LETTERS

Northern Hemisphere forcing of climatic cycles in Antarctica over the past 360,000 years

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The Milankovitch theory of climate change proposes that glacial–interglacial cycles are driven by changes in summer insolation at high northern latitudes[1]. The timing of climate change in the Southern Hemisphere at glacial–interglacial transitions (which are known as terminations) relative to variations in summer insolation in the Northern Hemisphere is an important test of this hypothesis. So far, it has only been possible to apply this test to the most recent termination[2,3], because the dating uncertainty associated with older terminations is too large to allow phase relationships to be determined. Here we present a new chronology of Antarctic climate change over the past 360,000 years that is based on the ratio of oxygen to nitrogen molecules in air trapped in the Dome Fuji and Vostok ice cores[4,5]. This ratio is a proxy for local summer insolation[5], and thus allows the chronology to be constructed by orbital tuning without the need to assume a lag between a climate record and an orbital parameter. The accuracy of the chronology allows us to examine the phase relationships between climate records from the ice cores[6–9] and changes in insolation. Our results indicate that orbital-scale Antarctic climate change lags Northern Hemisphere insolation by a few millennia, and that the increases in Antarctic temperature and atmospheric carbon dioxide concentration during the last four terminations occurred within the rising phase of Northern Hemisphere summer insolation. These results support the Milankovitch theory that Northern Hemisphere summer insolation triggered the last four deglaciations[3,10,11].

Antarctic deep ice cores have provided a wealth of information on glacial–interglacial climate change and atmospheric composition[6–9]. However, uncertainties of current timescales are large. For the Dome Fuji[7], Vostok[12] and Dome C[13] cores, timescales based on inverse methods using ice flow models disagree by up to ~5,000 years (~5 kyr), because of uncertainties in modelled accumulation rate and/or ice thinning[13]. Another approach is orbital tuning, in which a global atmospheric signal ($\delta^{18}O$ of O_2, or concentration of CH_4) is matched to presumed orbital forcing with an assumed lag[14,15]. This approach is generally affected by the difference between the ice age and the gas age, which adds error of the order of 1 kyr to the ice chronology[16], in addition to uncertainty in the assumed lag of several kyr (see below). Here we eliminate most of these uncertainties with a new orbital tuning that uses the O_2/N_2 ratio of trapped air, which is a proxy for the local summer insolation at the deposition sites[5].

Because O_2 molecules are preferentially excluded from freshly formed air bubbles during the snow–ice transition[17], the O_2/N_2 ratio of air trapped in Antarctic ice is depleted by 5–10‰ on average from the atmospheric ratio. The O_2/N_2 in the Vostok[5] and Dome Fuji[4] cores (Fig. 1) varies in accordance with local summer insolation with an amplitude of ~10‰, with stronger insolation leading to more O_2/N_2 depletion. Local summer insolation presumably controls the physical properties of near-surface snow, which in turn determines the magnitude of O_2/N_2 fractionation during the bubble close-off[5]. The O_2/N_2 records of Antarctic inland cores are expected to be in phase with summer solstice insolation. Support for this view comes from the present-day observation at Dome Fuji of maxima within several days of the solstice in snow temperature (10-cm depth) and vertical temperature gradient (between 10 and 50 cm) (Supplementary Fig. 1), both of which are important factors for snow metamorphism[5]. Positive feedback between snow grain size and absorption of solar radiation, as well as a negative relationship between surface albedo and solar zenith angle, increases the importance of the solstice[5].

Because of the physical link, the O_2/N_2 orbital tuning is distinct from traditional orbital tuning for ice and marine cores using global signals. The Dome Fuji and Vostok[5] O_2/N_2 records show no detectable climatic imprint (Supplementary Information), a feature that makes this proxy an ideal tuning tool. In particular, the O_2/N_2 records lack a nonlinear 100-kyr response, which has to be ignored or modelled during traditional orbital tuning. Our timescale is thus ideal for examining terminations. Another advantage is that our method gives the age of ice despite its use of the gas record, because the O_2/N_2 fractionation during bubble formation is determined by the physical properties of surrounding ice.

We derive chronological tie points between the O_2/N_2 record and local summer solstice insolation by peak-to-peak matching (Methods), to modify the original glaciological timescale[7] (DFGT-2003). We can use O_2/N_2 data only from the depths in which air is occluded as clathrate hydrates[4], which restricts the youngest tie point to ~82 kyr ago (all dates given are relative to AD 2000). The DFGT-2003 timescale, which has a constraint at ~41 kyr ago, is unchanged for 0–41.5 kyr ago and its agreement with Greenland timescales is probably within 1 kyr, based on a comparison with the Byrd-core isotope record on the GISP2 timescale[18]. We linearly interpolated the DFGT-2003 timescale between all the O_2/N_2 tie points, as well as between 41.5 and 82 kyr ago.

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We applied our method also to the Vostok core using published data[5,12] (Fig. 1). The new Vostok timescale (VkO-2006) resulted in an agreement with the Dome Fuji O_2/N_2 timescale to within 1 kyr for most of the overlapping period (~200–340 kyr ago), indicating that our method is robust. A large discrepancy was found only where the Dome Fuji O_2/N_2 data are too noisy to reliably constrain the timescale (220–230 kyr ago, Fig. 1b and c). To improve and extend the Dome Fuji O_2/N_2 timescale, we adopted tie points at 221.2, 230.8 and 334.9 kyr ago (Fig. 1c) from the Vostok core, by transferring them to the Dome Fuji core by visual matching of the isotopic records[7,12] (the final Dome Fuji timescale is named DFO-2006). We note that an independent study[19], limited to Vostok, used new[19] and published[5] O_2/N_2 data to derive a similar timescale, and discussed palaeoclimate implications.

The dating uncertainties at the tie points (Fig. 1c) due to the noise in the O_2/N_2 data (for example, from non-orbital natural variation, experimental error and gas loss correction) are estimated with a Monte Carlo method to range from 0.8 to 2.9 kyr (Methods). The uncertainty between the tie points originating in the DFGT-2003 timescale and its linear interpolation is estimated using an autocorrelated random walk model (Fig. 1e; Supplementary Information). The accuracy of the DFO-2006 timescale is validated through comparison with four radiometric-dated time markers spanning 92–131 kyr ago, all of which agree within the uncertainty (Supplementary Information).

This accurate chronology permits a critical examination of the phase relationship between Antarctic climate and orbital variations (Fig. 2). We examine the $\delta^{18}O_{ice}$ record of the Dome Fuji core and a preliminary reconstruction of 'site temperature' (ΔT_{site}) based on an isotopic inversion, which accounts for the influence of seawater $\delta^{18}O$ (hereafter $\delta^{18}O_{sw}$) and vapour source temperature[20] (Methods). Our chronology also indirectly gives the timing of the CO_2 rise at terminations, which occurs within ~1 kyr of the increase in Antarctic temperature[21].

A striking feature of Dome Fuji $\delta^{18}O_{ice}$ and ΔT_{site} records is that they appear to reflect every precessional-band variation of the northern insolation curve (Fig. 2), even for Marine Isotope Stages (MIS) 6c and 8c where changes are small, as well as the weak variation in MIS 3. The result confirms earlier observations[6,22] of the similarity of the patterns of northern insolation and the Antarctic climate, and further provides the phasing between them. The $\delta^{18}O_{ice}$ power spectrum (Fig. 3) shows strong power in the precession band (23-kyr period) with a mean lag of 1.0 ± 0.5 kyr behind precession. The site temperature ΔT_{site} shows smaller obliquity and precession components than $\delta^{18}O_{ice}$, and the precession component is as strong as the obliquity component (Fig. 3). The lag of ΔT_{site} behind precession is 1.8 ± 0.5 kyr. At 41-kyr period, $\delta^{18}O_{ice}$ and ΔT_{site} lag behind obliquity by 2.1 ± 0.7 and 4.7 ± 1.1 kyr, respectively. Importantly, the southern summer insolation (Fig. 2) is in anti-phase (or is completely out of phase) with Antarctic climate, so does not appear to have a strong influence on the orbital-scale Antarctic climatic changes. These phasings suggest that the Antarctic climate on orbital timescales is paced by northern summer insolation presumably through northern ice sheet variation, although minor influence by southern insolation cannot be entirely ruled out. Our data therefore disfavour hypotheses that call for early Southern Hemisphere warming by southern summer insolation to trigger the northern deglaciation[23].

Another remarkable feature of the data is that the last four Antarctic terminations occur within the rising phase of northern summer insolation. We examine the onset of rapid warming towards interglacial warmth, excluding prior gradual warming for terminations II and III (Fig. 4). Because the gradual warming is attributable to a linear climatic response to northern summer insolation for all climatic substages, they should be separated from terminations when examining hypotheses[10,14,23,24]. The onsets are distinctly defined in the Antarctic climate records for terminations I, II and IV (Fig. 4). We identify the onset of termination III as the start of the continuous warming at 247 kyr ago after a brief temperature reversal, but our conclusion is unchanged if the prior small peak is included. These onsets occur 6, 3, 7 and 2 kyr later than the 65° N summer solstice insolation minima, respectively, for terminations I, II, III and IV (Supplementary Table 1), and the entire duration of these warmings also fit within the rising phase of northern summer insolation (Fig. 4). The timing of the last four terminations is thus fully consistent with Milankovitch theory[3,10]. One possible explanation for the difference in the timings of terminations relative to insolation is the hypothesis

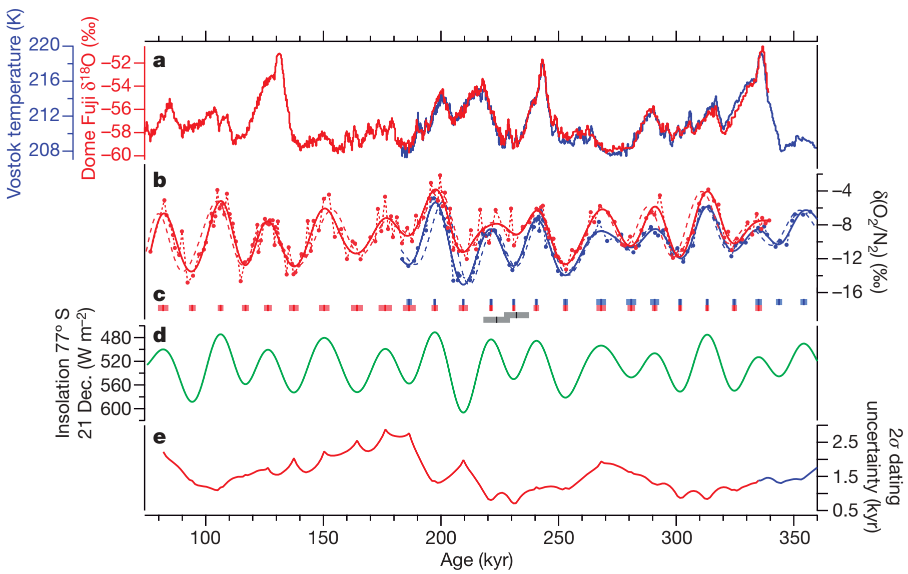

Figure 1 | Orbital tuning of the Dome Fuji and Vostok timescales using O_2/N_2 records. a, Dome Fuji $\delta^{18}O_{ice}$ (red, ref. 7) and Vostok temperature[12] (blue, converted from δD) on the respective O_2/N_2 timescales (DFO-2006 and VkO-2006). **b**, O_2/N_2 records of the Dome Fuji[4] (filled red circles) and Vostok[5] (filled blue circles) cores and filtered curves (solid lines) on the O_2/N_2 timescales, and filtered curves on the original glaciological timescales (dashed lines). **c**, Age tie points for the Dome Fuji (red) and Vostok (blue) cores with 2σ error bars. The tie points at 221.2, 230.8 and 334.9 kyr ago for DFO-2006 are adopted from the Vostok data (see text). Two discarded points of the Dome Fuji core are shown (grey markers). **d**, Summer solstice insolation at 77° S as the tuning target (inverted axis scale). **e**, 2σ dating uncertainty of DFO-2006 (red) and VkO-2006 (blue, for oldest part) (see Supplementary Information).

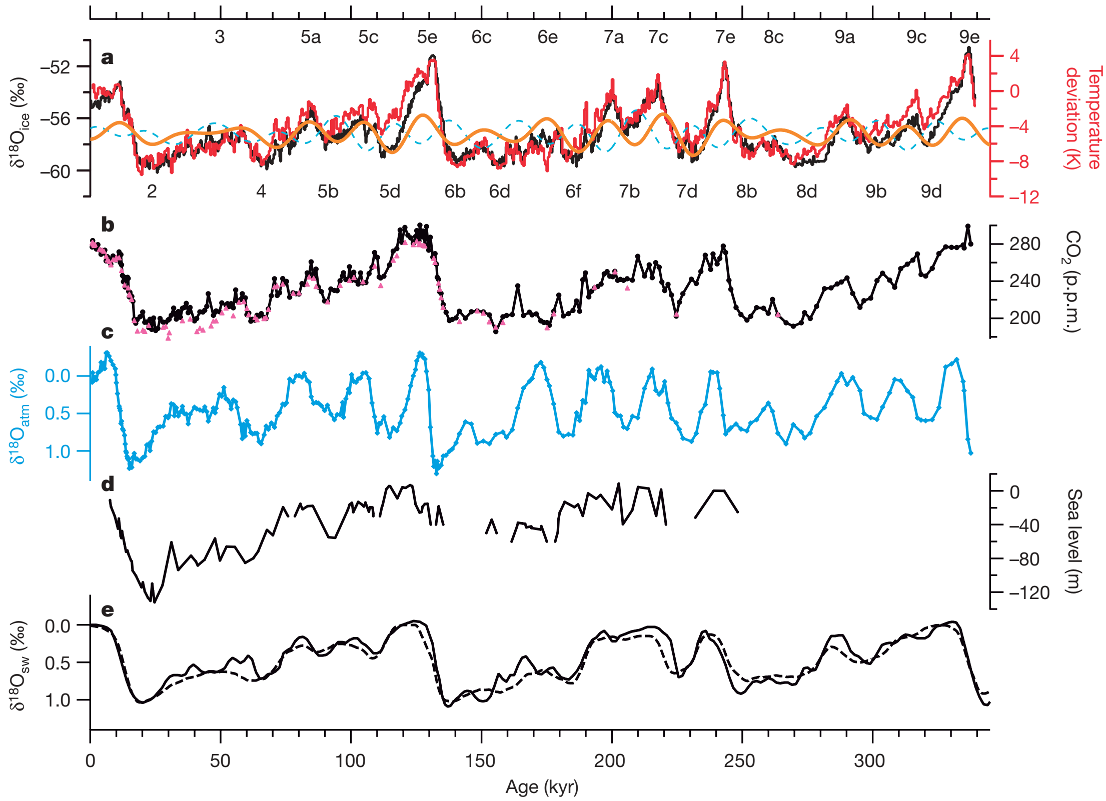

Figure 2 | Comparison of Dome Fuji climatic and atmospheric records with insolation and sea level records. **a**, Dome Fuji $\delta^{18}O_{ice}$ (black, ref. 7) and temperature deviation (ΔT_{site}) from the mean of the last 10 kyr (red, this study) on the DFO-2006 timescale, and summer solstice insolation at 65° N (orange) and 65° S (light blue dashed line). The insolation curves are scaled to match 65° N insolation with low-pass filtered $\delta^{18}O$ (not shown) for the MIS 5b–a transition. Corresponding MIS numbers[27] are shown without implying their exact timings. The ΔT_{site} and $\delta^{18}O_{ice}$ records are scaled to visually match the changes from the LGM to early Holocene. **b**, Atmospheric CO_2 concentration from the Dome Fuji core by wet extraction[8] (black circles) and dry extraction (pink triangles, this study). (Two wet-extraction CO_2 outliers[8] (at 149.8 and 365.0 m, in Holocene ice) are excluded from this plot. Experimental error rather than chemical reaction in meltwater[8] may be suspected for these samples, which were measured on the first day of ~80 measurement days.) Note that the dry extraction values are too low in the transition zone from air bubbles to clathrate hydrates (for ~30–55 kyr ago), and the wet-extraction values are too high in the LGM[8] (around 20 kyr ago). **c**, Dome Fuji $\delta^{18}O_{atm}$ record (this study, inverted axis scale). **d**, Sea level reconstruction based on radiometric dating of fossil corals with open system correction[28]. **e**, Orbitally tuned $\delta^{18}O_{sw}$ (inverted axis scale) reconstructed from isotope records of marine sediment cores through regression analyses[29] (solid line) and ice sheet modelling[30] (dashed line).

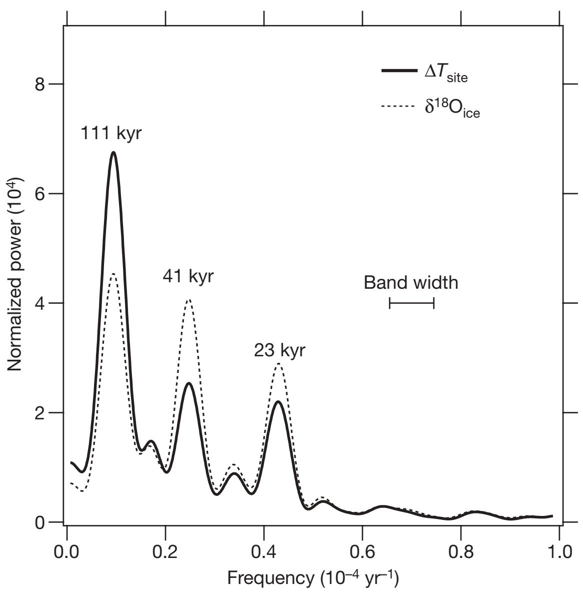

Figure 3 | Power spectra of $\delta^{18}O$ and ΔT_{site} records calculated by the Blackman–Tukey method with 50% lags. The results (obtained using AnalySeries software) are normalized so that the integrated power over periods longer than 10 kyr becomes unity. Periods of three spectral peaks are shown.

that obliquity and ice volume may modulate the exact phasing between climate and northern summer insolation[11].

The onset of the last Antarctic termination was possibly linked to the 19-kyr meltwater pulse, which ended the Last Glacial Maximum (LGM)[2,25,26]. If the same mechanism holds for the previous terminations, the initial rapid ice volume decrease is predicted to coincide with or precede the onset of Antarctic terminations. A U–Th-dated marine core places the onset of $\delta^{18}O_{sw}$ change at ~138 kyr ago[23] for termination II, in agreement with the Antarctic timing (137.0 ± 2.2 kyr ago). Therefore, contrary to hypotheses ascribing the trigger of glacial terminations to CO_2 (ref. 14), obliquity[24], or southern summer insolation[23], our chronology implicates northern summer insolation as the primary trigger[3,10,11].

The major part of post-interglacial Antarctic cooling (MIS 9e–d, 7e–d, and 5e–d transitions) also coincides with decreasing 65° N summer solstice insolation (Fig. 4). Comparison of the Dome Fuji ΔT_{site} with the CO_2 concentration and a radiometric-dated sea level reconstruction (Fig. 2d) as well as orbitally tuned $\delta^{18}O_{sw}$ records (Fig. 2f) reveals that the Antarctic cooling for the MIS 5e–d and 7e–d transitions began earlier by several millennia than the corresponding CO_2 and sea level drops (Fig. 2). This suggests that post-interglacial cooling began in the Northern Hemisphere with ice area growth and was transferred to Antarctica quickly through modulation of poleward heat transport and methane concentration decrease[9]—before the reduced CO_2 forcing, or the sea level drop caused by northern ice volume growth, became significant.

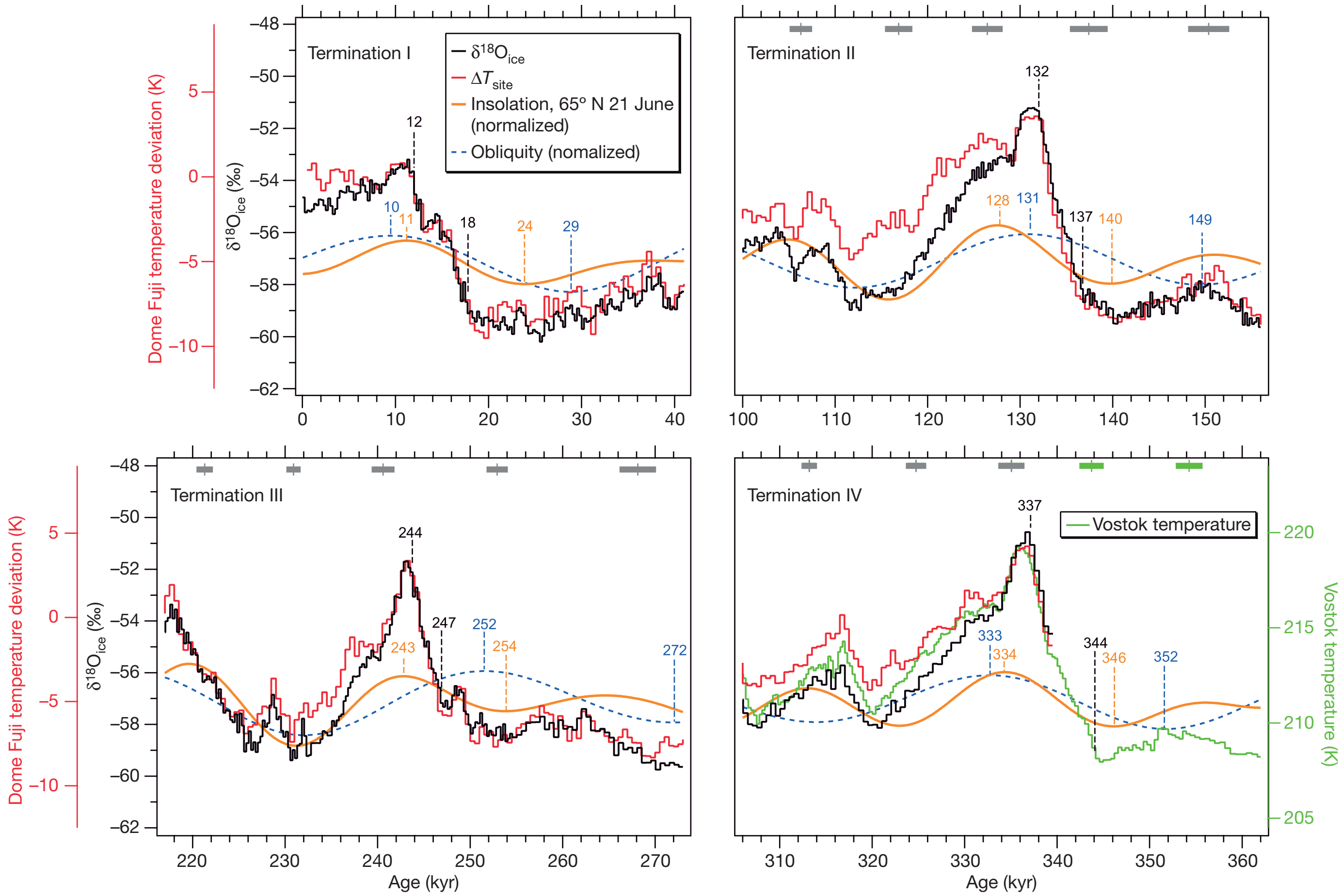

Figure 4 | Comparison of Antarctic parameters with insolation and obliquity around the last four terminations. Shown are Dome Fuji $\delta^{18}O_{ice}$ (black, ref. 7) and temperature ΔT_{site} (red, this study) on the DFO-2006 timescale compared with 65° N summer solstice insolation (orange) and obliquity (blue dashed line). The Vostok temperature[12] on the VkO-2006 timescale is also plotted for termination IV (green). Numbers indicate key timings (minima and maxima of insolation and obliquity, and onset and end of $\delta^{18}O_{ice}$ increases). The age tie points with 2σ uncertainties are shown at the top.

The phasing between the isotopic composition of atmospheric molecular oxygen ($\delta^{18}O_{atm}$) and northern summer insolation varies by ±6 kyr (Fig. 2), showing that our timescale is inconsistent with $\delta^{18}O_{atm}$-orbital tuning[14]. In addition, prior studies have used $\delta^{18}O_{atm}$ as a proxy for $\delta^{18}O_{sw}$ and thus sea level[6,14]. However, the fact that $\delta^{18}O_{atm}$ lagged the 19-kyr meltwater pulse by 4 kyr indicates a large change of the fractionation between atmospheric and seawater $\delta^{18}O$ (known as the Dole effect) in the interval 19–15 kyr ago, creating the false impression[6] that the onset of sea level rise lagged the Antarctic temperature and CO_2 increases. The Dole effect also drove $\delta^{18}O_{atm}$ change after ~7 kyr ago, when $\delta^{18}O_{atm}$ shows a clear shift to glacial-like values despite the stable sea level (Fig. 2). The DFO-2006 timescale reveals another Dole-effect-forced lead of $\delta^{18}O_{atm}$ to $\delta^{18}O_{sw}$ by ~10 kyr at the MIS 5e–d transition (Fig. 2). These observations indicate that the Dole effect involves complex glacial–interglacial changes, and thus orbital tuning and sea level reconstruction based on $\delta^{18}O_{atm}$ (ref. 14) are not accurate enough to assess the causality of glacial cycles[3]. Rather, $\delta^{18}O_{atm}$ should be used for reconstruction of the past Dole effect based on the accurate timescale, which is of biogeochemical interest[5].

In summary, the mean phasing of Antarctic climate, as well as the timing of the last four terminations and three post-interglacial coolings, are consistent with the hypothesis that high northern latitude summer insolation is the trigger of glacial–interglacial cycles. The role of CO_2 as conveyor and amplifier of the orbital input should be quantified with climate models run using our new timescale; this quantification is important for future climate change predictions. Our timescale should be validated further with new radiometric age markers, as well as by process studies for complete understanding of the physical link between O_2/N_2 and local insolation. With future O_2/N_2 measurements, it may be possible to apply this method to the Dome Fuji and Dome C cores for termination V and older terminations, to investigate the phasing of climate and atmospheric composition with respect to orbital forcing further back in time.

METHODS SUMMARY

All gas measurements on the Dome Fuji core were carried out at Tohoku University. Ice-core air was extracted with a melting technique[8] and measured with a mass spectrometer (Finnigan MAT Delta-S) with 1σ uncertainty of 0.2‰, 0.02‰ and 0.04‰ for $\delta(O_2/N_2)$, $\delta^{15}N$ of N_2 and $\delta^{18}O$ of O_2, respectively. The O_2/N_2 and $\delta^{18}O$ data were corrected for gravitational enrichment in the firn by subtracting $4\times\delta^{15}N$ and $2\times\delta^{15}N$, respectively. Because O_2/N_2 in clathrate ice gradually becomes depleted over years in a warm freezer (−25 °C) owing to diffusive gas loss[4], the data were corrected according to the storage period, and outliers were rejected. The new CO_2 data were obtained with a dry extraction method and gravitationally corrected.

Dome Fuji tuning tie points were made for ~82–335 kyr ago at all maxima and minima of a low-pass filtered O_2/N_2 curve and summer solstice insolation at 77° S. The DFGT-2003 glaciological timescale[7] was modified according to the tie points (see text). The Vostok tie points were made for ~186–354 kyr ago to modify the Vk-FGT1 timescale[12]. Uncertainties at and between the tie points were estimated by Monte Carlo simulations (Supplementary Information).

Gas age of the Dome Fuji core for CO_2 and $\delta^{18}O_{atm}$ was derived by subtracting ice age–gas age difference (Δage) from the DFO-2006 ice age, using a densification model. The Δage depends primarily on accumulation rate, ranging from 1.6 to 4.7 kyr with an accuracy better than 20%.

The Dome Fuji temperature ΔT_{site} was reconstructed with an isotopic inversion using a low-pass filtered deuterium-excess (d-excess) record. Note that the

ΔT_{site} presented here is not intended to resolve millennial changes because of the filtering as well as averaging of ΔT_{site} to 500-yr intervals. Future improvements to the ice core data and traverse snow isotope data from the coast to Dome Fuji will be conducted.

Full Methods and any associated references are available in the online version of the paper at www.nature.com/nature.

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Supplementary Information is linked to the online version of the paper at www.nature.com/nature.

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METHODS

Data correction and selection. Because O_2/N_2 in the clathrate ice becomes gradually depleted over years during sample storage in a warm freezer (for example, $-25\,^{\circ}C$) due to diffusive gas loss[4], the O_2/N_2 data were corrected according to the storage period assuming a linear O_2/N_2 depletion with time up to 639 days (Supplementary Fig. 2). The samples stored for more than 800 days do not appear to deplete further, thus those data were corrected by the same correction for 639 days. When plotted against age, the corrected O_2/N_2 record shows a smoother and more regular variation than the uncorrected data (Supplementary Fig. 3), and it resembles the independent Vostok record[5], indicating that the correction is reasonable. We note that using different fits to the depletion trend (exponential function or linear trend throughout all storage durations) result in identical timescales within 2σ uncertainty. Because the post-coring fractionation of bubble ice is not currently understood[4], the data from shallower depths ($<$1,200 m or younger than $\sim$75 kyr ago) are not suitable for dating[31].

After the correction, outliers in the O_2/N_2 data were rejected on the basis of deviation from a low-pass filtered curve (Supplementary Fig. 4). 28 points out of the original 185 points were rejected with a criterion of 3.2‰. The high noise levels for the period younger than $\sim$230 kyr ago are attributable to errors in the large gas-loss corrections for the samples stored at $-25\,^{\circ}C$ for a long time. Most of the outliers indeed consist of the longest-stored samples. The Dome Fuji O_2/N_2 data can be improved in the future by measuring ice samples stored in a cold freezer ($-50\,^{\circ}C$; ref. 4) or at Dome Fuji. Measuring O_2/N_2 within a few months after drilling would be ideal for future studies.

Orbital tuning of O_2/N_2 record to local summer solstice insolation. The tuning tie points were made between the O_2/N_2 record and summer solstice insolation at 77° S by peak-to-peak matching (Fig. 1). Note that use of average insolation for a few months around the summer solstice as a tuning target produces identical results within 0.2 kyr. For detecting the maxima and minima in the O_2/N_2 data, as well as for the outlier rejection, the O_2/N_2 data on a glaciological timescale[7] (DFGT-2003) were linearly interpolated at 0.1-kyr intervals, and then smoothed by a 559-order finite-duration impulse response (FIR) filter with a Kaiser window (sharp cut-off from 16.7 to 10.0 kyr period; pass-band ripple of less than 1%; attenuation at stop-band of more than 99.9%), which was designed with the Filter Design and Analysis Tool of the MATLAB software. To avoid phase modulation, the filter was applied twice in the opposite direction (filtfilt function in MATLAB). The peak positions in the filtered curve were used for the matching. We checked the filter performance by applying it to the insolation curve, and confirmed that the peak positions are not altered by more than 0.1 kyr. The timings of the peaks in the filtered O_2/N_2 curve were tied to those in the local summer solstice insolation. Mid-transitions were not used because the relationship between O_2/N_2 and insolation may be nonlinear. Also, the use of mid-transition instead of peaks does not improve the precision by our method. The DFGT-2003 timescale was linearly interpolated between the tie points to derive the O_2/N_2 timescale (DFO-2006). Two tie points at $\sim$221 and $\sim$230 kyr ago are transferred from the Vostok O_2/N_2 data because the Dome Fuji data are too noisy in this part. We also transferred the tie point at $\sim$333 kyr ago from the Vostok data to better constrain the timescale in the oldest part of the current record.

Ice age–gas age difference. The age of gas in the Dome Fuji core for the $\delta^{18}O_{atm}$ and CO_2 records was derived by subtracting an estimate of ice age–gas age difference (Δage) from the DFO-2006 ice age. The Δage was estimated by a semi-empirical firn densification model[32] after adjustment of model parameters to reproduce the present density profile (similar to ref. 8). The uncertainty of Δage is estimated by comparing Antarctic Isotope Maxima[33] (AIM) in the Dome Fuji $\delta^{18}O_{ice}$ with the abrupt increases of CH_4 (ref. 31), assuming that these two signals are synchronous[18,33]. The estimated 2σ uncertainties are 8–18% of Δage at AIM 1, 2, 8, 14, 17, 23 and 24 (the sampling intervals of CH_4 data for other AIM are too low to usefully constrain the uncertainty). We thus conservatively assign a constant fraction of $\pm$20% as the 2σ uncertainty of Δage (for example, 0.9 kyr in the LGM, 0.5 kyr for the Holocene).

Dry extraction method for CO_2 measurements. The dry extraction method for the CO_2 concentration data has been described elsewhere[34]. Slight modifications were made as follows. First, the extraction apparatus was pre-conditioned by hot humid air to minimize contamination[8]. Second, a helium cycle cooler (Janis Research) was used instead of liquid helium for cryogenic collection of extracted air. Third, shorter sample tubes were used. Forth, the ice vessel was evacuated for 90 min.

Site temperature reconstruction. An isotopic inversion technique[20,35] was used to correct the raw isotopic record for both vapour source temperature and marine isotopic composition variations. Owing to the high level of noise in the deuterium excess data especially between 20 and 80 kyr ago, we use a filtered record (passband $>$ 5,000 yr, stopband $<$ 2,000 yr), as we only discuss the large climate variations. We used a recent marine $\delta^{18}O_{sw}$ reconstruction[30]. We estimate the sensitivities of isotopic values[36] to site (Dome Fuji) and source temperatures and marine isotopic composition by using a Rayleigh-based model. We tune this model with the Dome Fuji present-day surface data. Unfortunately, we currently lack isotopic data of surface snow from the coast to Dome Fuji to correctly define the air mass trajectories. Our modelling thus does not offer comprehensive constraints, and future improvements are necessary to interpret ΔT_{site} in detail. Nevertheless, the sensitivities we found are close to those obtained for Vostok[20] and Dome C[35] (see Supplementary Information).

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