1 **Abrupt Ice Age Shifts in Southern Westerlies and Antarctic Climate Forced from**

2 **the North**

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24   The Southern Hemisphere (SH) mid-latitude westerly winds play a central role in the global

25   climate system via Southern Ocean upwelling[1], carbon exchange with the deep ocean[2],

26   Agulhas Leakage[3], and possibly Antarctic ice sheet stability[4]. Meridional shifts of the SH

27   westerlies have been hypothesized in response to abrupt North Atlantic Dansgaard-Oeschger

28   (DO) climatic events of the last ice age[5,6], in parallel with the well-documented shifts of the

29   intertropical convergence zone (ITCZ)[7]. Shifting moisture pathways to West Antarctica[8] are

30   consistent with this view, but may represent a Pacific teleconnection pattern[9]. The full SH

31   atmospheric-circulation response to the DO cycle and its impact on Antarctic temperature

32   remain unclear[10]. Here we use five volcanically-synchronized ice cores to show that the

33   Antarctic temperature response to the DO cycle can be understood as the superposition of

34   two modes: a spatially homogeneous oceanic "bipolar seesaw" mode that lags Northern

35   Hemisphere (NH) climate by about 200 years, and a spatially heterogeneous atmospheric

36   mode that is synchronous with NH abrupt events. Temperature anomalies of the atmospheric

37   mode are similar to those associated with present-day Southern Annular Mode (SAM)

38   variability, rather than the Pacific South America (PSA) pattern. Moreover, deuterium excess

39   records suggest a zonally coherent migration of the SH westerlies over all ocean basins in

40   phase with NH climate. Our work provides a simple conceptual framework for understanding

41   the circum-Antarctic temperature response to abrupt NH climate change. We provide

42   observational evidence for abrupt shifts in the SH westerlies, which has previously-

43   documented[1-3] ramifications for global ocean circulation and atmospheric CO_2. These coupled

44   changes highlight the necessity of a global, rather than a purely North Atlantic, perspective on

45   the DO cycle.

46   During the glacial DO cycle, abrupt variations in northward heat transport by the Atlantic

47   Meridional Overturning Circulation (AMOC) affect Greenland and Antarctic temperature

48   oppositely (Fig. 1), via an oceanic teleconnection called the bipolar seesaw[6,11]. Antarctica warms

49   during Greenland cold phases (stadials), and cools during Greenland warmth (interstadials),

50   with the gradual nature of Antarctic climate change reflecting buffering by a large heat

51   reservoir[11] – likely the global ocean interior[6]. The DO cycle affects atmospheric circulation also;

52   the ITCZ shifts southwards during stadials, and northwards during interstadials[7]. General

53   Circulation Model (GCM) simulations suggest parallel shifts of the SH westerlies[5,6,12], but the

54   available observational evidence (a deuterium excess record from West Antarctica[8]) cannot

55   distinguish between such shifts and Pacific-only teleconnections[9]. Furthermore, the impact of

56   the atmospheric circulation changes on Antarctic climate remains unknown, and models are

57   inconclusive on this question[10,13].

58   We use water stable isotope ratios, a proxy for site temperature[14], from five Antarctic ice cores:

59   WAIS (West Antarctic Ice Sheet) Divide (WDC), EPICA (European Project for Ice Coring in

60   Antarctica) Dronning Maud Land (EDML), EPICA Dome C (EDC), Dome Fuji (DF) and Talos Dome

61   (TAL). WDC is synchronized to Greenland ice cores at high precision via atmospheric methane

62   (Fig. 1a, b)[15]; here we synchronize WDC to the other cores in the 10-57 ka before present (BP)

63   interval using volcanic markers (Methods; Extended Data Figure 1), greatly improving our ability

64   to study the timing of regional Antarctic climate variations relative to Greenland. The Antarctic

65   response to DO events is investigated using a stacking technique, in which 19 individual events

66   are aligned at the midpoint of their abrupt methane transition in the WDC core, and averaged

67   to obtain the shared climatic signal (Methods).

68   Antarctica cools in response to DO warming (Fig. 2a, b), consistent with the bipolar seesaw

69   theory[6,11]. In the Antarctic mean $\delta^{18}O$ stack, the cooling onset occurs about two centuries after

70   the abrupt NH event, providing validation of earlier results from West Antarctica[15]. There is a

71   spatial pattern to the Antarctic response, however. A step-like divergence from the mean signal

72   is seen around $t \approx 0$ yr (i.e., synchronous with NH climate), with the interior East Antarctic

73   Plateau sites (DF and EDC) warming, and EDML cooling (Fig. 2c). This instantaneous warming

74   over the Plateau is particularly pronounced at DO events 1, 8, 12 and 14 (Fig. 1d, red curve).

75   Using principal component analysis (PCA, see Methods), we find that two modes of variability

76   explain over 96% of signal variance in the five stacked records (Fig. 2d). The first principal

77   component (PC1, 83% of variance explained) has the triangular shape of the Antarctic Isotope

78   Maximum events – the classic thermal bipolar seesaw signal[11] – with a spatially homogeneous

79   expression (Fig. 2f). The two-century lag behind Greenland warming identifies PC1 as an ocean-

80   propagated response[15].

81   The second principal component (PC2, 13% of variance explained) is a step-like function with a

82   heterogeneous spatial pattern (Fig. 2g). This mode is very different from the bipolar seesaw.

83   The PC2 response is synchronous with NH warming within precision (28 ± 40 year lag); this

84   timing, and additional evidence presented below, suggest this mode represents an atmospheric

85   teleconnection. The PCA does not necessarily separate physical processes. We assume two

86   underlying teleconnections: oceanic (two-century lag) and atmospheric (synchronous). Some

87   amount of each process is included in PC1, as evident by some immediate warming around t=0.

88   We perform a rotation of the PCA vectors (Methods) to isolate the "purely" oceanic and

89   atmospheric responses (Fig. 2e). The associated estimate of the atmospherically-forced

90   temperature anomaly (Fig. 2h) is cooling at EDML, warming at DF, EDC and TAL, and a negligible

91   response at WDC; this pattern is robustly reproduced using different methods (Extended Data

92   Fig. 6). The magnitude of the Antarctic atmospheric response is roughly proportional to the

93   Greenland ice core $\delta^{18}O$ perturbation (Extended Data Fig. 4).

94   The Antarctic response to DO cooling is qualitatively similar to the DO warming case. The ocean

95   seesaw warming response is delayed by 226 ± 44 years and the EOF2 spatial pattern has the

96   opposite sign – i.e. additional warming at EDML, and cooling on the interior East Antarctic

97   Plateau (Extended Data Fig. 7). The atmospheric signal over Antarctica is much weaker for the

98   DO cooling case, with PC2 explaining only 9% of variance. This difference is likely due to the fact

99   that DO warmings are more abrupt and of larger magnitude than DO coolings.

100   To better understand the atmospheric mode, we turn to deuterium excess (d), a proxy for

101   vapor source conditions[16] commonly used to identify changes in atmospheric circulation and

102   vapor transport pathways[8,17,18]. In isotope-enabled GCM simulations, Antarctic d is anti-

103   correlated with the Southern Annular Mode (SAM) index[8,19]. This anti-correlation can be

104   understood conceptually: when the SH westerlies are displaced equatorward (negative SAM

105   phase), Antarctic moisture will originate further north where sea-surface temperature (SST) is

106   higher and relative humidity lower (Extended Data Fig. 8b), both of which act to make d more

107   positive[16]. We use the logarithmic definition of deuterium excess (d_{ln}), which better preserves

108   isotopic moisture source information than the linear definition[8,20].

109   The Antarctic mean d_{ln} response (Figs. 3a, b) lags NH climate by 8 ± 48 years for DO warming,

110   and 9 ± 42 years for DO cooling, respectively, consistent with previous results for WDC[8]. The

111 observed d_{ln} response is consistent with a shift in the meridional position of the SH westerly

112 winds and vapor origin, such that they move equatorward in response to NH warming, and

113 poleward in response to NH cooling. The timing suggests propagation to the SH high-latitudes

114 via an atmospheric teleconnection. The d_{ln} response is largest for the interior Plateau sites (DF,

115 EDC), possibly because their vapor source areas are more distant from confounding local effects

116 such as the sea-ice edge[21]. The response is weak or absent at EDML; possibly because SH

117 westerlies' variability is relatively weak in the Atlantic sector (Extended Data Fig. 9), or because

118 of regional effects such as wind-driven changes to the sea ice, gyre circulation or Weddell Sea

119 deep convection[22]. Critically, the four cores that do show a clear d_{ln} response collectively

120 sample water vapor from all ocean basins (Extended Data Fig. 8a), suggesting the changes to

121 the SH atmospheric circulation are zonally coherent and involve all ocean basins (rather than

122 just the Pacific basin as demonstrated previously with WDC).

123 Figure 4a compares the two independent signals we attribute to a change in atmospheric

124 circulation: PC2 of the $\delta^{18}O$ response, and the Antarctic mean d_{ln} response. Their time evolution

125 is nearly identical, suggesting they are distinct but consistent manifestations of the atmospheric

126 circulation change. The SAM and Pacific-South American (PSA) pattern are the leading modes of

127 large-scale SH atmospheric variability with strong influence on Antarctic temperature[9,23]. We

128 focus our analysis on East Antarctica, where we infer the largest response. The SAM (Fig. 4b)

129 clearly impacts East Antarctic surface air temperature (SAT) strongly (correlation $|r|$ up to

130 0.65), and with the correct sign to explain the warming seen at EDC, DF and TAL. East Antarctic

131 warming is seen for a more negative SAM index, driven primarily by anomalous atmospheric

132 heat advection[24] (the observed cooling at EDML is discussed below). The PSA (Fig. 4c), on the

133   other hand, is not meaningfully correlated with SAT at the East Antarctic sites ($|r|$ at or below

134   0.15). We further create a synthetic index that is the projection of the atmospheric loadings

135   (Fig. 2h) onto the reanalysis SAT anomaly at the core locations (Methods). The patterns in SAT

136   and geopotential height associated with this index (Fig. 4d) closely resemble those of the SAM,

137   with warming in East Antarctica, and roughly annular geopotential height anomalies.

138   These tests suggest that the SAM is the closest present-day analog to the temperature

139   response we identify in the ice core record, corroborating our independent evidence from the

140   d_{ln} data. While the PSA pattern may have been active during the DO cycle, it does not dominate

141   the Antarctic response.

142   Our data-based inferences on the timing and sign of changes to the SH westerlies/SAM are

143   consistent with coupled atmosphere-ocean GCM simulations in which AMOC transitions are

144   induced by North Atlantic fresh-water forcing[6,12,25]. Such model simulations show a positive

145   shift in SAM index in response to AMOC shutdown and vice versa (Fig. 3a, b); this shift is

146   synchronous with the applied forcing within uncertainty (Extended Data Table 1). Our observed

147   atmospheric response is more gradual than the model-simulated SAM shift, possibly because of

148   (multi-decadal) data resolution in some cores and the fact that the d_{ln} signal integrates over a

149   large moisture source area extending to 20°S.

150   Next, we address differences between the ice-core data and the modern-day correlation

151   pattern (Fig. 4b), most notably at EDML. The reanalysis correlation pattern captures the SAT

152   response to monthly internal SAM variability, representing atmospheric heat advection

153   anomalies[24]. The ice cores, on the other hand, record the response to a persistent long-term

154 shift in SAM[13,26], driving changes in SST, stratification and sea ice extent[22,26]. We speculate that

155 on longer timescales the oceanic influence of the Weddell Sea drives the cooling at EDML, due

156 to e.g. enhanced sea ice cover[22] and stratification, and a weakening of the wind-driven Weddell

157 Gyre. The negligible warming response at WDC is consistent with the relatively weak influence

158 of the SAM in West Antarctic seen in monthly reanalysis (Fig. 4b). Our observations may help

159 constrain the long-term response to a persistent SAM shift, on which GCMs disagree[13].

160 Last, we want to highlight additional structure in the Antarctic $\delta^{18}O$ stacks that is currently not

161 part of scientific discourse on interhemispheric climate coupling. Most notably, Antarctic

162 warming appears to slow down around $t=-400$ yr (Fig. 2b), forming a secondary change point

163 that precedes the abrupt DO warming events[27]. Likewise, the rate of Antarctic cooling appears

164 to increase 200 years prior to the abrupt DO cooling events (Extended Data Fig. 7b). These

165 secondary change points are subtler than the ones analyzed in this work, have no apparent

166 corresponding features in Antarctic d_{ln} or Greenland climate, and remain unexplained.

167 In conclusion, our results show that Antarctica is influenced by NH abrupt climate change on

168 two distinct time scales, representing a slow oceanic and a fast atmospheric teleconnection. In

169 particular, we provide observational evidence for zonally-coherent meridional shifts in the SH

170 westerly winds in phase with Greenland DO events, and its impact on Antarctic temperature.

171 Such shifts have implications for global ocean circulation, Southern Ocean upwelling and

172 productivity, and atmospheric CO_2[1-3]. It is therefore paramount to consider the DO cycle from a

173 global, rather than a purely North Atlantic perspective.

174

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270   **Figure 1 | Records of abrupt glacial climate variability. a,** Greenland Summit (average of GISP2

271   and GRIP[28]) ice core δ^{18}O. **b**, WDC methane[29]. **c**, Antarctic 5-core average d_{ln} anomaly. **d**,

272   Antarctic δ^{18}O anomaly at EDML (blue), the Antarctic Plateau (average of DF and EDC, red) and

273   5-core average (black); offset for clarity. All records are synchronized to the WAIS Divide

274   WD2014 chronology; Antarctic data are shown as anomalies relative to present. DO interstadial

275   periods marked in grey and numbered; Heinrich stadials marked in blue. Isotope ratios are on

276   the VSMOW (Vienna Standard Mean Ocean Water) scale. Thin curves show records at original

277   resolution (ranging from ~5 to ~50 years), with the thick lines a moving average (300 and 150

278   year window for Antarctic and Greenland data, respectively).

279

280   **Figure 2 | The Antarctic climate response to DO warming. a**, Stack of NGRIP δ^{18}O. **b**, Stack of

281   Antarctic δ^{18}O at indicated locations, with "Plateau" the average of DF and EDC. **c**, As in **b**, but

282   with 5-core mean subtracted. **d**, First two principal components of the Antarctic δ^{18}O stacks

283   (1500 yr window), with percentage of variance explained (offset for clarity). PC1 is strongly

284   correlated (r = 0.998) to the Antarctic mean. Linear fit to PC1 (t = -400 to t=0 interval) is shown

285   to highlight the response around t = 0. **e**, Rotated PC1 and PC2 vectors representing proposed

286   oceanic and atmospheric modes, with fits from change point analysis (Methods). The oceanic

287   mode lags by 211 ± 42 years; the atmospheric mode by 28 ± 40 years (1σ bounds; Extended

288   Data Table 1). **f-g**, Empirical Orthogonal Functions EOF1 and EOF2 associated with PC1 and PC2

289   in **d**, scaled to show the magnitude in permil. **h**, Spatial pattern associated with the atmospheric

290   mode as shown in **e**, scaled to permil. Isotope ratios are on the VSMOW scale.

291

292   **Figure 3| Deuterium excess and the SH westerlies. a**, DO warming: Greenland δ^{18}O stack

293   (turquoise); 5-core average Antarctic d_{ln} stack (orange with BREAKFIT result, see Extended Data

294   Table 1); SAM index (here the leading principal component of sea level pressure variability

295   south of 20°S) following a freshwater-forced AMOC perturbation in CCSM3 (Community Climate

296   System Model version 3) model simulations (grey with fit from change-point analysis, see

297   Extended Data Table 1). **b**, as **a**, but for DO cooling. **c**, Magnitude of Antarctic d_{ln} response to DO

298   warming in permil. The weak d_{ln} trend before and after the abrupt jump likely reflects the SST

299   of SH vapor source waters following the thermal bipolar seesaw[8,11]. **d**, as **c**, but for DO cooling.

300   Isotope ratios are on the VSMOW scale.

301

302   **Figure 4 | Attribution of the atmospheric mode of Antarctic temperature variability. a**,

303   Comparison of PC2 of the 5 Antarctic $\delta^{18}O$ stacks as in Fig. 2d (pink, left axis) and the Antarctic

304   mean d_{ln} stack as in Fig. 3a (black, right axis). Isotope ratios are on the VSMOW scale. **b**,

305   Correlation between a standardized monthly SAM index and SAT (2-meter temperature) in ERA-

306   Interim[30] for 1979-2017 (shading, scale bar on right) with superimposed 850 hPa geopotential

307   height regressions (10 m contours) and the ice core atmospheric temperature mode from Fig.

308   2h (circles, scale bar from Fig. 2). Note that regions of anti-correlation are colored red (i.e.

309   warming in response to negative SAM shift). **c**, as panel **b**, but for a standardized PSA index. The

310   SAM and PSA are here taken to be PC1 and PC2 of the 850 hPa geopotential height field south

311   of 20°S, respectively. **d**, as panel **b**, but for a synthetic index of the atmospheric mode created

312   by regressing ERA-Interim SAT anomalies at the ice core sites onto the coefficients in Fig. 2h

313   (Methods).

314

Methods

Volcanic ice core synchronization: Volcanic reference horizons provide the most precise way to synchronize ice-core age scales[31-35]. Within the last decade, progress has been made in volcanically linking the EPICA (European Project for Ice Coring in Antarctica) Dome C (EDC) ice core to the EPICA Dronning Maud Land (EDML) core[31], the Talos Dome (TAL) core[32], and the Dome Fuji (DF) core[33]. Here we provide new volcanic stratigraphic links between the WAIS (West Antarctic Ice Sheet) Divide ice core (WDC) and the EDML, EDC and TAL cores, based on pattern matching of volcanic peaks in high-resolution records of either sulfur (WDC) or sulfate (EDML, EDC, TAL). Extended Data Fig. 1b summarizes the various synchronizations, with previously published ones indicated with grey arrows, and the new ones presented here indicated with colored arrows.

Volcanic synchronization via sulfur/sulfate and/or ECM (electrical conductivity measurements) records is a commonly-used technique for Antarctic ice cores, and we rely on previously established methods described in detail elsewhere[31-35]. Matches are made by identifying sequences of sulfur peaks that have the same relative spacing in both cores[31-33,35]. Additional confidence in the match points comes from approximately proportional acid concentration levels, smooth variations in the resulting annual layer thickness between stratigraphic tie-points, and in some cases a distinctive shape of the common signals in the different ice cores. We identify 773 volcanic ties between WDC and EDML, 396 between WDC and EDC, and 425 between WDC and TAL (Source Data).

Stratigraphic matching was performed independently by two authors (M.Si. and M.Se.), and then compiled and compared by a third author (C.B.). Both analysts use an iterative approach, in which they first identify the major, unambiguous events. After marking these events (or clusters thereof), they re-align the data sets, and replot them with the marked events now overlapping. At this point, usually several of the smaller events are nearly on top of each other. These events are then marked, and the data is replotted with the newly marked events overlapping, et cetera.

We distinguish three cases: "doubly", "singly", and "inconsistently" identified events. Doubly identified matches are cases where both workers identify the same stratigraphic match between two cores for a given volcanic event (within a margin of a few cm). Singly identified matches are cases where only one of the two workers identified a stratigraphic match. Inconsistently identified matches are cases where a single volcanic event in one core is linked to two different volcanic events in the other core. Around 99% of the singly identified events were found by M.Si., demonstrating a difference in event detection threshold. For example, in the WDC-EDC synchronization the median volcanic peak sizes in WDC non-sea salt Sulfur (nssS) were 67.7 and 31.4 ppb for the doubly and singly identified matches, respectively, demonstrating that analyst M.Se. is more conservative in assigning match points. For comparison, the background nssS level is 15±4 ppb.

353 Here, all the doubly identified events are all assumed correct, and retained. In the (relatively
354 rare) case of an inconsistently matched event, the stratigraphic links suggested by both workers
355 are discarded to avoid ambiguity. Sequences of singly identified events are retained only if they
356 occur in between two doubly identified match points. In the case of an inconsistently identified
357 event (which is discarded), all singly identified matches adjacent to it are discarded also, until
358 another doubly identified match is encountered. In all three synchronizations, the vast bulk of
359 the singly identified matches come from the same worker, suggesting a difference in style and
360 event detection threshold between the two workers. We give some examples below of
361 hypothetical sequences of tie points, and how they are dealt with. Let "**d**", "**s**" and "**i**" denote
362 doubly, singly, and inconsistently identified tie points, respectively.

363 **d-s-s-s-d**: This is a hypothetical series of three **s** tie points in between two **d** tie points. Because
364 both workers agree on the tie points on either end, there is no reason to assume the **s** tie
365 points are incorrect; it simply reflects the fact that one worker (M.Se) is much more
366 conservative in assigning tie points than the other worker (M.Si). Therefore all tie points are
367 retained in the final synchronization (**d-s-s-s-d**). Singly identified ties are retained in such cases
368 irrespective of how many are in the series (for example a series of 10 **s** ties would be retained if
369 bracketed on either side by a **d**).

370 **d-i-d:** In this case the **i** tie is removed, but the **d** ties are retained resulting in the sequence **d-d**
371 in the final synchronization.

372 **d-s-s-i-s-d**: The **i** tie is removed in all cases. However, in this example the **s** tie points occur
373 adjacent to an **i** tie point, which casts doubt on their reliability. Therefore they are removed
374 together with the inconsistent tie, and only the sequence **d-d** is retained in the final
375 synchronization.

376 The matching is described below for the 10-61ka interval of interest (the WDC-EDC
377 synchronization extends back only to 57 ka BP).

378 In synchronizing WDC and EDC, the two workers identified 473 matches, out of which 103 were
379 doubly, 8 inconsistently, and 362 singly identified (all but two of which by M.Si.). The final
380 selection has retained the 103 doubly identified matches, and 293 of the singly identified
381 matches; 69 singly identified matches were discarded as they were bracketed on either side by
382 an inconsistent match. The 8 inconsistently identified matches all differed by less than 70 cm in
383 EDC (or about 65 years).

384 In synchronizing WDC and EDML, the two workers identified a total of 793 matches, out of
385 which 247 were doubly identified, 3 matches were inconsistently identified (all adjacent, and
386 not separated by doubly identified events), and 543 were singly identified (all by M.Si.). The
387 final selection has retained the 247 doubly identified matches, and 526 of the singly identified
388 matches; 17 singly identified matches were discarded as they were bracketed on either side by
389 an inconsistent match. The 3 inconsistently identified matches all differed by less than 1.6 m in
390 EDML (or about 65 years).

391 Talos Dome proved to be the most difficult of the cores to synchronize, presumably because
392 the layers are more strongly compressed in this intermediate depth core. The first attempt at
393 synchronization yielded 55 doubly and 5 inconsistently identified matches, with most of the
394 errors in the 770 to 905 m depth range (up to 5 m offsets in TAL). In light of this inconsistency,
395 the workers reviewed their volcanic ties throughout the core, with particular focus on the
396 problematic interval. In the revised synchronization, the workers identified a total of 437
397 matches, out of which 253 were doubly identified, 4 matches were inconsistently identified (all
398 adjacent, and not separated by doubly identified events), and 180 were singly identified (all but
399 9 by M.Si). The final selection has retained the 253 doubly identified matches, and 172 of the
400 singly identified matches; 8 singly identified matches were discarded. The 4 inconsistently
401 identified matches all differed by less than 75 cm in TAL (or about 75 years). So while the final
402 synchronization shows good agreement between both workers, we feel obliged to also report
403 the first, less successful attempt. The TAL depth range that is hardest to synchronize to WDC
404 (between 770 and 905 m TAL depth, or 14.9 to 25.2 ka BP) also yielded no matches to EDC in a
405 published study[32] (see yellow triangles at bottom of Extended Data Fig. 1a). Note that this
406 problematic 15-25 ka time interval does not influence the main results of this work – none of
407 the abrupt DO events used in the stacking lie in this interval (DO 2 is excluded from the stacking
408 because the absence of an abrupt CH_4 signal precludes precise synchronization to Greenland[15]).

409 The number of doubly and inconsistently identified events provides one way to assess the
410 reliability of the volcanic synchronization; the percentage of inconsistently identified events
411 (out of the pool of doubly and inconsistently identified events) ranges from 1% to 7% for the
412 three cores. Errors tend to occur in clusters of adjacent picks, due to the misidentification of a
413 sequence of peaks; seen in this light the WDC-EDML and WDC-TAL synchronizations each only
414 contain a single inconsistently identified sequence. The observed inconsistencies were always
415 relatively small, and on the order of a few decades. A second method of assessing the reliability
416 is via the redundancy offered by having multiple cores. Whenever three ice cores in Extended
417 Data Fig. 1b are connected via three independent synchronizations (i.e., whenever the arrows
418 form a triangle), this offers the possibility to test the internal consistency of the
419 synchronization. Over the age interval of interest (the last 61 ka) 213 ties had previously been
420 identified between EDML and EDC[36], as well as 102 ties between TAL and EDC. This introduces a
421 degree of redundancy that allows testing the internal consistency of the synchronization. EDC,
422 TAL and EDML are volcanically synchronized within the AICC2012 (Antarctic Ice Core
423 Chronology)[36], while WDC uses the independent WD2014 time scale[37,38]. For each volcanic tie
424 point, the difference in WD2014 age minus the AICC2012 age is shown in Extended Data Fig. 1a.
425 EDC and EDML are volcanically synchronized over the last 60ka (blue triangles), and therefore
426 the excellent agreement between the blue (WDC age minus EDML age) and red (WDC age
427 minus EDC age) curves in Extended Data Fig. 1a shows the volcanic framework to be internally
428 consistent. Likewise, for the period 25-42 ka BP and <13 ka BP, where TAL and EDC are
429 volcanically synchronized (yellow triangles), the yellow (WDC age minus TAL age) and red (WDC
430 age minus EDC age) curves agree well, suggesting internal consistency. Beyond 42 ka BP the

431 yellow curve deviates from the blue and red ones, suggesting the TAL core is imperfectly
432 synchronized within the AICC2012 chronology (due to an absence of volcanic ties at the time).

433 All ice cores used in this study were synchronized to the WAIS Divide WD2014 chronology[37,38].
434 For the four non-WDC cores, we start from their original ice core chronologies; this is the
435 AICC2012 chronology[36] for the EDML, EDC and TAL ice cores, and the DF2006 chronology for
436 the DF core[39]. For each core we add a time-variable offset to the WD2014 chronology that is
437 obtained using linear interpolation between the volcanic tie points. For the EDML, EDC and TAL
438 cores we have direct synchronizations to WDC as described above. For the Dome Fuji core,
439 synchronization was indirect via the EDC core (Extended Data Fig. 1b). Previously, 297 tie points
440 have been identified[33] between EDC and DF in the interval of interest (mean spacing of 173
441 years). These volcanic ties are transferred to WDC using the WDC-EDC synchronization
442 described above.

443 Over the time interval of interest, the offset with the AICC2012 cores (EDML, EDC, TAL) ranges
444 roughly from -330 years to + 430 years, with an average offset of -10 years (with negative
445 values meaning WD2014 is younger than AICC2012). For the DF core the range is from -230 to
446 1884 years, with an average of +739 years.

447 **Uncertainty in volcanic synchronization:** There are two types of uncertainty to consider. First,
448 the volcanic ties themselves may be incorrect, and second, in between the ties the
449 interpolation strategy introduces an error. The first type is difficult to quantify. Either the ties
450 have been correctly identified, and the relative age uncertainty is zero, or the ties are false, and
451 the relative age uncertainty is infinite (i.e., we have learned nothing). Past studies have
452 sometimes assigned Gaussian errors to volcanic tie points; while this is a practical necessity for
453 applications that optimize the fit to multiple age constraints[36,40-42], it does not reflect the true
454 uncertainty meaningfully and is not applied here.

455 We have high confidence in the correctness of the volcanic ties because of the internal
456 consistency of the new volcanic ties with previously published ones (Extended Data Fig. 1a), and
457 the fact that doubly identified ties greatly outnumber the inconsistently identified ones. We
458 have removed inconsistent matches from the synchronization, and we here assume the
459 remaining matches to be correct.

460 The second type of uncertainty is due to interpolation between volcanic ties. This introduces an
461 age uncertainty that depends on the age difference between adjacent ties, L. We estimate the
462 interpolation uncertainty using the layer-counted section of the WD2014 chronology, which
463 goes back to 31.2 ka BP. To estimate the interpolation uncertainty for two volcanic markers
464 that are, say, L=100 years apart, we can randomly pick thousands of 100-year intervals from the
465 WAIS Divide WD2014 chronology. Within each of these, the age evolution deviates from the
466 assumed linear interpolation; 1σ standard deviation of this age deviation is used as the
467 interpolation uncertainty. Typical results are shown in Extended Data Fig. 2a for several values
468 of L. We find that the maximum uncertainty scales as $\propto L^2$. Compared to East Antarctica, West

469   Antarctica receives a larger contribution of its snowfall from storm systems and synoptic
470   activity, making accumulation rates more variable[43,44]; the estimates given here should thus be
471   considered conservative when used in interior Antarctica. The volcanic interpolation
472   uncertainty for the four cores is plotted in Extended Data Fig. 2b. For DF, synchronization to
473   WDC is done via EDC as an intermediary core, and therefore the two synchronization errors are
474   added in quadrature.

475   In our synchronization we use both the singly and doubly identified tie points, and treat them
476   equally. We acknowledge, however, that the doubly identified ties are more reliable. Therefore,
477   we have repeated the analyses described in the main text using only the doubly identified tie
478   points (as opposed to both singly and doubly identified tie points). We find that the conclusions
479   of this work do not depend on this choice of tie points. Those tests are not shown here, but the
480   alternative chronologies that use only the doubly identified tie points, and alternative versions
481   of the manuscript figures showing those analyses were presented to the reviewers and are
482   available from the corresponding author upon request.

483   **Water stable isotope data:** A combination of previously published and unpublished ice core
484   water isotope data (δ^{18}O and δD) are used in this study. Deuterium excess (d_{ln}) is calculated
485   from the δ^{18}O and δD isotope ratios using the logarithmic definition of Uemura et al.[20]:

486    $$d_{ln} = \ln(1+\delta D) - 8.47 \ln(1+\delta^{18}O) + 0.0285\, [\, \ln(1+\delta^{18}O)\,]^2$$

487   For WDC we use previously published δ^{18}O and δD data[8,15,45], measured using laser
488   spectroscopy. Data have a typical depth resolution of 1m for the 0 to 2.3 ka BP interval, of 0.5
489   m for 2.3 to 56 ka BP, and of 0.25 m for 56-68 ka BP; this corresponds to an average time
490   resolution of 17 years for the study period (11 to 61 ka BP). Using the cm-scale CFA (continuous
491   flow analysis) record of WDC δ^{18}O instead gives identical results to those presented here[46].

492   For EDML we use previously published δ^{18}O and δD data[47,48], measured using conventional
493   IRMS (isotope ration mass spectrometry). Data have a typical depth resolution of 0.5 m,
494   corresponding to an average time resolution of 24 years for the study period.

495   For EDC we use previously published δ^{18}O and δD data[48,49], measured using conventional IRMS.
496   Data have a typical depth resolution of 0.55 m, corresponding to an average time resolution of
497   44 years for the study period.

498   For DF we use new and published water isotope data[39,50,51]. Two data sets are used. The first is
499   a data set of δ^{18}O measured using IRMS in the 300 to 1151 m depth range (10 to 71 ka BP) at
500   0.5 m resolution[50]. The second is a data set of δ^{18}O and δD measured using IRMS in the 550 to
501   849 m depth range (23 to 45 ka BP); this data set was measured at 0.1 m resolution, and
502   averaged into 0.5 m bins. Note that d_{ln} is only available from the second data set, which spans
503   DO/AIM (Antarctic Isotopic Maximum) evens 2 through 11. In the depth range of overlap, δ^{18}O
504   data from both data sets are averaged (after correcting the second data set by + 0.213‰ to

505　account for a calibration offset) to produce the final time series. The combined $\delta^{18}O$ record has
506　an average time resolution of 35 years for the study period.

507　For TAL we use a combination of new and previously published[10,52,53] data. Bag average, 1 m
508　resolution $\delta^{18}O$ and δD data measured using IRMS are available for the entire core. High-
509　resolution (0.1 m) $\delta^{18}O$ data measured using IRMS are available in the 598 to 786 m (10 to 16 ka
510　BP) and 1030 to 1282 m (37 to 65 ka BP) depth ranges. High-resolution $\delta^{18}O$ are averaged into
511　0.5 m bins, and combined with bag-average data for remaining depth intervals. The combined
512　$\delta^{18}O$ record has an average time resolution of 50 years for the study period.

513　**Stacking procedure:** The stacking procedure used to investigate the Antarctic climate response
514　to abrupt DO variability is described in detail elsewhere[15]. In short, the individual Greenland
515　events are aligned at the midpoint of their abrupt $\delta^{18}O$ transitions (either DO warming or DO
516　cooling). All Antarctic events (on their volcanically-synchronized WD2014 timescales) are
517　aligned at the midpoints of the abrupt WDC CH_4 transitions. We then average over events to
518　obtain the shared climatic signal; to derive the north-south phasing we use the independently
519　established CH_4 delay of 56 ± 19 years (1σ) behind $\delta^{18}O$ in Greenland[54].

520　For one of the abrupt events (DO-10 / AIM-10), improved inter-polar synchronization data are
521　available from [10]Be variations during the Laschamp event (41 ka BP) between the Greenland
522　NGRIP core, and the Antarctic EDC and EDML cores[27]; these timing constraints were
523　incorporated into our stacking procedure (DO-10 only). The EDML, EDC, DF and TAL ice cores
524　have much higher Δage uncertainty and lower resolution CH_4 records than WDC, therefore the
525　north-south phasing precision cannot be improved by considering CH_4 synchronization for
526　those cores also.

527　In this work we consider DO-0 (i.e. the YD-Holocene transition) through DO-16; DO-17 falls
528　outside the volcanic synchronization for the EDC and DF cores. DO-2 is omitted due to the
529　absence of a clear CH_4 response, precluding synchronization. All stacked records are shown in
530　Extended Data Fig. 3.

531　Uncertainty in the timing of the stacked records comes from these sources: (1) The gas age-ice
532　age difference (Δage) in the WDC core[37;] (2) DO midpoint detection in the abrupt NGRIP $\delta^{18}O$
533　record; (3) DO midpoint detection in the abrupt WDC CH_4 record; (4) Interpolation of the WDC
534　chronology between tie points; (5) The climatic lag of atmospheric CH_4 behind Greenland $\delta^{18}O$
535　of 56 ± 19 years (1σ)[54]; (6) Trend analysis in the BREAKFIT[55] and RAMPFIT[56] fitting routines; and
536　(7) volcanic synchronization onto the WD2014 chronology.
537
538　The combined uncertainty due to the first 5 items was assessed previously using a Monte-Carlo
539　routine, suggesting a 1σ timing uncertainty of 38 and 41 years in the WDC stacks for DO
540　warming and DO cooling, respectively[15]. The uncertainty in the trend analysis is given in
541　Extended Data Table 1. The uncertainty in the volcanic synchronization is shown in Extended

542 Data Fig. 2; averaged over the stacked events the 1σ synchronization uncertainty is 0 years at
543 WDC, 2 years at EDML, 6 years at EDC, 13 years at DF, and 2 years at TAL.

544 By stacking only the most prominent DO/AIM events (those following Heinrich events; i.e., DO-
545 4, 8, 12, 14 and 17) or just the minor ones (the remainder), we find that the magnitude of the
546 atmospherically-forced Antarctic response is larger for the former, suggesting proportionality
547 with the climate perturbation (Extended Data Fig. 4a-4f). Proportionality of the atmospheric
548 response is further seen for individual events (Extended Data Fig. 4g); see the figure caption for
549 details.

550 **Principal Component Analysis:** Principal component analysis (PCA) allows different climatic
551 modes to be identified in (paleoclimatic) time series from different locations[57]. Here we
552 perform PCA on the stacked δ^{18}O records at the 5 individual sites using the MATLAB function
553 "pca". The δ^{18}O stacks in the main manuscript combine 17 individual events – all those that fall
554 within the volcanic synchronization interval. To get a sense for the sensitivity to including or
555 excluding individual events, we perform additional experiments in which we stack only n events
556 (rather than all 17). The n events are randomly sampled without replacement (i.e. any given
557 event cannot be picked twice for each stacking). We then perform PCA of these stacked records
558 (the same events are stacked for each of the cores), and standardize the PC vectors by taking
559 the z-score (or standard score). Extended Data Fig. 5 shows typical results for n=2 and n=8,
560 where for each n we repeat the experiment 50,000 times to obtain reliable statistics. Extended
561 Data Figs. 5a and 5b show PC1 and PC2, respectively, with the solid line showing the mean of
562 50,000 experiments, and the shaded envelope the associated +/- 1σ standard deviation.
563 Extended Data Fig. 5c shows a histogram of the percentage of variance explained by each of the
564 modes.

565 We find that even by stacking as few as just n=2 events the method can, on average, identify
566 the oceanic and atmospheric components described in the main manuscript. Perhaps not
567 surprisingly, we find that when including fewer events the signal-to-noise ratio decreases: with
568 fewer events the estimated signal amplitude is smaller in both PC1 and PC2, and the
569 uncertainty envelope is wider. As more events are included in the stacking, the percentage of
570 variance explained by PC1 increases as the coherence between the various Antarctic cores
571 increases due to an improved signal-to-noise ratio.

572 Analysis in the main manuscript uses a 1500-year window (centered around t = 100 yr), which is
573 chosen because it corresponds to the recurrence time of the shortest DO cycles[58-61]. The
574 variance explained by the oceanic and atmospheric modes depends on the window length, as
575 shown in Extended Data Fig. 6a. For window lengths exceeding ~750 years, the "oceanic" mode
576 explains most of the variance (PC1), with the "atmospheric" mode explaining less. However, at
577 short window lengths (<750 year) the atmospheric mode explains most of the variance, making
578 it PC1. Comparing PC1 at 400-year window to PC2 at 2000-year window (Extended Data Fig. 6b)
579 illustrates the ability of the PCA to identify the atmospheric mode at different window lengths.

580   The cross-over behavior (i.e. the atmospheric mode shifts from being PC2 to PC1 as a function
581   of window length) is due to the fact that the signal variance of the step-like atmospheric mode
582   occurs chiefly within the $t = 0$ to $t = 100$ interval; signal variance of the oceanic mode depends
583   strongly on the window length, due to its gradual nature.

584   Lag time analysis of PC1 and PC2 using the RAMPFIT[56] and BREAKFIT[55] routines confirms the
585   cross-over behavior. At short window-length (<700 years), PC1 is characterized by the
586   instantaneous /fast response of the hypothesized atmospheric teleconnection, whereas at large
587   window-length (>800 years) it shows the 200-year delayed hypothesized oceanic
588   teleconnection. PC2 shows the opposite behavior, though note that for PC2 at 400-year
589   window length no meaningful solution can be found with either routine. At window lengths
590   >700 years the lag times remain stable and vary only within the uncertainty bound.

591   Unless specified otherwise, a 1500-year window is used in this work.

592   **Robustness of the atmospheric spatial pattern:**

593   The spatial pattern we identify for the atmospheric mode is one of the main results of this
594   work, and we here test its robustness. In Extended Data Figs. 6d—6f we compare three
595   alternative ways of identifying the spatial pattern; the Pearson correlation coefficient (r)
596   between the shown pattern and the pattern identified in the main text (EOF 2 at 1500 yr
597   window) is given for each. Details on the three methods are given in the figure caption.
598   Correlation coefficients ranging from $r = 0.92$ to $r = 0.999$ suggest that identification of the
599   atmospheric pattern is both qualitatively and quantitatively robust.

600   **Rotated PCA vectors:** PCA aims to explain the largest amount of variance, whereas our goal is
601   to distinguish between the oceanic and atmospheric modes. While PC1 is clearly dominated by
602   the 200-year-delayed oceanic bipolar seesaw (Fig. 2d), it appears that PC1 also captures some
603   abrupt warming around $t = 0$, apparently due to the fact that atmospherically-induced warming
604   is more prevalent over Antarctica than cooling (NH DO warming case). We therefore construct
605   (admittedly somewhat subjective) oceanic and atmospheric response functions (Fig. 2e), which
606   are derived from the principal components in the following way. Let PC1 and PC2 be the first
607   two principal components through time; these vectors are perpendicular. We let the
608   atmospheric response ATM be identical to PC2. The oceanic response OCE is found by rotating
609   the PC1 vector in the plane spanned by vectors PC1 and PC2 over an angle of -13 degrees:

610    $$\text{OCE} = \cos(-13°) \times \text{PC1} + \sin(-13°) \times \text{PC2}$$

611   The rotation angle of -13 degrees is picked such that the $\delta^{18}O$ shift at $t = 0$ is minimized. The
612   spatial pattern associated with the ATM response (Fig. 2h) is found by multiple linear regression
613   of the $\delta^{18}O$ stacks at the individual sites to OCE and ATM using the MATLAB function "regress".

614   **SAM, PSA and synthetic "atmospheric" indices.** The SAM and PSA indices were calculated from
615   climate model and reanalysis data as respectively the first and second mode of variability in

616   PCA/EOF analysis (MATLAB function "pca") of sea-level pressure (model) or 850-hPa
617   geopotential height (reanalysis) south of 20°S, after subtracting the long term mean and scaling
618   the anomalies by the square root of the cosine of latitude to account for the decreased surface
619   area closer to the poles. A synthetic index of the inferred atmospheric teleconnection is created
620   by projecting reanalysis SAT anomalies at the ice core locations onto the atmospheric loading
621   coefficients from Fig. 2h, the spatial signature of which is shown in Fig. 4d. Mathematically, this
622   projection is done by multiplying the SAT anomalies with the loading coefficients, and summing
623   over them at each monthly time step.

624   It is worth noting that all these inferences with respect to reanalysis data are reliant upon just
625   five atmospheric loading coefficients from the limited (from the perspective of the large-scale
626   circulation) spatial domain of Antarctica, so it is difficult to rigorously exclude non-SAM
627   atmospheric influences from the temperature pattern alone.

628   In Extended Data Fig. 9 we compare variability in the SH westerly winds in the glacial CCSM3
629   (Community Climate System Model version 3) climate simulations[25,62] to the ERA-Interim
630   reanalysis[30]. Both show greater variability in the Indian and Pacific sectors, relative to the
631   Atlantic sector. Compared to ERA-interim, the CCSM3 SAM is more zonal/annular in its
632   structure; the CCSM3 westerlies also appear to have smaller variability (some of this difference
633   could be due to comparing annual mean with decadal mean data). In CCSM3 the forced
634   response of the SH westerlies (at 19ka in response to increased North Atlantic freshwater, right
635   panel) is very similar to the internal variability of the SH westerlies (middle panel), suggesting
636   that the change in the SH atmospheric circulation induced by freshwater forcing in the North
637   Atlantic is analogous to the existing mode of internal variability.

638   To estimate the magnitude of SAM shifts of the DO cycle, we analyze the changes in central
639   East Antarctica where the signal is largest and most consistent with the present-day observed
640   relationship (Fig. 4). Using an isotope sensitivity of 0.8 ‰ K^{-1}, the atmospherically-induced
641   temperature anomaly in central East Antarctica (DF, EDC) is in the range of 0.20°C to 0.45°C
642   during a DO-warming (lower and upper bound reflecting typical minor and major DO/AIM
643   events, respectively). Regression of ERA-interim 2m temperature at DF and EDC to the SAM
644   index shows a slope of around -1.2°C per unit of normalized SAM (the normalized SAM index
645   time series has a standard deviation of 1), implying a shift in SAM index of around 0.2 to 0.4
646   normalized (modern-day) SAM units (rounded to one significant figure). This estimate assumes
647   (1) a linear isotope-temperature response using the modern-day spatial slope, and (2) that the
648   monthly SAM-SAT regression from monthly internal SAM variability also applies to persistent
649   SAM shifts during the glacial; both assumptions are subject to uncertainties that we do not
650   address here. The CCSM3 model simulates a persisting SAM shift of the same magnitude as the
651   internal SAM variability in decadal averaged data (Fig.3); because internal SAM variability will
652   be larger in monthly than in decadally-averaged data, the model and data-based estimates may
653   be in agreement. The reanalysis time period is too short to derive robust estimates of decadally
654   averaged SAM variability.

655 **GCM simulations**. We used the TraCE-21k transient climate model simulations done with the
656 Community Climate System Model version 3 (CCSM3)[25,62-65]. AMOC collapse and resumption are
657 triggered using freshwater forcing in the North Atlantic. The "DGL19k" run is used for the
658 AMOC collapse, the "DGL-overshoot-C" run for the AMOC resumption case[64].

659 **Moisture origin analysis**. We use two separate experiments that trace moisture origin of the
660 precipitation at the coring sites.

661 The first method uses a Lagrangian moisture source diagnostic[66] based on a previously
662 published data set[67]. Using the winds, temperature and humidity of the ERA-Interim reanalysis
663 data set[30] covering the years 1980-2013, 5 million air parcel trajectories have been calculated
664 covering the global atmosphere at 6-hourly resolution using the Lagrangian particle dispersion
665 model FLEXPART[68]. From the analysis of specific humidity changes over time along the air
666 parcel trajectories, moisture sources have been identified whenever specific humidity in the air
667 parcels increased more than a threshold value of 0.1 g kg^{-1} over a 6-hour period. The fractional
668 contribution of each moisture source to the final precipitation at the target location (an area of
669 300×300 km centered on each ice core site) is obtained from calculating the amount
670 contributed by a moisture source to the humidity already present in an air parcel. During
671 precipitation en route, the previous sources' contributions are proportionally discounted. This
672 results in a quantitative estimate of the contribution of surface areas to the precipitation in the
673 target region in units of evaporation (water mass per unit area per time), including their
674 position in terms of latitude and longitude. The moisture source contributions for each site and
675 precipitation event have been composited into mass-weighted annual mean values and scaled
676 with respect to the total amount of water deposited at the target region.

677 The second method uses water tagging in a 50-year simulation with the Community
678 Atmosphere Model (CAM), with prescribed seasonally varying SST and modern boundary
679 conditions. The experiment is set up to evaluate the meridional moisture source distribution,
680 with further details and figures given in Ref. [8]. In short, evaporation is tagged in 11 bins. One bin
681 is the Antarctic continent (re-evaporation) and ocean and ice shelfs south of 70°S; 10 are zonal
682 bins of 5° latitude each, ranging from 20°S to 70°S. For each core, the moisture source
683 distribution is found by evaluating the relative contributions from each of the bins for the last
684 30 years of the run. Two moisture source distributions were created, one for all year in which
685 the mean annual SAM index was positive, and one for all years in which it was negative
686 (Extended Data 8b).

687 **Change point detection**. We use two well-documented and widely-used change-point detection
688 methods: BREAKFIT[55] and RAMPFIT[56], with results given in Extended Data Table 1. The choice of
689 which one to use is based on the shape of the time series $x(t)$. BREAKFIT finds a single change-
690 point, and fits a linear slope on either side; these features make it suitable for the oceanic
691 mode / PC1 in the main manuscript. RAMPFIT finds two change-points; it is assumed $x(t)$ has
692 constant value x_1 for $t<t_1$, then a ramp up or down until it reaches value x_2 at time t_2, after
693 which it remains constant at value x_2 for $t>t_2$. These features make RAMPFIT suitable for the

atmospheric mode / PC2 in the main manuscript. To find the two change-points in the d_{ln} stacks, we applied both the RAMPFIT algorithm, and the BREAKFIT algorithm twice (once for each change-point). The results are comparable (Extended Data Table 1), and in the main text we report the average of both methods.

Methods References:

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794

Data Availability. Source Data (WDC sulfur data, volcanic tie points and water isotope data on synchronized chronologies) and derived products (stacks, PCA results, etc.) are available with the online version of the paper, and via the NOAA paleoclimate data archive.

798

Code Availability. MATLAB code used for the stacking procedure can be found in the supplementary information of ref.[15], and is available from the corresponding author upon request.

802

Extended Data Figure 1 | Volcanic synchronization of Antarctic ice cores. a, Age offset between the WD2014[37,38] (WDC) and AICC2012[36] (TAL, EDML, EDC) age scales, with each dot representing a volcanic tie point. Yellow and blue triangles denote the timing of TAL-EDC and EMDL-EDC volcanic ties[32,36], respectively. **b,** Overview of synchronizations between the ice cores used in this study. Previously published synchronizations are in grey; synchronization presented here are in color. Synchronizations within Antarctica are based purely on volcanic links; synchronization between WDC and NGRIP (Greenland) are based on atmospheric CH_4 (green arrow).

811

Extended Data Figure 2 | Age uncertainty due to volcanic synchronization. a, Interpolation uncertainty (1σ) for 4 different values of *L* (the spacing between two adjacent volcanic tie points), based on the layer-counted WD2014 age scale[38]. **b**, Interpolation uncertainty in synchronizing the 4 ice cores to WAIS Divide. Grey vertical lines give the timing of DO events.

816

Extended Data Figure 3 | Site-specific stacks of δ^{18}O and d_{ln}. a, Stack of NGRIP δ^{18}O (teal, left axis) and WDC CH_4 (green, right axis) during DO warming. **b**, As in **a**, but for DO cooling. **c**, Stack of Antarctic δ^{18}O at indicated locations (see color legend) during DO warming. **d**, As in **c**, but for

820   DO cooling. **e**, Stack of Antarctic d_{ln} at indicated locations (see color legend) during DO

821   warming. **f**, As in **e**, but for DO cooling. Isotope ratios are on the VSMOW scale.

822

823

Extended Data Figure 4| Proportionality of the atmospheric response. Panels **a-f** compare stacks of just the major DO/AIM events (those following Heinrich events, namely DO/AIM0, DO/AIM1, DO/AIM4, DO/AIM8, DO/AIM12 and DO/AIM14, left panels), and just the minor DO/AIM events (the remainder, right panels). **a,b,** stacks of NGRIP $\delta^{18}O$ (left axis) and CH4 (right axis). **c,d,** Stacks of Antarctic $\delta^{18}O$ at indicated locations. **e,f,** Same as panels **c** and **d**, but with Antarctic mean subtracted. **g,** Proportionality of the atmospheric response for individual events (numbered). NGRIP event size is found via regression of individual NGRIP events to the multi-event NGRIP $\delta^{18}O$ stack normalized to unit variance (Fig. 2a). The Antarctic atmospheric response is found via multiple linear regression of single-site, individual events to the atmospheric and oceanic modes (Fig. 2e). Shown are average (dots) and standard deviation (grey error bars) of the response at EDC, DF and EDML (the cores with the strongest atmospheric response); the EDML response is multiplied by -1 as it has the opposite sign of the response at DF and EDC. Red and blue dots denote the major and minor DO/AIM events, respectively. Isotope ratios are on the VSMOW scale.

838

Extended Data Figure 5 | Reducing the number of events in the $\delta^{18}O$ stacks. a, PC1 when stacking 2 or 8 randomly selected events; thick line and shaded area represent the mean and +/- 1σ standard deviation of 50,000 runs, respectively. Vertical yellow bars denote the 200-year period after the abrupt DO event at $t = 0$. **b,** as panel **a**, but for PC2. **c,** Histogram of percentage of signal variance explained by PC1 and PC2, when stacking 2 or 8 randomly selected events. Color coding as in panels **a** and **b**.

845

Extended Data Figure 6| Robustness of the atmospherically-forced warming pattern. a, Principal Component analysis as function of window length, with percent variance explained by PC1 and PC2. **b,** Comparison of PC1 at 400 year window length, to PC2 at 2000 year window length, to show the cross-over of the atmospheric response (i.e. from PC1 to PC2) as a function of window length. **c,** Lag time of the Antarctic PC1 and PC2 response as a function of window length, assessed using BREAKFIT[55] and RAMPFIT[56] routines (see Methods for explanation of which one is used). **d,** EOF1 at 400 year window length scaled to permil units. **e,** EOF2 at 2000 year window length scaled to permil units. **f,** Slope of linear fit to $\delta^{18}O$ stacks in the $t = 0$ to $t =$ +200 year interval, shown as the change in permil during the 200 years. Isotope ratios are on the VSMOW scale.

856

Extended Data Figure 7 | The Antarctic climate response to DO cooling. a, Stack of NGRIP $\delta^{18}O$. **b,** Stack of Antarctic $\delta^{18}O$ at indicated locations. **c,** As in **b**, but with Antarctic mean subtracted. **d,** First two principal components of the Antarctic $\delta^{18}O$ stacks, with percentage of

860   variance explained (offset for clarity) Lines show the BREAKFIT (PC1) and RAMPFIT (PC2) fits. **e-**
861   **f**, Empirical Orthogonal Functions EOF1 and EOF2 associated with PC1 and PC2 in **d**, scaled to
862   show the magnitude in permil. Isotope ratios are on the VSMOW scale.

863

864   **Extended Data Figure 8 | Moisture sources of Antarctic ice core and the SAM. a,** Mass-
865   weighted probability distribution functions (PDFs) of Antarctic moisture sources for the five ice
866   cores of interest (5×10^{-5} deg^{-2} contour lines; the area integrated PDF equals 1). Distributions are
867   calculated from reanalysis data[30,67] using a Lagrangian source diagnostic[21,66] (Methods).
868   Parallels are plotted in 15° increments of latitude; meridians in 45° increments of longitude. **b**,
869   Sea surface temperature (SST, black)[69] and relative humidity (RH, grey)[70] as a function of
870   latitude. The colored curves give the latitudinal source distribution during negative SAM phase
871   (solid curves, SAM index < 0) and positive SAM phase (dashed curves, SAM index > 0). The solid
872   and open dots show the first moment of the source distribution during negative and positive
873   SAM phase, respectively. Note that during a positive SAM phase moisture sources for all core
874   locations are located closer to the Antarctic continent. Source distribution data were obtained
875   using water tagging experiments[8] in the Community Atmosphere Model (Methods).

876

877   **Extended Data Figure 9 | Southern Annular Mode-like variability in zonal near-surface winds.**
878   **a,** ERA-Interim reanalysis (1979-2016 annual means) zonal wind speed at 10 m height regressed
879   onto SAM index (here the first principal component of sea level pressure variability south of
880   20°S), expressed in units of m s^{-1} per standard deviation in the index. **b**, Same as panel **a**, but for
881   internal variability in the CCSM3 SynTraCE model simulation[25,62] during glacial climate prior to
882   freshwater forcing of Heinrich stadial 1 (19.5ka to 19.01ka BP, decadal means). **c**, Same as panel
883   **a**, but for the response to North Atlantic freshwater forcing in the CCSM3 SynTraCE model
884   (19.1ka to 18.9ka BP, decadal means, with the freshwater forcing applied at 19ka).

885

886   **Extended Data Table 1 | Change-point analysis on Antarctic response.** Change-points are
887   found using the BREAKFIT[55] or RAMPFIT[56] routine as indicated; the parameter t_{OCE} is the single
888   change-point of the oceanic mode; t_{ATM1} and t_{ATM2} are the two change-points of the
889   atmospheric mode, representing the onset and ending of the shift, respectively (Methods).
890   Stated uncertainties give the 1σ value in the fitting routine only, found using a Monte Carlo
891   moving block bootstrap analysis with 1000 iterations[55,56]; the full uncertainty in the combined
892   interpolar CH_4 synchronization and stacking procedure[15] is estimated to be around 40 years
893   (1σ), which can mostly be attributed to uncertainty in the WAIS Divide ice age-gas age
894   difference Δage[37]. The $\delta^{18}O$ PC2 and $\delta^{18}O$ atmospheric modes are identical.

Figure axes and labels:

- Top axis / bottom axis: WD2014 age (ka BP), ranging 60 to 10
- a) Summit δ18O (‰): −36, −38, −40, −42
- b) WDC CH4 (ppb): 400, 500, 600, 700
- c) Antarctic d_ln (‰): −2, −4, −6, −8, −10, −12
- d) Antarctic δ18O (‰): 2, 0, −2, −4, −6, −8 (EDML, Plateau)

Heinrich Stadials labeled: HS5, HS4, HS3, HS2, HS1

DO: 16 15 14 13 12 11 10 9 8 7 6 5.2 5.1 4 3 2 1 0