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Antarctic and global climate history viewed from ice cores

Edward J. Brook[1]* & Christo Buizert[1]

A growing network of ice cores reveals the past 800,000 years of Antarctic climate and atmospheric composition. The data show tight links among greenhouse gases, aerosols and global climate on many timescales, demonstrate connections between Antarctica and distant locations, and reveal the extraordinary differences between the composition of our present atmosphere and its natural range of variability as revealed in the ice core record. Further coring in extremely challenging locations is now being planned, with the goal of finding older ice and resolving the mechanisms underlying the shift of glacial cycles from 40,000-year to 100,000-year cycles about a million years ago, one of the great mysteries of climate science.

The origins of the Antarctic ice sheet are in late Eocene–early Oligocene time, about 34 million years ago; the last major phase of growth began at about 14 million years BP (before present)[1,2]. Coring in the present ice sheet (Box 1) has recovered records up to 800,000 years (800 kyr) old and there are hopes of extending the continuous record further. New studies on the ice sheet margin have provided unique but discontinuous samples of older ice[3]. Owing to the flow of ice from the interior to the ice sheet margin, ice deposited during the earliest history of the ice sheet is not likely to have been preserved, but the maximum age of extant ice is unknown.

The Antarctic ice sheet preserves a history that links Antarctica to the rest of Earth. Richly detailed and uniquely preserving the composition of the atmosphere, the record in the ice underpins much of global climate change research. Drilling and recovering long ice cores (Box 1) require specialized engineering owing to the low temperatures and the high pressures inside the ice sheet, the need to preserve the ice for analysis, and the remote locations of deep-drilling sites. The stratigraphically continuous dataset from cores in the dry ice sheet interior that extends to 800 kyr is complemented by more detailed, but younger, datasets from coastal sites with higher snowfall rates (Box 1). Collectively, Antarctic ice cores provide fundamental insights into the nature of global climate cycles driven by orbital variations, internal climate variability on sub-orbital timescales, changes in global biogeochemical cycles, Antarctic climate dynamics, abrupt climate change, and a host of other topics.

One of the strengths of the ice core record is that a wide variety of parameters, reflecting different aspects of the Earth system, can be measured in great detail in the same core. The isotopic composition of the ice is a proxy for local temperature[4], while the chemical composition records the input of dust, sea salt, volcanic material, pollutants, other aerosol material, and even extraterrestrial dust[5–8]. Past snowfall rate, a fundamental climate parameter, can be derived from layer counting and other age constraints[9]. The temperature of the ice sheet itself retains a memory of past climate and can be measured in the ice core borehole[10]. Gradual compaction of the firn converts snow to ice, trapping small samples of the atmosphere in a matrix that is remarkably resistant to gas loss. The trapped air provides the only direct record of changes in atmospheric composition prior to modern atmospheric measurements.

Over the last decade, drilling and analysis of Antarctic ice cores have uncovered a large amount of new information. This paper reviews the myriad ways in which these data show how connected Antarctica is to the rest of the world. Deep ice cores from the Antarctic interior show that on long timescales Antarctic climate closely follows variations in solar insolation that drive climate change globally and that it exhibits major temperature changes at the termination of ice ages. New results from high-resolution cores in high-accumulation regions provide unprecedented detail about the millennial-scale climate 'seesaw' between Antarctica and the Northern Hemisphere, a signature of variations in ocean heat transport related to shifts in Atlantic Ocean circulation. The dust content of ice cores reveals enhancements in dust flux in cold climates, with possible contributions to ocean productivity. Greenhouse gas data from trapped air show how global biogeochemical feedbacks contribute to climate change on long and short timescales, and that climate and greenhouse forcing are extremely tightly coupled (Box 2).

The long view: ice age cycles

Two deep ice cores from low-accumulation regions of the East Antarctic Plateau, one at Dome Concordia[11] (Dome C) and a second at Dome Fuji[12] (Dome F, or Valkyrie Dome) provide now-iconic records of multiple glacial cycles as far back as 800 kyr and 720 kyr BP, respectively (Fig. 1). These results complement the pioneering 420-kyr dataset from the Vostok ice core[13]. Viewed broadly, almost all climate-related parameters in these records show major, synchronous variations across the glacial cycles, which have an average duration of about 100 kyr. The Antarctic records closely mirror other global environmental proxies, most notably the oxygen isotopic composition of benthic foraminifera in deep-ocean sediments (Fig. 1g), which is widely used as an index for global ice volume and glacial–interglacial conditions[14]. All Antarctic records show a characteristic sawtooth pattern on these timescales, with a gradual cooling trend from glacial inception to peak glacial conditions, followed by a relatively fast glacial termination.

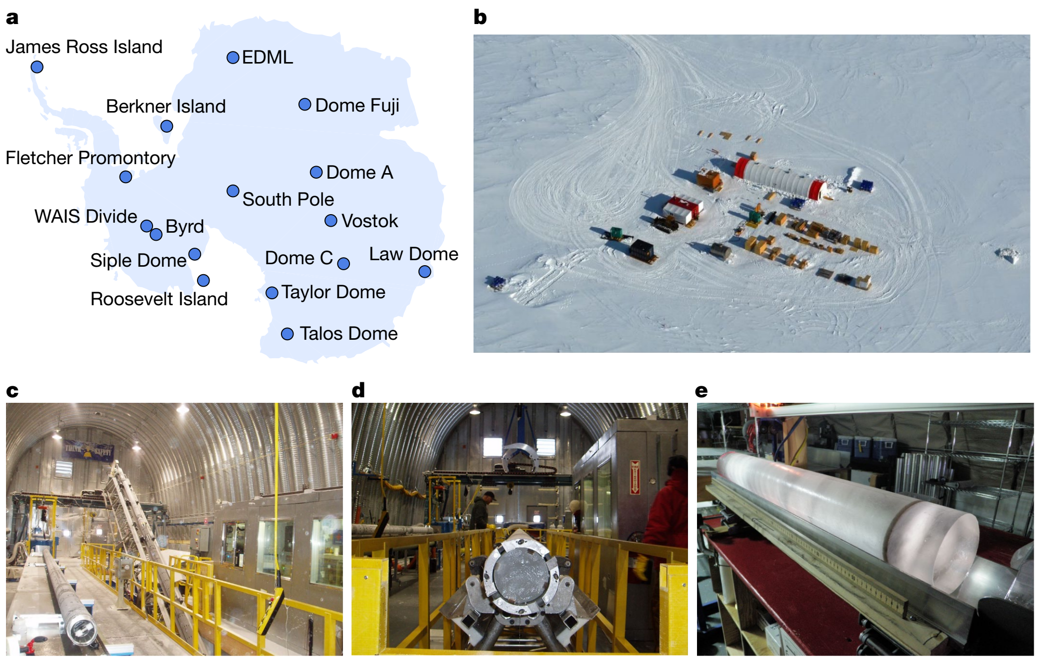
The stable water isotope ratios (δD or $\delta^{18}O$) traditionally used as temperature proxies suggest that the amplitude of East Antarctic temperature change over the glacial cycles ranges from about 6 °C to 13 °C, with the largest-amplitude oscillations[11] during the last 450 kyr. Ice isotope records are an indirect measure of temperature, however, and are influenced by other phenomena related to moisture transport and deposition. The ice sheet has a long thermal memory, and in locations with thick ice and high accumulation rate borehole thermometry can be used to circumvent these problems and estimate glacial–interglacial temperature change. Borehole-based temperature

[1]College of Earth, Ocean, and Atmospheric Sciences, Oregon State University, Corvallis, OR, USA. *e-mail: brooke@geo.oregonstate.edu

Box 1

Ice core drilling

The science of ice core drilling originated in the 1950s and the first deep core in Antarctica was at Byrd Station in 1968, during the International Geophysical Year[111]. Since then, numerous deep-coring projects have been completed (blue circles in **a** in the figure). The longest core, at Vostok Station, reaches 3,700 m below the ice surface. The oldest so far, the EPICA Dome C ice core, extends to 800,000 years[11,112]. The technology for deep ice coring has gradually evolved, with recent developments in larger volume and replicate coring drills[113], and new access tools that will allow quick sampling of deep ice sections, coring within bedrock and analysis in situ[101–103]. National archive facilities retain samples from all deep ice cores, preserving a unique resource for the scientific community. At an ice core drilling camp at the South Pole (**b**), the long arch structure contains the drill. The tipping tower of the US Deep Ice Coring Drill is shown (**c**). An ice core section in the WAIS Divide Drill (**d**) is shown immediately after a drilling run. A core section from the WAIS Divide site (**e**) shows a visible volcanic ash layer. The core is 12 cm in diameter. Image credits: **b**, US National Science Foundation (http://spicecore.org/photos.shtml); photographs in **c** and **d** were taken by Jay Johnson and in **e** by Heidi Roop.



reconstructions at the WAIS Divide site in west Antarctica[10] indicate a glacial–interglacial temperature change of 11.3 ± 1.8 °C for the last termination, consistent with estimates based on stable water isotopes alone. Globally, glacial–interglacial temperature change has been estimated[15] at about 3.5 °C; the higher Antarctic values confirm the theory that polar temperatures change more than tropical temperatures do (polar amplification)[10].

It is understood that the glacial cycles are paced by variations in Earth's orbit[16], with insolation changing due to orbital eccentricity (with an approximately 100-kyr period), axial tilt (with an approximately 41-kyr period) and precession of the equinoxes (with an approximately 19–23-kyr period). In terms of radiative forcing at the top of the atmosphere, the insolation changes by themselves are too weak to drive global temperature changes, and feedbacks such as changes in greenhouse gases and the high albedo of extensive (Northern Hemisphere) glaciation are required to explain the observed climate variations[10].

Several challenges remain to our understanding of glacial–interglacial dynamics, foremost of which is the lack of a clear explanation for the apparent 100-kyr periodicity of the glacial cycle, given that the 100-kyr eccentricity cycle produces negligible variations in either seasonal or annual radiative forcing (the so-called '100-kyr problem')[17–19]. Closer inspection shows that individual cycles are not uniform in length; accurate U/Th dating of speleothems suggests that glacial terminations are actually spaced by four to five precession cycles, supporting the theory that changes in Northern Hemisphere insolation driven by precession are the predominant driver of glacial terminations[20]. Several models have been put forward to 'predict' which Northern Hemisphere insolation maxima lead to glacial terminations and which do not[19,21]. Most proposed solutions to the 100-kyr problem rely on the inertia of gradually growing Northern Hemisphere ice sheets that have the ability to survive several insolation maxima, typically with a nonlinear response of ice volume to insolation[17,19,22–24].

A second challenge concerns the remarkable coherence of Antarctic temperature variability and global climate change on orbital timescales, although orbital forcing due to precession acts with opposite effect in each hemisphere[25]. The canonical view holds that global climate responds to summer solstice insolation at 65° N, the latitude band of the large Northern Hemisphere ice sheets. Indeed, accurate dating of the Dome Fuji ice core confirms that Antarctic orbital-scale climate change follows Northern Hemisphere insolation closely, with the largest warmings concurrent with rising Northern Hemisphere summer insolation[26].

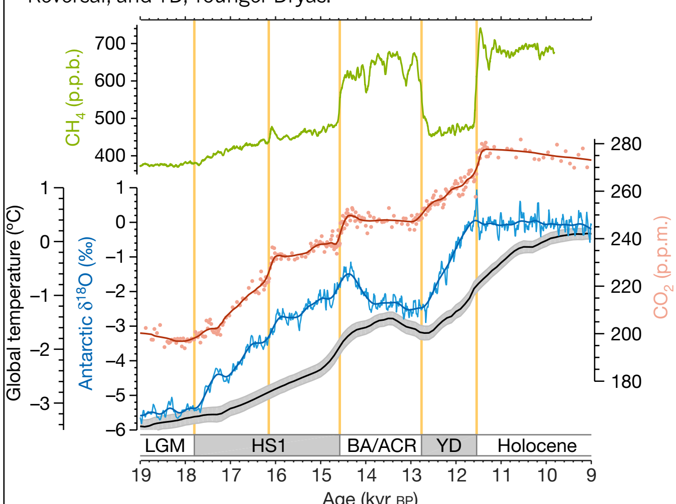
The most obvious mechanism to explain this bi-hemispheric coherence is through the globally well mixed greenhouse gases, although their link to Northern Hemisphere insolation remains incompletely understood. Rising Northern Hemisphere insolation could further drive enhanced Laurentide ice sheet meltwater runoff into the North Atlantic to reduce the Atlantic overturning circulation, which would in turn warm Antarctica via the so-called 'bipolar seesaw' (see section 'The close view: millennial-scale variability'), suggesting a role for ocean circulation in synchronizing the hemispheres[5,25,27]. Alternatively, it has been suggested that synchroneity at precession frequencies arises from the fact that Antarctic temperature is sensitive to local summer duration (which changes in-phase with the Northern Hemisphere

Box 2

CO_2 and temperature phasing during the last deglaciation

The question of the relative timing of greenhouse gas and (Antarctic) temperature change at the glacial terminations has generated substantial interest. Evaluating this phasing is complicated by the difference in age between the ice and the gas trapped inside it. Early studies suggested a substantial (600- to 1,200-year) lag of the CO_2 concentration rise behind Antarctic warming[114–117]. More recent work (see figure), based on high-accumulation cores and improved treatment of firn compaction, finds CO_2 and Antarctic temperature to be more or less synchronous (within uncertainty) for the last two glacial terminations[84,118,119]. The close relationship between Antarctic temperature and CO_2 obviously reflects important feedbacks and interactions between the global carbon cycle and the climate system. The early studies suggesting a long lag of CO_2 increase behind Antarctic warming have been misinterpreted as proof of a negligible warming effect of greenhouse gases; this is incorrect because (1) the onset of Antarctic warming is driven by changes in interhemispheric heat exchange, rather than by CO_2, and (2) the global temperature rise lags CO_2, rather than leads it[15]. However, what then does this phasing tell us? One interpretation is that the close phasing reflects the dominant role of Southern Ocean ventilation in setting atmospheric CO_2 levels[119]. Alternatively, the rise in both CO_2 levels and Antarctic temperature may be caused by reduced North Atlantic Deep Water formation, in which case their synchroneity reflects a common driver, rather than interdependence. Much as in Douglas Adams' *The Hitchhiker's Guide to the Galaxy*, although the answer to our question is now apparent, precisely what it signifies remains to be revealed.

The figure shows atmospheric CO_2 change from the WAIS Divide (Antarctica) ice core for the period 19–9 kyr ago, global temperature reconstruction[15], the east Antarctic oxygen isotope stack (the water $^{18}O/^{16}O$ isotope ratio anomaly relative to the present)[119], and the atmospheric CH_4 record from the WAIS Divide ice core[92]. Vertical yellow bars show the timing of major inflection points in the CO_2 record. Grey shading around the black trace indicates uncertainty in the temperature reconstruction. LGM, Last Glacial Maximum; HS1, Heinrich Stadial 1; BA, Bølling–Allerød; ACR, Antarctic Cold Reversal; and YD, Younger Dryas.



summer solstice insolation intensity in the precession band), rather than summer peak insolation (which changes out-of-phase with the same)[28]. Although local insolation may play a role at some ice core sites[29], the precise timing and magnitude of temperature variations suggest that the bipolar seesaw and greenhouse gas variations are the dominant influences on orbital-scale Antarctic climate.

A third challenge is that the amplitude of the glacial cycles in Antarctic temperature, CO_2 and ice volume increased around 450 kyr BP, at the so-called Mid-Brunhes event, through an intensification of the interglacial climates (Fig. 1). Atmospheric CH_4 and other markers of tropical hydrology appear to be only weakly affected by this transition[20,30]. It is debated whether the Mid-Brunhes is truly a single event, or whether this apparent transition simply emerges from the orbital forcing without changes in the underlying coupling between insolation and climate[19,21,31]. Ocean sediments show that the current '100-kyr world' was preceded by a '41-kyr world' in which glacial cycles were paced by Earth's axial tilt[14]. The transition between these two occurred between 1,200 kyr and 800 kyr BP, a period not covered in the ice core record. A major goal of the international ice coring community is to recover a continuous ice core through this key transition (see section 'The future of Antarctic ice core science'), primarily to examine changes in greenhouse forcing and the temporal patterns of Antarctic climate.

In addition to the palaeotemperature data, the long ice cores provide co-registered records of the flux of dust, sea salt and other atmospheric aerosols, and the concentrations of long-lived atmospheric gases (Fig. 1). Glacial periods are characterized by much higher levels of mineral dust deposition (Fig. 1c and d)[32,33], commonly attributed to changes in both source strength and atmospheric transport, with the former term dominating[34]. Climate-driven changes in Southern Hemisphere dust source regions (Patagonia and possibly Australia) include more exposure of sources due to aridity or glaciation, and increases in wind strength[32]. Changes in iron delivery to the ocean from dust[32] are probably involved in changes in ocean productivity and atmospheric CO_2 levels[35,36]. There has been considerable interest in developing an (aerosol-based) ice core tracer of past sea ice extent, given the important role of sea ice in the climate system. Initially promising candidates include sea salt sodium and methanesulphonic acid. Further work suggested that quantitative interpretations are difficult because of a variety of confounding effects in the transport and production of both tracers[37].

The three major greenhouse gases (carbon dioxide (CO_2), methane (CH_4) and nitrous oxide (N_2O)) also show large variations on the same timescale as ice volume and Antarctic climate over the glacial cycles (Fig. 1e and f)[30,38,39]. The largest variations are systematically associated with the glacial terminations, although the imprint of axial tilt and precession is also evident. These greenhouse gas changes act as a positive feedback on the orbitally paced glacial cycles and account for about 40% of the glacial–interglacial change in Earth's radiative balance[10].

Atmospheric CO_2 concentrations range from roughly 170 to 300 parts per million (p.p.m.) over the glacial–interglacial cycles (Fig. 1e). In addition to their impact on Earth's radiative budget, CO_2 variations reflect global carbon cycling and climate–biosphere interactions. A satisfactory accounting for the full magnitude of the glacial atmospheric CO_2 reduction is still lacking[40,41], although it is fairly clear that multiple processes operated. Changes in carbon storage on land and changes in CO_2 solubility in the ocean are obvious players, but the sum of these effects results in little net change[40]. Therefore, a higher glacial oceanic carbon inventory must also be involved and both physical and biological processes in the ocean probably contributed.

The Southern Ocean is a key region of interest. The dense, deep water mass originating around Antarctica that fills the abyssal oceans occupied more volume during the last glacial period and was more isolated and poorly ventilated, allowing it to store large quantities of respired carbon[42–44]. Increased efficiency of the biological pump, for example, through iron fertilization[45] or Southern Ocean stratification[46], could further draw down atmospheric CO_2. Wind-driven upwelling in the Southern Ocean brings deep waters rich in respired carbon near the surface, allowing carbon exchange with the atmosphere; low nutrient utilization in the Southern Ocean's surface makes this CO_2 'leak' to the atmosphere more effective[41,46]. Processes that may control the deglacial

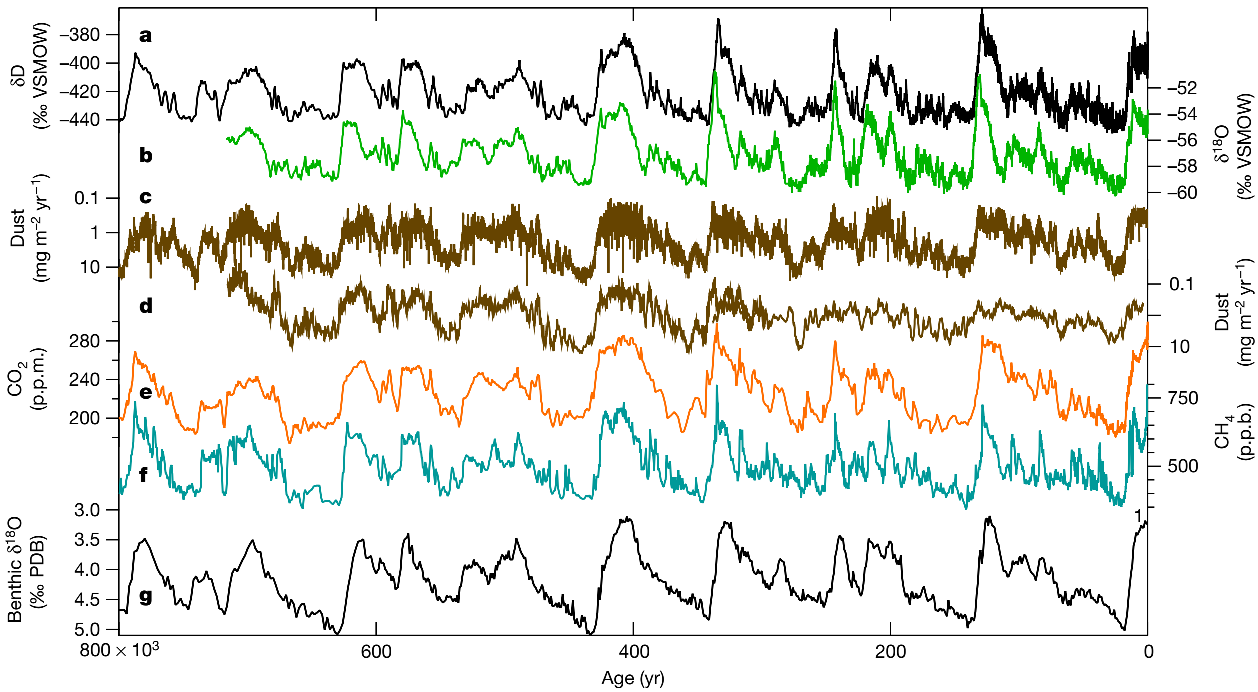


Fig. 1 | Data covering the last 800,000 years from long Antarctic ice core records, and the benthic isotope stack, a proxy for global glacial–interglacial cycles, with upward direction corresponding to warm interglacial conditions. a, The EPICA Dome C δD ($^2H/^1H$ isotopic ratio of water)[11]. VSMOW, Vienna Standard Mean Ocean Water. **b**, The Dome Fuji $\delta^{18}O$ ($^{18}O/^{16}O$ isotopic ratio of water)[12]. **c**, **d**, The EPICA Dome C[81] and Dome Fuji[12] dust records. **e**, **f**, The EPICA Dome C/Vostok CO_2[109] and CH_4[30] records. **g**, Benthic oxygen isotope stack[14]. PDB, Pee-Dee belemnite standard.

release of carbon through the Southern Ocean include the extent of Antarctic summer sea ice[42] or the strength and position of the Southern Hemisphere westerly winds[47]. The atmospheric carbon isotopic composition measured in ice cores is consistent with deglacial ventilation of respired carbon from the deep ocean[48]. The uniformity of glacial minimum CO_2 levels is an interesting challenge, suggesting a consistent negative feedback, the nature of which is not yet clear[49].

CH_4 rose by about 300 parts per billion (p.p.b.) at glacial terminations, with smaller changes during glacial cycles linked to insolation variations driven by orbital tilt and precession[30] (Fig. 1f). On millennial timescales, CH_4 is tightly coupled to Dansgaard–Oeschger events (see section 'The close view: millennial-scale variability'), although the response appears to be modulated by insolation[50]. A number of factors have been invoked to explain CH_4 variations on these timescales, including changes in emissions from wetlands (the major modern source), release of CH_4 from sea floor hydrates or permafrost, and changes in the atmospheric sink (primarily hydroxyl radical). Isotopic tracers and modelling do not support a dominant role for the last two factors[51–55]. Instead, CH_4 variations are consistent with changes in tropical and mid-latitude hydroclimate, driven by both insolation and shifts in the position of the Inter-Tropical Convergence Zone (ITCZ) during abrupt climate events. Additional boreal contributions may have existed during interglacials when Northern Hemisphere ice sheets retreated[56]. The ice core record contains no indications of very large bursts of CH_4 during interglacial periods, even during those warmer than the Holocene, suggesting that 'CH_4 time bomb' scenarios of a runaway positive carbon cycle feedback for the Arctic are unlikely in the near term. A slower, more chronic release of CH_4 from melting permafrost, thermokarst lakes or marine hydrates is expected as a result of global warming[57].

N_2O also varies on glacial–interglacial timescales, with typical glacial and interglacial values of around 210 p.p.b. and 270 p.p.b., respectively[39]. Although the associated radiative forcing is small, the history of this gas is of interest as an integrative tracer of changes in the global nitrogen cycle and because of likely positive feedbacks of global warming on the modern N_2O budget. The primary natural sources are microbial nitrification and denitrification in both marine and terrestrial ecosystems. Changes in the ocean source are linked to circulation-driven changes in upper ocean oxygen that affect denitrification[58]. Changes in nitrification on land are linked to temperature and rainfall, primarily in the tropics. Stable isotope data indicate that both the marine and terrestrial sources increased during the last deglaciation, with fast changes[59] in the terrestrial source that paralleled those in CH_4.

Apart from the changes in the main greenhouse gases, other aspects of the atmospheric evolution revealed by Antarctic ice cores add to our understanding of Earth system change on long timescales; the list below is by no means exhaustive. First, the isotopic composition of atmospheric O_2 shows a strong orbital precession signal, with superimposed millennial-scale variations linked to abrupt climate change. On both timescales this reflects the strength of the global monsoon, which is modulated by both insolation and the position of the ITCZ[13,60]. Second, atmospheric O_2 concentrations have been falling at a rate[61] of 8.4‰ per million years, suggesting that at present O_2 sinks (the oxidation of sedimentary organic carbon and pyrite) exceed O_2 sources (the burial of the same) by about 2%. Third, ice core measurements demonstrate the steady accumulation of radiogenic ^{40}Ar (from ^{40}K decay in the crust) in the atmosphere, allowing an estimate of the contemporary crustal degassing rate[62] and providing a dating technique for old ice. Last, the atmospheric Kr/N_2 ratio is a proxy for global mean ocean temperature[63], owing to the temperature dependence of solubility. The reduced atmospheric krypton inventory during the Last Glacial Maximum suggests that the mean oceanic temperature was 2.57 ± 0.24 °C lower than at present[64]. The trend of mean ocean warming during the last deglaciation follows Antarctic temperature and atmospheric CO_2 closely, further demonstrating the close link between the high-latitude Southern Hemisphere and the global climate system.

The close view: millennial-scale variability

Ice cores also preserve the impact of millennial-scale and shorter climate variability, increasingly recognized as important for understanding Earth system feedbacks and the potential for abrupt future change. The now well known, abrupt Dansgaard–Oeschger events

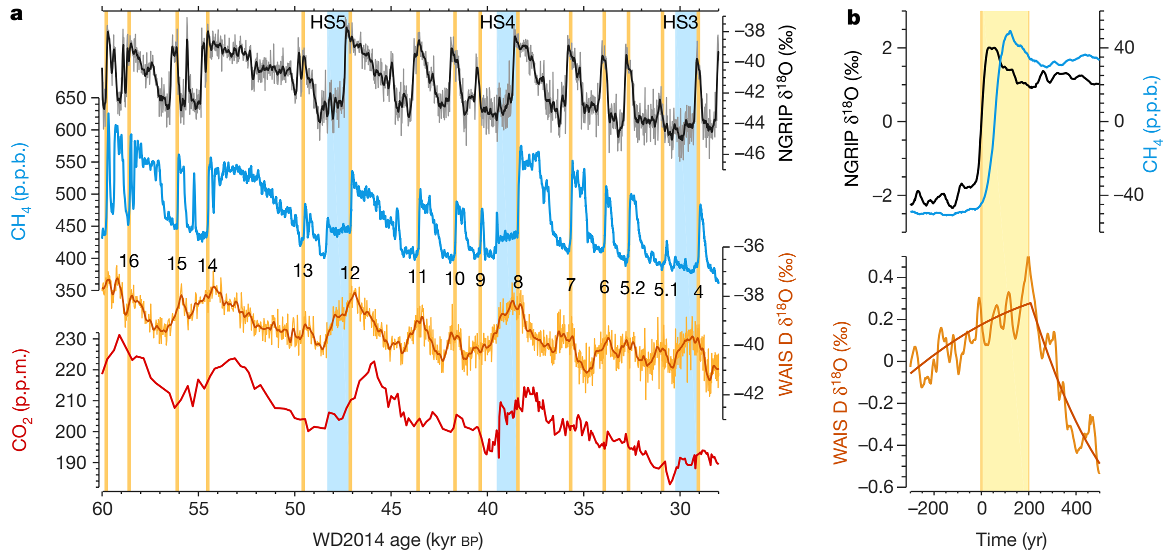


Fig. 2 | Abrupt climate variability of the last Ice Age. a, Records of abrupt climate variability. From top to bottom, the traces show Greenland water isotope ratios from the NGRIP core[67], atmospheric CH_4 from the Antarctic WAIS Divide ice core[85], Antarctic ice core water isotope ratios from the WAIS Divide core[73] and atmospheric CO_2 from a multi-core compilation[110]. Water isotope ratios are measured relative to VSMOW. Blue bars show the approximate timing of Heinrich Stadials 5 to 3. Numbers indicate AIM events. **b**, Interpolar phasing of abrupt climate change. The records from **a** are aligned at the abrupt Northern Hemisphere transitions (yellow vertical lines), and averaged to obtain the shared climatic signal. The Antarctic cooling response of the bipolar seesaw is delayed by around two centuries behind the abrupt Northern Hemisphere events[73].

observed in Greenlandic ice cores (Fig. 2) and other Northern Hemisphere climate records covering the last ice age[65–67] have well documented counterparts in Antarctica[68–73], termed Antarctic Isotope Maxima (AIM) events. The precise relative timing of events in Greenland and Antarctica has been established by using well mixed atmospheric gases as stratigraphic markers[68–70,73,74]. Antarctica warmed during Northern Hemisphere cold periods and cooled when Greenland was warm (Fig. 2). This bi-polar linkage is well documented for the last ice age and deglaciation. It very probably operated in prior ice ages but does not have obvious counterparts during interglacials such as the present Holocene. There are regional differences in the expression of the AIM events, which may ultimately provide more information about mechanisms[72].

The concept of the ‘bi-polar seesaw’ emerged from these observations of asynchronous temperature variations between the hemispheres[27,75]. The basic theory is that perturbations to the northward, cross-equatorial heat transport of the Atlantic Ocean exert opposite temperature effects on both hemispheres, with the Antarctic counterpart damped by a large heat reservoir, commonly assumed to be the Southern Ocean. Climate models in which the strength of the Atlantic Meridional Overturning Circulation (AMOC) is perturbed, usually through freshwater input into the North Atlantic, can reproduce the seesaw pattern[76–78].

Recently, the WAIS Divide site in West Antarctica provided a climate record with temporal resolution similar to that of the central Greenland cores, although extending only to 68 kyr (the longest stratigraphically ordered Greenland record is NGRIP, at about 123 kyr long). This new ice core combines high accumulation rate, very low gas/ice age difference, and high-precision atmospheric records. These attributes allow for the most precise investigation of the interpolar phasing of the bipolar seesaw yet[73]. Comparison of Antarctic and Greenlandic events reveals on average an approximately 200-year-long lag of the Antarctic response behind both Greenland abrupt warming and cooling events (Fig. 2). The origin of this delay must lie in the climate coupling between the hemispheres, and indicates a north-to-south propagation of the climate signal dominated by oceanic processes (given that atmospheric propagation would be much faster). Further work on the WAIS Divide ice core[79] explored interhemispheric atmospheric teleconnections by analysing the deuterium excess record, a parameter believed to reflect source moisture conditions and transport pathways to the site. This study showed that a component of the WAIS Divide deuterium excess variations is closely correlated with Greenland climate at zero time lag, implying a fast (atmospheric) link between abrupt warming in the north and shifts in Southern Hemisphere moisture pathways to Antarctica. A similar response had been observed in the Dome C core for termination II[80]. These studies indicate that the two polar regions are coupled via both oceanic and atmospheric teleconnections, each operating on their own timescale.

Tracers of mineral dust deposition in Antarctic ice cores also vary on millennial timescales[6,72,81] with lower dust deposition during warmer periods. Changes in dust transport and generation of dust in source areas are both likely to be involved. Sea salt sodium does not display as clear a link to the millennial-scale climate events, but as discussed above, interpretation of this proxy is complicated. Deposition of both dust and sea salt aerosol appear to be regionally variable within Antarctica.

Atmospheric gas records from Antarctic ice cores are providing more detail about millennial-scale variability in global biogeochemical cycles. Changes in CO_2 tend to follow Antarctic AIM events closely (Fig. 2), particularly for the larger events, as well as for the deglacial warming where the two are essentially synchronous (see Box 2). This close association of Antarctic temperature and CO_2 suggests that Southern Ocean processes are also critical to atmospheric CO_2 variability on these timescales. Changes in export production or ventilation in the Southern Ocean are candidate mechanisms. Both can be linked to AMOC changes, for example, via increased upwelling during Heinrich stadials due to southward migration of the Southern Hemisphere westerly winds[47], or via more direct impacts of AMOC changes on ocean circulation[82] or the efficiency of the biological pump[58].

High-resolution CO_2 data reveal that very rapid carbon cycle variations are superimposed on the slower millennial changes during the last ice age and deglaciation[83,84]. There appear to be two types of rapid CO_2 increase. The first is exemplified by increases during two major warming events in the Northern Hemisphere during the deglaciation: the Younger Dryas termination and Bølling–Allerod warming (Box 2). In each case, CO_2 increases of about 10 p.p.m. took place over periods of 100–200 years[84]. Their coincidence with northern warming is unambiguous, given coincident increases in atmospheric CH_4 that mark the two abrupt northern events. The second type is associated with Heinrich stadials, periods of ice-rafted debris discharge into the

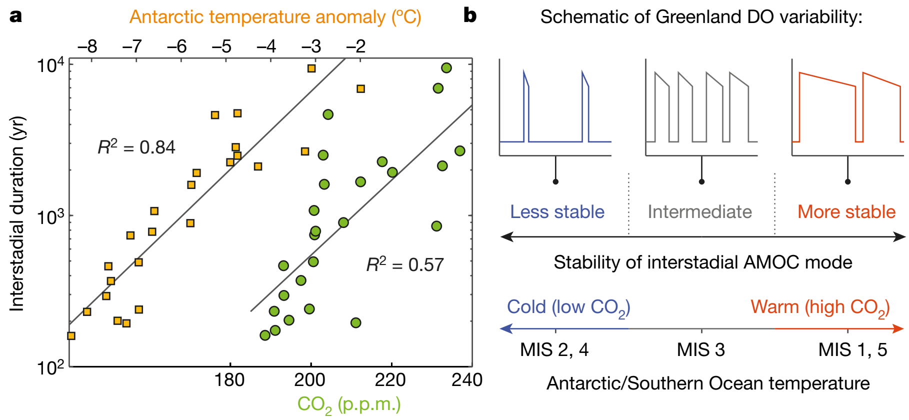


Fig. 3 | Dependence of millennial-scale variability on the background climate state. a, Dansgaard–Oeschger interstadial duration (logarithmic scale) from Greenland ice core records plotted against Antarctic temperature[73] and atmospheric CO_2, with the coefficient of determination R^2 listed for each case. **b**, Schematic of Greenland Dansgaard–Oeschger variability during various states of the background climate. On the left, during cold (low-CO_2) periods such as Marine Isotope Stages (MIS) 2 and 4, Dansgaard–Oeschger interstadials are infrequent and of short duration, suggesting that the interstadial AMOC mode is unstable. In the middle, during intermediate climates such as MIS3, Dansgaard–Oeschger interstadials are frequent and of medium duration, resulting in high event frequency. On the right, during warm (high CO_2) periods such as MIS 5, Dansgaard–Oeschger interstadials are of long duration, resulting in a lower event frequency, suggesting that the interstadial AMOC mode is very stable. Figure modified from ref. [89] (Wiley).

North Atlantic that occurred during several of the Greenland stadial periods. Clear centennial-scale CO_2 rises are seen during Heinrich Stadial 1 (16.3 kyr BP)[84] (Box 2) and Heinrich Stadial 4 (39.5 kyr BP)[83]. These events are synchronous with short-duration oscillations[85] in atmospheric CH_4, suggesting that the abrupt shifts in both gases are synchronous with changes in tropical hydrology, for example, due to southward shifts in the position of the ITCZ.

Rapid changes in terrestrial carbon storage are a potential explanation for these abrupt increases in CO_2. Warming in the Northern Hemisphere at the onset of the Bølling period and end of the Younger Dryas (Box 2) could conceivably alter the carbon balance in terrestrial ecosystems. However, stable isotope data[48] do not support this as a primary mechanism for the CO_2 increases associated with northern warming, and instead suggest that sea surface temperature change is probably at least part of the explanation. The 16.3-kyr CO_2 rise during HS1 is associated with a negative carbon isotopic excursion, implying the release of respired carbon[48]. Drying in the tropics and release of respired carbon associated with the southward migration of the ITCZ could explain this rapid CO_2 change and an analogous CO_2 change during HS4 and possibly HS5 (Fig. 2). This hypothesis is consistent with the suggestion that the small increases in atmospheric CH_4 registered at precisely the same time as the HS1 and HS4 CO_2 increases were caused by a southward shift of ITCZ rainfall[85].

At least one further aspect of Antarctic ice core data on short timescales deserves mention—the record of volcanism provided by anomalies in non-marine salt sulphate and the presence of volcanic debris (tephra). Ice core records provide highly detailed information about the eruption frequency needed to understand the role of volcanic forcing on climate. Much progress has been made recently in refining these records. For example, identification of synchronous volcanic sulphate deposition in the Antarctic and Greenland indicates tropical eruptions with the potential to affect climate globally[86]. Using the new WAIS Divide record and new data from Greenland, ref. [86] identified five eruptions in the last 2,500 years that were larger than the 1815 Tambora event. Volcanic events also provide critical stratigraphic links that are improving ice core timescales[87,88], a critical factor for making a unified Antarctic ice core chronology that will allow examination of regional differences in climate.

The middle ground: orbital–millennial interaction

Commonly, orbital-scale and millennial-scale climate change are treated and discussed in the literature as separate topics. For example, in their seminal paper on the bipolar seesaw, Stocker and Johnson[27] start by filtering out the orbital signal in order to investigate the abrupt Dansgaard–Oeschger events. Likewise, studies that seek to understand the link between ice volume and insolation (see ref. [19] for example) do not always consider the influence of abrupt events. This approach is of course valid, and much has been learned by studying these timescales in isolation. However, capturing the full dynamics of the climate system requires consideration of the myriad ways in which millennial- and orbital-scale climate change interact, and influence each other.

For example, the duration of Dansgaard–Oeschger interstadial phases scales strongly with the mean climate state[89]. In Fig. 3a, Dansgaard–Oeschger duration is plotted against Antarctic temperature and CO_2. During relatively warm, high-CO_2 periods such as MIS 5, Dansgaard–Oeschger interstadials tend to be long, suggesting a greater stability of the interstadial (or strong) AMOC mode; conversely, during cold, low-CO_2 periods such as MIS 2 and 4, Dansgaard–Oeschger interstadials are infrequent and of short duration, suggesting a weak interstadial mode that readily collapses (Fig. 3b). Early studies hypothesized a controlling influence of continental ice mass[90,91], but this is contradicted by global climate model simulations showing a stronger AMOC overturning upon increasing Laurentide ice volume[92]. Viable explanations for this state dependence (Fig. 3) include sea ice dynamics of the North Atlantic[93,94], the state of the Southern Ocean[89] and CO_2 levels (Fig. 3a)[12], none of which are mutually exclusive.

Conversely, the bipolar seesaw appears to play an important part in the orbitally paced glacial cycle. As discussed above, data[95] and models[58] suggest that atmospheric CO_2 levels, thought to be the global amplifier of the glacial cycles, are closely linked to (millennial-scale) changes in ocean circulation. Moreover, speleothem records indicate a strongly weakened East Asian monsoon during all of the last seven glacial terminations, interpreted as a southward shift in the ITCZ driven by North Atlantic cooling and AMOC cessation[20]. In this view, glacial terminations are simply the most powerful realizations of the (millennial-scale) Antarctic AIM events[5]. It thus appears that the bipolar seesaw has an important role in the machinery of glacial cycles, and should be considered in answering questions that are commonly placed in the 'orbital' realm, such as the aforementioned problems of the 100-kyr cycle and the interhemispheric climate symmetry at the obliquity and precession timescales.

The interdependence of orbital and millennial-scale climate change highlights the need to consider and develop theories that are applicable to both timescales. Transient climate model simulations of the last deglaciation that incorporate both orbital and millennial-scale AMOC forcings have been highly successful in fitting observations[96];

however, the freshwater fluxes were still prescribed rather than simulated. Similar coupled ocean–atmosphere climate model experiments are needed on all timescales, although the computational cost of such an endeavour is a challenge.

Lessons for a warming world from Antarctic ice cores

Perhaps the most fundamental message from ice cores is just how profound the anthropogenic impact on our atmospheric composition has been in the context of long-term natural variability. Levels of the primary greenhouse gases CO_2, CH_4 and N_2O, all of which are directly affected by human emissions, are at higher levels than at any time[30,38,39] in the last 800 kyr. At the time of writing, concentrations of these three gases are elevated by about 45%, 155% and 22% over pre-industrial values, respectively. Modern changes in CO_2 are also much more rapid than in the ice core record. The fastest pre-industrial increases in CO_2 were about 0.1 p.p.m. per year[84], compared to consistent recent growth rates close to 2 p.p.m. per year since the early 1990s[97]. For CH_4, the modern growth rate is also faster than observed through most of the ice core record, although certain natural abrupt transitions match the speed of the current increase[85,97]. Nitrous oxide is not generally sampled closely enough to make such comparisons, but its past growth rate is unlikely to be as rapid as the current rise. The strong correlation of variations in these gases with climate proxies over the last 800 kyr verifies the importance of the greenhouse effect in global climate. As discussed above, the temperature records themselves confirm the theory of polar amplification[10], an important feature in future warming with implications for Arctic societies and wildlife.

Recent work shows that rapid changes in climate and global biogeochemical cycles during the last ice age and deglaciation affected both the Southern Hemisphere and the tropics[60,73,84,85], with a strong imprint on tropical systems during Heinrich stadials[60,85]. However, it is not clear what this tells us about what to expect in the future. On the one hand, much of the major millennial variability in the ice core record occurred under different (that is, glacial) boundary conditions, including lower sea level, larger ice sheets and considerably lower CO_2 levels. The lack of those conditions now suggests that in the near future completely analogous events are unlikely. On the other hand, there is a considerable need to understand the strengths of fast feedbacks in the climate system, including the possibility of rapid mass loss at ice sheet margins, the possibility that Greenland meltwater runoff and surface warming could affect the Atlantic overturning circulation, and the potential for fast changes in the global carbon cycle.

The ice core record suggests a more vigorous and stable AMOC during warmer background climate states (Fig. 3); if this palaeo-observation from the last ice age were to apply to future climates also (which has not been rigorously tested), it would imply that the short-term, transient AMOC weakening driven by freshening of the surface North Atlantic may in the long term be offset by an increase in equilibrium AMOC strength in a warmer world[89,98]. Furthermore, changes in the ITCZ and westerly winds appear to be part of the process that transmits millennial signals to Antarctica. It is therefore important to understand their dynamics and impacts better, considering the potential for future changes in regions currently supporting a major fraction of the global human population.

The future of Antarctic ice core science

The continuous ice core record to 800 kyr is a remarkable achievement. An extension of this record to earlier times, with a goal of reaching back to 1.5 million years BP, is a major new international priority, with ongoing searches for appropriate sites[99]. Several fundamental questions drive this quest. As discussed, ocean sediment records show that before the so-called Mid-Pleistocene Transition (1,200–800 kyr ago), global climate was dominated by a strong 41-kyr period, the cycle associated with the variations in Earth's tilt. After the Mid-Pleistocene Transition the quasi-100-kyr variability dominated. One fundamental question is whether global temperature was warmer at this time, perhaps resulting in smaller and more mobile ice sheets, and shorter ice age cycles[21]. A second question is whether Antarctic temperature before the Mid-Pleistocene Transition tracked the benthic isotope record, as it does in later times, or varied on some other timescale[100]. A third question concerns whether changes in the long-term mean atmospheric concentrations of greenhouse gases play a part in changing the frequency of glaciation.

The search for a suitable location at which to drill for very old ice involves major investments in radar remote sensing and rapid-access drills[101–103] with which to test probable sites, and the development of better measurement methods. Several international groups are setting their sights on this goal, and it is likely that more than one record will be needed to confirm results. 'Snapshots' of older time periods can also be achieved by shallow drilling in ice margin regions[3], with the potential for finding ice older than 2 million years[104].

There is also much more to learn about long-term climate and biogeochemical cycles in the existing 800,000-year record through more detailed measurements, including completion of long isotopic records for greenhouse gases and improved (volcanic) synchronization of ice cores both within Antarctica and from Greenland. Understanding the processes that lead to warm interglacials, the nature of abrupt change throughout the record, and the controls on greenhouse gas variability and its links to climate change are some obvious goals. The Holocene history of Antarctica is also critical. Although we have a fairly comprehensive ice core view of the entire Holocene[29,105], the data needed to put current global warming in the context of Antarctic changes could be improved[106]. Increasing the spatial coverage of high-quality, well dated records would add a great deal to our understanding of this issue.

One of the largest questions about Antarctica for the near future concerns the stability of the West Antarctic Ice Sheet (WAIS). This ice sheet is believed to be vulnerable to collapse as it is grounded below sea level with a retrograde bedrock slope. A critical question that ice core science might answer is whether the ice sheet collapsed during the last interglacial, when temperatures and sea level were higher than today. No cores in the present WAIS penetrate this period, but given the limited number of drilling projects and the high basal melt rate in the area, this cannot be taken as evidence that the WAIS was not there. Modelling suggests that WAIS collapse could be recorded by distinctive patterns in ice core temperature proxies adjacent to the WAIS, related to altered atmospheric circulation[107]. The Mount Moulton blue ice site, at 2,820 m in Marie Byrd Land[108], provides the only interglacial stable isotope record from West Antarctica; the data are consistent with collapse scenarios[107]. Given uncertainties in models and interpretations of blue ice sites, and the appropriate boundary conditions for collapse scenarios, this result is by no means definitive. Additional coring in and next to the WAIS will be needed to provide better constraints—sites identified for this purpose include Hercules Dome and Skytrain Ice Rise.

Advancing understanding and measurement of ice core proxies will continue to be important. New techniques, for example, deep-ocean temperature proxies from noble gases, 'clumped isotopes' in atmospheric gases, mass-independent isotope fractionation in water, sulphate, oxygen and other systems, water isotope diffusion, palaeo-data assimilation techniques, better measurements and more detail in heavy isotope proxies of dust sources, continuous, centimetre-scale water isotope and gas analysis, more detailed isotopic measurements of greenhouse gases, and a host of other methods hold promise for revealing much more about Antarctic and global environmental change.

Finally, although there are ice core records throughout Antarctica (Box 1), the overall spatial coverage is in fact quite thin considering the size and complexity of the continent, and new drilling will continue to be needed to advance scientific goals. International cooperation in ice core drilling and prioritizing science is strong. The technical expertise required to accomplish challenging field programmes is available in many nations, placing ice core science in an excellent position to improve our understanding of the history of Antarctica and its links to the larger Earth system.

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1. Galeotti, S. et al. Antarctic Ice Sheet variability across the Eocene-Oligocene boundary climate transition. *Science* **352**, 76–80 (2016).
2. Flower, B. P. & Kennett, J. P. Relations between Monterey Formation deposition and middle Miocene global cooling: Naples Beach section. *Calif. Geol.* **21**, 877–880 (1993).
3. Higgins, J. A. et al. Atmospheric composition 1 million years ago from blue ice in the Allan Hills, Antarctica. *Proc. Natl Acad. Sci. USA* **112**, 6887–6891 (2015). **This study reports the first greenhouse gas data from ice older than 800,000 years.**
4. Jouzel, J. et al. Validity of the temperature reconstruction from water isotopes in ice cores. *J. Geophys. Res. Oceans* **102**, 26471–26487 (1997).
5. Wolff, E., Fischer, H. & Röthlisberger, R. Glacial terminations as southern warmings without northern control. *Nat. Geosci.* **2**, 206–209 (2009).
6. Schüpbach, S. et al. High-resolution mineral dust and sea ice proxy records from the Talos Dome ice core. *Clim. Past* **9**, 2789–2807 (2013).
7. McConnell, J. R. et al. Antarctic-wide array of high-resolution ice core records reveals pervasive lead pollution began in 1889 and persists today. *Sci. Rep.* **4**, 5848 (2014).
8. Brook, E. J., Kurz, M. D. & Curtice, J. Flux and size fractionation of ^{3}He in interplanetary dust from Antarctic ice core samples. *Earth Planet. Sci. Lett.* **286**, 565–569 (2009).
9. Van Ommen, T. D., Morgan, V. & Curran, M. A. Deglacial and Holocene changes in accumulation at Law Dome, East Antarctica. *Ann. Glaciol.* **39**, 359–365 (2004).
10. Cuffey, K. M. et al. Deglacial temperature history of West Antarctica. *Proc. Natl Acad. Sci. USA* **113**, 14249–14254 (2016). **This work provides the first accurate borehole-based temperature reconstruction from Antarctica, indicating a glacial–interglacial temperature change of 11.3 $\pm$ 1.8 °C.**
11. Jouzel, J. et al. Orbital and millennial Antarctic climate variability over the past 800,000 years. *Science* **317**, 793–796 (2007). **This paper reports the full 800,000-year temperature reconstruction from the EPICA Dome C ice core, the longest such record.**
12. Dome Fuji Project Members. State dependence of climatic instability over the past 720,000 years from Antarctic ice cores and climate modeling. *Sci. Adv.* **3**, e1600446 (2017).
13. Petit, J. R. et al. Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* **399**, 429–436 (1999).
14. Lisiecki, L. E. & Raymo, M. E. A Pliocene-Pleistocene stack of 57 globally distributed benthic δ^{18}O records. *Paleoceanography* **20**, 1–17 (2005).
15. Shakun, J. D. et al. Global warming preceded by increasing carbon dioxide concentrations during the last deglaciation. *Nat. Geosci.* **484**, 49–55 (2012).
16. Hays, J. D., Imbrie, J. & Shackleton, N. J. Variations in the Earth's orbit: pacemaker of the ice ages. *Science* **194**, 1121–1132 (1976).
17. Imbrie, J. et al. On the structure and origin of major glaciation cycles. Paleoceanogr. *Paleoclimatol.* **8**, 699–735 (1993).
18. Raymo, M. The timing of major climate terminations. *Paleoceanogr. Paleoclimatol.* **12**, 577–585 (1997).
19. Paillard, D. The timing of Pleistocene glaciations from a simple multiple-state climate model. *Nature* **391**, 378–381 (1998).
20. Cheng, H. et al. The Asian monsoon over the past 640,000 years and ice age terminations. *Nature* **534**, 640–646 (2016).
21. Tzedakis, P., Crucifix, M., Mitsui, T. & Wolff, E. W. A simple rule to determine which insolation cycles lead to interglacials. *Nature* **542**, 427–432 (2017).
22. Bintanja, R. & Van de Wal, R. North American ice-sheet dynamics and the onset of 100,000-year glacial cycles. *Nature* **454**, 869–872 (2008).
23. Abe-Ouchi, A. et al. Insolation-driven 100,000-year glacial cycles and hysteresis of ice-sheet volume. *Nature* **500**, 190–193 (2013).
24. Ganopolski, A. & Brovkin, V. Simulation of climate, ice sheets and CO_2 evolution during the last four glacial cycles with an Earth system model of intermediate complexity. *Clim. Past* **13**, 1695–1716 (2017).
25. Broecker, W. S. & Denton, G. H. The role of ocean-atmosphere reorganizations in glacial cycles. *Quat. Sci. Rev.* **9**, 305–341 (1990).
26. Kawamura, K. et al. Northern Hemisphere forcing of climatic cycles in Antarctica over the past 360,000 years. *Nature* **448**, 912–916 (2007). **This study links variations in the O_2/N_2 ratio of trapped air in the Dome Fuji ice core to local summer insolation, thereby dating the core and showing that glacial terminations in Antarctica closely followed Northern Hemisphere summer insolation.**
27. Stocker, T. F. & Johnsen, S. J. A minimum thermodynamic model for the bipolar seesaw. *Paleoceanogr. Paleoclimatol.* **18**, https://doi.org/10.1029/2003PA000920 (2003).
28. Huybers, P. & Denton, G. Antarctic temperature at orbital timescales controlled by local summer duration. *Nat. Geosci.* **1**, 787–792 (2008).
29. WAIS Divide Project Members. Onset of deglacial warming in West Antarctica driven by local orbital forcing. *Nature* **500**, 440–444 (2013).
30. Loulergue, L. et al. Orbital and millennial-scale features of atmospheric CH_4 over the past 800,000 years. *Nature* **453**, 383–386 (2008). **This paper reports the full 800,000-year atmospheric methane record from the EPICA Dome C ice core, showing variations on orbital and millennial timescales.**
31. Yin, Q. Insolation-induced mid-Brunhes transition in Southern Ocean ventilation and deep-ocean temperature. *Nature* **494**, 222–225 (2013).
32. Wolff, E. W. et al. Southern Ocean sea-ice extent, productivity and iron flux over the past eight glacial cycles. *Nature* **440**, 491–496 (2006). **Chemical measurements from the EPICA Dome C ice core indicate that both the flux of iron (from wind-blown dust) and sea ice extent increased during glacial periods over the past 740,000 years.**
33. Lambert, F. et al. Dust–climate couplings over the past 800,000 years from the EPICA Dome C ice core. *Nature* **452**, 616–619 (2008).
34. Fischer, H., Siggaard-Andersen, M. L., Ruth, U., Röthlisberger, R. & Wolff, E. Glacial/interglacial changes in mineral dust and sea-salt records in polar ice cores: sources, transport, and deposition. *Rev. Geophys.* **45**, 1–26 (2007).
35. Martínez-Garcia, A. et al. Links between iron supply, marine productivity, sea surface temperature, and CO_2 over the last 1.1 Ma. *Paleoceanogr. Paleoclimatol.* **24**, PA1207 (2009).
36. Jaccard, S. et al. Two modes of change in Southern Ocean productivity over the past million years. *Science* **339**, 1419–1423 (2013).
37. Abram, N. J., Wolff, E. W. & Curran, M. A. A review of sea ice proxy information from polar ice cores. *Quat. Sci. Rev.* **79**, 168–183 (2013).
38. Lüthi, D. et al. High-resolution carbon dioxide concentration record 650,000–800,000 years before present. *Nature* **453**, 379–382 (2008). **This paper completed the 800,000-year-long EPICA Dome C CO_2 record shown in Fig. 1, demonstrating the variation of CO_2 maxima during interglacial times.**
39. Schilt, A. et al. Glacial-interglacial and millennial-scale variations in the atmospheric nitrous oxide concentration during the last 800,000 years. *Quat. Sci. Rev.* **29**, 182–192 (2010).
40. Sigman, D. M. & Boyle, E. A. Glacial/interglacial variations in atmospheric carbon dioxide. *Nature* **407**, 859–869 (2000).
41. Sigman, D. M., Hain, M. P. & Haug, G. H. The polar ocean and glacial cycles in atmospheric CO_2 concentration. *Nature* **466**, 47–55 (2010).
42. Stephens, B. B. & Keeling, R. F. The influence of Antarctic sea ice on glacial–interglacial CO_2 variations. *Nature* **404**, 171–174 (2000).
43. Skinner, L., Fallon, S., Waelbroeck, C., Michel, E. & Barker, S. Ventilation of the deep Southern Ocean and deglacial CO_2 rise. *Science* **328**, 1147–1151 (2010).
44. Ferrari, R. et al. Antarctic sea ice control on ocean circulation in present and glacial climates. *Proc. Natl Acad. Sci. USA* **111**, 8753–8758 (2014).
45. Martin, J. H. Glacial-interglacial CO_2 change: the iron hypothesis. *Paleoceanogr. Paleoclimatol.* **5**, 1–13 (1990).
46. Franois, R. et al. Contribution of Southern Ocean surface-water stratification to low atmospheric CO_2 concentrations during the last glacial period. *Nature* **389**, 929–935 (1997).
47. Anderson, R. et al. Wind-driven upwelling in the Southern Ocean and the deglacial rise in atmospheric CO_2. *Science* **323**, 1443–1448 (2009).
48. Bauska, T. K. et al. Carbon isotopes characterize rapid changes in atmospheric carbon dioxide during the last deglaciation. *Proc. Natl Acad. Sci. USA* **113**, 3465–3470 (2016).
49. Galbraith, E. & Eggleston, S. A lower limit to atmospheric CO_2 concentrations over the past 800,000 years. *Nat. Geosci.* **10**, 295–298 (2017).
50. Brook, E. J., Sowers, T. & Orchardo, J. Rapid variations in atmospheric methane concentration during the past 110,000 years. *Science* **273**, 1087–1091 (1996).
51. Petrenko, V. V. et al. Minimal geological methane emissions during the Younger Dryas–Preboreal abrupt warming event. *Nature* **548**, 443–446 (2017).
52. Sowers, T. Late quaternary atmospheric CH_4 isotope record suggests marine clathrates are stable. *Science* **311**, 838–840 (2006).
53. Bock, M. et al. Hydrogen isotopes preclude marine hydrate CH_4 emissions at the onset of Dansgaard-Oeschger events. *Science* **328**, 1686–1689 (2010).
54. Levine, J. et al. Reconciling the changes in atmospheric methane sources and sinks between the Last Glacial Maximum and the pre-industrial era. *Geophys. Res. Lett.* **38**, L23804 (2011).
55. Murray, L. T. et al. Factors controlling variability in the oxidative capacity of the troposphere since the Last Glacial Maximum. *Atmos. Chem. Phys.* **14**, 3589–3622 (2014).
56. Fischer, H. et al. Changing boreal methane sources and constant biomass burning during the last termination. *Nature* **452**, 864–867 (2008).
57. Brook, E., Archer, D., Dlugokencky, E., Frolking, S. & Lawrence, D. in *Abrupt Climate Change. A report by the U.S. Climate Change Science Program and the Subcommittee on Global Change Research* (ed. McGeehin, J. P.) Ch. 5, 163–201 (US Geological Survey, Reston, 2008).
58. Schmittner, A. & Galbraith, E. D. Glacial greenhouse-gas fluctuations controlled by ocean circulation changes. *Nature* **456**, 373–376 (2008).
59. Schilt, A. et al. Isotopic constraints on marine and terrestrial N_2O emissions during the last deglaciation. *Nature* **516**, 234–237 (2014).
60. Severinghaus, J. P., Beaudette, R., Headly, M. A., Taylor, K. & Brook, E. J. Oxygen-18 of O_2 records the impact of abrupt climate change on the terrestrial biosphere. *Science* **324**, 1431–1434 (2009).
61. Stolper, D., Bender, M., Dreyfus, G., Yan, Y. & Higgins, J. A Pleistocene ice core record of atmospheric O_2 concentrations. *Science* **353**, 1427–1430 (2016).
62. Bender, M. L., Barnett, B., Dreyfus, G., Jouzel, J. & Porcelli, D. The contemporary degassing rate of ^{40}Ar from the solid Earth. *Proc. Natl Acad. Sci. USA* **105**, 8232–8237 (2008). **Precise measurements of argon isotope ratios in trapped air are used in this study to develop a chronometer for old ice based on the accumulation of ^{40}Ar in the atmosphere from ^{40}K decay in the crust.**
63. Headly, M. A. & Severinghaus, J. P. A method to measure Kr/N_2 ratios in air bubbles trapped in ice cores and its application in reconstructing past mean ocean temperature. *J. Geophys. Res.* **112**, D19105 (2007).
64. Bereiter, B., Shackleton, S., Baggenstos, D., Kawamura, K. & Severinghaus, J. Mean global ocean temperatures during the last glacial transition. *Nature* **553**, 39–44 (2018). **Measurements of Kr, Xe, Ar, and N are used in this study to make the first precise estimates of changes in global deep-ocean temperature across the last glacial–interglacial transition, showing that these are mostly synchronous with changes in Antarctic air temperature and atmospheric CO_2.**

65. Grootes, P., Stuiver, M., White, J., Johnsen, S. & Jouzel, J. Comparison of oxygen isotope records from the GISP2 and GRIP Greenlandice cores. *Nature* **366**, 552–554 (1993).
66. Dansgaard, W. et al. Evidence for general instability of past climate from a 250-kyr ice-core record. *Nature* **364**, 218–220 (1993).
This paper reports the first detailed record of abrupt changes in temperature in Greenland.
67. Andersen, K. K. et al. High-resolution record of Northern Hemisphere climate extending into the last interglacial period. *Nature* **431**, 147–151 (2004).
68. Blunier, T. & Brook, E. J. Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science* **291**, 109–112 (2001).
This paper used methane variations to make a common timescale for ice cores in Greenland and Antarctica, and showed the bi-polar seesaw pattern for the larger climate variations during the last ice age.
69. Brook, E. J. et al. Timing of millennial-scale climate change at Siple Dome, West Antarctica, during the last glacial period. *Quat. Sci. Rev*. **24**, 1333–1343 (2005).
70. EPICA Community Members. One-to-one coupling of glacial climate variability in Greenland and Antarctica. *Nature* **444**, 195–198 (2006).
This paper demonstrated that all abrupt climate events in the Greenland record have counterparts in Antarctica.
71. Stenni, B. et al. Expression of the bipolar see-saw in Antarctic climate records during the last deglaciation. *Nat. Geosci*. **4**, 46–49 (2011).
72. Landais, A. et al. A review of the bipolar see–saw from synchronized and high resolution ice core water stable isotope records from Greenland and East Antarctica. *Quat. Sci. Rev*. **114**, 18–32 (2015).
73. WAIS Divide Project Members. Precise interpolar phasing of abrupt climate change during the last ice age. *Nature* **520**, 661–665 (2015).
Using very precise chronological constraints, this paper demonstrated that millennial-scale warming and cooling in Antarctica lagged counterpart events in Greenland by about 200 years on average.
74. Bender, M. et al. Climate correlations between Greenland and Antarctica during the past 100,000 years. *Nature* **372**, 663–666 (1994).
75. Broecker, W. S. Paleocean circulation during the last deglaciation: a bipolar seesaw? Paleoceanogr. *Paleoclimatol*. **13**, 119–121 (1998).
76. Vellinga, M. & Wood, R. A. Global climatic impacts of a collapse of the Atlantic thermohaline circulation. *Clim. Change* **54**, 251–267 (2002).
77. Schmittner, A., Saenko, O. & Weaver, A. Coupling of the hemispheres in observations and simulations of glacial climate change. *Quat. Sci. Rev*. **22**, 659–671 (2003).
78. Stouffer, R. J. et al. Investigating the causes of the response of the thermohaline circulation to past and future climate changes. *J. Clim*. **19**, 1365–1387 (2006).
79. Markle, B. R. et al. Global atmospheric teleconnections during Dansgaard-Oeschger events. *Nat. Geosci*. **10**, 36–40 (2017).
80. Masson-Delmotte, V. et al. Abrupt change of Antarctic moisture origin at the end of Termination II. *Proc. Natl Acad. Sci. USA* **107**, 12091–12094 (2010).
81. Lambert, F., Bigler, M., Steffensen, J. P., Hutterli, M. & Fischer, H. Centennial mineral dust variability in high-resolution ice core data from Dome C, Antarctica. *Clim. Past* **8**, 609–623 (2012).
82. Menviel, L., England, M. H., Meissner, K., Mouchet, A. & Yu, J. Atlantic-Pacific seesaw and its role in outgassing CO_2 during Heinrich events. *Paleoceanogr. Paleoclimatol*. **29**, 58–70 (2014).
83. Ahn, J., Brook, E. J., Schmittner, A. & Kreutz, K. Abrupt change in atmospheric CO_2 during the last ice age. *Geophys. Res. Lett*. **39**, GL053018 (2012).
84. Marcott, S. A. et al. Centennial-scale changes in the global carbon cycle during the last deglaciation. *Nature* **514**, 616–619 (2014).
The most detailed CO_2 record for the deglaciation to date is reported from the WAIS Divide ice core in this paper, showing the tight coupling of CO_2 and Antarctic climate (Box 2).
85. Rhodes, R. H. et al. Enhanced tropical methane production in response to iceberg discharge in the North Atlantic. *Science* **348**, 1016–1019 (2015).
86. Sigl, M. et al. Timing and climate forcing of volcanic eruptions for the past 2,500 years. *Nature* **523**, 543–549 (2015).
87. Veres, D. et al. The Antarctic ice core chronology (AICC2012): an optimized multi-parameter and multi-site dating approach for the last 120 thousand years. *Clim. Past* **9**, 1733–1748 (2013).
88. Bazin, L. et al. An optimized multi-proxy, multi-site Antarctic ice and gas orbital chronology (AICC2012): 120–800 ka. *Clim. Past Discuss*. **8**, 5963–6009 (2012).
89. Buizert, C. & Schmittner, A. Southern Ocean control of glacial AMOC stability and Dansgaard-Oeschger interstadial duration. *Paleoceanogr. Paleoclimatol*. **30**, 1595–1612 (2015).
90. McManus, J. F., Oppo, D. W. & Cullen, J. L. A. 0.5-million-year record of millennial-scale climate variability in the North Atlantic. *Science* **283**, 971–975 (1999).
91. Schulz, M., Berger, W. H., Sarnthein, M. & Grootes, P. M. Amplitude variations of 1470-year climate oscillations during the last 100,000 years linked to fluctuations of continental ice mass. *Geophys. Res. Lett*. **26**, 3385–3388 (1999).
92. Muglia, J. & Schmittner, A. Glacial Atlantic overturning increased by wind stress in climate models. *Geophys. Res. Lett*. **42**, 9862–9868 (2015).
93. Oka, A., Hasumi, H. & Abe-Ouchi, A. The thermal threshold of the Atlantic meridional overturning circulation and its control by wind stress forcing during glacial climate. *Geophys. Res. Lett*. **39**, GL051421 (2012).
94. Wang, Z. & Mysak, L. A. Glacial abrupt climate changes and Dansgaard-Oeschger oscillations in a coupled climate model. *Paleoceanogr. Paleoclimatol*. **21**, PA001238 (2006).
95. Ahn, J. & Brook, E. J. Atmospheric CO_2 and climate on millennial time scales during the last glacial period. *Science* **322**, 83–85 (2008).
96. Liu, Z. et al. Transient simulation of last deglaciation with a new mechanism for Bølling-Allerød warming. *Science* **325**, 310–314 (2009).
97. Dlugokencky, E. *Trends in Atmospheric Methane*. http://www.esrl.noaa.gov/gmd/ccgg/trends_ch4/ (Earth System Research Laboratory, 2018).
98. Toggweiler, J. & Russell, J. Ocean circulation in a warming climate. *Nature* **451**, 286–288 (2008).
99. Fischer, H. et al. Where to find 1.5 million yr old ice for the IPICS "Oldest-Ice" ice core. *Clim. Past* **9**, 2489–2505 (2013).
100. Raymo, M., Lisiecki, L. & Nisancioglu, K. H. Plio-Pleistocene ice volume, Antarctic climate, and the global $\delta^{18}O$ record. *Science* **313**, 492–495 (2006).
101. Schwander, J., Marending, S., Stocker, T. & Fischer, H. RADIX: a minimal-resources rapid-access drilling system. *Ann. Glaciol*. **55**, 34–38 (2014).
102. Alemany, O. et al. The SUBGLACIOR drilling probe: concept and design. *Ann. Glaciol*. **55**, 233–242 (2014).
103. Goodge, J. W. & Severinghaus, J. P. Rapid Access Ice Drill: a new tool for exploration of the deep Antarctic ice sheets and subglacial geology. *J. Glaciol*. **62**, 1049–1064 (2016).
104. Yan, Y. N. J. et al. 2.7-Million-Year-Old Ice from Allan Hills Blue Ice Areas, East Antarctica Reveals Climate Snapshots Since Early Pleistocene. ***Goldschmidt Conf. (Paris, France)*** 4359, https://goldschmidtabstracts.info/2017/4359.pdf (European Association of Geochemistry and the Geochemical Society, 2007).
105. Masson-Delmotte, V. et al. A comparison of the present and last interglacial periods in six Antarctic ice cores. *Clim. Past* **7**, 397–423 (2011).
106. Pages 2k Consortium. Continental-scale temperature variability during the past two millennia. *Nat. Geosci*. **6**, 339–350 (2013).
107. Steig, E. J. et al. Influence of West Antarctic Ice Sheet collapse on Antarctic surface climate. *Geophys. Res. Lett*. **42**, 4862–4868 (2015).
108. Korotkikh, E. V. et al. The last interglacial as represented in the glaciochemical record from Mount Moulton Blue Ice Area, West Antarctica. *Quat. Sci. Rev*. **30**, 1940–1947 (2011).
109. Bereiter, B. et al. Revision of the EPICA Dome C CO_2 record from 800 to 600 kyr before present. *Geophys. Res. Lett*. **42**, 542–549 (2015).
110. Bereiter, B. et al. Mode change of millennial CO_2 variability during the last glacial cycle associated with a bipolar marine carbon seesaw. *Proc. Natl Acad. Sci. USA* **109**, 9755–9760 (2012).
111. Gow, A. J., Ueda, H. T. & Garfield, D. E. Antarctic ice sheet: preliminary results of first core hole to bedrock. *Science* **161**, 1011–1013 (1968).
112. EPICA Community Members. Eight glacial cycles from an Antarctic ice core. *Nature* **429**, 623–628 (2004).
113. Slawny, K. R. *et al*. Production drilling at WAIS Divide. *Ann. Glaciol*. **55**, 147–155 (2014).
114. Neftel, A., Oeschger, H., Staffelbach, T. & Stauffer, B. CO_2 record in the Byrd ice core 50,000–5,000 years BP. *Nature* **331**, 609–611 (1988).
115. Barnola, J. M., Pimienta, P., Raynaud, D. & Korotkevich, Y. S. CO_2-climate relationship as deduced from the Vostok Ice Core—a reexamination based on new measurements and on a reevaluation of the air dating. *Tellus B* **43**, 83–90 (1991).
116. Fischer, H., Wahlen, M., Smith, J., Mastroianni, D. & Deck, B. Ice core records of atmospheric CO_2 around the last three glacial terminations. *Science* **283**, 1712–1714 (1999).
117. Monnin, E. et al. Atmospheric CO_2 concentrations over the last glacial termination. *Science* **291**, 112–114 (2001).
118. Pedro, J. B., Rasmussen, S. O. & van Ommen, T. D. Tightened constraints on the time-lag between Antarctic temperature and CO_2 during the last deglaciation. *Clim. Past* **8**, 1213–1221 (2012).
119. Parrenin, F. et al. Synchronous change of atmospheric CO_2 and Antarctic temperature during the last deglacial warming. *Science* **339**, 1060–1063 (2013).

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