, C., Stadnitskaia, A., Cherkashov, G., and Sinninghe Damsté, J. S.: Branched
 1408 glycerol dialkyl glycerol tetraethers and crenarchaeol record post-glacial sea
 1409 level rise and shift in source of terrigenous brGDGTs in the Kara Sea (Arctic
 1410 Ocean), Organic Geochemistry 92, 42-54,
 1411 doi:10.1016/j.orggeochem.2015.11.009, 2016.



- 1412 De Jonge, C., Stadnitskaia, A., Hopmans, E. C., Cherkashov, G., Fedotov, A., and
 1413 Sinninghe Damsté, J. S.: In situ produced branched glycerol dialkyl glycerol
 1414 tetraethers in suspended particulate matter from the Yenisei River, eastern
 1415 Siberia, *Geochimica et Cosmochimica Acta*, 125, 476-491,
 1416 doi:10.1016/j.gca.2013.10.031, 2014b.
- 1417 De Jonge, C., Stadnitskaia, A., Hopmans, E. C., Cherkashov, G., Fedotov, A., Streletskaya,
 1418 I. D., et al.: Drastic changes in the distribution of branched tetraether lipids in
 1419 suspended matter and sediments from the Yenisei River and Kara Sea (Siberia):
 1420 Implications for the use of brGDGT-based proxies in coastal marine sediments,
 1421 *Geochimica et Cosmochimica Acta* 165, 200-225, doi:10.1016/j.gca.2015.05.044,
 1422 2015.
- 1423 De Jonge, C., Radujković, D., Sigurdsson, B. D., Weedon, J. T., Janssens, I., and Peterse,
 1424 F.: Lipid biomarker temperature proxy responds to abrupt shift in the bacterial
 1425 community composition in geothermally heated soils, *Org Geochem*, 137, 894
 1426 103897, <https://doi.org/10.1016/j.orggeochem.2019.07.006>, 2019.
- 1427 Dearing Crampton-Flood, E., Peterse, F., and Sinninghe Damsté, J. S.: Production of
 1428 branched tetraethers in the marine realm: Svalbard fjord sediments revisited,
 1429 *Organic Geochemistry* 138 doi:10.1016/j.orggeochem.2019.103907, 2019.
- 1430 Dearing Crampton-Flood, E., Tierney, J. E., Peterse, F., Kirkels, F M S A, and Sinninghe
 1431 Damsté, J. S. BayMBT: A bayesian calibration model for branched glycerol dialkyl
 1432 glycerol tetraethers in soils and peats, *Geochimica Et Cosmochimica Acta*, 268,
 1433 142-159, doi:10.1016/j.gca.2019.09.043, 2020.



- 1434 Douglas, P. M. J., Affek, H. P., Ivany, L. C., Houben, A. J. P., Sijp, W. P., Sluijs, A.,
 1435 Schouten, S., Pagani, M.: Pronounced zonal heterogeneity in Eocene southern
 1436 high latitude sea surface temperatures, PNAS 111 (18), 6582-6587, 2014.
- 1437 Elling, F. J., Kanneke, M., Lipp, J. S., Becker, K. W., Gagen, E. J., and Hinrichs, K.-U.: Effects
 1438 of growth phase on the membrane lipid composition of the thaumarchaeon
 1439 Nitrosopumilus maritimus and their implications for archaeal lipid distributions
 1440 in the marine environment, Geochim Cosmochim Acta, 141, 579-597,
 1441 <https://doi.org/10.1016/j.gca.2014.07.005>, 2014.
- 1442 Evans, D., Sagoo, N., Renema, W., Cotton, L. J., Müller, W., Todd, J. A., et al.: Eocene
 1443 greenhouse climate revealed by coupled clumped isotope-Mg/Ca thermometry,
 1444 Proceedings of the National Academy of Sciences of the United States of America,
 1445 115(6), 1174-1179. doi:10.1073/pnas.1714744115, 2018.
- 1446 Exon, N. F., Kennett, J. P., and Malone, M.: Proceedings of the Ocean Drilling Program,
 1447 initial reports, volume 189, College Station, Texas: U.S. Government Printing
 1448 Office., 2001.
- 1449 Feng, X., Vonk, J.E., van Dongen, B.E., Gustafsson, Ö., Semiletov, I.P., Dudarev, O.V.,
 1450 Wang, Z., Montluçon, D.B., Wacker, L., Eglinton, T.I.: Differential mobilization of
 1451 terrestrial carbon pools in Eurasian Arctic river basins, PNAS 110 (35) 14168-
 1452 14173; <https://doi.org/10.1073/pnas.1307031110>, 2013.
- 1453 Fensome, R. A., Taylor, F. J. R., Norris, G., Sarjeant, W. A. S., Wharton, D. I., and
 1454 Williams, G. L.: In Dinkins G. (Ed.), A classification of modern and fossil
 1455 dinoflagellates, Salem: Micropalaeontology, Special Paper, 1993.



- 1456 Fisher, R. A., Corbet, A. S., and Williams, C. B.: The relation between the number of
 1457 species and the number of individuals in a random sample of an animal
 1458 population, *Journal of Animal Ecology*, 12(1), 42-58, doi:10.2307/1411, 1943.
- 1459 Foster, G. L., Royer, D. L., and Lunt, D. J.: Future climate forcing potentially without
 1460 precedent in the last 420 million years, *Nature Communications*, 8, 14845, 2017.
- 1461 Frieling, J., Gebhardt, H., Huber, M., Adekeye, O. A., Akande, S. O., Reichart, G.J., et al.:
 1462 Extreme warmth and heat-stressed plankton in the tropics during the Paleocene-
 1463 Eocene Thermal Maximum. *Science Advances*, 3(3) doi:10.1126/sciadv.1600891,
 1464 2017.
- 1465 Frieling, J., Huurdeman, E. P., Rem, C. C. M., Donders, T. H., Pross, J., Bohaty, S. M., et al.:
 1466 Identification of the Paleocene-Eocene boundary in coastal strata in the Otway
 1467 Basin, Victoria, Australia, *Journal of Micropalaeontology*, 37(1), 317-339,
 1468 doi:10.5194/jm-37-317-2018, 2018.
- 1469 Frieling, J., Iakovleva, A. I., Reichart, G. J., Aleksandrova, G. N., Gnibidenko, Z. N.,
 1470 Schouten, S., et al.: Paleocene–Eocene warming and biotic response in the
 1471 epicontinental west Siberian Sea, *Geology*, doi:10.1130/G35724.1, 2014.
- 1472 Frieling, J., and Sluijs, A.: Towards quantitative environmental reconstructions from
 1473 ancient non-analogue microfossil assemblages: Ecological preferences of
 1474 Paleocene – Eocene dinoflagellates, *Earth-Science Reviews*, 185, 956-973,
 1475 doi:10.1016/j.earscirev.2018.08.014, 2018.



- 1476 Fuller, M., and Touchard, Y. On the magnetostratigraphy of the East Tasman Plateau,
 1477 timing of the opening of the Tasmanian Gateway and paleoenvironmental
 1478 changes, In N. Exon, J. P. Kennett and M. Malone (Eds.), The Cenozoic Southern
 1479 Ocean. tectonics, sedimentation and climate change between Australia and
 1480 Antarctica (pp. 127-151), Washington: American Geophysical Union (AGU)
 1481 Geophysical Monograph series, 2004.
- 1482 Gibbs, S. J., Bown, P. R., Murphy, B. H., Sluijs, A., Edgar, K. M., Pälike, H., et al.: Scaled
 1483 biotic disruption during early Eocene global warming events, Biogeosciences,
 1484 9(11), 4679-4688, 2012.
- 1485 Guasti, E., Speijer, R. P., Brinkhuis, H., Smit, J., and Steurbaut, E.: Paleoenvironmental
 1486 change at the Danian-Selandian transition in Tunisia: Foraminifera, organic-
 1487 walled dinoflagellate cyst and calcareous nannofossil records, Marine
 1488 Micropaleontology, 59, 210-229, 2006.
- 1489 Hartman, J. D., Bijl, P. K., Sangiorgi, F., Peterse, F., Schouten, S., Salabarnada, A., et al.:
 1490 Paleooceanography and ice sheet variability offshore Wilkes Land, Antarctica –
 1491 part 3: Insights from Oligocene–Miocene TEX₈₆-based sea surface temperature
 1492 reconstructions, Climate of the Past, 14, 1275–1297, 2018.
- 1493 Hill, P. J., and Exon, N. F.: Tectonics and basin development of the offshore Tasmanian
 1494 area; incorporating results from deep ocean drilling, In N. F. Exon, J. P. Kennett
 1495 and M. Malone (Eds.), The Cenozoic Southern Ocean; tectonics, sedimentation
 1496 and climate change between Australia and Antarctica (Geophysical Monograph



- 1497 Series 151 ed., pp. 19-19), Washington, D.C., U.S.A.: American Geophysical Union,
 1498 2004.
- 1499 Hill, P. J., and Moore, A. M. G.: Geological framework of the South Tasman Rise and
 1500 East Tasman Plateau, Geoscience Australia, 2001/40, 2001.
- 1501 Hines, B. R., Hollis, C. J., Atkins, C. B., Baker, J. A., Morgans, H. E. G., and Strong, P. C.:
 1502 Reduction of oceanic temperature gradients in the early Eocene southwest
 1503 Pacific Ocean, Palaeogeography, Palaeoclimatology, Palaeoecology, 475, 41-54,
 1504 doi:10.1016/j.palaeo.2017.02.037, 2017.
- 1505 Holdgate, G. R., McGowran, B., Fromhold, T., Wagstaff, B. E., Gallagher, S. J., Wallace, M.
 1506 W., et al.: Eocene-Miocene carbon-isotope and floral record from brown coal
 1507 seams in the Gippsland Basin of southeast Australia, Global and Planetary
 1508 Change, 65(1-2), 89-103, doi:10.1016/j.gloplacha.2008.11.001, 2009
- 1509 Hollis, C. J., Crouch, E. M., Morgans, H. E. G., Handley, L., Baker, J. A., Creech, J., et al.:
 1510 Tropical sea temperatures in the high latitude South Pacific during the Eocene,
 1511 Geology, 37(2), 99-102, 2009.
- 1512 Hollis, C. J., Dunkley Jones, T., Anagnostou, E., Bijl, P. K., Cramwinckel, M. J., Cui, Y., et
 1513 al.: The DeepMIP contribution to PMIP4: Methodologies for selection,
 1514 compilation and analysis of latest paleocene and early Eocene climate proxy data,
 1515 incorporating version 0.1 of the DeepMIP database, Geoscientific Model
 1516 Development, 12(7), 3149-3206, doi:10.5194/gmd-12-3149-2019, 2019



- 1517 Hollis, C. J., Tayler, M. J. S., Andrew, B., Taylor, K. W., Lurcock, P., Bijl, P. K., et al.:
 1518 Organic-rich sedimentation in the south Pacific Ocean associated with late
 1519 Paleocene climatic cooling, *Earth-Science Reviews*, 134, 81-97, 2014.
- 1520 Hollis, C. J., Taylor, K. W. R., Handley, L., Pancost, R. D., Huber, M., Creech, J. B., et al.:
 1521 Early Paleogene temperature history of the southwest Pacific Ocean: Reconciling
 1522 proxies and models, *Earth and Planetary Science Letters*, 349–350(0), 53-66,
 1523 doi:10.1016/j.epsl.2012.06.024, 2012
- 1524 Hopmans, E. C., Schouten, S., and Sinninghe Damsté, J. S.: The effect of improved
 1525 chromatography on GDGT-based palaeoproxies, *Organic Geochemistry*, 93, 1-6,
 1526 doi:10.1016/j.orggeochem.2015.12.006, 2016.
- 1527 Hopmans, E. C., Weijers, J. W. H., Schefuß, E., Herfort, L., Sinninghe Damsté, J. S., and
 1528 Schouten, S.: A novel proxy for terrestrial organic matter in sediments based on
 1529 branched and isoprenoid tetraether lipids, *Earth and Planetary Science Letters*,
 1530 224, 107-116, 2004.
- 1531 Houben, A. J. P., Bijl, P. K., Sluijs, A., Schouten, S., and Brinkhuis, H.: Late Eocene
 1532 Southern Ocean cooling and invigoration of circulation preconditioned Antarctica
 1533 for full-scale glaciation, *Geochemistry, Geophysics, Geosystems*, 20,
 1534 <https://doi.org/10.1029/2019GC008182>, 2019.
- 1535 Huber, M., Brinkhuis, H., Stickley, C. E., Döös, K., Sluijs, A., Warnaar, J., et al.: Eocene
 1536 circulation of the Southern Ocean: Was Antarctica kept warm by subtropical
 1537 waters? *Paleoceanography*, 19, 4026, 2004.



- 1538 Huber, M., and Caballero, R.: The early Eocene equable climate problem revisited,
 1539 Climate of the Past, 7, 603-633, 2011.
- 1540 Huber, M., and Thomas, E., Paleooceanography: The greenhouse world, in Encyclopedia
 1541 of Ocean Sciences, pp 319–329, 2010.
- 1542 Hurley, S. J., Elling, F. J., Kanneke, M., Buchwald, C., Wankel, S. D., Santoro, A. E., Lipp, J.
 1543 S., Hinrichs, K.-U., and Pearson, A.: Influence of ammonia oxidation rate on
 1544 thaumarchaeal lipid composition and the TEX86 temperature proxy, Proceedings
 1545 of the National Academy of Sciences, 113, 7762-7767,
 1546 10.1073/pnas.1518534113, 2016.
- 1547 Huurdeman, E.P., Frieling, J., Reichgelt, T., Bijl, P.K., Bohaty, S.M., Holdgate, G.R.,
 1548 Gallagher, S.J., Peterse, F., Greenwood, D.R., Pross, J.: Rapid expansion of meso-
 1549 megathermal rain forests into the southern high latitudes at the onset of the
 1550 Paleocene-Eocene Thermal Maximum. Geology doi:
 1551 <https://doi.org/10.1130/G47343.1>, 2020.
- 1552 Inglis, G. N., Farnsworth, A., Lunt, D., Foster, G. L., Hollis, C. J., Pagani, M., et al.: Descent
 1553 toward the icehouse: Eocene sea surface cooling inferred from GDGT
 1554 distributions, Paleooceanography, 30(7), 1000-1020,
 1555 doi:10.1002/2014PA002723, 2015.
- 1556 Inglis, G. N., Bragg, F., Burls, N. J., Cramwinckel, M. J., Evans, D., Foster, G. L., Huber, M.,
 1557 Lunt, D. J., Siler, N., Steinig, S., Tierney, J. E., Wilkinson, R., Anagnostou, E., de Boer,
 1558 A. M., Dunkley Jones, T., Edgar, K. M., Hollis, C. J., Hutchinson, D. K., and Pancost, R.
 1559 D.: Global mean surface temperature and climate sensitivity of the early Eocene



- 1560 Climatic Optimum (EECO), Paleocene–Eocene Thermal Maximum (PETM), and
 1561 latest Paleocene, *Clim. Past*, 16, 1953–1968, [https://doi.org/10.5194/cp-16-](https://doi.org/10.5194/cp-16-1953-2020)
 1562 1953-2020, 2020.
- 1563 Kim, J. -, Meer, J. v. d., Schouten, S., Helmke, P., Willmott, V., Sangiorgi, F., et al.: New
 1564 indices and calibrations derived from the distribution of crenarchaeal isoprenoid
 1565 tetraether lipids: Implications for past sea surface temperature reconstructions,
 1566 *Geochimica et Cosmochimica Acta*, 74, 4639-4654, 2010.
- 1567 Kirkels, F. M. S. A., Ponton, C., Galy, V., West, A. J., Feakins, S. J., and Peterse, F.: From
 1568 Andes to Amazon: Assessing Branched Tetraether Lipids as Tracers for Soil
 1569 Organic Carbon in the Madre de Dios River System, *Journal of Geophysical*
 1570 *Research: Biogeosciences*, 125, e2019JG005270, [10.1029/2019jg005270](https://doi.org/10.1029/2019jg005270), 2020.
- 1571 Korasidis, V. A., Wallace, M. W., Dickinson, J. A., and Hoffman, N.: Depositional setting
 1572 for Eocene seat earths and related facies of the Gippsland Basin, Australia,
 1573 *Sedimentary Geology*, doi:10.1016/j.sedgeo.2019.07.007, 2019
- 1574 Kozdon, R., Penman, D. E., Kelly, D. C., Zachos, J. C., Fournelle, J. H., & Valley, J. W.:
 1575 Enhanced poleward flux of atmospheric moisture to the Weddell Sea region (ODP
 1576 Site 690) during the Paleocene-Eocene Thermal Maximum, *Paleoceanography*
 1577 and *Paleoclimatology*, 35, e2019PA003811,
 1578 <https://doi.org/10.1029/2019PA003811> , 2020
- 1579 Lattaud, J., Dorhout, D., Schulz, H., Castañeda, I. S., Schefuß, E., Damsté, J. S. S., et al.:
 1580 The C32 alkane-1,15-diol as a proxy of late Quaternary riverine input in coastal



- 1581 margins, *Climate of the Past*, 13(8), 1049-1061, doi:10.5194/cp-13-1049-2017,
 1582 2017.
- 1583 Lauretano, V., Zachos, J. C., and Lourens, L. J.: Orbitally paced carbon and deep-sea
 1584 temperature changes at the peak of the Early Eocene Climatic Optimum.
 1585 *Paleoceanography and Paleoclimatology*, 33(10), 1050-1065,
 1586 doi:10.1029/2018PA003422, 2018.
- 1587 Leutert, T. J., Auderset, A., Martínez-García, A., Modestou, S., and Meckler, A. N.:
 1588 Southern Ocean temperature evolution coupled to middle Miocene ice sheet
 1589 expansion, *Nature Geoscience* 13, 634-639, 2020.
- 1590 Liu, X., Lipp, J. S., Simpson, J. H., Lin, Y., Summons, R. E., and Hinrichs, K.: Mono- and
 1591 dihydroxyl glycerol dibiphytanyl glycerol tetraethers in marine sediments:
 1592 Identification of both core and intact polar lipid forms. *Geochimica et*
 1593 *Cosmochimica Acta*, 89, 102-115. doi:<https://doi.org/10.1016/j.gca.2012.04.053>,
 1594 2012.
- 1595 Lunt, D. J., Jones, T. D., Heinemann, M., Huber, M., LeGrande, A., Winguth, A., et al.: A
 1596 model-data comparison for a multi-model ensemble of early Eocene atmosphere-
 1597 ocean simulations: EoMIP, *Climate of the Past*, 8(5), 1717-1736, 2012.
- 1598 Lunt, D. J., Bragg, F., Chan, W.-L., Hutchinson, D. K., Ladant, J.-B., Morozova, P.,
 1599 Niezgodzki, I., Steinig, S., Zhang, Z., Zhu, J., Abe-Ouchi, A., Anagnostou, E., de Boer,
 1600 A. M., Coxall, H. K., Donnadieu, Y., Foster, G., Inglis, G. N., Knorr, G., Langebroek, P.
 1601 M., Lear, C. H., Lohmann, G., Poulsen, C. J., Sepulchre, P., Tierney, J. E., Valdes, P. J.,
 1602 Volodin, E. M., Dunkley Jones, T., Hollis, C. J., Huber, M., and Otto-Bliesner, B. L.:



- 1603 DeepMIP: model intercomparison of early Eocene climatic optimum (EECO)
 1604 large-scale climate features and comparison with proxy data, *Clim. Past*, 17, 203–
 1605 227, <https://doi.org/10.5194/cp-17-203-2021>, 2021.
- 1606 Macphail, M. K.: Australian palaeoclimates, cretaceous to tertiary, volume 1: Review
 1607 of palaeobotanical and related evidence up to 2000, Canberra: Geology
 1608 Department, Australian National University, 2000.
- 1609 Macphail, M. K.: ODP Leg 189 initial results: Terrestrial plant microfossils. Canberra:
 1610 Geoscience Australia, 2002
- 1611 Mertens, K. N., Dale, B., Ellegaard, M., Jansson, I. -, Godhe, A., Kremp, A., et al.: Process
 1612 length variation in cysts of the dinoflagellate *protoceratium reticulatum*, from
 1613 surface sediments of the baltic-kattegat-skagerrak estuarine system: A regional
 1614 salinity proxy, *Boreas*, 40(2), 242-255, 2011
- 1615 Moore, D. H., Betts, P. G., and Hall, M.: Towards understanding the early Gondwanan
 1616 margin in southeastern Australia, *Gondwana Research*, 23(4), 1581-1598,
 1617 doi:10.1016/j.jgr.2012.08.006, 2013.
- 1618 Müller, R. D., Gaina, C., and Clark, S.: Seafloor spreading around Australia, In J. Veevers
 1619 (Ed.), *Billion-year earth history of Australia and neighbours in gondwanaland*
 1620 (2000) – BYEHA (pp. 1-1) School of Geosciences, University of Sydney, 2000
- 1621 Naafs, B. D. A., Gallego-Sala, A. V., Inglis, G. N., and Pancost, R. D.: Refining the global
 1622 branched glycerol dialkyl glycerol tetraether (brGDGT) soil temperature



- 1623 calibration, *Organic Geochemistry*, 106, 48-56,
 1624 doi:10.1016/j.orggeochem.2017.01.009, 2017.
- 1625 Naafs, B. D. A., Inglis, G. N., Blewett, J., McClymont, E. L., Lauretano, V., Xie, S., et al.: The
 1626 potential of biomarker proxies to trace climate, vegetation, and biogeochemical
 1627 processes in peat: A review, *Global and Planetary Change*, 179, 57-79,
 1628 doi:10.1016/j.gloplacha.2019.05.006, 2019.
- 1629 Naafs, B. D. A., Inglis, G. N., Zheng, Y., Amesbury, M. J., Biester, H., Bindler, R., et al.:
 1630 Introducing global peat-specific temperature and pH calibrations based on
 1631 brGDGT bacterial lipids, *Geochimica et Cosmochimica Acta*, 208, 285-301,
 1632 doi:10.1016/j.gca.2017.01.038, 2017.
- 1633 Naafs, B. D. A., McCormick, D., Inglis, G. N., and Pancost, R. D.: Archaeal and bacterial
 1634 H-GDGTs are abundant in peat and their relative abundance is positively
 1635 correlated with temperature, *Geochimica Et Cosmochimica Acta*, 227, 156-170,
 1636 doi:10.1016/j.gca.2018.02.025, 2018.
- 1637 Naafs, B. D. A., Rohrsen, M., Inglis, G. N., Lähteenoja, O., Feakins, S. J., Collinson, M. E.,
 1638 et al.: High temperatures in the terrestrial mid-latitudes during the early
 1639 Palaeogene, *Nature Geoscience*, 11(10), 766-771, doi:10.1038/s41561-018-
 1640 0199-0, 2018.
- 1641 O'Brien, C. L., Robinson, S. A., Pancost, R. D., Sinninghe Damsté, J. S., Schouten, S., Lunt,
 1642 D. J., et al.: Cretaceous sea-surface temperature evolution: Constraints from TEX₈₆
 1643 and planktonic foraminiferal oxygen isotopes, *Earth-Science Reviews*, 172, 224-
 1644 247, doi:10.1016/j.earscirev.2017.07.012, 2017.



- 1645 O'Brien, C. L., Huber, M., Thomas, E., Pagani, M., Super, J. R., Elder, L. E., et al.: The
 1646 enigma of Oligocene climate and global surface temperature evolution,
 1647 Proceedings of the National Academy of Sciences, 202003914,
 1648 doi:10.1073/pnas.2003914117, 2020.
- 1649 O'Connor, L. K., Robinson, S. A., Naafs, B. D. A., Jenkyns, H. C., Henson, S., Clarke, M., et
 1650 al.: Late Cretaceous temperature evolution of the southern high latitudes: A
 1651 TEX₈₆ perspective, Paleceanography and Paleoclimatology, 34(4), 436-454,
 1652 doi:10.1029/2018PA003546, 2019.
- 1653 Oksanen, J., Blanchet, F. G., Friendly, M., Kindt, R., Legendre, P., McGlinn, D., et al.:
 1654 Vegan: Community ecology package, 2015.
- 1655 Pancost, R. D., Taylor, K. W. R., Inglis, G. N., Kennedy, E. M., Handley, L., Hollis, C. J., et
 1656 al.: Early Paleogene evolution of terrestrial climate in the SW Pacific, southern
 1657 New Zealand, Geochemistry, Geophysics, Geosystems, 14(12), 5413-5429, 2013.
- 1658 Passchier, S., Bohaty, S. M., Jiménez-Espejo, F., Pross, J., Röhl, U., Van De Flierdt, T., et
 1659 al.: Early Eocene to middle Miocene cooling and aridification of east Antarctica,
 1660 Geochemistry, Geophysics, Geosystems, 14(5), 1399-1410, 2013
- 1661 Passchier, S., Ciarletta, D. J., Miriagos, T. E., Bijl, P. K., and Bohaty, S. M.: An Antarctic
 1662 stratigraphic record of stepwise ice growth through the Eocene-Oligocene
 1663 transition, Bulletin of the Geological Society of America, 129(3-4), 318-330,
 1664 doi:10.1130/B31482.1, 2017.



- 1665 Peterse, F., Kim, J. -, Schouten, S., Kristensen, D. K., Koç, N., and Sinninghe Damsté, J. S.:
 1666 Constraints on the application of the MBT-CBT paleothermometer at high
 1667 latitude environments (Svalbard, Norway), *Organic Geochemistry*, 40, 692-699,
 1668 2009.
- 1669 Peterse, F., Meer, J. v. d., Schouten, S., Weijers, J. W. H., Fierer, N., Jackson, R. B., et al.:
 1670 Revised calibration of the MBT-CBT paleotemperature proxy based on branched
 1671 tetraether membrane lipids in surface soils. *Geochimica Et Cosmochimica Acta*,
 1672 2012.
- 1673 Pross, J., & Brinkhuis, H., Organic-walled dinoflagellate cysts as paleoenvironmental
 1674 indicators in the paleogene; a synopsis of concepts, *Palaeontologische Zeitschrift*,
 1675 Band 79, 53-59, 2005.
- 1676 Pross, J., Contreras, L., Bijl, P. K., Greenwood, D. R., Bohaty, S. M., Schouten, S., et al.:
 1677 Persistent near-tropical warmth on the Antarctic continent during the early
 1678 Eocene epoch, *Nature*, 488, 73-73, 2012.
- 1679 Qin, W., Carlson, L. T., Armbrust, E. V., Devol, A. H., Moffett, J. W., Stahl, D. A., and
 1680 Ingalls, A. E.: Confounding effects of oxygen and temperature on the TEX₈₆
 1681 signature of marine Thaumarchaeota, *Proceedings of the National Academy of*
 1682 *Sciences*, 112, 10979-10984, 10.1073/pnas.1501568112, 2015.
- 1683 Reichgelt, T., West, C.K., Greenwood, D.R.: The relation between global palm
 1684 distribution and climate, *Scientific Reports* 8(1), DOI: 10.1038/s41598-018-
 1685 23147-2, 2018.



- 1686 Robert, C.: Cenozoic environments in the Tasmanian area of the Southern Ocean (ODP
 1687 Leg 189): Inferences from bulk and clay mineralogy, Geophysical Monograph
 1688 Series, 151, 127-151, 2004.
- 1689 Rochon, A., Lewis, J., Ellegaard, M., and Harding, I. C.: The gonyaulax spinifera
 1690 (dinophyceae) “complex”: Perpetuating the paradox? Review of Palaeobotany
 1691 and Palynology, 155(1), 52-60,
 1692 doi:<https://doi.org/10.1016/j.revpalbo.2008.12.017>, 2008.
- 1693 Röhl, U., Brinkhuis, H., Stickley, C. E., Fuller, M., Schellenberg, S. A., Wefer, G., et al.: Sea
 1694 level and astronomically induced environmental changes in middle and late
 1695 Eocene sediments from the East Tasman Plateau. Geophysical Monograph Series,
 1696 151, 127-151, 2004a.
- 1697 Röhl, U., Brinkhuis, H., Sluijs, A., and Fuller, M.: On the search for the
 1698 Paleocene/eocene boundary in the Southern Ocean: Exploring ODP Leg 189
 1699 Holes 1171D and 1172D, Tasman Sea, Geophysical Monograph Series, 151, 113-
 1700 124, 2004b.
- 1701 Röhl, U., Westerhold, T., Bralower, T. J., and Zachos, J. C.: On the duration of the
 1702 Paleocene-Eocene Thermal Maximum (PETM), Geochemistry, Geophysics,
 1703 Geosystems, 8, Q12002, 2007.
- 1704 Salamy, K.A., Zachos, J.C.: Latest Eocene-Early Oligocene climate change and Southern
 1705 Ocean fertility: Inferences from sediment accumulation and stable isotope data,
 1706 Palaeogeogr., Palaeoclimatol., Palaeoecol. 145, 61–77, 1999.



- 1707 Schefuß, E., Eglinton, T., Spencer-Jones, C. et al.: Hydrologic control of carbon cycling
 1708 and aged carbon discharge in the Congo River basin, *Nature Geosciences* 9, 687–
 1709 690, 2016.
- 1710 Schellenberg, S. A., Brinkhuis, H., Stickley, C. E., Fuller, M., Kyte, F. T., and Williams, G.
 1711 L.: The Cretaceous/Paleogene transition on the East Tasman Plateau,
 1712 southwestern Pacific, In N. Exon, J. P. Kennett and M. Malone (Eds.), *The Cenozoic*
 1713 *Southern Ocean; tectonics, sedimentation and climate change between Australia*
 1714 *and Antarctica* (pp. 93-112), Washington: Geophysical Monograph Series, 2004.
- 1715 Schouten, S., Hopmans, E. C., Schefuß, E., and Sinninghe Damsté, J. S.: Distributional
 1716 variations in marine crenarchaeotal membrane lipids: A new tool for
 1717 reconstructing ancient sea water temperatures? *Earth and Planetary Science*
 1718 *Letters*, 204, 265-274, 2002.
- 1719 Schouten, S., Hopmans, E. C., and Sinninghe Damsté, J. S., The organic geochemistry of
 1720 glycerol dialkyl glycerol tetraether lipids: A review, *Organic Geochemistry*, 54,
 1721 19-61, doi:10.1016/j.orggeochem.2012.09.006, 2013.
- 1722 Seton, M., Müller, R. D., Zahirovic, S., Gaina, C., Torsvik, T., Shephard, G., et al.: Global
 1723 continental and ocean basin reconstructions since 200 Ma, *Earth-Science*
 1724 *Reviews*, 113(3-4), 212-270, 2012.
- 1725 Shannon, C. E.: A mathematical theory of communication, *Bell System Technical*
 1726 *Journal*, 27(3), 379-423, doi:10.1002/j.1538-7305.1948.tb01338.x, 1948.



- 1727 Sijp, W. P., Von Der Heydt, A S, and Bijl, P. K.: Model simulations of early westward
 1728 flow across the Tasman Gateway during the early Eocene, *Climate of the Past*,
 1729 12(4), 807-817, doi:10.5194/cp-12-807-2016, 2016.
- 1730 Sijp, W. P., von der Heydt, A S, Dijkstra, H. A., Flögel, S., Douglas, P. M. J., and Bijl, P. K.:
 1731 The role of ocean gateways on cooling climate on long time scales, *Global and*
 1732 *Planetary Change*, 119, 1-22, doi:10.1016/j.gloplacha.2014.04.004, 2014.
- 1733 Simpson, E.: Measurement of diversity. *Nature*, 163, 688, 1949.
- 1734 Sinninghe Damsté, J. S.: Spatial heterogeneity of sources of branched tetraethers in
 1735 shelf systems: The geochemistry of tetraethers in the Berau River delta
 1736 (Kalimantan, Indonesia), *Geochimica et Cosmochimica Acta*, 186, 13-31,
 1737 doi:10.1016/j.gca.2016.04.033, 2016.
- 1738 Sluijs, A., Schouten, S., Pagani, M., Woltering, M., Brinkhuis, H., Sinninghe Damsté, J. S.,
 1739 et al.: Subtropical Arctic Ocean temperatures during the Palaeocene/Eocene
 1740 Thermal Maximum, *Nature*, 441, 610-613, 2006.
- 1741 Sluijs, A., Bijl, P. K., Schouten, S., Röhl, U., Reichert, G.J., and Brinkhuis, H.: Southern
 1742 Ocean warming and hydrological change during the Paleocene-Eocene Thermal
 1743 Maximum, *Climate of the Past*, 7, 47-61, 2011.
- 1744 Sluijs, A., and Brinkhuis, H.: A dynamic climate and ecosystem state during the
 1745 Paleocene-Eocene Thermal Maximum: Inferences from dinoflagellate cyst
 1746 assemblages on the New Jersey Shelf, *Biogeosciences*, 6(8), 1755-1781, 2009.



- 1747 Sluijs, A., Brinkhuis, H., Stickley, C. E., Warnaar, J., Williams, G. L., and Fuller, M.:
 1748 Dinoflagellate cysts from the Eocene - Oligocene transition in the Southern
 1749 Ocean: Results from ODP Leg 189, In N. Exon, and J. P. Kennett (Eds.),
 1750 Proceedings of the Ocean Drilling Program, scientific results, volume 189,
 1751 College Station, Texas: U.S. Government Printing Office, 2003.
- 1752 Sluijs, A., Frieling, J., Inglis, G. N., Nierop, K. G. J., Peterse, F., Sangiorgi, F., et al.: Late
 1753 Paleocene – early Eocene Arctic Ocean sea surface temperatures; reassessing
 1754 biomarker paleothermometry at lomonosov ridge, Clim. Past Discuss.,
 1755 doi:rg/10.5194/cp-2020-13, 2020.
- 1756 Sluijs, A., Pross, J., and Brinkhuis, H.: From greenhouse to icehouse; organic walled
 1757 dinoflagellate cysts as paleoenvironmental indicators in the Paleogene, Earth-
 1758 Science Reviews, 68, 281-315, 2005.
- 1759 Sluijs, A., van Roij, L., Harrington, G. J., Schouten, S., Sessa, J. A., LeVay, L. J., et al.:
 1760 Warming, euxinia and sea level rise during the Paleocene–Eocene Thermal
 1761 Maximum on the Gulf Coastal plain: Implications for ocean oxygenation and
 1762 nutrient cycling, Climate of the Past, 10(4), 1421-1439, doi:10.5194/cp-10-1421-
 1763 2014, 2014.
- 1764 Somme, T. O., Helland-Hansen, W., and Granjeon, D., Impact of eustatic amplitude
 1765 variations on shelf morphology, sediment dispersal, and sequence stratigraphic
 1766 interpretation: Icehouse versus greenhouse systems, Geology, 37(7), 587-590,
 1767 2009.



- 1768 Staub, J.R., Cohen, A.D.: Kaolinite-enrichment Beneath Coals; A Modern Analog,
 1769 Snuggedy Swamp, South Carolina, SEPM J Sediment Res Vol. 48(1):203–210,
 1770 1978.
- 1771 Stickley, C. E., Brinkhuis, H., McGonigal, K. L., Chapronière, G. C. H., Fuller, M., Kelly, D.
 1772 C., et al.: Late Cretaceous - Quaternary biomagnetostratigraphy of ODP Site 1168,
 1773 1170, 1171 and 1172, Tasmanian Gateway, In N. F. Exon, J. P. Kennett and M. J.
 1774 Malone (Eds.), Proceedings of the Ocean Drilling Program, scientific results,
 1775 volume 189, 2004a.
- 1776 Stickley, C. E., Brinkhuis, H., Schellenberg, S. A., Sluijs, A., Röhl, U., Fuller, M., et al.:
 1777 Timing and nature of the deepening of the Tasmanian Gateway,
 1778 Paleocceanography, 19, 4027, 2004b.
- 1779 Tang, X., Naafs, B. D., Pancost, R. D., Liu, Z., Fan, T., & Zheng, Y.: Exploring the
 1780 influences of temperature on “H-shaped” glycerol dialkyl glycerol tetraethers in a
 1781 stratigraphic context: Evidence from two peat cores across the late Quaternary,
 1782 Frontiers in Earth Science, 8, 477, 2021.
- 1783 Taylor, K. W. R., Huber, M., Hollis, C. J., Hernandez-Sanchez, M. T., and Pancost, R. D.:
 1784 Re-evaluating modern and Palaeogene GDGT distributions: Implications for SST
 1785 reconstructions, Global and Planetary Change, 108, 158-174, 2013.
- 1786 Thomas, D. J., Bralower, T. J., and Jones, C. E.: Neodymium isotopic reconstruction of
 1787 the late Paleocene - early Eocene thermohaline circulation, Earth and Planetary
 1788 Science Letters, 209(3-4), 309-322, 2003.



- 1789 Thomas, D. J., Korty, R., Huber, M., Schubert, J. A., and Haines, B.: Nd isotopic structure
 1790 of the Pacific Ocean 70–30 Ma and numerical evidence for vigorous ocean
 1791 circulation and ocean heat transport in a greenhouse world, *Paleoceanography*,
 1792 PA2535, 2014.
- 1793 Tierney, J. E., and Russell, J. M.: Distributions of branched GDGTs in a tropical lake
 1794 system: Implications for lacustrine application of the MBT/CBT paleoproxy,
 1795 doi:10.1016/j.orggeochem.2009.04.014, 2009.
- 1796 Tierney, J. E., Sinninghe Damsté, J. S., Pancost, R. D., Sluijs, A., and Zachos, J. C.: Eocene
 1797 temperature gradients, *Nature Geoscience*, 10(8), 538-539,
 1798 doi:10.1038/ngeo2997, 2017.
- 1799 Tierney, J. E., and Tingley, M. P.: A TEX₈₆ surface sediment database and extended
 1800 bayesian calibration, *Scientific Data*, 2 doi:10.1038/sdata.2015.29, 2015.
- 1801 Tierney, J. E., & Tingley, M. P.: BAYSPLINE: A new calibration for the alkenone
 1802 paleothermometer, *Paleoceanography and Paleoclimatology*, 33(3), 281-301,
 1803 doi:10.1002/2017PA003201, 2018.
- 1804 Torsvik, T. H., Van der Voo, R., Preeden, U., Niocaill, C. M., Steinberger, B., Doubrovine,
 1805 P. V., et al.: Phanerozoic polar wander, palaeogeography and dynamics, *Earth-*
 1806 *Science Reviews*, 114(3-4), 325-368, 2012.
- 1807 Truswell, E. M.: Palynomorph assemblages from marine Eocene sediments on the
 1808 west Tasmanian continental margin and the South Tasman Rise. *Australian*
 1809 *Journal of Earth Sciences*, 44, 633-654, 1997.



- 1810 Van Dijk, J., Fernandez, A., Bernasconi, S.M., Caves Rugenstein, J.K., Passey, S.R., White,
 1811 T.: Spatial pattern of super-greenhouse warmth controlled by elevated specific
 1812 humidity, *Nature Geoscience*, 13 (11), pp. 739-744, 2020.
- 1813 Van Hinsbergen, D. J. J., De Groot, L. V., Van Schaik, S. J., Spakman, W., Bijl, P. K., Sluijs,
 1814 A., et al.: A paleolatitude calculator for paleoclimate studies, *PLoS ONE*, 10(6),
 1815 2015.
- 1816 Warden, L., Kim, J. -, Zell, C., Vis, G. -, De Stigter, H., Bonnin, J., et al.: Examining the
 1817 provenance of branched GDGTs in the tagus river drainage basin and its outflow
 1818 into the Atlantic Ocean over the holocene to determine their usefulness for
 1819 paleoclimate applications, *Biogeosciences*, 13(20), 5719-5738, doi:10.5194/bg-
 1820 13-5719-2016, 2016.
- 1821 Warden, L., Moros, M., Weber, Y., and Sinninghe Damsté, J. S.: Change in provenance of
 1822 branched glycerol dialkyl glycerol tetraethers over the Holocene in the Baltic Sea
 1823 and its impact on continental climate reconstruction, *Organic Geochemistry*, 121,
 1824 138-154, doi:10.1016/j.orggeochem.2018.03.007, 2018.
- 1825 Warnaar, J., Bijl, P. K., Huber, M., Sloan, L. C., Brinkhuis, H., Röhl, U., et al.: Orbitally
 1826 forced climate changes in the Tasman sector during the middle Eocene,
 1827 *Palaeogeography, Palaeoclimatology, Palaeoecology*, 280, 361-370, 2009.
- 1828 Weijers, J. W. H., Lim, K. L. H., Aquilina, A., Damsté, J. S. S., and Pancost, R. D.:
 1829 Biogeochemical controls on glycerol dialkyl glycerol tetraether lipid distributions
 1830 in sediments characterized by diffusive methane flux, *Geochemistry, Geophysics,*
 1831 *Geosystems*, 12(10), doi:10.1029/2011GC003724, 2011.



- 1832 Weijers, J. W. H., Schouten, S., Spaargaren, O. C., and Sinninghe Damste, J. S.:
 1833 Occurrence and distribution of tetraether membrane lipids in soils: Implications
 1834 for the use of the TEX₈₆ proxy and the BIT index, *Organic Geochemistry*, 37,
 1835 1680-1693, 2006.
- 1836 Weijers, J. W. H., Schouten, S., van den Donker, J C, Hopmans, E. C., and Sinninghe
 1837 Damsté, J. S.: Environmental controls on bacterial tetraether membrane lipid
 1838 distribution in soils, *Geochimica Et Cosmochimica Acta*, 71, 703-713, 2007.
- 1839 Westerhold, T., Marwan, N., Drury, A. J., Liebrand, D., Agnini, C., Anagnostou, E., et al.:
 1840 An astronomically dated record of earth's climate and its predictability over the
 1841 last 66 million years. *Science*, 369(6509), 1383-1387,
 1842 doi:10.1126/science.aba6853, 2020.
- 1843 Willard, D. A., Donders, T. H., Reichgelt, T., Greenwood, D. R., Sangiorgi, F., Peterse, F.,
 1844 et al.: Arctic vegetation, temperature, and hydrology during early Eocene
 1845 transient global warming events, *Global and Planetary Change*, 178, 139-152,
 1846 doi:10.1016/j.gloplacha.2019.04.012, 2019.
- 1847 Williams, G. L., Fensome, R. A., and MacRae, R. A.: *Dinoflaj3*, American Association of
 1848 Stratigraphic Palynologists, Data Series, 2, 2017.
- 1849 Williams, S. E., Whittaker, J. M., Halpin, J. A., and Müller, R. D.: Australian-Antarctic
 1850 breakup and seafloor spreading: Balancing geological and geophysical
 1851 constraints, *Earth-Science Reviews*, 188, 41-58,
 1852 doi:10.1016/j.earscirev.2018.10.011, 2019



- 1853 Xie, S., Liu, X., Schubotz, F., Wakeham, S. G., & Hinrichs, K.: Distribution of glycerol
 1854 ether lipids in the oxygen minimum zone of the eastern tropical north pacific
 1855 ocean, *Organic Geochemistry*, 71, 60-71,
 1856 doi:<https://doi.org/10.1016/j.orggeochem.2014.04.006>, 2014.
- 1857 Zachos, J. C., Schouten, S., Bohaty, S., Quattlebaum, T., Sluijs, A., Brinkhuis, H., et al.:
 1858 Extreme warming of mid-latitude coastal ocean during the Paleocene-Eocene
 1859 Thermal Maximum: Inferences from TEX and isotope data, *Geology*, 34(9), 737-
 1860 740, 2006.
- 1861 Zeebe, R.E., Lourens, L.J.: Solar System chaos and the Paleocene–Eocene boundary age
 1862 constrained by geology and astronomy, *Science*, 365 (6456), pp. 926-929, 2019.
- 1863 Zell, C., Kim, J. -, Dorhout, D., Baas, M., and Sinninghe Damsté, J. S.: Sources and
 1864 distributions of branched tetraether lipids and crenarchaeol along the
 1865 Portuguese continental margin: Implications for the BIT index, *Continental Shelf*
 1866 *Research*, 96, 34-44, doi:10.1016/j.csr.2015.01.006, 2015.
- 1867 Zell, C., Kim, J. -, Hollander, D., Lorenzoni, L., Baker, P., Silva, C. G., et al.: Sources and
 1868 distributions of branched and isoprenoid tetraether lipids on the Amazon shelf
 1869 and fan: Implications for the use of GDGT-based proxies in marine sediments,
 1870 *Geochimica et Cosmochimica Acta*, 139, 293-312, doi:10.1016/j.gca.2014.04.038,
 1871 2014.
- 1872 Zell, C., Kim, J. -, Moreira-Turcq, P., Abril, G., Hopmans, E. C., Bonnet, M. -, et al.:
 1873 Disentangling the origins of branched tetraether lipids and crenarchaeol in the



- 1874 lower Amazon River: Implications for GDGT-based proxies, Limnology and
1875 Oceanography, 58(1), 343-353, doi:10.4319/lo.2013.58.1.0343, 2013
- 1876 Zhang, Y. G., Pagani, M., and Wang, Z.: Ring index: A new strategy to evaluate the
1877 integrity of TEX₈₆ paleothermometry, Paleoceanography, 31(2), 220-232,
1878 doi:10.1002/2015PA002848, 2016.
- 1879 Zhang, Y. G., Zhang, C. L., Liu, X. -, Li, L., Hinrichs, K. -, and Noakes, J. E.: Methane
1880 index: A tetraether archaeal lipid biomarker indicator for detecting the instability
1881 of marine gas hydrates, Earth and Planetary Science Letters, 307(3-4), 525-534,
1882 doi:10.1016/j.epsl.2011.05.031, 2011.
- 1883
- 1884