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Thresholds for irreversible decline of the Greenland ice sheet

Jeff Ridley · Jonathan M. Gregory · Philippe Huybrechts · Jason Lowe

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Abstract The Greenland ice sheet will decline in volume in a warmer climate. If a sufficiently warm climate is maintained for a few thousand years, the ice sheet will be completely melted. This raises the question of whether the decline would be reversible: would the ice sheet regrow if the climate cooled down? To address this question, we conduct a number of experiments using a climate model and a high-resolution ice-sheet model. The experiments are initialised with ice sheet states obtained from various points during its decline as simulated in a high-CO_2 scenario, and they are then forced with a climate simulated for pre-industrial greenhouse gas concentrations, to determine the possible trajectories of subsequent ice sheet evolution. These trajectories are not the reverse of the trajectory during decline. They converge on three different steady states. The original ice-sheet volume can be regained only if the volume has not fallen below a threshold of irreversibility, which lies between 80 and 90% of the original value. Depending on the degree of warming and the sensitivity of the climate and the ice-sheet, this point of no return could be reached within a few hundred years, sooner than CO_2 and global climate could revert to a pre-industrial state, and in that case global sea level rise of at least 1.3 m would be irreversible. An even larger irreversible change to sea level rise of 5 m may occur if ice sheet volume drops below half of its current size. The set of steady states depends on the CO_2 concentration. Since we expect the results to be quantitatively affected by resolution and other aspects of model formulation, we would encourage similar investigations with other models.

Keywords Greenland · Ice sheet · Climate model · Climate change

J. Ridley (✉) · J. M. Gregory · J. Lowe
Met Office Hadley Centre, FitzRoy Rd, Exeter, UK
e-mail: jeff.ridley@metoffice.gov.uk

J. M. Gregory
Department of Meteorology, Walker Institute for Climate System Research, University of Reading, Reading, UK

P. Huybrechts
Department of Geography, Earth System Sciences, Vrije Universiteit Brussel, Pleinlaan 2, 1050 Brussels, Belgium

1 Introduction

In the present day climate, the surface mass balance of the Greenland ice sheet (GIS) is positive (Hanna et al. 2005); accumulation (mainly snowfall) exceeds ablation (mainly melting followed by runoff). In a steady state, a positive surface mass balance is matched by solid ice discharge into the sea, together with a minor contribution from melting at the base of the ice sheet. In recent years, the net mass balance has been negative partly on account of increases in dynamical ice discharge (Lemke et al. 2007; Rignot et al. 2008).

General circulation models (GCMs) predict that both temperature and precipitation will be greater over Greenland in a climate of elevated CO_2, and the consequent increase in melting will dominate the increase in accumulation, so that the surface mass balance will tend to become less positive (Wild and Ohmura 2000; Kiilsholm et al. 2003; Huybrechts et al. 2004). With sufficient climate warming, the surface mass balance will become negative. This degree of warming has been suggested as a threshold of sustainability for the ice sheet: in these circumstances the ice sheet would eventually be eliminated, because its net mass balance would be negative even if it retreated inland and ice discharge into the ocean was eliminated. The threshold of sustainability of the ice-sheet may be regarded as a "tipping point": below the threshold, an ice-sheet would persist of some reduced size which would depend on the climate forcing; beyond the threshold, the ice sheet would be lost (perhaps over a long timescale) with no requirement for further increase in climate forcing.

Evaluation of the surface mass balance of the GIS using an ice-sheet model with climate perturbations from a range of atmosphere–ocean general circulation models (AOGCMs) indicates that the threshold of sustainability would be reached with a global-mean temperature rise in the range 1.9–4.6 K relative to pre-industrial (Gregory and Huybrechts 2006). AOGCMs project that unmitigated anthropogenic greenhouse-gas emissions would lead to global-mean warming of several degrees during the twenty-first century. Likely ranges of warming relative to pre-industrial are 1.6–6.9 K (based on Meehl et al. 2007). It is thus possible to describe the probability of exceeding the threshold of sustainability of the Greenland ice sheet as a function of the uncertainty in temperature rise associated with the atmospheric greenhouse gas concentration as shown in Fig. 1. This was produced by combining the temperature threshold from Gregory and Huybrechts (2006) with an estimate of uncertainty in climate sensitivity from Murphy et al. (2004).

To study the evolution of the ice sheet under a transient future greenhouse climate, a number of AOGCMs have been modified to incorporate three-dimensional thermo-mechanical ice sheet models (ISMs). Such models allow changes in climate simulated by the AOGCM to interact with the ISM through surface mass balance feedbacks. The feedbacks include changes in surface albedo and elevation, changes in the atmospheric circulation induced by

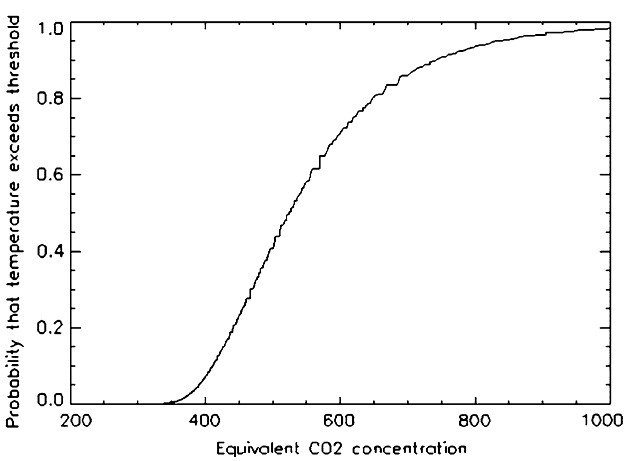

Fig. 1 The probability, as a function of atmospheric equivalent CO_2 concentration, that the global temperature rise exceeds the threshold of sustainability of the Greenland ice sheet, above which eventual deglaciation is expected

topography change, and changes in oceanic circulation caused by changes in freshwater runoff (Huybrechts et al. 2002; Ridley et al. 2005; Mikolajewicz et al. 2007). It was found that the increased runoff can cause a slight (1 Sv) slowdown in the North Atlantic thermohaline circulation. The reduction in mean elevation of the ice sheet changes the Northern annular mode, causing a warming to the west of Greenland and a cooling over Northern Europe. The albedo-temperature feedback is effective after about 30% of the ice sheet has been lost and results in an increase in annual mean temperature of the ice-free regions of 5–8°C. The coupled AOGCM-ISMs (henceforth "AIOGCMs") confirm the surface mass balance implications of Gregory and Huybrechts (2006), that with sufficient increases in CO_2 concentration the GIS declines and may be eliminated. In particular, Ridley et al. (2005) found that with an atmospheric CO_2 concentration of four times pre-industrial (1,120 ppm) the entire GIS is melted within 3,000 years. The loss of the GIS could take much longer with moderate warming (Greve 2000; Driesschaert et al. 2007).

The question we address here is whether the ice sheet could recover as the climate cooled following a reduction in CO_2 concentration. Because of the positive climate feedbacks on ice-sheet decline, cooling below the original threshold of sustainability might not be sufficient for ice-sheet regrowth; there could be hysteresis in the coupled system of ice-sheet and climate. Toniazzo et al. (2004) found evidence for this. The climate simulated by their AOGCM for pre-industrial CO_2 with GIS completely removed did not give the positive surface mass balance required for regrowth. However, their AOGCM spatial resolution may not have been sufficient to model ice-sheet inception, and Lunt et al. (2004) found on the other hand that regrowth could occur using a high-resolution ISM. Charbit et al. (2008) have recently addressed this question using an Earth system model of intermediate complexity coupled to a high-resolution ISM. They considered various emissions of CO_2 and the carbon cycle in their model simulated its gradual removal. With cumulative emissions above 3,000 GtC, climate warming leads to irreversible removal of the ice sheet, whereas for smaller emissions it recovers over several thousand years. However, the low climate sensitivity of their model meant that no scenario exceeded a global temperature rise of 4°C.

In this work we examine whether recovery of the ice-sheet would be possible from various states of decline by transplanting them into a climate of pre-industrial atmospheric CO_2 (280 ppm). Obviously, since CO_2 could not instantaneously be reduced in this way, our experimental design is an idealisation. Our aim is to look for a threshold of irreversibility in the size of the ice-sheet, above which it could regrow in pre-industrial CO_2, and below which it would be inevitably lost. The existence of such a threshold is implied by the results of Toniazzo et al. (2004) and Charbit et al. (2008). This threshold is of practical significance, because it is a "point of no return", which would set a constraint on acceptable future timelines of CO_2 concentration (Lowe et al. 2008) in order to avoid the eventual loss of the ice sheet, which is an outcome that might well be regarded as a dangerous consequence of climate change. If there is a threshold of irreversibility in size, it implies that a warm climate must not be allowed to persist for more than a certain time while the ice sheet is losing mass.

2 Method

In this paper we use the HadCM3 AOGCM (Pope et al. 2000; Gordon et al. 2000), having an atmosphere resolution of 3.75° in longitude and 2.5° latitude, and a Greenland ice sheet model of 20 km resolution (Huybrechts and de Wolde 1999). These components are the same as used in the AOIGCM described by Ridley et al. (2005). However, in this work we couple them asynchronously, rather than using synchronous coupling as in the ice sheet decline experiment. This is because the rate of growth of ice sheets through accumulation is typically an order of magnitude slower than their melt (Zweck and Huybrechts 2005) and consequently the computational resource to simulate regrowth with an AIOGCM is prohibitive. In other respects, the coupling between the AOGCM and ISM follows that described in Ridley et al. (2005). Near surface temperature anomalies and precipitation are bilinearly interpolated from the AOGCM to ISM with a modification at the ice edge. In the AOGCM the grid cells are either classified as ice-covered or ice-free, whereas at the ice edge, the same area in the ISM will be only partially ice covered. To compensate for a low biased temperature when interpolating from the AOGCM to the ISM grids, the temperature on the AOGCM grid becomes the average between that of the ice covered grid cell and the adjacent ice free grid cell (land or ocean) at the same latitude. The feedback from the ISM to the climate model is through the orography and surface characteristics, especially albedo, which are initialised at the start of each AOGCM simulation, and through changes in the volume of runoff from the ice sheet, which is routed through fixed networks, based on present-day orography, to the coast. Our experimental design is shown in Fig. 2.

In each simulation step, the current high-resolution ISM state is converted to orographic and surface type conditions (e.g. albedo) for the AOGCM, replacing the previous values in the region of Greenland. Next the AOGCM is used to simulate 30 years of climate. It appears from inspection that 4 years is enough for the AOGCM atmosphere to spin up to the new surface conditions; this short period indicates that the change between ISM states is not a large perturbation. The remaining 26 years of the simulation provide the surface forcing (surface air temperature anomalies and precipitation minus evaporation) for the next integration of the ISM. Time series comprising individual months are used, rather than a mean monthly climatology, in order to include the effect of inter-annual variability on mean surface mass balance. This matters because the positive-degree-day scheme, used by the ISM for ice ablation, is a non-linear function of temperature.

As in previous work, our melt and runoff model uses constant positive degree-day factors for snow and ice melting obtained from a calibration to the present-day mass balance and ice-sheet configuration (Janssens and Huybrechts 2000; Huybrechts et al. 2002). The model distinguishes between liquid and solid precipitation, and takes into account the process of meltwater retention by refreezing and the capillary suction effect in the snowpack.

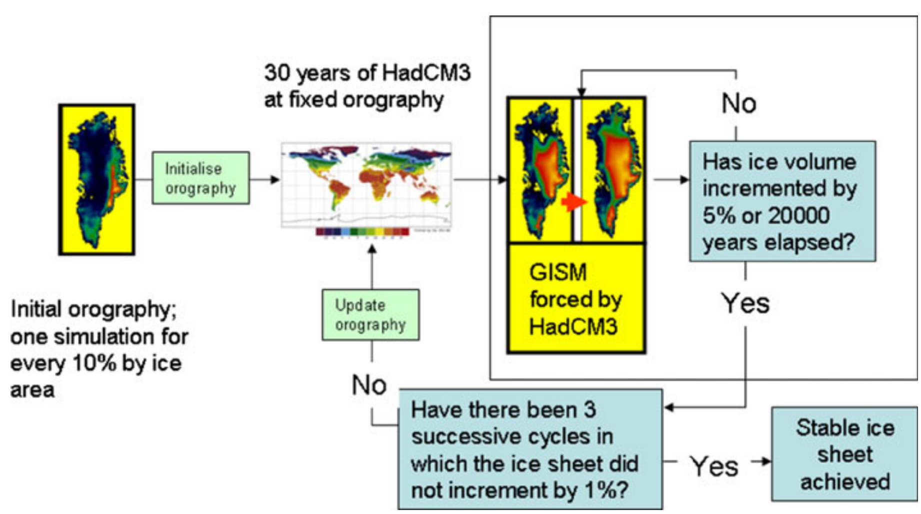

Fig. 2 Schematic diagram showing the experimental method. The ISM-derived surface type and topography determine the AOGCM surface boundary conditions in the region of Greenland. Iterations of climate and ice sheet continue until three cycles have passed with no significant change in the ice sheet volume

The melt model employed here has a somewhat larger sensitivity to temperature changes than older degree-day models, and its sensitivity also rises more quickly for positive perturbations. This is largely due to the separate treatment of the rain fraction, which is generally found to contribute more to runoff than to accumulation, and which is often neglected in ice-sheet modelling.

In order to simulate long time periods, the 26 years are recycled until a change in GIS area of 5% is realised or 20,000 simulated years have elapsed. The first condition is reached sooner when the ice sheet is rapidly evolving. The figure of 5% change in ice sheet is a somewhat arbitrary choice. Because it is a substantial change, it limits the number of ISM-AOGCM iterations and the computational requirement. Substantial changes in topography can occur without change of area, since the combination of ice dynamics and surface forcing can produce a significant redistribution of ice mass. However, an area threshold has the advantage of simplicity, being directly related to the climate state through the summer albedo. During the ISM simulation changes in ice sheet elevation occur as a consequence of surface mass balance, ice dynamics and isostatic balance of the lithosphere. Elevation changes are converted to surface temperature changes through a fixed lapse rate of 6.227°C km^{-1}.

After the ISM simulation step the process is repeated. If three successive iterations occur, each of which has not changed ice sheet volume by more than 1%, the ice sheet is considered to have reached a steady state. Subsequent iterations could still produce minor changes as a consequence of multi-decadal variability among the 30-year climatologies. This seems to be particularly true when northern Greenland is ice free, but in the majority of cases the total ice volume variability is less than 1%. Such small changes confirm that 30 years is usually a sufficiently long period to represent the climatology.

The weakness of asynchronous coupling is that internal variability on periods longer than 30 years, that relating to ocean circulation, is lost. Multi-decadal variability can cause significant variability over the north of Greenland where the precipitation and temperature are low and the balance between ice growth and ablation is delicate. The specific results described here may depend on the method of coupling between AOGCM and ISM, and in particularly the characterisation of the lapse rate when determining near-surface temperature anomalies.

3 Results

Eleven initial GIS states are selected from the simulation by Ridley et al. (2005). During that simulation under four times pre-industrial CO_2 the ice sheet declined from its present day volume to 3% of that volume over 3,000 years. The bedrock rebound and the vertical profile of ice sheet temperatures of our 11 initial ice sheet states represent instantaneous states (i.e. not equilibrium) during the ice sheet decline. The states are selected at 10% fractions of the area coverage of GIS, from 0% (fully depleted) to 100% (present day ice sheet). We choose the initial states on the basis of area coverage because of its climatic relevance through albedo, but the final state is described by the ice volume as this defines equivalent sea level change. Several ISM-AOGCM iterations are required to produce a stable ice sheet for each of the 11 sets of initial ice sheet conditions (Fig. 3).

Total ice volume for all the depleted ice sheets grows, when placed in a pre-industrial climate, including from an ice-free Greenland (0% initial state). The latter outcome to some extent contradicts the result of Toniazzo et al. (2004); since the AOGCM is the same, the difference arises from the ice-sheet model, as the initial growth occurs in the eastern and southern mountains which are poorly resolved in the AOGCM. Most of the growth in the highly depleted ice sheets is on the Greenland southern dome, where simulated precipitation is high. As the dome increases in elevation, topographically induced precipitation increases and growth accelerates, being eventually balanced by ice

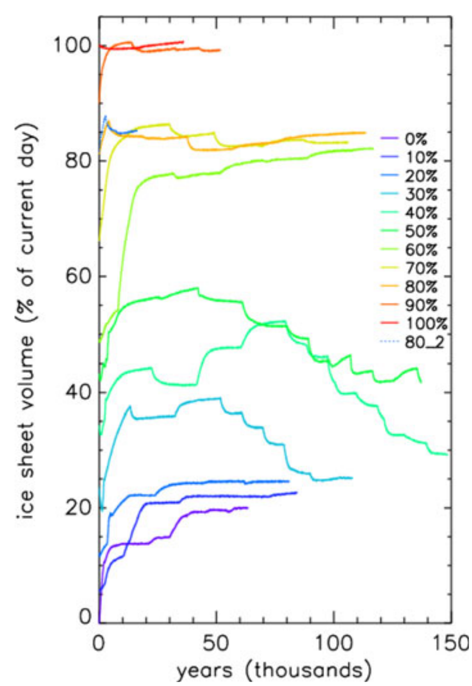

Fig. 3 Changes in ice sheet volume from initial states towards final equilibrium. The start of each ISM simulation step is indicated by a kink in the timeline, after which the ice sheet approaches a temporary equilibrium with the new GCM-generated climatology. Where the kinks are small the ice sheet is near a true coupled equilibrium state. The figure suggests convergence towards three equilibrium states, at about 100, 80 and 20% of present day volume

discharge. The decline in volume of some intermediate initial states (30–60% simulations in Fig. 3), before their growth, is due to an initial dominance of ablation rates in the low-albedo, low-elevation interior of Greenland, over the rate of growth of the south dome.

The asynchronous methodology results in discontinuities in the rate of change of volume (Fig. 3). Whether the ice sheet is growing or shrinking, the fixed climate during each ISM step allows the GIS by itself to approach a steady state. Particularly if the ice sheet is growing in one region and shrinking in another, significant local imbalances between ice sheet and forcing climate may develop before the 5% criterion is attained for the overall change in GIS area. The next climate may then be substantially different and produce rapid change in the ice sheet, possibly reversing some of the previous ISM step. During periods of rapid change a more frequent coupling to the climate might be desired. A synchronously coupled AOIGCM would result in a smooth time-series. However, our interest is not in the trajectories, but in the final steady states, and we do not think these are affected by the jerkiness on the way. It is possible to smooth the trajectories by reducing the integration time step to less than the 5% area changes. We assess this by reducing the asynchronous stepping to 2.5% of the ice sheet for the initial part of the 80% simulation (Fig. 3). Area changes of 5 and 2.5% in the climate coupling produce similar ice sheet simulations, suggesting that there may be strong local climate feedbacks embedded in some of the trajectories.

The volume time series are not a complete description for the state of the ice sheet. The trajectory of change in ice-sheet geographical thickness distribution during ice sheet growth can be quite different from that during its decline. Furthermore, an ice sheet which grows from an initial low-volume state to exceed the next-largest initial volume does not then follow the trajectory of the latter. Consequently, each ice sheet trajectory is unique, depending on the details of its initial state. This implies that our specific results may depend on the deglaciation trajectory from which the initial states were derived. A different AOGCM or ISM, perhaps with improved ice stream dynamics, might alter the ice sheet trajectories.

The ice sheet volumes eventually converge towards three equilibrium states at about 100, 80 and 20% of present day volume. The initial state of 90% grows to 100%, while that of 80% ends up at 80%. The ice sheet thickness for each of the equilibrium states are shown in Fig. 4a–c. The thickness distributions of the individual simulations which are tending towards each stable state are still statistically distinguishable, implying that they need longer to reach true equilibrium. We note that one of the trajectories, that with an initial state at 50% of present day ice volume has definitely not yet iterated to an equilibrium state. This ice sheet trajectory can be identified, in Fig. 3, by the moderate steps which are still occurring in the time series at the conclusion of the experiment. We think that this ice-sheet may be approaching the 20% state. An analysis of its behaviour suggests that the variability in the 26 year climates over north Greenland, is high and a longer simulated climate might be needed to obtain equilibrium.

It is possible that the equilibrium states obtained through asynchronous coupling are an artefact of the methodology. To test this we undertook a synchronously coupled simulation with pre-industrial atmospheric composition (using the AOIGCM of Ridley et al. 2005), starting from the final 20% state (specifically, the state which had resulted from the initial 10% ice sheet). A 100-year simulation showed

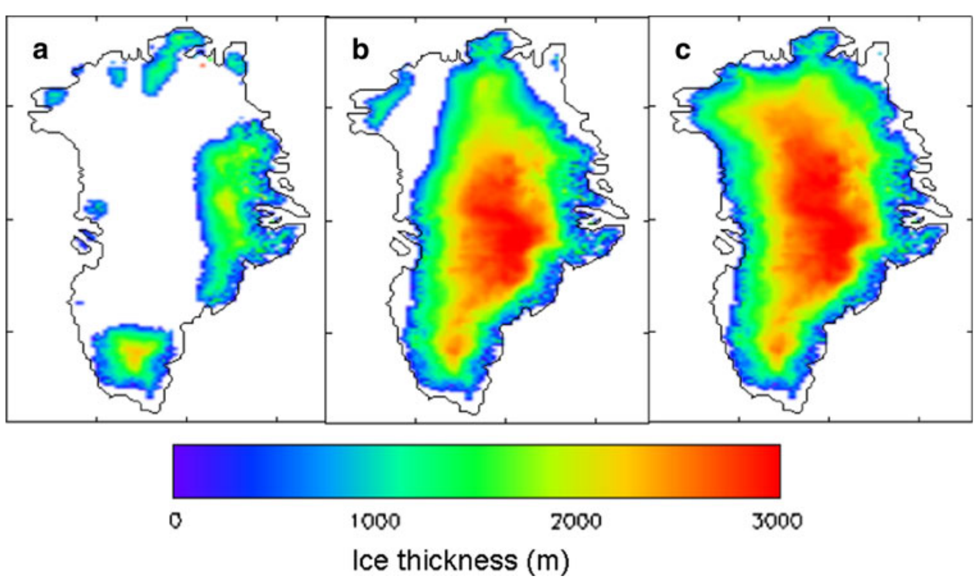

Fig. 4 The ice thickness at each of the equilibrium states by volume for pre-industrial atmospheric CO_2 forcing: **a** 20% (achieved from initial states of 0, 10, 20, 30 and 40%), **b** 80% (achieved by initial states of 60, 70 and 80%) and **c** 100% (achieved by initial states of 90 and 100%)

no mean drift in volume, supporting the stability of the final states.

The appearance of multiple distinct steady states is possible because of the influence the ice sheet has on its regional climate. We investigate this further regarding the different destinations of the initial 80 and 90% states, by looking at the west-east ISM surface elevation cross sections across the northern part of Greenland at 78°N (Fig. 5). The western margin of the 80% ice sheet lies east of the fringing mountains, and there is a topographic low point in the ISM surface elevation profile. The topography is not resolved when regridded to the atmosphere GCM grid, which represents the mountains only by increasing the orographic roughness. The predominately easterly winds, entering the rough region from a smooth ice sheet surface, experience increased surface wind stress. Increased wind stress results in more vertical turbulent mixing and a deepening of the boundary layer. A deepened turbulent boundary layer entrains sensible heat from the free troposphere and transports it to the surface. Such an interaction between orography and atmospheric dynamics raises the summer surface temperature sufficiently to result in net local ablation. The initial 90% ice sheet edge is further west and the enhanced orographic roughness is not present, so the feedbacks do not occur.

We confirm the hypothesis that ice sheet–climate feedbacks make the difference by using the AOGCM climate from the final 80% state to force an ISM integration starting from the 90% initial ice sheet state, and the climate from the 90% equilibrium state with the 80% initial ice sheet state. Figure 6 shows that under these conditions the 80% ice sheet grows everywhere, to 100%, whilst the 90% ice sheet is depleted. The principal depletion occurs in the north-west region, the same area where the ice sheet is completely absent in the 80% equilibrium state.

Where the initial ice sheet volume is 50% or less than present day, the albedo feedback is the dominant forcing mechanism to reduce the ice area (Ridley et al. 2005; Mikolajewicz et al. 2007). When free of ice, the near-surface air temperature over central Greenland is 10°C in July, and 5°C in August. The annual mean snow accumulation of 0.2 m has melted by the end of June. Summer ablation of the ice sheet, by the pool of warm air, causes its eastwards retreat.

If sustained indefinitely, a climate which is warmer than pre-industrial, but which does not pass the threshold of sustainability discussed in the introduction, would eventually produce a steady-state ice-sheet smaller than present, but still occupying much of Greenland (Greve 2000; Driesschaert et al. 2007). We demonstrate this situation by forcing the ISM, starting from the initial 100% state, with the climate of the SRES-B1 scenario. This produces a stable ice sheet (for a global warming of 2.7°C), after two iterations, with 77% of the initial volume and 82% of initial area. The surface mass balance remains positive, but is reduced, and ice discharge into the ocean is likewise reduced to 60% of its initial value. Since this ice sheet is less than 80% of its initial volume, it is smaller than the threshold of sustainability and would not regrow to 100% volume under a pre-industrial climate.

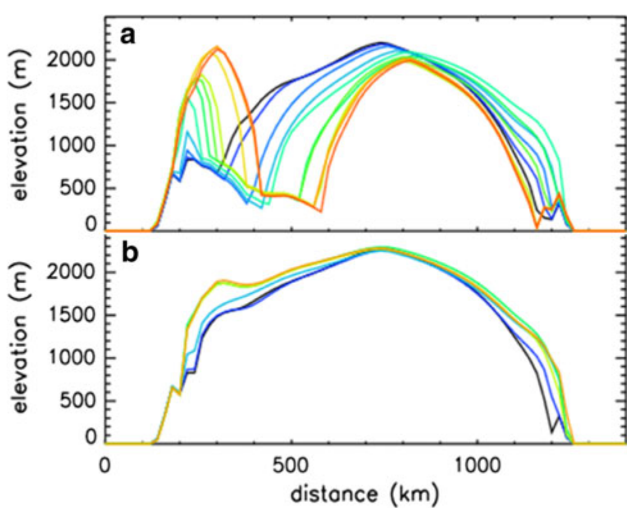

Fig. 5 Surface elevation profiles from West to East at 78°N during the iterative evolution from the 80 and 90% initial states. The initial state of the ice sheet is in *black* with colours (from *blue* to *red*) representing the state after subsequent iterations. **a** The initial 80% ice sheet shows ablation which exaggerates a topographic minimum on the western margin. **b** The orographic low is not present on the initial 90% ice sheet

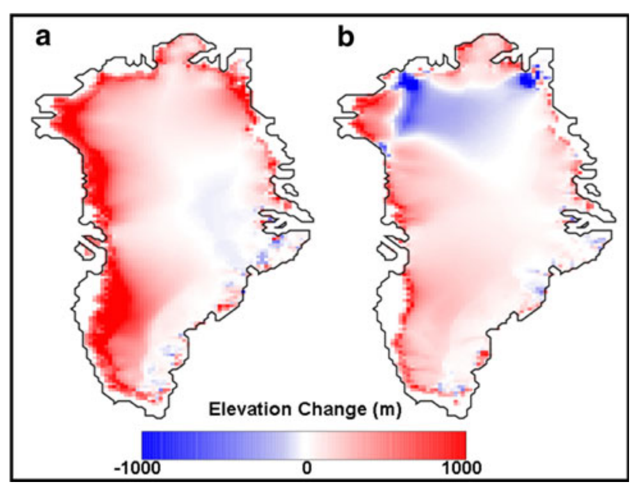

Fig. 6 Ice thickness change during two experiments: **a** The result of starting from the initial 80% ice sheet and forcing with the 90% climate. **b** The result of starting from the initial 90% ice sheet and forcing with the equilibrium 80% climate; this reveals a distinctive ablation pattern in the NW

4 Discussion

With stabilisation of greenhouse emissions an irreversible decline of the Greenland ice sheet will occur depending on the level of emissions as shown in Fig. 1. But our experiments show that feedback between ice-sheet and climate causes there to be more than one potential equilibrium state of the GIS for pre-industrial and present-day CO_2 concentration. For pre-industrial CO_2, our model indicates three steady states. Following complete deglaciation of Greenland, an ice-sheet of only about a quarter of the present volume could be regenerated under pre-industrial CO_2 conditions, implying an irreversible sea-level rise of more than 5 m. Other climate or ice sheets models, with different resolution or parameterisations, may produce different ice sheet states. However, if the result is qualitatively correct for the real world, it implies that the partial loss of the Greenland ice sheet as a consequence of anthropogenic climate change would be irreversible if the ice sheet is reduced below a critical size.

The future evolution of the Greenland ice sheet depends on the level of climatic warming and how long it is maintained, as shown schematically in Fig. 7. The ice sheet declines rapidly under warming and recovers slowly, through accumulation, on return to a pre-industrial climate. The steady state ice sheet achieved during recovery is dependent on the amount of ablation sustained during the warming phase.

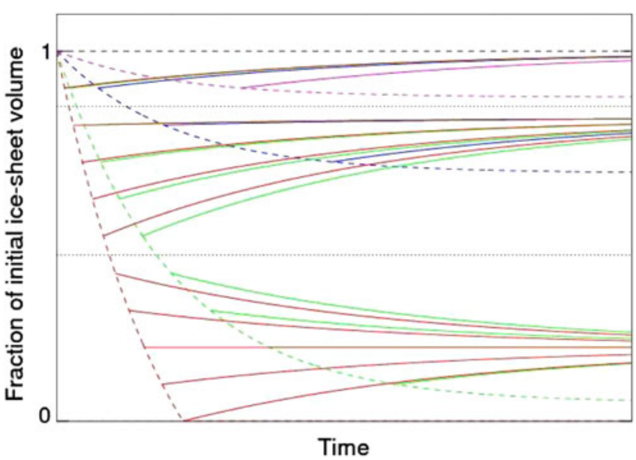

Fig. 7 Schematic diagram showing the future evolution of the Greenland ice sheet for four different levels of climatic warming, each maintained for a range of times, after which the climate reverts to pre-industrial. The colours indicate the climatic warming, *purple* for least and *red* for most; the ice sheet contracts faster in warmer climates. The *dashed lines* are the phase during which the warming is applied, and the *solid lines* the subsequent recovery phase towards three different steady states. The *dotted horizontal lines* are thresholds of irreversibility (points of no return). Thresholds of sustainability (tipping points) cannot easily be shown on this diagram; they are temperature thresholds, and correspond roughly to different initial rates of contraction. The diagram is not intended to be quantitative

In our model, the threshold of irreversibility lies between 80 and 90% of the initial ice sheet volume. If the threshold were passed, the ice-sheet would not recover to more than about 80% of its initial volume, implying an irreversible sea-level rise of at least 1.3 m, even if global climate otherwise reverted to a pre-industrial state.

For the ice sheet to reach the threshold size of irreversibility, a sufficiently warm climate has to be sustained for long enough. This means that the system has a temporary resilience and may allow some degree of warming overshoot above a dangerous threshold (Huntingford and Lowe 2007) for a limited period of time. Assuming the threshold is 90% of the initial size, and using the ice-sheet mass balance sensitivities of Gregory and Huybrechts (2006), we give the median time allowed for various global warmings (Table 1). If these time-limits were not exceeded, the 100% ice sheet could still be regained if the climate subsequently returned to pre-industrial.

For instance, at 4°C above pre-industrial, which is a mid-range warming for AOGCMs following the A2 scenario, the median time to reach the threshold of irreversibility is 900 years. If the climate system or the ice-sheet is more sensitive than the median model, the time is shorter. By comparison, the timescale on which the climate could return to pre-industrial is very long. Solomon et al. (2009) show that the warming after 1,000 years is still 40–60% of the equilibrium warming corresponding to the peak CO_2 concentration. We conclude, therefore, that the scope for a "safe" overshoot is limited; a warming substantially exceeding the threshold of sustainability is likely to lead to passing the point of no return (the threshold of irreversibility) for the Greenland ice-sheet before the climate could recover, and hence to irreversible sea-level rise.

Table 1 Time required in years, to the nearest decade, to reduce the volume of the Greenland ice sheet by 10%, for various steady-state global warmings with respect to pre-industrial climate

Global temperature rise (°C)	Median time to melt 10% of ice volume (years)
3	1,410
4	900
5	600
6	430

Relative to the average climate of 1980–1999, the warmings would be about 0.5°C less. For each warming, the rate of mass loss is calculated using the range of model sensitivities evaluated by Gregory and Huybrechts (2006), and we show the median of the model range

Since return to pre-industrial CO_2 would take millennia, it is also relevant to ask what ice-sheet steady states are produced under CO_2 concentrations greater than pre-industrial. We investigated this briefly by subjecting the final 80% steady state (specifically, the state which had resulted from the initial 80% state) to climates simulated by the AOGCM for atmospheric CO_2 concentrations of 350 ppm (+0.9°C) and 450 ppm (+2.0°C). The ice sheet remained stable with 350 ppm but commenced a decline with 450 ppm. Such behaviour is consistent with the thresholds indicated in Fig. 1. Evidently, for 450 ppm the corresponding steady state is absent or of smaller volume than for present-day or pre-industrial CO_2. We infer from Ridley et al. (2005) that above some threshold concentration of CO_2, no steady states exist with substantial ice volume. The number and size of ISM steady states is evidently dependent on CO_2 concentration; to quantify them would require a further extensive set of asynchronous integrations.

The idealised simulations conducted here are for a steady state climate and assume no change of insolation. Insolation changes ensure that the ice sheets rarely achieve a true steady state. The results could be somewhat different than the purely anthropogenic impact assessed here, although we note that summer insolation changes over Greenland will be very small during the next 20,000 years on account of a low eccentricity.

The climate model used, HadCM3, has a relatively coarse spatial resolution (105 by 250 km at the central latitude of Greenland). Consequently, climatic conditions at the ice sheet margin in particular, where gradients are large, are simulated rather crudely. This could be influential on ice-sheet–climate feedbacks involving ablation and orographic precipitation. The positive degree-day scheme, though widely used in ice-sheet modelling, makes large simplifications. The coupling methodology between climate and ice sheet models may influence the results. A model intercomparison exercise, especially involving high-resolution climate models, would provide a quantitative assessment of all such uncertainties.

This existence of multiple steady states raises the interesting possibility that the Greenland ice sheet in previous interglacials could have been in a different steady state from in the Holocene. In the last interglacial, the GIS had only 44–72% of its present-day volume (Lhomme et al. 2005). Perhaps this reduction could have resulted in part from the interaction of climate and ice-sheet, rather than from climate change alone.

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