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Southern Ocean temperature records and ice-sheet models demonstrate rapid Antarctic ice sheet retreat under low atmospheric CO₂ during Marine Isotope Stage 31

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1 **Southern Ocean temperature records and ice-sheet models demonstrate rapid**
2 **Antarctic ice sheet retreat under low atmospheric CO₂ during Marine Isotope**
3 **Stage 31.**
4

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18

19 **Highlights**

- 20 - We quantify precisely the magnitude of ocean warming using the **first ever**
21 Southern Ocean molecular paleo-temperature reconstructions during the warm
22 Marine Isotope Stage 31 (MIS31- 1.085 and 1.055 million years ago).
23 - Our data show a sustained surface Southern Ocean warming and a collapse of the
24 Antarctic and sub Antarctic ocean fronts during that warm period and under low
25 atmospheric CO₂ concentrations.
26 - We use sea surface temperature reconstructions to test the scenarios for the AIS
27 retreat during MIS31 using coupled ice-sheet/ice-shelf model.
28 - We propose a two-step model for deglaciating West Antarctica which involves
29 mild ocean warming (a new temperature threshold) which forces ice margin
30 retreat followed by rapid ocean warming as the ice sheet retreats.
31 - Our work shows that the Paris Agreement target temperature of 1.5°C is
32 sufficient to drive runaway retreat of the West Antarctic Ice Sheet. We derive this
33 conclusion from the robust, ocean temperature proxy record and ice sheet
34 simulation.
35

36

37 **Abstract**

38 Over the last 5 million years, the Earth's climate has oscillated between warm
39 (interglacial) and cold (glacial) states. Some particularly warm interglacial periods
40 (i.e. 'super-interglacials') occurred under low atmospheric CO₂ and may have
41 featured extensive Antarctic ice sheet collapse. Here we focus on an extreme super-
42 interglacial known as Marine Isotope Stage 31 (MIS31), between 1.085 and 1.055
43 million years ago and is the subject of intense discussion. We reconstructed the first
44 Southern Ocean and Antarctic margin sea surface temperatures (SSTs) from organic
45 biomarkers and used them to constrain numerical ice sheet-shelf simulations. Our
46 SSTs indicate that the ocean was on average 5 °C (±1.2 °C) warmer in summer than
47 today between 50 °S and the Antarctic ice margin. Our most conservative ice sheet
48 simulation indicates a complete collapse of the West Antarctic Ice Sheet (WAIS) with
49 additional deflation of the East Antarctic Ice Sheet. We suggest the WAIS retreated
50 because of anomalously high Southern Hemisphere insolation coupled with the
51 intrusion of Circumpolar Deep Water onto the continental shelf under poleward-
52 intensified winds leading to a shorter sea ice season and ocean warming at the
53 continental margin. In this scenario, the extreme warming we observed likely reflects
54 the extensively modified oceanic and hydrologic system following ice sheet collapse.
55 Our work highlights the sensitivity of the Antarctic ice sheets to minor oceanic
56 perturbations that could also be at play for future changes.

57

58 **Keywords**

59 Super-interglacial, Pleistocene, Paleoceanography, Paleoclimate modeling,
60 Antarctica, Southern Ocean, Organic Geochemistry

61

62 **1. Introduction**

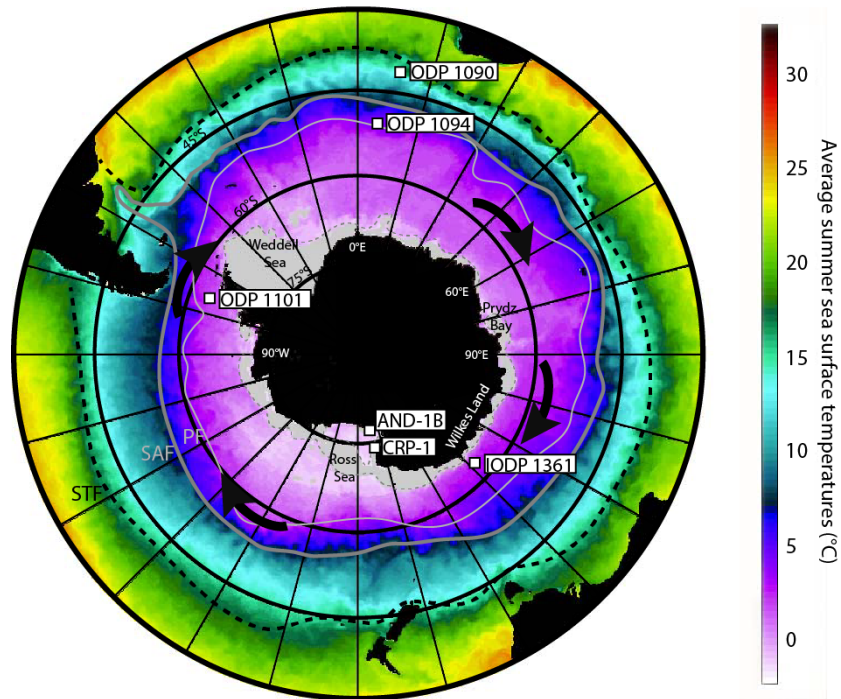
63 Some geological records suggest that during the early Pleistocene Marine Isotope
64 Stage 31 (MIS31 ~1.08 to 1.06 Ma), under preindustrial atmospheric CO₂ conditions
65 (Lüthi et al. 2008; Hönisch et al., 2009), global mean sea level rose by ~20 m (Raymo
66 et al., 2006). Recent climate simulations estimate that the global mean temperature
67 increased by 1.2 °C during boreal summers in the Northern Hemisphere (Justino et al.,
68 2019). The magnitude of the sea level high stand and geological evidence suggest a
69 major retreat of the West Antarctic Ice Sheet (WAIS), part of the East Antarctic Ice
70 Sheet (EAIS) and probably also the Greenland Ice Sheet (Hearty et al., 1999; Raymo
71 et al., 2006; Dutton and Lambeck, 2012; Dutton et al., 2015) during this time. In
72 sediment drill core records from around Antarctica, MIS31 is identified by the abrupt
73 appearance of foraminiferal oozes and coccolith-bearing sediments in the Weddell
74 Sea and Prydz Bay and of bioclastic limestone in the Ross Sea (Bohaty et al., 1998;
75 Scherer et al., 2003; Villa and Wise, 1998; Villa et al., 2008; Villa et al., 2012). This
76 indicates a significant southward migration of the Polar Front (PF), surface ocean
77 warming and open-marine conditions in the Ross and Weddell Seas, with a reduction
78 in sea ice extent relative to the present day (Becquey and Gersonde, 2002; Scherer et
79 al., 2008; Winter et al., 2010; Martinez-Garcia et al., 2010; McKay et al., 2012). To
80 date, elevated ocean surface temperatures have been inferred from paleoecological
81 and sedimentological changes observed in sediment cores around the Antarctic
82 margin (Bohaty et al., 1998; Scherer et al., 2003; 2008; Winter et al., 2010; Dunbar,
83 2012; Teitler et al., 2005). However, the scarcity of precise surface ocean temperature
84 data has so far limited models from realistically simulating the ice sheet and Southern
85 Ocean responses during MIS31 (Supplementary Information Section 1).
86 Consequently, it has also not been possible to isolate the cause of ice melt and sea

87 level rise although an unusual orbital configuration and elevated polar summer
88 insolation have been implicated (Laskar et al., 2004; Scherer et al., 2008).
89 Ice-sheet/ice-shelf and global climate simulations have highlighted the prominent role
90 the ocean plays in controlling the position of the grounding lines of the WAIS and
91 portions of the EAIS (Jacobs et al., 1992; Dinniman et al., 2016). Here we generate
92 the first SST time series from molecular paleothermometric methods (alkenone
93 unsaturation index U_{37}^k - Prahl et al., 1988; and the Long Chain Diol Index LDI -
94 Rampen et al., 2012) from Southern Ocean and Antarctic margin deep-sea sediment
95 cores. We then use these sea surface temperature reconstructions to test the scenarios
96 for the AIS retreat during MIS31 using coupled ice-sheet/ice-shelf model (Golledge,
97 et al., 2015).

98 **2. Material and Methods**

99 **2.1. Site locations and modern oceanography**

100 To document the evolution of the sea surface temperatures (SST) in the Atlantic
101 sector of the Southern Ocean during MIS31, we have selected the ODP Site 1094,
102 situated approximately 2° south of the present Antarctic Polar Front (Schneider-Mor
103 et al., 2008 - Fig. 1). To investigate the SSTs in the peri-Antarctic region, we have
104 selected the IODP Site 1361 (Wilkes Land - Escutia et al., 2011) and the ODP Site
105 1101 (Antarctic Peninsula – Barker et al., 1999). We supplement our latitudinal
106 temperature dataset with the high resolution alkenone-derived SSTs record from
107 Martinez-Garcia et al. (2010) for the ODP Site 1090 situated North of the modern
108 subantarctic front (SAF) and the Mg/Ca measurements on planktic foraminifera from
109 the Cape Roberts Project Site 1 (CRP-1) drilled beneath the northwestern Ross Ice
110 Shelf (Dunbar 2012 - Fig. 1). Sample sites were selected on the basis of previously
111 reported occurrence of MIS31 sediments in the literature.



112

113 **Figure 1:** Location of the principal ocean frontal systems in the Southern Ocean (Orsi
 114 et al., 1995) and drill sites chosen for this study. The modern summer sea surface
 115 temperature map was modified after HYCOM Consortium for Data-Assimilative
 116 Ocean Modelling Southern Ocean Atlas
 117 (<http://www7320.nrlssc.navy.mil/GLBHycom1-12/skill.html>). The general path for the
 118 Antarctic Circumpolar Current (ACC) is shown with the black arrows (Orsi et al.,
 119 1995). STF: subtropical Front (black dashed)- SAF: Subantarctic Front (dark grey)-
 120 PF: Polar Front (light grey).

121

122 **ODP Site 1090 (42°54.8'S, 8°53.9'E)** was drilled on the southern flank of the
 123 Agulhas Ridge, at a water depth of 3699 m. It lies within the modern subantarctic
 124 waters between the subtropical (STF) and the subantarctic (SAF) fronts in the Atlantic
 125 sector of the Southern Ocean. The modern annual SST at this location is 10 °C (9 °C
 126 in July-September - 11 °C in January-March; Levitus, 1994). Pleistocene sediment
 127 consists of foraminiferal nannofossil ooze and foraminifer-bearing nannofossil ooze.

128 **ODP Site 1094 (53°10.8'S, 5°7.8'E)** is located south of the Polar Front, north of the
 129 Weddell Gyre (Antarctic Circumpolar Current (ACC) -boundary) and within the
 130 biogenic silica belt (Gersonde et al., 1999). The site is located close to Bouvet Island,
 131 in the ice-free Antarctic Zone, at 2807 m water depth. The investigated sediments

132 (1094D 11H to 13H) mainly consist of carbonate bearing diatom ooze, with varying
133 amounts of foraminifera and nannofossils (between 20% and 45%). The mean annual
134 SST in this area is 0.7 °C, with mean temperatures of 0.2 °C in winter and of 2.5 °C
135 during austral summer (Levitus, 1994).

136 **IODP Site U1361 (64.4°S, 143.88°E)** is located on the Wilkes Land margin, on the
137 eastern levee of the Jussieu sub-marine channel. The mean annual SST in this area is
138 around 0 °C (Levitus, 1994). Sediments from 1361A-2H consist of silty clay and
139 calcareous silty clay. Sediments from 1361A-3H are foraminifera-bearing silty clay
140 and diatom-rich silty clay. The occurrence of calcareous nannofossils have been
141 reported in section 5W. We focused on the interval comprised between 1361A-2H-
142 6W and 1361A3H4W because in some discrete intervals planktonic foraminifers were
143 abundant (from Sample 2H-CC contains >90%).

144 **ODP Site 1101 (64.37°S, -70.26°E)** is located on the continental rise, west of the
145 Antarctic peninsula, 94 km offshore the shelf edge. The mean annual SST in this area
146 is around 0 °C (Levitus, 1994). Between 1101A9H1W100 and 1101A10H3W117 the
147 sediments consist of laminated clayey silts and foraminifera-bearing clayey silts
148 containing well preserved calcareous nannofossils.

149 **Cape Roberts Project Site 1 (CRP-1) (77.0°S, 163.8°E)** was drilled on the
150 continental shelf in the western Ross Sea near the Transantarctic Mountain belt.
151 Previous studies (Scherer et al., 2008) have focused on the unusual early Pleistocene
152 carbonate rich lithostratigraphic Unit 3.1 corresponding to the MIS31 interglacial.
153 This unique interval contains laminated sediments with abundant diatoms, planktic
154 foraminifera and thoracospherids (calcareous dinoflagellates - Scherer et al., 2008).

155

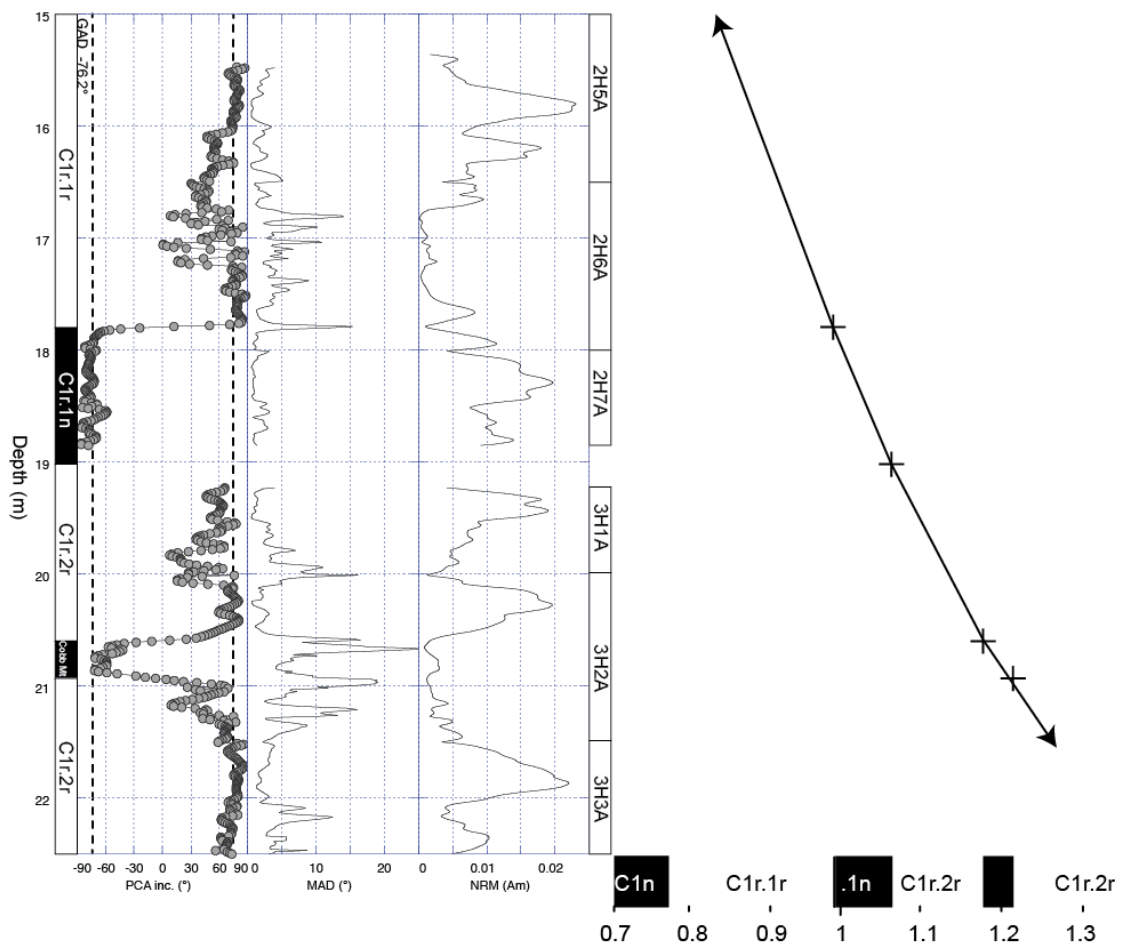
156 **2.2. Age models**

157 We developed a chronology for the IODP Site 1361 from new paleomagnetic
158 measurements performed on u-channel samples at the University of Otago
159 Paleomagnetic Research Facility (OPRF). Our LDI-derived SST data were then
160 converted from depth to the time by linear interpolation between geomagnetic
161 reversals. The alkenone-derived temperatures for the ODP Site 1094 and ODP Site
162 1101 were converted from depth to time using the chronologies from shipboard
163 paleomagnetic data.

164 **2.2.1. IODP Site 1361**

165 U-channel samples were collected at the IODP Core Repository at Texas A&M
166 University and shipped to the University of Otago's Paleomagnetic Research Facility
167 (OPRF). U-channel samples were collected from the IODP Site 1361 archive half
168 core to establish a more precise magnetostratigraphy. Archive half core sections were
169 demagnetized in fields of 20 mT during the expedition to build an initial age model
170 construction and core characterization. Magnetic moment measurements were made at
171 1 cm intervals using the 2G Enterprises pass-through superconducting rock
172 magnetometer which is housed in a 150 nT magnetically shielded room. U-channels
173 were demagnetized in alternating magnetic field increments of 5 mT between 20 mT
174 and 50 mT, then at 60 mT, 70 mT and finally at 100 mT. Demagnetization data were
175 analysed on orthogonal vector component plots using the PuffinPlot software
176 (Lurcock and Wilson, 2012). Polarity was determined from principal component
177 analyses (PCA, Kirschvink, 1980) of data between the 20 mT and c. 50 mT
178 increments which we interpret is the primary depositional remanent magnetizations
179 (DRM). PCAs were typically anchored to the origin with the exception of thin
180 intervals which were offset from the origin where a high coercivity magnetisation
181 prevents complete demagnetisation. Rock magnetic data of these intervals did not

182 reveal unusual changes in magnetic mineralogy therefore we suggest that strongly
 183 magnetised, high coercivity drop stones may be responsible for the offset origins.
 184 Our new magnetostratigraphy does not change the correlation with the GPTS (Tauxe
 185 et al., 2012) but does result in more precise placement of reversal boundaries (Fig. 2).
 186 Our new magnetostratigraphy places the C1r.1r-C1r.1n (988 ka) boundary (at 17.77m
 187 $\pm 0.03m$, the C1r.1n-C1r.2r (1072 ka) at $19.02m \pm 0.17m$, the top and base of the Cobb
 188 Mt (1178–1215 ka, Channell, 2017) at $20.54m \pm 0.06m$ and $20.92m \pm 0.04m$
 189 respectively (Fig. 2). The C1n-C1r.1r (781 ka) transition spans a core break between
 190 8.1 m and 14.34 m (11.22 m mid-point) where the lower contact on core 2H is
 191 affected by coring disturbance and the C1r.2r-C2n (1778 ka) transition occurs at
 192 $32.33m \pm 0.025m$ (Tauxe et al., 2012).



193
 194 **Figure 2:** IODP 1361 age model developed in this study.

195 **2.2.2. ODP Site 1094**

196 The age model of ODP Site 1094A and 1094D was constructed from shipboard
197 paleomagnetic data where the 20 mT demagnetisation step was taken as the
198 characteristic remanent magnetisation as confirmed from complete laboratory
199 demagnetisation of u-channel samples (Channell and Stoner, 2002). The C1n-C1r.1r
200 boundary (781 ka) occurs at 99.61m, the C1r.1r-C1r.1n (988 ka) boundary at 123m
201 ± 2.22 m, the C1r.1n-C1r.2r (1072 ka) at 129m ± 0.38 m, and the base of the Cobb Mt
202 (1178 ka) at 149.6 m.

203 **2.2.3. ODP Site 1101**

204 We adopted the age model (Acton et al., 2012) based on shipboard analyses and
205 shore-based laboratory analyses from u-channel samples. The C1n-C1r.1r (781 ka)
206 transition occurs at 55.05 m ± 0.01 m, the C1r.1r-C1r.1n (988 ka) at 71.15 m ± 0.19 m,
207 the C1r.1n-C1r.2r (1072 ka) at 76.12m ± 0.17 m, and the top and base of the Cobb Mt
208 (1178–1215 ka, Acton et al., 2012) at 81.61m ± 0.65 m and 82.53m ± 0.01 m
209 respectively.

210 **2.3. Temperature reconstructions**

211 To reconstruct the latitudinal SST gradient during the MIS31 super-interglacial along
212 a transect from Sub-Antarctic to Antarctic water masses of the Southern Ocean, we
213 used two complementary organic geochemical proxies for SSTs: the alkenone
214 unsaturation ratio (Prahl et al., 1988 - $U^{k'_{37}}$) and the Long Chain Diol Index (Rampen
215 et al., 2012 - LDI). The use of alkenone thermometry was motivated by the unusual
216 presence of surface-dwelling calcareous nannofossils in 1 Ma-old sediments at
217 extreme southern latitudes around the Antarctic (Scherer et al., 2008; Villa et al.,
218 2005; 2007; 2012). Specifically, representatives belonging to the alkenone (long-chain
219 unsaturated ketones) producer Noelarhabdaceae family (*Emiliana*, *Gephyrocapsa*,
220 *Reticulofenestra*, *Pseudoemiliana*, *Dictyococcites* sp. – Marlowe et al., 1990; Beltran
221 et al., 2007; 2011) were reported at drilling sites located south of the Polar Front (Fig.
222 1).

223 Although alkenones were detected at Site 1094 (concentrations between 2 and 25
224 ng/g), the concentrations were below detection limits at the Antarctic margin IODP
225 Site U1361. Thus, we chose to use the long chain Diol Index (LDI) to reconstruct the
226 SSTs at the Antarctic margin. This newly developed proxy for paleotemperatures (see
227 Lopes dos Santos et al., 2013; de Bar et al., 2018), is based on the relative abundances
228 of lipids with chain length of 28 and 30 carbon atoms containing two alcohol groups
229 (long chain alkyl 1,13- and 1,15-diols; $C_{28}1,13$ -diol, $C_{30}1,13$ -diol and $C_{30}1,15$ -diol). In
230 the marine environment, these long chain diols have been found in diatoms and
231 marine algae (Rampen et al., 2012). Although the source of the long chain diols is still
232 under discussion, these molecules have existed at least since the Mesozoic era and are
233 ubiquitous in the world ocean from the equator down to very high latitudes (74°S).
234 The LDI index correlates best with temperature for modern water depths of 0 to 30 m

235 during the warmest months of the year.

236 Alkenone unsaturation ratios were measured on samples from ODP Site 1094D-11H-
237 1W to 1094D-13H-3W (between 115.9 and 137.8 m) at a resolution of one sample
238 every 30 cm (temporal resolution of 2.2 kyrs) and from ODP Site 1101A-9H-2W to
239 1101A-10H-4W (between 68.7 and 80.4 m) at a resolution of one sample every 45 cm
240 (temporal resolution of 7.3 kyrs). For IODP Site 1361, we measured the Long chain
241 diol index (LDI) on 12 samples from 1361A-2H-6W to 1361A-3H-3W (between 17.7
242 and 21.7 m) at a resolution of one sample every 35 cm (temporal resolution of 23
243 kyrs). We found Long chain diols in 10 samples. For both methods, sediment samples
244 (6 to 9 g) were freeze-dried, homogenized, and extracted using an automated solvent
245 extractor a Dionex 300 Accelerated Solvent Extractor with dichloromethane and
246 methanol (9:1, v/v) at 1500 psi and 100°C to obtain a total lipid extract.

247 Alkenones were isolated by silica gel chromatography using solvents of increasing
248 polarity following the procedure of Sicre et al. (2001). The fractions containing
249 alkenones were concentrated, transferred to clean glass-vials and evaporated under
250 nitrogen. Gas chromatographic analyses were performed on a Hewlett Packard 6890
251 gas chromatograph with flame ionization detector using a fused silica capillary
252 column (Rxi-1ms, 50 m long, 0.32 mm internal diameter, 0.25 µm film thickness).
253 Helium was used as a carrier gas. We have converted the alkenone index into
254 temperature using the Prahl et al. (1988) calibration (see Supplementary Information
255 Section 1.3).

256 Long chain diols were isolated by silica gel chromatography following the procedure
257 of Rampen et al. (2012). Compound identification of the long chain diols was
258 conducted using an Agilent 6890N gas chromatograph (GC) equipped with an Agilent
259 5975B mass selective (MS) detector in selective ion monitoring mode operated at 100

260 eV (EI Source). They were separated using a fused silica capillary column (ZB-5MS,
261 30 m long, 0.32 mm internal diameter, 0.25 μ m film thickness). The relative
262 abundance of long chain diols was measured using single ion monitoring of m/z 299,
263 313, 327, and 411.

264 **2.4. Ice sheet/shelves simulations**

265 **2.4.1. General Circulation Model (GCM) set-up and simulations**

266 Our RegCM3 experiments were initialised from GENESIS3 GCM simulations
267 previously run at UMass Amherst (DeConto, et al., 2012; DeConto and Kowalewski
268 unpublished). The GCM was run for 30 years to equilibrate, and then for a further 11
269 years to provide input (initial / boundary conditions) for the RegCM3 regional climate
270 model. To provide appropriate input for RegCM3, GENESIS3 was tuned for orbital
271 forcing, CO₂ levels, and ice sheet configuration representative of MIS31. Ice sheet
272 configurations for Neogene interglacials were provided from model results from the
273 Pollard and DeConto (2009) ice sheet model (forced w/ GENESIS3 as well).

274

275 **2.4.2. Regional Climate Model (RCM) set-up and simulations**

276 The 11 years -period described above allows for a one-year spinup for the RCM,
277 followed by a ten years model evaluation period. GCM derived boundary conditions
278 were continuously updated every 6 hours of RCM simulation, with fields being
279 interpolated by the 'RegCM3_Polar' package onto the new model grid. No proxy data
280 were used during this phase of climate model set up, but RegCM3 simulations were
281 validated against (1) GCM performance conducted with GENESIS3, and (2) RegCM3
282 forced with NCEP/NCAR reanalysis. Reanalysis driven runs are for the 1990-1999
283 period using observed SST forcing while the Genesis-driven runs employ SST forcing
284 from a slab ocean model.

285

286 **2.4.3 Ice Sheet Model (ISM) set-up and simulations**

287 For our main experiments, we used climate model outputs from GCM experiments in
288 which the ice sheet configuration was similar to present, since our ice sheet model
289 experiments also start from a modern configuration. This avoids the circularity of
290 simulation WAIS collapse by imposing a climatology that has been produced under
291 conditions of a collapsed WAIS. We ran simulations with two different atmospheric
292 greenhouse gas concentrations (280 and 400 ppm CO₂) to quantify the sensitivity of
293 our results to different background climate states. The climatologies from the regional
294 climate model RegCM3_Polar simulations (RegCM3 adapted to Antarctica; Fig. S3)
295 were imposed at the beginning of each run and the ice sheet was allowed to freely
296 evolve to the new boundary conditions. Ice sheet experiments used the Parallel Ice
297 Sheet Model (PISM) version 0.6, whose dynamical core combines velocity solutions
298 of the shallow shelf and shallow ice approximations for the entire domain (see
299 Supplementary Information Section 2). We ran duplicate experiments both with, and
300 without, the sub-ice shelf grounding line melt interpolation scheme turned on. Once
301 afloat, we imposed two calving mechanisms – one based on horizontal strain rates
302 (Levermann et al., 2012), and another in which a simple minimum thickness criterion
303 (200m) must be satisfied. We ran experiments for two climate scenarios, reflecting
304 atmospheric CO₂ concentrations of 280 and 400 ppm, and for each of these, ran
305 duplicates with each of the two grounding-line parameterizations. All simulations
306 were run for 10000 years and were initialized from a thermally and dynamically
307 stable configuration used previously (Golledge, et al., 2015; 2017a; 2017b) which
308 closely resembles the present-day ice sheet (Fig. S4 panel a). Ice sheet configurations
309 after 5000 years for each of the experiments are shown in Figure S4.

310

311 **3. Results**

312 **3.4. Early Pleistocene Southern Ocean sea surface temperatures**

313 Our SST records were generated using the alkenone unsaturation proxy ($U_{37}^{k'}$ - Prahl
314 et al., 1988) in the ice-free Antarctic zone (ODP Site 1094) and in the polar zone ODP
315 Site 1101 and the Long Chain Diol Index (LDI – Rampen et al., 2012) at the IODP
316 Site 1361 (Fig. 3). Two temperature proxies were used because although $U_{37}^{k'}$ is well-
317 established, the calcareous nannofossils that produce alkenones (Marlowe et al., 1990;
318 Beltran et al., 2007; 2011) are not always detectable in the coldest high latitude ocean
319 regions. In contrast, the long chain diols used in the LDI index offer a good
320 alternative as they are ubiquitous in the ocean, including at high latitudes (74°S) and
321 LDI- and $U_{37}^{k'}$ -derived SSTs record similar water depths and seasons (Rampen et al.,
322 2012).

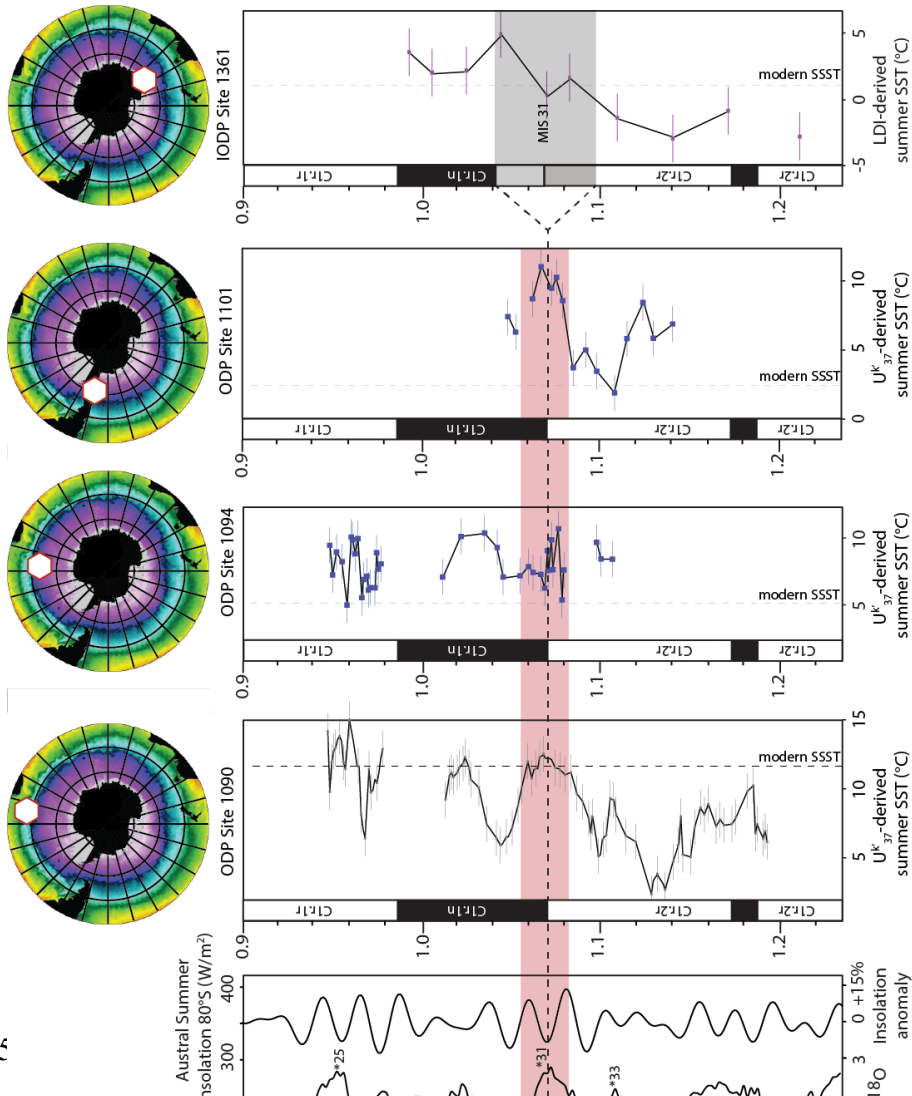
323 At ODP Site 1094, our record for the interval between 1.11 Ma and 0.95 Ma shows
324 that the summer SSTs were on average 8.6 °C (between 5.3 °C and 10.8 °C), which is
325 3 °C warmer than today (Southern Hemisphere Summer months - Levitus, 1994; Fig.
326 3). This indicates sustained long-term elevated ocean surface temperatures in this
327 sector of the Southern Ocean (Fig. 3A). At about 1.076 Ma, we identify a 10.8 ± 1 °C
328 temperature peak, i.e. 6 °C warmer than today, which we interpret as the MIS31
329 temperature optimum. Our data shows that during MIS31, SSTs closely track changes
330 in Southern Hemisphere summer insolation with the warmest conditions reached
331 during the highest Southern Hemisphere summer insolation maxima at 1.08 Ma
332 (Laskar et al., 2004). We also observe large magnitude rapid temperature variations,
333 e.g. 5.5 °C warming between 1082.69 ka and 1082.05 ka.

334 At ODP Site 1101 and IODP Site 1361, the age model and sampling resolution is

335 lower than at ODP Site 1094. However, at both sites, our data show that MIS31
336 corresponds to a summer SST increase of more than 5 °C compared to today (Levitus,
337 1994; Fig. 3A). The average SST through MIS31 at ODP Site 1101 is ~ 9.5 °C,
338 peaking at about ~11 °C (1.069 Ma), which, is 7 °C warmer than the modern
339 conditions. At IODP Site 1361, (the age model is less well constrained) the average
340 surface water temperature during MIS31 was between 0 and 1°C, reaching a
341 maximum value of 5 °C (6 °C warmer than today - Levitus, 1994). This result is in
342 agreement with the Mg/Ca measurements on planktic foraminifera (Dunbar, 2012)
343 and diatom based MIS31 temperature estimates (Scherer et al., 2008 and Scherer
344 comm. pers.) from the Unit 3.1 at CRP-1B who suggests summer SSTs in the Ross
345 Sea were warmer than today by 5-9°C.

346 Our paleotemperature reconstructions based on organic biomarkers, show that the
347 surface ocean, south of 50 °S, warmed on average by 5 °C (± 1 to 2 °C) (Fig. 3 and 4),
348 providing the first precise constraints on the maximum surface ocean temperature
349 during MIS31.

350



35

352 **Figure 3:** Alkenone (U^{k}_{37}) and LDI-based SSTs throughout MIS31.

353 We show time series of deep-sea benthic $\delta^{18}O$ record (‰) (Lisiecki and Raymo,
 354 2005), austral summer insolation at 80°S and insolation anomaly relative to present
 355 (%increase relative to present; Laskar et al., 2004). Some of the MIS peaks are
 356 indicated with the symbol *. MIS31 interglacial duration is highlighted in red. The
 357 black dotted horizontal line shows the age of MIS31's peak (*31). Due to the
 358 uncertainties on the age model at Site 1361, MIS31 peak is highlighted in grey. Each
 359 map indicates the site locations. The error bars correspond to the error of each
 360 temperature proxy.

361 Summer southern hemisphere summer season which is here the average temperature
 362 for the months DJF

363

364 Our temperature record from ODP Site 1094 also indicate strong high-frequency

365 variability within this warm period.

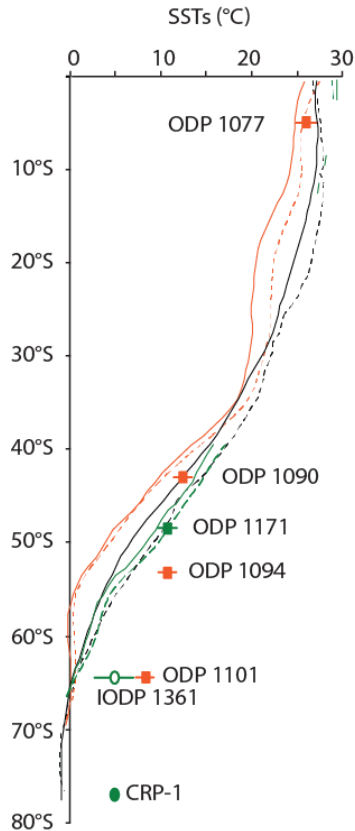


Figure 4: Comparison of the latitudinal sea surface temperature gradients during MIS31 warmth with modern conditions. MIS31 temperature data for Site ODP 1077 is from Schefus et al. (2004), for ODP Site 1171 from Beltran et al. (2016) and for CRP-1 from Dunbar (2012). The orange and green bars represent respectively the error bars for alkenone and LDI measurements. Black dashed line: Modern Annual Southern Ocean SST - Green dashed line: Modern Annual Atlantic Ocean SST - Orange dashed line: Modern Annual Pacific Ocean SST - Green line: Modern Austral Summer Southern Ocean SST - Orange line: Modern Austral Summer Atlantic Ocean SST - Black line: Modern Austral Summer Pacific Ocean SST.

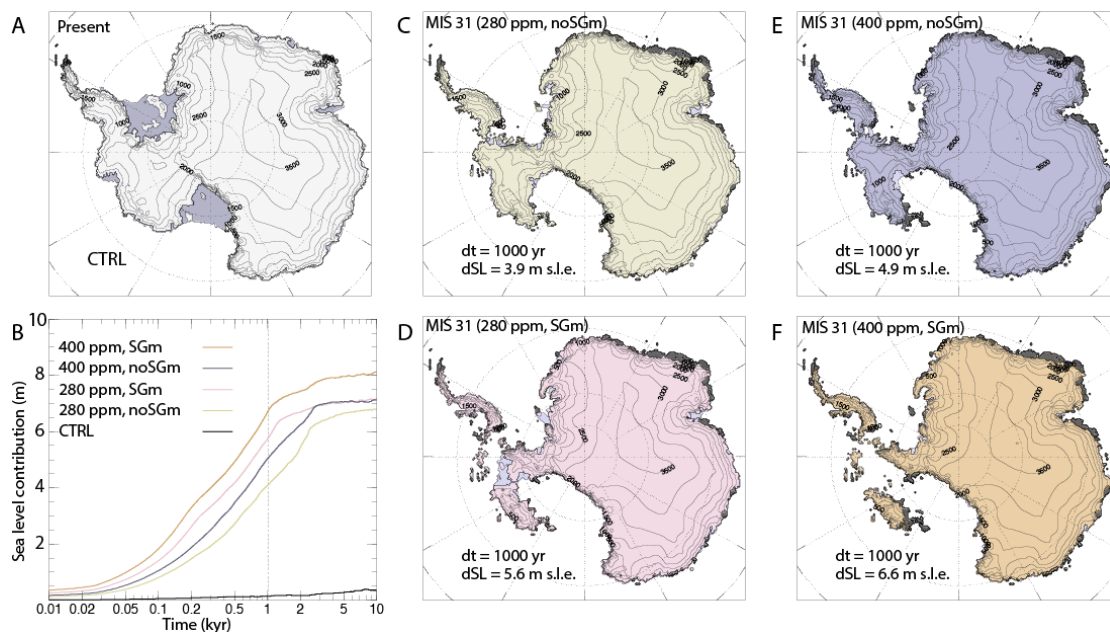
367

368 When comparing our high latitude records with the lower latitude alkenone-derived
 369 SSTs at the subantarctic ODP Site 1090 and the equatorial ODP Site 1077 (Schefus et
 370 al., 2004 - Fig. 4), we observe a significant polar amplification during MIS31,
 371 resulting in a reduced latitudinal SST gradient (Scherer et al., 2008; Martinez-Garcia
 372 et al., 2010; Dunbar, 2012; Beltran et al., 2016) of ~ -0.3 °C/°latitude (Fig. 4).

373 3.2. Modelling the AIS during MIS 31

374 To investigate the degree of ice sheet retreat in relation to our SST records we
 375 conducted a suite of numerical simulations that use climate fields from a regional
 376 climate model (DeConto et al., 2012 - RCM, Supplementary Section 2). Our proxy
 377 data were critical to the simulations because they were the quality assurance that the
 378 simulations were capturing the correct magnitude of warming. We used the Parallel
 379 Ice Sheet-Model, which is an open source three-dimensional thermodynamic coupled
 380 ice sheet/ice-shelf model. Our RCM simulation represents a time slice reconstruction

381 rather than a transient prediction, but considers two atmospheric CO₂ concentration
 382 scenarios - 280ppm and 400ppm - so that the sensitivity of the ice sheet to
 383 atmospheric composition can be assessed (Supplementary Section 2.2). Using these
 384 two scenarios as inputs, we also then combined each one with either a conservative
 385 grounding line scheme (no SGm) or a more dynamic grounding line (SGm). Because
 386 our climate forcing is not transient, we were interested primarily in seeing whether
 387 significant ice sheet impacts could occur within relatively short time periods
 388 (centuries to a millennium) of warmth, since our proxy data indicate that such
 389 temperatures are not likely to have been sustained for the entire interglacial.
 390 In all simulations, the WAIS melted within a few centuries (Fig. 5B, C, D, E and F)
 391 whereas the EAIS responded more slowly (Supplementary Section 2.3). The
 392 simulations revealed that regardless of the grounding line scenario and CO₂ forcing,
 393 no significant difference in the spatial extent of ice loss is observed after 5kyr (Fig.
 394 S4).



395

396 **Figure 5:** Snapshots of ice distribution for ice sheet experiment runs of 1000 years
 397 (dt). We present the results using a conservative grounding line behaviour (no SGm,
 398 C. and E.), a dynamic grounding line (SGm, D. and F.), each with two extreme CO₂
 399 scenarios (280 ppm and 400 ppm). A. represents the modern ice distribution and B. is

400 the sea level contribution from ice melting through time. dSL is the sea level increase
401 expressed in s.l.e (sea level equivalent).
402 Our results indicate that 2-m of sea-level rise from ice loss occur within a few
403 centuries (Fig. 5B) and that persistent elevated SSTs would cause complete collapse
404 of the WAIS, with additional thinning around the margins of the EAIS (i.e. the Wilkes
405 Land basin, Weddell Sea, Aurora Basin).

406 **4. Discussion and conclusions**

407 Our temperature reconstructions indicate proximal ice margin summer SSTs of 4-5 °C
408 in East and 9°C in West Antarctica (5-7 °C warmer than today) during MIS31, in
409 agreement with Scherer et al (2008), Villa et al. (2008, 2012). Those temperatures are
410 not compatible with the presence of a significant West Antarctic ice sheet (Joughin
411 and Alley, 2011; Liu et al., 2015 - Fig. 5).

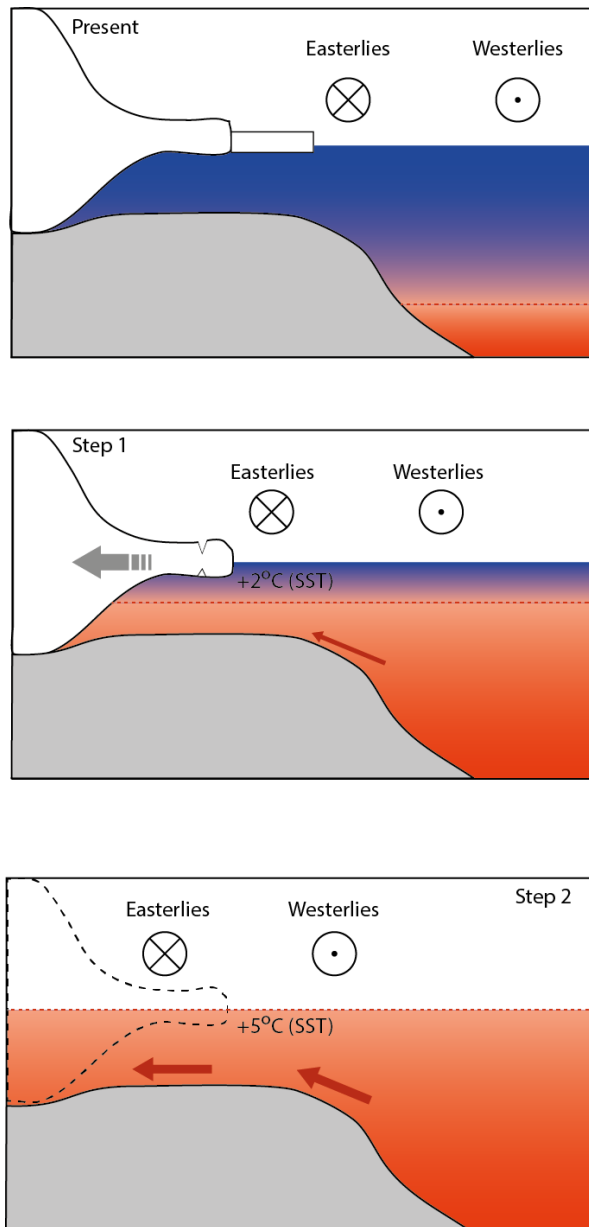
412 The causes of the MIS31 extreme warmth under pre-industrial atmospheric CO₂
413 conditions are still unclear. MIS31 interglacial coincides with a time of high obliquity
414 and eccentricity, which led to strong high latitude austral summer insolation
415 anomalies. How much of the ocean warming observed in our records can be attributed
416 to the high summer insolation? Simulations of previous interglacials (Yin and Berger,
417 2012) showed that reduction of summer sea ice increases the sensitivity of southern
418 high latitudes to solar radiation (Yin and Berger, 2012) resulting in warming of
419 surface waters. According to Yin and Berger (2012), during interglacials with similar
420 obliquity values as MIS31, increased insolation contributed to a maximum mean
421 annual warming of only 0.5 °C of the high southern hemisphere latitudes. As a
422 consequence, high insolation alone cannot account for the extreme ocean
423 temperatures we observe around Antarctica during MIS31. We argue here that the
424 exceptional warmth is the consequence of an increasingly stratified ocean due to
425 freshwater input from ice loss (Golledge et al., 2014) rather than a trigger for ice loss.

426 Reduced surface ocean salinity is supported by the diatom assemblage composition at
427 CRP-1 with increased abundance of *Chaetoceros* spp. correlated to the insolation
428 maximum at 1.08 Ma (Scherer et al., 2008). The calcareous nannofossil assemblage
429 composition at ODP Site 1094 also support this hypothesis with a clear increase of
430 *Helicosphaera carteri* and *Syracosphaera* spp. (Flores and Sierro, 2007; Scherer et
431 al., 2008).

432 We suggest ice margin retreat was caused by circulation changes in Antarctic coastal
433 waters, as a result of changes in wind patterns (Gille et al., 2016). A poleward shift of
434 the Southern Hemisphere Westerly Winds (SHWW) is seen in the decrease of aeolian
435 dust in the Subantarctic Zone (Martinez-Garcia et al., 2010 - ODP Site 1090), raising
436 the possibility that a wind-induced subsurface warming at the Antarctic margin took
437 place during MIS31. In the modern ocean, cold Antarctic shelf waters are separated
438 from warmer Circumpolar Deep Water (CDW) further offshore by the Antarctic
439 Slope Front (ASF – Jacobs, 1991), a well-marked density front that is primarily
440 maintained through coastal downwelling induced by the prevailing coastal easterly
441 winds (Sverdrup, 1953). Ocean model simulations have demonstrated that a poleward
442 shift of the SHWW reduces the extent and strength of the coastal easterlies and can
443 produce strikethrough advective warming of subsurface waters at the Antarctic
444 margin, via a weakening of the ASF that allows CDW to intrude onto the shelf
445 (Schmidtke et al., 2014; Gille et al., 2016). Heat from the CDW would subsequently
446 trigger basal melting of the ice shelves (Dinniman et al., 2016). Additional
447 observational (Schmidtke et al., 2014; Dutrieux et al., 2014) and modelling evidence
448 (Thoma et al., 2008; Dinniman et al., 2012) also implicates the SHWW in regulating
449 poleward transport of warm CDW onto the Antarctic continental shelf in the modern-
450 day ocean, and it therefore seems plausible that this onshore CDW transport was

451 enhanced by poleward-intensified winds during MIS31 (Martinez-Garcia et al., 2010).
452 Precise dynamics underlying such wind-driven CDW inflow however remain an area
453 of active research (Gille et al., 2016).

454 We propose that the exceptional surface ocean warming during the MIS31 super-
455 interglacial occurred in two steps (Fig. 6). [Step 1] The unusual MIS31 orbital
456 configuration resulted in an extended period (~ 2000 years) of warm and long-lasting
457 summer seasons that caused mild warming of Southern Ocean surface waters and the
458 reduction of sea ice development or survival. Concomitantly, the prevailing Southern
459 Hemisphere Westerly Winds and the easterly coastal winds migrated to the south
460 (Fig. 6). The poleward shifted winds warmed the coastal waters, thus bringing heat
461 southwards thereby initiating basal melting of ice shelves and the retreat of the
462 marine-based grounding lines. Our numerical ice sheet model demonstrates that a 0.5
463 °C ocean warming at the ice margin for c. 200 years is sufficient to cause ice retreat
464 and that most of it can occur in less than 2000 years. [Step 2] The surface warming in
465 the coastal regions (9.5 °C at ODP Site 1101 and 5 °C at IODP Site 1361) could only
466 take place after complete loss of the WAIS (Fig. 6) and would likely be amplified by
467 ocean stratification feedbacks.



468

469 **Figure 6:** Schematic evolution of ice retreat during MIS31 in response to poleward
 470 wind shift and subsequent ocean warming. Step 1 shows the increased advection of
 471 warmer deep waters (red arrow, see Pedro et al., 2016), Step 2 represents the run over
 472 ice melting at the maximum of MIS31 warmth.
 473

474 Our new data from the Antarctic sedimentary records demonstrate that a moderate
 475 surface ocean warming is sufficient to initiate a major West Antarctic Ice Sheet
 476 retreat. We also show that even after 10000 years of persistent elevated SSTs, marine
 477 basins of East Antarctica remain glaciated, with the majority of the EAIS sea-level
 478 contribution coming from coastal thinning around Dronning Maud Land. That the

479 marine basins of EAIS were not destabilized at MIS31 suggests that atmospheric
480 warming (as simulated in the RCM) at that time was insufficient to bring about
481 significant surface melting, thinning, and retreat from coastal pinning points
482 (Golledge et al., 2017b).

483 Future research should focus on further documenting high latitude Southern Ocean
484 SSTs using molecular thermometry to identify critical ocean temperatures that led to
485 major Antarctic ice retreats in the past under varying atmospheric CO₂ and orbital
486 conditions.

487

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502

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711 **Supplementary Information for:**

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713 **Southern Ocean temperature records and ice-sheet models demonstrate rapid**
714 **Antarctic ice sheet retreat under low atmospheric CO₂ during Marine Isotope**
715 **Stage 31.**
716

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Supplementary Information

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747 Section 1: Sea surface temperature proxies

748 Climate simulations require unambiguous, continuous, and well-dated SST records to
749 determine changes in latitudinal heat transport. Despite the emergence of new and
750 well calibrated paleotemperature proxies based on diatom morphometry (e.g. Allen,
751 2014; Kloster et al., 2018; Shukla and Romero, 2018), the few available high latitude
752 Southern Ocean SST reconstructions for MIS31 were either derived from 1) proxies
753 based on calcareous nannofossil and planktonic foraminifera assemblage
754 compositions that tend to be less reliable at high latitudes where the species diversity
755 is low 2) diatom assemblages which composition reflect not only temperatures but
756 also environmental parameters such as sea ice, nutrients and light etc... or 3)
757 uncorrelated temperature proxies possibly reflecting temperatures of different water
758 masses and/or seasons, which adds significant uncertainties to climate reconstructions.
759 The lack of accurate and consistent SSTs for the Sub-Antarctic and Antarctic zones of
760 the Southern Ocean during MIS31 greatly restricts our ability to precisely quantify the
761 global latitudinal ocean heat transfer and subsequent sub-ice ocean heat flux,
762 impacting our ability to simulate the AIS response to ocean temperature changes. To
763 reconstruct the most continuous latitudinal record of the SSTs during the MIS31
764 super-interglacial along a transect from Sub-Antarctic to Antarctic water bodies in the
765 Atlantic sector of the Southern Ocean, we used two complementary organic
766 geochemical proxies for SSTs: the alkenone unsaturation ratio (U_{37}^k , Prahl et al.,
767 1988) and the Long Chain Diol Index (LDI, Rampen et al., 2012).

768 1.1. Alkenone thermometry

769 The C_{37} alkenone unsaturation index (U_{37}^k ; Eq. 1) is a robust proxy for SST,
770 providing surface ocean temperature reconstructions with a precision of +/-1.2 °C
771 (Sikes and Volkman, 1993; Muller et al., 1998; Conte et al., 2006).

772
$$U_{37}^{k'} = \frac{[C_{37:2}]}{[C_{37:2}] + [C_{37:3}]} \quad \text{Eq. 1}$$

773

774 Global core-top studies demonstrate that $U_{37}^{k'}$ correlates best with mean annual
 775 temperatures at 0–10 m water depth range (Conte et al., 1993) and there is little
 776 evidence that the SST signal is affected by other environmental parameters such as
 777 productivity or salinity.

778 Previous studies (Sikes and Volkman, 1993; Conte et al., 2006;) showed that south of
 779 40°S in the Pacific and Atlantic sectors of the Southern Ocean, alkenone production is
 780 limited to late spring-summer. As both ODP Sites 1090 and 1094 chosen in our study
 781 were drilled south of the subtropical front, the $U_{37}^{k'}$ -derived SST reconstructions are
 782 likely to reflect the mixed-layer conditions during the warmest months of the year.

783 1.2. Long chain Diol Index thermometry for IODP Site UI361

784 The Long-chain diol index (LDI; Eq. 2) is defined as:

785
$$\text{LDI} = \frac{[C_{301,15\text{-diol}}]}{[C_{281,13\text{-diol}}] + [C_{301,13\text{-diol}}] + [C_{301,15\text{-diol}}]} \quad \text{Eq. 2}$$

786

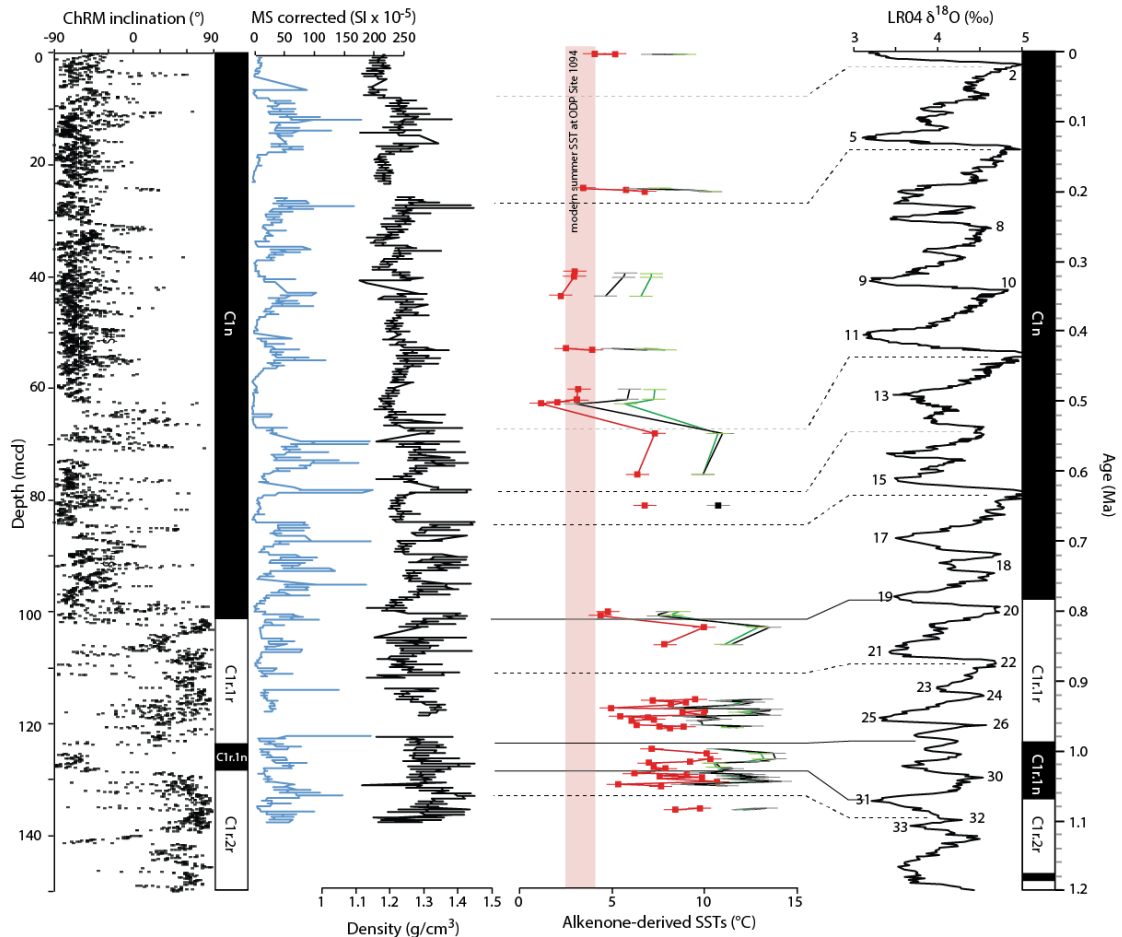
787 Core top studies indicate that the LDI index is linearly correlated to SSTs (+/-2°C) in
 788 a range of temperatures from -3°C to 27°C (Eq. 3) with salinity and nutrient
 789 conditions having little to no influence (Rampen et al., 2002).

790
$$\text{LDI} = 0.033 \times \text{SST} + 0.095 \quad \text{Eq. 3}$$

791 1.3. Alkenone Calibration

792 Most alkenone-derived SST reconstructions use the calibration of Prahl et al.(1988)
 793 based on the empirical linear relationship between $U_{37}^{k'}$ and SST obtained from
 794 laboratory cultures. However, several studies show that the $U_{37}^{k'}$ measured in
 795 particulate matter of surface waters disagree with this calibration (Prahl et al.,1988).
 796 These observations led to the development of alternate calibrations either global such

797 as the polynomial U_{37}^k -SST relationship of Conte et al. (2006) or region-specific
 798 (Muller et al., 1998; Sikes and Volkman, 1993).
 799 To select the most accurate U_{37}^k calibration, we have extended the ODP Site 1094
 800 sampling (Fig. S1) to cover the long term evolution of SSTs until today.



801
 802 **Fig. S1.** Long term evolution of the alkenone-derived sea surface temperatures at ODP Site 1094 using
 803 three different calibrations : Prahl et al. (1988) in red; Sikes and Volkman (1993) in green; Conte et al.
 804 (2006) in black. Our additional low resolution dataset shows that the Prahl et al.(1988) calibration
 805 leads to SSTs that are similar to the modern ones at the same location.
 806

807 When comparing the core top SSTs with the Levitus monthly modern SSTs it appears
 808 that the Prahl et al. (1988) calibration leads to values that are consistent with the
 809 modern February SSTs at the same location (Fig. S1). We have thus converted the
 810 alkenone index into temperature using the Prahl et al. (1988) calibration.

811 Section 2: Ice sheet/shelves simulations

812 2.1. Global Climate Modelling.

813

814 The relationship between air and ocean temperatures under different CO₂ conditions

815 evolves through time (Rugenstein et al., 2016), such that in Antarctica, short-term (<

816 100 year) sea-surface temperature anomalies are considerably less than the

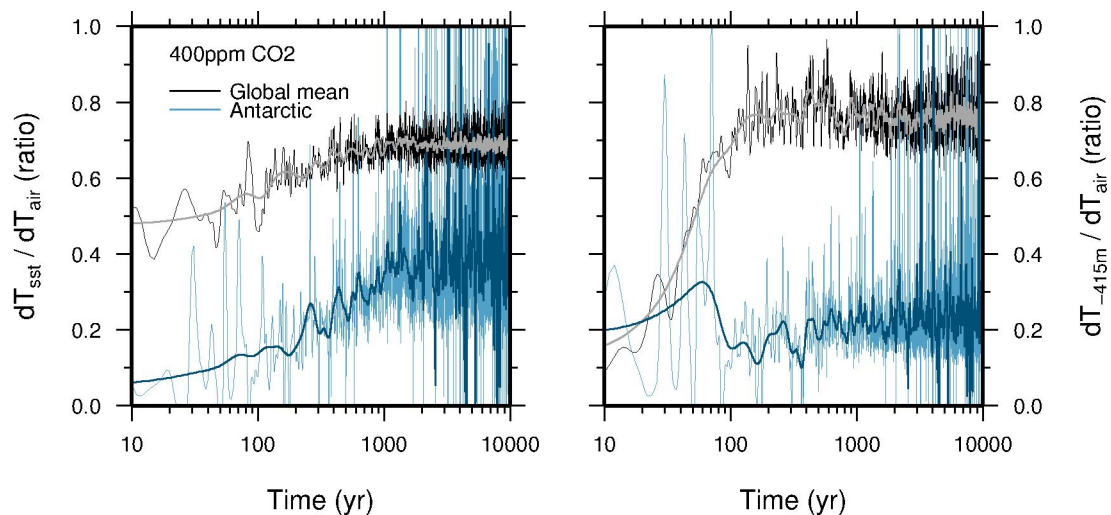
817 equilibrium state (Fig. S2 below). When attempting to infer an atmospheric warming

818 from our oceanic proxy data, therefore, we conservatively assume that the ocean /

819 atmosphere warming has achieved balance (taking c. 1000 years), and thus mean

820 annual sea temperature anomalies are likely to be c. 2.5 times the inferred mean

821 annual air temperature anomaly (SST to air ratio of 0.4, Fig. S2).



822

823 **Fig. S2.** Evolution of the relationship between mean annual surface air temperature anomaly and that

824 of the sea surface (left) and ocean at 415 m depth (right) for Antarctica, compared to the global mean

825 (Rugenstein et al., 2016). An equilibrium SST to air ratio of 0.4 implies a 2.5 times greater air

826 temperature anomaly than that felt by the sea surface.

827

828 2.2. Regional Climate Modelling

829 Paleoclimate simulations were run using RegCM3_Polar. The regional climate model,

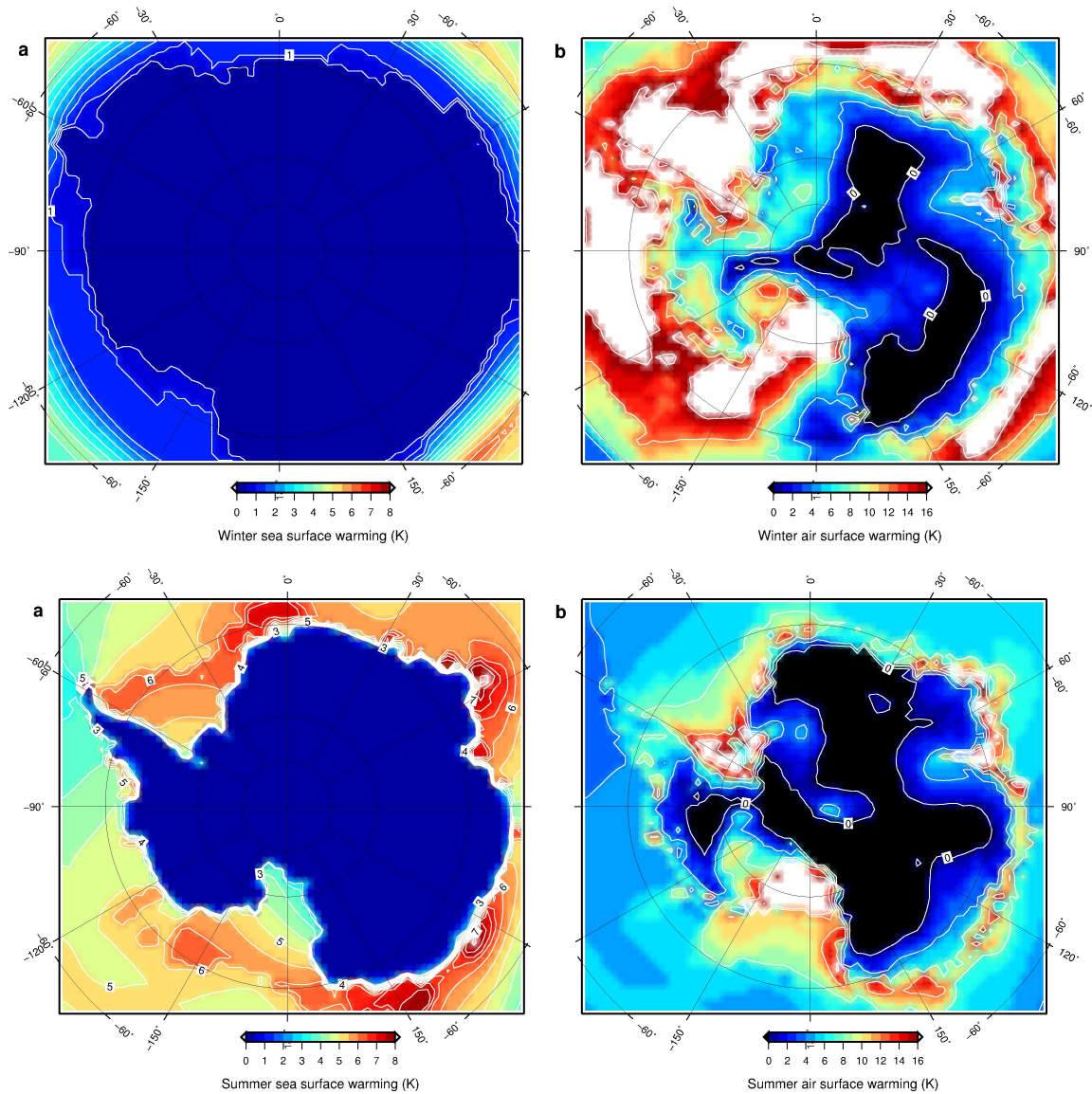
830 RegCM3, was adapted to Antarctica (RegCM3_Polar) centered on the South Pole for

831 Antarctic reconstructions. Boundary conditions for the regional model were taken

832 from the library of existing GCM Pliocene simulations (GENESIS v. 3.0.; Pollard and

833 DeConto, 2009). Simulations varied greenhouse gas, orbital, sea ice and SST

834 conditions; changes in environmental conditions were prescribed in both the RCM
835 and GCM. The boundary forcing from GENESIS GCM was interpolated horizontally
836 and vertically to the RegCM3_Polar grid and topography and was applied at 6-hour
837 intervals for both 40km and 80km horizontal resolution regional climate simulations.
838 Regional climate simulations were run using existing ice configuration and a MIS31
839 modeled ice sheet configuration imported from the Pollard and DeConto ice sheet /
840 shelf model output (Pollard and DeConto, 2009).
841



842

843

844 *Fig. S3. Seasonal climatologies for Antarctica during MIS31, as simulated by the Regional Climate*
845 *Model, as anomalies from present-day.*

846

847 2.3. Ice Sheet Modelling

848 For our ice sheet experiments we use the climatologies from the RCM simulations

849 (Fig. S3) as anomalies to present-day air temperature and precipitation fields

850 (Lenaerts et al., 2012), in the same manner as employed previously (Golledge et al.,

851 2017a). We use monthly climate fields in order to capture the high-latitude changes in

852 seasonality brought about by the different orbital configuration of MIS31, compared

853 to present. A positive-degree day model translates temperatures above freezing into

854 surface melt, of which a proportion (60%) is retained in the snowpack through

855 percolation refreezing processes. This value is difficult to constrain precisely (Kuipers

856 Munneke et al., 2011) but here we use a constant value for both our control and our

857 perturbation experiments, thus differences in the results are not likely to arise from

858 errors in this parameterisation.

859 We use the Parallel Ice Sheet Model (PISM) version 0.6. Fast flow of grounded ice is

860 facilitated by the plastic failure of subglacial sediments that arises as a consequence of

861 an initially prescribed till friction angle, representing sediment strength, and its

862 saturation state, which evolves dynamically in concert with the thermal properties of

863 the basal layer of overlying ice. Thus where ice reaches sufficient thickness to allow

864 basal melting to occur, meltwater begins to weaken the substrate until driving stresses

865 exceed the evolving cohesiveness of the till. Failure of the substrate and the

866 concomitant acceleration of overlying ice obeys a pseudo-plastic law (Schoof, 2006;

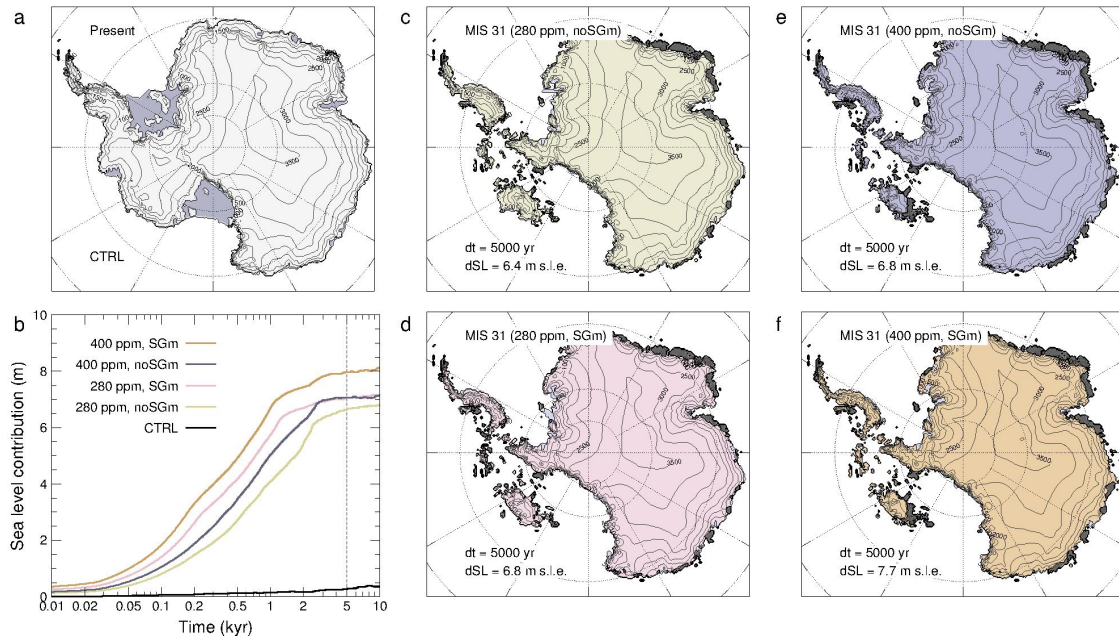
867 Bueler and Brown, 2009), such that small increments of stress above the shear

868 strength of the substrate lead to an increasing velocity response that ultimately thins

869 the ice, reducing its driving stress and leading to deceleration of the ice. The cyclic

870 behaviour of ice streams that occurs as a consequence of this mechanism is described
871 in more detail elsewhere (Van Pelt and Oerlemans, 2012). PISM employs a sub-grid
872 grounding line scheme (Feldmann et al., 2014) in which sub-ice shelf melt may be
873 interpolated across the grounded-floating ice junction, or not. The former approach
874 tends to accelerate ice sheet retreat in marine basins, whereas the latter produces a
875 slower response (Golledge et al., 2015). Equilibrium states tend to differ less between
876 the two approaches than do retreat rates, however. For completeness, we run duplicate
877 experiments both with, and without, the sub-ice shelf grounding line melt
878 interpolation scheme turned on. Once afloat, we impose two calving mechanisms –
879 one based on horizontal strain rates (Levermann et al., 2012), and another in which a
880 simple minimum thickness criterion (200m) must be satisfied.

881 With the model setup described above, we undertook a series of simulations to
882 explore the range of likely ice sheet responses to MIS31 climate states as simulated by
883 the RCM. We ran experiments for two climate scenarios, reflecting atmospheric CO₂
884 concentrations of 280 and 400 ppm, and for each of these ran duplicates with each of
885 the two grounding line parameterisations described above. We also ran a control
886 experiment in which a present-day climate field was maintained throughout. All
887 simulations were run for 10000 years and were initialised from a thermally and
888 dynamically stable configuration used previously (Golledge et al., 2015; Golledge et
889 al., 2017a; 2017b) which closely resembles the present-day ice sheet (Fig. S4 panel a).
890 RCM-derived climatologies were imposed at the beginning of each run and the ice
891 sheet was allowed to freely evolve to the new boundary conditions. Ice sheet
892 configurations after 5000 years for each of the experiments are shown in Figure S4.



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Fig. S4. Ice sheet geometry and contribution to global sea level after 5000 years of simulated MIS31 climates, using either 280 ppm or 400 ppm atmospheric CO_2 , and with one of two different grounding-line schemes (noSGm, SGm; see text for details). Note the rapid response of the ice sheet under all scenarios (panel b), corresponding to the collapse timescale of WAIS.

Timeseries data illustrate that mass loss more-or-less stabilises after approximately 5000 years in most cases. In all MIS31 experiments, WAIS is lost relatively rapidly (first 1-2 millennia), with subsequent mass loss (from EAIS) taking place more slowly.

904

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