Seasonal temperatures in West Antarctica during the Holocene

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The recovery of long-term climate proxy records with seasonal resolution is rare because of natural smoothing processes, discontinuities, and limitations in measurement resolution. Yet insolation forcing, a primary driver of multi-millennial-scale climate change, acts through seasonal variations with direct impacts on seasonal climate[1]. Whether the sensitivity of seasonal climate to insolation matches theoretical predictions has not been assessed over long timescales. Here, we analyze a continuous record of water-isotope ratios from the West Antarctic Ice Sheet (WAIS) Divide ice core to reveal summer and winter temperature changes through the last 11,000 years. Summer temperatures in West Antarctica increased through the early-to-mid Holocene, reached a peak at 4.1 ka, and then decreased to the present. Climate model simulations show that these variations primarily reflect changes in maximum summer insolation, confirming the general connection between seasonal insolation and warming, and demonstrating the importance of insolation intensity rather than seasonally integrated insolation or season duration[2,3]. Winter temperatures varied less overall, consistent with predictions from insolation forcing, but also fluctuated in the early Holocene, likely owing to changes in meridional heat transport. The magnitudes of summer and winter temperature changes constrain the lowering of the WAIS surface since the early Holocene to less than 162 m, and most likely less than 58 m, consistent with geological constraints elsewhere in West Antarctica[4,5,6,7].

Milankovitch famously postulated that variations of Earth's orbit and axis drive climate changes over tens of thousands of years by altering the seasonal cycle of insolation[1]. By controlling summer temperatures and ice ablation, summer insolation in the northern high latitudes is thought to drive global ice volume changes over glacial-interglacial timescales[8]. While modeling studies support this idea[9,10], empirical evidence of the specific climate response to insolation changes derives almost entirely from mean-annual temperature reconstructions[11,12] or from indirect effects on, for example, trapped gases and melt layers in polar ice[13,14] and marine eolian deposits[15]. The absence of seasonal temperature reconstructions has precluded direct evidence of insolation forcing on seasonal climate, a relationship that may vary geographically. In Antarctica, long records of multiple glacial-interglacial cycles have supported different claims about whether summer insolation's effects relate most strongly to its maximum intensity, its seasonal integral, or to duration above a threshold[2,3,16,17]. Site-specific empirical determinations would provide valuable tests of such competing ideas.

Seasonal Temperature Reconstructions

We reconstructed seasonal temperature variability in West Antarctica through the Holocene (the last 11,000 years) and performed new model experiments to understand its physical controls. The Holocene offers a window of time for assessing the influence of orbital forcing without the complicating effects of Northern Hemisphere deglaciation[18]. Our reconstruction (Figures 1,2) uses the high-resolution water isotope record from the WAIS Divide ice core (WDC)[18,19,20] (Methods Section 1, ED Figure 1a,b), obtained with a continuous-flow technique that provides millimeter-scale depth resolution[21]. Layer ages were determined previously[2,22].

Records of seasonal temperatures from ice cores are limited by measurement resolution and information loss from water-isotope diffusion. In Greenland, the longest records separating summer and winter variability extend to only 2 ka[23,24], while only climate model simulations are available for older periods[10]. For Antarctica, prior to the present study, the longest records spanned only a few centuries[25]. A combination of three factors account for the considerably greater scope of our reconstruction: exceptional depth resolution of measurements, conditions at WAIS Divide (high accumulation, low temperature, thick ice) that allow for preservation of sub-annual information through the entire Holocene[26], and an analysis strategy that circumvents interannual noise by evaluating millennial averages of the seasonal parameters.

Our method corrects water-isotope variations for diffusion[26,27,28] and assesses uncertainties including preservation bias and precipitation intermittency (Methods Sections 2 and 3). The diffusion correction operates on the high-resolution data and produces isotopic time-series from which seasonal summer-winter amplitudes were extracted. These were converted to temperature using a model-derived scaling[29] (6.96 per mil δD/°C; Methods Section 2.2) and added to previously reconstructed annual-mean temperatures[30] to obtain summer and winter histories.

Seasonal Trends

Summer temperatures at WAIS Divide (Figure 2a) generally rose through the early and middle Holocene, persisted at a maximum between ~5 and 1.5 ka, then decreased toward the present, with a total Holocene range of around 2°C. These variations broadly correlate with local maximum insolation, rather than with integrated summer insolation or the duration of summer (Figure 3d,e). Winter temperatures varied less than summer ones overall (~1°C range), but also fluctuated at ~10 to 8 ka, a variation too rapid to attribute to orbital forcing.

Annual mean WAIS Divide temperature changes[30] (Figure 2e) were considerably influenced by winter variability in the early Holocene, whereas summer variability dominates the overall Holocene pattern (Methods Section 3.4, ED Table 2). Summer variability also accounts for most of the cooling in the last 2 kyr, indicating that the ~1°C annual-average cooling of the entire West Antarctic during this period[31,32] likewise reflects this season. Neither season at WDC experienced the early Holocene optimum nor overall Holocene cooling that appears in some global temperature reconstructions[33,34]. To assess the significance of the dominant multi-millennial trends in each season, we performed Monte Carlo analysis (Methods Section 3.5) using 4 ka as a demarcation point in summer (this is the timing of maximum summer temperature) and 6 ka in winter (when winter temperatures plateau). For summer (Figure 2b) this indicates a >95% chance that warming from 11 to 4 ka and cooling from 4 ka to present exceeded 0.7°C and 0.6°C, respectively. For winter, the trend from 11 to 6 ka is indistinguishable from zero, while cooling of greater than ~0.3°C from 6 to 0 ka occurred with >95% likelihood (Figure 2d).

Moist Energy Balance Model

To evaluate how orbitally-driven insolation changes may explain the WAIS Divide reconstructed temperatures (Figure 2), we first simulated temperature history at 80°S using a global, zonal-mean (2° resolution) moist energy balance model (MEBM) accounting for incoming and outgoing radiation, albedo, and meridional atmospheric heat transport (Methods Section 4). The model is driven by top-of-atmosphere seasonal insolation changes (Figure 3a-e); for this latitude, the maximum summer insolation increases until ~2.5 ka and annual mean and annual- and summer-integrated values mostly decline through the Holocene. The calculations yield summer maximum temperatures and seasonal temperature amplitudes (Figure 3g) that covary with local maximum summer insolation (Figure 3e), and with the general pattern of our reconstructed summer temperatures (ED Figure 7). While heating at lower latitudes can influence Antarctic temperature through atmospheric and oceanic heat transport, modeled maximum summer temperatures at WAIS Divide correlate best with local insolation (70 to 90°S, $R^2 = 0.9$, $p << 0.001$ during 0 to 6 ka) rather than insolation anywhere in the subtropical through subpolar latitudes (20 to 60°S, $R^2 = 0.33$ to 0.55, $p < 0.05$). Indeed, models indicate heat export from WAIS Divide in summer (ED Figure 4k), rather than import from more northern locations. Since December is always the month of maximum insolation (Figure 3a-c), variability of December insolation dominates the response of maximum summer temperature. For winter, modeled temperatures are less variable than those of summer at 80°S (Figure 3g) due to the lack of direct insolation (Figure 3b), and have an opposite trend. Winter minima are a function of three factors: changes in the length of the zero-insolation season, the effective cooling rate of the surface, and convergent heat transport from lower latitudes. Lower minimum winter temperatures occur at times when the zero-insolation season is longer. However, neither the length of the zero-insolation season, modeled minimum temperatures, nor winter heat divergence correlate well with reconstructed winter temperatures.

HadCM3 Simulations

To investigate the role of more-complex geography and mechanisms, including topographical changes not accounted for in the MEBM, we simulated Holocene climate with a fully-coupled general circulation model, HadCM3[36] (Methods Section 5). Simulations forced solely by changes in orbital parameters produce summer maximum temperatures (for approximately the December solstice) at 80°S similar to our reconstructed values and to the MEBM: increasing over the Holocene, peaking at 4 to 3 ka, and decreasing into the modern (ORBIT, Figure 2a). This pattern reflects a strong role of maximum summer insolation in determining observed summer temperatures. The similarity of the early- to mid-Holocene (11-6 ka) summer temperature increase in the orbitally forced HadCM3 simulations and our reconstruction suggests little influence of changing ice-sheet elevation and extent. A similar comparison for winter yields a ~1.25°C decrease of model ORBIT temperature (Figure 2c) compared to a possible small increase in temperature in the reconstruction (Figure 2d; >90% chance of >0.1°C), suggesting some warming due to a lowering ice sheet.

Next, as boundary conditions in the HadCM3 simulations, we prescribed variable greenhouse gas concentrations and two different ice sheet histories, GLAC1D and ICE-6G, which entail net surface lowerings of ~83 m and ~208 m, respectively, from 11 to 7 ka at the WDC site (Figure 4a). These elevation scenarios substantially affect simulated temperatures (Figure 2a,c). Much of the elevation-induced warming in these models, which occurs primarily in the early Holocene, can be attributed directly to the surface lapse-rate effect (Figure 4b). However, comparison to the orbital-only runs (Figure 4c) reveals a remaining temperature anomaly (Figure 4d), attributable to greenhouse gases, ice sheet extent, and nonlinear responses to simultaneously imposed forcings. Sea ice has only a small impact on the temperature at 80°S in summer (Methods Section 5.4, ED Figure 6).

Inconsistencies exist between the different ice sheet scenarios (Figure 4d) and the summer vs. winter seasons, but differences are minor enough to permit a bounded estimate of the true Holocene elevation decrease. This calculation is made by comparing the excess of the reconstructed temperature increase over the orbital-only simulation to the same excess for the ice sheet model simulations, and scaling to the elevation changes used in the latter (Methods Section 6). We find central estimates for elevation decrease of 23 m and 53 m from comparison to the GLAC-1D and ICE-6G scenarios, respectively, over the period 10 to 3.5 ka (Table 1). Accounting for uncertainties in the seasonal temperature reconstructions (Figure 2) allows for elevation changes ranging from 33 m increase to 131 m decrease (2σ) from 10 to 3.5 ka, or 54 m increase to 162 m decrease (2σ) if the time interval is narrowed to 10 ka to 6.5 ka (Table 1). Our results thus are consistent with geological observations of ice high-stands on mountain nunataks, which indicate less than 100 m of Holocene surface lowering[4,5,6].

Winter temperatures on the Antarctic mainland must respond to insolation forcings indirectly, via heat transport from lower latitudes. Orbital forcing models predict winter cooling across the Holocene, mostly from 11-6 ka (Figure 2c, 3g). Both models and reconstructed winter temperatures lack a late Holocene maximum. But in the earlier Holocene, the winter reconstruction does not display the cooling trend expected from models and is dominated by prominent millennial variations. The mismatch with insolation at lower latitudes and absence of local forcings suggests variations in the efficacy of meridional atmospheric heat transport.

Discussion

Diverse and numerous proxies are used to reconstruct globally-averaged surface temperatures for evaluating climate models and distinguishing natural from anthropogenic climate variability[33,34,41,42,43]. How these proxies depend on seasonal factors has been assessed in only a few cases[44]. Our West Antarctic study provides a cautionary example, as the mean-annual temperature history reflects different controlling factors of summer and winter temperatures whose importance varies with time. In such a situation, important seasonal dynamics may be missed, or proxies misinterpreted, when only mean climate is considered. In addition, incorporating more information from the Southern polar regions should help global temperature assessments avoid biases associated with weighting of temperature reconstructions toward northern sites, which have produced differing interpretations of the relationship between global climate and forcings in the Holocene, even including opposing trends[34,45,46].

Prior analyses with simplified atmospheric models[3] identified the duration of Southern Hemisphere summer as a key driving variable of Antarctic climate at orbital timescales. Some paleoclimate findings validate this claim; for example, the onset of deglacial warming in West Antarctica corresponds with increasing integrated summer insolation[2]. Our results – spanning about half a precession-cycle – reveal a dominant role for annual maximum insolation in determining West Antarctic summer climate during the Holocene, without precluding a greater role for duration or integrated summer insolation in other periods, such as glacial terminations.

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Table 1 | WAIS Elevation Change. *Elevation decrease estimates in meters (2σ, positive values correspond to a lowering ice sheet) for the intervals 10 to 6.5 ka and 10 to 3.5 ka (Methods Section 6).*

	GLAC1D			ICE-6G		
Interval	-2σ	Nominal	2σ	-2σ	Nominal	2σ
10 to 6.5 ka	-9.93	25.67	58.75	-54.00	57.52	161.96
10 to 3.5 ka	-5.63	22.99	48.49	-33.23	52.63	130.53

Figure 1 | Water isotope seasonal variability. ***a,*** *Example section of the diffusion-corrected (solid line) and raw[20] (dashed line) WDC δD records, with annual maxima (red circles) and minima (blue circles) determined algorithmically (Methods Section 2.1). Extended Data (ED) Figure 1 provides the full high-resolution WDC δD record, diffusion lengths, and extrema.* ***b,*** *50-yr annual-amplitude averages (summer minus winter divided by 2), with 2σ uncertainty; horizontal line indicates Holocene mean.* ***c, d, e,*** *50-yr δD averages for summer (red), mean (purple), and winter (blue); horizontal line indicates Holocene mean; shaded regions are 2σ bounds for combined analytical and diffusion-correction uncertainty.*

Figure 2 | Seasonal temperature reconstruction. ***a,c,*** *Reconstructed summer and winter temperatures at WDC for 1,000-year averaging (solid red and blue lines). Shaded regions are 1σ and 2σ uncertainty ranges for combined uncertainties arising from analysis, diffusion-correction, seasonality of accumulation, precipitation intermittency, isotope-temperature scaling, and reconstructed mean temperatures (Methods Section 3). Also shown are Moist Energy Balance Model (MEBM) calculated temperatures for 80°S (maximum and minimum annual values) and HadCM3 zonal temperatures for 80°S (late-December for summer, mid-August for winter) (ORBIT, GLAC1D, ICE-6G). The 0 ka ORBIT simulation uses pre-industrial settings, a calculation not available for GLAC1D or ICE-6G. Normalization is done at 1 ka when all model runs intersect within 0.05K and the ice sheet configuration is well known. The ICE-6G values at 11 ka for summer and winter (not shown on plots) are -3.93 K and -10.82 K, respectively. Coefficient of determinations for model results vs. WDC temperatures (ED Figure 7) are high for summer (HadCM3 ORBIT R^2 = 0.93, $p << 0.001$; MEBM R^2= 0.80, $p << 0.001$), but not for winter (HadCM3 ORBIT R^2= 0.00, p = 0.85; MEBM R^2= 0.05, p = 0.30). The winter agreement improves if only the period 0 – 6 ka is considered (HadCM3 ORBIT R^2= 0.74, p = 0.01; MEBM R^2= 0.39, p = 0.02).* ***b,d,*** *Histograms of net temperature changes over the specified time intervals, derived by Monte Carlo analysis accounting for systematic and non-systematic uncertainties (Methods Section 3.5).* ***e,*** *WDC mean-annual temperature with 1σ and 2σ uncertainty bounds[30]. ED Table 2 shows the amount of variability in the mean-annual temperature that can be explained by the summer and winter temperatures.*

Figure 3 | Temporal and Spatial Variability in Insolation and Model Temperatures. ***a,*** *Insolation change through the Holocene[35] for December and January, and their average. December best resembles the WDC summer reconstruction.* ***b,*** *The full seasonal cycle of insolation at 80°S for 500-yr snapshots over the Holocene. Line colors in (b) and (c) correspond to age.* ***c,*** *Zoom of summer insolation. The maximum always occurs in the latter half of December (grey shading), migrating across 8 days over the course of the Holocene.* ***d,*** *Holocene trends of annual mean insolation (black), annual integrated insolation (dashed red line), and summer-integrated insolation (red line).* ***e,*** *Maximum summer insolation intensity (black line) and summer duration (red lines), defined as the number of days above a threshold insolation value each year.* ***f,*** *Anomaly in maximum insolation colored by latitude in the Southern Hemisphere. The thick blue line shows the latitude of the WDC site.* ***g,*** *Calculated temperatures for 80°S using the MEBM, including maximum summer value (red), minimum winter value (blue), and amplitude of the seasonal temperature cycle (black).*

Figure 4 | Possible ice elevation histories and the corresponding modeled temperatures. ***a,*** *Elevation histories used in HadCM3. GLAC1D[37] is 96 meters higher at 11ka compared to present, and ICE-6G[38] is 222 meters higher.* ***b,*** *Temperature anomalies from elevation change (GLAC1D solid lines; ICE-6G dashed lines) using an atmospheric* *lapse rate of 9.8°C/km, and spatial lapse rates for interior West Antarctica of 12°C/km[39] and 14°C/km[40].* **c,** *HadCM3 residual-temperature anomalies for December (summer) calculated by subtracting the ORBIT run from the GLAC1D and ICE-6G runs in Figure 2a, highlighting the portion attributable to changing elevation rather than insolation.* **d,** *Residual temperature change in (b) subtracted from the results in (c), showing the component driven from processes besides the direct lapse-rate effect and orbital forcing.*

Methods

We measured WAIS Divide core (WDC) water isotopes using continuous flow analysis (Section 1), and then corrected for cumulative diffusion using spectral techniques to determine diffusion lengths and restore pre-diffused amplitudes within 140-yr sliding windows (Section 2, ED Figure 1c). Summer maxima and winter minima (Figure 1a) identified in these corrected data were then used to calculate summer and winter amplitudes for each year. We converted the isotope-amplitudes to temperature-amplitudes using a model-determined scaling factor (Section 2.2), and added them to previously reconstructed mean annual temperatures[30] to recover summer and winter values. Significant seasonal noise processes required multi-centennial to millennial averaging to reduce uncertainty (Section 3). To elucidate physical controls on sub-annual temperatures, we used a simple energy balance model and HadCM3, a General Circulation Model, to calculate expected changes in seasonal and monthly surface temperatures through time under varying boundary conditions (Sections 4 and 5). Finally, using both observations and modeling, we estimated the change in WAIS surface elevation through the Holocene (Section 6).

1 Water Isotopes

WDC water isotopes (ED Figure 1a) were analyzed on a continuous flow analysis system[21] using a Picarro Inc. cavity ring-down spectroscopy (CRDS) instrument, model L2130-i. Using permutation entropy[47], we identified data anomalies arising from laboratory analysis, which were corrected, including by resampling ice through 1035.4-1368.2 m depths (4517-6451 yrs)[48]. All other Holocene data are previously published[18,31], and available online[19,20]. Data are reported in 5-mm increments in delta-notation (per mil, ‰) relative to Vienna Standard Mean Ocean Water (VSMOW, $\delta^{18}O = \delta D = 0$‰), normalized to Standard Light Antarctic Precipitation (SLAP, $\delta^{18}O = -55.5$‰, $\delta D = -428.0$‰). WDC is annually dated, with accuracy better than 0.5% of the age between 0-12 ka[22]. For the Holocene, the temporal spacing of consecutive 5 mm samples is < 0.1 years, and the average < 0.05 years, ranging from ~2.6 weeks at 10 ka to one-half week from 1-0 ka[18].

2 Diffusion Corrections

Diffusion in the firn and deeper ice attenuates high-frequency water-isotope information in ice cores[26,27,49,50,51,52]. Diffusion length quantifies the statistical vertical displacement of water molecules from their original position[27,49]. We used diffusion-correction code developed by S. Johnsen, University of Copenhagen[23,24,27,28], which uses Maximum Entropy Methods (MEM) to invert an observed power density spectrum. As an input to these inversions, we determined diffusion lengths (ED Figure 1c) for 140-year windows using prior methods[18,26]. The power density spectrum observed in the ice core record $P(f)$, after diffusion, is $P(f) = P_o(f)exp[-(2\pi f\sigma_z)^2]$, where $P_o(f)$ represents the power spectrum of the undiffused signal (per mil^2·m), f is the frequency $\frac{1}{\lambda}$ (1/m), λ the signal wavelength (m), z the depth (m), and σ_z the diffusion length (m). The original, pre-diffusion power-density spectrum (i.e. diffusion-corrected) is calculated as $P_o(f) = P(f)\,exp(4\pi^2 f^2 \sigma_a^2)$, for diffusion length σ_a (yr). The $\sigma_a = \frac{\sigma_z}{\lambda_{avg}}$, where λ_{avg} is the mean annual layer thickness (m/yr) at a given depth. The diffusion-corrected spectrum takes the form of a series of complex numbers $X_R + iX_I$ vs. f. From this, the amplitude spectrum A is obtained by $A(f) =$

$\sqrt{X_R^2 + X_I^2}$ and the phase spectrum ϕ is obtained by $\phi(f) = tan^{-1}\left(\frac{X_I}{X_R}\right)$. The real components of the amplitude and phase spectrums give the diffusion-corrected water isotope signal $\delta_o(t)$ as:

$$\delta_o(t) = \sum_{i=1}^{N} A_i cos(2\pi f_i t + \phi_i)$$

Uncertainties on $\delta_o(t)$ are determined using the uncertainty range for diffusion lengths[26] calculated in each 140-year window. Before spectral analysis, the isotope data are linearly interpolated at a uniform time interval of 0.05 yr. Our determination of diffusive attenuation and correction arises from the observed frequency spectra themselves and therefore is entirely independent of firn diffusion and densification models.

2.1 Seasonal Water-Isotope Amplitudes

To select extrema (summers and winters) in the diffusion-corrected δD signal (Figure 1a, ED Figure 1b), we used the 'findpeaks' matlab function. Figure 1c,d show the resulting time series for summer and winter, averaged with a 50-year boxcar filter for clarity of trends. For every year defined in the WDC age-scale, we calculated the averaged diffusion-corrected δD. The difference between the two extrema and this mean define the summer and winter isotope-amplitudes.

2.2 Seasonal Temperatures

A linear scaling converted seasonal isotopic-amplitudes to seasonal temperature-amplitudes, using a sensitivity of isotopes to surface temperatures determined by the "Simple Water Isotope Model" (SWIM)[29]. Finally, to find summer and winter temperatures we added the individual seasonal temperature-amplitudes to the year's mean temperature obtained previously[30] by calibrating the water-isotope record against borehole temperatures and $\delta^{15}N$ constraints on firn thickness.

SWIM is based on earlier numerical Rayleigh-type distillation models[53,54], which simulate the transport and distillation of moisture down climatological temperature gradients. As moist air is transported towards the poles and cools, the saturated vapor pressure decreases nonlinearly, and moisture above saturation is removed by precipitation. The model keeps track of the isotopic fractionations at each step along this distillation process. In most previous simple models, there is an inconsistency in the calculation of the supersaturation that determines the point of condensation and that drives kinetic isotope fractionation. Modifications to these earlier models, employed in SWIM, ensure consistency in the calculation, which results in a smoother relationship between temperature and the δ-values of precipitation, and better agreement with observed spatial patterns of δD and $\delta^{18}O$. Given input of both δD and $\delta^{18}O$ data, SWIM calculates distributions of source temperatures, the temperature gradients of pseudo-adiabatic pathways, and condensation temperature. We used SWIM to derive sensitivities for surface isotope-temperature scalings using diffusion-corrected WDC data to obtain a surface scaling of 6.96‰ δD/°C. Using raw data, the surface scaling is 7.07‰ δD/°C. In comparison to other isotope-temperature scalings, Buizert et al.[55] obtain ~6.56‰ δD/°C and Cuffey et al.[30] ~7.10‰ δD/°C (both converted from $\delta^{18}O$ to δD using a factor of 8).

3 Uncertainties in Reconstructing Temperatures

We included uncertainties associated with the following factors: measurement analysis, diffusion correction, seasonality of accumulation, precipitation intermittency, modeled isotope-temperature scaling, and mean-temperature history. The 'analysis uncertainty' is 0.55‰ for δD (1σ)[21]. The 'diffusion-correction uncertainty' is described by Jones et al.[26]. The uncertainty of the mean temperature reconstruction, calculated previously[30], accounts for the majority of uncertainty in the early-Holocene, but a small fraction in the late-Holocene. Sections 3.1-3.3 explain the other uncertainty terms. Uncertainties for some factors (analysis, diffusion correction) can be treated as independent random variables so that, upon time-averaging, their magnitudes decrease as the inverse of the square root of the number of values. Uncertainties for other factors (intermittency, isotope-temperature scaling, mean temperature, seasonality) might be systematically biased, and therefore their magnitudes are taken to be invariant with respect to the interval of averaging. Based upon the 2σ uncertainties for summer and winter temperature (Figure 2a,c), we assessed the significance of dominant trends using Monte Carlo analysis (Figure 2b,d; Section 3.5).

3.1 Seasonal Preservation Bias Uncertainty

Unequal seasonal distribution of snowfall could result in different magnitudes of diffusion for winter and summer amplitudes[49]. The seasonal temperature cycle also impacts the magnitude of diffusion for all seasons. We utilized the Community Firn Model (CFM)[56,57], a firn-evolution model with coupled firn temperature, firn densification, and water-isotope modules, to test how seasonally weighted accumulation affects the diffusion of specified, hypothetical isotope records progressing from surface snow (δD_{snow}), to consolidated snowpack in the firn (δD_{firn}), to solid ice beneath the pore close-off depth (δD_{ice}). We applied the back-diffusion calculation (Section 2) to δD_{ice} to estimate the original δD_{snow}. We then assessed how reconstructions of δD_{snow} could be misinterpreted due to different seasonal-accumulation weightings (ED Figure 2a,b).

We performed five CFM runs using monthly time-steps for accumulation, temperature, and isotopes (ED Table 1). The seasonal cycle for δD_{snow} is based on the mean amplitude in Figure 1b (15.43‰). Five WAIS accumulation scenarios were tested based on monthly accumulation from the regional climate model MAR3.6 (Modèle Atmosphèrique Règional; ERA-Interim forced)[58], which spans the period January 1979 to December 2017. The mean accumulation over the entire 39-year period is 0.24745 m ice eq. a^{-1}, with about 1.6x as much snow in winter (Oct-Mar) as summer (Apr-Sep). The five scenarios are as follows. 1) 'constant': identical accumulation for all months (0.0206 m ice eq. $month^{-1}$; 1/12 the annual mean). 2**)** 'cycle': monthly accumulation equal to MAR monthly means. 3) 'noise': using the 'cycle' time series, we add noise to each time step in the 'cycle' series, we added a normal random variable of zero mean and the standard deviation for the month from MAR, 4) 'random': for each month, the accumulation is a normal random variable with mean and standard deviation equal to MAR monthly values. 5) 'loop': the entire 39-year MAR accumulation time series is repeated over and over again. For the temperature boundary condition, we used each month's mean 1979-2017 2m temperature predicted by MAR to create an annual temperature cycle. We repeated this 12-month time series for the duration of the model runs. This method ensures that model runs, which are designed to test accumulation seasonality, are not affected by interannual temperature variability, while also providing an estimate of the annual-temperature cycle, which affects the rate of isotope diffusion in the upper firn.

ED Figure 2a,b shows the results for the 'constant' and 'cycle' cases. The diffusion-correction technique accurately reconstructs δD_{snow} for summer and winter in the 'constant' snowfall scenario, but underestimates summer values in the MAR 'cycle' scenario by ~2.6‰, which is 11% of the full range of the observed WDC summer water-isotope values. Winter values are overestimated by only ~0.6‰, about 3% of the full winter range, since winter has 1.6x as much snow as summer. The 'noise', 'random', and 'loop' runs produce results within 0.3‰. These CFM experiments demonstrate that centennial trends in the summer and winter water isotopes on the order of a few per mil could arise from large changes in seasonal-accumulation weighting, while multi-millennial trends >>2.6‰ are unlikely to be caused by seasonal accumulation and can therefore be interpreted as climate signals of a different origin. For 1,000-year averaging (as in Figure 2), HadCM3 indicates seasonal accumulation weighting (winter:summer) of 1.3 to 1.7 throughout the Holocene (ED Figure 2f), which yields a 1σ uncertainty of 0.27‰ based on the CFM testing criteria.

To determine observationally if seasonal snowfall changed significantly across the Holocene, we used measured black carbon (BC) concentrations, the only age-scale-independent impurity. BC data are available from 0-2.5 ka and 6-11 ka. Seasonal-fire regimes in South America dominate BC concentrations at WDC, causing BC maxima and minima in fall and spring, respectively[59] (ED Figure 2c). We split each year into two parts, characterized by rising or falling BC. Call BC_1 and BC_2 the depth intervals of rising and falling BC. The duration of BC_2 is longer than BC_1 due to source characteristics[57] (ED Figure 2c), thus $BC_1/BC_2 < 1$ (ED Figure 2e). The BC_1/BC_2 ratio can change with time due to variability at the source, changes in atmospheric transport, or seasonality of snow deposition. We observe little change in BC_1/BC_2 resembling the multi-millennial trends seen in WDC summers and winters (ED Figure 2d,e). Unless there are competing and exactly compensating effects in seasonality (i.e. the source change exactly cancels the depositional and transport change, or other unlikely scenarios), the BC data provide evidence that changes in WDC seasonal snowfall were not large enough to affect our multi-millennial climate interpretations.

3.2 Intermittency of Precipitation Uncertainty

The episodic nature of snowfall creates an incomplete record of local climate variations[61], preventing interpretation of trends over short time intervals. We want to interpret isotopic variations averaged over a sufficiently long timescale that, to within a specified tolerance, trends are not likely to be random noise arising from the spread of distributions preserved in the ice. Using distributions of reconstructed annual amplitudes (Figure 1b, ED Figure 2g) for 1,000-year windows throughout the Holocene, we conducted Monte Carlo resampling simulations to determine that 250-year averaging-lengths are needed to achieve a standard error of 1‰, corresponding to a mean amplitude-to-noise ratio of 15. For the time period with greatest variability, centered on 4 ka, the standard error for a 1,000-

year average (as used in Figure 2) is 0.52‰ (ED Figure 2h). Since this is an amplitude uncertainty (rather than uncertainty associated with a season), we specify the 1σ uncertainties for summer and winter as half of 0.52‰.

3.3 Isotope-Temperature Scaling and Associated Uncertainty

The conversion of isotopic values (1,000-year averages) to temperature yields three curves for summer and three for winter: $T_{nominal}$, $T_{+1sigma}$, $T_{-1sigma}$. Each curve is normalized to the value at 1 ka (as done in Figure 2). The difference in the $T_{+1sigma}$ and $T_{-1sigma}$ curves gives the 1σ uncertainty range $+\sigma_{Tscale}$ to $-\sigma_{Tscale}$, which are then added in quadrature to the Cuffey et al.[30] mean-temperature uncertainties, yielding the final uncertainty estimates shown in Figure 2.

3.4 Relationship Between the Annual Mean and Individual Seasons

Using 1,000-year and 300-year averages of summer, winter, and mean temperature (ED Figure 3c-f), we determined R^2 values for summer and winter vs. the mean. We then subtracted the 1,000-year averages from the 300-year averages to obtain residuals, and then determined R^2 values again for summer and winter vs. the mean (ED Table 2). From 11-0 ka, the high summer correlations for the 1,000-year comparison indicate a strong association of the annual-mean temperature with the summer temperature at orbital timescales. At suborbital scales (i.e. the 300 to 1,000-year residuals), neither the summer nor the winter alone explains much of the mean-annual variability, and the annual mean is a random composite of the two seasons. If only 11-7 ka is considered, winter variability explains more of the mean at sub-millennial scales.

3.5 Trend Analysis

To assess the significance of dominant trends in our reconstructed seasonal temperatures, we conducted a Monte Carlo analysis founded on the assumption that all possibilities for the unknown time-dependence of errors are equally likely. The essential motivation for this approach is that we have determined the magnitudes of uncertainties as a function of age but that we have no information about whether the errors in our reconstruction persist at similar values for long periods of time (i.e., exhibit a bias) or whether they fluctuate at high-frequency.

We randomly generated a large number of alternative seasonal temperature histories governed by the uncertainties on 1000-year averages (ED Figure 3a,b), calculated the temperature trends for each alternate history over desired time intervals (such as 11-4 ka and 4-0 ka), and compiled the results into frequency distributions from which probabilities can be calculated (Figure 2b,d). Specifically, each alternative history deviates from the summer and winter temperature reconstruction by an amount that smoothly varies over time between random nodes whose values are a Gaussian random variable of zero mean and standard deviation for 1000-year averages at the age of the node. The number of nodes and the age of each node are random variables, uniformly distributed between 1 and 11 nodes and 0-11 ka, respectively. A small number of nodes produce an alternative temperature history for which the bias is serially correlated for millennia, while a large number of nodes produce a history for which the bias is uncorrelated from millennium to millennium.

4 Seasonal Moist Energy Balance (MEBM) Model

We used a simple global, zonal-mean moist energy balance model to calculate surface temperatures (ED Figure 4), accounting for top-of-atmosphere (TOA) insolation, temperature-dependent long-wave emission to space, temperature-dependent albedo to simulate brightening by snow and ice, and horizontal atmospheric-heat transport treated as diffusion of near-surface moist static energy[62,63,64]. The model has a 2-degree spatial resolution, a single surface and single atmospheric layer, and a surface-heat capacity based on the relative fraction of land and ocean surface in the zonal mean. Heat exchange between the surface and atmosphere layers arises from differences in blackbody radiation from each layer, and sensible and latent heat exchanges proportional to temperature and specific humidity contrasts (assuming a constant relative humidity of 80%), following bulk aerodynamic formulae. We calculated the annual TOA insolation-cycle at every latitude in 500-year time slices from 0-11 ka. For each time slice the model is run at 2-hour time resolution for 30 model-years to reach equilibrium.

ED Figure 4 compares the temporal evolution of annual mean and peak summer heat divergence by the atmosphere ($\nabla \mathbf{F}$) at the WDC site to the annual mean and summer maximum site temperature and insolation. While Holocene changes in $\nabla \mathbf{F}$ at the WDC site correlate with insolation forcing, the magnitude of changes in maximum direct insolation are much larger than those in atmospheric heat divergence. Further the Holocene changes in summertime heat divergence are of the wrong sign to cause net heating at the WDC site (positive divergence is an export of heat by the atmosphere from the site). Heat transport in the Antarctic is convergent in the annual mean, but divergent in mid-summer, as intense incoming insolation exceeds longwave emissions from the cold surface.

5 HadCM3 model simulations

5.1 Model setup

We used the fully coupled ocean-atmosphere model HadCM3[65,66], version HadCM3BM2.1, which well simulates tropical Pacific climate and its response to glacial forcing[67]. Our simulations are snapshots at 1-kyr intervals over the last 11 ka[36], with time-specific boundary conditions of orbital forcing[68], greenhouse gas (GHG) concentration[69,70], ice sheet topography, and sea level[37,38,71.72,73,74,75]. We used three simulations: 1) only orbital forcing changes (ORBIT), with all other boundary conditions set to the pre-industrial; 2) orbital/GHG forcing with GLAC1D ice-sheet elevation history; and 3) orbital/GHG forcing with ICE-6G ice-sheet elevation history. Elevation histories are shown in Figure 4a. Snapshot simulations were run for at least 500 years with analysis made on the final 100 years. Additional snapshot simulations for 10 ka allowed us to decompose the role of different forcings, described in the following sections. The large difference in forcings between 10 ka and the pre-industrial late Holocene make this comparison most instructive.

5.2 Summer Climate

We examined the zonal mean at 80°S in simulations for hypothetical 10 ka worlds, by changing the boundary conditions to compare to the pre-industrial/late-Holocene. These simulations are '10ka ORBIT-only'; two runs with only ice sheets at 10 ka and pre-industrial settings otherwise, called '10 ka GLAC1D-only' and '10ka ICE-6G-only'; and two runs with all 10 ka forcings, called '10ka GLAC1D-all', and '10ka ICE- 6G-all'. In '10ka ORBIT-only', reduced TOA shortwave radiation causes a large reduction in shortwave radiation at the surface (SW_d), and consequent cooling. Downward longwave radiation (LW_d) also decreases, likely due to atmospheric cooling. Sensible heat flux (SH_d) to the surface is increased indicating that the atmosphere and surface do not equally cool; one cause of this is the increased meridional heat convergence ($-\nabla\mathbf{F}$).

Changed ice sheets cause summertime cooling in both '10 ka GLAC1D-only' and '10ka ICE-6G-only', primarily via reduced LW_d. Increased SW_d, due to reduction of depth of the atmospheric column above the ice sheet surface, counteracts the reduced LW_d to some extent. (Reducing the atmospheric column reduces SW absorption and tends to cool the atmosphere, reducing LW_d). Both ice sheet scenarios also cause an increase in $-\nabla\mathbf{F}$, partly counteracting summertime cooling.

Using all 10 ka forcings causes cooling through both LW_d and SW_d. The decrease in SW_d is similar in '10ka GLAC1D-all' and '10ka ICE-6G-all' and slightly smaller than in '10ka ORBIT-only', likely because the thinner atmospheric column reduces absorption. The decrease in LW_d in '10ka GLAC1D-all' and '10ka ICE-6G- all' is larger than in '10ka ORBIT-only', '10 ka GLAC1D-only', and '10ka ICE-6G-only'. This indicates the importance of feedbacks in the atmosphere. Heat convergence $-\nabla\mathbf{F}$ increases in both simulations indicating remote feedbacks, in addition to local feedbacks related to the amount of water vapor in the atmosphere.

The preceding description of changes in the zonal mean in the 10 ka simulations compared to pre-industrial holds for the entire Holocene. Orbital forcing alone reduces SW_d and LW_d by roughly the same magnitude. With full forcing (including ice sheets), the reduction in LW_d is roughly three times the reduction in SW_d. Considering an energy budget over the WDC site (79.467°S 112.085°W), mechanisms are the same as for the zonal mean. Magnitudes of forcings change, but reduced SW_d still cools the surface, amplified by a LW_d feedback dependent on ice sheet size.

5.3 Winter Climate

During winter SW_d is no longer a factor as the sun is below the horizon, yet there is still surface warming caused by an increase in LW_d. An increase in $-\nabla\mathbf{F}$ in '10ka ORBIT-only' warms the atmospheric temperature, increasing LW_d and SH_d. With an ice sheet imposed the surface temperature cools. In both '10 ka GLAC1D-only' and '10 ka ICE-6G-only' there is a reduction in $-\nabla\mathbf{F}$, reducing LW_d. and SH_d. When all 10 ka forcings are introduced, the change in temperature is smaller than for ice sheet-only runs. In '10ka GLAC1D-all' we found no change in $-\nabla\mathbf{F}$, LW_d, or surface temperature. This suggests that the increase in $-\nabla\mathbf{F}$ from orbital forcing is almost perfectly balanced by the change in $-\nabla\mathbf{F}$ from the ice sheet configuration.

The processes controlling heat transport over Antarctica are complicated and HadCM3 may not be able to simulate them perfectly. Our simulations indicated that remote processes during winter alter the heat transport, affecting atmospheric and surface temperatures. Raising the topography of Antarctica tends to reduce such heat transport (ED Figure 5), producing an additional cooling on top of a pure lapse rate effect[76]. This cannot, however, explain the prominent millennial-scale changes at ~9.2 and ~7.9 ka (Figure 1e,2c). The intricacies of interpreting the early Holocene winter variability in West Antarctica necessitates further study.

5.4 Sea Ice

Sea ice changes may alter local energy fluxes from the ocean to the atmosphere. In HadCM3, sea-ice extent changes across the Holocene. We used two analyses (ED Figure 6) to show sea ice is not a primary control on the surface temperature at WDC (80°S): 1) Correlation analysis of sea-ice changes and surface temperature, and 2) Atmosphere-only model simulations in which we specified individual changes in the model boundary conditions (including sea ice).

We computed the dominant spatial patterns of sea-ice variability using EOFs across all of the HadCM3 simulations (ALL), and individually for three sub-set simulations (ORBIT, GLAC1D, ICE-6G), from 0-11 ka. We projected the model-simulated sea ice for each individual time-slice simulation onto these patterns to compute the amplitude of sea-ice variability in each simulation. The amplitude was compared to temperature at 80°S to understand how large-scale changes in the sea ice affect temperature for the months of December (summer) and July (winter) (ED Figure 6).

In winter, we found negligible correlations between sea-ice change and temperature in all sets of simulations (ALL: 0.02, ORBIT: 0.04; GLAC1D: 0.16; ICE-6G: 0.06). This suggests that winter sea ice is not an important factor in determining the temperature at 80°S. In summer, only the ORBIT simulation has meaningful correlations between temperature and sea ice variability (ALL: 0.59, ORBIT: 0.84, GLAC1D: -0.60; ICE-6G: 0.36). The sign of the correlation changes between simulations despite the sea-ice change pattern being the same in all simulations. From this we concluded that sea ice is not a dominant control on temperature at 80°S in summer. The correlation in the ORBIT simulations suggests that there may be some relationship between sea ice and temperature; we investigated this with atmosphere-only simulations.

In atmosphere-only simulations at 0 ka and 10 ka, we specified the top of the atmosphere insolation, SST, and sea ice from the ORBIT simulations. Over land areas and sea-ice regions, the model calculates the surface temperature using the land-surface scheme in the model. The atmosphere model is identical to the model used within the coupled-model. We ran a series of experiments varying the orbital configuration, SST, or sea ice (summarized in ED Table 3).

The zonal mean of the change in the sea ice that we prescribed can be seen in ED Figure 6e, and the change in the SST can be inferred from ED Figure 6f-h. ED Figure 6f shows that the atmosphere-only model replicates the change in temperature of the coupled-model. ED Figure 6g shows that the effect of the 10 ka orbital configuration ('Atmos_10k_insol') is to cool Antarctica considerably by about 0.5°C. North of 65°S there is no change in the surface temperature, primarily a response to the imposed SST and sea ice, which are the same in the 'Control' and 'Atmos_10k_insol'. Imposing the SST and sea ice from 10 ka ('Atmos_10k_ice_SST'), we find very little change in the surface temperature over Antarctica, but there are some large changes in the surface temperature north of 70°S. ED Figure 6h shows the result of imposing 10 ka SST or 10 ka sea ice. The 10 ka SST ('Atmos_10k_SST') tends to warm Antarctica, consistent with the large increases in SST north of 65°S. Changing sea ice ('Atmos_10k_ice') tends to cool Antarctica. Both effects are small, approximately 0.1°C, and of opposite sign. This explains the small net change in the surface-temperature change over Antarctica when SST and sea ice are changed simultaneously, as shown by 'Atmos_10k_SST_ice'. It should be noted that in the coupled system a change in sea ice can not be decoupled from a change in the SST, so not only is the effect of sea ice on the climate small, it is also likely associated with a compensating change in SST. From these simulations we concluded that sea ice and SST changes are not a dominant driver of the change in the surface temperature over Antarctica.

ED Figure 6e shows that the change in sea ice at 10 ka in ORBIT is much larger than the change in either GLAC1D or ICE-6G. The ORBIT simulations do not account for all of the changes in the boundary conditions at 10 ka and are therefore less realistic than either ICE-6G or GLAC1D. Since ICE-6G and GLAC1D both show much smaller changes in the sea ice and SST than ORBIT, we expect that in reality there is also a much smaller change in the sea ice and SST than in ORBIT. We thus concluded that sea ice has a small impact on the temperature at 80°S in summer.

We also performed a similar analysis of the winter season (not shown). We found the atmosphere-only model does not compare well with the coupled model, simulating very little change in the surface temperature. Doing a term-by-term decomposition of the atmosphere model is not, therefore, particularly useful as it tells us more about the model rather than the physical climate. The failure of the atmosphere-only model to capture the changes at 10 ka suggests that the importance of the SST and sea ice is in their day-to-day coupling with the atmosphere and not in any long-term mean change in this season.

6 Estimating elevation changes

Temperatures simulated by HadCM3 for ORBIT provide a control scenario against which observations can be compared to identify the signal of elevation change. For a chosen time interval, the net reconstructed warming ΔT_R exceeds that of ORBIT by an amount $\Delta T_R - \Delta T_O$. This can be compared to the effective lapse-rate $(\Delta T_M - \Delta T_O) / \Delta Z_M$ defined by a HadCM3 simulation including topographic change (GLAC1D or ICE6G models) and all forcings, for model warming ΔT_M and model elevation decrease ΔZ_M. Specifically, the estimated elevation decrease is $\Delta Z_R = \Delta Z_M [(\Delta T_R - \Delta T_O) / (\Delta T_M - \Delta T_O)]$. Summer and winter reconstructions offer two separate assessments, for which we calculate the algebraic average.

Accounting for uncertainties in ΔT_R requires recognizing that uncertainties of summer and winter reconstructions are not independent, while also recognizing that they emerge from two independent sources: uncertainty in the mean annual temperature history (calculated in Cuffey et al.[30]) and uncertainty in the seasonal amplitude (calculated in the present study). In general, the uncertainty of seasonal temperature at a specified time is the quadrature sum of annual and amplitude uncertainties, which gives 1σ and 2σ uncertainties for a given season and time. However, if the true value of annual temperature is shifted by an amount $\alpha\sigma$ from the nominal reconstruction, this must be true for both summer and winter. And if the true value of amplitude is shifted by an amount $\beta\sigma$ from the nominal reconstruction, the temperature shift must be $+\beta\sigma$ in one season but $-\beta\sigma$ in the other.

To define bounding cases on elevation change in a specified time interval, we calculated the maximal (or minimal) temperature change ΔT_R for a season by differencing the upper (or lower) limit at one end of the interval with the lower (or upper) limit at the other end, and also calculating the corresponding ΔT_R for the opposite season required by the correlated errors. The elevation decreases ΔZ_R were then calculated by comparison to HadCM3 simulations, as specified previously, and the summer and winter values averaged. This process was completed four times for each time interval and HadCM3 model, corresponding to four different initial ΔT_R (maximum and minimum ΔT_R for summer, and maximum and minimum ΔT_R for winter), and the most extreme case taken as the result (this proved to be the one starting with maximum summer ΔT_R). Table 1 lists results for two time intervals and the two HadCM3 simulations with variable topography.

Data Availability

The WDC water isotope datasets analyzed during the current study are available in the U.S. Antarctic Program Data Center (USAP-DC) repository, https://doi.org/10.15784/601274 and https://doi.org/10.15784/601326. The impurity datasets analyzed during the current study are available in USAP-DC repository, https://doi.org/10.15784/601008.

The data generated in this study is available in the USAP-DC repository, https://doi.org/10.15784/601603, including raw and diffusion-corrected water isotopes, seasonal water isotopes (maximum summer and minimum winter values), and seasonal temperature reconstructions. Additional data are available in the online source files linked to this publication.

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Code Availability

The MATLAB code used for the diffusion correction of water isotope data and the subsequent selection of seasonal extrema (summer and winter) is available online at Zenodo, https://doi.org/10.5281/zenodo.7042035.

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Author Contributions

T.R.J. designed the project. T.R.J. K.M.C., and W.H.G.R. led the writing of the paper, with help from B.R.M. and E.J.S. High-resolution water isotope measurements were contributed by T.R.J. and J.W.C.W., the analysis of the water isotopes was led by V.M., B.H.V., T.R.J., and J.W.C.W. HadCM3 simulations were conducted by W.H.G.R. and P.J.V., while W.H.G.R. developed the methodology for determining how modeled sea ice affects West Antarctic temperature. B.R.M. conducted MEBM simulations. C.M.S. and T.R.J. conducted CFM simulations. T.R.J., K.M.C., and B.M.V. developed the diffusion-correction calculations, with help from A.G.H., and C.A.B. T.R.J. developed the methodology for the selection of extrema (summer and winter) in the diffusion-corrected water isotope data. T.R.J. developed the methodology for quantifying the effect of seasonal accumulation on water-isotope diffusion using the CFM and chemistry data. T.R.J. and K.M.C. determined the methodology for the uncertainty of seasonal temperature reconstruction. K.M.C. and E.J.S determined the methodology for the uncertainty of multi-millennial temperature trends, with help from T.R.J. B.R.M. provided isotope-temperature scaling uncertainty using SWIM. K.M.C. determined the methodology for constraints on WAIS Divide elevation changes in the Holocene. T.J.F. and M.S. provided chemistry data. T.R.J., V.M., B.H.V., J.G., and K.S.R. assisted with the development of the water isotope dataset over depths of 1035.4 to 1368.2 m. All authors discussed the results and contributed input to the manuscript.

Author Information

Reprints and permissions information is available at www.nature.com/reprints. The authors declare no competing financial interests. Readers are welcome to comment on the online version of the paper. Correspondence and requests for materials should be addressed to T.R.J. (tyler.jones@colorado.edu).

ED Table 1 | CFM simulation inputs. *Mean* $\underline{\dot{b}}$ *and standard deviation* $\sigma_{\dot{b}}$ *of each month's accumulation rate and monthly mean temperature at WAIS Divide for 1979 to 2017 predicted by MAR*[56] *and the isotope values used for each month during the CFM simulations.*

ED Table 2 | Seasonal vs. mean correlations. R^2 *values for 1,000-year and 300-year averages of summer and winter vs. mean, as well as for residuals.*

ED Table 3 | Atmosphere-only climate model experiments. *Variations of boundary conditions in each experiment.*

ED Figure 1 | WDC water-isotope data. ***a,*** *The raw, high-resolution WDC δD water isotope record*[18,19,20] *(grey), the raw 50-yr running mean (white), and the diffusion-corrected signal (black).* ***b,*** *The WDC diffusion-corrected δD record with extrema picks for summer (red) and winter (blue).* ***c,*** *The high-resolution diffusion length record (black; 140-yr windows, 70-yr time steps; 1σ uncertainty bounds in light grey) compared to prior estimates*[26] *(red; 500-yr windows, 500-yr time steps).*

ED Figure 2 | Precipitation uncertainties.. *Uncertainties for seasonally weighted accumulation are shown in a-f, and for precipitation intermittency in g-h.* ***a,*** *The diffusion envelope of CFM output data (50-yr avg.), based on an input sine wave with* $f=1yr^{-1}$ *and amplitude=15.43‰. The original amplitude of the signal (based on the dotted dashed lines) decreases as time passes due to downward advection of the firn, as shown by the decay of the maximum (red) and minimum (blue) lines, while the mean values of the 'constant' and 'cycle' scenarios do not change, dependent on the seasonal weighting of snowfall.* ***b,*** *Diffused CFM output data from beneath the pore close-off depth (>200 yr) (black lines; smaller amplitude), with diffusion-corrected data shown with grey lines (larger amplitude). Red circles are the annual maximum value, and blue circles the annual minimum, selected using the same algorithm as Figure 1a.* ***c,*** *A zoom of black carbon concentrations at ~6.5 ka*[57]*. The maxima (red circles) and minima (blue circles) can be used to separate approximate depth intervals corresponding to winter (*BC_1*) and summer (*BC_2*); vertical blue lines correspond to nominal January 1, as defined by the peak of nssS/Na*[22]*.* ***d,*** *The 140-yr averages for* BC_1 *(blue) and* BC_2 *(red). The grey line is WDC annual accumulation*[60]*, orange circles are* BC_1 + BC_2*, which should equal annual accumulation.* ***e,*** *Black carbon seasonality* BC_1/BC_2 *(black), based on (d).* ***f,*** *Accumulation seasonality for HadCM3 seasonal snowfall (red line) compared to the range of seasonality tested using the CFM (dashed blue lines) and modern MAR seasonality*[58] *(blue diamond).* **g,** *Distribution of annual amplitudes for water isotopes for a 1,000-year window centered at 4 ka.* **h,** *Standard errors are determined for 1,000-realizations of random sampling of the distribution in (a) to determine a standard deviation of the residuals of the true mean minus the mean of the random sampling.*

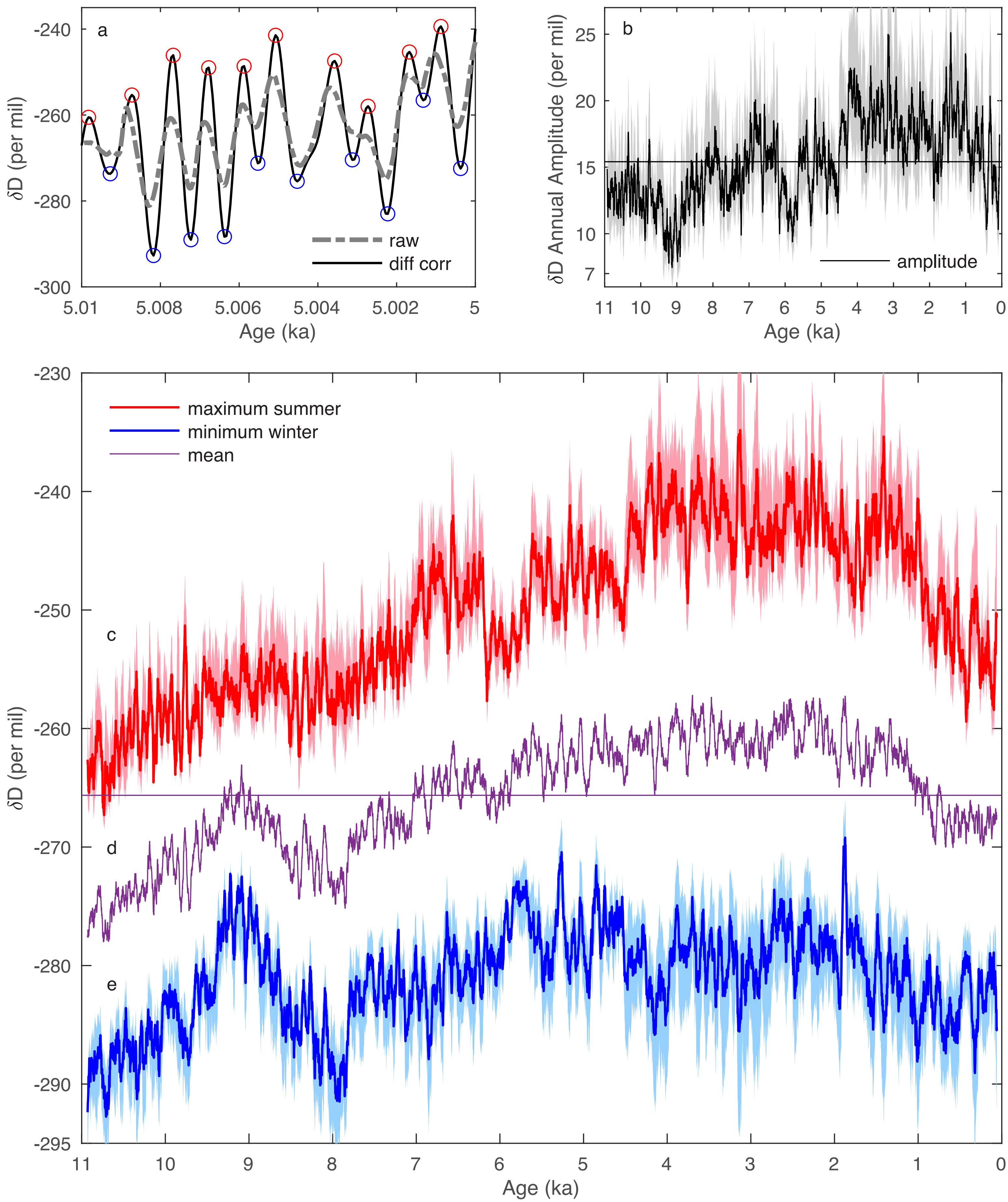
ED Figure 3 | Trend analysis. ***a,b,*** *The first 10 of 10,000 randomly generated, alternative seasonal temperature histories for summer (a) and winter (b), used in Figure 2b,d to generate probability distributions of temperature trends in the Holocene. Thick, solid lines are mean values, and thick, dashed lines are 2σ uncertainty ranges.* ***c-f,*** *The 1,000-year (thin line) and 300-year (thick line) averages normalized to the 11-0 ka mean (c-e) and residuals (1,000 year minus 300 year) (f) of summer (red), winter (blue), and the mean (black), used to calculate* R^2 *values in ED Table 2.*

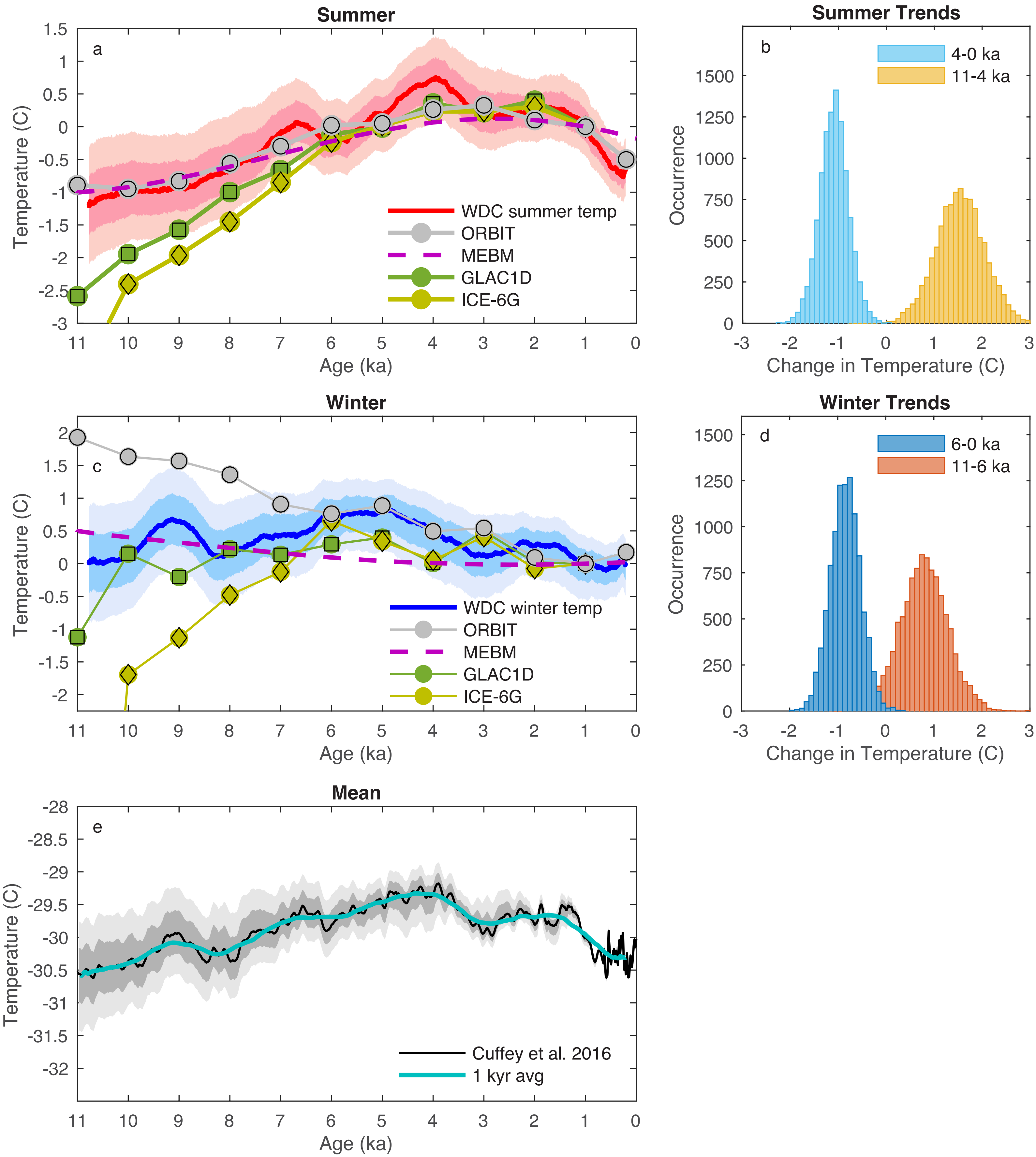
ED Figure 4 | MEBM results. *MEBM seasonal surface temperatures are shown in a-c.* ***a,*** *Modeled seasonal surface temperature cycle at 80°S, colored by age.* ***b, c,*** *Zoom in of modeled summer and winter temperature. MEBM annual results for the mean, maximum, and minimum are shown in d-l. Plots for temperature anomaly normalized to the mean (blue), insolation (red), and heat divergence (black) for the annual mean (d-f), annual max (g-i), and annual min (j-l). Note the sign of heat divergence; negative values correspond to heat convergence at the site.*

ED Figure 5 | 80°S energy balance at 10 ka. *Bar charts of HadCM3 energy balance terms at 10 ka for* ***a,*** *December (summer) and* ***b,*** *July (winter), including model runs for 'orbit only' (purple, ORBIT), 'ice sheet only' (blue, GLAC1D; green, ICE-6G), and 'all forcings' (orange, GLAC1D; yellow, ICE-6G). Positive values all indicate a surface or atmospheric warming. Variables include surface temperature (T_{surf} in °K), latent heat (LH in Wm^{-2}), sensible heat to the surface (SH_d in Wm^{-2}), shortwave radiation (SW in Wm^{-2}), downward LW radiation (LW_d in Wm^{-2}), and change in heat transport ($\nabla \cdot \boldsymbol{F}$ in 10^7 W).*

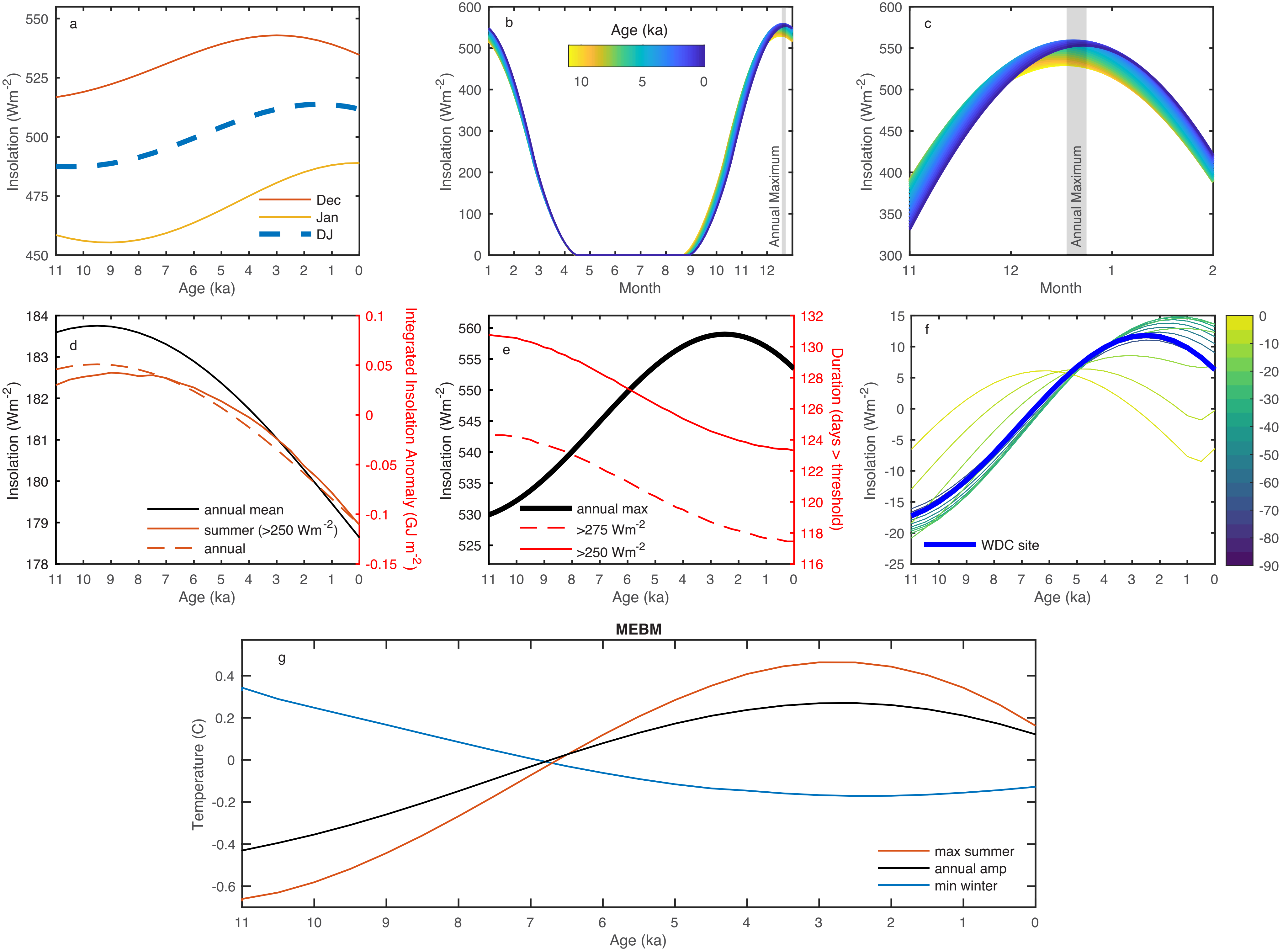
ED Figure 6 | Sea ice variability and temperature in HadCM3 simulations. *Maps of the dominant pattern of variability in each of the seasons and scatter plots of amplitude of the pattern against temperature. Maps were created using Python's package cartopy.* ***a,c,*** *The EOF of sea-ice variability in the Southern Hemisphere for December and July in ALL of the simulations: this is the dominant pattern of sea ice variability.* ***b,d,*** *The amplitude of the patterns in (a) and (c) vs. the temperature at 80°S for each simulation. Plots e-h show the zonal mean temperature and sea ice in HadCM3 simulations for December average.* ***e,*** *Change in sea-ice fraction for coupled-model simulations from 10k to 0k.* ***f,*** *Change in surface temperature between 10k and 0k for the coupled-model simulations (dotted line) and atmosphere-only simulations (solid line).* ***g,*** *Change in surface temperature for atmosphere-only runs from 10k to 0k.* ***h,*** *Change in surface temperature for atmosphere-only runs from 10k to 0k. See ED Table 3 for descriptions of the model simulations used in panels (g) and (h).*

ED Figure 7 | Model results vs. WDC temperatures. *Coefficient of determination and p-values for comparison of HadCM3 (1-kyr resolution, n=12) or MEBM (0.5-kyr resolution, n=23) model results with WDC summer and winter temperatures.*

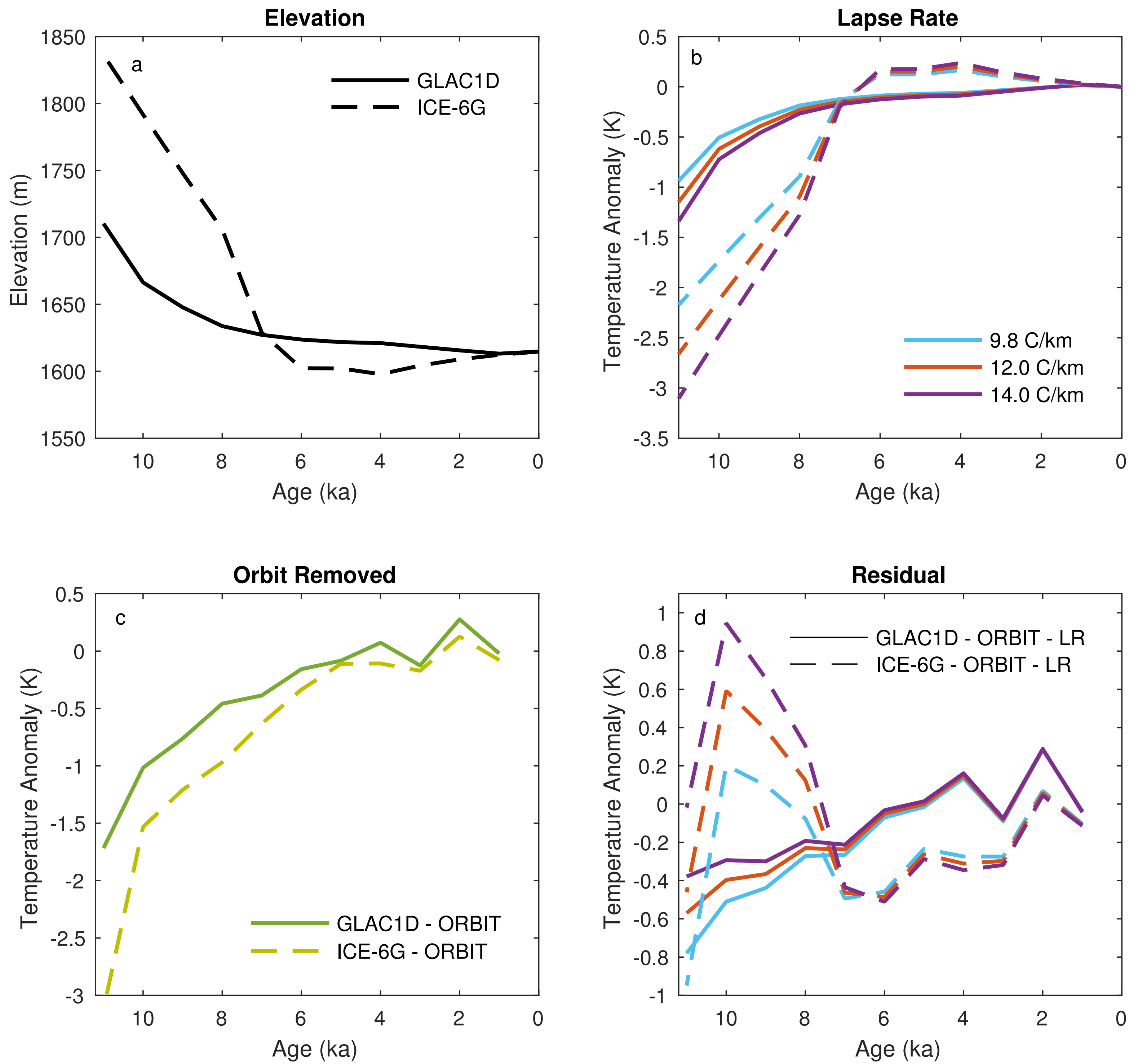

a
raw
diff corr
δD (per mil)
Age (ka)
b
amplitude
δD Annual Amplitude (per mil)
Age (ka)
maximum summer
minimum winter
mean
c
d
e
δD (per mil)
Age (ka)



Summer
a
WDC summer temp
ORBIT
MEBM
GLAC1D
ICE-6G
Temperature (C)
Age (ka)
Summer Trends
b
4-0 ka
11-4 ka
Occurrence
Change in Temperature (C)
Winter
c
WDC winter temp
ORBIT
MEBM
GLAC1D
ICE-6G
Temperature (C)
Age (ka)
Winter Trends
d
6-0 ka
11-6 ka
Occurrence
Change in Temperature (C)
Mean
e
Cuffey et al. 2016
1 kyr avg
Temperature (C)
Age (ka)



a
Dec
Jan
DJ
Insolation (Wm-2)
Age (ka)
b
Age (ka)
10
5
0
Annual Maximum
Insolation (Wm-2)
Month
c
Annual Maximum
Insolation (Wm-2)
Month
d
annual mean
summer (>250 Wm-2)
annual
Insolation (Wm-2)
Integrated Insolation Anomaly (GJ m-2)
Age (ka)
e
annual max
>275 Wm-2
>250 Wm-2
Insolation (Wm-2)
Duration (days > threshold)
Age (ka)
f
WDC site
Insolation (Wm-2)
Age (ka)
MEBM
g
max summer
annual amp
min winter
Temperature (C)
Age (ka)


Elevation
a
GLAC1D
ICE-6G
Elevation (m)
Age (ka)
Lapse Rate
b
9.8 C/km
12.0 C/km
14.0 C/km
Temperature Anomaly (K)
Age (ka)
Orbit Removed
c
GLAC1D - ORBIT
ICE-6G - ORBIT
Temperature Anomaly (K)
Age (ka)
Residual
d
GLAC1D - ORBIT - LR
ICE-6G - ORBIT - LR
Temperature Anomaly (K)
Age (ka)