

1 **Enhanced terrestrial carbon export from East Antarctica during the early Eocene**

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45 **Key points:**

- 46 • Terrestrial organic carbon is an important CO₂ sink when transported via rivers
47 to the ocean and sequestered in coastal marine sediments.
- 48 • Biomarkers indicate enhanced terrestrial organic carbon burial during early
49 Eocene - this could have acted as a negative feedback mechanism
- 50 • Hydrology regulates organic carbon export from the terrestrial biosphere at this
51 site

52

53 **Abstract:**

54 Terrestrial organic carbon (TerrOC) acts as an important CO₂ sink when transported
55 via rivers to the ocean and sequestered in coastal marine sediments. This mechanism
56 might help to modulate atmospheric CO₂ levels over short- and long timescales (10³
57 to 10⁶ years), but its importance during past warm climates remains unknown. Here
58 we use terrestrial biomarkers preserved in coastal marine sediment samples from
59 Wilkes Land, East Antarctica (~67°S) to quantify TerrOC burial during the early
60 Eocene (~54.4 to 51.5 Ma). Terrestrial biomarker distributions indicate the delivery of
61 plant-, soil- and peat-derived organic carbon (OC) into the marine realm. Mass
62 accumulation rates of plant- (long-chain *n*-alkane) and soil-derived (hopane)
63 biomarkers dramatically increase between the earliest Eocene (~54 Ma) and the early
64 Eocene Climatic Optimum (EECO; ~53 Ma). This coincides with increased OC mass
65 accumulation rates and indicates enhanced TerrOC burial during the EECO. Leaf wax
66 δ²H values indicate that the EECO was characterised by wetter conditions relative to
67 the earliest Eocene, suggesting that hydroclimate exerts a first-order control on
68 TerrOC export. Our results indicate that TerrOC burial in coastal marine sediments

69 could have acted as an important negative feedback mechanism during the early
70 Eocene, but also during other warm climate intervals.

71

72 **1. Introduction**

73 Over long-timescales (10^4 to 10^6 years), the evolution of atmospheric carbon dioxide
74 (CO_2) concentrations primarily reflects the balance between CO_2 sources (e.g., solid
75 earth degassing, oxidation of rock-derived OC and sulphide minerals; Plank and
76 Manning, 2019; Hilton et al., 2014; Torres et al., 2014) and CO_2 sinks (e.g., silicate
77 weathering, organic carbon burial) (Berner 1990; Gaillardet et al., 1999; Galy et al.,
78 2007). Between ~20 to 50% of total OC burial in modern marine sediments is derived
79 from the terrestrial biosphere (TerrOC; i.e. soils and vegetation) (Hilton & West 2020).
80 Consequently, TerrOC burial in marine sediments may act as an important CO_2 sink
81 over short- to long timescales (10^3 to 10^6 years) (Hilton et al 2015).

82 The erosion and subsequent burial of TerrOC in marine sediments is
83 determined by a variety of climatic and tectonic controls (Eglinton et al 2021). Steep
84 river catchments are typically characterised by enhanced TerrOC export (e.g., Hilton,
85 2017; Kao et al., 2014). However, climate also exerts a key control on TerrOC export
86 (Hilton, 2017; Galy et al 2015) and there is a close link between mean annual run-off
87 and TerrOC export in modern river systems (Hilton, 2017; Wang et al., 2020). Although
88 TerrOC export and burial may help to modulate atmospheric CO_2 over geological
89 timescales, the importance of this process in the past is unknown and the majority of
90 work has been on modern erosive systems (e.g., Hilton, 2017; Galy et al., 2015; Wang
91 et al., 2020).

92 The early Eocene Climatic Optimum (EECO; 53.3-49.1 Ma) is characterised by
93 high temperatures (~10 to 16°C warmer than pre-industrial) (Inglis et al 2020), an

94 intensified hydrological cycle (Carmichael et al., 2016), and lacked continental-scale
95 ice sheets (Francis and Poole, 2002). As such, it can serve as a natural laboratory to
96 understand how TerrOC export and burial operates when it is significantly warmer and
97 wetter than present. Here we use samples recovered from the East Antarctic margin
98 (Integrated Ocean Drilling Program (IODP) Leg 318; IODP Site U1356; ~ 67°S
99 paleolatitude) (Escutia et al 2011b) to determine the amount of TerrOC transported
100 along the land-sea continuum during the earliest Eocene (54.3 to 51.5 Ma), including
101 one of the three DeepMIP intervals (i.e., the Early Eocene Climatic Optimum; 53.3 to
102 49.1 Ma) (Hollis et al 2019). A combination of bulk and molecular proxies are used to
103 fingerprint the type of OC in the East Antarctic margin. We then apply a mixing model
104 approach to quantify TerrOC burial along the East Antarctic margin. We also develop
105 new leaf wax $\delta^2\text{H}$ records and compare these alongside isotope-enabled modelling
106 simulations to determine the relationship between the hydrological cycle and TerrOC
107 export during the early Eocene.

108

109 **2. Methods and Materials**

110 2.1. Site description

111 Site U1356 (modern coordinates: 63°18.6138' S, 135°59.9376' E) is located ~300 km
112 off Wilkes Land, Antarctica, at the transition between the continental rise and the
113 abyssal plain. Current water depth is 3992 meters (Escutia et al 2011a) (Figure 1).
114 Samples were obtained from lithological Unit XI (~948 to 1000 mbsf) that consists of
115 bioturbated hemipelagic early Eocene claystones. Occasional laminated siltstone and
116 sandstone interbeds indicate sporadic gravity flows or bottom current activity reaching
117 the site (Escutia et al 2011a). Lithological Unit XI is dated as early Eocene based upon

118 a combination of published dinoflagellate cyst (dinocyst) biostratigraphy and
119 magnetostratigraphy (Bijl et al 2013, Tauxe et al 2012).

120

121 2.2. Bulk isotope geochemistry

122 Total organic carbon (TOC) measurements were carried out at MARUM, University of
123 Bremen. Approximately 10 cm³ of sediment (n = 360) was freeze-dried and ground to
124 a fine powder using an agate mortar. Total carbon (TC), total organic carbon (TOC),
125 and total sulphur (TS) were measured using a LECO CS-200 Carbon-Sulphur
126 analyzer. Approximately 65 mg of the homogenized sample was weighed in a ceramic
127 cup and burnt in an induction furnace. The resultant CO₂ and SO₂ were then measured
128 with a nondispersive infrared detector to provide a measure of the sedimentary TC
129 and TS content. To determine the TOC content, sediments (~65 mg) were decalcified
130 using 12.5% hydrochloric acid (HCl) to remove carbonate species, followed by a
131 subsequent rinse of the residue with de-ionized water. Samples were subsequently
132 analyzed as described above. Total inorganic carbon (TIC) was determined by
133 subtracting TOC from TC. All data are reported in weight percent (wt. %) dry sample.

134 Bulk carbon isotope analyses of the OC fraction ($\delta^{13}\text{C}_{\text{org}}$) were carried out at
135 MARUM, University of Bremen at 2 cm resolution throughout Cores 318-U1356A-
136 103R to -106R (n = 368). The bulk samples were decalcified with 1 N HCl solution,
137 followed by a subsequent rinse of the residue with de-ionized water. The sample was
138 then combusted using the oven of an HERAEUS CHN-Analyser. The $\delta^{13}\text{C}$ value was
139 subsequently measured with a Finnigan MAT Delta E Mass-spectrometer.
140 Measurements were calibrated using a house standard (milled and decalcified

141 organic-rich surface sediment from the German Wadden Sea), itself calibrated to
142 international standards. Repeatability of the house standard is typically $\pm 0.15\%$.

143

144 2.3. Organic geochemistry

145 A total of 118 sediment samples (~10-15g dry mass) from IODP Site 1356 (948.96 to
146 998.1 mbsf) were freeze-dried and extracted using Accelerated Solvent Extraction with
147 dichloromethane (DCM) and methanol (MeOH) (9:1, v/v, respectively) as the organic
148 solvent. An internal standard (5 α -andosterone) with a known concentration was added
149 to each sample prior to extraction. Excess solvent was removed using rotary
150 evaporation under vacuum. The total lipid extract (TLE) was separated into 'apolar',
151 'ketone', 'ethyl acetate' (EtOAc), and 'polar' fractions over an activated alumina (Al₂O₃)
152 column using hexane:DCM (9:1, v/v; N1), hexane:DCM (1:1, v/v; N2), EtOAc:DCM
153 (1:1 v/v; N3), and DCM:MeOH (1:1, v/v; N4), respectively. Al₂O₃ was activated by
154 oven-drying for at least 4 h at 150 °C.

155 For a subset of earliest Eocene samples (n = 11), we performed urea adduction
156 on the N1 fractions to separate cyclic (i.e. non-adduct) and aliphatic (i.e. adduct)
157 hydrocarbons. To achieve this, 200 μ l of hexane, acetone and urea (10% in MeOH)
158 were added to the saturated hydrocarbon fraction. The sample was frozen for ca. 60
159 minutes until urea crystals formed. Solvent was then removed under a stream of N₂
160 and extracted (x5) with ca. 1ml of *n*-hexane (non-adduct fraction). The urea crystals
161 were then dissolved in 500 μ l of MeOH and 500 μ l of water and the adduct fraction
162 was extracted (x5) with ca. 1ml of *n*-hexane. The adduction procedure was repeated
163 on the adduct fraction once more to ensure all non-adduct material was removed
164 (Inglis et al 2019).

165

166 2.3.1 GC and GC-MS analysis

167 N1 fractions were redissolved in hexane and analyzed using a gas chromatograph
168 (GC; Shimadzu 2010) with a flame ionization detector (FID) and hydrogen as carrier
169 gas at constant pressure (190 kPa). Separation of the different compounds was
170 achieved using the following column: length: 60 m, diameter: 0.25 mm, film thickness:
171 0.25 μm , coating: 100% dimethyl-polysiloxane. The gas chromatograph temperature
172 program increased from 50 to 120 $^{\circ}\text{C}$ at 30 $^{\circ}\text{C min}^{-1}$, then 120-310 $^{\circ}\text{C}$ at 5 $^{\circ}\text{C min}^{-1}$,
173 with a final isothermal time of 20 min at 300 $^{\circ}\text{C}$. Compound identification was
174 confirmed by GC-MS (Shimadzu QP2010-Plus Mass Spectrometer (MS) interfaced
175 with a Shimadzu 2010 GC) using the same column described above. Compounds
176 were identified using retention times and mass spectra. The abundance of *n*-alkanes
177 and hopanes are normalised to TOC. An in-house external standard comprised of 20
178 *n*-alkanes (ranging from C_{16} to C_{37}) was run every 10 samples to monitor analytical
179 performance.

180

181 2.3.2. GC-C-IRMS analysis

182 Compound specific hydrogen isotopic compositions were determined for a subset of
183 samples from the earliest Eocene ($n = 11$) and EECO ($n = 5$). We analysed long-chain
184 (C_{27} , C_{29}) *n*-alkanes via gas chromatography-combustion-isotope ratio mass
185 spectrometry (GC-C-IRMS) using a ThermoFisher Trace GC Ultra coupled to a
186 ThermoFisher Scientific Delta V Isotope Ratio MS. The GC column used was a
187 30 m \times 0.25 mm i.d. fused silica column with ZB1 stationary phase. The H_3 factor was
188 measured daily allowing isotope values to be corrected for protonation reactions

189 occurring within the ion source of the mass spectrometer (Sessions et al 1999). The
190 H_3 factor was typically below 5 ppm mV^{-1} and had a rate of change of less than 0.1
191 day^{-1} . The GC program was as follows: starting temperature 70 °C, rising at 10 °C/min
192 to 300 °C, at which point the oven temperature was held stable for 8 min, giving a total
193 analysis-time of 32 min. Triplicate runs of each sample were performed. The average
194 standard error of the mean (SEM) value for a triplicate measurement was typically 2
195 to 7‰. Each individual sample was co-injected with sacrificial compounds consisting
196 of *n*-pentadecane and ethyl caprylate to condition the reactor. Measured isotope
197 values were normalised by comparing the instrument's response to a suite of *n*-
198 alkanes with a known isotopic composition (B5 mix; supplied by A. Schimmelmann,
199 Indiana University, USA) injected before and after each triplicate of sample runs. The
200 root mean square error determined by replicate measurements of the standard across
201 the course of analyses was between 4 and 7 ‰. Normalised results are reported in
202 standard per mil (‰) notation as δ^2H values relative to Vienna Standard Mean Ocean
203 Water (VSMOW) and calculated against calibrated H_2 gas, introduced directly into the
204 ion source.

205

206 2.4. Lipid biomarker proxies

207 We use biomarkers ratios to assess the source and maturity of organic matter. The
208 average chain length (ACL) expresses the average number of carbon atoms for the
209 long-chain odd-carbon numbered *n*-alkanes (Pancost & Boot 2004). The ACL is
210 defined for *n*-alkanes using the following equation (Eglinton & Hamilton 1967):

211

212 $ACL = (25 \times C_{25}) + (27 \times C_{27}) + (29 \times C_{29}) + (31 \times C_{31}) + (33 \times C_{33}) / (25 + 27 + 29 + 31$
213 $+ 33)$ [e.q. 1]

214

215 The carbon preference index (CPI) reflects the dominance of odd-carbon-numbered
216 relative to even-carbon-numbered *n*-alkane homologues (Bray & Evans 1961).
217 Modern sediments exhibit high CPI values (> 3 to 30) (Diefendorf & Freimuth 2017).
218 This value decreases over time due to diagenesis, approaching values of unity (= 1)
219 in mature rocks and oils. The CPI is defined using the following equation (Bray & Evans
220 1961):

221

222 $CPI = 0.5 \times ((C_{25} + C_{27} + C_{29} + C_{31} / C_{26} + C_{28} + C_{30} + C_{32}) + (C_{27} + C_{29} + C_{31} + C_{33} /$
223 $C_{26} + C_{28} + C_{30} + C_{32}))$ [eq. 2]

224

225 The hopane $\beta\beta/(\alpha\beta+\beta\beta)$ ratio is used to evaluate changes in sediment maturity.
226 Modern sediments exhibit high hopane $\beta\beta/(\alpha\beta+\beta\beta)$ ratios (up to 1) whereas thermally
227 mature rocks and oils are characterised by lower hopane $\beta\beta/(\alpha\beta+\beta\beta)$ ratios
228 (Mackenzie et al 1980), although there can be exceptions (e.g. acidic peats) (Inglis et
229 al 2018). With increasing maturation, C_{31} to C_{35} hopanes also
230 undergo isomerisation at the C-22 position. An increase in $22S/(22R + 22S)$ values is
231 characteristic of increasing thermal maturation (Mackenzie et al., 1980).

232

233 2.5. Mass accumulation rates

234 Biomarker mass accumulation rates (MARs; in $\text{ng}/\text{cm}^2/\text{kyr}^{-1}$) are calculated using
235 linear sedimentation rate (LSR; cm/kyr), dry bulk density (ρ ; g/cm^3) (Dadey et al 1992)
236 and biomarker abundance (ng/g dry sediment):

237

238 $\text{MAR} = \text{LSR} * \rho * \text{biomarker abundance}$ [eq. 3]

239

240 Linear sedimentation rates (LSRs) are calculated between the age tie-points. To avoid
241 age inversions (i.e., within Core 105R-1) we constructed LSRs based on a selection
242 of tie points rather than incorporate all bio- and magnetostratigraphic datums in the
243 age model (see Supplementary Information).

244

245 2.6. Continental shelf area calculations

246 We calculate continental shelf area using the Eocene-Oligocene (ca. 34 Ma)
247 paleotopography published elsewhere (Hochmuth et al 2020a, Hochmuth et al 2020b,
248 Paxman et al 2019). The same calculations are also conducted using Bedmap2
249 (Fretwell et al 2013) for the present day. Continental shelf area for the Eocene is
250 defined as being from the 0 m contour (e.g. shore of an ice free Antarctica) to the -
251 1000m bathymetric contour (Figure S1). We calculate continental shelf area for the
252 region directly adjacent to Wilkes basin and for the whole Antarctic continent
253 (Appendix) in order to understand potential (Terr)OC sequestration.

254

255 2.7. Modelling simulations

256 $\delta^2\text{H}_{\text{wax}}$ is usually employed as a proxy for the isotopic composition of precipitation
257 ($\delta^2\text{H}_{\text{w}}$). However, this integrates a combination of climatic changes including
258 temperature, rainfall amount and characteristics, and atmospheric vapour transport.
259 Here we use modelled $\delta^2\text{H}_{\text{w}}$ obtained via the isotope-enabled Community Earth
260 System Model version 1.2 (iCESM1.2) (Zhu et al 2020, Zhu et al 2019) to compare
261 with our proxy reconstructions and to aid climatic interpretation of $\delta^2\text{H}_{\text{w}}$ (Lee et al
262 2007, Schmidt et al 2007). The iCESM1.2 simulations were performed following
263 protocols of the Deep-time Model Intercomparison Project (Lunt et al 2021, Lunt et al
264 2017) with the Eocene paleogeography and vegetation (56.0 - 47.8 Ma) (Herold et al
265 2014) and atmospheric CO_2 levels of $\times 1$, $\times 3$, $\times 6$ and $\times 9$ the preindustrial value (284.7
266 ppmv). The different atmospheric CO_2 levels span the range of proxy-derived CO_2
267 estimates for the early Eocene (Anagnostou et al, 2020). We also carried out a single
268 pre-industrial simulation to isolate the non- CO_2 component on $\delta^2\text{H}_{\text{w}}$ (i.e., changes due
269 to paleogeography, vegetation, aerosols, and lack of continental ice sheets). All
270 simulations were run at a horizontal resolution of $1.9 \times 2.5^\circ$ (latitude \times longitude) for
271 the atmosphere and land, and a nominal 1° for the ocean and sea ice. Of particular
272 relevance to this study is the capability of iCESM1.2 to simulate the transport and
273 transformation of oxygen and hydrogen isotope ratios ($\delta^{18}\text{O}$ and $\delta^2\text{H}$) in the climate
274 system (Brady et al 2019). Seawater $\delta^{18}\text{O}$ and $\delta^2\text{H}$ in the simulations were initialized
275 from constant values of -1.0‰ and -8.0‰ , respectively, to account for the absence
276 of ice sheets in a hothouse climate (Hollis et al 2019). The simulations were run for
277 more than 2000 years with the surface climate and water isotopes close to
278 equilibration. The simulations capture the magnitude of Eocene global warmth, the
279 reduction in the meridional gradient of sea-surface temperature, and the overall values
280 and distribution of marine planktic foraminifera $\delta^{18}\text{O}$. Readers are referred to Zhu et

281 al. (2019) and Zhu et al. (2020) for details of the experimental setup, equilibration
282 state, and assessment of the simulation results.

283

284 **3. Results**

285 3.1. Bulk geochemistry

286 Total organic carbon (TOC) content ranges between 0.1 and 1.5% (Figure 2). There
287 is a gradual increase in TOC values between 54.4 and 53.5 Ma (Core 106R to 105R)
288 followed by a gradual decrease thereafter (53.5-52.7 Ma, Core 104R to 101R). Bulk
289 $\delta^{13}\text{C}_{\text{org}}$ values range between -25.7 and -28.1 ‰ (Figure 2). The early Eocene record
290 from Site U1356 exhibits several $\delta^{13}\text{C}_{\text{org}}$ variations. These $\delta^{13}\text{C}_{\text{org}}$ cycles and
291 excursions have maximum values of up to -26‰ (in one case -25‰). The average
292 $\delta^{13}\text{C}_{\text{org}}$ values of -26 to -28 ‰ are typical for terrestrial organic matter (Hayes 1993),
293 although this value may be different in the Eocene (Sluijs & Dickens 2012). The
294 cyclicity in the $\delta^{13}\text{C}_{\text{org}}$ record is relatively pronounced in the lower part of the studied
295 section (Cores U1356-105R and -106R; Figure 2), where TOC values are high enough
296 (0.5 – 1.2 wt. %) to provide robust data. In the upper part of the early Eocene section
297 (e.g., Cores 103R and -104R), TOC contents are often lower than 0.5 wt. % and yield
298 a lower fidelity $\delta^{13}\text{C}_{\text{org}}$ signal.

299

300 3.2. Biomarker distributions and mass accumulation rates

301 The N1 fraction contains a range of *n*-alkanes (C_{17} to C_{33}) and is dominated by long-
302 chain homologues (C_{27} to C_{33}). The abundance of long-chain *n*-alkanes ranges
303 between ~1 and 250 ng/g (average: 92 ng/g; Fig. 2d). Long-chain *n*-alkane (plant-
304 derived OC; Figure 3d) mass accumulation rates (MARs; in $\text{ng}/\text{cm}^2/\text{kyr}^{-1}$) increase by

305 two orders-of-magnitude between the earliest Eocene (~54 Ma) and the onset of the
306 EECO (~53 Ma; Figure 3). CPI values (average: 2.8; Figure 4a) indicate that long-
307 chain *n*-alkanes have retained a biological odd-over-even predominance. In
308 combination with high ACL values (~27 to 30; Supplementary Information), this
309 confirms they are derived from terrestrial higher plants (Diefendorf & Freimuth 2017).

310 The apolar fraction also contains a range of bacterial-derived hopanes (C₂₇ to
311 C₃₂) and hopenes (C₂₇ to C₃₀). The abundance of hopanes ranges from ~1 to 191 ng/g
312 (average: 54 ng/g) and exhibits a significant linear correlation with long-chain *n*-alkane
313 abundance ($r^2 = 0.81$, $p < 0.001$; Figure S2). Hopane (soil- and peat-derived OC;
314 Figure 3d) mass accumulation rates (MARs; in ng/cm²/kyr⁻¹) increase by two orders-
315 of-magnitude between the earliest Eocene (~54 Ma) and the onset of the EECO (~53
316 Ma; Figure 3). The dominant hopanes are (22R)-17 β ,21 β (H)-hopane (C₃₀), (22R)-
317 17 α ,21 β (H)-homohopane (C₃₁) and (22R)-17 β ,21 β (H)-homohopane (C₃₁). The C₃₁
318 hopane $\beta\beta/(\alpha\beta+\beta\beta)$ isomerisation ratio ranges between 0.31 and 0.82 (average: 0.65;
319 Figure 4b). The earliest Eocene (54.3 to 53.7 Ma) samples (Core 106R) are
320 characterised by the lowest C₃₁ $\beta\beta/(\alpha\beta+\beta\beta)$ indices (0.4 to 0.6). C₃₁ $\beta\beta/(\alpha\beta+\beta\beta)$ indices
321 increase during the early Eocene (~53.7 Ma) and remain higher (~0.5 to 0.8) thereafter
322 (Cores 105R to 101R) with the exception of two relatively low values between 949 and
323 950 mbsf (Core 101R; Figure 4b). C₃₀ $\beta\beta/(\alpha\beta+\beta\beta)$ indices were stable and high (~0.7-
324 0.8). C₃₁ hopane 22S/(22S + 22R) ratios were stable and low (<0.1).

325 Our analyses are combined with previously published Branched-versus-
326 Isoprenoid Tetraether (BIT) values (Bijl et al 2013). BIT values range between 0.16
327 and 0.62 (average: 0.35) and exhibit higher values (> 0.4 to 0.5) during the EECO
328 (Core 105R to 102R; Supplementary Information) compared to the earliest Eocene
329 (Core 106R).

330

331 3.3. Long-chain *n*-alkane $\delta^2\text{H}$ values

332 Long-chain *n*-alkane $\delta^2\text{H}$ ($\delta^2\text{H}_{\text{wax}}$) values were determined for a subset of samples (*n*
333 = 15) deposited during the earliest Eocene (54.3 to 53.9 Ma) and across the EECO
334 (53.3 to 49.1 Ma). C_{27} *n*-alkane $\delta^2\text{H}$ values average -190‰ (*n* = 15, standard error of
335 the mean [SEM]: $\pm 4.0\text{‰}$) with a range from -140 to -222‰ (Figure 5a). C_{29} *n*-alkane
336 $\delta^2\text{H}$ values average -194‰ (*n* = 15, SEM: $\pm 2.9\text{‰}$) with a range from -158 to -224‰
337 (Figure 5a). Both the C_{27} and C_{29} *n*-alkanes co-vary and exhibit a positive linear
338 relationship ($r^2 = 0.78$, $p < 0.001$), indicating that they are derived from a similar source
339 (i.e. higher plants) (Figure S3).

340

341 4. Discussion

342 4.1. Enhanced TerrOC burial on the East Antarctic margin during the EECO

343 The export of terrestrial OC along the terrestrial-aquatic continuum can be evaluated
344 by analysing the distribution of lipid biomarkers within the Wilkes Land sediment
345 samples. The hydrocarbon distribution is dominated by long-chain *n*-alkanes with an
346 odd-over-even predominance. High CPI values (~ 3 ; Figure 4a) together with the
347 dominance of long-chain *n*-alkanes (relative to shorter-chain homologues) confirms
348 that they are derived from the epicuticular wax of terrestrial higher plants (Eglinton &
349 Hamilton 1967). The hydrocarbon assemblage also contains abundant bacterial-
350 derived hopane biomarkers. Hopanes can be produced in terrestrial and marine
351 environments by a range of bacteria (Talbot & Farrimond 2007). However, the positive
352 linear relationship between long-chain *n*-alkanes (terrestrial higher plants) and

353 hopanes ($r^2 = 0.81$; $p < 0.001$; Figure S2) indicates that here, both are sourced from
354 the terrestrial biosphere.

355 Long-chain *n*-alkane (plant-derived OC; Figure 4d) and hopane (soil-derived
356 OC; Figure 4c) mass accumulation rates (MARs; in $\text{ng}/\text{cm}^2/\text{kyr}^{-1}$) increase by two
357 orders-of-magnitude between the earliest Eocene (~ 54 Ma) and the onset of the EECO
358 (~ 53 Ma). This significant increase indicates enhanced burial of plant- and soil OC
359 during the early Eocene, with highest accumulation rates during the EECO. Note that
360 we performed the same calculations assuming a constant LSR (1.82 cm/year, i.e., the
361 average LSR rate) and find similar MARs during the EECO (~ 53 Ma; Figure S4-5).
362 Delivery of TerrOC into the marine realm is consistent with the high abundance of
363 terrestrial palynomorphs (i.e., pollen and spores; (Contreras et al 2013, Pross et al
364 2012) and other terrestrial biomarkers (including high BIT indices; Pross et al., 2012;
365 Bijl et al., 2013; Figure 3a) in the same sediments. These findings are analogous to
366 those observed in modern (high-latitude) environments, where large river systems in
367 northern Eurasian and North American (Hilton et al., 2015; Sparkes et al., 2015;
368 Semiletov et al., 2011; Guo et al., 2004) are capable of transporting large quantities of
369 TerrOC from soils into the marine realm.

370 The earliest Eocene sediments also contain an unusually high abundance of
371 the 'thermally-mature' C_{31} $17\alpha,21\beta(\text{H})$ hopane stereoisomer and exhibit relatively low
372 C_{31} hopane $\beta\beta/(\alpha\beta+\beta\beta)$ ratios (0.4-0.7; Figure 4b). This implies delivery of biospheric
373 (i.e., thermally immature) and petrogenic OC (i.e., thermally mature) into the marine
374 realm. Input of petrogenic OC should be associated with the delivery of other
375 thermally-mature hopanes (e.g., C_{30} $\alpha\beta$ hopanes). However, C_{30} hopane $\beta\beta/(\alpha\beta+\beta\beta)$
376 values remain relatively high (>0.7), implying minimal petrogenic OC input. Modern
377 acidic bogs can be dominated by the 'thermally mature' C_{31} $17\alpha,21\beta(\text{H})$ isomer (Inglis

378 et al 2018), but typically lack other thermally-mature hopanes (e.g., C₃₀ αβ hopanes;
379 Inglis et al., 2018). We therefore argue that the occurrence of the ‘thermally mature’
380 C₃₁ 17α,21β(H) stereoisomer within an otherwise immature hopane assemblage
381 represents input of acidic peat into the marine realm. The occurrence of the ‘thermally
382 mature’ C₃₁ 17α,21β(H) stereoisomer within recent coastal marine sediments has also
383 been interpreted to reflect input of peat from the surrounding catchment (Smittenberg
384 et al 2004). This hypothesis helps explain why C₃₀ and C₂₉ hopane ββ/(αβ+ββ) values
385 remain relatively high throughout (>0.7). It is also consistent with the lack of other
386 thermally-mature biomarkers (e.g. αβ 22S-hopanoid isomers, bisnorhopanes,
387 benzohopanes, triaromatic steroids) and relatively high CPI values (>3; Figure 4a).

388 Input of acidic peat into the marine realm is consistent with the occurrence of
389 *Stereisporites* spores in the Site U1356 sediments (Pross et al., 2012; Figure S6) that
390 can be attributed to the Sphagnaceae family, which today only comprises the genus
391 *Sphagnum* (commonly known as peat moss). Despite the low abundances registered,
392 the occurrence of *Stereisporites* spores in the Site U1356 sediments appears
393 noteworthy for various reasons. Firstly, *Stereisporites* spores are relatively rare or
394 absent in early Cenozoic sediments (see Inglis et al., 2015 and ref. therein). Secondly,
395 spores are generally strongly underrepresented in sporomorph assemblages from
396 marine sediments due to the selective nature of sporomorph transport as a function of
397 transport distance (Contreras et al 2014, Moss et al 2005). As such, even their
398 occurrence in low numbers (e.g., ~1% of total sporomorph assemblage) can have
399 substantial ecological significance and points towards the existence of peatlands
400 and/or forested mires within the Site U1356 catchment area. The input of acidic peat
401 into the marine realm is also consistent with the results of biome-based modelling

402 suggesting that ~15–20% of Antarctica was capable of supporting wetland and peat
403 formation during the Eocene (DeConto et al 2012).

404

405 4.2. Quantifying TerrOC burial on the East Antarctic margin during the EECO

406 Lipid biomarkers indicate that a considerable fraction of the OC pool on the early
407 Eocene Wilkes Land continental margin is derived from the terrestrial biosphere.
408 However, the relative proportion of terrestrial vs marine OC must be known in order to
409 quantify TerrOC burial (Weijers et al 2009). In modern settings, bulk OC $\delta^{13}\text{C}$ values
410 ($\delta^{13}\text{C}_{\text{org}}$) can discriminate between marine OC (^{13}C -enriched) and vascular plant
411 and/or soil OC (^{13}C -depleted). However, marine OC can be relatively ^{13}C -depleted
412 during the Eocene (up to 7‰ relative to modern) (Sluijs & Dickens 2012) due to
413 enhanced fractionation under higher CO_2 (Freeman & Hayes 1992).

414 Here we use BIT indices (Hopmans et al 2004) to characterise the fraction of
415 terrestrial vs marine OC and employ a two end-member mixing-model:

416

$$417 \quad f_{\text{terrestrial}} = (X_{\text{sample}} - X_{\text{marine}}) / (X_{\text{terr}} - X_{\text{marine}}) * 100 \text{ (eq. 4)}$$

418

419 where $f_{\text{terrestrial}}$ is the terrestrial OC fraction and X_{sample} is the BIT index for the sediment
420 sample (following Weijers et al., 2009). The marine- (X_{marine}) and terrestrial end-
421 member (X_{terr}) for BIT is 0 and 0.91, respectively (Hopmans et al., 2004). The BIT
422 index can reach a theoretical value of 1. However, BIT values in globally-distributed
423 soil samples average 0.91 (Weijers et al 2006). Using our mixing-model approach, we
424 estimate that the proportion of terrestrial OC in the Wilkes Land sediments ranges
425 between 20% (minimum) and 65% (maximum). As the BIT index captures the
426 proportion of soil OC (rather than soil and plant OC), this approach may underestimate

427 the proportion of terrestrial OC in the Wilkes Land sediments. The BIT index has been
428 shown to be controlled strongly by crenarchaeol—rather than brGDGT—
429 concentrations, and may lead to a decoupling between BIT values and other terrestrial
430 OC tracers (Smith et al 2012). However, the fractional abundance of crenarchaeol is
431 stable throughout the early Eocene (Bijl et al., 2013), arguing that BIT is largely
432 controlled by changes in brGDGT abundance. The BIT index can also be influenced
433 by in-situ marine or fluvial brGDGT production. However, brGDGTs reflect typical soil-
434 distributions, with no abnormal Isomerization Ratio or #Rings_{tetra} values (Sinninghe
435 Damsté 2016). This suggests negligible marine in situ or river production) and is
436 consistent with other early Eocene sites in the southwest Pacific Ocean (e.g., Bijl et
437 al., 2021).

438 The proportion of TerrOC can then be multiplied alongside TOC MARs to
439 provide a first-order approximation of TerrOC sequestration along the Wilkes Land
440 margin. This first requires an estimate of the Wilkes Land continental shelf area
441 (118,675 km²). Here we assume that the early Eocene shelf was similar to that of the
442 latest Eocene (Paxman et al 2019) and was significantly smaller than the present day
443 continental shelf area (170,731 km²). We also assume that the sediment package
444 thickness (and thus, %TerrOC) is uniform. Assuming that 20 to 65% of the OC pool in
445 the Wilkes Land sediments is terrestrial-derived (see above), this implies the burial of
446 20 to 72 PgC of TerrOC the depositional interval (54.3 and 51.5 Ma). If we assume
447 similar climate, vegetation distributions and transport mechanisms across the entire
448 Antarctic shelf area (3,947,850 km²), the potential mass of TerrOC sequestered is
449 much larger (~410 to 1300 PgC) and equates to a burial flux of 0.14-0.46 TgC/yr.
450 These estimates are relatively modest compared to highly-erosive mountainous river
451 catchments (e.g., Taiwan; > 8 TgC/yr; Kao et al., 2014) and large river systems (e.g.,

452 the Mackenzie River; 2.2 (+1.3/-0.9) TgC/yr; Hilton et al., 2015). To refine these
453 estimates, additional studies at different localities are required. Additional seismic data
454 may help to improve estimates of offshore sediment thickness, whilst the application
455 of binary (or ternary) mixing models based on other geochemical proxies (e.g., C/N
456 ratios, *n*-alkane/alkenones ratios) will help to refine our estimates of TerrOC burial
457 (see Weijers et al., 2009). Until then, our estimates should be treated with caution.

458 Although these estimates incorporate various uncertainties, our work highlights
459 the potential importance of continental margins as depocentres for TerrOC burial
460 during the early Eocene. This is consistent with High Arctic continental shelf sediments
461 deposited between the Palaeocene-Eocene Thermal Maximum (PETM; ~56 Ma) and
462 Eocene Thermal Maximum 2 (ETM2; ~53 Ma), which are dominated by long-chain *n*-
463 alkyl lipids and low hydrogen indices (Boucsein & Stein 2009, Weller & Stein 2008).
464 Terrigenous input—which is strongly correlated with TerrOC export in modern settings
465 (e.g., Galy et al., 2015)—also increases during the EECO, including in New Zealand
466 (e.g. Mead Stream) (Nicolo et al 2007, Slotnick et al 2012) and the North Iberian
467 margin (Payros et al 2015). Enhanced TerrOC burial may be further promoted by
468 larger, flatter drowned shelf areas during the Eocene (Bowen 2013, Sømme et al
469 2009). Taken together, this suggests an important role for continental shelves in
470 storing terr(OC) during the Eocene. This is especially important given the evidence for
471 low (terr)OC burial in open ocean Eocene environments (Olivarez Lyle & Lyle 2006).

472

473 **4.3. Hydroclimate regulates TerrOC export during early Eocene**

474 Enhanced TerrOC burial in coastal marine sediments could be driven by expansion of
475 the terrestrial biosphere during the EECO. However, TerrOC export is largely

476 controlled by precipitation (Hilton 2017) and the capacity of rivers to erode and
477 transport TerrOC (Galy et al 2015). Here we explore whether hydroclimate regulates
478 OC export from the terrestrial biosphere. To assess this, we use the hydrogen isotopic
479 composition ($\delta^2\text{H}$) of leaf wax (C_{27} and C_{29} *n*-alkane) biomarkers in a subset of
480 samples deposited during the earliest Eocene (54.3 to 53.9 Ma) and EECO. C_{27} and
481 C_{29} *n*-alkane $\delta^2\text{H}$ values average -190 and -195‰, respectively, and are comparable
482 to those reported from other early Eocene-aged sites from similar latitudes (Speelman
483 et al 2010). The isotopic composition of precipitation ($\delta^2\text{H}_w$) is estimated by assuming
484 an apparent fractionation factor (${}^2\text{H}\epsilon_{\text{wax/w}}$) which integrates a range of variables (e.g.
485 soil- and/or leaf water evapotranspiration, biosynthetic effects, vegetation type).
486 Theoretical models have incorporated evaporative ${}^2\text{H}$ -enrichment of soil water into
487 predictions of ${}^2\text{H}\epsilon_{\text{wax/w}}$ (Konecky et al 2019, Smith & Freeman 2006). However, in
488 settings with extensive evaporation, woody plants take up rainwater opportunistically
489 via different rooting strategies (Fan et al 2017), resulting in minimal xylem water ${}^2\text{H}$ -
490 enrichment. Although leaf water ${}^2\text{H}$ -enrichment is more variable (Daniels et al 2017,
491 Feakins et al 2016, Kahmen et al 2013), it is significantly reduced in warm and wet
492 climates (Feakins et al 2016). As the East Antarctic margin was characterised by
493 elevated temperatures ($> 16 \pm 5$ °C) and high precipitation (> 1300 - 1600 mm/year)
494 (Pross et al 2012), the impact of leaf water ${}^2\text{H}$ - enrichment is likely minimal.

495 Changes in the plant community can also influence the apparent fractionation
496 between $\delta^2\text{H}_{\text{wax}}$ and $\delta^2\text{H}_w$ values. To account for the impact of vegetation isotope
497 effects upon leaf wax $\delta^2\text{H}$ values, we developed a pollen-corrected fractionation factor
498 (${}^2\text{H}\epsilon_{\text{wax/w-corr}}$) using a three endmember mixing model (following Feakins 2013). This
499 includes C_3 gymnosperms, C_3 dicots and pteridophytes (i.e. ferns) and is defined as:

500

501 ${}^2\text{H}\epsilon_{\text{wax/w-corr}} = f_{\text{gymnosperm}} * \epsilon_{\text{gymnosperm}} + [f_{\text{dicot}} * \epsilon_{\text{dicot}}] + [f_{\text{pteridophytes}} * \epsilon_{\text{pteridophytes}}]$ (eq. 5)

502

503 Where ϵ is based on the mean fractionation factor in modern plant types (Sachse et
504 al 2012) and f is based upon the percentage of pollen in adjacent samples (Contreras
505 et al 2013, Pross et al 2012). Our pollen-corrected fractionation factor ranges between
506 -107 and -113‰ and is higher than assumed in some Eocene studies (-130‰)
507 (Handley et al 2012, Pagani et al 2006) but similar to an updated global dataset of
508 ${}^2\text{H}\epsilon_{\text{wax/w}}$ in modern systems (e.g., -121‰ for C_{29} *n*-alkane; McFarlin et al., 2019). When
509 our pollen-corrected fractionation factor is applied to our leaf wax dataset, we obtain
510 $\delta^2\text{H}_{\text{w-corr}}$ values which average -82‰ (C_{27} *n*-alkane; Figure 5b) and -83‰ (C_{29} *n*-
511 alkane; Figure 5b). These values are ${}^2\text{H}$ -enriched relative to modern values (-99‰;
512 based on iCESM1.2 pre-industrial x1 CO_2 simulation; see also below). There are
513 several reasons for this. Firstly, local air temperatures will yield more ${}^2\text{H}$ -enriched
514 water vapour (a temperature effect) (Poulsen et al 2007). Secondly, under warmer
515 climates, decreased rainout efficiency at the low-latitudes (due to a reduction in vertical
516 moisture transport) will result in more ${}^2\text{H}$ -enriched precipitation at the mid-to-high
517 latitudes (e.g., Pagani et al 2006). Our pollen-corrected $\delta^2\text{H}_{\text{w}}$ values exhibit
518 considerable variability in the earliest Eocene interval (ca. 30 to 40‰; Figure 5). This
519 may represent changes in the hydroclimate regime. However, it could also record
520 changes in OC source region. Indeed, sporomorph distributions indicate that TerrOC
521 is derived from both the lowlands and higher-altitude hinterlands (Pross et al., 2012).
522 Although leaf waxes typically provide a spatially-integrated perspective, the spatial and
523 temporal integration may vary by compound type and/or change through time. Future
524 work using other compound classes (e.g., long-chain *n*-alkanoic acids) may help to

525 differentiate between lowland (proximal) and highland (distal) OC sources (see
526 Feakins et al 2018, Hemingway et al 2016).

527 Our pollen-corrected $\delta^2\text{H}_w$ values indicate higher average $\delta^2\text{H}_w$ values (e.g., -
528 66‰ for the C_{29} *n*-alkane) during the EECO compared to the earliest Eocene (e.g.,
529 e.g., -90‰ for the C_{29} *n*-alkane; Figure 5b). This is consistent with an increase in
530 poleward moisture transport in response to higher CO_2 and higher temperatures. To
531 confirm that the higher average $\delta^2\text{H}_w$ values are due to higher CO_2 and/or higher
532 temperatures, we calculate the precipitation-amount weighted $\delta^2\text{H}_w$ in isotope-enabled
533 CESM simulations with a range of atmospheric CO_2 concentrations ($\times 1$, $\times 3$, $\times 6$ and
534 $\times 9$ the preindustrial level; see section 2.7). The modelled $\delta^2\text{H}_w$ exhibits a monotonic
535 increase with the CO_2 concentrations in the simulations with values of -106‰ in $\times 1$
536 CO_2 , -96‰ in $\times 3$ CO_2 , -91‰ in $\times 6$ CO_2 and -81‰ in $\times 9$ CO_2 (Figure 6b). This $\delta^2\text{H}_w$
537 sensitivity to CO_2 levels in isotope-enabled climate models has been attributed to
538 warmer temperatures and/or the associated reduction in isotopic rainout at low-
539 latitudes (Poulsen et al., 2007; Speelman et al., 2010). The qualitatively consistent
540 results between model simulations and proxy data supports our interpretation that the
541 observed increase in proxy-inferred $\delta^2\text{H}_w$ values between the earliest Eocene and
542 EECO (~ 20 to 40 ‰) is attributable to a CO_2 -induced warming. There is also close
543 data-model agreement in the absolute $\delta^2\text{H}_w$ values (Figure 6), supporting our data-
544 based reconstructions. The impact from non- CO_2 boundary conditions on $\delta^2\text{H}_w$ can
545 also be evaluated by comparing iCESM1.2 Eocene $\times 1$ simulations against the pre-
546 industrial simulation. This shows that the combined non- CO_2 boundary conditions (i.e.,
547 paleogeography, vegetation, aerosols, removal of continental ice sheets) decrease
548 $\delta^2\text{H}_w$ by ~ 7 ‰. This implies that non- CO_2 boundary conditions exert a small secondary
549 control on $\delta^2\text{H}_w$ values at our site. The iCESM1.2 simulations ($\times 3$, $\times 6$ and $\times 9$ PI CO_2

550 simulations) also exhibit high MAP estimates (~1310, 1520, and 1720 mm/year,
551 respectively; Figure 7c-d) during the early Eocene. These are higher than previous
552 simulations (Carmichael et al 2016) but more consistent with existing sporomorph-
553 based estimates (>1300 mm/year [paratropical rainforest biome] and >1600 mm/year
554 [temperate rainforest biome]; Figure 7a-b) (Pross et al 2012). Slightly lower values
555 (800–1350 mm/year) have been reconstructed for Site U1356 based on climofunctions
556 applied to detrital geochemistry (Passchier et al 2013) and likely reflects the
557 differences in the catchment sourcing.

558 Taken together, our results reveal that EECO was characterised by enhanced
559 poleward moisture transport relative to the earliest Eocene (~54 Ma; Figure 5-6). The
560 EECO also coincides with an interval of enhanced delivery of terrestrial OC into the
561 marine realm (Figure 3), implying a causal link between the hydrological cycle and
562 TerrOC burial at our site. Crucially, TerrOC burial can help to remove the atmospheric
563 CO₂ and could serve as a negative feedback to decrease the surface temperature
564 during intervals of global warmth. Compared to the silicate weathering feedback
565 (which operates on the timescales of a million years), the organic carbon cycle
566 operates over shorter timescales. As such, it could also be an important negative
567 feedback during shorter-term warming events (e.g., hyperthermals).

568

569 **5. Conclusions**

570 The transport and burial of biospheric organic carbon in coastal marine sediments is
571 an important carbon sink in modern settings. However, the relative importance of this
572 feedback in past warm climates remains a major gap in our understanding. Using a
573 multi-proxy approach, we find a significant increase in the accumulation rate of
574 terrestrial biomarkers in early Eocene coastal marine sediments from offshore East

575 Antarctica. As the early Eocene Antarctic continent was covered by a vast terrestrial
576 OC reservoir ($> 10^{15}$ PgC), the transport and subsequent burial of biospheric organic
577 carbon in coastal marine sediments could have acted as a key CO₂ sink. This study
578 highlights the importance of the terrestrial biosphere during past warm climates and
579 its potential role as a negative feedback to stabilize the surface temperature.

580 **Open Research**

581 The geochemical and numerical data used in the study are available at Pangaea
582 (<https://www.pangaea.de/>) and associated with a CC-BY 4.0 license.

583

584 **Authorship tiers and contributions**

585 G.N.I and J.B. designed the study. GNI drafted the manuscript and designed the
586 figures. G.N.I., J.B. and J.L.T conducted biomarker analysis. U.R. conducted total
587 organic carbon (TOC) and bulk organic carbon isotope analysis. J.P provided pollen
588 data. C.P and J.Z. provided CESM1.2 model results. S.S.R.J provided continental
589 shelf estimates. All the authors discussed the results and contributed to the final text.

590

591 **Declaration of competing interest**

592 The authors declare that they have no known competing financial interests or personal
593 relationships that could have appeared to influence the work reported in this paper.

594

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612

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914 **Captions**

915

916 **Figure 1:** Continental configurations of the Australian sector of the Southern Ocean
917 during the early Eocene (subchron C24n; 53.8 Ma). Map shows the position of the
918 study site (U1356). Dark grey areas reflect present day shorelines, and light grey areas
919 are submerged continental blocks above 3,000 m water depth. Modified from Bijl et
920 al., 2013.

921

922 **Figure 2:** Organic carbon burial along the East Antarctic margin during the early
923 Eocene. a) $\delta^{13}\text{C}_{\text{org}}$ (‰), b) total organic carbon (TOC; wt. %), c) bacterial-derived
924 hopanoids (ng/g), d) plant-derived long-chain *n*-alkanes (ng/g). Black segments
925 indicate intervals of no core recovery.

926

927 **Figure 3:** Enhanced terrestrial organic carbon burial along the East Antarctic margin
928 during the Eocene. a) branched-versus-isoprenoidal tetraether (BIT) index (Bijl et al.,
929 2013), b) total organic carbon (TOC; g/cm²/kyr⁻¹) mass accumulation rate (MAR), c)
930 hopanioid MAR (ng/cm²/kyr⁻¹), d) long-chain *n*-alkane MAR (ng/cm²/kyr⁻¹). Note that

931 (a), (b) and (c) are on a logarithmic scale. Black segments indicate intervals of no core
932 recovery.

933

934 **Figure 4:** Lipid biomarker thermal maturity ratios in early Eocene-aged Wilkes Land
935 sediments. a) long-chain *n*-alkane carbon preference index (CPI), b) C₃₁ hopane
936 $\beta\beta/(\alpha\beta+\beta\beta)$ indices. Black segments indicate intervals of no core recovery.

937

938 **Figure 5:** Enhanced poleward moisture transport to the East Antarctic margin during
939 the early Eocene). a) δ^2H_{wax} estimates inferred from the C₂₇ *n*-alkane (orange) and C₂₉
940 *n*-alkane (blue). Error bars represent $\pm 1\sigma_{\bar{x}}$ (following Polissar and D'Andrea, 2014).
941 b) pollen-corrected proxy-derived δ^2H_w estimates inferred from the C₂₇ *n*-alkane
942 (orange) and C₂₉ *n*-alkane (blue). Error bars indicate $\pm 1\sigma$ propagated errors of the
943 analytical error and the fractionation uncertainty (following Feakins, 2013). Here we
944 use the standard deviation from modern plant type groups (i.e., C₃ angiosperms, C₃
945 gymnosperms, ferns; Sachse et al., 2012) to calculate the fractionation uncertainty
946 (see Appendix).

947

948 **Figure 6:** Reconstructed and simulated δ^2H_w on the East Antarctic margin during the
949 early Eocene: a) δ^2H_{w-corr} estimates inferred from the C₂₇ *n*-alkane (light grey) and C₂₉
950 *n*-alkane (dark grey). Dashed line represents median value. Box represents 1st and 3rd
951 interquartile range. Whiskers represent 95% confidence interval. b) model-derived
952 δ^2H_w estimates inferred from iCESM1.2. Open triangle represents pre-industrial x1
953 CO₂ simulation. Closed circles represent Eocene simulations at four different CO₂
954 levels (x1, x3, x6 and x9 pre-industrial CO₂).

955

956 **Figure 7:** Long-term hydroclimate variability along the East Antarctic margin during
957 the early Eocene. a) mean annual precipitation (MAP) estimates for the temperate
958 rainforest biome based on bioclimatic analysis (Pross et al., 2012), b) MAP estimates
959 for the paratropical rainforest biome based on bioclimatic analysis (Pross et al., 2012),
960 c) and d) MAP estimates obtained via iCESM1.2 for four different CO₂ levels (x1, x3,
961 x6 and x9 pre-industrial CO₂). Whiskers represent precipitation estimates during
962 summer (JJA) and winter (DJF) months. Black segments indicate intervals of no core
963 recovery.

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