Substantial contribution to sea-level rise during the last interglacial from the Greenland ice sheet

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During the last interglacial period (the Eemian), global sea level was at least three metres, and probably more than five metres, higher than at present[1,2]. Complete melting of either the West Antarctic ice sheet or the Greenland ice sheet would today raise sea levels by 6–7 metres. But the high sea levels during the last interglacial period have been proposed to result mainly from disintegration of the West Antarctic ice sheet[3], with model studies attributing only 1–2 m of sea-level rise to meltwater from Greenland[4,5]. This result was considered consistent with ice core evidence[4], although earlier work had suggested a much reduced Greenland ice sheet during the last interglacial period[6]. Here we reconsider the Eemian evolution of the Greenland ice sheet by combining numerical modelling with insights obtained from recent central Greenland ice-core analyses. Our results suggest that the Greenland ice sheet was considerably smaller and steeper during the Eemian, and plausibly contributed 4–5.5 m to the sea-level highstand during that period. We conclude that the high sea level during the last interglacial period most probably included a large contribution from Greenland meltwater and therefore should not be interpreted as evidence for a significant reduction of the West Antarctic ice sheet.

The substantial impact that a 5-m sea-level rise would have on coastal regions provides strong motivation for understanding the origins of the Eemian historical highstand. The peculiar dynamics[7,8] of the West Antarctic ice sheet (WAIS) (flow dominated by basal processes and grounding below sea level) may have, in theory, enabled it to rapidly disintegrate in response to a minor and transient sea-level forcing[3]. In contrast, it is accepted that complete melting of the Greenland ice sheet (GIS) would have required a sustained climatic forcing[9]. A substantial portion of the mass loss from the GIS occurs by surface melting in the low-elevation zone around its margin, and a modest climatic warming could rapidly expand this zone of net melt across the gently sloping ice-sheet surface. The GIS is thus recognized to be moderately vulnerable to climatic warming, and the warming necessary to sustain destruction of the entire ice sheet is estimated to be 5–8 °C (ref. 10).

Here we argue that the current 1–2 m estimate[4,5] for the Eemian GIS sea-level contribution is too small and that a contribution more than twice this large is plausible. The difference results primarily from a quantitative re-evaluation of the isotopic palaeothermometer ($\delta^{18}O$) used to construct the climate histories driving the evolution of the model ice sheets.

Generally, a $\delta^{18}O$ history measured from ice core samples is used to construct a climatic temperature history by assuming proportionality between deviations of mean annual temperature (ΔT) and isotopic content ($\Delta\delta^{18}O$) from modern values, so that $\Delta\delta^{18}O = \alpha\Delta T$. The earlier modelling studies assumed, as was standard practice, that the sensitivity of isotopic content to temperature $\alpha = d\delta^{18}O/dT \approx 0.67$ per mil °C^{-1}, which is the modern value for α evaluated as a spatial derivative on the ice-sheet surface. Recently, palaeotemperature reconstructions based on temperature–depth distribution[11,12] and on gas-phase isotopic composition[13] have revealed that the temporal isotopic sensitivity (α evaluated at one location through time) is much smaller and was, for example, approximately 0.4 per mil °C^{-1} for the last glacial–interglacial transition in central Greenland, when corrected for changes in ocean water composition. Several reasons for the lower sensitivity have been recognized, including changes in the seasonal timing of precipitation[14] and covariation of low- and high-latitude temperature changes[15].

These factors are also expected to cause low α for millennial-scale climate changes within interglacial periods. For example, changes in seasonal timing of precipitation are modelled to occur in correlation with temperature changes in the Holocene epoch, because both

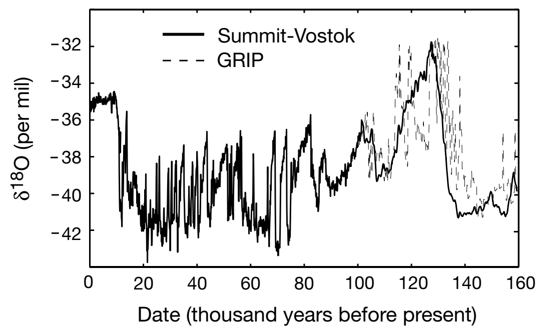

Figure 1 Assumed isotopic history. The solid line is our 'standard case' isotopic history used as model forcing. It was fabricated for use in the present study and should not be mistaken for a Greenland climate reconstruction. For comparison, the dashed line shows the GRIP record[22] which has been used in recent modelling studies[5], and which we do not use because the chronology is known to be disturbed by folding of the ice[23].

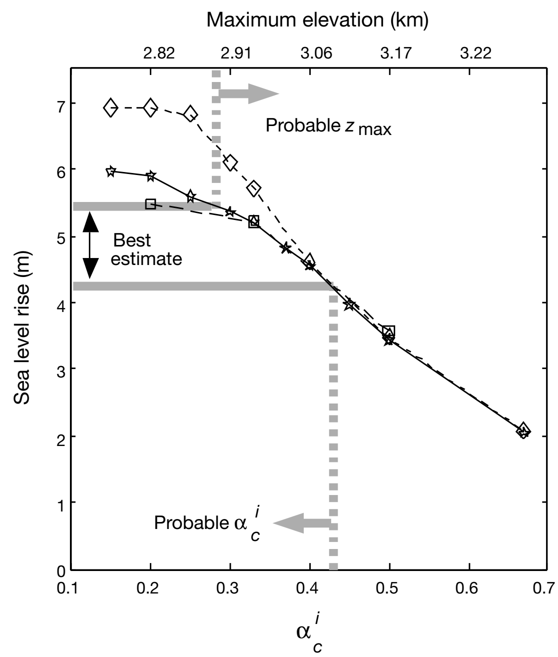

Figure 2 Calculated Eemian sea-level rise due to Greenland melting. Black stars show the maximum Eemian sea-level rise, as a function of the isotopic sensitivity α_c^i, using the 'standard case' isotopic forcing. The corresponding summit elevations are shown on the top axis. Diamonds show results of the same calculations, but with the isotopic lapse rate correction (see Methods) removed, to illustrate its importance. Squares represent calculations applying $\alpha_c = 0.33$ to all the pre-Eemian forcing with $\delta \leqslant -35$ per mil (which implies that the penultimate glacial climate was very similar to that of the last glacial). The black stars were calculated with $\alpha_c = \alpha_c^i$, which is physically less plausible but which allows direct comparison to earlier modelling work in the limit $\alpha_c^i = 0.67$, and gives a minimum lower bound for the 'best estimate' sea-level contribution.

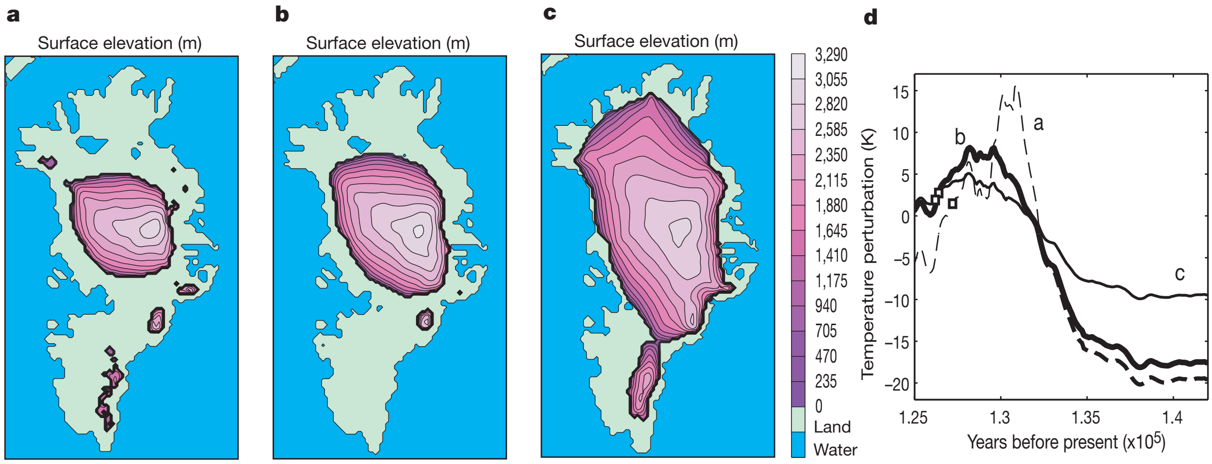

Figure 3 Calculated surface elevation maps for the Greenland ice sheet at its Eemian minimum. **a**–**c**, Maps of three different values for α_c^i: $\alpha_c^i = 0.2$ (with pre-Eemian $\alpha_c = 0.33$) (**a**), $\alpha_c^i = 0.4$ (with pre-Eemian $\alpha_c = 0.4$) (**b**), $\alpha_c^i = 0.67$ (with pre-Eemian $\alpha_c = 0.67$) (**c**). Map **b** is preferred. Map **c** is entirely consistent with previous modelling efforts[4,5]. **d**, The constant-elevation climatic temperature perturbations (relative to modern temperature) generated by these three model calculations, with the squares indicating time of minimum ice sheet volume: thick line (map **b**), thin line (map **c**), dashed lines (map **a**). The thick portion of the dashed line is also the pre-Eemian temperature history used in the three calculations shown by squares in Fig. 2.

depend on the strength of the North Atlantic thermohaline circulation[16], which itself responds to variations of the Atlantic basin moisture balance[17]. Thus an unlikely constancy of numerous climate variables would be required in order to have α equal to its modern spatial gradient.

Indeed, a low sensitivity, α_c^i, for millennial-scale climate changes within interglacial periods is implied by Holocene climate history, as recorded in borehole temperatures and the Greenland cores' isotopic records. Central Greenland has cooled over the last 5 kyr, by 1.5–2.5 °C according to several independent borehole temperature analyses[12,11,18] and this cooling is recorded as a 0.375 per mil drop in average $\delta^{18}O$ (refs 19, 20). Thus the observed α has been $\alpha_{obs} = \Delta\delta_{obs}/\Delta T_{obs} \approx 0.15$ to 0.25 per mil °C^{-1}. This indicates α_c^i is in the range 0.15 to 0.43 after accounting for possible elevation changes (see Methods), and therefore is similar to α for the glacial–interglacial transition, or possibly lower.

Such low values for α imply that the Eemian climate in Greenland was warmer than previously thought, and the consequent melting of the GIS more significant. To investigate this possibility, we model the evolution of the GIS in the changing climate of the last 160,000 years, using a three-dimensional coupled-ice-and-heat-flow model[21], developed using ref. 9. The model includes elevation changes due to isostatic response of the crust, and has a horizontal grid of resolution 20 km by 20 km.

The climatic forcing for the model consists of prescriptions for temperature, accumulation rate and ablation rate at all grid points on the ice-sheet surface, as functions of time. These forcings are all governed by a history of temperature change at constant elevation $\Delta T_c(t)$ (relative to modern), using standard relationships (see Methods) incorporating surface elevation and slope, and latitude–longitude. The ΔT_c spans 160 kyr, representing the cold climate of the penultimate glacial period, the warming to the Eemian interglacial (110–130 kyr ago), and the return to glacial conditions followed by warming to the Holocene. The model is initialized by first calculating a steady-state ice sheet for modern climate, and then imposing 200 kyr of constant climate 6 °C colder than modern. Memory of this initial state is largely erased within the first 15 kyr of the model calculation; results for the Eemian are insensitive to plausible changes in the initialization temperature.

Our interest here is the Eemian fate of the ice sheet, but we run the model for another 100 kyr to the present to show that the modern GIS is accurately reproduced. The model ice sheet accurately reproduces all gross features of GIS geography, including the volume (to within 1%), and the summit location (to within 50 km in the horizontal and 40 m in the vertical). For the most recent 98 kyr we use the ΔT_c inferred from the borehole temperature calibration[11,18] of the GISP2 project ice core $\delta^{18}O$ record[19].

Earlier forcings must be treated differently. There are at present no reliable palaeoclimate histories for the Eemian or earlier periods from any of the Northern Hemisphere ice caps. We circumvent this difficulty by fabricating a plausible climate history (meaning a history of isotopic content) and performing appropriate sensitivity tests. The crucial constraint on the history is the maximum $\delta^{18}O$ value for Eemian ice in the Summit cores, which is heavier than modern accumulation by an amount $\Delta\delta_E \approx 3.3$ per mil (ref. 22). This value fixes the maximum Eemian $\delta^{18}O$ in our fabricated history. Although the chronology of Eemian ice in the Summit cores is uncertain, there is no reason to think the $\delta^{18}O$ composition has been altered, and the combined CH_4 and $\delta^{18}O$ of O_2 gas compositions show conclusively that this ice accumulated during the Eemian[23].

Furthermore, we know the approximate duration of the Eemian as a whole based on a globally significant variety of independent but mutually consistent climate proxy records, including the Vostok ice core deuterium-isotope history which reflects southern high-latitude climate[24], the global atmospheric methane content which reflects low-latitude climate[23], and the Devil's Hole isotopic record of mid-latitude climate[25]. Most directly relevant to Greenland climate are proxy records for the deep Atlantic circulation, which show a similar duration, and a persistent interglacial maximum for approximately 10 kyr (ref. 26).

For a 'standard case' fabricated isotopic chronology before 98 kyr ago ($\Delta\delta^*$) we use the Vostok ice core deuterium history[24] as a template (see Methods and Fig. 1), which is scaled to the measured Greenland $\Delta\delta_E$ maximum. The $\Delta\delta^*$ yields climatic temperature changes (Methods) using the constant α_c^i. Using the constraints $\alpha_c^i = 0.15$–0.43 per mil °C^{-1} and $\Delta\delta_E = 3.3$ per mil, we investigated the extent to which the GIS shrank during the last interglacial. For all values $\alpha_c^i < 0.45$, we find that the GIS was substantially smaller than at present, and contributed more than 4 m to sea-level rise (Figs 2 and 3).

The volume reduction of the model ice sheet occurs rapidly in response to the peak warmth of the climate history. Therefore the calculated maximum sea-level contribution does not decrease substantially if we reduce the duration of the fabricated Eemian climate by reasonable amounts (Fig. 4). In many other respects, the sea-level rise is not particularly sensitive to the imposed chronology and accumulation sensitivity (Fig. 4). This is due in part to the rapid response time (a few millennia) of the ice sheet to changes in margin position and accumulation rate, which are dominant controls on ice-sheet evolution at the timescale most relevant here. It is also due to the action of systematic buffers. Most important is the trade-off between changes of climatic temperature and those of elevation. The fixed constraint $\Delta\delta \leqslant 3.3$ means that if the ice-sheet summit elevation is very high (due to high accumulation rate, for example) then the climatic temperature (and hence melt rates at the margins) must also be very high. Likewise, as the summit elevation decreases in response to marginal melt, the climatic temperature must decrease too, reducing melt rates. The latter constrains the maximum sea-level contribution against increases in the duration of isotopically heavy precipitation (curves 5, 6 and 7 in Fig. 4).

Measurements of the total gas content of the GRIP core Eemian ice[27] provide a rough constraint on the upper limit for the sea level contribution. The highest gas contents of Eemian ice are only slightly higher than late Holocene values, indicating that the elevation of the Eemian ice dome was not more than several hundred metres lower in elevation than at present. High frequency variability of the total gas measurement is at least $0.01\,\mathrm{cm^3\,g^{-1}}$, which corresponds to a $\sim$350 m elevation uncertainty. In our model results the corresponding constraints are $\alpha_c^i > 0.25$ and a sea-level contribution less than 5.4 m (Fig. 2).

The preceding results indicate that a sea-level contribution greater than 4 m from the GIS should be adopted as a 'most likely' scenario. Important uncertainties exist, in particular due to our lack of information about changes in the spatial pattern of accumulation rate, and also our inability to predict a priori the isotopic sensitivity. Our answer is by no means definitive, and we cannot firmly exclude a much smaller GIS contribution.

Nevertheless, the plausibility of substantial GIS melt[6] implies that the Eemian highstand should not be invoked as evidence for major reduction of WAIS volume. Further, this Northern Hemisphere melt would provide a rather large sea-level forcing at the WAIS margins, and the Eemian history of the WAIS (unknown at present) should be viewed in this context. A more general conclusion is that the magnitude of the Eemian highstand is a poor measure of the behaviour of the ice sheets, because contributions from both hemispheres could be obscured by asynchronous changes, or by growth of the vast East Antarctic ice sheet. □

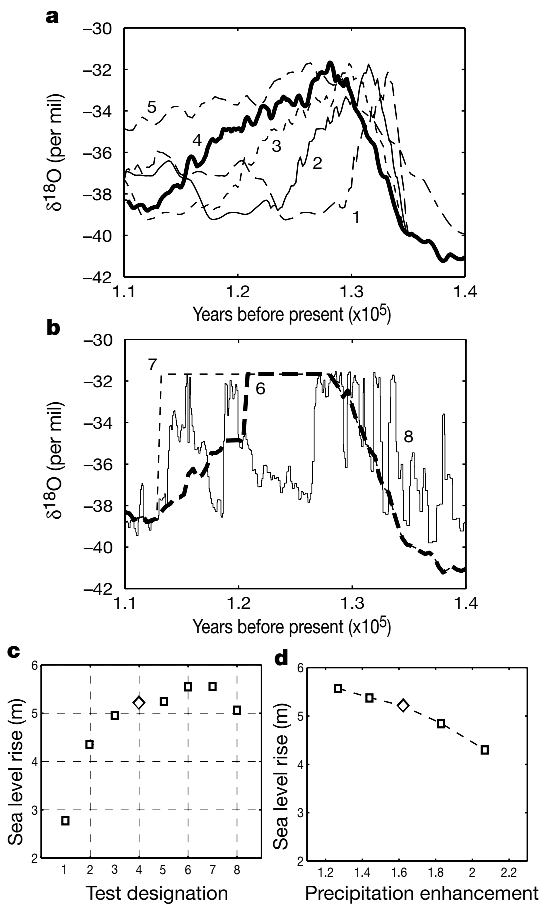

Figure 4 Sensitivity tests. Eight separate interglacial isotopic histories were used as model forcings, and calculations were performed with $\alpha_c^i = 0.33$. **a**, $\delta(t)$ for 'standard case' (Test 4) and compressed/stretched versions such that interglacial durations are $\frac{1}{4}$ (Test 1), $\frac{1}{2}$ (Test 2), $\frac{3}{4}$ (Test 3), and 2 (Test 5) times the standard case. **b**, $\delta(t)$ for a 7.5-kyr constant maximum (Test 6), a 15-kyr constant maximum (Test 7), and an unmodified GRIP forcing (Test 8). **c**, Corresponding sea-level maxima, with standard case (Test 4) indicated by a black diamond. **d**, Sensitivity to accumulation-rate parameter a, shown as sea-level maxima calculated for various values of a, which we have converted to the fractional precipitation increase due to a 7K warming. The diamond corresponds to the a value inferred from ice core studies for the glacial–interglacial transition. The average a within the Holocene interglacial climate has been markedly lower[18], suggesting that the lower precipitation enhancement values are more likely.

Methods

Inferring α_c^i

Temperature change at the ice core site ΔT_{obs} has components due to climatic change (ΔT_c) and elevation change (ΔZ) such that $\Delta T_{\mathrm{obs}} = \Delta T_c + \lambda_T \Delta Z$ where λ_T is the thermal lapse rate $-7.5\,^{\circ}\mathrm{C\,km^{-1}}$. The corresponding isotopic change is $\Delta\delta_{\mathrm{obs}} = \alpha_c \Delta T_c + \alpha_z \lambda_T \Delta Z$; in which the sensitivities α_c and α_z are not necessarily equal. The α_z value depends primarily on the evolution of single air masses[28] and therefore, unlike α_c, may be insensitive to temporal changes of precipitation seasonality and source region temperature. We therefore make the reasonable assumption (which minimizes GIS melt in our calculations) that $\alpha_z \lambda_T$ pertaining to temporal ΔZ equals the modern spatial gradient -6.2 per mil $\mathrm{km^{-1}}$ (ref. 28) rather than being reduced in correlation with α_{obs}, the α value observed through time at the ice core site. Thus we calculate

$$\alpha_c = \frac{\alpha_{\mathrm{obs}} - \alpha_z \lambda_T \Delta Z / \Delta T_{\mathrm{obs}}}{1 - \lambda_T \Delta Z / \Delta T_{\mathrm{obs}}} \qquad (1)$$

Elevation change of the Greenland summit over the last 5 kyr is not known precisely but model-based bounds are $\Delta Z = 0$ to -100 m (refs 4, 5, 18 and 27) which gives α_c^i in the range 0.15–0.43. This is lower than the 0.52 inferred for the end of the Little Ice Age[29], presumably because some elements of the climate system that change over millennia do not change during higher frequency climate-change events such as the Little Ice Age termination.

Mass balance parametrizations

Precipitation rates $P(x,y,t)$ are perturbed from present day ($t = t_o$) fields as a function of air temperature T, as[5]:

$$P(x,y,t) = P(x,y,t_o)\exp[a(T(x,y,t) - T(x,y,t_o))] \times \min\left[1.5, \frac{\nabla h(x,y,t) + 0.001}{\nabla h(x,y,t_o) + 0.001}\right] \qquad (2)$$

The final factor incorporates the influence of changes in ice surface slope ∇h, of importance for orographically-driven precipitation at the margins. Present-day precipitation fields $P(x,y,t_o)$ are from refs 9 and 30. The standard case exponential factor $a = 0.0693$.

An annual mass balance is then calculated for all points and times as the difference between accumulation and melt rates. Surface melt and the fraction of precipitation to fall as snow are calculated from the annual degree–day method[5]. Air temperatures are assumed to be normally distributed about the monthly mean values, with a standard deviation of 5 °C.

Fabricated isotopic history

Greenland Summit δ^{18}O and Vostok δD are strongly correlated for averages of several millennia or longer. In terms of the Vostok isotopic history $\Delta\delta_{\mathrm{V}}$ (ref. 24) and Vostok Eemian isotopic maximum $\Delta\delta_{\mathrm{V}}^{\mathrm{E}}$ (Δ indicates deviations from present values) we define $\Delta\delta^* = (\Delta\delta_{\mathrm{V}}/\Delta\delta_{\mathrm{V}}^{\mathrm{E}})\Delta\delta_{\mathrm{E}}$ (Fig. 1) for the Eemian. The preceding glacial is likewise scaled to the Vostok and Greenland minima.

Climate perturbation with isotopic lapse rate correction

The large measured value for $\Delta\delta_{\mathrm{E}}$ may be partly due to reduced elevation of the ice-sheet surface. We assume that all Eemian ice beneath the modern summit originated near the high point of the Eemian ice sheet. Therefore the elevation change ΔZ is for the summit dome. Climate perturbations are then $\Delta T_c = (\Delta\delta^* - \alpha_z\lambda_T\Delta Z)/\alpha_c$. For every location on the ice-sheet surface, temperature perturbations are modulated by the thermal lapse rate and thus $T(x,y,t) - T(x,y,t_o) = \Delta T_c + \lambda_T\Delta Z(x,y,t)$. In some runs we have set $\alpha_c = \alpha_c^i$ for all ages greater than 98 kyr ago, and in others set $\alpha_c = 0.33$ for the penultimate deglaciation ($\delta^{18}O < -35$ per mil and age >120 kyr ago). The resulting difference proves to be trivial in the present context (Fig. 2).

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