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Multistability and critical thresholds of the Greenland ice sheet

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Recent studies have focused on the short-term contribution of the Greenland ice sheet to sea-level rise, yet little is known about its long-term stability. The present best estimate of the threshold in global temperature rise leading to complete melting of the ice sheet is 3.1 °C (1.9–5.1 °C, 95% confidence interval) above the preindustrial climate[1], determined as the temperature for which the modelled surface mass balance of the present-day ice sheet turns negative. Here, using a fully coupled model, we show that this criterion systematically overestimates the temperature threshold and that the Greenland ice sheet is more sensitive to long-term climate change than previously thought. We estimate that the warming threshold leading to a monostable, essentially ice-free state is in the range of 0.8–3.2 °C, with a best estimate of 1.6 °C. By testing the ice sheet's ability to regrow after partial mass loss, we find that at least one intermediate equilibrium state is possible, though for sufficiently high initial temperature anomalies, total loss of the ice sheet becomes irreversible. Crossing the threshold alone does not imply rapid melting (for temperatures near the threshold, complete melting takes tens of millennia). However, the timescale of melt depends strongly on the magnitude and duration of the temperature overshoot above this critical threshold.

Like many components of the climate system, ice sheets are believed to exhibit hysteresis behaviour[2–4]. The theoretical foundation for the possibility of bifurcations between different states rests on the existence of the strongly positive elevation and albedo climate feedbacks. As an ice sheet begins to melt around the margin, the local elevation decreases and temperatures rise. A decrease in area of the ice sheet additionally contributes to warming through a reduction in the surface and planetary albedo. This results in further melting until a new equilibrium is reached. The level of warming needed to melt an ice sheet completely is considered to be a critical threshold, or tipping point[5]. Although precise knowledge of this threshold may not aid centennial timescale predictions of sea-level rise, it is important for assessing the probability of irreversible changes to the cryosphere given the extremely long lifetime of anthropogenic CO_2 in the atmosphere[6].

Coupled climate–ice-sheet models show that the Antarctic ice sheet exhibits hysteresis behaviour for a certain range of atmospheric CO_2 concentrations[7] and that multiple stable states exist for the continental Northern Hemisphere ice sheets under different levels of orbital forcing[8]. For the Greenland ice sheet (GIS), it has been shown that its decline is likely to be irreversible beyond a certain threshold[9–11]. Although previous studies support the idea of multistability of the GIS, the critical temperature leading to the transition to an ice-free Greenland remains highly uncertain. The last Intergovernmental Panel on Climate Change report[12] cites one comprehensive study[1], in which the global mean temperature change leading to total loss of the GIS was estimated to be 3.1 ± 0.8 °C (95% confidence interval of 1.9–5.1 °C). The range was determined by forcing a surface mass balance (SMB) model (with a fixed topography—that is, no dynamic-ice-sheet model) with temperature and precipitation anomalies from atmosphere–ocean general circulation model (AOGCM) simulations. An ensemble of climatic forcing was used to produce a probabilistic estimate of when the SMB would turn negative, used as the criterion for loss of GIS stability. The SMB was calculated using the semi-empirical positive degree day approach, which allows simulation of the present-day SMB with a sufficient degree of accuracy, but it is unclear how appropriate it is for simulations under very different climate conditions[13]. Furthermore, the criterion of a negative SMB implying a loss of stability is theoretically sound as a sufficient condition for melting the GIS, but its accuracy for inferring the critical temperature threshold has not been investigated.

To study the long-term response of the GIS to rising temperatures more directly, we used an intermediate complexity, regional climate model[14] coupled to an ice-sheet model[15] (see Methods). This new approach has been shown to reproduce the regional climate well compared to observations[14] and, most importantly, it explicitly incorporates the albedo and elevation feedbacks that are important for proper simulation of the long timescale response of the GIS to climate change. In a complementary study[16], we carried out simulations over the last glacial–interglacial cycle. Not only do the resulting ice-sheet configurations provide proper initial conditions for the present study, the simulations allowed us to evaluate the sensitivity of the model to a wide range of parameter combinations and to limit the parameter values to those consistent with empirical constraints. Therefore, here we carried out an ensemble of simulations in which we explicitly account for uncertainties in the SMB scheme and regional hydrological sensitivity. We used 11 evenly spaced values of the melt-model parameter, c, that span the widest accepted range of melt-model uncertainty[16], with each value assumed to be equally likely (see Methods). Additionally, we accounted for uncertainty in the regional hydrological sensitivity by applying nine values of a scaling factor that affects the sensitivity of the modelled precipitation to the regional temperature change. The range and weighting of the scaling factor values were chosen to be consistent with results from the statistical analysis by Frieler *et al.*[17] of the World Climate Research Programme's Coupled Model Intercomparison Project phase 3 (CMIP3) multimodel AOGCM data set[18]. Thus, a total of 99 model versions were used to carry out fully coupled simulations of both the transient and quasi-equilibrium evolution of the GIS and to assess the probability distributions of its critical temperature thresholds.

To facilitate such long-timescale simulations, we applied a spatially constant temperature anomaly to the boundary climate over the ocean around Greenland. AOGCM simulations of future warming scenarios over the next century have been used previously to represent the spatial variability of anomalies[1]. However,

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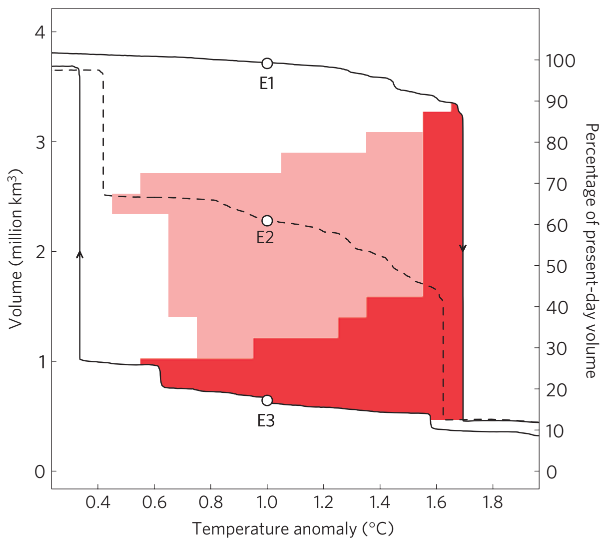

Figure 1 | Stability analysis. Equilibrium stability diagram of the GIS versus the applied regional summer temperature anomaly for the representative model version. The upper branch shows the GIS volume as the temperature increases, starting from the complete ice sheet; the lower branch shows the volume as the temperature decreases, starting from ice-free conditions. The shading shows the modelled basins of attraction in the multistable region. Simulations that start with an intermediate volume in the red regions inevitably approach the lower branch of the diagram in equilibrium for the given temperature anomaly. Simulations that start with an intermediate volume in the light red region reach an intermediate equilibrium volume. Those that start in the white region fully regrow to reach the upper branch of the diagram in equilibrium. The dashed line corresponds to the intermediate equilibrium branch. The open circles labelled E1, E2 and E3 correspond to the equilibrium states shown in Fig. 4.

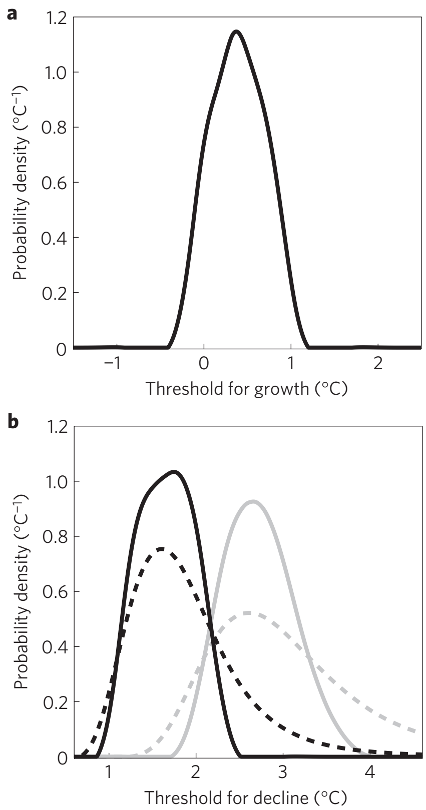

Figure 2 | Threshold estimates. Probability distributions of the regional summer temperature threshold for **a**, growth and **b**, decline of the GIS using the fully coupled climate–ice-sheet model (black solid lines) and the negative SMB criterion (grey solid line). The dashed black and grey lines show the distributions of the gobal mean temperature threshold for decline of the GIS using each method, respectively.

in this case, it should be remembered that the patterns of climate change in transient AOGCM simulations are not identical to the quasi-equilibrium response. Furthermore, there is poor agreement between the models on a local (that is, smaller than the Greenland domain) scale[19]. By using a spatially constant temperature anomaly, we assume higher confidence in the regional rather than local changes.

The CMIP3 AOGCM simulations show that the annual mean warming around Greenland is expected to be at least 50% higher than the globally averaged temperature change, but the summer warming, which is most important for melting, should remain close to the global mean[12,20]. Our model therefore includes a seasonal variation of temperature anomalies that reflects the most likely climatic scenario: an anomaly in winter twice as strong as that in summer. The summer temperature anomalies reported here can also be considered to be relative to preindustrial temperatures, as the mean summer temperature for the period 1958–2001 (used for the boundary conditions of the model) is not considerably different from the preindustrial average[21]. To relate the regional summer temperature to the global mean, we then used the probabilistic distribution of the ratio between regional and global temperatures obtained from the statistical analysis of the CMIP3 AOGCM simulations by Frieler *et al.*[17] (see Supplementary Information).

To be confident of our approach, we compared a transient simulation of the GIS to results from a fully coupled ice-sheet AOGCM under similar warming conditions[22]. Our ice sheet (forced with the anomalies described above) responded in a very similar way to that forced by the more detailed AOGCM fields (see Supplementary Information). From this we can conclude that the climatic changes simulated by our model over Greenland are reasonable and that the local scale differences play only a minor role in the long-timescale evolution of the ice sheet. Thus, for such experiments, these differences can be ignored.

We traced the stability of the GIS relative to temperature changes with a technique routinely used to analyse the stability of the Atlantic meridional overturning circulation[23]—namely, we applied a temperature anomaly changing very slowly in time (that is, at a rate much slower than the response time of the ice sheet), so as to maintain a quasi-equilibrium state at all times. We initialized the simulations using the present-day GIS configurations and applied the slowly increasing temperature anomaly until complete deglaciation was achieved; the anomaly was then decreased slowly until regrowth completed, thus tracing both branches of the stability diagram.

Figure 1 shows the resulting stability diagram for a representative model version (with $c = -55\,\mathrm{W\,m^{-2}}$ and a hydrological sensitivity of $\sim 7.5\%\,{}^\circ\mathrm{C}^{-1}$) of the 99 tested in our ensemble. In this case, the regional summer temperature threshold leading to GIS decline is 1.7 °C. Above the threshold, the ice sheet retreats to an essentially ice-free state (about 10% of the modern GIS volume). Conversely, when starting from warm, ice-free conditions, the ice sheet builds volume in the high-elevation regions until a critical point is reached around 0.3 °C and the ice sheet switches to the fully glaciated state. All simulations showed this hysteresis behaviour, offset to higher or lower temperatures, but with the width of the multistable region consistently being approximately 1.4 °C (individual stability curves for all simulations, along with animations, are given in the Supplementary Information).

Results from the entire ensemble of simulations show that the best estimate for the regional summer temperature threshold for GIS growth from ice-free conditions is 0.4 °C, with a 95% credible

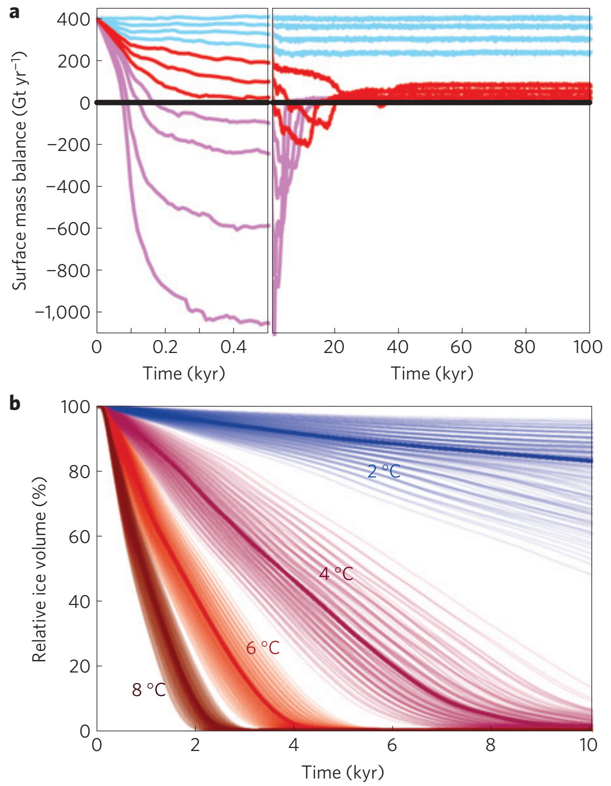

Figure 3 | Transient GIS evolution. a, The modelled transient SMB for the representative case with an applied constant regional summer temperature anomaly ranging from 0 to 6 °C. Simulations are separated into those that melt the ice sheet completely and exhibit negative SMB within the first 200 years (purple), those that melt completely but do not exhibit negative SMB in the first 200 years (red) and those that do not melt substantially (blue). **b**, Relative GIS volume change in transient simulations over the next 10,000 years, for all model versions (thin lines) and the representative model version (thick lines). Each colour denotes the constant regional summer temperature anomaly applied in each case (blue, 2 °C; purple, 4 °C; red, 6 °C; dark red, 8 °C).

interval of −0.2–1.0 °C (Fig. 2a). It is, therefore, still possible that the GIS is bistable for preindustrial conditions. The regional summer temperature threshold for GIS decline is estimated to be 1.8 °C, with a 95% credible interval of 1.1–2.3 °C (Fig. 2b). Using the AOGCM-estimated scaling coefficients[17], we can convert from a regional summer temperature anomaly into the global mean temperature anomaly (see Supplementary Information). We find the best estimate for the global mean temperature threshold for decline of the GIS to be 1.6 °C with a 95% credible interval of 0.8–3.2 °C (Fig. 2b).

Our results indicate that the threshold temperature for decline is likely to be significantly lower than the previous best estimate of 3.1 °C obtained using the negative SMB criterion. The existence of competing growth and melt processes on different time and spatial scales means that the present-day SMB is not a representative indicator of equilibrium ice-sheet stability. Furthermore, most precipitation occurs in the southeast over the high-elevation coastal regions that are hydrologically isolated[24,25] from more sensitive regions of the ice sheet. In transient simulations under warming, the GIS always disappears when negative SMB is reached in the first few centuries (Fig. 3a). However, several other simulations also result in GIS decline, yet they do not exhibit negative SMB throughout the first millennia. This shows that the threshold in the SMB for the GIS that leads to decline will be greater than zero (in our simulations, it lies in the range of 150–340 Gt yr^{-1}). The ice sheet will thus still seem stable for a temperature anomaly approximately 1 °C higher than predicted using our coupled climate–ice-sheet model (Fig. 2b), whereas deglaciation processes in more sensitive regions will probably already have been initiated.

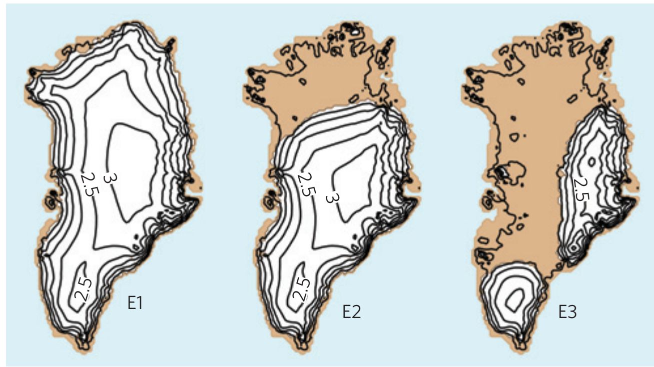

Figure 4 | Equilibrium states of the GIS. Three example equilibrium states of the GIS obtained for the representative model version with an applied temperature anomaly of 1 °C. These states correspond to those denoted by the open circles in Fig. 1.

Crossing the deglaciation threshold, however, does not imply a rapid collapse of the ice sheet. In fact, the response time of the GIS is extremely long for temperatures near the threshold. For higher temperatures, rapid decline of the order of hundreds of years is possible. Figure 3b shows the results from a set of transient experiments with a constant temperature anomaly applied in each simulation. For 2.0 °C regional summer warming, which is just above the deglaciation threshold in the representative case, complete melting of the GIS takes about 50,000 years. In contrast, with warming of 4.0 °C, the ice sheet needs about 8,000 years to melt completely, and for warming of 8 °C, 20% of the ice sheet melts in just 500 years and the entire ice sheet melts within about 2,000 years. Thus, the time needed to melt a significant portion of the GIS is strongly dependent on the level of warming.

We address the question of irreversibility by mapping the basins of attraction towards the stable branches of the hysteresis diagram (for the representative case). Following a methodology from previous work[11], we produced a set of reduced-volume ice-sheet configurations by applying a constant temperature anomaly just above the deglaciation threshold and saving the output after every 5% volume reduction. Using a reduced-volume configuration as the initial state, we then applied a different constant temperature anomaly and ran the model until equilibrium was reached. When a temperature anomaly outside the multistable region is applied for any initial conditions, the GIS inevitably evolves towards one of the two stable branches shown in Fig. 1. For temperatures above the deglaciation threshold, the GIS always melts completely (for example, state E3 in Figs 1 and 4) and for temperatures below the glaciation threshold, the GIS always approaches its full size (for example, state E1 in Figs 1 and 4). For temperatures inside the multistable region, the equilibrium state of the GIS is strongly dependent on the initial conditions. A reduced-volume ice sheet that initially lies within the red area for a given temperature anomaly (Fig. 1) will eventually approach the lower branch of the diagram (that is, almost complete deglaciation). Ice-sheet configurations that are initially located within the light red region evolve towards an intermediate equilibrium state (for example, state E2 in Figs 1 and 4), which is qualitatively similar to that found by Ridley and colleagues[11]. As long as the GIS volume remains in the attraction domain of the lower branch of the stability diagram, it will continue to melt irreversibly, even if the temperature anomaly later decreases.

Our study shows that a temperature threshold for melting the GIS exists and that this threshold has been overestimated until now. For a sufficiently large temperature anomaly, a significant portion of the ice sheet may be lost within several centuries, and the GIS will continue to melt even if temperatures later drop below the threshold value (but stay in the multistable range). Therefore, if anthropogenic CO_2 emissions in the coming century drive the temperature considerably above the deglaciation threshold, irreversible total loss of the GIS will be difficult to avoid, ensuring continued substantial sea-level rise for millennia.

Methods

The regional energy-moisture balance model REMBO (ref. 14) simulates daily temperature and precipitation fields over Greenland under the assumptions that temperatures over Greenland are largely controlled by lateral energy and moisture transport dominated by synoptic-scale processes, which can be parameterized as a diffusive process, and that precipitation is mostly orographically controlled. As the boundary conditions for REMBO, we use the monthly ERA-40 reanalysis climatological (1958–2001) temperature and relative humidity outside Greenland[26]. Temperature anomalies are prescribed over the boundary ocean points. The model variables, including SMB, snowpack thickness and surface temperature, are calculated daily to track changes in surface albedo.

REMBO is coupled bidirectionally to the three-dimensional, polythermal shallow-ice approximation ice-sheet model SICOPOLIS (ref. 15). SICOPOLIS includes a locally deforming lithosphere model to account for bedrock deformation. The SMB and surface temperature are input as boundary conditions to SICOPOLIS and changes in topography and ice-sheet extent calculated by the ice-sheet model are input to REMBO. The climate and SMB fields are updated every ten ice-sheet model years to provide accurate surface forcing to the ice sheet. Most importantly, REMBO coupled to SICOPOLIS explicitly captures elevation and albedo feedbacks in the climate–ice-sheet system at relatively high resolution (20 km) compared with general circulation models.

The modelled SMB and related variables (total precipitation, melting, runoff, refreezing) lie within the range of published state-of-the-art results from regional climate models[14]. In particular, the first-order dependence of melt on elevation produced by energy balance modelling is captured with the SMB equation used here[14]:

$$M_s = \frac{\Delta t}{\rho_w L_m}[\tau_a(1-\alpha_s)S + c + \lambda T]$$

where M_s is the daily melt rate, Δt is the conversion factor for the time step (seconds per day), ρ_w is the density of water and L_m is the latent heat of melting. The short-wave radiation reaching the surface is calculated from S, the insolation at the top of the atmosphere, and α_s and τ_a, the surface albedo and atmospheric transmissivity, respectively. Surface albedo is obtained through a simple parameterization based on the type of ground surface present and the snow thickness[14]. The long-wave radiation and sensible and latent heat fluxes are incorporated as a linear approximation, where T is the daily temperature and λ and c are empirical coefficients. Based on simulations over the past two glacial cycles[16], we found that the choice of the free parameter, c, largely determined the sensitivity of the modelled GIS to changes in climate.

To generate the ensemble of model versions, we first varied the parameter c in the melt equation in the constrained range of -60 to $-50\,\mathrm{W\,m^{-2}}$ in $1\,\mathrm{W\,m^{-2}}$ increments (producing 11 model versions). All of these model versions are consistent with observational and palaeoclimate constraints[16]. Furthermore, results from an analysis[17] of the CMIP3 AOGCM data set[18] show that the expected increase in precipitation over Greenland per degree of regional summer warming is of the order of $6.3 \pm 3.3\%\,^{\circ}\mathrm{C}^{-1}$ (see Supplementary Information). The standard version of REMBO exhibits an increase in precipitation relative to the regional summer warming of approximately $9\%\,^{\circ}\mathrm{C}^{-1}$, which lies at the high end of the range of AOGCM predictions. To better account for the uncertainty in the sensitivity of the regional precipitation to temperature changes, we include an additional precipitation scaling factor. Using this factor, we modified the magnitude of precipitation calculated by REMBO such that the sensitivity of the simulated regional precipitation to the summer temperature anomaly varies in the broad range of 2–$12\%\,^{\circ}\mathrm{C}^{-1}$ (producing 9 model versions). The result of these parameter perturbations is an ensemble of 99 model versions. Each model version was then weighted based on the AOGCM sensitivities, to produce the temperature-threshold probability distributions (see Supplementary Information for details).

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Author contributions

All authors contributed equally to this work.

Additional information

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