Ocean forcing of penultimate deglacial and last interglacial sea-level rise

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Sea-level histories during the two most recent deglacial-interglacial intervals experienced significant differences[1-3] despite both periods having similar changes in global mean temperature[4,5] and forcing from greenhouse gases[6]. Although the last interglaciation (LIG) experienced stronger boreal summer insolation forcing than during the present interglaciation[7], understanding why LIG global mean sea level may have been 6-9 m higher than present has proven particularly challenging[2]. During glacial as well as interglacial periods, extensive areas of polar ice sheets were grounded below sea level, with grounding lines and fringing ice shelves extending onto continental shelves[8], suggesting that oceanic forcing by subsurface warming may also have contributed to ice-sheet loss[9-12] analogous to ongoing changes by the Antarctic[13,14] and Greenland[15] ice sheets. Such forcing would have been especially effective during glacial periods when the Atlantic Meridional Overturning Circulation (AMOC) experienced large variations on millennial timescales[16], with a reduction of the AMOC causing subsurface warming throughout much of the Atlantic basin[9,12,17]. Here we show that greater subsurface warming induced by the longer duration of reduced AMOC during the penultimate deglaciation can explain the more-rapid sea-level rise than during the last deglaciation. This greater forcing also contributed to excess loss from the Greenland and Antarctic ice sheets during the LIG, causing global mean sea level to rise at least 4 m above modern. When accounting for the combined influences of penultimate and last-interglacial deglaciation on glacial isostatic adjustment, this excess loss of polar ice during the LIG can explain much of the relative sea level recorded by fossil coral reefs and speleothems at intermediate- and far-field sites.

Climate evolution over the last two terminations shares a number of similarities (Extended Data Fig. 1). Proxy records of ocean circulation show that the last two terminations were accompanied by large reductions of the AMOC. Climate responses to these reductions show the characteristic bi-polar seesaw due to reduced northerly ocean heat transport and the weakening of the Asian monsoon due to the cooling of the Northern Hemisphere. Other similarities include an increase in the rate of sea-level rise when the AMOC begins to decrease and the occurrence of a Heinrich event during the period of reduced AMOC. Similar climate changes accompanied earlier terminations over the last 640 ka[18], suggesting that an AMOC reduction is a characteristic feature of these periods of rapid deglaciation.

There are also several notable differences between the last two terminations (Extended Data Figs. 1, 2). First, proxy data suggest that the AMOC during T-II remained in a reduced state for ~7,000 years before recovering at the start of the LIG. In contrast, during T-I, the AMOC only remained weak for ~3,500 years before recovering to nearly full strength during the 1,500-year Bølling-Allerød warm interval. It then decreased again during the 1,200-year Younger Dryas cold interval, with its final recovery at the start of the present interglaciation. Second, the full T-II sea-level rise occurred during the 7-kyr sustained "one-step" period of reduced AMOC, whereas only ~50% of the T-I sea-level rise occurred during the ~6.5-kyr "two-step" period of reduced AMOC[3] (Fig. 1). Third, ice-rafted debris (IRD) suggests that Heinrich event 11 (H11), which is nearly twice as long as Heinrich event 1 (H1), was sourced from more than just the Hudson Strait Ice Stream (HSIS), which was the primary source for H1[19].

A transient simulation of T-I climate used an atmosphere-ocean general circulation model (the National Center for Atmospheric Research Community Climate System Model version 3; NCAR CCSM3) forced by changes in insolation, CO_2, ice sheets, and freshwater fluxes that, while not in full agreement with reconstructions, were designed to cause the two-step reduction of the AMOC[17] (Fig. 1k). The simulation successfully captured many aspects of the climate evolution through T-I as recorded by proxy records[17,20,21]. Among the responses to the AMOC reduction was subsurface warming throughout much of the Atlantic basin[17] (Fig. 1k, 1l), which is supported by proxy temperature records from intermediate-depth (1,000-1,500 m) North Atlantic core sites[12].

We used the same climate model to conduct a transient simulation that spans T-II and the LIG (140-115 ka) (Methods). We applied freshwater forcing consistent with reconstructions that reproduced the 7-kyr "one-step" reduction in the AMOC suggested by proxy records of ocean circulation (Extended Data Figs. 1,3) in order to quantify associated changes in subsurface temperatures during T-II and into the LIG and thus allow direct comparison with subsurface warming simulated for T-I.

Fig. 1 compares forcing of ice-sheet surface mass balance from insolation, greenhouse gases (GHGs), and low-latitude Pacific sea-surface temperatures (SSTs) for T-II and T-I to representative examples of the simulated oceanic forcing at sites in the North (30°W, 45°N) (Fig. 1e, 1k) and South Atlantic (45°W, 70°S) (Fig. 1f, 1l). Changes in GHGs and SSTs are similar during the two terminations, with increases of ~2 W m^{-2} from GHGs and ~2°C warming from low-latitude Pacific SSTs, which strongly influence Northern Hemisphere ice-sheet surface mass balance[10]. Despite these similarities, sea level reached modern by the end of T-II while it remained ~50% below modern at the end of T-I (Fig. 1a, 1g). Some have attributed the faster rate of sea-level rise during T-II to the greater boreal summer insolation forcing[3], but that forcing only exceeds that of T-I after the majority (~80 m) of T-II sea-level rise had occurred (Fig. 1a, 1b, Extended Data Fig. 2). Otherwise, insolation forcing during the first 8,000 years of each termination is similar (~55 W m^2), whereas the associated 80 m of sea-level rise during T-II is much greater than the ~35 m during T-I.

We attribute this contrast in sea-level response to the similar radiative forcing and temperature changes of the last two terminations to the greater subsurface warming during T-II associated with the "one-step" reduction in the AMOC than during the T-I "two-step" reduction, leading to greater oceanic forcing of marine ice-sheet margins in the North (Fig. 1e, 1k, Extended Data Fig. 4) and South (Fig. 1f, 1l, Extended Data Fig. 4) Atlantic. Moreover, the Eurasian Ice Sheet during the Penultimate Glacial Maximum (PGM, ~140 ka) was larger than during the Last Glacial Maximum (LGM, ~21 ka), with most of the excess mass located in low-lying areas south-southeast of the glaciated Barents and Kara Seas[22] that, from isostatic depression, was also marine based (Extended Data Fig. 5). We thus hypothesize that collapse of this large marine-based ice complex triggered by oceanic forcing would have also contributed to the rapid T-II sea-level rise, with the associated IRD contribution to H11 diluting the contribution from the HSIS[19]. In general, this greater FW flux from deglaciating Northern Hemisphere ice sheets during T-II provided an important positive feedback on that deglaciation through its influence on the AMOC and subsurface temperatures.

Fig. 2 compares forcings during the last two interglaciations. Peak global mean SSTs were similar (Fig. 2d, 2i) while LIG radiative forcing from CO_2 was only slightly higher (~0.25 W m^{-2}) than during the present interglaciation (Fig. 2c, 2j). The main difference is in the higher boreal and lower austral summer insolation forcing during the LIG (Fig. 2b, 2h). Modeling studies show that this forcing would cause excess mass loss from the Greenland Ice Sheet during the LIG, but the estimated 1-3 m of global mean sea-level equivalent (GMSLE) is too small to explain the LIG highstand, thus requiring a contribution from the Antarctic Ice Sheet[2]. Lower austral summer insolation forcing during the LIG (Fig. 2b), however, results in surface cooling over most of Antarctica, suggesting an important role for oceanic forcing, with warming hypothesized to originate from an AMOC reduction during the LIG[11] or from a lagged ice-sheet response to warming from a change in the strength and/or position of the Southern Ocean westerlies associated with the T-II AMOC reduction[3]. One ice-sheet model simulates up to 6.7 m of sea-level rise when specifying a uniform increase of Southern Ocean temperatures by 3°C[11].

Our transient climate simulation shows that T-II oceanic forcing in the Southern Ocean as well as the North Atlantic continued into the early LIG (Fig. 2e, 2f, Extended Data Fig. 4). We use the Parallel Ice Sheet Model (PISM) to assess the response of the Antarctic and Greenland ice sheets to this oceanic as well as surface forcing through T-II and into the LIG as simulated by our transient climate run (Methods). The Greenland Ice Sheet starts to deglaciate from its PGM extent when adjacent ocean temperatures begin to warm at ~137.5 ka (Fig. 1e, Fig. 3g). It reaches its present extent at 131.5 ka and then loses an additional 0.88 m GMSLE by 119.5 ka largely by oceanic forcing of those sectors of the ice sheet that remain marine based, causing drawdown of the ice-sheet interior (Fig. 3c). The majority (3.42 m) of the total sea-level rise (3.88 m) occurs between 136-129 ka (Fig. 3g), corresponding to the period of rapid rise in global mean sea level (GMSL) (Fig. 1a). Sensitivity tests in which ocean temperatures are held constant at either PGM or LIG values show that the simulated deglaciation is controlled entirely by oceanic forcing (Methods, Extended Data Fig. 6), supporting our hypothesis that oceanic forcing contributed to deglaciation of other Northern Hemisphere ice sheets (Extended Data Fig. 4).

Our simulations also show that the major deglacial phase of the Antarctic Ice Sheet from its PGM extent closely coincides with the onset of warming of adjacent ocean temperatures at ~137.5 ka induced by the AMOC slowdown (Fig. 1f, Fig. 3g). In particular, the ice sheet retreats to its present extent at ~128 ka, with the majority (6.25 m) of the total (6.65 m) sea-level rise also occurring during the rapid T-II rise in global sea level (Fig. 3g). Sea-level rise then slows beginning at 128 ka, followed by an acceleration starting at 126.5 ka, with a total of 2.99 m of LIG sea-level rise occurring by 116 ka (Fig. 3g). The majority of this LIG deglaciation is associated with collapse of the Amundsen Sea sector of the West Antarctic Ice Sheet (WAIS) (Fig. 3e) largely in response to oceanic forcing (Extended Data Fig. 6), similar to what is suggested by observed recent changes and projected for future ice-sheet recession in this area[13,14]. This destabilization leads to retreat that continues after the period of peak oceanic forcing at a rate that is determined largely by the retrograde gradient of the bed beneath WAIS, followed by a slowing of retreat as the Southern Ocean cools (Fig. 2f).

We next apply an ice-age sea-level model[23] to predict how our simulated changes in LIG ice-sheet mass would be recorded at three widely distributed sites with well-dated corals that provide minimum estimates of relative sea level (RSL) during the LIG[24,25], and a speleothem record that bounds RSL during the same period[26] (Methods). Of the five adopted ice histories, the two (LAM and HYB) based on studies that use a significantly larger Eurasian Ice Sheet during the PGM relative to the LGM[22,27] best predict RSL histories during the LIG that are consistent with the elevation of the corals from the Bahamas, Western Australia and the Seychelles (Fig. 4a-c, Extended Data Fig. 7). However, all simulations tend to underestimate the first half (prior to 122 ka) of the LIG RSL inferred from the speleothem record in Mallorca (Fig. 4d, Extended Data Fig. 7).

In the absence of melting of polar ice during the LIG, predictions of RSL at the Bahamas and at Mallorca would show a monotonic rise, while those at the Seychelles and Western Australia would tend to show a monotonic fall[28]. Our ice-sheet simulations, however, are characterized by excess melt from WAIS (relative to present day) that increases from 0 to 3 m GMSLE between 127 ka and 124 ka (Fig. 3g). This signal is responsible for the accentuated RSL rise over the same period in the prediction for Bahamas and Mallorca and the reversal in the RSL trend at Western Australia and the Seychelles (Fig. 4).

Our results do not account for several other processes that may have caused LIG global mean and relative sea level to have been even higher than modeled here. For example, our atmospheric modeling may underestimate surface melting around the lower-elevation margins of the Greenland Ice Sheet. The greater boreal summer insolation forcing during the LIG relative to the present interglaciation likely caused an even greater loss of glaciers, which today account for 0.41 m GMSLE. Warmer-than-present LIG temperatures would have caused additional thermosteric sea-level rise[5]. Finally, we note that any additional melt near the start of the LIG would change the preferred Earth models identified in our analysis (Methods). LIG RSL can also be influenced by dynamic topography due to mantle convection, introducing meter-scale displacement on these 10^5-year timescales[29].

In summary, several lines of evidence suggest that the greater oceanic forcing during T-II than during T-I, as simulated by our climate modeling, contributed to the more-rapid sea-level rise during T-II. First, forcing of ice-sheet surface mass balance was similar during the two terminations, indicating that an additional forcing was required to explain the differences in rates of sea-level rise. Second, our GIA modeling demonstrates that the larger PGM Eurasian ice sheet caused a significantly larger fraction of the ice-sheet bed to be below sea level, and thus be more vulnerable to oceanic forcing, than during T-I. Third, our ice-sheet modeling shows that >85% of the volume loss of the PGM Greenland and Antarctic ice sheets to their present sizes occurs in response to oceanic forcing during T-II. Although additional modeling of the deglaciation of the former NH ice sheets during T-II will be required to further support this hypothesis, our ice-sheet modeling does show that oceanic forcing was the primary driver of excess ice loss from the Greenland and Antarctic ice sheets during the LIG. Our sea-level predictions demonstrate that the modeled 4 m of GMSLE from LIG deglaciation of Greenland and Antarctic ice sheets may explain much of the LIG RSL at intermediate- and far-field sites when GIA from T-II deglaciation is included, although additional melt is required to fully reconcile these data.

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Fig. 1. Sea-level change and climate forcings during the penultimate and last deglaciations. a, g, Records of relative and global mean sea level (Methods, Extended Data Fig. 2). Uncertainty on blue line is 1σ; uncertainty on coral data (circles) is 2σ. **b, h,** June 21 insolation for 65ºN[7]. **c, i,** Radiative forcing from greenhouse gases (CO_2, CH_4, and N_2O)[6]. Uncertainty is square root of the sum of squares of the uncertainties of the individual greenhouse gases. **d, j,** Tropical (23.5°N-S) mean annual sea-surface temperature stack with 2 s.d. relative to the HadISST1.1 1870–1889 data[4,5]. **e, k,** Changes in the model maximum Atlantic meridional overturning transport (below 500 m) (black line) and of temperature as a function of time and depth at 30ºW, 45ºN relative to 140 ka (**e**) and 22 ka[17] (**k**) as simulated by the NCAR CCSM3 (Methods). **f, l,** Evolution of temperature as a function of time and depth at 45ºW, 70ºS relative to 140 ka (**f**) and 22 ka[17] (**l**) as simulated by the NCAR CCSM3 (Methods).

Fig. 2. Sea level and climate forcings during the last and present interglaciations. a, g, Records of relative and global mean sea level (Methods, Extended Data Fig. 2). Uncertainty on coral data (circles) is 2σ. **b, h,** June 21 insolation for 65ºN[7]. **c, i,** Radiative forcing from greenhouse gases (CO_2, CH_4, and N_2O)[6]. Uncertainty is the square root of the sum of squares of the uncertainties of the individual greenhouse gases. **d, j,** Global mean annual sea-surface temperature stack with 2 s.d. relative to the HadISST1.1 1870–1889 data[4,5]. **e, k,** Evolution of temperature as a function of time and depth at 30ºW, 45ºN relative to 140 ka (**e**) and 22 ka[17] (**k**) as simulated by the NCAR CCSM3 (Methods). **f, l,** Evolution of temperature relative to 140 ka as a function of time and depth at 45ºW, 70ºS relative to 140 ka (**f**) and 22 ka[17] (**l**) as simulated by the NCAR CCSM3 (Methods).

Fig. 3. Simulations of the Greenland and Antarctic ice sheets. a, The Greenland Ice Sheet at the Penultimate Glacial Maximum. Logarithmic scale bar for velocity fields shown by color scheme; contours on ice sheet are for surface elevation in meters. **b,** The Greenland Ice Sheet at 116 ka. Logarithmic scale bar for velocity fields shown by color scheme; contours on ice sheet are for surface elevation in meters. **c,** Difference in thickness of the Greenland Ice Sheet between 116 ka and present day. Contours on ice surface and color scheme show change in thickness. **d,** The Antarctic Ice Sheet at the Penultimate Glacial Maximum. Logarithmic scale bar for velocity fields shown by color scheme; contours on ice sheet are for surface elevation in meters. **e,** The Antarctic Ice Sheet at 116 ka. Logarithmic scale bar for velocity fields shown by color scheme; contours on ice sheet are for surface elevation in meters. **f,** Change in thickness of the Antarctic Ice Sheet between 116 ka and present day. Contours on ice surface and color scheme show change in thickness. **g,** Contributions of the Greenland and Antarctic Ice Sheets to global mean sea level between 140 ka and 116 ka.

Fig. 4. Predictions of relative sea level at four far-field sites. a, Relative sea-level (RSL) data from the Bahamas based on well-dated corals compared to a prediction of RSL using our simulated LIG loss from the Greenland and Antarctic Ice Sheets and the LAM ice history[22] (solid green line) and the HYB ice history (solid blue line) (see Methods). Also shown are predictions of RSL using just the LAM ice history[22] (dashed green line) and the HYB ice history (dashed blue line). **b, c,** As in **a**, except for Western Australia (**b)** and the Seychelles **(c)**. Uncertainties are for age (2σ), elevation (downward) and coral-depth habitat (upward). **d,** RSL data from Mallorca based on speleothem records. Uncertainties are for age (2σ) and growth. The Earth models used in the calculations are characterized by a lithospheric thickness, and upper and lower mantle viscosity

of: 140 km, 0.3 X 10^{21} Pas, 8.0 X 10^{22} Pa s (**a**), 96 km, 0.3 X 10^{21} Pas, 5.0 X 10^{22} Pa s (**b**), 30 km, 0.5 X 10^{21} Pas, 3.0 X 10^{22} Pa s (**c**), and 120 km, 2.0 X 10^{21} Pas, 8.0 X 10^{22} Pa s (**d**). Each of the coral records are comprised of data collected from multiple sites and the RSL predictions are shown for the following representative locations: 24.05°N, 285.47°E (**a**), 21.97°S, 113.93°E (**b**), 4.28°S, 55.73°E (**c**), and 39.61°N, 3.38°E (**d**). The consistency between the data and the predictions would be unaffected if we plotted RSL histories at each location that accounted for the variable collection sites.

Methods

Transient climate modeling. We use the fully-coupled configuration of the Community Climate System Model version 3 (CCSM3) in T31 resolution[30] for the transient simulation of the penultimate deglaciation and last interglaciation (LIG) from 140 to 120 ka. CCSM3 was used in the TraCE-21K transient simulation of the past 21,000 years spanning the last deglaciation and the current interglaciation[17,21,31]. The transient simulation of the penultimate deglaciation was initialized with a 600-year equilibrium simulation of the penultimate glacial maximum that branched off the TraCE-21K last glacial maximum (LGM) simulation with orbital configuration[7] and greenhouse gas contribution (CO_2) for 140 kyr[32]. The transient simulation of penultimate deglaciation with CCSM3 was integrated from 140 ka to 129 ka with changing atmospheric greenhouse gas concentrations[32], Earth's orbit[7], and continental ice sheets based on ICE-5G[33] but with the timing of the corresponding sea-level rise adjusted to closely follow the Waelbroeck et al.[1] and Grant et al.[34] sea-level reconstructions for the penultimate deglaciation (Extended Data Figs. 2, 3). We note that our sea-level modeling suggests that the sizes of the penultimate glacial maximum Northern Hemisphere ice sheets differed from the LGM and thus our climate modelling does no account for these important difference between the two terminations. Further climate modeling is thus needed to assess how these differences may have affected atmospheric circulation over the North Atlantic Ocean and the AMOC.

To simulate the impact of freshwater forcing from Heinrich event 11 on the Atlantic Meridional Overturning Circulation, freshwater is added at the surface of the North Atlantic in the area between 50°-70°N, being ramped to 0.17 Sverdrups (Sv; 10^6 m^3 s^{-1}) from 138 ka to 135.5 ka where it remains until 129.7 ka when it is shut off (Extended Data Fig. 3). The transient simulation of the LIG with CCSM3 was integrated from 129 ka to 116 ka with changing orbits and atmospheric greenhouse gases under present-day ice-sheet configuration. No additional freshwater fluxes were applied during the transient simulation of the LIG.

Ice-sheet modeling. We use version 0.7.1 of the Parallel Ice Sheet Model (PISM), in which the dynamical core superposes velocity fields from the shallow shelf and shallow ice approximations across the entire domain. Fast flow ("streaming") of grounded ice is enabled by plastic failure of subglacial sediments, which depends on a prescribed but spatially-variable till friction angle, representing sediment strength and its degree of saturation. The till friction angle is based primarily on topography, so that deeper areas have lower friction angles. This mimics the effect of weaker sediments accumulating in deeper basins. The parameterization follows the form, *phi min / phi max / elevation min / elevation max*, in which the *phi min* is the friction angle applied below *elevation min*, *phi max* is the friction angle applied above *elevation max*, and values in between are linearly interpolated. For our Greenland simulations we prescribe values of 10 / 30 / -300 / 300, and for Antarctica 6 / 30 / -700 / -100. These values are based on, but modified from, previous work (Aschwanden et al.[35] for Greenland; Golledge et al.[36] for Antarctica), but these values are uncertain. Our values were chosen following exploratory simulations that sought to best capture the broad-scale geometric and dynamic features of the ice sheets.

Sediment strength evolves dynamically depending on the basal ice temperature. Where ice is sufficiently thick to allow basal melting, meltwater weakens the substrate until driving stresses exceed till cohesion. Failure of the substrate that results in acceleration of overlying ice follows a pseudo-plastic law[37,38], such that a small increase in stress above the shear strength of the substrate leads to an increasing velocity response. This ultimately thins the ice, which reduces the gravitational driving stress and results in a deceleration of the ice sheet. The cyclic behavior of ice streams that occurs as a consequence of this mechanism is described in more detail elsewhere[39].

PISM uses a sub-grid grounding line scheme[40] in which the interpolation of sub-ice shelf melt across the grounded to floating transition may be turned on or off. When turned on, the scheme tends to accelerate ice-sheet retreat in marine basins, whereas when it is off, the scheme produces a slower response[36]. This difference in behavior results in differences in retreat rates, but equilibrium states (for example, ice volume) are less affected. In our experiments we investigated both approaches, and found that interpolating sub-shelf melt across the grounding line produced simulations that were most closely in keeping with geological constraints for Termination I (T-I) (see below).

We also used a range of enhancement factors for the shallow ice (SIAe) and shallow shelf (SSAe) equations (SIAe = 1, 2, 3; SSAe = 0.5, 1), and different values for the basal sliding exponent that controls how plastic or linear the substrate deformation response to applied driving stresses ($q = 0.25, 0.6$). Floating ice is controlled by two calving mechanisms – one based on horizontal strain rates[41] and another that prescribes a minimum thickness criterion (50 m for Greenland, 200 m in Antarctica).

We run separate simulations for the Greenland and Antarctic ice sheets, both at 20-km resolution. To drive our ice-sheet model, we use output climatologies from the transient CCSM3 simulations described above for T-I and Termination II (T-II). Atmospheric outputs are applied as anomalies to present-day air temperature and precipitation fields[42,43], in the same manner as employed previously[44]. We employ a positive-degree day (PDD) model to translate temperatures above freezing into surface melt, of which 60% remains in the snowpack as a consequence of refreezing during percolation. The proportion of refreezing that takes place even under present conditions is difficult to constrain precisely[45] so we use a uniform value both for the control and perturbation experiments, in order to minimize the effects of this parameterization. That is, differences in the simulation outputs are unlikely to arise from uncertainty in this aspect of model parameterization.

During our model tuning process, we explored a wide range of degree-day values (from 1 mm $^{o}C^{-1}$ day^{-1} up to 64 mm $^{o}C^{-1}$ day^{-1}) independently for both snow and ice. We tried the more usual melt threshold of 273 K and the lower value of 270 K following van den Broeke et al.[46]. The latter method yields more widespread melt, mimicking the possible melt arising from short-wave radiation under sub-freezing conditions, and thus degree-day factors are typically lower (van den Broeke et al., 2010). We also allow for stochastic variability in daily temperatures using a zero-mean white noise component whose standard deviation is set at 5 K. Although the choice of PDD parameters did exert some control on the geometry of the evolving ice mass, the basic shape of the ice sheet evolved in a similar manner regardless of either the melt forcing or the glaciological parameterization, suggesting that the dominant control on ice-sheet geometry is the climate forcing from the GCM. Recent work has shown that our simulations of surface mass balance (SMB) of the Greenland Ice Sheet under the high boreal summer insolation of the LIG may be sensitive to climate model resolution and SMB model type (i.e., PDD, surface energy balance)[47].

Oceanic fields for temperature and salinity at 500 m depth were used as inputs to a thermodynamic ocean model that calculates basal melt from salt and heat-flux gradients across the ice/ocean interface, according to the scheme described in[48]. As with the atmospheric variables, we apply the oceanic fields as anomalies from a present-day ocean configuration that for Antarctica is tuned to reproduce observed melt-rate patterns[49]. Since such constraints are not currently available for Greenland, we use a spatially uniform melt factor instead, which is iteratively refined so that both LGM and present-day ice-sheet extents are reproduced (see below). Ice thickness and bed topography for the two ice sheets are taken from the most recent compilations[50,51].

With the model set-up as described above, we ran a series of time-evolving experiments that first focused on T-I, rather than T-II. The rationale for this approach is that substantial geological data exist with which to constrain the evolving ice-sheet geometry through the last deglaciation, whereas there are few constraints for the preceding T-II. Therefore, in order to optimize our parameter settings, we undertook >500 experiments of T-I for both ice sheets until a good fit to empirical constraints was found. For Antarctica, our guiding constraints are that the ice sheet at the LGM, immediately prior to T-I, should occupy the majority of the continental shelf, and have an ice-volume excess above present that is within the range of 5.6-14.5 m represented by previous simulations[52-55]. Furthermore, we required that the evolution of the simulated ice sheet must reproduce the glacial maximum thickening of West Antarctica and thinning of East Antarctica inferred from ice-core analyses[55], and exhibit a pattern of mass loss that is consistent with geologically inferred deglacial changes in ice discharge[56]. In Greenland, geological constraints on the offshore extent of the LGM ice sheet are sparse, but the ice volume excess is thought to have been in the range 2 – 5 m global mean sea level equivalent[57,58]. We use this range as our target (Extended Data Fig. 8). Finally, both ice sheets are required to reproduce present-day grounded ice extent and volume as closely as possible at the end of the T-I simulations.

Once this phase of parameter optimization is complete, we run our experiments for T-II using the exact same settings, changing only the input climatology based on outputs from CCSM3. This dual approach allows for the robust simulation of a period, such as T-II, for which little data exist to constrain outputs. In addition, this methodology allows for the direct comparison of model outputs for the two periods, allowing any differences to be attributed solely to the imposed climate forcing rather than to uncertainties in the modelling procedure. Finally, by tuning the model to fit relatively well-known constraints such as LGM and present-day extent and volume, we reduce the influence of any inaccuracies in the climate model representation of air or ocean temperatures during the periods of simulation. Thus, if CCSM3 under- or overestimates the magnitude of past climate anomalies with respect to present, the internal consistency between the T-I and T-II climate simulations coupled with the data-constrained simulation of T-I mean that the reliability of the T-II simulation is unaffected.

A novelty of our ice-sheet simulations compared to previous studies[59,60] is that we use a fully evolving T-I experiment to constrain our model parameterizations. This includes components such as degree-day factors for the PDD scheme. For Greenland, we run an ensemble of tuning experiments that explore a range of snow and ice melt factors as well as ice-flow enhancement coefficients (Extended Data Fig. 8). By then selecting the parameterization that at the end of the T-I simulation most closely reproduces present-day ice volume and geometry we ensure that the surface melt fields we generate are realistic. We then apply this setup to our T-II experiments. Our annual temperature range is defined by the CCSM3 outputs. However, we also experimented with duplicate simulations in which we modified our Greenland climatologies to incorporate summer temperatures from Fausto et al.[42]. These simulations resulted in only minor differences in mass change, suggesting that in our experiments, atmospheric forcing plays a lesser role than oceanic forcing (Extended Data Figs. 6, 9). This is supported by experiments in which we also explored alternative grounding line schemes to make the ice sheets either more or less sensitive to ocean temperature change. In the less sensitive experiments, the ice sheet failed to advance sufficiently far offshore, and was thus incompatible with geological constraints.

Predictions of relative sea level. Calculations of glacial isostatic adjustment described in the text are based on a pseudo-spectral sea-level theory[23] for the case of spherically symmetric (i.e., rheology varies with depth alone), Maxwell viscoelastic Earth models, with a truncation at spherical harmonic degree and order 256. The theory incorporates time-varying coastlines, changes in the perimeter of grounded, marine-based ice sheets, and the impact on sea level of load-induced perturbations to the Earth's rotation axis, where these perturbations are computed using the rotational stability theory of Mitrovica et al.[61]. Profiles of the density and elastic structure of the Earth model are taken from the seismic Preliminary Reference Earth Model[62]. The viscosity structure of the Earth models is defined by three layers, a lithospheric zone of infinite viscosity, and sub-lithospheric upper and lower mantle regions, where the boundary between the latter two regions is taken to be 670 km depth. The thickness of the lithosphere and the viscosity of the upper and lower mantle are free parameters of the modeling and are varied, respectively, within the following ranges: 30 – 140 km; 2-20 X 10^{20} Pa s; and 2-100 X 10^{21} Pa s.

The set of five ice histories adopted in this study are based, in part, on histories constructed by Dendy et al.[28] in their investigation of the sensitivity of LIG sea level predictions to variations in the timing and geometry of ice cover during Marine Isotope Stage 6. We begin by summarizing these ice histories.

All models in Dendy et al.[28] use the ICE6G ice history[63] for the period extending from the LGM to present day and they extend back four full glacial cycles. The models are constrained to have interglacial ice volumes and geometry identical to present-day ice cover on the Earth (i.e., there is no excess ice melting during previous interglaciations, including the LIG). The so-called WAE ice model adopts the eustatic sea-level curve estimated by Waelbroeck et al.[1] on the basis of benthic foraminifera isotope records. In the period prior to the LGM, the ice geometry is constrained to be identical to the geometry post-LGM whenever the eustatic values are identical. The LAM and COL models in Dendy et al.[28] also adopt the pre-LGM eustatic curve of Waelbroeck et al.[1], but are distinguished from WAE by their ice history during the penultimate glacial cycle.

In particular, these models adopt the ice geometry during the penultimate glacial maximum (PGM) inferred by Lambeck et al.[22] and Colleoni et al.[27], which are both characterized by more significant ice cover over Eurasia during the PGM than the LGM. Since the difference in peak Eurasian ice volume during the PGM in the LAM and COL models is large (55 m and 71 m, respectively, in units of equivalent GMSL), we have constructed an intermediate ice history (HYB) that is essentially the average of these models (peak volume of 66 m GMSL equivalent during the PGM) The increased ice cover of the LAM, HYB and COL models relative to the WAE model is compensated, in large part, by a reduction of the volume of the Laurentide Ice Sheet during the PGM relative to the LGM[22,27]. All four models, WAE, LAM, HYB and COL, converge to the same ice geometry (i.e., the present-day ice geometry) at the beginning of the model LIG. We note that we have adapted the WAE, LAM and COL models described by Dendy et al.[28] to more closely follow the eustatic curve of Waelbroeck et al.[1]. Finally, the model SHA in Dendy et al.[28] is constructed in a manner identical to WAE, with the exception that the model adopts the eustatic curve derived by Shakun et al.[64] in the period prior to the LGM.

The ice histories considered in the present study combine the five models described above with the Antarctic and Greenland Ice Sheet histories discussed in the main text. Specifically, the difference in ice height during the period from 140 ka to 116 ka relative to the present day in the ice-sheet simulations of the main text are applied to each of the Dendy et al.[28] models. The net result is that the five models constructed in this manner are characterized, in contrast to those in Dendy et al.[28], with excess melting of the Antarctic and Greenland Ice Sheets during the LIG relative to present-day. We ran 337 Earth models for each of the five ice histories (total of 1685 simulations) in which parameters defining the Earth model were varied over plausible ranges.

In exploring the fit of the relative sea level (RSL) predictions to the coral record, we considered three sites that have the largest data sets of well-dated corals (Bahamas, Seychelles and Western Australia) and a relatively new speleothem data set from Mallorca (*43*) (Extended Data Fig. 7). Given that corals provide a minimum bound on sea level, our metric for fit for these data was the number of coral records that any specific RSL prediction bounded from above. In contrast, we interpret the height uncertainties associated with the published speleothem data to represent a two-way bound on peak RSL.

None of the 1685 simulations (i.e., our sampling of 337 Earth models and 5 ice histories) were successful in bounding all coral records from above. As an indication of performance, Extended Data Fig. 7 shows predictions from the full suite of simulations that satisfy the following criteria: (1) all coral data from Western Australia and Bahamas, with the exception of the earliest datum at the latter site (at ~131 ka), fall below the prediction; and (2) the prediction at the Seychelles falls above all three coral records at an elevation of ~4 m. The various lines on the figure represent the different Earth models for each ice history that satisfy these constraints. For each ice history (i.e., each column of Extended Data Fig. 7), the Earth models sampled on each frame (i.e., each site) represent a discrete set that may or may not overlap with the set from a different site. As an example, in the case of the LAM and COL ice histories, no single Earth model appears on the results for all three sites. This is reflected in Fig. 4 in the main text, where the simulation highlighted in each frame is the result for a distinct Earth model. This variation is justified by the fact that the Earth's mantle is subject to large amplitude variations in viscoelastic structure and so it would be unexpected if the sea-level response at each of the three sites preferred the same Earth model.

Note that the number of simulations that satisfy our plotting criterion for the Bahamas increases as one moves to ice histories with larger Eurasian ice cover at the PGM (i.e., from the WAE to the COL results), but the number of simulations that satisfy the criteria for the Seychelles decreases in the same sense. While not apparent from Extended Data Fig. 7, the predicted highstand at Seychelles increases as one considers Earth models with progressively thinner elastic lithospheres (see Ref. 28, Fig. 9A), and the simulations that predict RSL highstands above the Seychelles records are those based on a lithospheric thickness of 30 km (as in Fig. 4 of the main text) or, in a couple of cases for the WAE ice history, 50 km. This raises two important issues. First, none of the simulations that yield RSL above all the coral elevations at Seychelles also satisfy the geological constraints at the Bahamas. Second, since the predicted highstand at the Seychelles is sensitive to the adopted lithospheric thickness, there is a trade-off between the preferred value of this parameter and the level of excess melting during the LIG. That is, increasing polar ice sheet melting above the ~4 m GMSL equivalent adopted in the simulations in Extended Data Fig. 7, would increase the range of lithospheric thickness that would satisfy the Seychelles coral record, and thus bring the inference into better accord with other GIA-based estimates of this Earth model parameter.

This issue may also have relevance in regard to the results for Mallorca (Extended Data Fig. 7) where simulations are only plotted if the misfit between the GIA predictions and the speleothem observations is within 50% of the minimum misfit achieved in all simulations. In this case, fewer of the simulations provide a reasonable fit to the speleothem record as one considers ice histories with progressively larger volumes over Eurasia at PGM and, indeed, no simulations based on the COL ice history satisfy our plotting criterion. However, regardless of the adopted ice history, none of the simulations fit the highstand constraints before 125 ka. Bringing the GIA predictions in Extended Data Fig. 7 into accord with the Mallorca observations would require additional excess melting that is limited to the earliest phase of the LIG.

As a final point, simulations based on the SHA ice history yielded misfits significantly larger than predictions shown in Extended Data Fig. 7.

The sea-level simulations described above yield changes in sea level and topography at each time slice of the ice history. As an example, Extended Data Fig. 5 shows the reconstructed topography for the area covered by the Scandinavian Ice Sheet at 131 ka, near the end of the MIS6 deglaciation, for a simulation based on the LAM ice history and a specific Earth model (see caption). The map supports the suggestion in the main text that the margin of grounded ice complexes in this region across MIS 6 through 5e were marine based.

Evidence for warming over the Greenland Ice Sheet during the Last Interglaciation. Here we evaluate the evidence for warming over the Greenland Ice Sheet during the Last Interglaciation (LIG). This supports our climate model simulation that while the LIG atmosphere was warmer than pre-Industrial, it largely remained below freezing and did not lead to significant mass loss from surface melting.

Regarding the reconstructed LIG temperatures at the NEEM[65] and GISP2[66] ice-core sites, there is uncertainty in which $d\delta^{18}O_{ice}/dT$ relationship should be used to reconstruct LIG temperatures, and this uncertainty is exacerbated when applying the modern $d\delta^{18}O_{ice}$-dT relationship to past climates, where differences in orbital forcing, moisture transport pathways, ice-sheet topography, and sea-ice extent can change the relationship[67-72]. To illustrate some of these uncertainties, we have compared our simulated temperatures for the NEEM and GISP2 ice-core sites with the temperature reconstructions for these sites based on $\delta^{18}O_{ice}$ (Extended Data Fig. 10). These reconstructions span the interval 127-120 ka, which is the warmest interval in the ice-core records for the LIG suggested by this proxy. The published reconstructed temperatures for GISP2 (blue symbols on upper panel)[66,73] and NEEM (dark blue line on lower panel)[65] are based on the relation $d\delta^{18}O_{ice}/dT$ = ~0.5‰ C^{-1} which is derived from Greenland ice-core sites elsewhere[74]. During the LIG, the precipitation-weighted $\delta^{18}O$ is likely biased to summer months rather than mean annual temperature (van de Berg et al., 2013), so we compare this reconstruction with our simulated summer temperature (JJA) (grey line on each panel). This suggests that our simulated JJA temperatures are underestimating the mean of the reconstructions by 4-5°C. This difference is reduced when we account for our simulated ice-surface lowering of ~200 m at NEEM and ~400 m at GISP2 (see Fig. 3C) and assume the lapse rate of 7.5°C km^{-1} used by Dahl-Jensen et al.[65], thus placing our results within the published uncertainties of the reconstructions (green line on each panel).

However, following the publication of Dahl-Jensen et al.[65], Masson-Delmotte et al.[75] established that the $d\delta^{18}O_{ice}/dT$ relation at the NEEM site is ~1.1‰ C^{-1}, suggesting that the NEEM and GISP2 LIG summer temperatures are about half of the originally published values based on the Vinther et al.[74] $d\delta^{18}O_{ice}/dT$ relation (red symbols on upper panel, red line on lower panel). Masson-Delmotte et al.[75] (p. 1500) conclude that "For the last interglacial period, the observed $\delta^{18}O$ anomaly of 3.6‰ at NEEM deposition site would then translate into 3.6$\pm$0.7 °C warming, instead of the estimate of 7.5$\pm$1.8 °C (NEEM, 2013) that was obtained using the Greenland average Holocene isotope–temperature relationship (Vinther et al., 2009)."

Our simulated JJA temperatures (grey line on each panel) are thus only 1-2°C colder than the mean reconstructions for GISP2 and NEEM based on this new calibration, but they are in excellent agreement with the mean values when accounting for our modeled ice-surface lowering (green line on each panel).

Landais et al.[76] used $\delta^{15}N$ from the NEEM core to reconstruct temperatures that were 8.5° ± 2.5 °C warmer during the LIG compared to preindustrial (PI). However, the $\delta^{15}N$ reconstruction represents annual temperature whereas the $\delta^{18}O_{ice}$ temperatures are biased to the summer, which is the critical season for influencing changes in surface mass balance through melting. The two temperature reconstructions are thus not directly comparable. Moreover, Landais et al.[76] identify "large uncertainties" (p. 1944) in their temperature reconstruction, including in the firn model used, in the assumed accumulation rates, and in the potential influence of surface melt on firn depth.

We thus conclude that when using the most suitable temperature calibration for the ice-core sites and within the uncertainties of the ice-core proxy reconstructions, our climate model successfully captures the LIG summer (JJA) temperature anomaly relative to pre-Industrial at NEEM and GISP2. Consistent with this model-data agreement for warmer LIG JJA temperatures, we find that the LIG surface mass balance of the GrIS is more negative than present day (Extended Data Fig. 9).

Dahl-Jensen et al.[65] stated "during our NEEM field campaigns (2007-2012), the mean surface air temperature in July reached -5.4°C." However, Box[77] reported the average JJA temperature for 2007-2012 at NEEM site as -10.9 ± 0.3°C, suggesting that Dahl-Jensen et al.[65] are reporting a maximum July temperature value during their period of record rather than climatology. But the JJA temperature that matters for comparing to the LIG is the pre-Industrial, which Box[77] found to be -12.6 ± 0.6°C for 1840-1870 (period of record closest to pre-Industrial). (Dahl-Jensen et al.[65] compared to the average of the last millennium.) Thus, even if the Dahl-Jensen et al.[65] LIG temperature reconstruction is correct (7.5±1.8 °C warmer than the mean of the past millennium), average LIG summer temperatures would still be well below freezing (~ -5°C). More likely, however, they are even further below freezing when using the $d\delta^{18}O_{ice}/dT$ relation established for the NEEM site[75], i.e., 3.6±0.7 °C warmer than the mean of the past millennium, with average LIG summer temperatures thus being -9°C.

The evidence for surface melt at the NEEM ice core site is based on: (1) a low-resolution record showing that out of 73 samples, seven have elevated CH_4 and N_2O during the interval 118-127 ka, and (2) a high-resolution CH_4 record that suggests five melt events in the 123.5-122.5 ka interval, or one every 200 years[65,78]. Noble gases that were measured at the times of four of the five elevated CH_4 events in the high-resolution record confirm melting at these times[78]. This alone makes it clear that these were infrequent periods of melting rather than continuous melting throughout the LIG. According to Anais Orsi (personal communication, March, 2019), during a melt event, such as the 2012 event[79], the melt percolates and refreezes in the top 1m of the firn, often in many layers, so one melt event may be represented by more than one melt layer. Moreover, although the noble gas results clearly identify four periods of enhanced melting, one cannot exclude the possibility that each sample represents a single 2012-like melt event.

In summary, ice-core proxies suggest that Greenland LIG temperatures were warmer than present, but constraining the amount of warming from these proxies remains uncertain. However, even the highest estimates of warming still suggest that average JJA temperatures remained well below freezing relative to pre-Industrial, and based on the more-appropriate $\delta^{18}O_{ice}$-temperature calibration from Masson-Delmotte et al.[75], are in good agreement with our simulated temperatures for the ice-core sites. Consistent with this model-data agreement for warmer LIG JJA temperatures, we find that the LIG surface mass balance of the GrIS is more negative then present day. Rare episodes of melting occurred, but while their frequency may increase under higher mean temperatures and insolation, such as is recorded in the Holocene section of the GISP2 ice core[80], we conclude that with a frequency of perhaps only one melt event every 200 years, they had a negligible influence on long-term surface mass balance, and average summer temperatures at the NEEM site otherwise remained well below freezing during the LIG.

Influence of freshwater forcing from modeled mass loss from the Greenland and Antarctic Ice Sheets on ocean circulation during the last interglaciation. We did not include additional freshwater (FW) forcing after 129.5 ka, but we show here that the FW fluxes from our modeled mass loss from the Greenland and Antarctic ice sheets (GrIS and AIS) after 129.5 ka (during the LIG) were too small to have influenced the Atlantic Meridional Overturning Circulation (AMOC) or Antarctic Bottom Water (AABW) formation. Since global sea level reached modern at 129-130 ka, and our modeled AMOC resumes at 129.5 ka, we only consider the FW fluxes from the GrIS and AIS since 129.5 ka.

From 129.5 to 127 ka, modeled GrIS mass loss was 0.2 m of sea-level equivalent, which is equivalent to a FW flux of 0.0009 Sv. From 127 ka to 117.5 ka, GrIS mass loss was 0.09 m, which is equivalent to 0.0001 Sv. For reference, Bakker et al.[81] showed that a FW flux of 0.01 Sv from Greenland for the RCP4.5 scenario (see their Fig. SI3) results in a median reduction in the AMOC of ~5% (their Fig. 2, GrIS only). The FW fluxes from LIG loss of the GrIS in our model are two orders of magnitude smaller than this, and thus would have no impact on the AMOC, and thus on our ice-sheet model simulations.

From 129.5ka to 123.5 ka, AIS mass loss was 4.1 m, which is equivalent to a FW flux of 0.008 Sv. Bakker et al.[82] found that a FW flux of 0.12 Sv from the AIS increases variability in AABW by ~10% and in AMOC by ~5%. The FW fluxes from LIG loss of the AIS in our model is a factor of 15 smaller than this, and thus would have no impact on AABW or the AMOC, and thus on our ice-sheet model simulations.

Code availability.

CCSM3 is freely available as open-source code from http://www.cesm.ucar.edu/models/ccsm3.0/

PISM is freely available as open-source code from https://github.com/pism/pism.git.

Data availability

Antarctic bedrock topography and ice thickness data are from the BEDMAP2 compilation, available at https://secure.antarctica.ac.uk/data/bedmap2/. Greenland topography and ice thickness data are from BedMachine v3, available at https://nsidc.org/data/idbmg4. Greenland mass balance and geothermal heat flux data are available from the seaRISE website: http://websrv.cs.umt.edu/isis/index.php/Data. Information on Antarctic surface mass balance data are available at http://www.projects.science.uu.nl/iceclimate/models/antarctica.php#racmo23. Antarctic geothermal heat flux data are available at the Open Science Framework https://doi.pangaea.de/10.1594/PANGAEA.882503. The datasets generated and used for this study (Figs. 1,2,3,4, Extended Data Figs. 3,4,5,6,7,8,9) are available from the Open Science Framework (DOI 10.17605/OSF.IO/FX7WK).

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Extended Data Fig. 1. Climate and sea-level records for Termination II and Termination I. **(a)** εNd records from the North Atlantic Ocean as proxies of Atlantic meridional overturning circulation (AMOC)[16,83]. **(b)** CCSM3 maximum AMOC transport (below 500 m) (this study). **(c)** EPICA Dome C δD record on AICC2012 age model as proxy of Antarctic temperature[84] (blue line) and percentage of warm planktonic foraminiferal species as proxy of North Atlantic sea surface temperatures[83] (grey line). **(d)** $\delta^{18}O$ record from Chinese stalagmite as proxy of Asian monsoon strength[18]. **(e)** Rate of sea-level change derived from a relative sea level (RSL) reconstruction based on benthic foraminifera isotopes[1]. **(f)** A stack of North Atlantic ice-rafted debris records recording Heinrich event 11 (H11)[83,85-87]. **(g)** εNd[88] (brown, orange symbols) and Pa/Th[89] (purple, green symbols, 1 sigma uncertainty) records from the North Atlantic Ocean as proxies of AMOC. **(h)** CCSM3 maximum AMOC transport (below 500 m) (this study). **(i)** EPICA Dome C δD record on AICC2012 age model (dark blue line)[84] as proxy of Antarctic temperature and temperature reconstruction from the Greenland GISP2 ice core (light blue line)[70]. **(j)** $\delta^{18}O$ record from Chinese stalagmite as proxy of Asian monsoon strength[18]. **(k)** Rate of sea-level change derived from a RSL reconstruction based on benthic foraminifera isotopes[1]. **(l)** A stack of North Atlantic ice-rafted debris records recording Heinrich event 1 (H1)[90].

Extended Data Fig. 2. Sea-level records for the last two terminations and interglaciations. (a) Sea-level reconstructions for the penultimate deglaciation and the last interglaciation (the latter identified by the grey-shaded area). Eustatic sea-level record is based on benthic foraminifera isotopes (blue line with 1σ uncertainty)[1] and relative sea-level (RSL) record is based on Red Sea isotopes (gray crosses; green line, 1-kyr moving Gaussian filter)[34] placed on a revised age model[91]. Also shown are RSL data from U-series dated corals at Tahiti (sky blue circles)[92], Huon Peninsula (light blue green circle; altered samples shown by gray circles)[93], the Seychelles (light green circles)[25], western Australia (blue circles)[24], and the Bahamas (cyan circles)[24]. All of the U-series ages have been recalculated to normalize them with the same set of decay constants for ^{234}U and ^{230}Th[94] and are shown with 2σ age uncertainty. We note that the offset between the Red Sea record (green line) and the benthic foraminifera record (blue line) may reflect the complex 3-dimensional Earth structure in the vicinity of the Red Sea rift[95,96]. The variability in the Red Sea and Huon Peninsula RSL records may reflect a sea-level reversal at ~137 ka[91] which, if it existed, was too small to be recorded by the benthic foraminiferal record. The rate of sea-level change based on the benthic foraminiferal record is also shown. **(b)** Sea-level reconstructions for the last deglaciation and the present interglaciation (the latter identified by the grey-shaded area). The record of global mean sea level is based on benthic foraminifera isotopes (blue line with 1σ uncertainty)[1]. Also shown are individual sea-level estimates (black circles, 2σ uncertainty) that have been corrected for glacial isostatic adjustment[97]. Rate of sea-level change based on the benthic foraminiferal record is also shown. **(c)** Upper panel shows eustatic sea-level reconstructions for the penultimate deglaciation (blue line with 1σ uncertainty) and the last deglaciation (black line with 1σ uncertainty)[1]. Lower panel shows June 21 insolation for 65°N for the penultimate deglaciation (blue line) and the last deglaciation (black line)[7].

Extended Data Fig. 3. Comparison of our freshwater forcing during T-II with other estimates. (**a**) Our simulated changes in AMOC. (b) Our FW forcing. (**c**) Reconstruction of freshwater (FW) flux from sea-level reconstructions from Waelbroeck et al.[1]. (**d**) Reconstruction of FW flux from sea-level reconstructions from Marino et al.[3] (**e**) Our stack of ice-rafted for Heinrich event 11 (H11) (Extended Data Fig. 1), which shows that the H11 interval of iceberg discharge is in good agreement with the timing of our FW forcing. (**f**) The sea-level change associated with our FW flux into the North Atlantic (grey line), the sea-level change associated with the ICE-5G ice sheets[33] used as a boundary condition in our climate model (green line), and a reconstruction of global sea-level change[1] (blue line with 1 sigma uncertainty). The timing of sea-level change in the ICE-5G time series shown here was adjusted from its chronology for T-I by adjusting the corresponding sea-level rise to closely follow the Waelbroeck et al.[1] and Grant et al.[34] sea-level reconstructions for the penultimate deglaciation.

Extended Data Fig. 4. Maps of the evolution of temperature at 400-m water depth in the North Atlantic, Arctic, and Southern Oceans between 138 ka and 124 ka relative to temperature at 140 ka. a-h, Maps of the evolution of temperature at 400-m water depth in the North Atlantic and Arctic Oceans for (**a**) 138-140 ka, (**b**) 136-140 ka, (**c**) 134-140 ka, (**d**) 132-140 ka, (**e**) 130-140 ka, (**f**) 128-140 ka, (**g**) 126-140 ka, and (**h**) 124-140 ka. **i-p,** Maps of the evolution of temperature at 400-m water depth in the Southern Ocean for (**i**) 138-140 ka, (**j**) 136-140 ka, (**k**) 134-140 ka, (**l**) 132-140 ka, (**m**) 130-140 ka, (**n**) 128-140 ka, (**o**) 126-140 ka, and (**p**) 124-140 ka.

Extended Data Fig. 5. Predicted topography for the area covered by the Scandinavian Ice Sheet at 131 ka. The calculation is based on the LAM ice history (see text) and an Earth model characterized by a lithosphere of thickness 100 km, upper mantle viscosity of 3×10^{20} Pa s, and lower mantle viscosity of 5×10^{22} Pa s. The white zone in (**a**) represents coverage of grounded ice extent at this time and the dashed white line on this frame is the shoreline location. Frame (**b**) is identical to (**a**), except the area of ice coverage is removed**.** It is clear from frame (**a**) that all but the southeast section of the perimeter of the Scandinavian ice sheet is predicted to be marine based at this time, and from frame (**b**) that much of the interior of the ice sheet was also marine based.

Extended Data Fig. 6. Results of sensitivity tests to oceanic forcing of the Greenland and Antarctic ice sheets. (a) Response of Greenland Ice Sheet to atmospheric forcing from CCSM3 with fixed ocean temperatures for the Penultimate Glacial Maximum (PGM) (blue line) and for the Last Interglaciation (LIG) (orange line) compared to ice-sheet response to atmospheric and oceanic forcing (black line). Present interglacial ice volume shown by horizontal dashed line. **(b)** Response of Antarctic Ice Sheet to atmospheric forcing from CCSM3 with fixed ocean temperatures for the PGM (blue line) and for the LIG (orange line) compared to ice-sheet response to atmospheric and oceanic forcing (black line). Present interglacial ice volume shown by horizontal dashed line. **(c)** As in **a**, response of Greenland Ice Sheet to atmospheric forcing from CCSM3 with fixed ocean temperatures for the LIG (orange line), but the vertical scale (grounded ice volume) has been increased to better illustrate the response. The initial ice-sheet size used in this experiment (and the comparable one for Antarctica) was the LIG ice sheet, whereas the climate forcing used was for the penultimate deglaciation and the LIG, i.e., from colder-than-present to LIG climate, resulting in a small response to the atmospheric forcing, since the LIG ice-sheet size had already adjusted to the combined atmospheric and oceanic forcing, as shown by the black line in **a**.

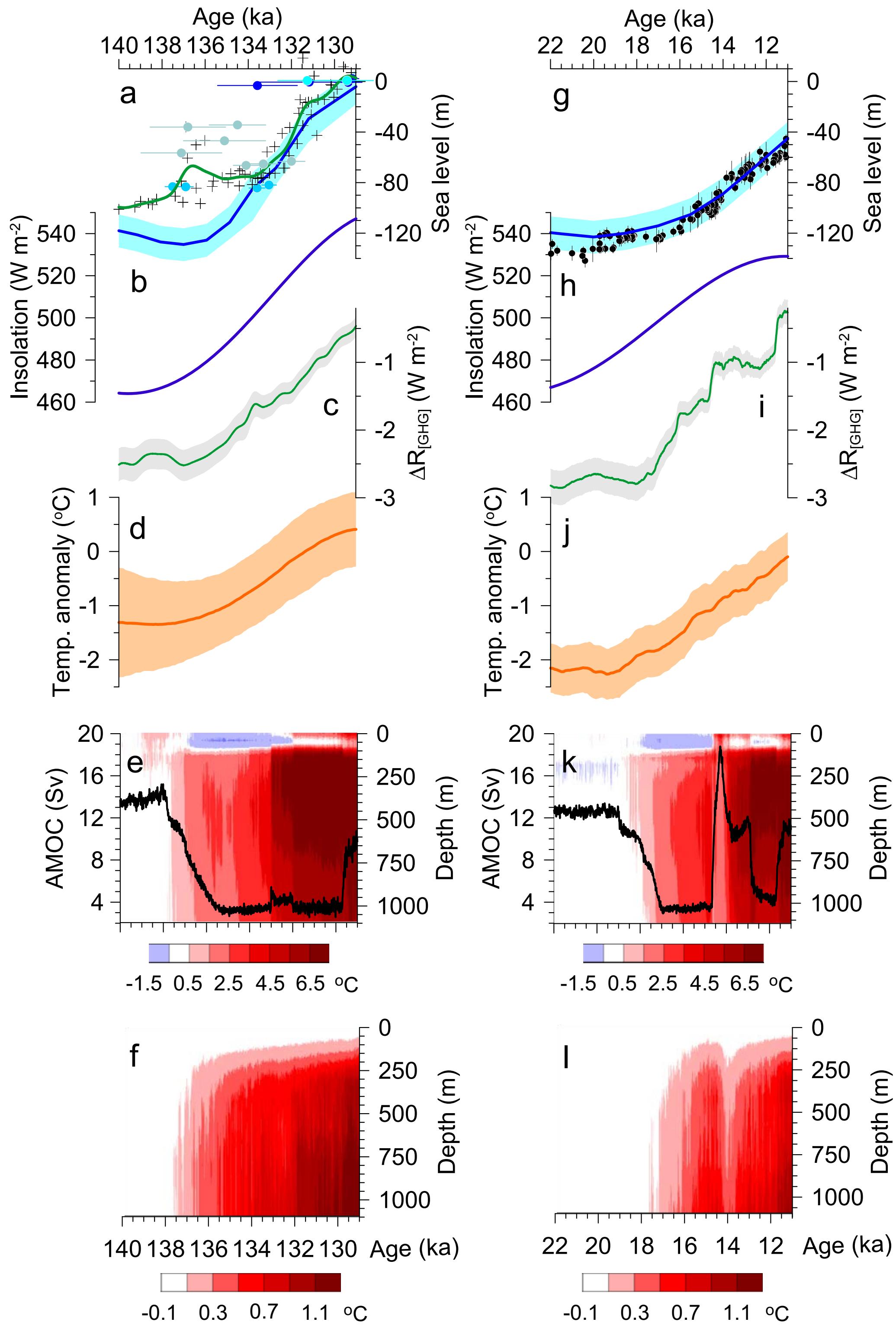
Extended Data Fig. 7. Predictions of relative sea level (RSL) at three far-field sites (the Seychelles, Western Australia, and Mallorca) and one intermediate-field site (Bahamas). a-

d, RSL predictions for the Bahamas from the full suite of simulations that bound from above all coral data with the exception of the earliest datum (at ~131 ka) for (**a**) the COL ice history, (**b**) the LAM ice history, (**c**) the HYB ice history, and (**d**) the WAE ice history. Age uncertainty is 2σ, and depth uncertainty reflects uncertainty in habitat depth. **e-h**, RSL predictions for the Seychelles from the full suite of simulations that lie above the three coral records with an elevation of ~4 m for (**e**) the COL ice history, (**f**) the LAM ice history, (**g**) the HYB ice history, and (**h**) the WAE ice history. Age uncertainty is 2σ, and depth uncertainty reflects uncertainty in habitat depth. **i-l**, RSL predictions for western Australia from the full suite of simulations that bound from above all coral data for (**i**) the COL ice history, (**j**) the LAM ice history, (**k**) the HYB ice history, and (**l**) the WAE ice history. Age uncertainty is 2σ, and depth uncertainty reflects uncertainty in habitat depth. **m-p**, RSL predictions for Mallorca from the full suite of simulations that fit the data within 50% of the minimum misfit achieved for all simulations for (**m**) the COL ice history, (**n**) the LAM ice history, (**o**) the HYB ice history, and (**p**) the WAE ice history. Age uncertainty is 2σ, and depth uncertainty reflects uncertainty in speleothem water depth.

Extended Data Fig. 8. Sensitivity of Greenland ice sheet model to melt parameterisation. (**a**) Time series of tuning experiments for the Greenland Ice Sheet with the preferred run in blue and three runs used for **b-d** shown in green, orange, and red. **b-d**. Surface elevation differences under a present-day climatology at the end of the 40,000-year T-I parameter tuning experiments, using degree-day factors drawn from our ensemble that (**b**), give a low amount of surface melting, (**c**), medium amount of surface melting, and (**d**), high amount of surface melting. Values shown are differences from the reference experiment. These experiments are identical to the T-I reference experiment used to parameterise the T-II simulations (Fig. 3) except for the degree-day factors used. The results show that our ice-sheet model is sensitive to the way in which surface mass balance is parameterised by controlling the amount of surface melting.

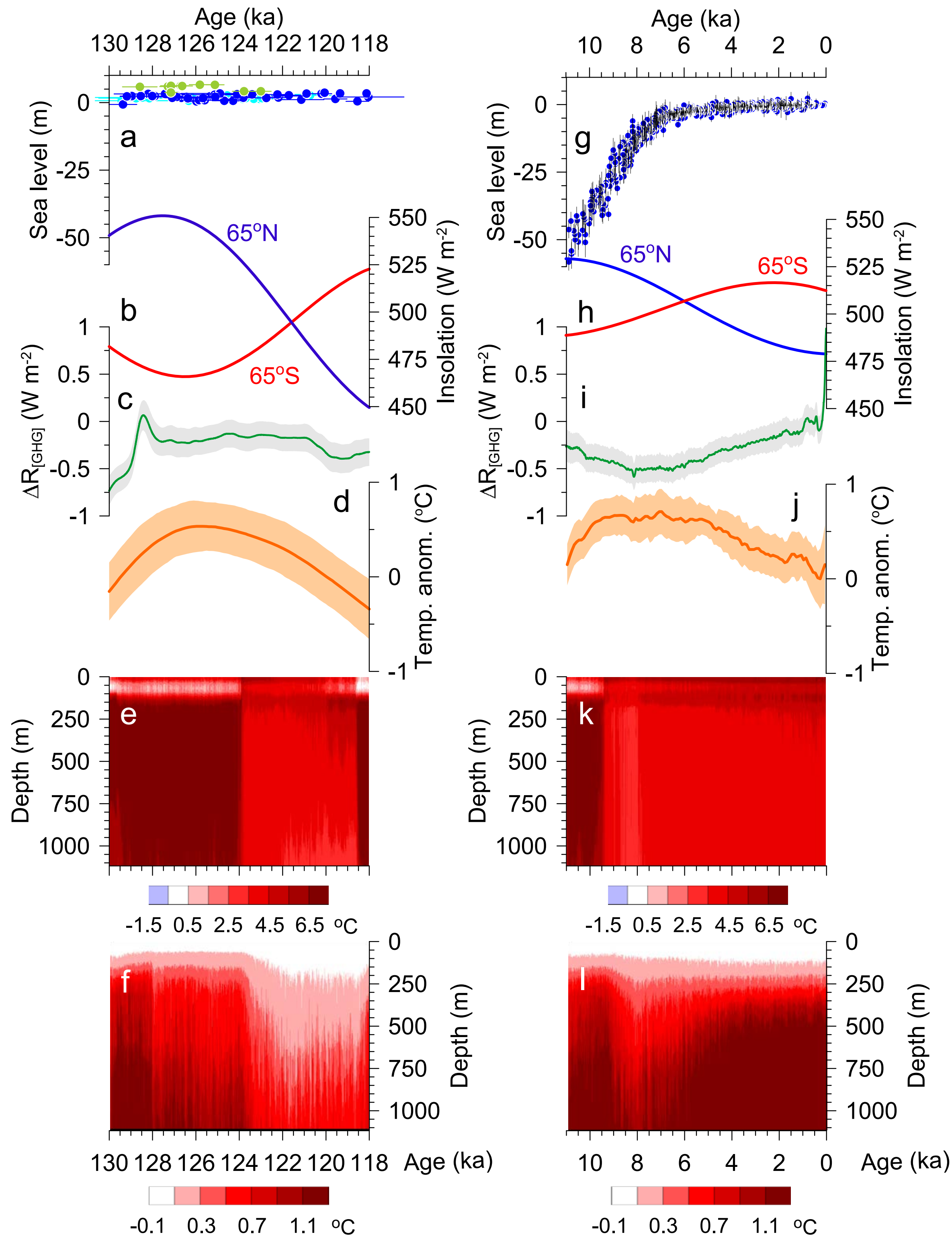
Extended Data Fig. 9. Simulated ice-volume changes and components of the mass balance for the Greenland Ice Sheet. (**a**) Simulated changes in ice volume for T-I. (**b**) Simulated changes in mass-balance components for T-I. (**c**) Simulated changes in ice volume for T-II. (**d**) Simulated changes in mass-balance components for T-II. (**e**) Modelled surface mass balance anomaly during the Last Interglaciation (129-120 ka) with respect to modelled present day.

Extended Data Fig. 10. Comparison of our simulated summer temperature for Greenland ice-core sites with the temperature reconstructions for these sites based on $\delta^{18}O_{ice}$. (a) The simulated summer temperature (JJA) (grey line) and lapse-rate corrected JJA temperature (green line) compared to reconstructed temperatures for the GISP2 ice-core site (blue symbols, 1 sigma uncertainty)[66,73] based on the relation $d\delta^{18}O_{ice}/dT = \sim 0.5‰ \, C^{-1}$ which is derived from Greenland ice-core sites elsewhere[74]. Also shown are the reconstructed temperatures using the $d\delta^{18}O_{ice}/dT$ relation established for the NEEM site ($\sim 1.1‰ \, C^{-1}$)[75] (red symbols, 1 sigma uncertainty), suggesting that the GISP2 LIG summer temperatures are about half of the originally published values based on the Vinther et al.[74] $d\delta^{18}O_{ice}/dT$ relation and in good agreement with our model results. **(b)** The simulated JJA temperature (grey line) and lapse-rate corrected JJA temperature (green line) compared to reconstructed temperatures for the NEEM ice-core site (dark blue line, gray shading is uncertainty)[65] based on the relation $d\delta^{18}O_{ice}/dT = \sim 0.5‰ \, C^{-1}$ which is derived from Greenland ice-core sites elsewhere[74]. Also shown are the reconstructed temperatures using the $d\delta^{18}O_{ice}/dT$ relation established for the NEEM site ($\sim 1.1‰ \, C^{-1}$)[75] (red line, pink shading is uncertainty), suggesting that the NEEM LIG summer temperatures are about half of the originally published values based on the Vinther et al.[74] $d\delta^{18}O_{ice}/dT$ relation and in good agreement with our model results. These reconstructions span the interval 127-120 ka, which is the warmest interval in the ice-core records for the LIG suggested by this proxy.



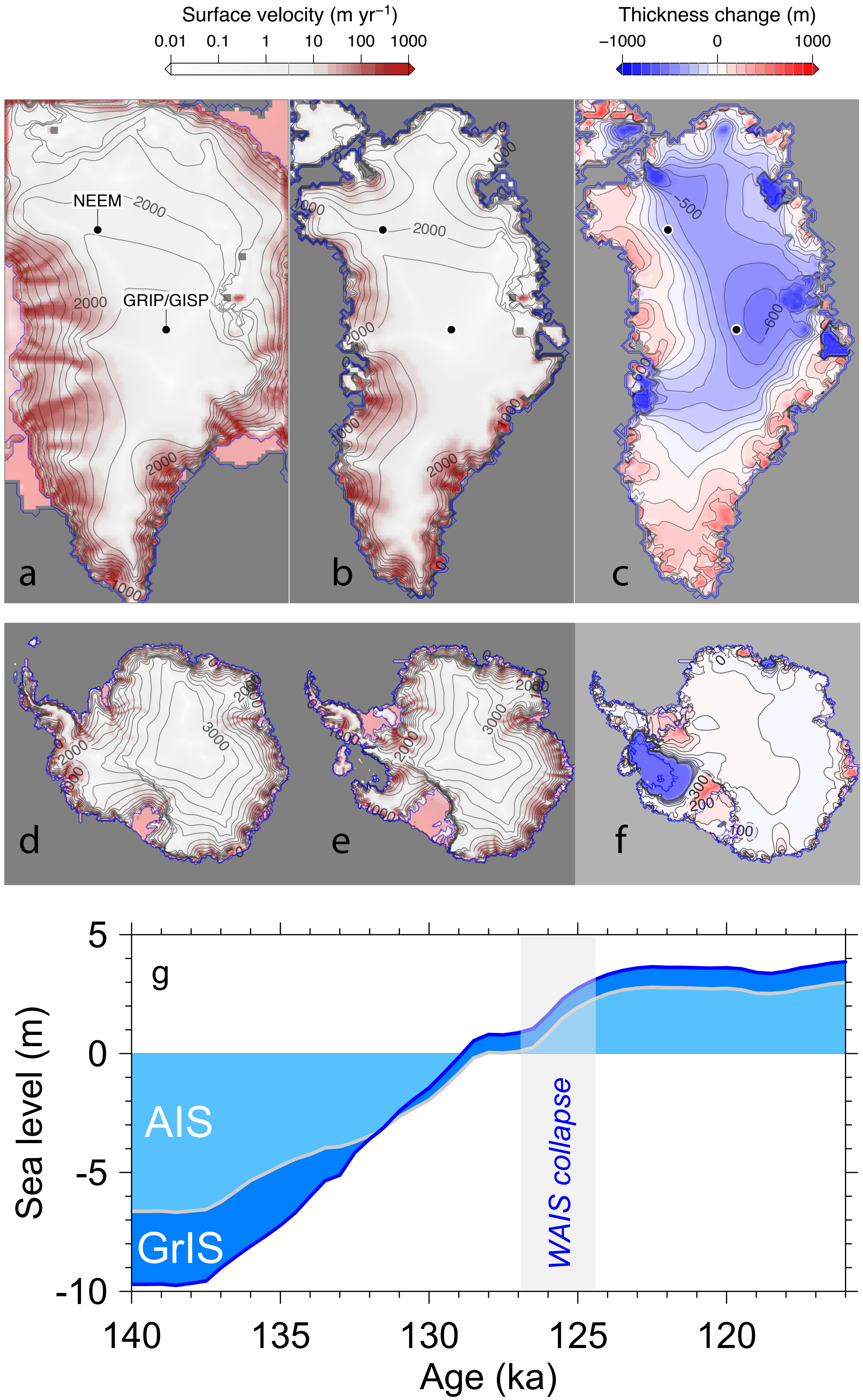
Age (ka)
140 138 136 134 132 130
a
b
c
d
e
f
Sea level (m)
0
-40
-80
-120
Insolation (W m-2)
540
520
500
480
460
ΔR[GHG] (W m-2)
-1
-2
-3
Temp. anomaly (oC)
1
0
-1
-2
AMOC (Sv)
20
16
12
8
4
Depth (m)
0
250
500
750
1000
-1.5 0.5 2.5 4.5 6.5 oC
-0.1 0.3 0.7 1.1 oC
Age (ka)
22 20 18 16 14 12
g
h
i
j
k
l

Figure 1



Age (ka)
130 128 126 124 122 120 118
Sea level (m)
a
65°N
b
65°S
Insolation (W m-2)
ΔR[GHG] (W m-2)
c
d
Temp. anom. (°C)
Depth (m)
e
-1.5 0.5 2.5 4.5 6.5 °C
f
Age (ka)
-0.1 0.3 0.7 1.1 °C
10 8 6 4 2 0
g
h
i
j
k
l

Figure 2



Surface velocity (m yr^{-1})
0.01 0.1 1 10 100 1000
Thickness change (m)
−1000 0 1000
NEEM
GRIP/GISP
2000
1000
-500
-600
3000
300
200
100
a
b
c
d
e
f
g
AIS
GrIS
WAIS collapse
Sea level (m)
Age (ka)
5
0
-5
-10
140
135
130
125
120

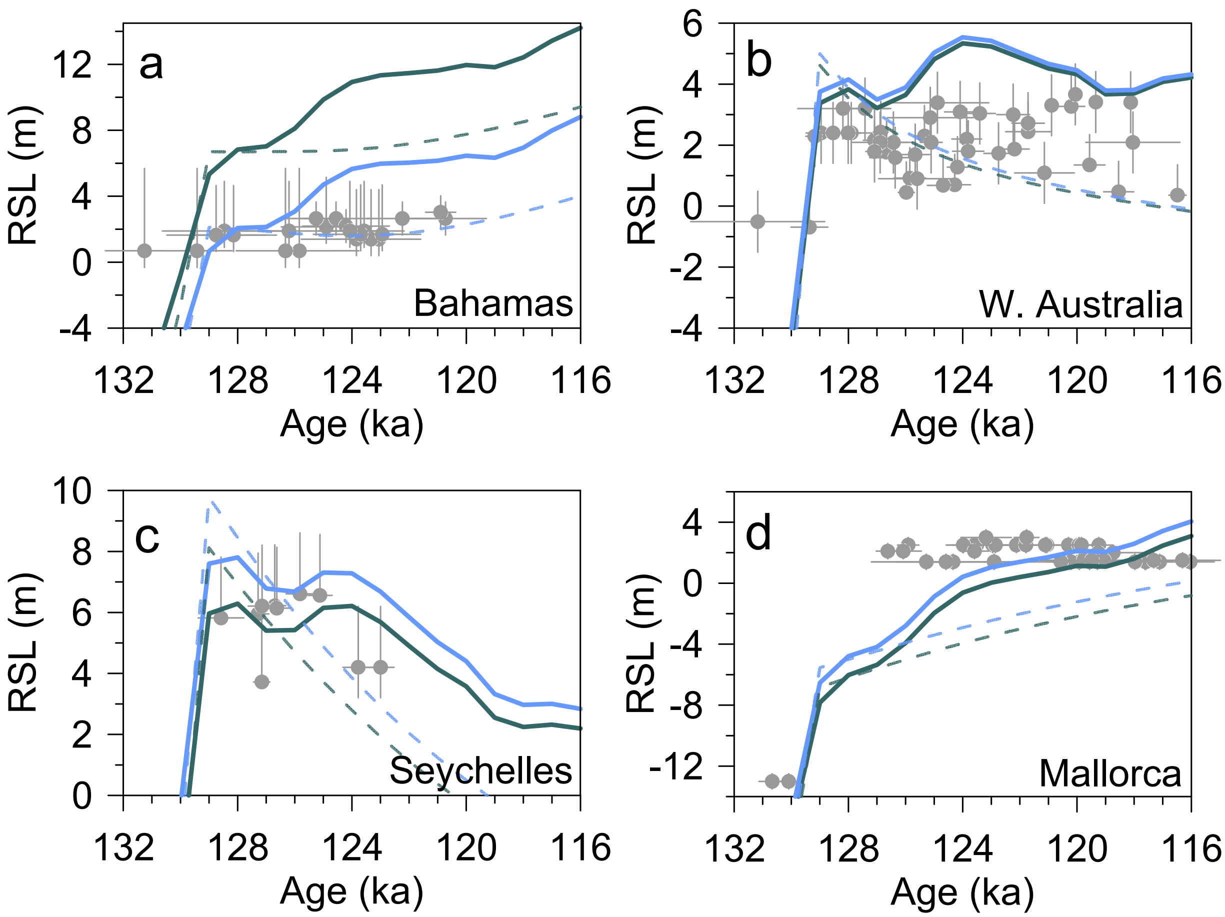
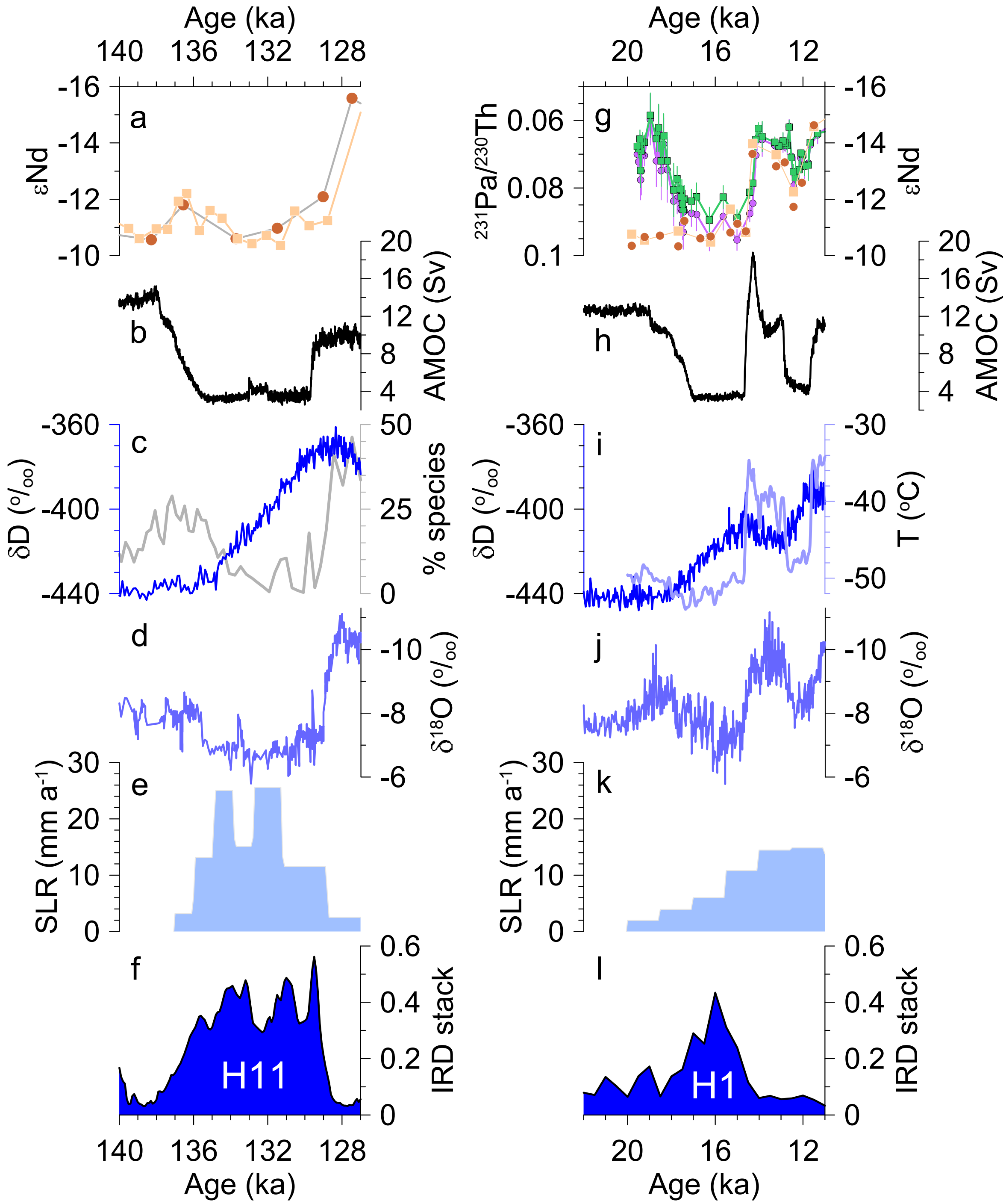
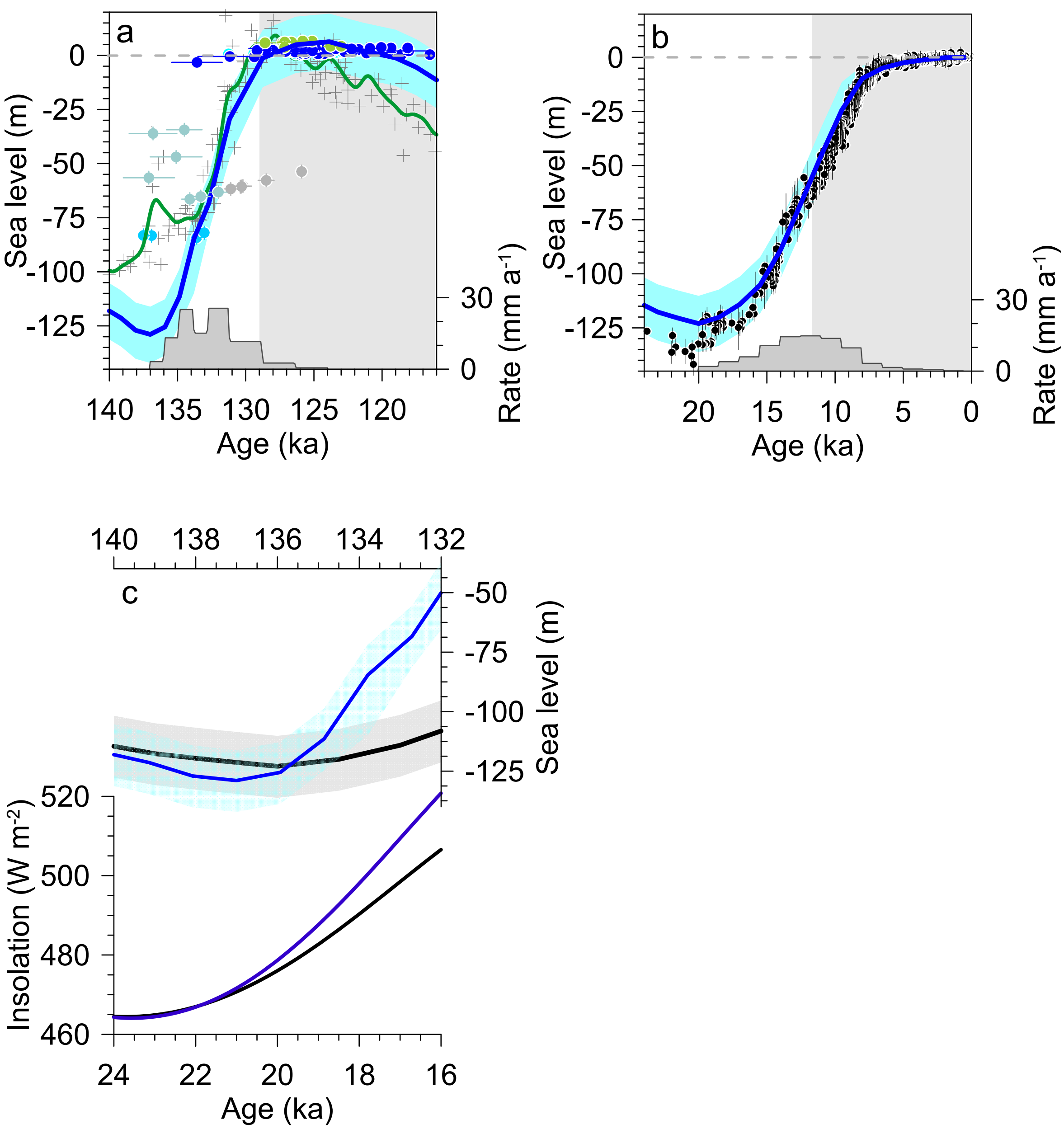

a
Bahamas
b
W. Australia
c
Seychelles
d
Mallorca
RSL (m)
Age (ka)
132
128
124
120
116
12
8
4
0
-4
6
2
-2
10
-8
-12

Figure 4



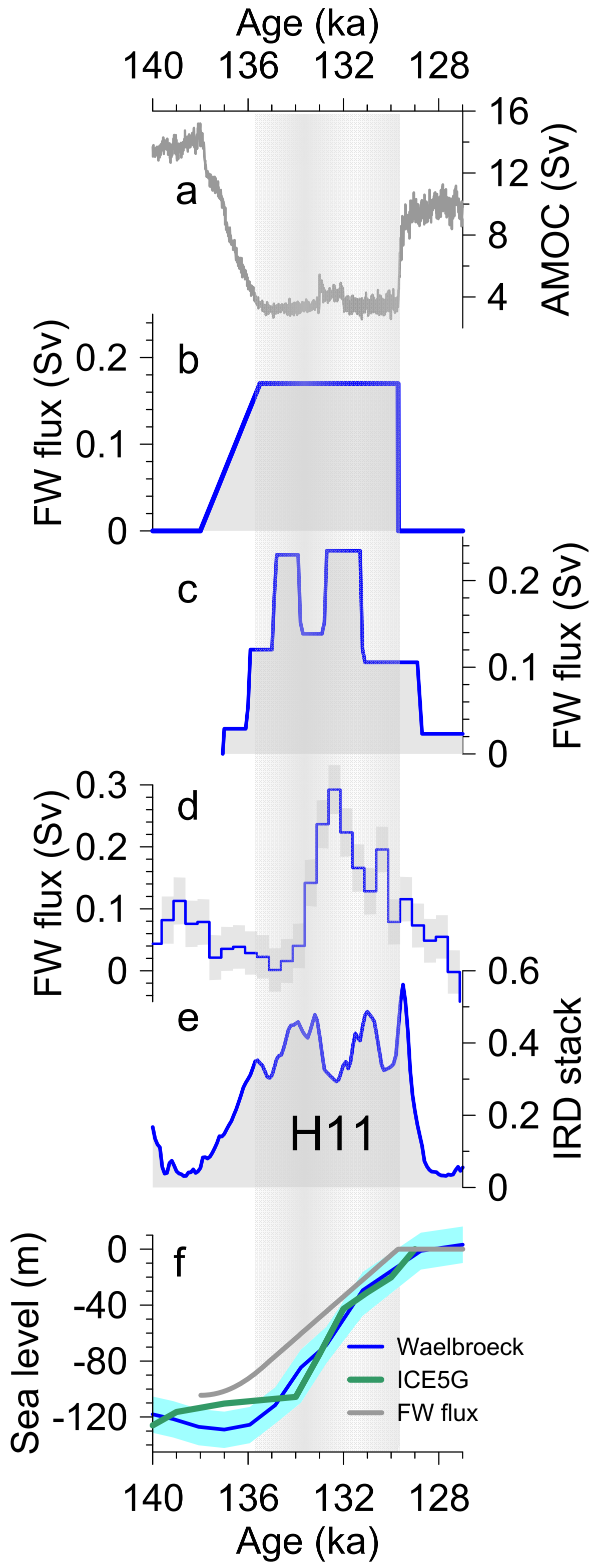
Age (ka)
140 136 132 128
a
εNd
-16 -14 -12 -10
b
AMOC (Sv)
20 16 12 8 4
c
δD (‰)
-360 -400 -440
% species
50 25 0
d
δ18O (‰)
-10 -8 -6
e
SLR (mm a-1)
30 20 10 0
f
H11
IRD stack
0.6 0.4 0.2 0
Age (ka)
20 16 12
g
231Pa/230Th
0.06 0.08 0.1
h
i
T (°C)
-30 -40 -50
j
k
l
H1

ED Fig. 1

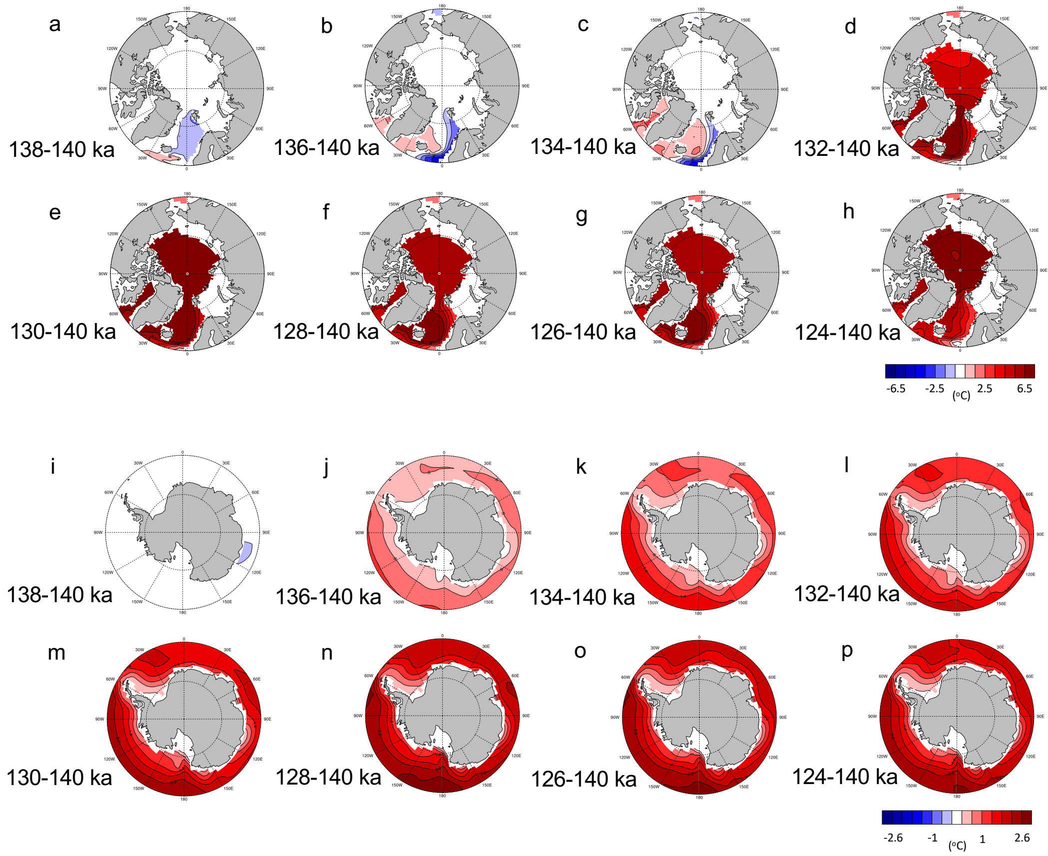


a
Sea level (m)
0
-25
-50
-75
-100
-125
30
0
Rate (mm a-1)
140 135 130 125 120
Age (ka)
b
Sea level (m)
0
-25
-50
-75
-100
-125
30
0
Rate (mm a-1)
20 15 10 5 0
Age (ka)
140 138 136 134 132
c
-50
-75
-100
-125
Sea level (m)
520
500
480
460
Insolation (W m-2)
24 22 20 18 16
Age (ka)

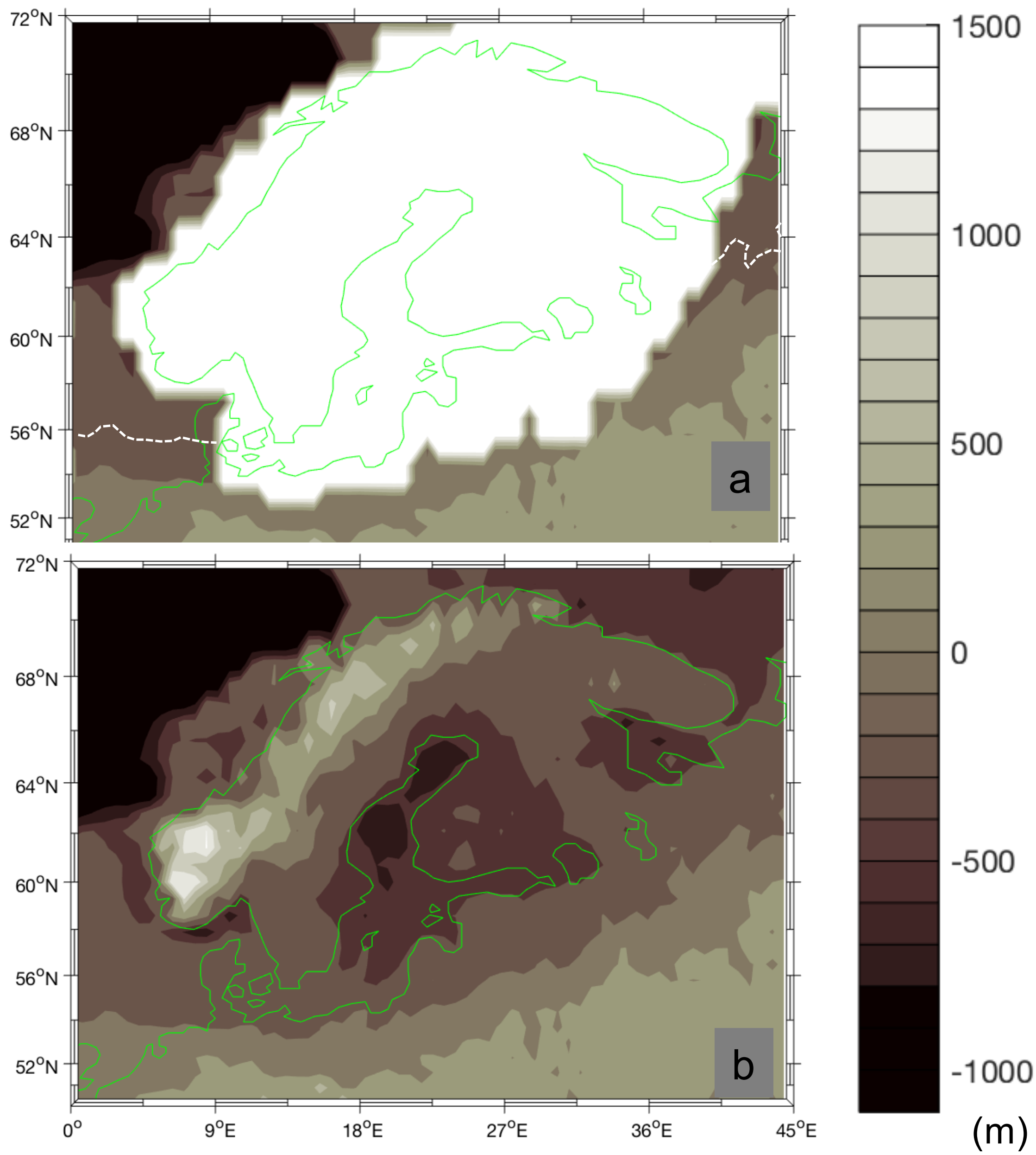
ED Fig. 2


Age (ka)
140
136
132
128
16
12
8
4
AMOC (Sv)
a
b
0.2
0.1
0
FW flux (Sv)
c
0.2
0.1
0
FW flux (Sv)
d
0.3
0.2
0.1
0
FW flux (Sv)
e
0.6
0.4
0.2
0
IRD stack
H11
f
0
-40
-80
-120
Sea level (m)
Waelbroeck
ICE5G
FW flux
140
136
132
128
Age (ka)

ED Fig. 3

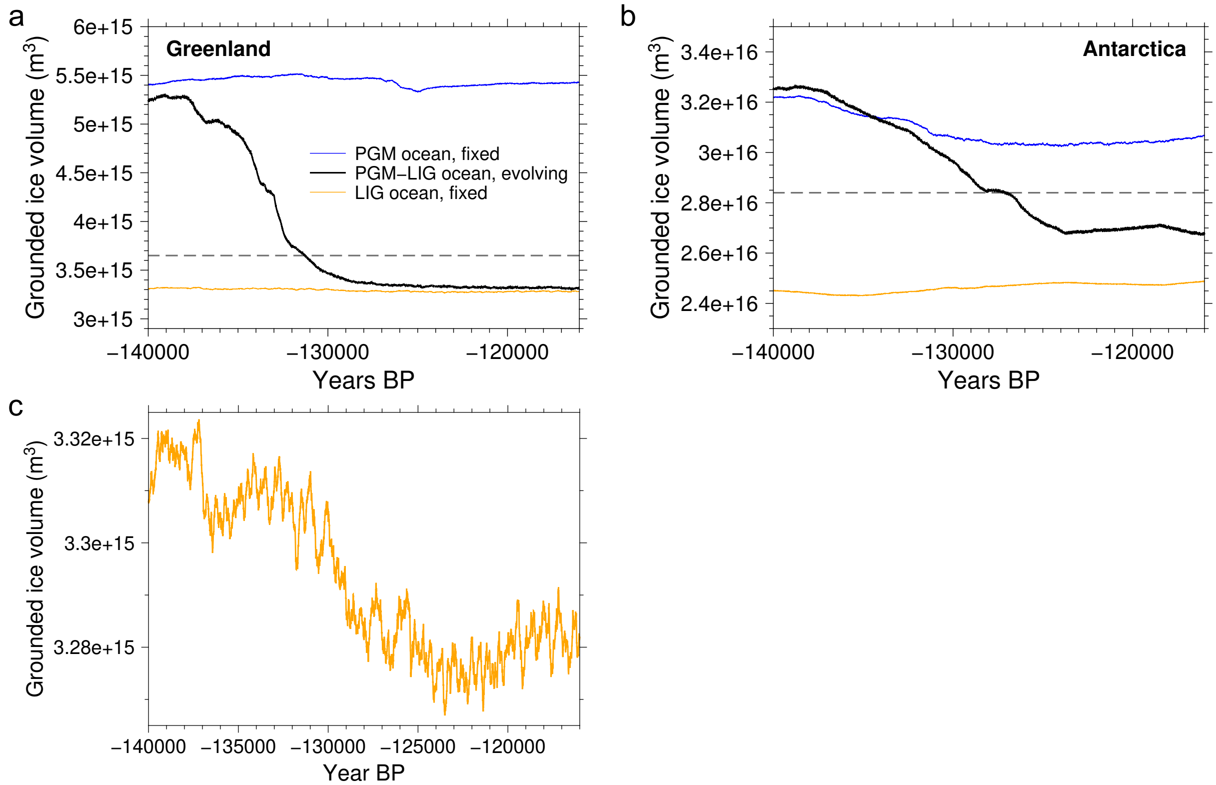

a
138-140 ka
b
136-140 ka
c
134-140 ka
d
132-140 ka
e
130-140 ka
f
128-140 ka
g
126-140 ka
h
124-140 ka
-6.5
-2.5
(°C)
2.5
6.5
i
138-140 ka
j
136-140 ka
k
134-140 ka
l
132-140 ka
m
130-140 ka
n
128-140 ka
o
126-140 ka
p
124-140 ka
-2.6
-1
(°C)
1
2.6

ED Fig. 4



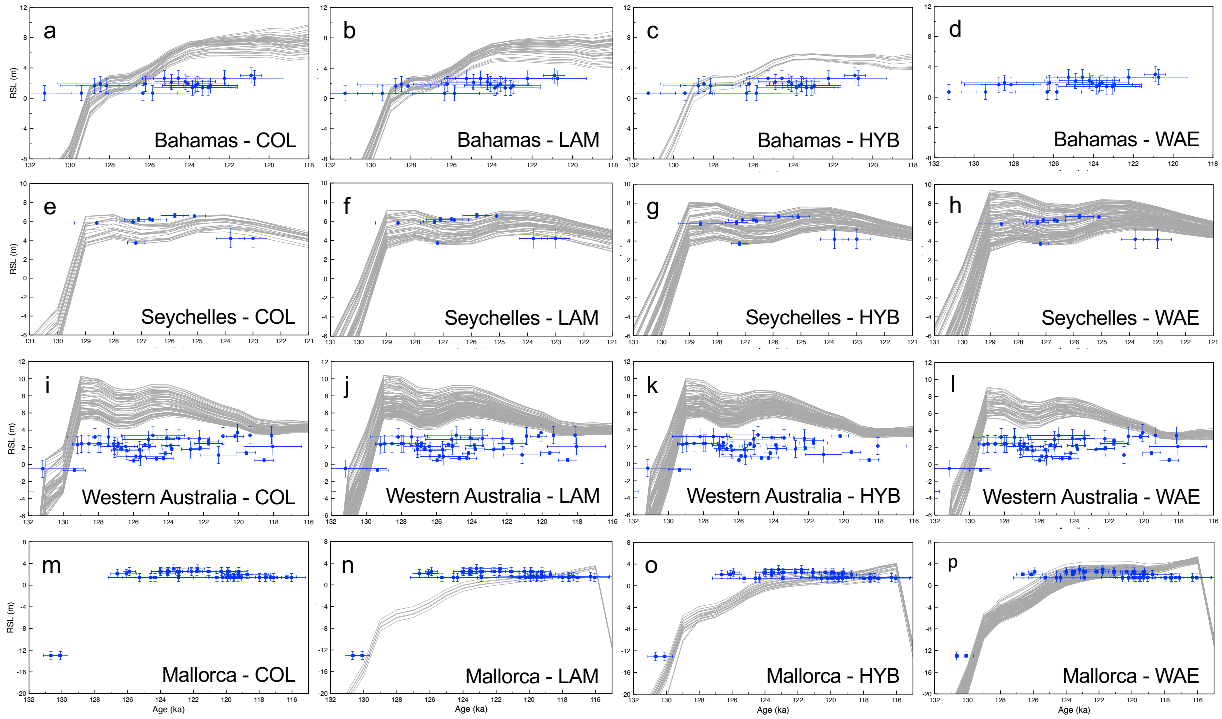
72°N
68°N
64°N
60°N
56°N
52°N
a
b
0°
9°E
18°E
27°E
36°E
45°E
1500
1000
500
0
-500
-1000
(m)

ED Fig. 5

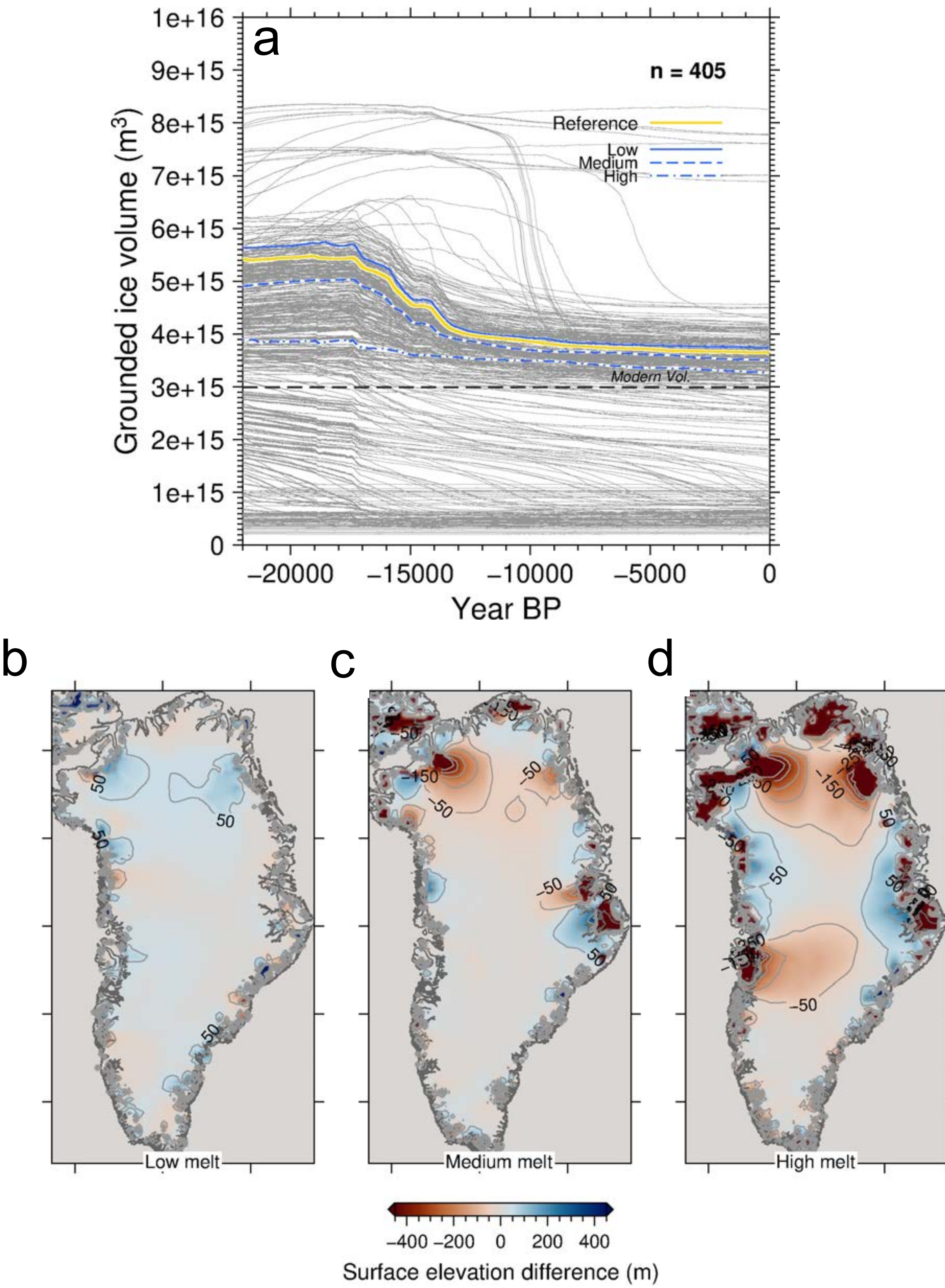


a
Greenland
Grounded ice volume (m3)
6e+15
5.5e+15
5e+15
4.5e+15
4e+15
3.5e+15
3e+15
PGM ocean, fixed
PGM–LIG ocean, evolving
LIG ocean, fixed
–140000
–130000
–120000
Years BP
b
Antarctica
Grounded ice volume (m3)
3.4e+16
3.2e+16
3e+16
2.8e+16
2.6e+16
2.4e+16
–140000
–130000
–120000
Years BP
c
Grounded ice volume (m3)
3.32e+15
3.3e+15
3.28e+15
–140000
–135000
–130000
–125000
–120000
Year BP

ED Fig. 6

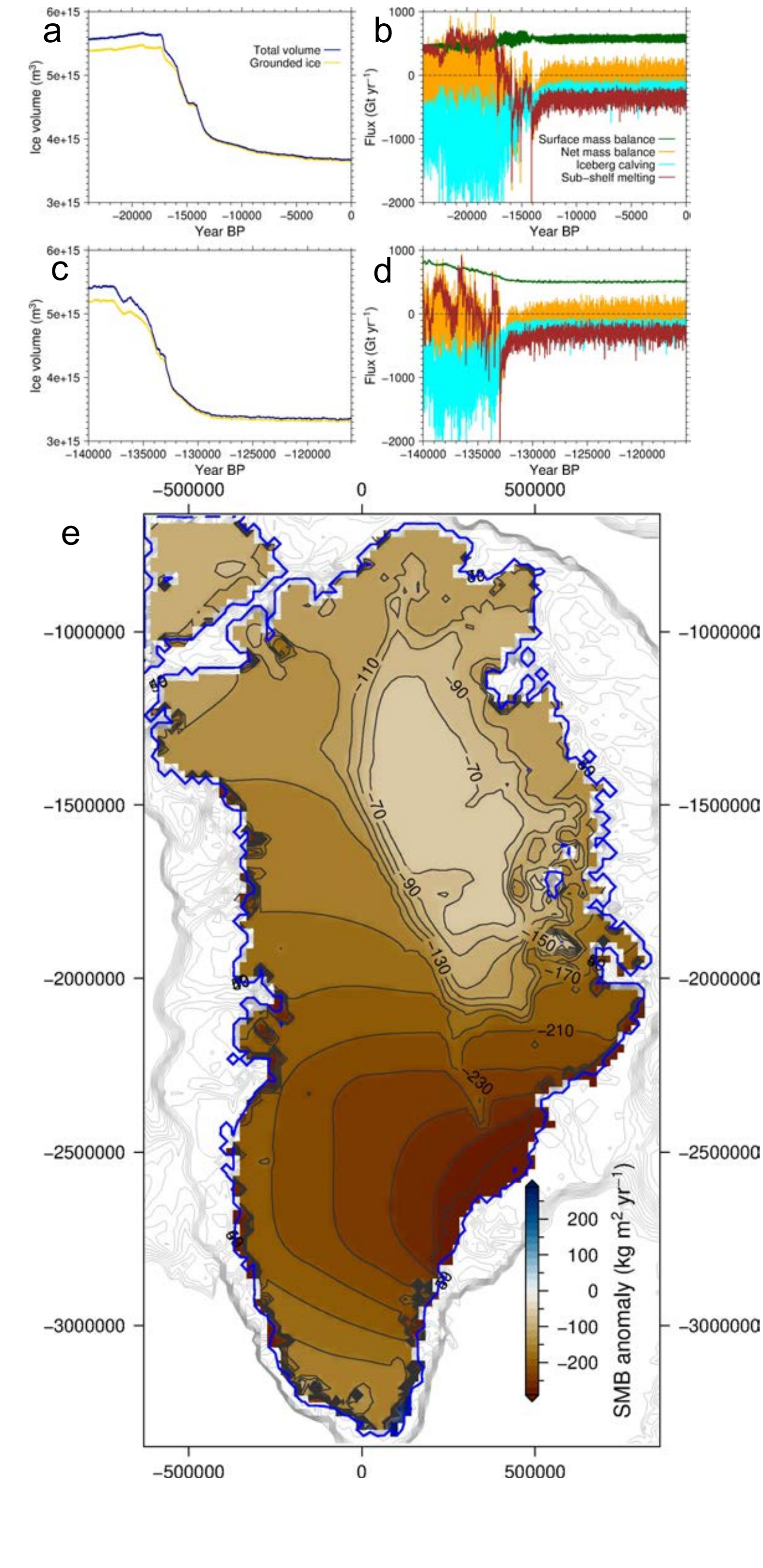

a
Bahamas - COL
b
Bahamas - LAM
c
Bahamas - HYB
d
Bahamas - WAE
e
Seychelles - COL
f
Seychelles - LAM
g
Seychelles - HYB
h
Seychelles - WAE
i
Western Australia - COL
j
Western Australia - LAM
k
Western Australia - HYB
l
Western Australia - WAE
m
Mallorca - COL
n
Mallorca - LAM
o
Mallorca - HYB
p
Mallorca - WAE
RSL (m)
Age (ka)

ED Fig. 7



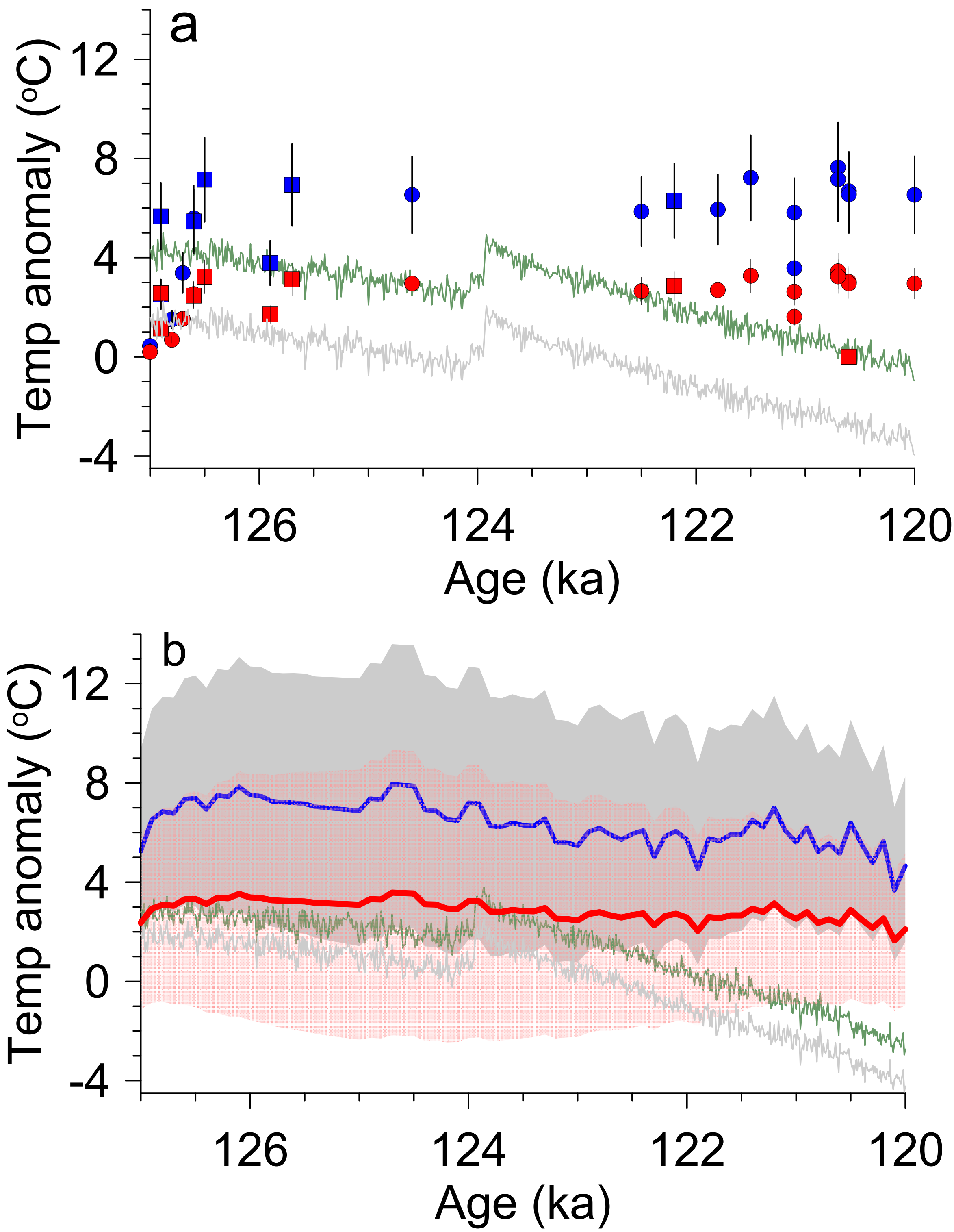
a
n = 405
Reference
Low
Medium
High
Modern Vol.
Grounded ice volume (m^3)
Year BP
b
c
d
Low melt
Medium melt
High melt
−400 −200 0 200 400
Surface elevation difference (m)

ED Fig. 8



a
b
c
d
e
Total volume
Grounded ice
Ice volume (m^3)
Year BP
Flux (Gt yr^{-1})
Surface mass balance
Net mass balance
Iceberg calving
Sub-shelf melting
SMB anomaly (kg m^2 yr^{-1})

ED Fig. 9


a
Temp anomaly (ºC)
12
8
4
0
-4
126
124
122
120
Age (ka)
b
Temp anomaly (ºC)
12
8
4
0
-4
126
124
122
120
Age (ka)

ED Fig. 10