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► **To cite this version:**

Christo Buizert, T. Fudge, William Roberts, Eric Steig, Sam Sherriff-Tadano, et al.. Antarctic surface temperature and elevation during the Last Glacial Maximum. *Science*, 2021, 372 (6546), pp.1097-1101. 10.1126/science.abd2897 . hal-03249154

**HAL Id: hal-03249154**

**<https://hal.science/hal-03249154>**

Submitted on 1 Jul 2021

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# 1 **Antarctic surface temperature and elevation during the Last**

## 2 **Glacial Maximum**

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45 **Abstract:** Water stable isotopes in polar ice cores are a widely used temperature proxy in  
46 paleoclimate reconstruction, yet calibration remains challenging in East Antarctica. Here we  
47 reconstruct the magnitude and spatial pattern of Last Glacial Maximum surface cooling in  
48 Antarctica using borehole thermometry and firn properties in seven ice cores. West Antarctic  
49 sites cooled  $\sim 10^{\circ}\text{C}$  relative to preindustrial. East Antarctic sites show a range from  $\sim 4$  to  $\sim 7^{\circ}\text{C}$   
50 cooling, consistent with results of global climate models when the effects of topographic changes  
51 indicated by ice-core air-content data are included, but less than indicated by use of water stable  
52 isotopes calibrated against modern spatial gradients. An altered Antarctic temperature inversion  
53 during the glacial reconciles our estimates with water isotope observations.

54 **One Sentence Summary:** Temperature reconstructions based on borehole thermometry and firn  
55 properties, suggest that interpretation of ice core water isotopes using modern spatial slopes  
56 overestimates last glacial maximum surface cooling in central East Antarctica.

57 **Main Text:**

58 Using oxygen and hydrogen isotope ratios in ancient polar ice as records of past site temperature  
59 requires a calibration (1). Surface temperature and isotopic composition of precipitation correlate  
60 spatially in Antarctica, with a regression coefficient  $\alpha_s$  (spatial slope) of  $0.80 \text{ } \text{‰K}^{-1}$  for  $\delta^{18}\text{O}$  (2).  
61 Reconstructing past temperatures requires regression over time, and this temporal slope  $\alpha_T$  may  
62 differ from  $\alpha_s$ . In East Antarctica where the longest continuous ice core records, going back to 800  
63 ka BP (thousands of years before present), have been extracted (3), independent temperature  
64 estimates are not available and the spatial slope is commonly used to convert isotopic ratios to  
65 temperature (1); this approach gives a surface temperature difference  $\Delta T_s$  between the Last Glacial  
66 Maximum (LGM, 26-18 ka BP) and preindustrial of around  $-9^{\circ}\text{C}$  (1, 4, 5).

67 Antarctic LGM-preindustrial isotope changes depend on many factors including hemispheric sea-  
68 surface temperatures (6), sea ice extent (7), ice sheet elevation (8), vapor origin and transport,  
69 precipitation seasonality, and post-depositional isotopic exchange (9). Isotope-enabled general  
70 circulation models seek to capture these physical processes, making them an invaluable tool for  
71 studying isotopic variations. Such models simulate LGM-preindustrial  $\alpha_T$  ranging from 0.3 to 1.4  
72 ‰K<sup>-1</sup> in central East Antarctica (implied  $\Delta T_s$  of 4 to 20°C), implying several aforementioned  
73 processes are poorly constrained (8, 10-12).

74 We distinguish three temperatures: (i) the climatic temperature  $T_{CLIM}$  at constant elevation (relative  
75 to the present-day geoid); (ii) the surface temperature  $T_s$ , which may differ from the climatic  
76 temperature due to changing ice sheet topography; and (iii) the vapor condensation temperature  
77  $T_c$ , which is warmer than the surface due to the strong Antarctic inversion (2, 13).

78 Here, we empirically reconstruct LGM surface temperature across Antarctica (Fig. 1) using two  
79 independent methods. We investigate five East Antarctic ice cores: EPICA (European Project for  
80 Ice Coring in Antarctica) Dome C (EDC), EPICA Dronning Maud Land (EDML), Dome Fuji  
81 (DF), Talos Dome (TAL), and South Pole (SP); and two West Antarctic cores: West Antarctic Ice  
82 Sheet (WAIS) Divide (WD), and Siple Dome (SDM).

83 First, we estimate  $\Delta T_s$  at EDC and DF from the measured borehole temperature profiles (Fig. 2)  
84 using a method similar to that employed recently at WD (14). Due to the downward ice flow and  
85 low thermal diffusivity, the ice sheet maintains an imprint of its past surface temperature history.  
86 The large ice sheet thickness at EDC and DF is favorable for preserving past temperatures, yet the  
87 low accumulation rate is not. Consequently, the relative uncertainty in the EDC and DF borehole  
88 reconstructions is larger than that at WD. To constrain the problem better we use downward ice

89 velocities measured via phase-sensitive radio echo sounding (EDC only), and accurate age  
90 constraints derived via volcanic synchronization to the layer-counted WD timescale.

91 We force a 1-D heat transport-ice flow model at the surface boundary with a temperature history  
92 based on the  $\delta^{18}\text{O}$  record scaled with a constant  $\alpha_T$  value (10). Applying traditional isotope scaling  
93 ( $\alpha_T \approx 0.7 \text{ ‰ K}^{-1}$ , yielding  $\Delta T_s = -9^\circ\text{C}$  at EDC and  $-7.5^\circ\text{C}$  at DF) simulates temperature profiles that  
94 do not fit the borehole observations at either site (Fig. 2). At EDC the model-data fit is optimized  
95 for  $\alpha_T = 1.14 \text{ ‰ K}^{-1}$  consistent with  $\Delta T_s = -5.5^\circ\text{C}$  (95% confidence range is  $-6.9^\circ\text{C}$  to  $-3.1^\circ\text{C}$ ). At  
96 DF the optimal  $\Delta T_s$  is in the  $-2.0^\circ\text{C}$  to  $-5.4^\circ\text{C}$  range; we provide a range without a best estimate  
97 because at DF there are no direct constraints on the downward ice velocity. In Fig. 1 the WD, EDC  
98 and DF borehole estimates are marked “BH”.

99 Second, we reconstruct past climate at all seven sites using the dependence of firn densification,  
100 the gradual transformation of polar snow to ice, on  $T_s$  and accumulation rate ( $A$ ). Air bubbles are  
101 isolated from the atmosphere at the lock-in depth (50-120 m below the surface), an event preserved  
102 in two ice core signals (15):  $\delta^{15}\text{N}$  of  $\text{N}_2$  which records past firn column thickness via gravitational  
103 enrichment, and the gas age-ice age difference or  $\Delta\text{age}$ . Critically, the  $\delta^{15}\text{N}$  and  $\Delta\text{age}$ -isopleths are  
104 perpendicular in  $T_s$ - $A$  space (Fig. 3A), meaning that if  $\delta^{15}\text{N}$  and  $\Delta\text{age}$  are independently known, a  
105 unique climatic [ $T_s$ ,  $A$ ] solution exists (subject to the uncertainties of the firn model).

106 Synchronization using both volcanic deposits and globally synchronous abrupt atmospheric  
107 methane variations, allows us to estimate  $\Delta\text{age}$  empirically for the Antarctic ice cores (10, 16). We  
108 use an inverse dynamical firn densification-heat transport model (17, 18) to reconstruct  $T_s$  and  $A$   
109 histories that optimize the fit to  $\Delta\text{age}$  and  $\delta^{15}\text{N}$  data (Fig. 3B-C). Reconstructed accumulation rates  
110 agree (within uncertainty) with independent estimates (Fig. S8). Methodological biases and

111 uncertainties are estimated using a Monte-Carlo approach (10). The histograms in Fig.1 give the  
112  $\Delta T_s$  distribution of the  $\Delta$ age-based reconstruction.

113 In East Antarctica,  $\Delta T_s$  ranges from  $-3.8 \pm 2.0^\circ\text{C}$  (DF) to  $-7.1 \pm 1.7^\circ\text{C}$  (TAL); at DF, EDC and EDML,  
114  $\Delta T_s$  is substantially lower than estimates from isotope scaling using  $\alpha_s$ . The two West Antarctic  
115 sites have similar  $\Delta T_s$  of  $-10.2 \pm 2.4^\circ\text{C}$  (SDM) to  $-10.3 \pm 1.3^\circ\text{C}$  (WD). The  $\Delta$ age- and borehole-based  
116 reconstruction methods agree within uncertainty at all sites (Fig. 1). Allowing for more flank-like  
117 ice flow at EDC during the glacial period (which would occur if the divide position were different  
118 than at present), improves the agreement by changing the borehole estimate to around  $-4.5^\circ\text{C}$  (10);  
119 we choose to report the  $-5.5^\circ\text{C}$  value to keep both methods independent. PMIP4 (Paleoclimate  
120 Modeling Intercomparison Project phase 4) simulations (19) find a seven-site-mean  $\Delta T_s$   
121 magnitude that is  $1.2 \pm 4.6^\circ\text{C}$  larger than our  $\Delta$ age-based reconstructions (mean and spread of ten  
122 climate models; Fig. 1).

123 We emphasize that the firm method is primarily constrained by the empirical  $\Delta$ age estimates.  
124 Because  $T_s$  and  $A$  broadly co-vary via the saturation vapor pressure, the deglacial climatic changes  
125 run parallel to the  $\delta^{15}\text{N}$ -isopleths (Fig. 3A). Therefore,  $\delta^{15}\text{N}$  data alone do not constrain the  
126 magnitude of climate change meaningfully. The effects of  $T_s$  and  $A$  are additive in  $\Delta$ age, however,  
127 making  $\Delta$ age a sensitive proxy for climate change (Fig. 3D), as first noted by Jakob Schwander  
128 (20). The empirical  $\Delta$ age at 24ka is larger than at 18 ka BP for all five cores where both are  
129 available, and coldest conditions in Antarctica occur around 27-24 ka BP in our reconstructions  
130 (Fig. S8h); this follows expectations from local insolation (21).

131 We propose that elevation changes explain the spatial differences in  $\Delta T_s$  (8). Let  $\Delta z$  be the LGM  
132 elevation anomaly relative to present. We present new WD and DF total air content data (Fig.  
133 S12), and interpret them in terms of elevation change (22). These data suggest a 420 m (range: 280  
134 – 590 m) contrast in  $\Delta z$  between WD and central East Antarctica (here DF and EDC) – for example  
135  $\Delta z = +300$  m at WAIS and  $\Delta z = -120$  m in central East Antarctica (Fig. 4B). Our estimate is broadly  
136 in agreement with LGM ice sheet reconstructions that suggest a West-East  $\Delta z$  contrast between  
137 160 and 560 m (10). Although the implied  $\Delta z$  at WAIS exceeds the observed highstand at ice  
138 margin nunataks (23), such data do not strongly constrain the elevation at WD over 500 km away.  
139 The corresponding  $\Delta T_s$  contrast (WD  $\Delta T_s$  minus the average  $\Delta T_s$  at DF and EDC) is  $-6.2 \pm 2.3^\circ\text{C}$   
140 in the  $\Delta$ age-based reconstructions,  $-6.0 \pm 2.0^\circ\text{C}$  in the borehole reconstructions, and  $-5.9 \pm 2.7^\circ\text{C}$   
141 in the PMIP4 model ensemble; the level of agreement suggests this is a robust feature of Antarctic  
142 LGM climate. This temperature contrast is thus plausibly linked to  $\Delta z$  via the (spatial) lapse rate  
143 in interior of Antarctica of around  $-12^\circ\text{C km}^{-1}$  (2, 24).

144 To further assess the elevation impact on  $\Delta T_s$  we perform an atmosphere-ocean general circulation  
145 model (AOGCM) sensitivity study of Antarctic LGM climate using the MIROC and HadCM3  
146 models and a series of LGM topographic reconstructions (10). We first estimate climatic LGM  
147 cooling using full LGM boundary conditions (including LGM albedo) but preindustrial Antarctic  
148 topography; this yields a seven-site average  $\Delta T_{\text{CLIM}}$  of  $-4.7^\circ\text{C}$  and  $-7.0^\circ\text{C}$  in the MIROC and  
149 HadCM3 models respectively, but stronger albedo-driven cooling is found over the Ross and  
150 Weddell Seas due to ice growth onto the continental shelf (Fig. 4A). Note that simulated climatic  
151  $\Delta T_{\text{CLIM}}$  is similar in interior West and East Antarctica in the absence of topographic change.

152 Next, we perform climate simulations with five Antarctic LGM topographic reconstructions. These  
153 reconstructions suggest  $\Delta z$  of +100 to +600 m in interior WAIS and down to -250 m in interior  
154 East Antarctica (Fig. 4B). These changes result in greater  $\Delta T_s$  in West than in central East  
155 Antarctica (Fig. 4C), in agreement with our reconstructions. By comparing the various topographic  
156 reconstructions, we find that  $\Delta T_s$  is closely linked to  $\Delta z$  in both models following the dry adiabatic  
157 lapse rate of  $-9.8^\circ\text{C km}^{-1}$  (Fig. 4D). There is also a fraction of the variance that cannot be explained  
158 by lapse rate effects that is due to the topography altering the atmospheric circulation around  
159 Antarctica. We find a correlation  $r = 0.96$  between the reconstructed and the simulated site  $\Delta T_s$   
160 pattern (averaged across the five topographic reconstructions and both models); for the PMIP4  
161 multi-model mean this correlation is  $r = 0.95$ . We conclude that LGM ice sheet topography change  
162 plausibly explains the  $\Delta T_s$  spatial variability in our reconstruction (8).

163 Our findings have implications for the interpretation of water isotopes in Antarctic ice cores. We  
164 find  $\alpha_T$  in the range of 0.9 to  $1.4 \text{ ‰K}^{-1}$  in East Antarctica and therefore  $\alpha_T > \alpha_S$ , opposite to  
165 Greenland where  $\alpha_T < \alpha_S$  (17, 25). We compare our  $\alpha_T$  with those from LGM and preindustrial  
166 simulations using the latest generation isotope-enabled Community Earth System Model (iCESM,  
167 Fig. 4E). The good agreement ( $r = 0.91$ ;  $0.06 \text{ ‰ K}^{-1}$  mean offset) demonstrates our reconstructed  
168  $\alpha_T$  are consistent with isotope physics, yet the large inter-model spread in simulated  $\alpha_T$  (see section  
169 S3.5 in (10) for a review) prevents us from claiming it validates our results. While the  $\alpha_T$  agree  
170 well, iCESM simulates a  $\Delta T_s$  and LGM-preindustrial  $\delta^{18}\text{O}$  change that are both too large  
171 (compared to our reconstructions and ice core data, respectively).

172 Last, we investigate changes to the strong surface-based inversion in the Antarctic boundary layer  
173 (Fig. 4F). The condensation temperature  $T_C$  is higher than  $T_s$ , and they correlate spatially with a

174 slope  $dT_C/dT_S$  in the 0.63-0.67 range (2, 13, 26).  $T_C$  controls precipitation  $\delta^{18}\text{O}$ , with a present-day  
175 spatial sensitivity of  $d\delta^{18}\text{O}/dT_C = d\delta^{18}\text{O}/dT_S \times dT_S/dT_C \approx 0.80/0.65 = 1.23 \text{ ‰ K}^{-1}$ . We now assume  
176 that, unlike  $\Delta T_S$ , the LGM-preindustrial change  $\Delta T_C$  can be estimated using this spatial slope via  
177  $\Delta T_C = \Delta\delta^{18}\text{O}/1.23$  (Fig. 4F). At WD and SDM the  $\alpha_T \approx \alpha_S$  assumption holds, suggesting the ratio  
178  $\Delta T_C/\Delta T_S$  is close to the present-day ratio of 0.65; in central East Antarctica the ratio  $\Delta T_C/\Delta T_S$   
179 exceeds 0.65 consistent with  $\alpha_T > \alpha_S$ . We plotted simulated  $\Delta T_S$  vs.  $\Delta T_C$  across interior Antarctica  
180 from a wide range of AOGCMs and topographies; we find the ratio  $\Delta T_C/\Delta T_S$  ranges from 0.48 to  
181 1.3 (95% interval, grey lines) with our empirical reconstructions falling within the model data  
182 cloud (Fig. 4F). In aggregate these simulations find that  $\Delta T_C/\Delta T_S$  tends to exceed the present-day  
183 ratio of 0.65 (~79% of model data points) – such a change to the inversion structure would result  
184 in  $\alpha_T > \alpha_S$  for  $\Delta T_S$ . In the iCESM simulations the  $\Delta T_C/\Delta T_S$  and  $\alpha_T$  fields look similar, with the  
185  $\Delta T_C/\Delta T_S = 0.65$  contour line broadly aligning with the  $\alpha_T = 0.8 \text{ ‰K}^{-1}$  contour line (Fig. S11). We  
186 conclude that physically plausible changes to the inversion (27, 28) may reconcile our  
187 reconstructions with previous work on Antarctic LGM water isotopes.

188 Our reconstructions improve the LGM Antarctic temperature estimation and provide a benchmark  
189 for testing the ability of (isotope-enabled) climate models to simulate climate states radically  
190 different from the late Holocene. For surface temperature, the spatial isotopic slope is not always  
191 a good approximation of the temporal slope, challenging the prevalent interpretation of ice core  
192 water isotopes in Antarctica.

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261

262 **Acknowledgments:** The idea of weak East Antarctic LGM cooling was suggested a decade ago  
263 by Sigfús J. Johnsen (1940–2013) and Jakob Schwander, but never published. We thank Emilie  
264 Capron and Amaëlle Landais for help in gathering  $\delta^{15}\text{N}$  data; Takeo Hondoh and Takao Kameda  
265 for support in Dome Fuji air content analyses; Camille Bréant for sharing model output; Vasileios  
266 Gkinis for useful discussions; Wing-Le Chan for assistance with model analysis; Jean-Yves  
267 Peterschmitt for compiling PMIP4 model data, and all PMIP4 groups for sharing model output;  
268 Chris Adams of Corvidopolis for providing a toddler-free workspace during the COVID-19  
269 pandemic.

270 **Funding:** This work was supported by the US National Science Foundation (NSF) through grants  
271 1643394 (to C.B.), 1643355 (to T.J.F. and E.J.S), 1443472, 1643722 (to E.J.B), and 1738934 (to  
272 R.B.A.); the University of Washington Royalty Research Fund (to T.J.F); MEXT and the Japan  
273 Society for the Promotion of Science KAKENHI Grant Numbers 18749002, 26241011,  
274 15KK0027, 17H06316 and 20H00639 (to K.K), 20H04327 (to I.O.), 22310003 (to T.N.),  
275 15J12515, 17H06104, 17H06323 and 20K14552 (to A.A-O); the National Research Foundation  
276 of Korea (NRF) through grants NRF-2018R1A2B3003256 and NRF-2018R1A5A1024958 (to

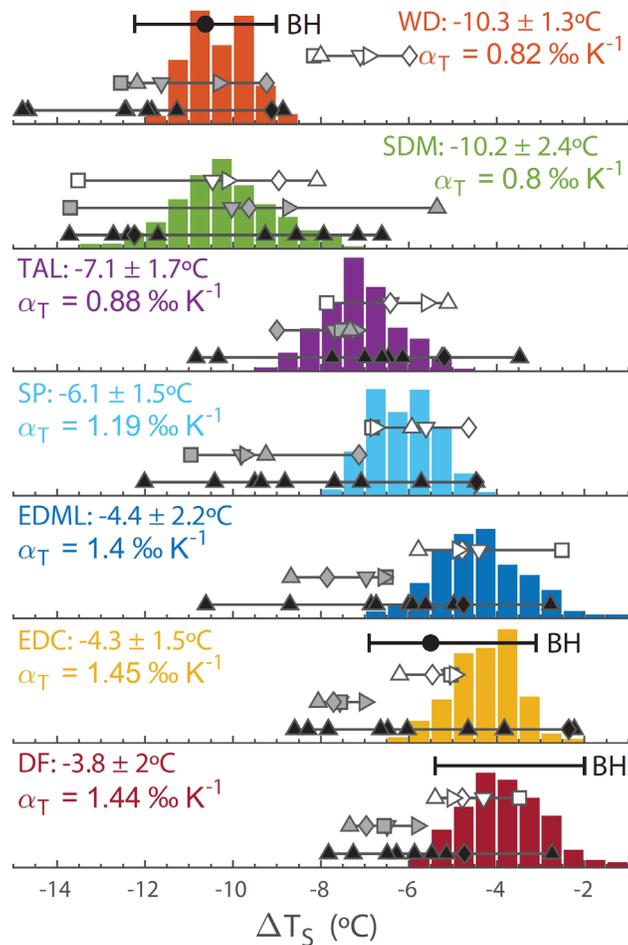
277 J.A.); the European Research Council (ERC) under the European Union's Horizon 2020 research  
278 and innovation programme through grant agreement No 820047 (to M.Si.); IT-MIUR-PNRA  
279 (Italian Antarctic Research Program) through the BE-OI (PNRA16\_00124) project (to M.Se.); the  
280 Villum Investigator Project IceFlow (NR. 16572, to A.S.); Beyond EPICA - Oldest Ice Core EU  
281 Coordination and Support Action; measurements at EDC (Concordia Station) were supported by  
282 the French Polar Institute (IPEV, prog 902) and the Italian Antarctic Program (PNRA and ENEA).

283 **Author Contributions:** C.B., E.J.S., and W.H.G.R. conceived of the study;  $\Delta$ age-based  
284 reconstructions by C.B.; ice flow and borehole temperature modeling by T.J.F. and C.R.; GCM  
285 modeling by W.H.G.R., S.S.-T., T.O., A.A.-O., C.H., Z.L., J.Z., and B.O.-B.; PMIP4 model  
286 compilation by M.K.; methane data by E.J.B., J.E., J.A., K.M., J.C., T.A.S., K.K., S.A. and T.N.;  
287  $\delta^{15}\text{N-N}_2$  data by I.O., K.K., R.B., J.P.S., S.A. and T.N.; borehole thermometry data by C.R., E.L.  
288 and H.M.; ApRES data by C.M. and H.C.; SP isotope data by E.J.S., E.C.K. and T.R.J.; air content  
289 data and corrections by J.E., K.K., V.L., J.F. and R.B.A.; SDM chronology by T.J.F. and C.B.;  
290 SDM tephra matches by N.W.D.; volcanic matching by M.Si., M.Se., A.S., T.J.F. and C.B.; J.Sc.  
291 (and S.J.J.) independently performed a similar study that reached the same conclusion around a  
292 decade ago; all authors contributed towards the final manuscript.

293 **Competing interests:** The authors declare no competing financial interests.

294 **Data and materials availability:** All new ice core data from this study are available in the  
295 supplementary material data sheet and online at <https://www.ncdc.noaa.gov/paleo/study/32632>;  
296 previously published data are available with their original publications and/or in publicly  
297 accessible online data archives.

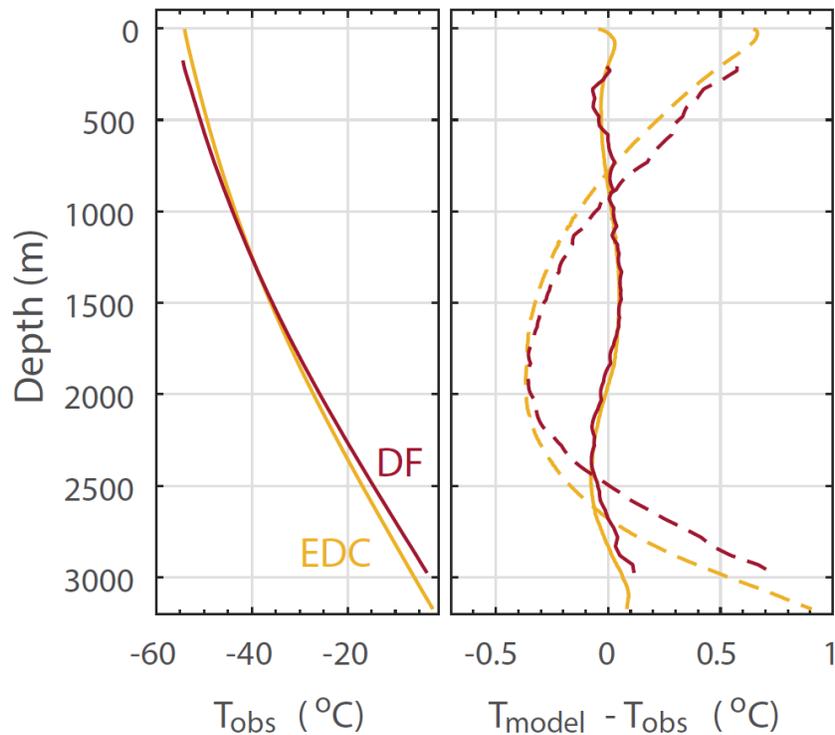
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299

300 **Figure 1. Summary of Antarctic LGM cooling estimates.** Black markers with horizontal  
 301 errorbars marked “BH” give borehole estimates; WD results from ref. (14). Histograms give  
 302 distribution of  $\Delta$ age-based temperature reconstructions from a Monte Carlo sampling ( $N = 1000$ )  
 303 of model parameters; listed are mean and  $2\sigma$  standard deviation of the distribution, as well as the  
 304 implied temporal isotope slope  $\alpha_T$ .  $\Delta T_S$  is the LGM (18-21.4 ka BP) minus preindustrial (0.5-2.5  
 305 ka BP) condition. White (MIROC), grey (HadCM3) and black (PMIP4) show AOGCM-simulated  
 306  $\Delta T_S$ , with symbols denoting different LGM topography reconstructions (10): Pollard and Deconto  
 307 2009 (downward triangle); Whitehouse et al. 2012 (square); Glac-1D (diamond, (29)); Golledge  
 308 et al. 2014 (rightward triangle); Ice-6G (upward triangle).

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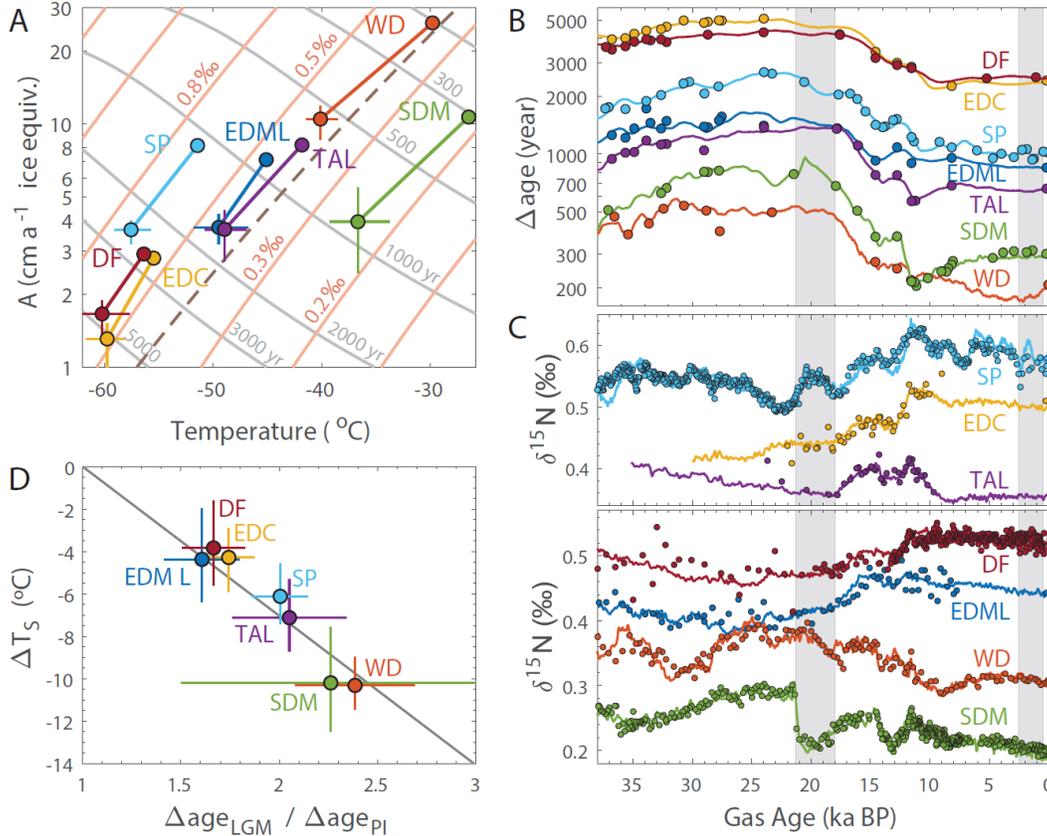


310

311 **Figure 2. Borehole temperature reconstruction for EDC and DF.** Left panel: Site borehole  
 312 temperature observations at EDC (yellow) and DF (red). At both sites the ice-bedrock interface is  
 313 at the pressure melting point ( $-2.2^{\circ}\text{C}$ ). Right panel: model-data mismatch at EDC (yellow) and DF  
 314 (red) for an ice flow-heat transport model forced by the optimized temperature histories (solid  
 315 lines,  $\Delta T_s$  of  $-5.5^{\circ}\text{C}$  at EDC and  $-3.2^{\circ}\text{C}$  at DF), and forced with water-isotope scaling of  $0.7\text{‰K}^{-1}$   
 316 <sup>1</sup> (dashed lines,  $\Delta T_s$  of  $-9.0^{\circ}\text{C}$  at EDC and  $-7.5^{\circ}\text{C}$  at DF).

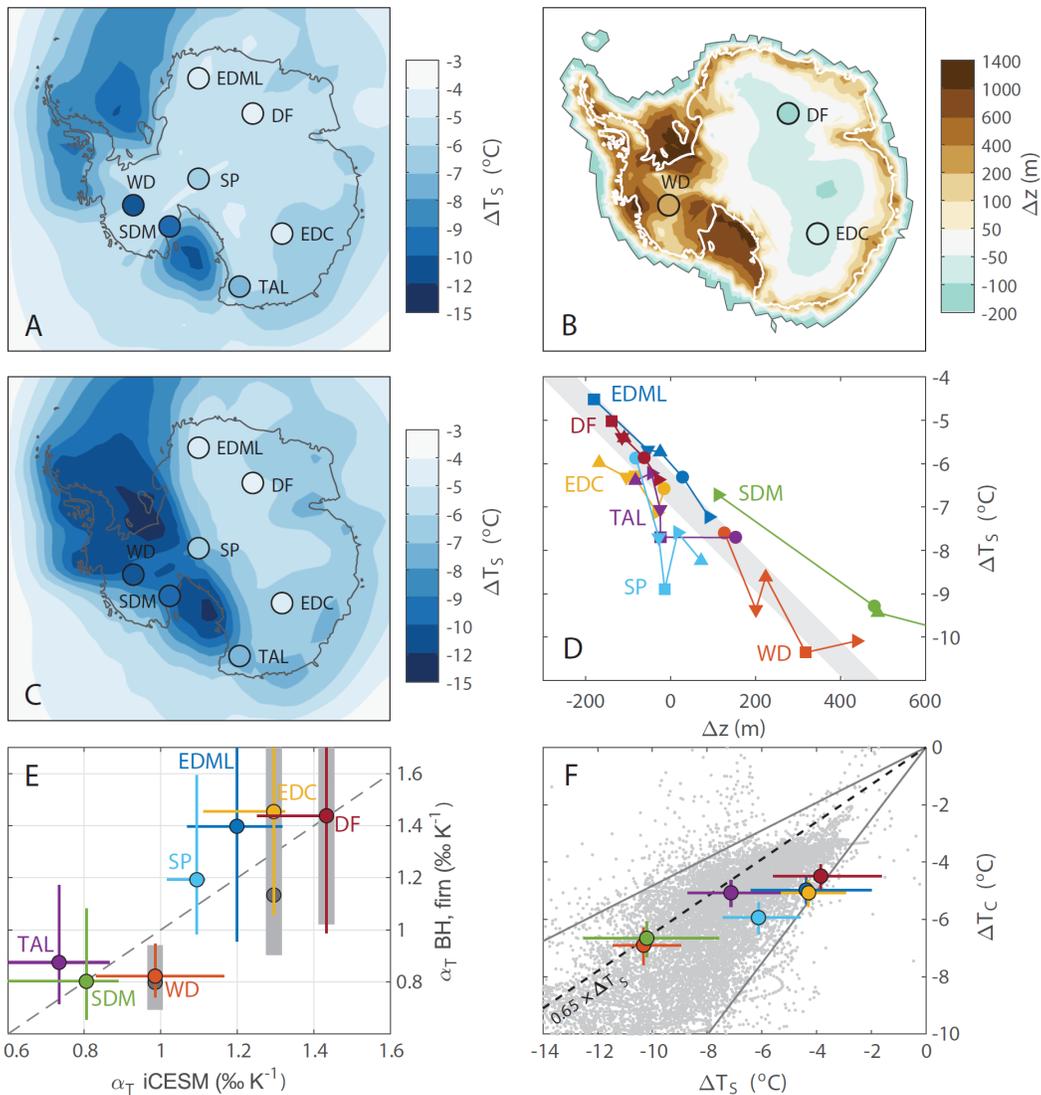
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320 **Figure 3.  $\Delta\text{age}$ -based temperature reconstructions.** (A)  $\Delta\text{age}$  and  $\delta^{15}\text{N}$ -isopleths (grey and  
 321 salmon, respectively) in the steady-state Herron-Langway firm densification model as a function  
 322 of  $T_s$  and  $A$ . Dashed line shows accumulation scaling via the saturation vapor pressure at the site  
 323 (ignoring the atmospheric inversion). Reconstructed preindustrial and LGM conditions at the  
 324 seven sites are indicated. (B) Model fit to empirical  $\Delta\text{age}$  constraints. Grey vertical bars denote the  
 325 LGM (21.4-18 ka BP) and preindustrial (2.5-0.5 ka BP) periods. (C) Model fit to  $\delta^{15}\text{N}$  data, divided  
 326 over two panels to prevent overlapping curves. Data shown on the WD2014 timescale (30, 31).  
 327 (D) Reconstructed  $\Delta T_s$  versus ratio of LGM  $\Delta\text{age}$  over preindustrial  $\Delta\text{age}$  (with linear fit), showing  
 328 the utility of  $\Delta\text{age}$  as a climate proxy.



329

330 **Figure 4. Climate models and Antarctic topography.** (A) AOGCM simulations of  $\Delta T_s$  using  
 331 preindustrial ice topography in Antarctica (average of MIROC and HadCM models), with  $\Delta$ age-  
 332 based  $\Delta T_s$  reconstructions for the seven sites. (B) simulated LGM elevation anomaly (shaded,  
 333 average of five topographies) with LGM elevation anomaly of +310 m, -80 m, and -140 m at WD,  
 334 EDC and DF (10) (C) As in panel (A), but using LGM ice topography in Antarctica (average of  
 335 five LGM topographies and both MIROC and HadCM models). (D) Elevation change vs.  $\Delta T_s$  in  
 336 the AOGCM simulations (average of MIROC and HadCM models); symbols denote the different

337 LGM topographic reconstructions (see Fig. 1 caption for legend). The grey bar shows the dry  
338 adiabatic lapse rate. **(E)** Temporal isotope slope  $\alpha_T$  from the iCESM model against our  
339 reconstructions (borehole in grey,  $\Delta$ age-based in colors). **(F)**  $\Delta T_s$  vs  $\Delta T_C$  from  $\Delta$ age-based  $\Delta T_s$  and  
340 isotope-based  $\Delta T_C$  (large dots with error bars) and from LGM-preindustrial AOGCM simulations  
341 (small grey dots, grey lines enclose the central 95% of estimates); black dashed line gives modern  
342 spatial slope (2). Models plotted are PMIP3 (except CNRM-CM5 that simulates  $\Delta T_s > 0^\circ\text{C}$ ),  
343 PMIP4 (all model output publicly available), and all iCESM, MIROC and HadCM3 simulations  
344 used in this work; we show interior Antarctica (surface pressure  $> 800$  hPa);  $T_C$  is taken to be the  
345 annual mean troposphere temperature maximum (typically around 500hPa). The models have an  
346 average preindustrial spatial  $dT_C/dT_s$  of 0.68 (range: 0.31 to 0.89) in interior Antarctica.

347

348 **Content of the Supplementary Materials:**

349 Materials and Methods

350 Figures S1 to S12

351 Tables S1 to S7

352 References 32-154 are only called out in the supplementary materials.

353

# Science



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## Supplementary Materials for

### **Antarctic surface temperature and elevation during the Last Glacial Maximum**

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Eric Lefebvre, Jon Edwards, Kenji Kawamura, Ikumi Oyabu, Hideaki Motoyama, Emma C. Kahle,  
Tyler R. Jones, Ayako Abe-Ouchi, Takashi Obase, Carlos Martin, Hugh Corr, Jeffrey P. Severinghaus,  
Ross Beaudette, Jenna Epifanio, Edward J. Brook, Kaden Martin, Jérôme Chappellaz, Shuji Aoki,  
Takakiyo Nakazawa, Todd A. Sowers, Richard Alley, Jinho Ahn, Michael Sigl, Mirko Severi,  
Nelia W. Dunbar, Anders Svensson, John Fegyveresi, Chengfei He, Zhengyu Liu, Jiang Zhu,  
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#### **This PDF file includes:**

- Materials and Methods
- Figures S1 to S12
- Tables S1 to S7
- Additional References 32-154

#### **Other Supplementary Material for this manuscript includes the following:**

- Data Excel file S1

## 26 **Materials and Methods**

27

### 28 **S1 Borehole temperature reconstructions at Dome C and Dome F**

#### 29 **S1.1 Models and data**

##### 30 **S1.1.1 Forward ice flow-heat transport model**

31 We use a transient one-dimensional ice-flow model to compute the vertical-velocity profile through time:

$$32 \quad w(\hat{z}) = -(\dot{b} - \dot{m} - \dot{H})\psi(\hat{z}) - \dot{m} - \left(\frac{\rho_{ice}}{\rho_{firm}} - 1\right)\dot{b} \quad (S1)$$

33 where  $\hat{z}$  is the non-dimensional height above the bed,  $\dot{b}$  is the accumulation rate,  $\dot{m}$  is the melt rate,  $\dot{H}$  is  
34 the rate of ice-thickness change,  $\rho_i$  is the density of ice,  $\rho_{firm}$  is the density profile and  $\psi(\hat{z})$  is the vertical  
35 velocity shape function computed as:

$$36 \quad \psi(\hat{z}) = \left(1 - \frac{p+2}{p+1}(1 - \hat{z}) + \frac{1}{p+1}(1 - \hat{z})^{p+2}\right) \quad (S2)$$

37 following Lliboutry (1979) where  $p$  is vertical velocity shape parameter (32). Firm compaction is  
38 incorporated through the right hand term in equation S1 and assumes a density profile that does not vary  
39 with time.

40 The heat equation following Cuffey and Paterson (2010) is (33):

$$41 \quad \rho c_p \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left( k \frac{\partial T}{\partial z} \right) - \rho c_p w \frac{\partial T}{\partial z} + Q \quad (S3)$$

42 where  $c_p$  is the heat capacity,  $T$  is temperature,  $k$  is the thermal conductivity, and  $Q$  is the heat production  
43 term, and  $z$  is the height above the bed. The firm density profile is modeled using the Herron and Langway  
44 model (18) matched to modern measurements to assure a smooth vertical velocity profile. Values of  $c_p$  are  
45 temperature dependent and calculated at each time step using the relationships from ref. (34). For  $k$ , we use  
46 three different relationships as described below (Section S1.2.1). The thermal conductivity in the firm is  
47 found using the relationship in ref. (35).

##### 48 **S1.1.2 Borehole temperature measurements**

49 The EDC borehole temperature measurements were made in 2008 to a depth of 3235 m where the probe  
50 could no longer descend, possibly due to partial borehole closure. The temperature was extended to the  
51 pressure melting temperature of  $-2.18^\circ\text{C}$  at the ice-rock interface at 3275 m. The principle of the probe is a  
52 four-wires circuit to measure the resistance of the temperature sensor (36). The upper 100 m of  
53 measurements are less reliable because they were made in the cased portion of the firm and have been  
54 excluded from our analysis. The data have been smoothed with a 5 m Hanning window. The uncertainty of  
55 the temperature measurements is 1 cK. The temperature profile is shown in Fig. 1 of the main text.

56 The Dome Fuji borehole temperature measurements were made in January 2013 on the DF-2 borehole using  
57 a Pt-100 resistance thermometer. As at EDC, the upper 100 m of measurements are less reliable because  
58 they were made in the cased portion of the firm and have been excluded from our analysis. The data are  
59 averaged over 50 m intervals to improve the signal-to-noise ratio. The uncertainty of the temperature  
60 measurements is 5 cK. The temperature profile is shown in Fig. 2 of the main text.

##### 61 **S1.1.3 Vertical velocity measurement**

62 Phase sensitive radar measurements (ApRES) allow the determination of englacial vertical velocities (37).  
63 Three sets of measurements have been made with one year repeat intervals at Dome C. We focus on the  
64 highest quality measurements made in 2016 and 2017 (Figure S1). We use the ApRES measurements to  
65 constrain the Lliboutry  $p$  parameter by finding the best fit where the surface vertical velocity is also a free  
66 parameter (38). We do not fit the upper 200 m of measurements because these are affected by firn  
67 compaction which introduces additional uncertainty.

68 The uncertainty in the difference in vertical position of the ApRES antennas is larger than the changes in  
69 vertical position of the reflectors, such that a choice about the vertical velocity at the basal interface must  
70 be made. If the bed is frozen, then the basal vertical velocity can be assumed to be zero such that the ApRES  
71 measurements can be uniformly shifted (38). However, Dome C is potentially melting at the bed (39), so  
72 the basal velocity may not necessarily be zero. The deepest 15 m of ApRES measurements show near zero  
73 values, but from 15 to 30 m above the bed, the average is  $-0.0035 \text{ m a}^{-1}$ , with an average of all 30 m of  
74  $-0.0022 \text{ m a}^{-1}$ . Therefore, there are a number of different ways to fit the vertical velocity, by either shifting  
75 the ApRES measurements, assuming a basal melt rate in the Lliboutry model fit, or both. Note that because  
76 of the way a basal melt rate affects the vertical velocity profile, shifting the ApRES measurements by a  
77 given value does not yield the same fit as imposing the same value as the modeled basal melt rate. We use  
78 5 combinations of assumptions to find the range of potential of  $p$  values which are described in table S1.  
79 This give a range of values from  $p = 1.2$  to  $3.2$  with the surface vertical velocities ranging from  $0.03$  to  
80  $0.033 \text{ m a}^{-1}$ . The range of  $p$  values is similar to the range found from inverse modeling of the depth-age  
81 relationship (39). We use a value of  $p = 2$  as our base scenario and  $p = 1$  and  $3.5$  to define the range of  
82 uncertainty. The surface velocities found in the fits are a few  $\text{mm a}^{-1}$  greater than the modern accumulation  
83 rate. This could indicate a small component of vertical thinning (a meter to a few meters per 1000 years) or  
84 the uncertainty associated with the ApRES measurements.

85 No ApRES data are available for the Dome Fuji site.

#### 86 **S1.1.4 Ice-flow model forcing: accumulation rate**

87 The accumulation rate history is found using the best-fit vertical velocity profile to compute the thinning  
88 function and the timescale for EDC. We use a hybrid timescale where the most recent 57 ka use the  
89 timescale synchronized to WAIS Divide (40); prior ages are from AICC2012 (41, 42).

90 For Dome Fuji we use a similar approach in which the accumulation history is found using the thinning  
91 function in combination with the timescale of ref. (43).

#### 92 **S1.1.5 Ice-flow model forcing: surface temperature**

93 The surface temperature history is found from the modern temperature by scaling the temperature change  
94 to the deuterium record (3, 43). The  $\delta\text{D}$  record is first interpolated to even spacing and smoothed with a  
95 1000-year 7<sup>th</sup> order low pass filter. The surface temperature history,  $T_s$ , is found by

$$96 \quad T_s = T_0 + \Delta T_s \quad (\text{S4})$$

97 Where  $T_0$  is the modern temperature and the change in surface temperature,  $\Delta T_s$ , is found using equation  
98 S5 following ref. (14):

$$99 \quad \Delta T_s(\delta) = \alpha_T^{-1} \int_t \delta(t) dt \quad (\text{S5})$$

100

101 where  $\dot{\delta}$  is the time derivative of the filtered temporal  $\delta D$  record,  $t$  is time, and  $\alpha_T$  the isotope sensitivity  
102 (temporal slope). In all model results presented here,  $\alpha_T$  is a constant and is not allowed to vary through  
103 time.  
104

### 105 **S1.1.6 Inverse procedure**

106 We use a gradient descent inverse method by defining the mismatch index:

$$107 \quad J = \int_0^H \frac{[T(z) - \Theta(z)]^2}{\sigma_{\Theta(z)}^2} dz \quad (\text{S6})$$

108 where  $T(z)$  are the modeled borehole temperatures,  $\Theta(z)$  are the measured borehole temperatures, and  $\sigma_{\Theta(z)}$   
109 are the uncertainties in the measured borehole temperatures. The temperatures in the firn are excluded  
110 (S1.1.2). The inverse procedure finds the values of  $T_{\text{surface}}$ ,  $T_{\text{bed}}$ , and  $\alpha_T$ , which all have a single optimum  
111 value (i.e. unimodal distribution).

## 112 **S1.2 Results of temperature optimizations and uncertainty at Dome C**

113 We first discuss the borehole temperature reconstruction at Dome C, where we have the better data  
114 constraints; we apply the lessons from Dome C to the Dome F reconstruction in Section S1.3.

115 We performed a series of optimizations for different prescribed ice properties and ice-flow parameters to  
116 evaluate the most likely temperature change and its uncertainty. As described below, we consider two types  
117 of uncertainty: (1) the temperature dependence of thermal conductivity (Section S1.2.1); and (2) parameter  
118 and forcing history (Section S1.2.2). Our best estimate of the LGM (18-21.4 ka BP) minus PI (0.5-2.5 ka  
119 BP) temperature change is  $\Delta T_s = -5.5^\circ\text{C}$  based on the average of three relationships for the thermal  
120 conductivity and our preferred forcings:

- 121 • a vertical velocity profile with  $p=2$ , based on ApRES vertical velocities (Section S1.1.3)
- 122 •  $0.0005 \text{ m a}^{-1}$  basal melting based on ref. (39)
- 123 • 100 m of deglacial thickening between 17 and 7 ka (Section S3.3)

124 The total uncertainty range is approximately  $\pm 1.9^\circ\text{C}$ , which is the sum of the thermal conductivity  
125 uncertainty (Section 1.2.1, Table S2) and parameter/history uncertainty (Section S1.2.2, S1.2.3, Table S3  
126 and Table S4).

### 127 **S1.2.1 Thermal conductivity**

128 The temperature dependence of the thermal conductivity ( $k$ ) of ice is an important term for the inference of  
129 the LGM-Holocene temperature change at low accumulation sites. The standard relationship in Cuffey and  
130 Paterson (2010, p. 400) is from the review of Yen (1981), ref. (34). To evaluate the uncertainty due to the  
131 thermal conductivity, we use the relationship from two data sets that were not used in the Yen (1981)  
132 compilation (44, 45): Ross et al. (1978) and Waite et al., (2006). The thermal conductivity relationships are  
133 plotted in Figure S2, with the full temperature range in the left panel and only the temperature range of  
134 Dome C ice experiences in the right panel. Ross et al. (1978) reported only the fit, so the underlying data  
135 cannot be plotted.

136 The inferred temperature change and misfit are given in Table S2 and shown in Figure S3. The choice of  
137 relationship results in a  $1^\circ\text{C}$  difference in the inferred temperature change. Much of the model-data  
138 temperature misfit occurs in the bottom 500m. The near-bed temperature gradient, where the temperature  
139 gradient is nearly constant with depth, is not easily modeled. We test the impact of better fitting the near-  
140 bed temperature by decreasing the temperature dependence of the thermal conductivity. We also performed

141 tests in which we add strain heat in the basal ice to better match the basal temperature; because these last  
142 tests have the same impact as the changes in thermal conductivity on the inferred  $\Delta T_s$ , we do not discuss  
143 these model runs further.

144 To assess the influence of the thermal conductivity of the deepest ice, we perform tests in which we adjust  
145  $k$  for all ice with a temperature above  $-13^\circ\text{C}$ . This temperature was chosen because it corresponds to the  
146 depth at which the ice grain size and fabric shows a distinct change (46), which is thought to be associated  
147 with the onset of migration recrystallization. The model-data fit is improved (Table S2) when we increase  
148  $k$  linearly by up to 5% for the basal ice ( $-2.18^\circ\text{C}$ ). It is not clear how the changes in grain properties might  
149 impact  $k$ . Yen (1981) notes that  $k$  may be 5% greater along the c-axis of single crystals, which would be in  
150 the opposite direction to what we have applied because the c-axis fabric is becoming less vertically oriented  
151 near the bed (46); however, a study from the food science community suggests that the larger grain sizes  
152 (associated with slower freezing rates in that study) can cause a decrease  $k$  of 10%, although they do not  
153 report the grain sizes (47). While it is not clear how the change in grain properties should affect  $k$ , our  
154 adjustment provides a useful way to assess its impact on the inferred  $\Delta T_s$ . The closer match to the basal  
155 temperature gradient allows the temperature misfit to be reduced. For each of the three published thermal  
156 conductivity relationships, applying the 5%  $k$  adjustment changes  $\Delta T_s$  by around  $0.25^\circ\text{C}$ .

157 The temperature measurement uncertainty is  $0.01^\circ\text{C}$  at EDC. Therefore, we consider RMS values within  
158  $0.02^\circ\text{C}$  of each other as equally viable solutions, and thus cannot reject any of the thermal conductivity  
159 relationships. We take the average of the six scenarios to define our best estimate of the magnitude of the  
160 LGM-Holocene temperature change. We conservatively use the full range of the six scenarios ( $1.2^\circ\text{C}$ ) as  
161 our uncertainty estimate.

### 162 **S1.2.2 Uncertainty from parameter choices**

163 We performed additional optimizations to assess the influence of the vertical velocity profile; the results  
164 are given in Table S4 and Figure S4. The range of possible  $p$  values for the vertical velocity profile from  
165 the ApRES measurements (S1.1.3) is very similar to published estimates (39). We use  $p = 1$  and  $3.5$  as the  
166 lower and upper bounds and a preferred  $p$  value of  $2$ . A smaller  $p$  value, which produces a vertical velocity  
167 profile more like that beneath a stable divide where a Raymond Arch develops (48), yields a larger LGM-  
168 Holocene temperature change, while a larger  $p$  value (more flank-like flow) yields a smaller temperature  
169 change. The change in  $\Delta T_s$  upon varying  $p$  values within the specified range is around  $0.68^\circ\text{C}$  (Table S4).  
170 Note that surface accumulation rates are coupled to the vertical velocity profile via the thinning function.  
171 When we alter the velocity profile, we alter the surface accumulation rates along with it to ensure internal  
172 consistency. Therefore, the uncertainty in past accumulation rates is also included in our error estimates  
173 (see Fig. S9 for the range of LGM accumulation rates in the borehole study).

174 The impact of basal melting was evaluated by assuming no basal melt or  $1\text{ mm/yr}$  of basal melt instead of  
175  $0.5\text{ mm a}^{-1}$  based on Parrenin et al. (2007). No basal melt increases the magnitude of the LGM-Holocene  
176 temperature change compared to the base scenario, while increasing the basal melt does the reverse. The  
177 change in  $\Delta T_s$  upon varying melt rates within the specified range is around  $0.37^\circ\text{C}$  (Table S4).

178 We base our choice of thickness change on published simulated thickness histories (49, 50). The impact of  
179 deglacial ice sheet thickness changes was tested with alternative scenarios of either no thickness change, or  
180 doubling the amount of thickening between  $17$  and  $7\text{ ka}$  to  $200\text{m}$ , roughly spanning the range of responses  
181 seen in ice sheet model simulations (Section S3.3). Imposing no thickness change yields a  $0.1^\circ\text{C}$  larger  $\Delta T_s$ .  
182 Given the low sensitivity to the prescribed elevation, we did not evaluate specific inferences from individual  
183 ice sheet models.

### 184 **S1.2.3 Temporal Variations in the shape of the vertical velocity**

185 The shape of the vertical velocity profile is not necessarily fixed in time. A potential influence on the shape  
186 of the vertical velocity profile are changes in the temperature profile in the ice sheet; however, the low  
187 accumulation rates lead to heat transfer being dominated by conduction and thus only minor temperature  
188 variations deep in the ice sheet. Another potential influence is the development of a preferred crystal  
189 orientation fabric; however, the ice fabric at interior East Antarctic sites evolves slowly because of the low  
190 deformation rates (46) such that the evolution of the ice fabric is unlikely affect the vertical velocity pattern  
191 during the glacial termination.

192 Instead, the most likely reason for the shape of the vertical velocity profile to change is the position of the  
193 ice divide. As shown by the ApRES measurements and noted elsewhere (39), the shape of the vertical  
194 velocity profile has more of a quadratic character, consistent with the influence of lower deviatoric stresses  
195 near an ice divide (48), than the more linear shape predicted by Lliboutry (Eq. S2). This difference in shape  
196 of the vertical velocity profile beneath divides is what gives rise to so-called Raymond Arches commonly  
197 observed within coastal ice rises (51). No Raymond Arch is observed at EDC, likely due to the long  
198 characteristic time (100 ka) for which the divide position would need to be stable. The core and the ApRES  
199 measurements are near the ice divide today, such that any movement of the divide in the past would result  
200 in the more linear (larger  $p$  value) shape typical for sites away from a divide.

201 We evaluate the impact of a temporal variation in the vertical velocity shape function with a range of  
202 scenarios where we vary the onset time of the current vertical velocity profile ( $p = 2$  for these scenarios)  
203 and the shape of the previous vertical velocity profile. We allow the transition in vertical velocity shapes to  
204 occur in 1 ka, at three different ages in the Holocene: beginning at 10 ka, 6 ka, and 2 ka. The previous shape  
205 of the vertical velocity profile is not known, other than it should be more linear (more typical of a flank  
206 site). We use  $p = 4$ , based on the vertical velocity profiles outside of the region of divide influence at  
207 Roosevelt Island (38);  $p = 7$  based on recent measurements from Hercules Dome, another interior Antarctic  
208 dome site; and  $p = 15$  which approximates a shape profile with deformation concentrated in the warm ice  
209 near the bed. The inferred  $\Delta T_S$  for the nine scenarios is shown in Table S3; these scenarios use the Yen  
210 (1981) thermal conductivities and the standard forcing choices (S1.2); using the Waite et al., (2006) or Ross  
211 et al., (1978) thermal conductivities gives the same change in the inferred  $\Delta T_S$  to within hundredths of a  
212 degree (34, 44, 45).

213 The inferred reduction in  $\Delta T_S$  magnitude ranges from 0.44°C to 2.02°C with a mean reduction of 1.07°C.  
214 Because the shape of the vertical velocity profile is always more linear (or flank-like) in the past, the  
215 inferred  $\Delta T_S$  is always smaller in magnitude. As the change in  $\Delta T_S$  increases, the misfit to the borehole also  
216 increases. Thus, it is not clear that a variable vertical velocity shape function yields a more robust solution.  
217 Because we have no constraints on the shape of the vertical velocity profile for older ages and the timing  
218 of a transition, it is difficult to produce a quantitative estimate of the potential impact. Therefore, we do not  
219 directly include a temporally variable vertical velocity profile scenario in our “best estimate” independent  
220 borehole  $\Delta T_S$  estimate and uncertainty. However, the sensitivity test described in this section suggests an  
221 asymmetrical source of uncertainty in the EDC borehole reconstruction, which we represent via an  
222 estimated uncertainty range that is not centered around our best estimate of  $\Delta T_S = -5.5^\circ\text{C}$  (but rather around  
223  $\Delta T_S = -5.0^\circ\text{C}$ , which is halfway between our best estimate and the mean of the tests described in this section).  
224 We do note that the influence of time-variable  $p$  can reconcile the borehole- and  $\Delta\text{age}$ -based  $\Delta T_S$   
225 reconstructions.

#### 226 **S1.2.4 Additional temperature variability in borehole reconstructions**

227 As part of the borehole reconstruction, we performed a variety of optimizations with  $\alpha_T$  allowed to vary for  
228 different time periods. For example, in optimizations with two  $\alpha_T$  scalings (one for the Holocene and one  
229 for the remainder of the record which is dominated by the LGM-Holocene change), the Holocene scaling

230 was often very large (resulting in little temperature variability) while the glacial scaling remained similar  
231 to that found in the single-scaling scenarios. Since the Holocene temperature variations are likely small,  
232 they have diffused away in the borehole temperature profile and cannot be resolved. In addition, the near-  
233 surface temperatures were less reliable such that the upper 150m were not included in the optimization,  
234 which may also inhibit the ability to infer an independent Holocene scaling. This lack of high frequency  
235 temperature information is a primary challenge of using borehole temperatures from low accumulation sites  
236 compared to sites such as WAIS Divide (15). Only the broadest information of glacial temperature is  
237 preserved, and we cannot reconstruct high frequency variations in the temperature history (25, 52).

### 238 **S1.3 Results of temperature optimizations and uncertainty at Dome Fuji**

239 We have performed a similar analysis for Dome Fuji as for Dome C; however, there are no ApRES vertical  
240 velocity measurements at Dome Fuji, and temperature measurements have a five times higher uncertainty  
241 at 0.05K. In addition, the ages and ice flow near the bed are not easily modeled, adding uncertainty in the  
242 vertical strain pattern. In particular, Dome F has a reversal of the thinning function near the bed, which  
243 layers in the deepest few 100 m having experienced less cumulative thinning than layers above it. In  
244 addition, the DF core has inclined (up to 50°) internal layers in the deepest few 100 m, implying  
245 inhomogeneous bottom melting in the vicinity (53).

246 Therefore, the confidence in the inference of LGM-Holocene temperature change is lower than at Dome C.  
247 Our primary goal with the Dome Fuji borehole temperature measurements is to determine if they are  
248 consistent with the smaller  $\Delta T_S$  inferred with our firn-based reconstruction method.

249 We report two model runs that envelope a plausible range of the LGM-Holocene temperature change at  
250 Dome Fuji based on the information from Dome C and previous modeling of Dome Fuji (39). Many more  
251 runs were performed, but their  $\Delta T_S$  falls within this envelope. Using parameters for a small  $\Delta T_S$  – Yen  
252 (1981) conductivity and no increase in basal thermal conductivity, a vertical velocity with  $p=4$  and 0.3 mm  
253  $a^{-1}$  of basal melt – we obtain  $\Delta T_S = -1.8^\circ C$ . Using parameters for a large  $\Delta T_S$  – Waite et al. (2006) thermal  
254 conductivity with a basal increase up to 5%,  $p=1$ , and no basal melt – we obtain  $\Delta T_S = -5.4^\circ C$ . We thus  
255 report this range of  $-1.8^\circ C$  to  $-5.4^\circ C$  as the  $\Delta T_S$  that is consistent with the DF borehole data, without making  
256 an attempt to provide a central or best estimate within this range as we feel that the problem is insufficiently  
257 constrained to do so. The reader may be tempted to view the midpoint of the interval ( $\Delta T_S = -3.6^\circ C$ ) as a  
258 best estimate of DF cooling, but we do not endorse this interpretation.

259 The inferred DF  $\Delta T_S$  range overlaps well with the firn-based reconstruction (Figure 1). The borehole  
260 reconstruction suggests that  $\Delta T_S$  at DF was likely smaller than  $\Delta T_S$  at EDC, as also seen in the firn-based  
261 reconstruction and climate model simulations. Importantly, the DF borehole reconstruction is inconsistent  
262 with a large  $\Delta T_S$  of around  $-7.5$  to  $-8^\circ C$ , as suggested by traditional interpretation of the DF water isotope  
263 data (8, 43).

264

## 265 **S2 Firn-based temperature reconstructions**

### 266 **S2.1 Ice Core data**

#### 267 **S2.1.1 $\delta^{15}N$ - $N_2$ data**

268 Ice core  $\delta^{15}N$ - $N_2$  data used in this study are all available from published work (30, 54-56), with the  
269 exception of new Dome Fuji data first published with this paper (supplementary data). We combine ice  
270 sample data and lock-in zone firn air data (where available). Site  $\delta^{15}N$ - $N_2$  data are shown in Fig. 3C of the  
271 main text. All  $\delta^{15}N$  data use the modern atmosphere as the reference scale.

272 Unpublished Dome Fuji  $\delta^{15}\text{N}$  data are generated in three measurement campaigns. The first campaign (at  
273 Tohoku University, before 2001 (57)) covers 38 points from 563.3 to 992.9 m (LGM and older) with an  
274 uncertainty of 0.02 per mil (1 sigma). The second campaign (at Scripps Institution of Oceanography, 2007,  
275 (58)) covers 3 points from 462.5 to 478.5 m (Antarctic Cold Reversal) with uncertainty of 0.005 per mil.  
276 The third campaign (at the National Institute of Polar Research, 2016 - 2018) covers 293 points from 113.0  
277 to 558.0 m (Holocene and Termination I) with uncertainty of 0.006 per mil (manuscript for method in  
278 preparation). There are offsets between the different datasets, thus 0.020 per mil is added to the Tohoku  
279 dataset and 0.010 per mil is added to the SIO dataset to match them with the NIPR dataset where they are  
280 overlapped.

281 For the SDM site we furthermore use published  $\delta^{40}\text{Ar}$  data to calculate  $\delta^{15}\text{N}$ -excess to constrain abrupt  
282 warming at the 22ka event (59). An anomalous short-duration interval around 15.3 ka BP where SDM  $\delta^{15}\text{N}$   
283 approaches zero, thought to reflect a short-lived ablation event or catastrophic firn-break up (59), is  
284 removed from the SDM  $\delta^{15}\text{N}$  data set and not interpreted here.

### 285 **S2.1.2 CH<sub>4</sub> data**

286 Ice core atmospheric CH<sub>4</sub> mixing ratio data are used for ice core synchronization in the gas phase (Fig.  
287 S5). Site CH<sub>4</sub> data were previously published for all sites except DF (16, 42, 60-62).

288 Dome Fuji CH<sub>4</sub> concentrations are measured at Tohoku University and Laboratoire de Glaciologie et  
289 Géophysique de l'Environnement (LGGE, currently L'Institut des Géosciences de l'Environnement). The  
290 Tohoku University data are collected in the same campaign as the  $\delta^{15}\text{N}$  data using the method of ref (57)  
291 with analytical uncertainty of ~6 ppb (1 standard deviation). For the LGGE data, the air is extracted from  
292 the ice samples of ~50 g using an established melt-refreeze method, and analyzed by a gas chromatograph  
293 using a standard gas (499 ppb) for calibration, and an analytical uncertainty of 10 ppb (1 standard deviation)  
294 (60, 63). Before the synchronization, the LGGE dataset is shifted by +18 ppb to account for systematic  
295 offset in calibration scale relative to the Tohoku University scale; note that we use CH<sub>4</sub> only to constrain  
296 the event timing, and therefore the absolute concentrations are not important.

297 The SDM data used here are a compilation of previously published data (62, 64-68) measured at Oregon  
298 State University and Seoul National University using the same methods described in refs. (64-67).

299 The full DF and SDM CH<sub>4</sub> records are provided in the data supplement.

### 300 **S2.1.3 Water isotope data**

301 Water isotopic  $\delta^{18}\text{O}$  (rather than  $\delta\text{D}$ ) data are used at all sites, reported on the Vienna standard mean ocean  
302 water (V-SMOW) reference scale. These data are publicly available (3, 43, 64, 69-75). The ice core  $\delta^{18}\text{O}$   
303 data are corrected for the mean ocean  $\delta^{18}\text{O}$  using ref. (76). The high-frequency structure in the firn-based  
304 reconstructions (Fig. S8) comes from the  $\delta^{18}\text{O}$  data, while our firn-based reconstruction method only  
305 constrains magnitude of the long-term temperature change.

### 306 **S2.1.4 Siple Dome volcanic ties**

307 Volcanic tie points between the Siple Dome and WAIS Divide ice cores are given in the supplementary  
308 data files. We identify a total of 18 such volcanic ties based on established techniques: nine ties are found  
309 via matching Electrical Conductivity Measurement (ECM) records of volcanic activity (55, 77), and nine  
310 ties via the geochemical identification of tephras (78, 79). The volcanic tie points are shown on Fig S5. The  
311 oldest of these volcanic tie points is from 32ka BP, and thus the LGM and deglaciation are covered. Beyond  
312 32ka we identify several ice-ice ties based on  $\delta^{18}\text{O}_{\text{ice}}$  to extend the WD2014 chronology further back – these  
313 tie points have no impact on the  $\Delta T_s$  presented here.

## 314 **S2.2 Empirical $\Delta$ age reconstruction**

315 For all sites other than WD, we estimate  $\Delta$ age empirically by combing ice-phase volcanic matching and

316 gas-phase atmospheric methane ( $\text{CH}_4$ ) matching to WD (16). Site  $\Delta$ age estimates are shown in Fig. 3B of  
317 the main text. WD was chosen for this purpose because of the high resolution  $\text{CH}_4$  record (80), and small  
318  $\Delta$ age at the site (30). The matching procedure is shown in Fig. S5, where we show volcanic ties (dots) and  
319 the site  $\text{CH}_4$  records plotted on top of the WD  $\text{CH}_4$  record. Volcanic ties to WD are published for all sites  
320 except SDM (40, 42, 81-83); SDM ties are provided in the data supplement and described below.

321 Conspicuous  $\text{CH}_4$  features – such as Dansgaard-Oeschger (DO) oscillations and the onset of the deglacial  
322  $\text{CH}_4$  rise – are matched between the cores, thereby assigning a WD2014 gas age to the core being matched;  
323 the WD2014 ice age at that same depth is found from the volcanically synchronized time scales (40),  
324 allowing the gas age-ice age difference to be calculated. The sources of  $\Delta$ age uncertainty are: (1) uncertainty  
325 in the WD2014  $\Delta$ age; (2) uncertainty in the WD  $\text{CH}_4$  tie point determination; (3) uncertainty in the matched  
326 core  $\text{CH}_4$  tie point determination (for example due to the finite time resolution of the  $\text{CH}_4$  record); (4)  
327 uncertainty in age scale interpolation between volcanic tie points; these 4 terms are added in quadrature to  
328 estimate the  $\Delta$ age uncertainty. Note that WD has the smallest  $\Delta$ age of all cores considered here. The stated  
329  $2\sigma$  WD  $\Delta$ age uncertainty is 120 years during the LGM; which is a 23% relative uncertainty. However, at  
330 EDC the maximum  $\Delta$ age during the LGM is around 5000 years, in which case the same 120-year  
331 uncertainty only represents a relative uncertainty of 2.4%.

332 Figure S6 compares our empirical  $\Delta$ age estimates (dots with errorbars) to the  $\Delta$ age in the AICC2012 ice  
333 core chronology (grey curves) for the cores where this is available (TAL, EDML and EDC), as well as the  
334  $\Delta$ age simulated in this study using a firn densification model (colored curves). We find good agreement.

335 The approach outlined above provides WD2014-consistent  $\Delta$ age estimates for all cores except WD; to apply  
336 our firn-based approach to WD we have to generate independent empirical  $\Delta$ age estimates. To do so we  
337 use published bipolar volcanic markers for Holocene and glacial period (42, 84). We match the WD  $\text{CH}_4$   
338 record to the NGRIP  $\delta^{18}\text{O}$  at the midpoint of abrupt DO transitions, assuming a 25-year lag of atmospheric  
339  $\text{CH}_4$  behind Greenland climate (30, 85, 86); this provides a GICC05 gas age for each matched event, which  
340 is combined with a GICC05 ice age at the same depth (found from the bipolar volcanic synchronization) to  
341 calculate empirical  $\Delta$ age WD estimates. The uncertainty in these  $\Delta$ age markers is calculated as above for  
342 the other cores. The most prominent  $\text{CH}_4$  feature around DO-2 is actually associated with Heinrich event 2  
343 rather than with DO-2 (80). A comparison of the WD  $\text{CH}_4$  record to GISP2  $\text{CH}_4$  identifies the location of  
344 the DO-2  $\text{CH}_4$  feature that is concurrent with the DO-2 thermal  $\delta^{15}\text{N-N}_2$  fractionation signal (Fig S7B); this  
345 allows us to establish an empirical WD  $\Delta$ age constraint around this time by matching the WD  $\text{CH}_4$  feature  
346 to Greenland records on the GICC05 time scale. Figure S7A compares the WD2014 and GICC05  $\Delta$ age  
347 reconstructions for WD; during the glacial the WD2014  $\Delta$ age is around 70 years smaller on average (84).

348 The differences between the two  $\Delta$ age estimates for WD can be transferred to all the cores in this study,  
349 and thus we have two sets of empirical  $\Delta$ age estimates for each core: one consistent with the WD2014  
350 chronology, and one with the GICC05 chronology. The influence of the choice of base chronology on  $\Delta T_s$   
351 is estimated in the Monte-Carlo procedure (see below). The  $\text{CH}_4$  match points, and the empirical WD2014  
352 and GICC05  $\Delta$ age estimates (with uncertainty) derived from it, are provided in the data supplement.

353 For the SDM site we furthermore derive a  $\Delta$ age constraint from the abrupt 22ka warming anomaly, which  
354 is seen in both the ice phase ( $\delta^{18}\text{O}_{\text{ice}}$ ) and gas phase ( $\delta^{15}\text{N-N}_2$  and  $\delta^{40}\text{Ar}$ ).

355 In the remainder we will refer to the empirical  $\Delta$ age constraints as “ $\Delta$ age data” for the sake of brevity.

### 356 S2.3 Firn densification modeling

357 The rate of firn densification is sensitive to the temperature and the overburden pressure – with the latter  
358 determined by the time-integrated snow accumulation rate since the deposition of a layer. The time-variably  
359 surface temperature  $T_s(t)$  and accumulation rate  $A(t)$  histories are the primary controls on firn densification

360 rates, and firn densification models can be used to simulate the evolution of the firn layer through time (17,  
 361 20, 87-91). The  $\delta^{15}\text{N}-\text{N}_2$  and  $\Delta\text{age}$  become fixed at the lock-in depth, providing observational constraints  
 362 on past densification rates (15, 88, 92). Critically, the  $\delta^{15}\text{N}$ - and  $\Delta\text{age}$ -isopleths run perpendicular to each  
 363 other in  $T_S, A$ -space (Main text Fig. 3A), meaning that if  $\delta^{15}\text{N}$ - and  $\Delta\text{age}$  are both independently known, a  
 364 unique temperature and accumulation solutions exist (20); however such solutions are obviously subject to  
 365 any biases present in the firn densification physics used.

366 To simulate past Antarctic firn evolution, we use a dynamical firn densification – heat transport model  
 367 described elsewhere in the literature (17, 30, 93, 94). For the firn densification physics we use the  
 368 overburden-pressure formulation of the Herron-Langway firn model; given by Equation 4c in Ref. (18).  
 369 The model can be run in an inverse mode, where an automated routine is used to find the  $T_S(t)$  and  $A(t)$   
 370 solutions that optimize the fit to the  $\delta^{15}\text{N}$  and  $\Delta\text{age}$  data; this approach was previously applied to Greenland  
 371 records (17).

372 The inverse model adjusts the  $T_S(t)$  and  $A(t)$  input to minimize the cost function:

$$373 \quad S = \sqrt{\frac{1}{N_\delta} \sum_i \frac{(d_i - m_i)^2}{u_i^2}} + \sqrt{\frac{1}{N_\Delta} \sum_i \frac{(D_i - M_i)^2}{U_i^2}} \quad (\text{S7})$$

374 where  $d_i$  ( $D_i$ ) are the  $\delta^{15}\text{N}$  ( $\Delta\text{age}$ ) data,  $m_i$  ( $M_i$ ) the interpolated modeled values at the same depth,  $u_i$  ( $U_i$ )  
 375 the data uncertainty, and  $N_\delta$  ( $N_\Delta$ ) the total number of data. The climate forcings used in the modeling  
 376 procedure are described by:

$$377 \quad T_S(t) = T_{\text{init}}(t) + f_T(t) \quad (\text{S8})$$

$$378 \quad A(t) = A_{\text{init}}(t) \times [1 + f_A(t)] \quad (\text{S9})$$

379 where  $T_{\text{init}}$  and  $A_{\text{init}}$  are the initial temperature and accumulation estimates;  $f_T$  and  $f_A$  are functions to correct  
 380 our initial estimate, which are being optimized in the inverse modeling procedure.  $T_{\text{init}}$  is based on linear  
 381 scaling of the site  $\delta^{18}\text{O}$  (corrected for mean-ocean  $\delta^{18}\text{O}$ ), using the isotope sensitivity constant  $\alpha_{\text{init}}$  listed in  
 382 table S5. The  $\alpha_{\text{init}}$  values are chosen to be intermediate between the spatial slope  $\alpha_S$ , and the LGM-PI  $\alpha_T$   
 383 values reconstructed here (main text, Fig. 1). The reconstructed  $\Delta T_S$  is independent of the choice of  $\alpha_{\text{init}}$  in  
 384 the 0.4 to 2.8 ‰  $\text{K}^{-1}$  range that we have tested; the values used were found iteratively in two steps after a  
 385 first attempt using the spatial slope isotope sensitivity as  $\alpha_{\text{init}}$ . The choice of  $\alpha_{\text{init}}$  mainly dictates the  
 386 magnitude of high-frequency (decadal to millennial) temperature variability in the final reconstruction.  
 387 Where available,  $A_{\text{init}}$  is based on de-strained layer-counted annual layer thickness data; this is only the  
 388 Holocene at SP, and the last 31ka at WD. Elsewhere,  $A_{\text{init}}$  is likewise based on linear scaling of the water  
 389 isotope record. All six variables in Eqs. (S8) and (S9) are plotted in Fig. S8.

390 The functions  $f_T$  and  $f_A$  are each defined using a series of control points  $a_i$  (white dots in Fig. S8), such that  
 391  $f(t_i) = a_i$ . At times in between the control points, the value of  $f$  is found via linear interpolation between the  
 392 two adjacent control points. Because modern-day climatic conditions at the sites are known, we furthermore  
 393 let  $f(t=0) = 0$ . Because our interest is in reconstructing  $\Delta T_S$ , we decided to keep  $f_T$  (and in many cases  $f_A$ )  
 394 constant through the duration of the LGM. This is implemented by requiring two adjacent control points to  
 395 have the same value ( $a_{i+1} = a_i$ ). This is indicated by thick grey horizontal bar connecting the two control  
 396 points in Fig. S8. The values for  $a_i$  that minimize the cost function Eq. (S8) are found in an automated  
 397 gradient method (17).

398 The timing of the control points was chosen to coincide with climatic change points in the  $\delta^{18}\text{O}_{\text{ice}}$  records;  
 399 for the deglaciation these are the onset of Heinrich Stadial (17.8 ka), the onset and termination of the  
 400 Antarctic Cold Reversal (14.7ka and 12.8 ka), and the Holocene onset (11.6 ka). In other cases large features  
 401 in the  $\delta^{15}\text{N}$  record were used to select control points. Note that the number and timing of the control points

402 is different for each of the cores, as they were adjusted based on the unique characteristics and data  
403 availability at each of the cores. In general, cores with more data ( $\delta^{15}\text{N}$  and  $\Delta\text{age}$ ) have a greater number of  
404 control points.

405 As a first example of why certain control points were selected, we examine EDC and TAL where no  $\delta^{15}\text{N}$   
406 data are available older than  $\sim 23$  ka BP. This means the  $T_s(t)$  and  $A(t)$  solutions are no longer uniquely  
407 constrained at those ages, and therefore we require  $f_T$  to have a constant value for all  $t$  older than 18ka BP  
408 (grey horizontal bars in Fig S8a and S8e). The model fits the EDC and TAL  $\Delta\text{age}$  observations prior to the  
409 LGM through making adjustments to  $f_A$  instead (keeping  $f_T$  constant).

410 As a second example, consider the SDM core, which has a large upward shift in  $\delta^{18}\text{O}$  around 21.4 ka BP  
411 possibly driven by ice dynamics (59). We use control points on either side of the shift in both  $f_T$  and  $f_A$   
412 to allow the model to adjust both the temperature and accumulation change across this transition. The  
413 transition is characterized by a large drop in  $\delta^{15}\text{N}$  (the single largest feature in all  $\delta^{15}\text{N}$  considered here),  
414 and both warming and a reduction in accumulation rate are needed to fit this feature (Fig. S8g).

415 As a third example, consider the SP core. Due to its flank-flow configuration the accumulation rate is much  
416 more variable than it would be at a dome, as the deposition site moves across spatial gradients in surface  
417 accumulation (55, 95). For this reason, SP requires more accumulation control points than the other cores  
418 do in order to fit the SP  $\delta^{15}\text{N}$  data (Fig. S8d).

419 For the purpose of estimating  $\Delta T_s$  the control points used during the deglaciation are most important. For  
420 all sites we conducted sensitivity studies in which the number of control points during the deglaciation is  
421 varied, and the changes in reconstructed  $\Delta T_s$  are well within the spread estimated from the Monte Carlo  
422 uncertainty approach (Section S2.4). Therefore, we do not report on these experiments separately.

423 Besides the  $T_s(t)$  and  $A(t)$  histories, the densification model requires other site-specific, user-defined  
424 parameters (Table S5). The convective zone (CZ) thickness is estimated from firm air sampling data at all  
425 sites except TAL where such data is not available and we use a generic value of 5m. The surface snow  
426 density  $\rho_0$  is estimated from density data at the sites. Following established methods (96), the lock-in density  
427 is estimated by  $\rho_{\text{LI}} = \rho_{\text{CO}} - \rho_{\text{diff}}$ , with the temperature-dependent close-off density  $\rho_{\text{CO}}$  given by the  
428 Martinerie equation (97), and  $\rho_{\text{diff}}$  a site-dependent lock-in zone thickness. We estimate  $\rho_{\text{diff}}$  from firm  
429 density and firm air sampling data, and fine-tune the value to best fit modern-day  $\Delta\text{age}$  and  $\delta^{15}\text{N}$  at the site.  
430 It is established that lock-in zone thickness is proportional to site accumulation (98), and indeed we find  
431 low  $\rho_{\text{diff}}$  values at the low-accumulation EDC and DF sites ( $4 \text{ kg m}^{-3}$  and  $3 \text{ kg m}^{-3}$ , respectively) and higher  
432 values at the high-accumulation WD site ( $10 \text{ kg m}^{-3}$ ). The recommended value for Greenland Summit is  $14$   
433  $\text{kg m}^{-3}$  (96). Intermediate-accumulation site SP has a thick lock-in zone, and we find an optimal  $\rho_{\text{diff}}$  of  $15$   
434  $\text{kg m}^{-3}$ . The ice sheet thickness  $H$  is taken as reported in the literature, and the site geothermal heat flux  
435 (GHF) is estimated by fitting the borehole temperature profile – note that the GHF estimation given in Table  
436 S5 is not optimized in any way, and therefore we recommend against interpreting these numbers. The GHF  
437 estimates in table S5 are lower bounds to the true GHF, because the effect of ice melting at the bed is not  
438 taken into account.

439 It has been hypothesized that dust (or perhaps calcium or fluorine/chlorine) may soften the ice, enhancing  
440 densification rates (99-102). The effect is not included here, see section S2.5 for a justification.

441 The optimal solutions presented in the main manuscript are derived using the dynamical description of the  
442 Herron-Langway (HL) densification model (18). However, other physical equations of firm densification  
443 are available, and we want to assess how the choice of model influences the result (Fig. S9). We compare  
444 the range of our dynamical HL Monte Carlo simulation (histogram, see also section S2.3 below) to results  
445 from running the Arnaud (88, 103), Barnola (20, 87) and Bréant (89) firm densification models in the inverse  
446 mode; we focus on the WD, EDC and DF sites that have borehole thermometry estimates (dots with error

447 bars). For all models we use the exact same experimental design: the same present-day  $T_S$  and  $A$ ; the same  
448  $T_{\text{init}}$  and  $A_{\text{init}}$ ; the same  $\Delta\text{age}$  and  $\delta^{15}\text{N}$  data; and the same  $T_S$  and  $A$  control points. We adjusted the  $\rho_{\text{diff}}$  model  
449 parameter for each model to obtain a good fit to the present-day  $\Delta\text{age}$  and  $\delta^{15}\text{N}$  data, with the values given  
450 in Table S6. The model-specific  $\rho_{\text{diff}}$  adjustments are similar at all sites, suggesting they reflect model biases  
451 that are stable across a range of climatic conditions.

452 At WD, all models surveyed find a  $\Delta T_S$  at WD in the -9.8 to -10.6°C range, consistent with the borehole  
453 reconstruction. At EDC the HL and Bréant models find a  $\Delta T_S$  that agrees with the borehole within  
454 uncertainty; the Arnaud and Barnola models reconstruct a somewhat small  $\Delta T_S$ . At DF all models are in  
455 agreement within the borehole uncertainty range; yet again the Barnola and Arnaud models find smaller  
456  $\Delta T_S$ . We find at both cold sites that the HL and Bréant models reconstructs a larger  $\Delta T_S$  than the Arnaud  
457 and Barnola models do, which is consistent with the fact that the former models have a lower effective  
458 activation energy than the latter models (Fig. S9, right panel), meaning that it requires a larger temperature  
459 change to induce the same change in densification rates. The Bréant model has a temperature-dependent  
460 effective activation energy (implemented as the sum of three Arrhenius terms) and is a modification of the  
461 Arnaud model (89); note that the Bréant model provides a much better fit to the borehole estimates than the  
462 Arnaud model does suggesting it is indeed an improvement over the Arnaud model.

463 Past accumulation is well constrained in our method – better in fact than the temperature is (compare the  
464  $T_S$  and  $A$  envelopes in Fig. S8). The reason is an elemental physical one. The ice-equivalent lock-in depth  
465 (LIDIE), accumulation rate, and  $\Delta\text{age}$  are linked via the simple equation  $A = \text{LIDIE} / \Delta\text{age}$ . LIDIE is a  
466 scaled version of the lock-in depth (LID), and the scaling (of around 0.7) is very stable across a wide range  
467 of climatic conditions (104). Because in our method both  $\Delta\text{age}$  and the LID (from  $\delta^{15}\text{N}$ ) are known, the  
468 accumulation rate is constrained very strongly. This is clearly visible in Fig. S9, where the four densification  
469 models find almost identical LGM accumulation rates. Uncertainty in  $\delta^{15}\text{N}$ ,  $\Delta\text{age}$ ,  $\rho_0$ ,  $\rho_{\text{diff}}$  and CZ does  
470 impact the reconstructed  $A$  slightly, which is investigated in the Monte-Carlo study (S2.4). The  $A = \text{LIDIE}$   
471 /  $\Delta\text{age}$  relationship is so fundamental that is independent of the firm densification model used to first order  
472 – the firm model choice only controls the scaling between LIDIE and LID, which have a ratio very close to  
473 0.7 in all models and all climates (104).

474 This comparison suggests that the dynamical HL and Bréant models provide results in good agreement with  
475 the borehole temperature reconstructions over a large temperature range, and should therefore be given  
476 preference in simulating Antarctic firm dynamics at cold locations.

## 477 S2.4 Monte-Carlo uncertainty estimation

478 To estimate the uncertainty in our  $\Delta T_S$  estimates we use a Monte-Carlo (MC) approach in which we  
479 randomly disturb the  $\delta^{15}\text{N}$  and  $\Delta\text{age}$  data, and draw the user-defined model parameters randomly from a  
480 prescribed distribution (Table S5). For 1000 such iterations we perform the model inversion to estimate the  
481  $T_S(t)$  and  $A(t)$ , allowing us to describe the uncertainty in  $\Delta T_S$ . Because inversion of the full dynamical HL  
482 model is computationally expensive, we instead use the steady-state version of the HL model for the MC  
483 uncertainty estimation. The procedure is identical to the dynamical model described above and uses Eqs.  
484 (S7) to (S9) to solve for  $T_S(t)$  and  $A(t)$ . The steady-state model does not calculate the temperature profile,  
485 and we use a constant firm temperature gradient to account for thermal  $\delta^{15}\text{N}$  fractionation by the GHF.

486 In each of the 1000 MC iterations we randomly draw each of the  $\delta^{15}\text{N}$  data points from a normal distribution  
487 with a mean equal to the observation, and a standard deviation equal to the uncertainty in that observation.  
488 The  $\Delta\text{age}$  data points are each perturbed in two steps. In a first step we draw a random number  $c$  from a  
489 uniform distribution between 0 and 1; all the  $\Delta\text{age}$  data points  $D_i$  for that iteration are set to a weighted sum  
490 of the WD2014 and GICC'05-based  $\Delta\text{age}$  constraints (Fig. S7):  $D_i = c \times D_i^{\text{WD2014}} + (1 - c) \times D_i^{\text{GICC'05}}$ .  
491 In the second step, we add a random perturbation to each of the  $D_i$  that is drawn from a normal distribution

492 of zero mean and a standard deviation equal to the uncertainty ( $U_i$ ). The  $\rho_0$  and  $\rho_{\text{diff}}$  parameters are drawn  
493 from a normal distribution with a mean and standard deviation listed in Table S5. It has been suggested that  
494 the convective zone may be climate dependent, and possibly thicker during the LGM. The convective zone  
495 thickness is therefore described as the sum of two parts: 1) a climate-independent part that is randomly  
496 drawn from a normal distribution with a mean and standard deviation as listed in Table S5 (in case of a  
497 negative value, we re-draw until a non-negative value is obtained); 2) a climate-dependent part that is equal  
498 to a scaled version of the site  $\delta^{18}\text{O}$  record (such that it equals zero in the PI and one in the LGM) multiplied  
499 by a random number drawn from a uniform distribution between 0 and 5 (the LGM CZ is thicker or equal  
500 to the PI CZ).

501 The bias of the steady-state approach is calculated by taking the difference between the  $T_s(t)$  and  $A(t)$   
502 solutions found in the dynamical HL inversion and the steady-state HL inversion, where both use the  
503 preferred parameter settings; all steady-state HL solutions in the MC study were corrected for this bias. In  
504 Figure 1 of the main text we report the mean and  $2\sigma$  standard deviation of the  $\Delta T_s$  distribution found in the  
505 MC study – this differs from the dynamical HL solutions by  $0.08^\circ\text{C}$  on average.

## 506 **S2.5 The glacial $\delta^{15}\text{N}$ data-model mismatch in previous studies**

507 For several of the East Antarctic sites (EDC, DF, SP, EDML; see Fig. 3 of the main text) the firn thickness  
508 during the glacial period as indicated by  $\delta^{15}\text{N}$  is thinner than it is at present. However, densification models,  
509 when forced with a  $\Delta T_s$  of around  $-9^\circ\text{C}$ , simulate a thicker glacial firn column (increased  $\delta^{15}\text{N}$ ) than at  
510 present. This has led to the notion that there is a  $\delta^{15}\text{N}$  data-model mismatch in the glacial period in East  
511 Antarctica (56, 89, 105, 106). Note however, that this behavior is not observed for the West Antarctic WD  
512 and SDM cores that have increased glacial  $\delta^{15}\text{N}$ , nor at the TAL site (based on the limited available  $\delta^{15}\text{N}$   
513 data). Moreover, densification models have been very successful at simulating Greenland firn evolution  
514 during the glacial time (17, 20, 90, 91, 107). Barring anomalous situations (such as surface melt and shear  
515 zones of ice streams), we are not aware of a single study in which firn models fail to simulate the basic  
516 properties of firn of interest here (meter-scale density, lock-in depth,  $\Delta\text{age}$ ) within a reasonable error  
517 margin.

518 One hypothesis for the low glacial  $\delta^{15}\text{N}$  is a thickened CZ during the LGM (105, 106). To explain the low  
519 glacial  $\delta^{15}\text{N}$  requires up to 40m of CZ thickness during the LGM at DF (108). Such deep convective mixing  
520 has not been observed in present-day Antarctica; the deepest documented CZ occurs at the extreme  
521 Antarctic Megadunes site, a zero-accumulation site where deep thermal cracks act as conduits for air mixing  
522 down to  $\sim 23$  m depth (108). Glacial records of chemistry, water isotopes, volcanic deposition and  
523 atmospheric composition of the cores used in this study all provide evidence for continuous and hiatus-free  
524 ice accumulation through the glacial period, precluding the kind of conditions that drive deep mixing at  
525 Megadunes. Also, low-accumulation dome sites (like e.g. Dome C) today tend to have thin rather than thick  
526 convective zones. Overall there is limited support or evidence for this hypothesis (105).

527 Another proposed solution to the glacial  $\delta^{15}\text{N}$  mismatch is linked to the hypothetical influence of dust (or  
528 perhaps calcium or fluorine/chlorine) in softening the firn, thereby enhancing densification rates (99-102).  
529 In this hypothesis, the high dust loading of glacial ice enhances densification rates, thereby thinning the firn  
530 column as reflected in low glacial  $\delta^{15}\text{N}$ . At the WD site, where past  $T_s(t)$  and  $A(t)$  are well-constrained by  
531 borehole thermometry and annual-layer counting respectively, it was found that including the effect of dust  
532 softening only acted to deteriorate the fit to observational data (30). Recently, a multi-site study found that  
533 while including dust softening improved the model-data agreement at EDC and EDML, it worsened the  
534 agreement at WD and the Greenland NGRIP site (89). Generally, in Greenland cores, where dust loading  
535 is an order of magnitude larger than in Antarctica, densification models are successful in simulating firn  
536 evolution without taking the hypothesized softening effect of dust into account (17, 20, 90, 91, 107). If the  
537 hypothesized dust softening effect were true, it should work at all locations and time periods; since it does  
538 not, we believe this hypothesis can be eliminated.

539 LGM climate in interior East Antarctica has no modern analogues, and therefore falls outside the calibration  
540 range of densification models – perhaps casting doubt on their ability to simulate such climates (note that  
541 this cannot be a complete solution to the “glacial  $\delta^{15}\text{N}$  problem”, because EDML, SP and Law Dome all  
542 exhibit low glacial  $\delta^{15}\text{N}$ , yet remain within the densification model calibration range). Bréant et al. provide  
543 an interesting variation on this idea, by suggesting that firn densification models are too sensitive to  
544 temperature at cold conditions (89). They re-tune the Arnaud model to reduce its sensitivity at low  
545 temperatures, thereby improving the fit to glacial  $\delta^{15}\text{N}$  data. There are two important caveats: (1) they  
546 invoke a hypothetical densification process with an activation energy of 1.5 kJ/mol – at least 10 times  
547 smaller than any known densification or vapor movement processes in firn; (2) the model (when forced  
548 with  $\Delta T_s = \sim -9^\circ\text{C}$ ) still requires dust softening to satisfactorily fit  $\delta^{15}\text{N}$  data at EDML and EDC (89), as  
549 well as at DF (Anaïs Orsi, personal communication 2019). In our analysis, the performance of the Bréant  
550 model at low temperatures is very comparable to the HL model (Fig. S9).

551 Our work suggests an alternative solution to the glacial  $\delta^{15}\text{N}$  data-model mismatch: all previous firn  
552 modeling work has overestimated glacial cooling in East Antarctica. We find that the HL densification  
553 model can fit  $\Delta\text{age}$  and  $\delta^{15}\text{N}$  data for all Antarctic core sites using  $T_s(t)$  forcing consistent with the borehole  
554 temperature profiles, and  $A(t)$  forcing consistent with de-strained layer thickness. The spatial  $\Delta T_s$  pattern  
555 we find correlates well ( $r=0.90$ ) with that in GCM simulations using realistic Antarctic LGM topography.

556 Our work uses the HL densification model, which is calibrated over a temperature range from  $-15^\circ\text{C}$  to  
557  $-57^\circ\text{C}$  (18). The lowest LGM temperatures in our reconstructions are  $-60^\circ\text{C}$  (Dome F), which is just outside  
558 the HL calibration range. However, for the two coldest sites (EDC and DF) our firn-based  $\Delta T_s$  is in good  
559 agreement with the borehole temperature profile (main text Fig. 2), suggesting the model performs  
560 adequately under such cold climatic conditions. Moreover,  $\Delta T_s$  at the cold DF site is very similar to  $\Delta T_s$   
561 at the nearby EDML site, where LGM temperatures of around  $-50^\circ\text{C}$  are well within the HL calibration range.

562

### 563 **S3 Climate model simulations**

#### 564 **S3.1 HadCM3 model**

565 We use the fully coupled ocean-atmosphere model HadCM3B M1 (109, 110). This model has a long history  
566 of use in paleoclimate simulations, more recently, has been found very useful for sensitivity testing past  
567 climates (111-114). Briefly, the atmosphere model has a horizontal resolution of  $96\times 73$  gridpoints  
568 ( $3.75^\circ$  longitude  $\times 2.5^\circ$  latitude) with 19 hybrid levels (sigma levels near the surface, changing smoothly to  
569 pressure levels near the top of the atmosphere). The ocean component has a horizontal resolution of  
570  $288\times 144$  grid points ( $1.25^\circ\times 1.25^\circ$ ) and in the vertical there are 20 depth levels.

571 The HadCM3 model simulations presented here are extended from a more than 5000 year long LGM  
572 simulation of the model using the PMIP2 protocol (111, 115). This is broadly similar to the PMIP3 (116)  
573 protocol (GHG, land sea mask, vegetation) but differs in the ice sheet reconstruction: PMIP2 uses ice5G,  
574 PMIP3 uses a composite (116). In our simulations, since we vary the Antarctic topography, the difference  
575 between the PMIP2 and PMIP3 lies only in the topography of the Northern Hemisphere ice sheets. HadCM3  
576 has a climate sensitivity (global warming in response to doubling of  $\text{CO}_2$ ) of around  $3.5^\circ\text{C}$ ; the LGM  
577 simulation has an Atlantic Meridional Overturning Circulation (AMOC) strength of  $\sim 17.2$  Sv.

578 All sensitivity simulations (section S3.3) are run for an additional 500 years (on top of the LGM spin-up)  
579 with the analysis performed on years 50-150. There are negligible differences between results for any 100  
580 year period after an initial 50 year spin up.

581 LGM surface atmosphere cooling in HadCM3 is  $-5.4^\circ\text{C}$  in the global mean;  $-3.2^\circ\text{C}$  in the tropics ( $30^\circ\text{S}$ -  
582  $30^\circ\text{N}$ );  $-10.6^\circ\text{C}$  in the northern hemisphere extratropics, and  $-5.0^\circ\text{C}$  in the southern hemisphere extratropics.

583 Northern hemisphere extratropical surface cooling is much greater than other zonal bands due to the large  
584 albedo and lapse-rate forcing of the large NH ice sheets (Laurentide and Fennoscandian).

### 585 **S3.2 MIROC model**

586 We further perform numerical experiments with the Model for Interdisciplinary Research on Climate 4m  
587 (MIROC4m) AOGCM (117). This model consists of an atmospheric general circulation model (AGCM)  
588 and an oceanic general circulation model (OGCM). The AGCM solves the primitive equations on a sphere  
589 using a spectral method. The horizontal resolution of the atmospheric model is  $\sim 2.8^\circ$  and there are 20 layers  
590 in the vertical. The OGCM solves the primitive equation on a sphere, where the Boussinesq and hydrostatic  
591 approximations are adopted. The horizontal resolution is  $\sim 1.4^\circ$  in longitude and  $0.56^\circ$  to  $1.4^\circ$  in latitude  
592 (latitudinal resolution is finer near the equator). There are 43 layers in the vertical. Note that the coefficient  
593 of horizontal diffusion of the isopycnal layer thickness in the OGCM is slightly increased to  $700 \text{ m}^2\text{s}^{-1}$   
594 compared with the original model version ( $300 \text{ m}^2\text{s}^{-1}$ ) that was submitted to PMIP2. The current model  
595 version has been used extensively for modern climate, paleoclimate (118), and future climate studies (119).  
596 The climate sensitivity of this model is  $4.3^\circ\text{C}$  (119); the model has an LGM AMOC strength of  $\sim 7 \text{ Sv}$ , a  
597 shoaled mode of northern source deep water in the Atlantic and expanded southern source deep water, in  
598 good agreement with reconstructions based on North-Atlantic  $\delta^{13}\text{C}$  and  $\Delta^{14}\text{C}$  marine sediment data (120).

599 Ice sheet topography sensitivity experiments are initiated from a previous LGM experiment (53, 120),  
600 which is forced with PMIP3 boundary conditions (116). Based on this original experiment, the topography  
601 of the Antarctic ice sheet is modified following multiple reconstructions (Section S3.3); the extent of  
602 Antarctic ice sheet is unchanged from the original PMIP3 LGM experiment. These sensitivity experiments  
603 are integrated for 1000 years and the climatology of year 401-500 is used for analysis. The main result does  
604 not depend on the choice of the period used for the analysis.

605 LGM cooling in MIROC and HadCM3 is respectively  $-5.2^\circ\text{C}$  in the global mean;  $-2.7^\circ\text{C}$  in the tropics  
606 ( $30^\circ\text{S}$ - $30^\circ\text{N}$ );  $-12.7^\circ\text{C}$  in the northern hemisphere extratropics, and  $-3.1^\circ\text{C}$  in the southern hemisphere  
607 extratropics. Northern hemisphere extratropical surface cooling is much greater than other zonal bands due  
608 to the large albedo and lapse-rate forcing of the large NH ice sheets (Laurentide and Fennoscandian).

### 609 **S3.3 Ice sheet topography sensitivity experiments**

610 We perform a series of sensitivity experiments with both models in which we vary the shape of the LGM  
611 ice sheet over Antarctica using a variety of reconstructions. For all of these LGM ice sheets we vary the  
612 topography but not the ice sheet extent (or ice mask), to ensure that we are only investigating the  
613 contribution of topography and not the albedo contribution to regional surface temperatures. For some  
614 reconstructions this results in a less extensive ice sheet in Antarctica, for others a more extensive ice sheet.  
615 In these latter cases the additional ice is set to 5m elevation. When adjusting the Antarctic ice sheet we  
616 follow the PMIP protocol of adding the LGM to preindustrial elevation anomaly to the models preindustrial  
617 topography. Since the PI topography in some reconstructions used is not the same as preindustrial  
618 topography in the climate model, this means that the absolute LGM topography in these cases differs  
619 between the climate model and the original reconstruction; the LGM-PI ice anomaly is the same as in the  
620 original publications, however.

621 To isolate the climatic and topographic contributions to LGM cooling, we furthermore run an LGM  
622 simulation in which we use full LGM boundary conditions (including LGM Antarctic ice mask), but PI  
623 topography in Antarctica. In this simulation we keep the ice sheet elevation constant relative to the modern  
624 geoid; since the LGM sea level was 120 m lower than during the PI, this means that in this simulation the  
625 ice sheet elevation is 120 higher relative to the contemporaneous sea level.

626 We perform LGM climate model simulations with 8 different ice sheet topographies; 7 from LGM ice sheet  
627 reconstructions (29, 50, 116, 121-125) and the last being the PI topography. All the LGM-PI ice elevation  
628 anomalies are shown in Fig. S10, expressed relative to the PI geoid rather than contemporaneous sea level

629 (hence the 120 m elevation drop over the Southern Ocean). Each of these topographies was used to force  
630 both the MIROC and HadCM3 climate models, with the simulated  $\Delta T_s$  shown in the right two columns of  
631 Fig. S10. Unless noted otherwise, we only consider the first five of the LGM topographies shown in Fig.  
632 S10 in our analyses. The other two are only shown here for completeness and historical reasons; the Ice-  
633 5G reconstruction has been superseded by Ice-6G, and the PMIP3 ice sheet has highly unrealistic ice  
634 loading (up to 2500 m surface elevation gain) over interior West Antarctica.

635 For all topographies (main text Fig. 1, Fig. S10), the HadCM3 model on average simulates 2.2°C more  
636 LGM cooling than MIROC does across Antarctica (except for SDM). The same difference is seen in the  
637 SH extratropical (90°-30°S) zonal averages given earlier. The core site average  $\Delta T_s$  is  $-6.3 \pm 0.7$  °C in  
638 MIROC and  $-8.5 \pm 1.1$  °C in HadCM3 ( $2\sigma$  spread between topographies); it is  $-6.6 \pm 1.8$  °C in our firn-based  
639 reconstructions (Fig. 1, main text). The MIROC model thus matches our reconstructions more closely on  
640 average – the only exception is the WD site. The PMIP4 multi-model ensemble has a core site average  $\Delta T_s$   
641 of  $-7.8 \pm 4.6$  °C using the Ice-6G topography.

642 Next we address the  $\Delta T_s$  spatial pattern, which is due to a combination of a lapse rate effect due to the  
643 elevated topography and a dynamical effect caused by changed atmospheric circulation again due to the  
644 elevated topography. The overall effect of ice sheet topography is comparable between the models, with  
645 both models showing a lapse-rate cooling with increased site elevation following a lapse rate of around -  
646  $10^\circ\text{C km}^{-1}$ , close to the dry adiabatic lapse rate (main text, Fig. 4d).

647 Figure 1 of the main text compares the simulated  $\Delta T_s$  from both models to the data-based reconstructions.  
648 In terms of the absolute changes, MIROC provides a better fit to the reconstructions than HadCM3 does,  
649 with the exception of the WD site where HadCM3 provides the closer fit. In terms of the spatial pattern  
650 both models perform equally well; we find a Pearson correlation of  $r = 0.86$  between the firn-based  $\Delta T_s$  and  
651 the HadCM3 simulation (average of 5 topographies), and for MIROC this is 0.88; when averaging both  
652 models, the correlation increases to 0.96. Of all the topographies considered, Pollard and DeConto (2009)  
653 and Whitehouse *et al.* (2013) give the best correlations (0.96 and 0.91, respectively). For the models in the  
654 PMIP4 LGM ensemble (section S3.6), we find correlations ranging from  $r = 0.46$  to  $r = 0.93$  (average  
655 single-model  $r = 0.79$ ); interestingly, when averaging the models we again find an increase in correlation  
656 to  $r = 0.95$ , which is higher than for any of the individual models. It is clear that AOGCMs broadly match  
657 the spatial pattern seen in the firn-based reconstructions.

### 658 **S3.4 Isotope-enabled CESM model simulations**

659 The LGM and preindustrial simulations used the water isotope-enabled Community Earth System Model  
660 version 1.3 (iCESM1.3) with a horizontal resolution of  $1.9 \times 2.5^\circ$  (latitude  $\times$  longitude) for the atmosphere  
661 and land, and a nominal  $1^\circ$  for the sea ice and ocean. The physical climate model of iCESM1.3 is an  
662 upgraded version of CESM1 (126) with small changes in the atmosphere component. The water isotope  
663 capability of iCESM has been documented and validated against present-day and paleoclimate observations  
664 (127).

665 The iCESM1.3 LGM simulation was performed following the protocols from the Paleoclimate Modelling  
666 Intercomparison Project phase 4 (PMIP4) with the LGM (at 21ka) values of greenhouse gas concentrations,  
667 Earth orbital parameters, and the ICE-6G reconstruction of land ice sheets (19, 125). Ocean state in the  
668 simulation was initialized from an existing equilibrated LGM simulation that used an older version of  
669 CESM (128). The isotopic composition of seawater was initialized from the Goddard Institute for Space  
670 Studies observations with a constant value of 1.05‰ added to account for the glacial enrichment due to the  
671 increased LGM ice sheets. The iCESM1.3 LGM was integrated for an additional  $\sim 1,000$  years. The TOA  
672 energy imbalance averaged over the last 100 years is approximately  $-0.1 \text{ W m}^{-2}$ , indicating the surface  
673 climate has reached a quasi-equilibrium glacial state. Readers are referred to ref. (129) for details of the  
674 LGM and corresponding preindustrial simulations.

### 675 **S3.5 Brief review of published isotope-enabled LGM simulations**

676 LGM to present water isotope changes in Antarctica depend on many key factors including (but not limited  
677 to) southern hemisphere SST patterns (6), sea ice extent (7), ice sheet elevation (8), atmospheric transport  
678 pathways (130, 131), precipitation seasonality (132), stratosphere-troposphere vapor exchange (133) and  
679 post-depositional snow redistribution and snow-vapor isotope exchange (9, 134). Isotope-enabled general  
680 circulation models (iGCMs) seek to capture the aforementioned physical processes making them an  
681 invaluable tool in understanding isotopic variations in Antarctica. However, most if not all of the above-  
682 mentioned factors have large uncertainties associated with them, and some processes are not typically  
683 simulated in iGCMs (such as stratospheric exchange and post-depositional isotopic exchange). iGCMs  
684 typically capture the present-day spatial slope well (2), yet  $\delta^{18}\text{O}$  offsets at individual sites of 10 ‰ or more  
685 (i.e. twice the LGM-PI  $\delta^{18}\text{O}$  difference) are not uncommon.

686 Owing to both the complexity of the system and the large uncertainty in individual components, iGCMs  
687 simulate a wide range of values for the temporal slope  $\alpha_T$  (here: the ratio of the LGM-PI change in  $\delta^{18}\text{O}$  of  
688 precipitation over  $\Delta T_S$ ). Considering only studies that investigate the LGM-PI difference, iGCMs have  
689 suggested  $\alpha_T$  values in central East Antarctica that are:

- 690 - in the 0.1 to 0.6 ‰  $\text{K}^{-1}$  range, and thus considerably lower than the spatial slope (6, 8, 11, 135)
- 691 - in the 0.6 to 0.9 ‰  $\text{K}^{-1}$  range and thus comparable to the spatial slope (5, 6, 8, 12, 136)
- 692 - in the 0.9 to 1.6 ‰  $\text{K}^{-1}$  range and thus considerably larger than the spatial slope (12, 137) (this study)

693 Note that it is not trivial to compile the temporal slopes consistently, given that papers report values at  
694 different locations and time intervals, and the fact that the simulated temporal slopes tend to be highly  
695 variable between nearby location (8, 138). A systematic iGCM intercomparison would be highly valuable  
696 to the field. Studies that are listed more than once may report results from different models, models run  
697 under different boundary conditions, or have large differences between central East Antarctic core sites.

698 It is very challenging to assess how the full uncertainty in all the aforementioned processes influences the  
699 simulated  $\alpha_T$ . However, the inter-model spread can serve as an uncertainty estimate of how well the various  
700 processes are constrained in the models. We remove the high and low extremes from the simulated  $\alpha_T$   
701 range, and estimate that central East Antarctic LGM-PI  $\alpha_T$  as simulated by iGCMs is in the 0.3 to 1.4 ‰  
702  $\text{K}^{-1}$  range. For EDC and DF this implies a surface temperature  $\Delta T_S$  of 4°C to 21°C, and 4°C to 18°C,  
703 respectively (rounded to nearest integer, and ice core  $\delta^{18}\text{O}$  corrected for mean ocean  $\delta^{18}\text{O}$ ).

### 704 **S3.6 Paleoclimate Modeling Intercomparison Project Phase 4 (PMIP4)**

705 Here we use the multi-model LGM ensemble from the PMIP4 project, which includes the following 10  
706 models: AWI-ESM1, AWI-ESM2, CCSM4-UofT, CESM1.2, iLOVECLIM1.1.1 (GLAC-1D),  
707 iLOVECLIM1.1.1 (ICE-6G-C), INM-CM4-8, IPSL-CM5A2, MIROC-ES2L, and MPI-ESM1.2. All  
708 models use the Ice-6G LGM ice sheet topography (125) except for iLOVECLIM1.1.1 (GLAC-1D), which  
709 uses the GLAC-1D model (29). For a description of the PMIP4 LGM experiment, a description of the  
710 various models, and an assessment of their performance, we refer to previous work (19, 139).

711

## 712 **S4 Elevation change from ice core Total Air Content data**

### 713 **S4.1 Data description and cut-bubble correction**

714 We use total air content (TAC) data as a paleo-elevation proxy from the WD, EDC and DF sites. Data from  
715 EDC were previously published (22), for WD and DF we present previously unpublished data.

716 For the DF core we use two separate data sets. A first series of measurements were carried out at LGGE/IGE,  
717 Grenoble, France, using an original barometrical method implemented with an experimental setup called

718 STAN (140). The STAN allows precise evaluation of the pressure and temperature of air extracted from an  
719 ice sample by its melting-refreezing under a vacuum in a volume-calibrated cell. After correcting the  
720 measured pressure for the partial pressure of saturated water vapor and of the calibrated volume for the  
721 volume occupied by refrozen bubble free ice, the TAC is calculated using the ideal gas law. The ice samples  
722 used in STAN have a mass of about 25 g and a regular shape of a rectangular parallelepiped or cube, which  
723 facilitates estimation of their specific surface. A second, independent dataset of DF TAC was generated at  
724 Tohoku University (TU), Sendai, Japan, as part of the gas chromatography measurements (57, 141). Briefly,  
725 an ice sample of ~300 g was melted in a vessel under vacuum, and the extracted air was transferred to a  
726 sample tube after removing water vapor. The pressure of the sample air was measured at room temperature  
727 in a combined volume consisting of the sample tube, gas chromatograph inlet and sample loops, with a  
728 semiconductor pressure transducer. The volumes of the sample tubes and other parts were calibrated prior  
729 to the ice core measurements, and the temperature and pressure of the laboratory were used for normalizing  
730 the results. The latter data set was scaled linearly by 0.975 to bring it into agreement with the calibrated  
731 STAN data set; the origin for the discrepancy is not clear, but we note that the two datasets agree excellently  
732 in terms of relative variations

733 Both DF datasets were corrected for the “cut-bubble effect”, which is the gas loss from air inclusions  
734 (bubbles, gas hydrates and relaxation features such as air cavities) cut at the surface of the sample. The  
735 correction depends on the specific surface of the ice sample and the size of air inclusions (142). The size of  
736 air inclusions was measured under a binocular microscope in 2-4 mm thick sections of ice cut in parallel  
737 with the samples used for STAN measurements; for the gas chromatography data set we used linear  
738 interpolation to find the bubble diameter at the sample depths.

739 The absolute precision of the STAN TAC measurements has been estimated to be within  $\pm 0.6\%$ , and of the  
740 gas chromatography TAC measurements better than 1%. However, the overall error of the obtained TAC  
741 values amounts to 1% due to the uncertainties in the cut-bubble correction. The average reproducibility of  
742 the STAN measurements performed in the same horizontal slice of an ice core has been confirmed to be  
743 better than 1%.

744 WAIS Divide TAC and methane were measured concurrently at Oregon State University using a wet-  
745 extraction process following ref. (143) with minor changes; the method is similar to those used elsewhere  
746 (97, 144). The measurements presented were made over intervals between 2006 to 2017. The sample size  
747 was 50 to 60 g, and the majority of samples were run in duplicate. The melt-refreeze process does not  
748 extract all air, and a solubility correction is applied. The size of the correction is found by performing a  
749 second melt-refreeze extraction on the frozen sample water, which allows estimating the magnitude of the  
750 dissolved air fraction. The solubility correction as a percent of the initial pressure is determined to be  $1.3 \pm$   
751  $0.2\%$  for Holocene samples and  $1.3 \pm 0.1\%$  for the deeper samples. Replicate analyses suggest a  
752 reproducibility of better than 0.5%. As for DF, we correct for the cut-bubble effect following ref. (142). We  
753 assume uniformly distributed spherical bubbles using published bubble radius data (145). Beyond 562m  
754 depth, radius values are extrapolated using a linear fit with the bubble correction estimated to be 0 at 1600m  
755 due to the completion of the bubble-clathrate transition.

756 Panel S12d compares TAC signal differences between the LGM and PI, where we have subtracted the PI  
757 mean TAC value from the records (thereby aligning them during the PI). We find that during the LGM,  
758 TAC at EDC and DF was elevated relative to the TAC at WD; the difference is around  $7 \text{ mL kg}^{-1}$  for DF-  
759 WD, and  $6 \text{ mL kg}^{-1}$  for EDC-WD.

## 760 **S4.2 Data corrections and Elevation change**

761 It is challenging to interpret TAC in terms of absolute elevation change; our approach instead aims to  
762 reconstruct the difference in LGM elevation anomaly  $\Delta z$  between WAIS and EAIS (which we term  $\Delta\Delta z$ ).  
763 Quantitative interpretation of TAC is possible because: (1) we investigate relative changes in TAC between

764 WD and DF/EDC, which means that the large influence of insolation on both records cancels out to first  
765 order; and (2) we only investigate the LGM and PI time slices, both of which are equilibrium firm states not  
766 impacted by transient TAC anomalies due to changes in overburden pressure and/or changing firm  
767 temperature gradients (146).

768 To convert the TAC data into elevation change, we first need to apply several corrections. The first and  
769 most critical is the correction for insolation (22). Because we are interpreting relative TAC changes (WD  
770 vs. EDC/DF) rather than absolute ones, most of the insolation signal is canceled out because it affects TAC  
771 records at both locations. However, the elevation component of TAC is recorded on the gas age scale, while  
772 the insolation component of TAC is recorded on the ice age scale, and hence the cancellation effect is  
773 incomplete. To account for this, we apply an insolation correction based on the integrated summer  
774 insolation (Fig. S12c) as is the standard in the literature (22, 146). Eicher et al. (2016) report a sensitivity  
775 of  $-5.7 \text{ mL kg}^{-1} \text{ GJ}^{-1}$  at NGRIP (146), while the data from Raynaud suggest a sensitivity of  $-6.6 \text{ mL kg}^{-1}$   
776  $\text{GJ}^{-1}$  at EDC. Here, we shall use the average of these two estimates as our optimal correction scenario. To  
777 assess the uncertainty in the correction, we furthermore construct a scenario with no insolation correction,  
778 and a scenario in which we use a correction that is twice as large as the optimal scenario.

779 A second correction is for the temperature at bubble close-off, which impacts TAC in two ways: (1) it  
780 changes the pore volume at close-off, which we calculate using ref (97); and (2) it impacts the number of  
781 moles trapped at a given volume and pressure via the ideal gas law. These two effects cancel each other out  
782 almost completely; the LGM TAC correction is 0.001, 0.0005, and 0.0004  $\text{mL kg}^{-1}$  at WD, EDC and DF,  
783 respectively.

784 Third, TAC reflects the elevation of the close-off depth, and not the ice surface of interest. We apply a  
785 correction using the simulated close-off depth from the firm model, which is well-constrained by  $\delta^{15}\text{N}$  data.  
786 The LGM close-off depth anomaly relative to the PI was +7 m, -12 m, and -8 m at WD, EDC and DF,  
787 respectively. We do not assess the uncertainty in the close-off depth correction given that it is negligible  
788 compared to the uncertainty in the insolation correction.

789 The TAC changes are then converted to relative elevation change ( $\Delta\Delta z$ ) using the pressure-elevation  
790 relationship over Antarctica (147), and listed in Table S7 – note that the three insolation correction scenarios  
791 are listed separately. For each insolation scenario, and for each combination of cores (either WD-EDC or  
792 WD-DF) we calculate a lower bound  $\Delta\Delta z$  by assuming the entire WAIS-EAIS air content change is solely  
793 due to elevation change at WD (with EDC/DF elevation stable), and an upper bound  $\Delta\Delta z$  by assuming the  
794 entire WAIS-EAIS air content change is solely due to elevation change at EDC/DF (with WD elevation  
795 stable) – the reason these two provide an lower and upper bound respectively is because a given TAC  
796 change represents a larger fractional change at the higher elevation EDC/DF sites than it does at WD. We  
797 also list a “weighted mean”, in which we convert the LGM-PI change in WD TAC as if it reflects elevation  
798 change at WD, and the LGM-PI change in EDC/DF TAC as if it reflects elevation change at EDC/DF.

799 We here report a best-estimate  $\Delta\Delta z$  of 420 m, with a conservative range of 280 – 590 m; these values are  
800 labeled with the letters C, L and U in Table S7, respectively, and given with two significant digits. In order  
801 to visually compare our TAC-based estimates to the model-simulated LGM topography anomaly (Fig. 4c),  
802 we added a constant elevation change to our TAC-based estimates such that it minimizes the root-mean-  
803 square (RMS) offset between the ice sheet model-simulated (Section S3.3) and TAC-based topography  
804 anomaly; this yields an LGM elevation anomaly of +310 m, -80 m, and -140 m at WD, EDC and DF,  
805 respectively (2 significant digits).

806 By calculating the RMS offset, we find that the ice sheet simulation by Whitehouse et al. (2012) provides  
807 the closest agreement (RMS offset of 13 m) to our TAC-based  $\Delta\Delta z$ , with a  $\Delta\Delta z$  of 403 and 457 m for WD-

808 EDC and WD-DF, respectively (compare to 388 and 445 m, table S7). The next-best model is ICE-6G,  
809 which has an RMS offset of 60 m.

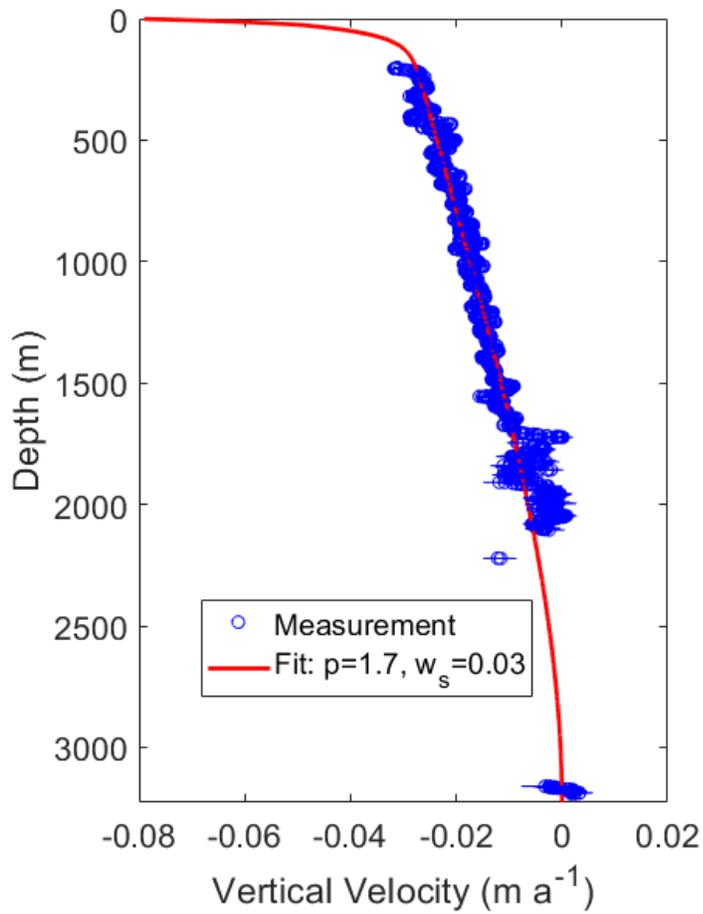
## 810 **S5 Occam's Razor**

811 Below we list reasons why we believe our reconstructions should be preferred in future interpretation of  
812 Antarctic LGM surface climate, rather than the traditional water isotope scaling method in which the  
813 modern spatial slope is used.

- 814 (1) Complexity: water isotopes are part of a complex system that involves the hemispheric-scale  
815 hydrological cycle including sea ice, SST, atmospheric transport, ice topography, etc; several of these  
816 changes are poorly constrained. By contrast, firn densification and ice flow are more simple physical  
817 systems with only local influences, that can be studied in detail today over a wide range of climatic  
818 conditions.
- 819 (2) Boundary layer inversion: Small fractional changes to the LGM inversion strength (as simulated in  
820 most climate models) complicate the reconstruction of  $\Delta T_S$  from water isotopes. Both our methods  
821 unequivocally reconstruct firn temperature, making the results easier to interpret. A small  $\Delta T_S$  in central  
822 East Antarctica is broadly consistent with water isotope observations via GCM-simulated changes to  
823 the LGM inversion (Fig. 4f, Fig. S11).
- 824 (3) Consistency: isotope-enabled models simulate LGM-PI temporal slopes that range roughly from 0.3 to  
825  $1.4 \text{ ‰K}^{-1}$  in central East Antarctica (Section S3.5) which would imply a  $\Delta T_S$  in the range of 4 to  $20^\circ\text{C}$ .  
826 By contrast, when comparing four different densification models in our  $\Delta\text{age}$ -based method, we find a  
827 range (largest minus smallest) of reconstructed  $\Delta T_S$  of  $0.78^\circ\text{C}$  at WD,  $1.70^\circ\text{C}$  at EDC, and  $1.18^\circ\text{C}$  at  
828 DF. The stated 95% confidence range (upper minus lower bound) is  $3.8^\circ\text{C}$  at WD,  $3.8^\circ\text{C}$  at EDC, and  
829  $3.4^\circ\text{C}$  at DF for the borehole method; and  $2.6^\circ\text{C}$  at WD,  $3.0^\circ\text{C}$  at EDC, and  $4.0^\circ\text{C}$  at DF for the  $\Delta\text{age}$   
830 method. The borehole- and  $\Delta\text{age}$ -based reconstruction methods agree within uncertainty at all sites. It  
831 is clear that our reconstruction methods show more consistency and agreement than isotope-enabled  
832 GCMs do.
- 833 (4) Elevation change: Air content data and ice sheet reconstructions all suggest that the LGM elevation  
834 anomaly in West Antarctica is several hundred meter higher than that in central East Antarctica (air  
835 content suggests 280 to 590 m); this implies a difference in  $\Delta T_S$  via the lapse rate. The PMIP4 model  
836 ensemble mean finds that  $\Delta T_S$  at WD minus the  $\Delta T_S$  in central East Antarctica (here mean of DF and  
837 EDC) is around  $-5.9 \pm 2.7^\circ\text{C}$ , in good agreement with our reconstructions ( $-6.1^\circ\text{C} \pm 2^\circ\text{C}$ ); using  
838 traditional water isotope scaling this difference is only  $-2$  to  $-3^\circ\text{C}$ , which is harder to reconcile with the  
839 elevation changes.
- 840 (5) Firn data: The traditional interpretation of isotope data ( $\Delta T_S \approx -9^\circ\text{C}$ ) leads to inconsistencies in  
841 simulated firn thickness ( $I05$ ), despite the fact that firn densification models are very skillful at  
842 simulating observed present-day firn density,  $\Delta\text{age}$  and  $\delta^{15}\text{N}$  over a wide range of climatic conditions.  
843 Despite efforts, nobody has been able to satisfactorily remove these inconsistencies via changes to  
844 densification physics (Section S2.5).

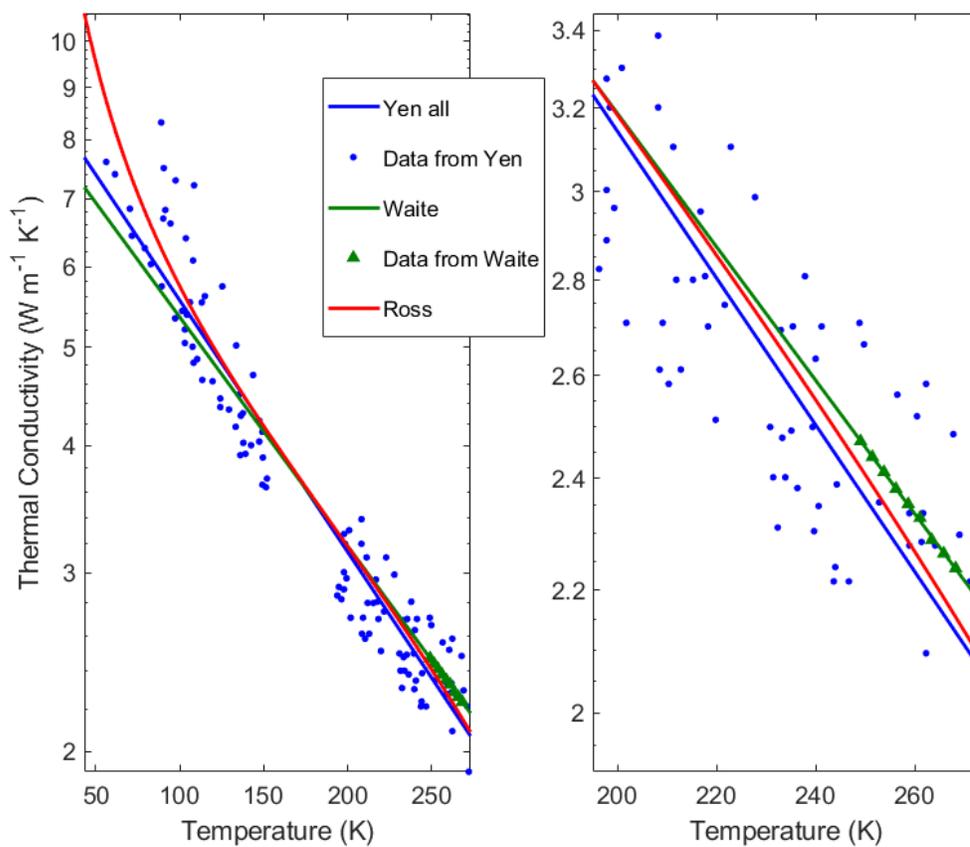
845 For these reasons, we believe that our solution of reduced  $\Delta T_S$  in central East Antarctica is the most  
846 parsimonious solution: it is consistent with the largest amount of observational and model-based evidence,  
847 while requiring the fewest number of assumptions. Following the principle of Occam's razor, this means  
848 that a small  $\Delta T_S$  in central East Antarctica should be the preferred scientific hypothesis.

849

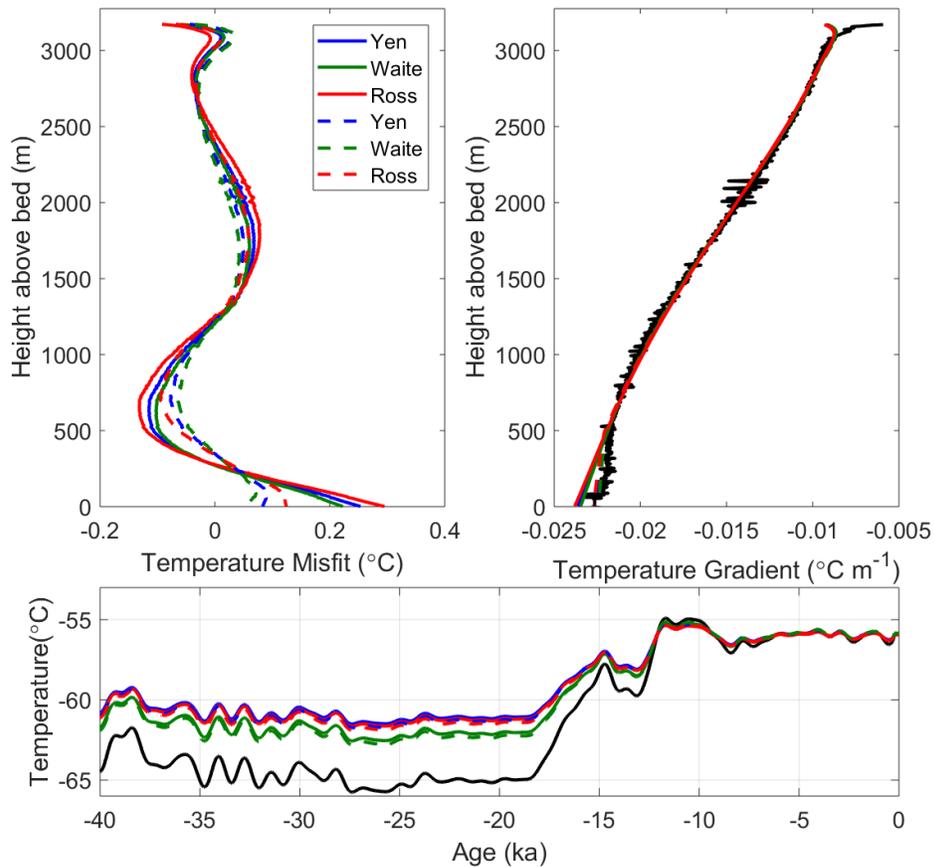


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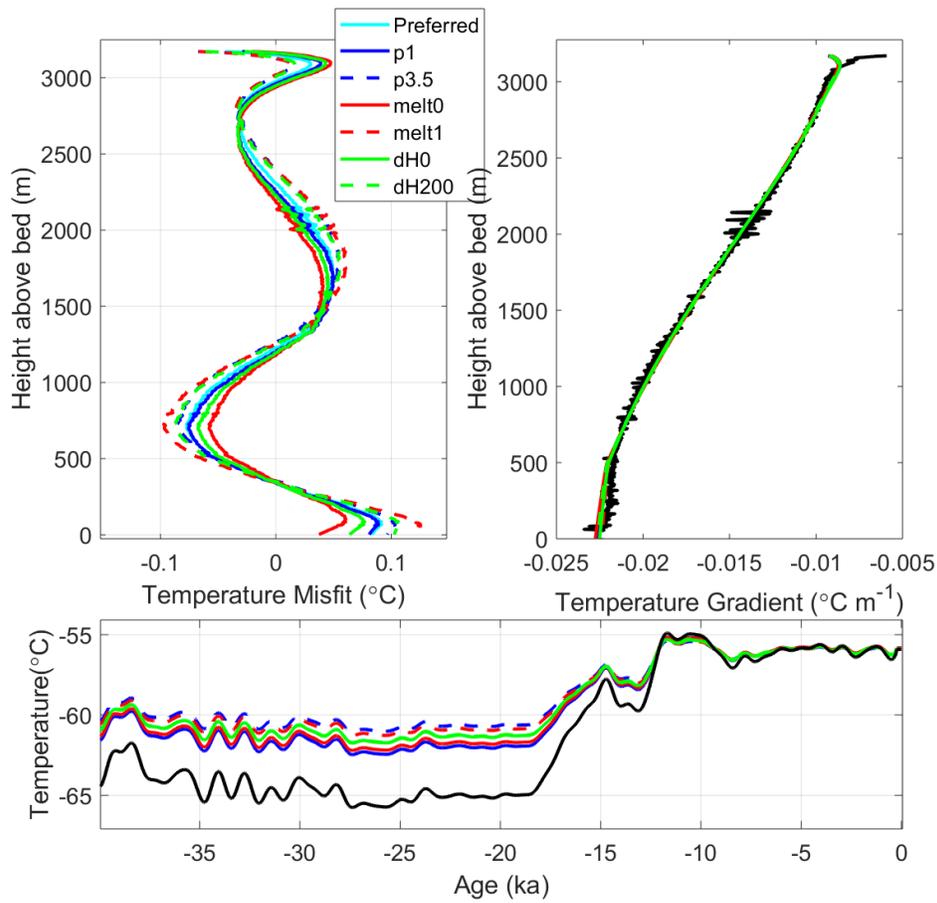
851 **Figure S1: Dome C vertical ice velocity in measurements and models.** Example fit for ApRES data with  
 852 Lliboutry approximation where  $p$  and  $w_s$  are free parameters. This fit has the ApRES measurements shifted  
 853 by the mean velocity at depths greater than 3000 m and no basal melt rate applied to the modeled profile.  
 854 Table S1 provides the fitted values for other combinations of parameters.



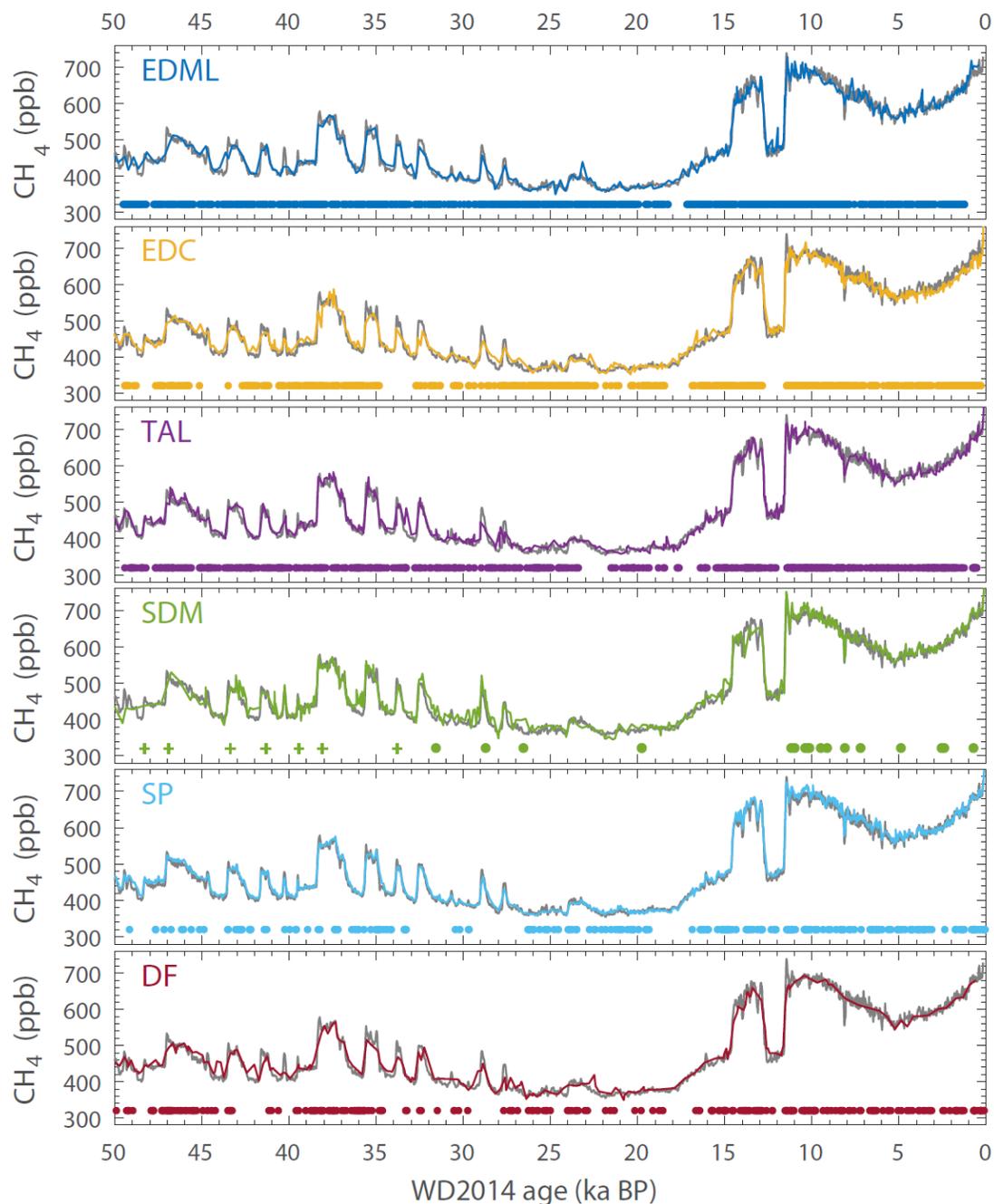
855  
 856 **Figure S2: Ice thermal conductivity.** Different relationships for the temperature dependence of thermal  
 857 conductivity. “Yen all” refers to the fit using all temperature data from Yen (1981) which is also in Cuffey  
 858 and Paterson, 2010 (33, 34). “Waite” refers to our fit to the data in Waite et al., 2006 (45). “Ross” refers to  
 859 the fit in Ross et al., 1978 in Table IIIb (44). The left panel shows the full range of the data; the right panel  
 860 shows the same focused on the temperature range relevant to the Dome C ice core.



861  
 862 **Figure S3: Fitting the Dome C borehole data – sensitivity to thermal conductivity.** Upper left panel:  
 863 Measured minus modeled temperature. Upper right panel: Modeled and measured (black) temperature  
 864 gradients. Lower panel: Inferred temperature history and temperature history from classical water isotope  
 865 scaling (black). The solid lines use the thermal conductivity parameterizations from the respective studies  
 866 (legend in top left panel); the dashed shows the same with a 5% thermal conductivity adjustment applied  
 867 to the deepest ice as a way to account for the reduced temperature gradient at these depths (top right).

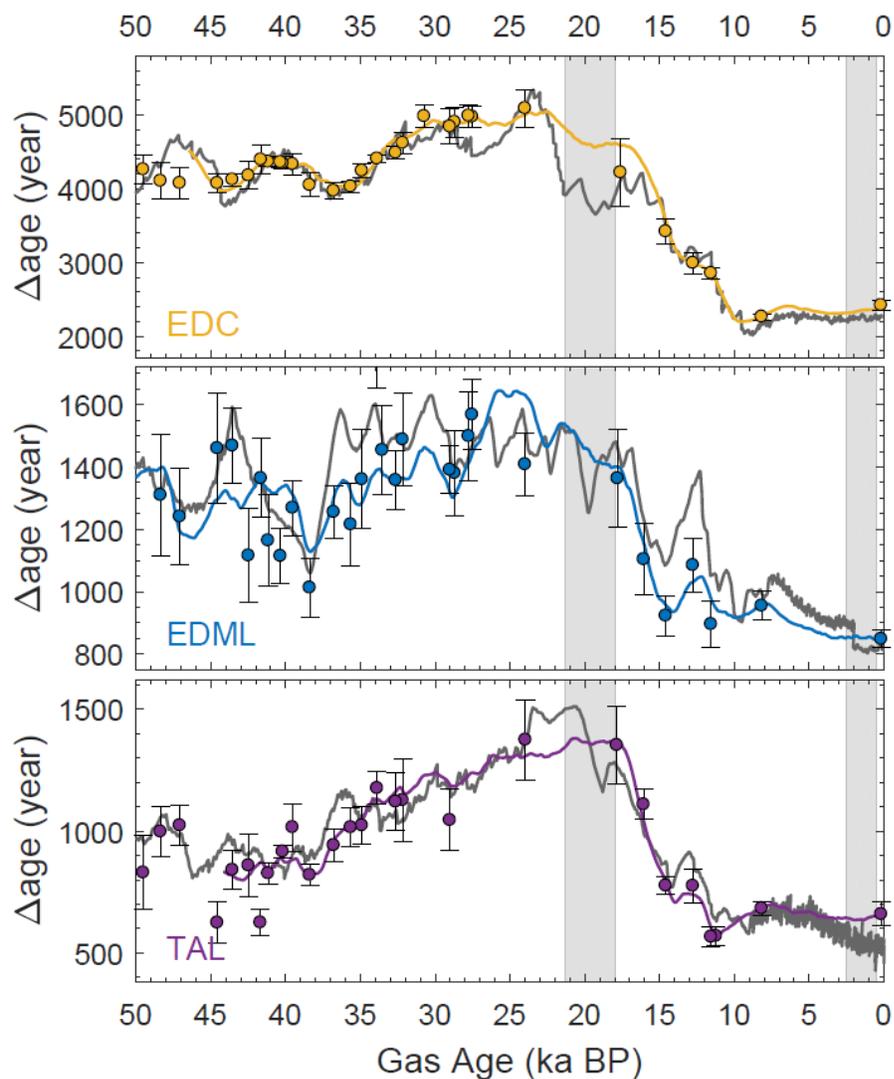


868  
 869 **Figure S4: Fitting the Dome C borehole data – sensitivity to ice flow model.** Upper left panel: Measured  
 870 minus modeled temperature. Upper right panel: Modeled and measured (black) temperature gradients.  
 871 Lower panel: Inferred temperature history and temperature history from classical water isotope scaling  
 872 (black). Legend is given in the upper left panel, with descriptions given in the text (Section S).



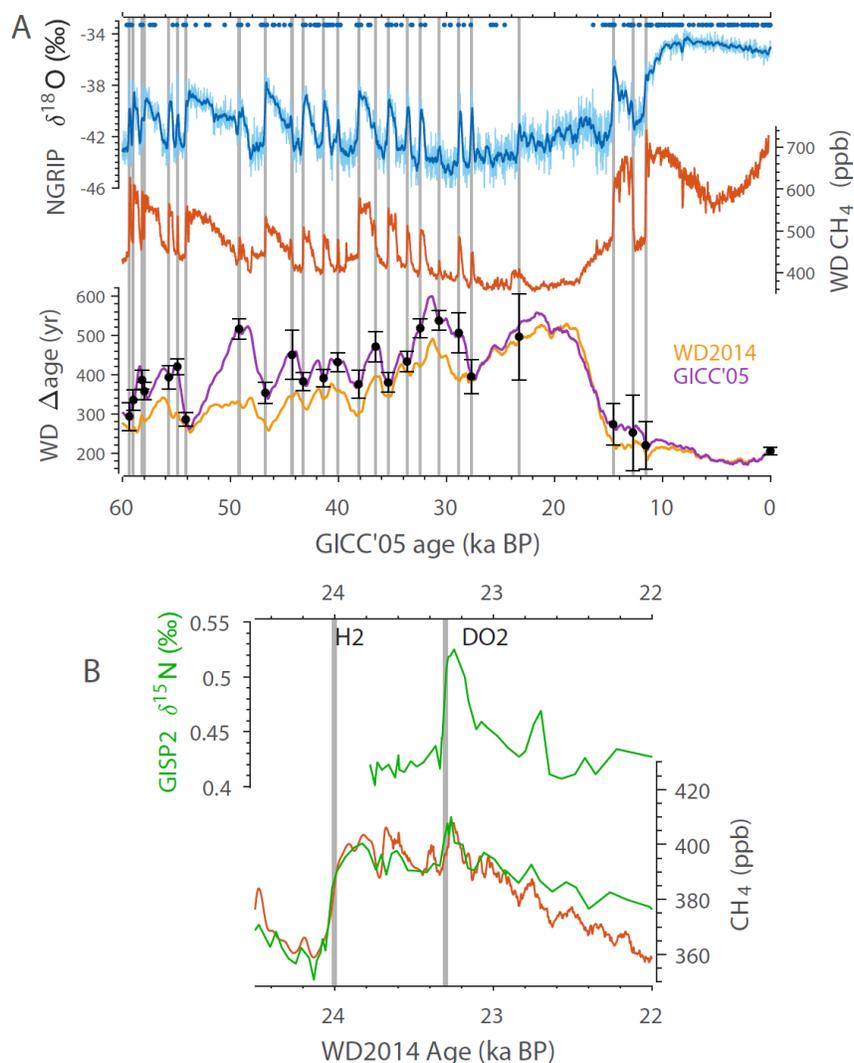
873

874 **Figure S5: Ice core synchronization via atmospheric methane and volcanic markers.** Ice core CH<sub>4</sub> in  
 875 the various cores (*57, 60, 64-68, 148*) as labeled (colored curve) synchronized to the high-resolution WD  
 876 CH<sub>4</sub> record (grey curve) on the WD2014 chronology (*30, 31, 80*). Dots represent the ages of volcanic tie  
 877 points, the + symbols (SDM only) represent ice-ice links based on features matched in the  $\delta^{18}\text{O}_{\text{ice}}$  records.



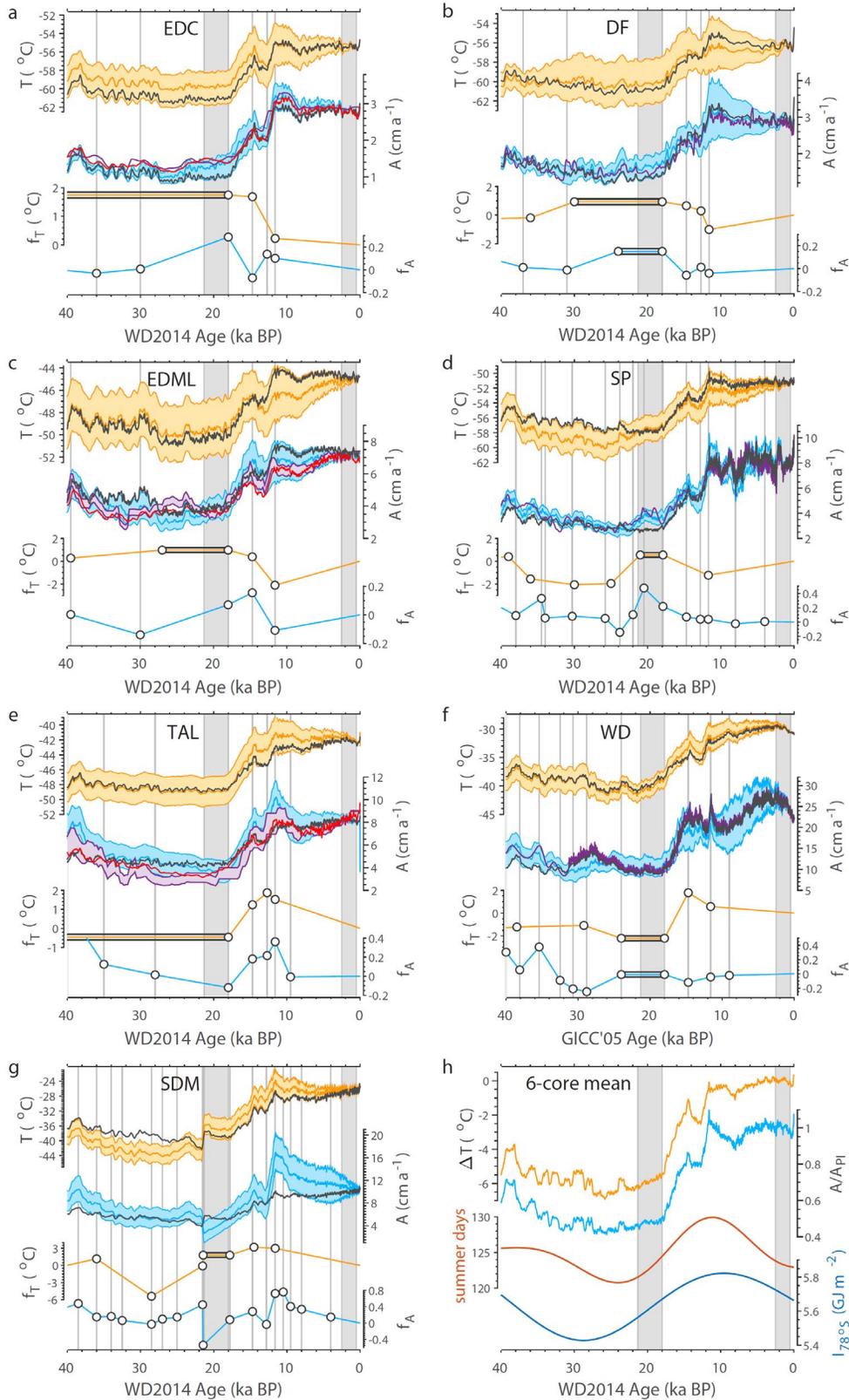
879

880 **Figure S6: Comparison of  $\Delta$ age to AICC2012.** Empirical  $\Delta$ age constraints derived here (dots) together  
 881 with  $\Delta$ age from firm densification modeling (colored curve) and  $\Delta$ age from the AICC2012 Antarctic Ice  
 882 Core Chronology (grey curve) for EDC, EDML and TAL (42). Note that in the AICC2012 approach  $\Delta$ age  
 883 is not explicitly modeled; consequently it has abrupt variations (including age inversions) that are likely to  
 884 be unphysical. The other four sites considered here (WD, SDM, SP and DF) are not part of the AICC2012  
 885 framework. Vertical grey shading denotes the LGM and PI periods.



886

887 **Figure S7: WAIS Divide Δage via bipolar volcanic and  $\text{CH}_4$  synchronization.** (a) Greenland NGRIP  
 888  $\delta^{18}\text{O}$  as a proxy for climate (upper, blue), with the location of bipolar volcanic tie points given as dots; WD  
 889 atmospheric  $\text{CH}_4$  (middle, orange); WD  $\Delta$ age on the WD2014 (30) and GICC'05 timescales (bottom, color  
 890 coded as shown). Black dots with errorbars give the empirical (GICC'05-compatible) WD  $\Delta$ age estimates  
 891 based on the inter-polar synchronization. (b) Synchronization at DO2. Top panel: Greenland GISP2  $\delta^{15}\text{N}$   
 892 (green); Bottom panel: GISP2  $\text{CH}_4$  (green) and WD  $\text{CH}_4$  (orange) data. Greenland warming at DO-2 is  
 893 indicated by GISP2  $\delta^{15}\text{N}$ , a gas-phase proxy for abrupt Greenland warming (149). Greenland DO-2  
 894 warming is coincident with a small  $\sim 15$  ppb  $\text{CH}_4$  feature in the GISP2 ice core – the same feature is visible  
 895 in the WD core. This allows us to assign a GICC'05 gas age to the WD DO2  $\text{CH}_4$  feature, using the DO-2  
 896 GICC'05 ice age. While the Heinrich H-2  $\text{CH}_4$  feature is more pronounced than the DO-2  $\text{CH}_4$  feature, it  
 897 cannot be used to synchronize WD to the GICC'05 ice chronology because the H-events are not recorded  
 898 in the Greenland ice phase. The H-2  $\text{CH}_4$  feature is used for synchronizing Antarctic  $\text{CH}_4$  records from  
 899 various cores, however (Fig. S5).



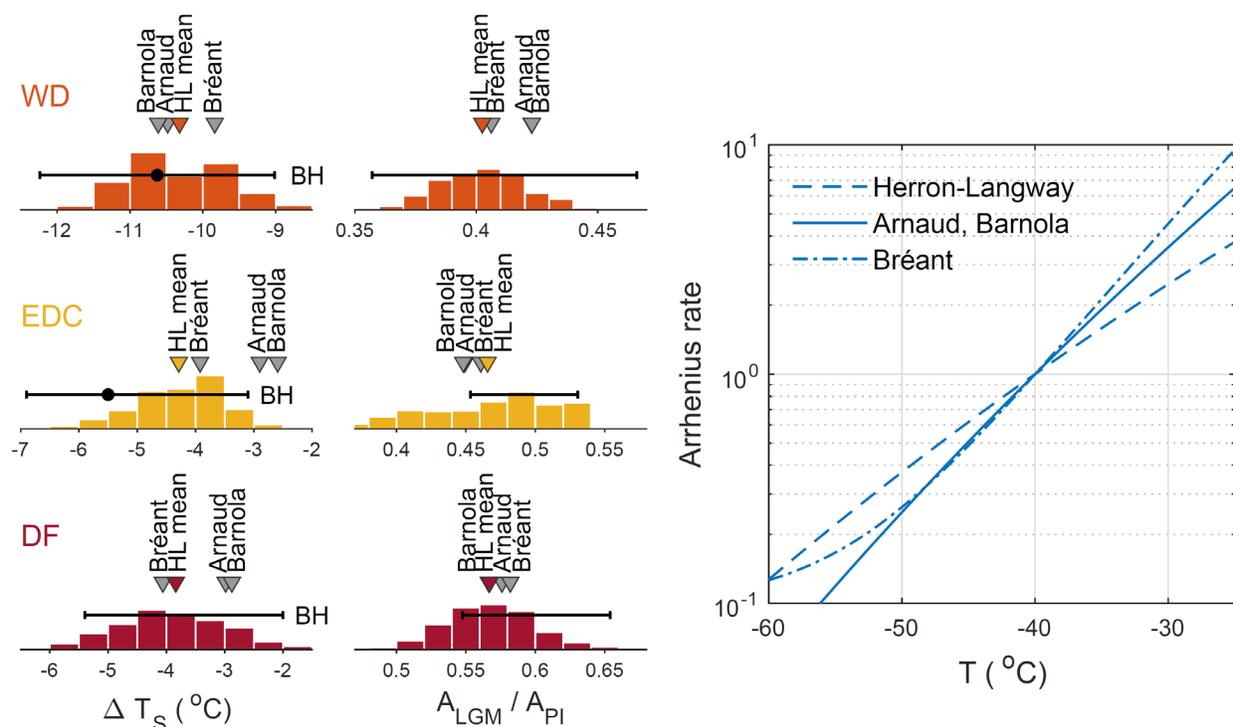
900

901 **Figure S8: Firn-based accumulation rate and temperature reconstructions. (a) EDC; (b) DF; (c)**  
 902 **EDML; (d) SP; (e) TAL; (f) WD; (g) SDM.** The initial guesses  $T_{init}(t)$  and  $A_{init}(t)$  are shown in grey; the

903  $T_{\text{site}}(t)$  and  $A_{\text{site}}(t)$  reconstructions in orange and blue, respectively, using the Monte Carlo estimation  
 904 (Section S2.4) with the envelope giving the 95% confidence range and the center line giving the distribution  
 905 mean. The high-frequency  $T$  and  $A$  variability comes from the  $\delta^{18}\text{O}$ , not from the optimization method. The  
 906 purple curves give accumulation estimates from de-strained layer thickness in the core, the red curves give  
 907 the accumulation estimates from the AICC2012 chronology (42). The  $f_T(t)$  and  $f_A(t)$  modification functions  
 908 are shown below in orange and blue, respectively, with the control points and intervals shown as white  
 909 circles and grey bars, respectively. (h) Six-core average  $T$  and  $A$  as anomalies relative to the PI. SDM is  
 910 withheld from the averaging because the abrupt 21ka feature is likely a local glaciological effect, and not  
 911 representative of Antarctic climate. Summer duration (red, days with insolation over  $250 \text{ Wm}^{-2}$ ) and  
 912 annually integrated insolation (blue) are both at  $78^\circ\text{S}$  (mean core latitude).

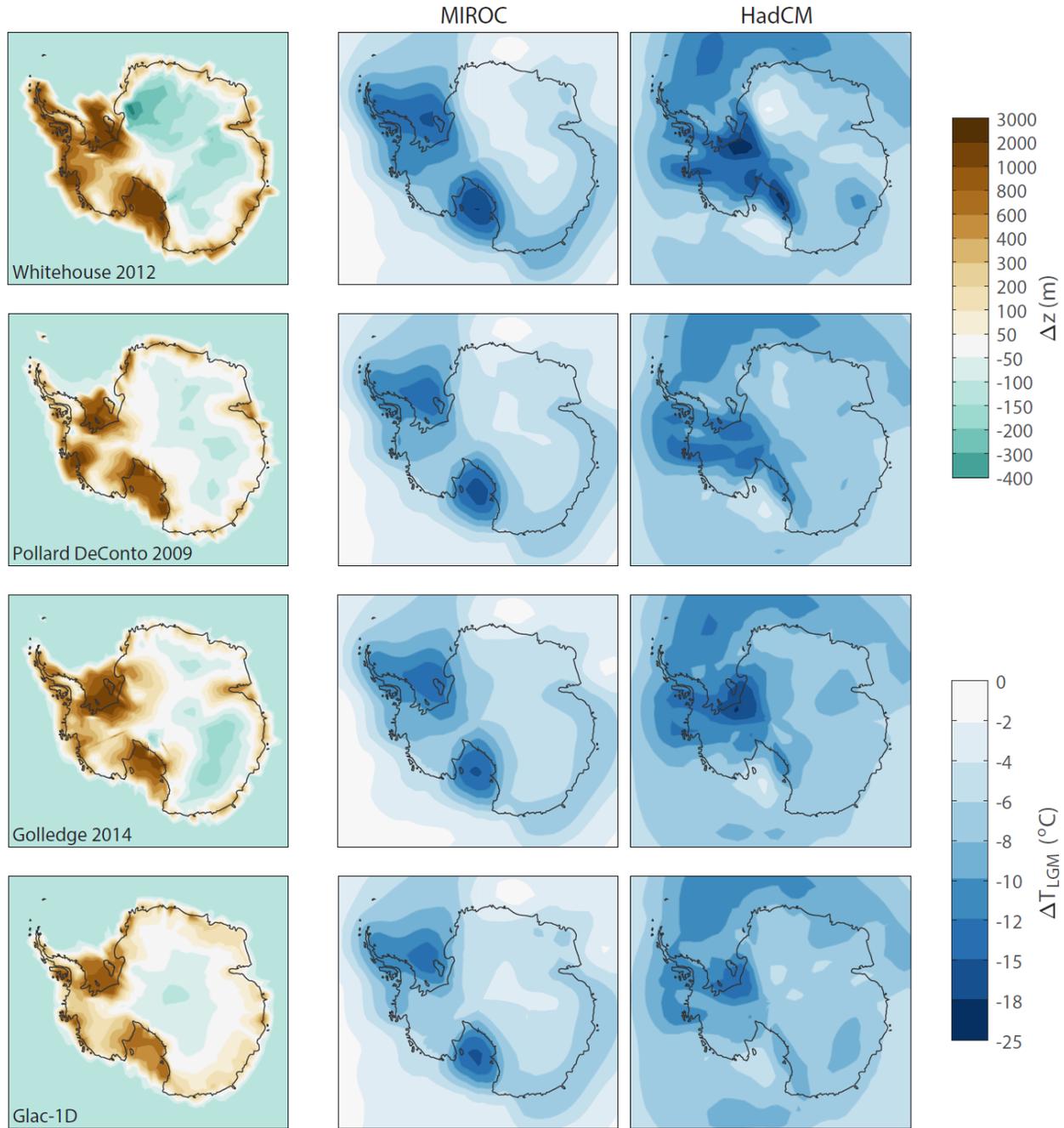
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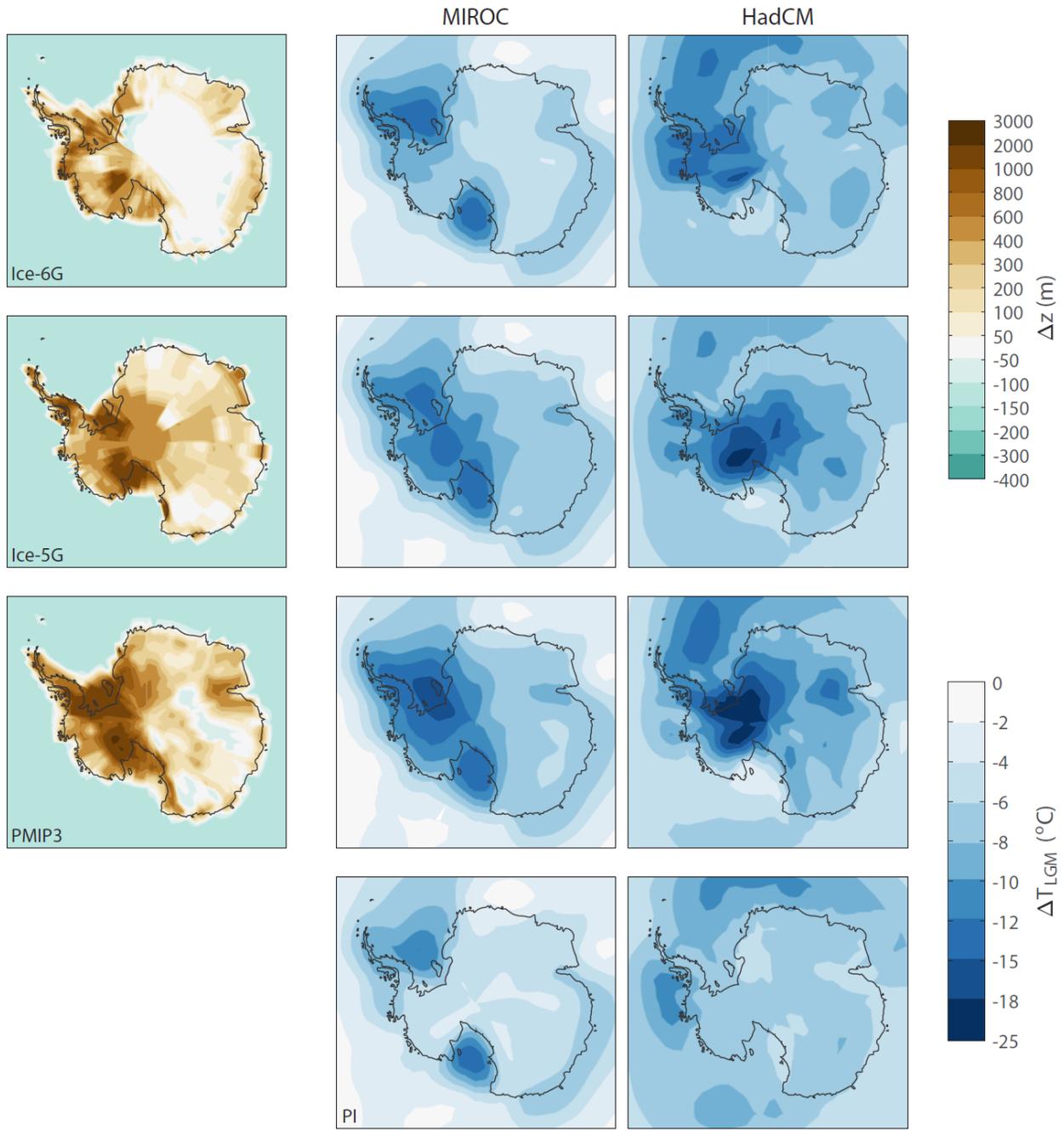
915

916 **Fig. S9. Firn-based climate reconstructions – a model comparison.** Histograms: ice core temperature  
 917 and accumulation reconstructions for the WD, EDC and DF sites. Histograms give the spread in the Herron-  
 918 Langway Monte Carlo sensitivity study (distribution mean indicated with colored triangles). Results from  
 919 the Arnaud (88, 103), Barnola (20, 87) and Bréant (89) firn densification models are indicated by the grey  
 920 triangles as marked. Black data and horizontal error bars give the range of borehole temperatures (marked  
 921 BH), and the range of accumulation rates consistent with the ice flow model uncertainty used in the borehole  
 922 temperature reconstructions. Right panel: plots of the Arrhenius-type activation energy term for the second  
 923 stage of firn densification in the firn models. The Arrhenius term has the form  $\exp(-Q/RT)$ , with  $R$  the gas  
 924 constant,  $T$  the Kelvin temperature and  $Q$  the activation energy; the Herron-Langway model uses  $Q_{\text{HL}} =$   
 925  $42.6 \text{ kJ/mol}$ ; the Arnaud and Barnola models use  $Q_A = Q_B = 60 \text{ kJ/mol}$ ; the Bréant model uses a weighted  
 926 sum of three activation energies ( $Q_1 = 110 \text{ kJ/mol}$ ,  $Q_2 = 75 \text{ kJ/mol}$ ,  $Q_3 = 1.5 \text{ kJ/mol}$ ). All values are  
 927 normalized to unit Arrhenius rate at  $-40^\circ\text{C}$ .



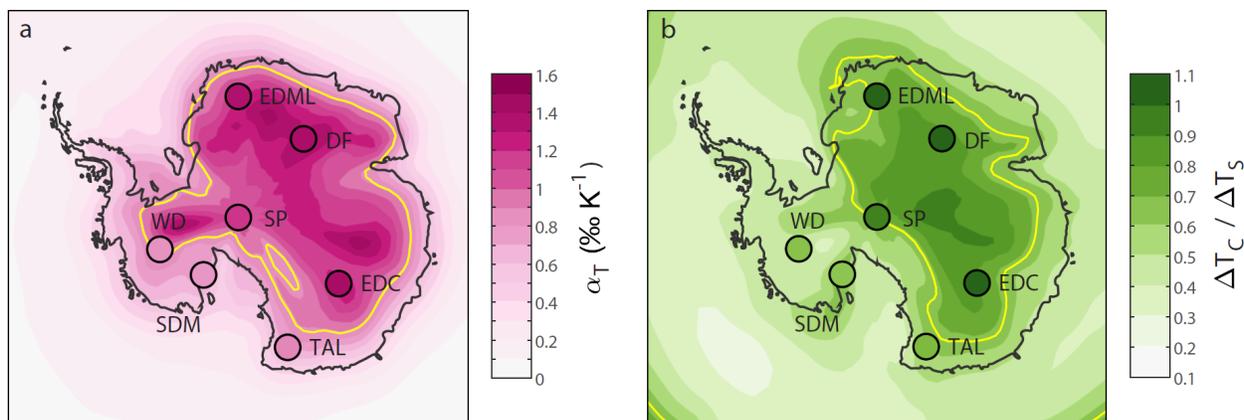
928

929 **Fig S10: Climate model ice sheet sensitivity study (continued on next page).** Left panels: Antarctic  
 930 LGM-PI surface topography anomalies used to force the climate model simulations. Topographies as  
 931 indicated from refs (29, 50, 116, 121-125). Note that the anomalies are shown relative to the modern geoid,  
 932 and as such the LGM sea level drop appears as a negative height anomaly over the oceans. Middle and right  
 933 panels: Antarctic cooling  $\Delta T_s$  simulated in the MIROC and HadCM AOGCMs, respectively, using the  
 934 topographic forcing shown in the left panels. The upper five topographies are used in the analyses of the  
 935 main manuscript; the Ice-5G and PMIP3 topographies are shown here for comparison purposes.



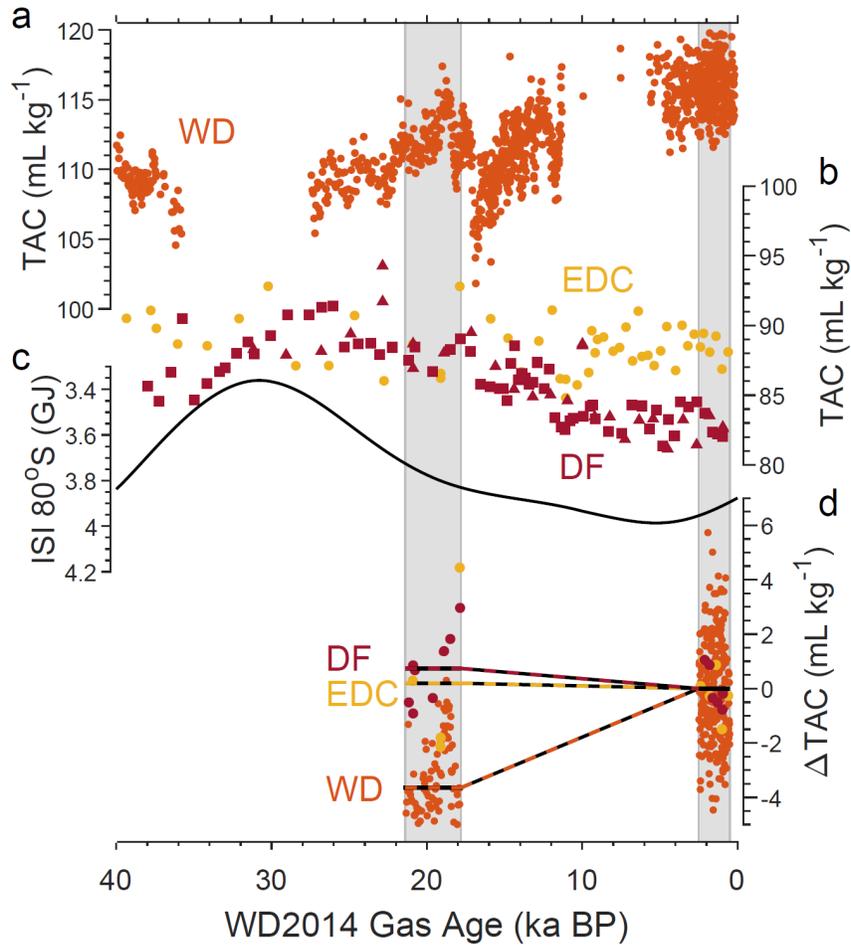
936

937 **Fig. S10 (continued).** Bottom row shows the Pre-industrial ice sheet forcing.



938

939 **Fig S11: Changes to the inversion strength.** (a) Map of the LGM-preindustrial temporal isotope slope in  
 940 isotope-enabled CESM simulations (shading), with the same based on the reconstructed  $\Delta T_S$  (dots). Data  
 941 and simulations are corrected for mean-ocean  $\delta^{18}\text{O}$ . The yellow contour line traces the modern spatial slope  
 942 value of  $0.8 \text{ ‰ K}^{-1}$  (2). (b) Map of the ratio  $\Delta T_C / \Delta T_S$  (both calculated as the LGM-preindustrial change)  
 943 in the CESM simulations (shading), with the same based on the data-based reconstructions (dots). The 500  
 944 hPa temperature is used as a proxy for  $T_C$  in CESM; this is the height with the warmest tropospheric  
 945 temperatures over interior Antarctica. The yellow contour line follows the modern spatial value  $0.65$  (2).



946

947 **Fig S12: Total air content records and site elevation change.** (a) WAIS Divide total air content data  
 948 (orange). (b) EPICA Dome C total air content data (yellow) and Dome Fuji total air content data from the  
 949 STAN setup (red triangles) and gas chromatography setup (red squares). (c) Integrated summer insolation  
 950 at 80°S, a key control on ice core air content (22). (d) LGM-PI changes in TAC at the three sites, with DF  
 951 and EDC showing a decrease in TAC, and WD showing an increase in TAC through time. These changes  
 952 can be interpreted in terms of relative surface elevation change (see text). All air content data are reported  
 953 in mL air (at standard temperature and pressure) per kg of ice.

954

955 **Table S1: Fitting the Liboutry equation to the ApRES data.** Fit uses Equation S2 where  $w_s$  is the  
 956 surface vertical velocity which is the sum of accumulation rate and ice thickness change. The ApRES data  
 957 are shown in Fig. S1.

ApRES uniform adjustment (m a <sup>-1</sup> )	melt rate in model fit (m a <sup>-1</sup> )	$p$	$w_s$ (m a <sup>-1</sup> )
0	0	3.2	0.032
0	-0.0005	2.85	0.032
0	-0.0022	1.2	0.033
0.0022	0	1.7	0.030
0.0015 (0.0022 to 0.0005)	-0.0005	1.35	0.031
	<b>Average</b>	<b>2.06</b>	<b>0.032</b>

958

959

960 **Table S2: The effect of thermal conductivity on reconstructed  $\Delta T_s$  at EDC.** We find the temperature  
 961 history (and  $\Delta T_s$ ) that provides the best fit to the EDC borehole data using three different parameterizations  
 962 of thermal conductivity. We either applied no adjustment to basal thermal conductivities, or up to 5%  
 963 increased conductivities (as marked). See Section S1.2.1 for more information. All scenarios use our best  
 964 estimate ice flow model.

	No Basal Adjustment		Basal Adjustment up to 5%	
	$\Delta T_s$ (°C)	RMS (°C)	$\Delta T_s$ (°C)	RMS (°C)
<b>Yen (1981)</b>	-5.04	0.0707	-5.29	0.0431
<b>Waite (2006)</b>	-6.01	0.0627	-6.27	0.0359
<b>Ross (1978)</b>	-5.14	0.0817	-5.39	0.0533
<b>Average (all 6)</b>	<b>-5.52</b>			
<b>Range (all 6)</b>	<b>1.23</b>			

965

966

967 **Table S4: Uncertainty estimation for EDC borehole reconstruction.**

$p$ value	2	4	4	4	7	7	7	15	15	15
onset (ka)	none	10	6	2	10	6	2	10	6	2
$\Delta T_s$ (°C)	-5.29	-4.85	-4.73	-4.52	-4.51	-4.27	-3.91	-4.14	-3.80	-3.27
$\Delta T_s$ change		0.44	0.56	0.77	0.78	1.02	1.38	1.15	1.49	2.02
RMS	.0431	.0468	.0481	.0496	.0505	.0539	.0598	.0553	.0623	.0762
RMS change		.0037	.0050	.0065	.0074	.0108	.0167	.0122	.0192	.0331
<b>Average <math>\Delta T_s</math> change</b>		<b>1.07°C</b>								
<b>Range <math>\Delta T_s</math> change</b>		<b>2.02°C</b>								

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970 **Table S4: Uncertainty estimation for EDC borehole reconstruction.** The values given represent the uni-  
 971 directional uncertainty; the full uncertainty range (interpreted as a 95% confidence interval) equals twice  
 972 the values listed. The full  $\Delta T_s$  uncertainty range stated in the manuscript equals 3.1 to 6.9°C, or  $5.0 \pm 1.9^\circ\text{C}$ .

	$\Delta T_s$ uncertainty $\sigma(\Delta T_s)$
Vertical velocity profile ( $p$ )	0.68
Basal melt rate	0.37
Ice thickness history	0.10
Time-variable vertical profile ( $p$ )	1.01
<b>Ice flow Total</b>	<b>1.3</b>
<b>Thermal conductivity (Table S2)</b>	<b>0.6</b>
<b>Total Uncertainty (Flow + thermal)</b>	<b>1.9</b>

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976 **Table S5: Firn densification model input parameters and ranges.** Preferred model input parameters  
 977 used in the dynamical Herron-Langway firn densification model with their range (noted with  $\sigma$ ) used in the  
 978 Monte Carlo sensitivity study. CZ is the convective zone thickness;  $\rho_0$  the firn surface density;  $\rho_{\text{diff}}$   
 979 expresses the density difference between the lock-in ( $r_{\text{LI}}$ ) where gases are effectively isolated from the  
 980 atmosphere, and the close-off ( $r_{\text{LI}}$ ) which is known from parameterizations (97), such that  $\rho_{\text{LI}} = \rho_{\text{CO}} - \rho_{\text{diff}}$ ;  $H$   
 981 is the ice sheet thickness; GHF is the geothermal heat flux – note that the model was not optimized to  
 982 reconstruct this parameter and we advise against using it in other applications.

Site	CZ [m]	$\sigma_1$ (CZ) [m]	$\sigma_2$ (CZ) [m]	$\rho_0$ [kg m <sup>-3</sup> ]	$\sigma$ ( $\rho_0$ ) [kg m <sup>-3</sup> ]	$\rho_{\text{diff}}$ [kg m <sup>-3</sup> ]	$\sigma(\rho_{\text{diff}})$ [kg m <sup>-3</sup> ]	GHF [mW m <sup>-2</sup> ]	H [m]	$\alpha_{\text{init}}$ [%K <sup>-1</sup> ]
<b>EAIS</b>										
EDC	0.0 <sup>a</sup>	2.0	5.0	340	50	4.0	4.0	48	3275	1.1
DF	4.0 <sup>b</sup>	2.0	5.0	335	50	3.0	4.0	54	3038	1.2
EDML	2.0 <sup>c</sup>	2.0	5.0	320	50	5.0	4.0	50	2590	1.15
TAL	5.0 <sup>d</sup>	2.0	5.0	320	50	9.0	4.0	62	1620	0.95
SP	6.0 <sup>e</sup>	2.0	5.0	380	50	15	4.0	56	2600	1.1
<b>WAIS</b>										
WD	3.5 <sup>f</sup>	2.0	5.0	420	50	10	4.0	n/a <sup>g</sup>	1000 <sup>g</sup>	0.88
SDM	2.0 <sup>h</sup>	2.0	5.0	340	50	3.5	4.0	72	1004	0.7

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<sup>a</sup>) See ref. (150)

<sup>b</sup>) Lower bound from model fitting in ref. (151)

<sup>c</sup>) Based on EDML firn air data (152)

<sup>d</sup>) Generic value used owing to lack of data

<sup>e</sup>) Firn-based estimates suggest the CZ is around 3 m at SP (153); increased here to fit ice core  $\delta^{15}\text{N}$  data

<sup>f</sup>) See ref. (154)

<sup>g</sup>) Due to the high accumulation rates at WD, the geothermal heat flux does not meaningfully penetrate into the firn column and only the upper 1000m are simulated in accordance with ref. (30).

<sup>h</sup>) See ref. (153).

992 **Table S6: Firn densification model input parameters for alternative firn model physics.** Preferred  
 993 model input parameters used in the various firn densification models. CZ is the convective zone thickness;  
 994  $\rho_0$  the firn surface density;  $\rho_{\text{diff}}$  expresses the density difference between the lock-in ( $r_{\text{LI}}$ ) where gases are  
 995 effectively isolated from the atmosphere, and the close-off ( $r_{\text{LI}}$ ) which is known from parameterizations  
 996 (97), such that  $\rho_{\text{LI}} = \rho_{\text{CO}} - \rho_{\text{diff}}$ ; H, GHF and  $\alpha_{\text{init}}$  are as in Table S5 and identical for the different models.

Site and Model	CZ [m]	$\rho_0$ [kg m <sup>-3</sup> ]	$\rho_{\text{diff}}$ [kg m <sup>-3</sup> ]
<b>WDC</b>			
Herron-Langway	3.5	420	10
Arnaud	3.5 <sup>a</sup>	390	-2.0
Barnola	3.5 <sup>a</sup>	420	10
Bréant	3.5 <sup>a</sup>	420	6.0
<b>EDC</b>			
Herron-Langway	0.0	340	4.0
Arnaud	1.0 <sup>b</sup>	340	-2.0
Barnola	1.0 <sup>b</sup>	340	4.0
Bréant	1.0 <sup>b</sup>	340	1.0
<b>DF</b>			
Herron-Langway	4.0	335	3.0
Arnaud	4.0	335	-3.0
Barnola	4.0	335	3.0
Bréant	4.0	335	-1.0

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 998 **Table S7: Relative elevation changes inferred from total air content records.** All elevation changes are  
 999 expressed as relative WAIS (WD) minus EAIS (EDC, DF) LGM elevation changes, and not absolute  
 1000 changes relative to the geoid. So a value of +400 m could for example indicate a 300 m elevation increase  
 1001 at WD, and a 100 m of elevation decrease at EDC/DF, or for example a 500m elevation increase at WD  
 1002 and a 100m elevation increase at EDC/DF. See the text (Section S4) for details. The values marked with  
 1003 letters C, L, and U, are used as the central estimate, lower bound and upper bound, respectively, reported  
 1004 in the main manuscript (in two significant digits).

	Lower bound [m]	Weighted mean [m]	Upper bound [m]
<b>No insolation correction</b>			
WD-EDC	418	496	516
WD-DF	453	563	588 U
mean	436	530	552
<b>Best-estimate insolation correction</b>			
WD-EDC	346	388	429
WD-DF	380	445	496
mean	363	417 C	462
<b>2 × best-estimate insolation correction</b>			
WD-EDC	275 L	278	342
WD-DF	308	325	404
mean	291	302	373

1005 **Supplementary references and notes**

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