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## Article

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# Transient AMOC weakening during overshoot warming masks committed Greenland ice sheet loss

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## Abstract

Climate mitigation strategies increasingly rely on temporarily exceeding the Paris Agreement’s warming targets (well below 2 °C, preferably 1.5 °C) before drawing temperatures back down through large-scale carbon dioxide removal (CDR). Whether such “overshoot” pathways can protect vulnerable components of the Earth system — including the Greenland Ice Sheet, which holds enough ice to raise global sea level by more than 7 meters — remains unclear. We use one-way coupled Earth system and ice sheet model simulations to show that temporary weakening of the Atlantic Meridional Overturning Circulation (AMOC) during overshoot warming cools the North Atlantic and temporarily mitigates ice loss in southwest Greenland, despite sustained global warming of up to 3 °C. However, once the AMOC recovers, renewed regional warming over Greenland triggers accelerated surface melt under global mean temperatures compatible with the Paris targets. In particular, after AMOC recovery, southwest Greenland becomes highly sensitive to small additional warming above 1.5°C, thus suggesting that this region may have a critical threshold within Paris-compatible temperature limits of 1.5 to 2°C. Our results suggest that, where overshoot pathways trigger substantial AMOC weakening, large-scale Greenland mass loss may emerge only after ocean circulation, and regional warming, recover — meaning AMOC weakening during overshoot would not prevent, but merely postpone, committed ice sheet destabilization.

**Keywords:** Greenland Ice Sheet, AMOC, Overshoot, Melt-elevation feedback

## Introduction

Limiting global warming to well below 2 °C above preindustrial levels, preferably to 1.5 °C, remains a central objective of international climate policy as set in the Paris agreement in 2015 [1]. Yet current emissions and pledged emission reductions remain incompatible with temperature stabilization at these warming levels [2, 3]. Many mitigation pathways therefore rely on temperature overshoot followed by large-scale Carbon Dioxide Removal (CDR) to return global mean temperatures toward Paris-compatible levels later in the century [4]. Although such pathways may ultimately limit long-term warming, they raise major uncertainties regarding irreversible or delayed responses in Earth system components with high inertia such as continental ice sheets [5].

The Greenland Ice Sheet (GrIS) holds  $\sim 7.4$  m of Sea-Level Equivalent (SLE) [6], and its meltwater threatens coastal populations and ecosystems through sea-level rise and potential disruption of the Atlantic Meridional Overturning Circulation (AMOC) [7]. Greenland has already undergone accelerated mass loss over recent decades due to increased surface melt and ocean-induced ice discharge through its marine-terminating outlet glaciers [8]. Numerical model simulations suggest that GrIS mass loss in the 21<sup>st</sup> century largely depends on future emission pathways [9]. Ice sheet models and paleoclimate evidence suggest that the GrIS may exhibit tipping point dynamic associated with the melt–elevation feedback, whereby surface lowering exposes the ice sheet to progressively warmer atmospheric conditions that further enhance melt [5, 10–14]. Model-based estimates of the critical threshold for large-scale GrIS melt fall in the range of 1.5 to 3.2 °C [10, 13, 15, 16].

Since the GrIS responds to climate change on centennial to millennial timescales [5], a temporary exceedance of these warming levels does not necessarily imply immediate ice sheet destabilization. Model simulations showed that stable GrIS long-term configurations can be achieved when overshooting 1.5 °C for a few centuries [17], even when considering peak temperatures well above 3.0 °C [14]. However, these studies are based on simplified ramp-up and ramp-down scenarios in temperature forcing and did not explicitly account for evolving atmosphere–ocean circulation changes associated with coupled carbon emission and removal pathways. One potentially important but poorly constrained process is the response of the AMOC to overshoots pathways [18]. CMIP6 climate models consistently show a weakening of the AMOC in response to anthropogenic warming, owing to polar amplification, increasing freshwater input and reduced deep-water formation in the North Atlantic, although the degree of weakening varies widely between models [19]. Those climate models that do project a strong decline of AMOC also show a strong effect on regional climate over Greenland through altered ocean heat transport, atmospheric circulation and moisture delivery [7, 20]. Because the Greenland Surface Mass Balance (SMB) is sensitive to both melt and snowfall, AMOC changes during transient temperature overshoots may substantially modulate the GrIS response to overshoot pathways.

Here we investigate the GrIS response to centennial-scale overshoots pathways simulated with the Norwegian Earth System Model (NorESM2-LM [21, 22]). NorESM2 belongs to the group of CMIP6 models that shows a strong decline of AMOC in response to anthropogenic forcing, and is therefore well suited to explore impacts of a

potential AMOC weakening on the GrIS. Indeed, the overshoot simulations used here [18, 23] show a strong northern hemisphere cooling around the North Atlantic Subpolar Gyre, which is amplified during the overshoots. These simulations are composed of idealized positive and negative CO<sub>2</sub> emissions phases [23] (Fig. 1a) producing warming pathways with and without temperature overshoot within the range of Paris targets (between 1.5 and 2 °C, Fig. 1b and Table T1). We assess the dynamic response of the GrIS to such NorESM2-LM pathways by forcing the Community Ice Sheet Model v.2 (CISM2 [24]) with SMB [13, 25] and ocean-induced thinning at marine-terminating margins [26]. Although CISM2 is not bi-directionally coupled in our simulations (*i.e.*, changes in ice-sheet geometry and melt-water fluxes do not feed back on climate), we exploit new ice sheet coupling features available in NorESM2-LM (see Materials and Methods) to explicitly account for the impact of the melt-elevation feedback on Greenland surface melt over multiple centuries and millennia [13, 25].

## Results and discussion

### Greenland mass loss under overshoot pathways

Under the reference simulation, in which global warming stabilizes near 1.7 °C above preindustrial without overshoot, the GrIS loses approximately 6 [5-8] cm SLE in 400 years of simulation (Fig. 1, Table T1). Mass loss accelerates throughout the first sixty years before slightly declining and stabilizing (Fig. 1d). By the end of the simulation, GrIS is experiencing more surface melt and ice discharge than under preindustrial conditions (Fig. S1). Simulated mass loss rates under the reference simulation are within the range of observed GrIS mass loss rates over the last two decades (0.3-0.8 mm SLE yr<sup>-1</sup> [27]). Unmitigated high-emission pathways strongly enhance Greenland mass loss. Simulations stabilizing around 3.4 °C global warming produce approximately three-fold increases in cumulative sea-level contribution relative to the reference pathway (Fig. 1, Table T1). Peak rates of Greenland mass loss occur during the rapid warming phase and approach upper-end values projected for high-emission 21st-century scenarios (0.9-2.4 mm SLE yr<sup>-1</sup>, [28]). Overshoot simulations that return temperatures toward Paris-compatible levels after reaching +2.9 °C warming during the peak of the overshoot lead to cumulative GrIS loss  $\sim$  30% to  $\sim$  65% higher than in the case without overshoot, depending on how quick CDR is applied (Fig. 1, Table T1). Spatial patterns of mass loss reveal pronounced regional contrasts (Fig. 2, S2). Most ice loss occurs in western Greenland, particularly in the southwest where surface melt dominates the mass budget (Fig. 3d-e, S3). By contrast, (limited) ice loss in north and southeast Greenland mainly owes to ocean-induced retreat in marine-terminating sectors (Fig. 3b-c, S3). These differences reflect the coexistence of distinct climatic (dry versus wet, high melt versus low melt) and dynamical (land-terminating versus marine-terminating) regimes across Greenland, with some basins primarily controlled by surface melting (southwest Greenland) and others by precipitation or ocean forcing (north and southeast Greenland) [29].

## AMOC weakening temporarily suppresses southwest Greenland melt

While there is a clear dependence of the total GrIS mass loss on both peak and landing warming levels (Fig. 1), there are substantial regional differences in the ice sheet response across Greenland (Fig. 2, S2). In southern Greenland, the ice sheet response is largely modulated by transient changes in North Atlantic circulation (Fig. 3). A defining climatic feature of the overshoot simulations is a pronounced weakening of the AMOC in response to anthropogenic carbon emissions, followed by gradual recovery during and after the negative-emission phase [18] (Fig. 3a). The magnitude and duration of AMOC weakening increase with peak warming, producing strong cooling over the North Atlantic and Greenland once emissions stabilize or CDR is applied (Fig. 3f-m). This regional cooling strongly modifies Greenland climate during the overshoot phase, with air temperature over Greenland remaining substantially cooler than expected from global mean warming alone due to reduced northward ocean heat transport. At the same time, reduced evaporation and atmospheric moisture transport decrease snowfall over parts of Greenland, especially in the southeast. The combination of these processes produces contrasting regional responses for the GrIS. Southwest Greenland, where surface melt dominates the mass budget, experiences a temporary stabilization during periods of weakened AMOC. Following an initial decline during rapid warming, SMB partially recovers as cooler regional conditions suppress melt rates, before dropping again after the AMOC recovers (Fig. 3d). By contrast, southeast Greenland responds differently because accumulation changes dominate over melt changes. Reduced moisture transport during AMOC weakening suppresses snowfall sufficiently to offset the cooling-induced reduction in melt, causing surface mass balance to remain anomalously low even while temperatures drop regionally (Fig. 3b). As a consequence, different Greenland basins evolve asynchronously under the same large-scale climate forcing. These results demonstrate that Greenland sensitivity under overshoot pathways cannot be inferred directly from global mean temperature trajectories alone. Instead, transient ocean circulation changes may strongly modulate regional melt and accumulation regimes.

## Melt acceleration in southwest Greenland after AMOC recovery

The temporary stabilization of southwest Greenland disappears once the AMOC gradually recovers toward present-day strength (Fig. 4a-c). As northward ocean heat transport resumes, regional atmospheric warming over Greenland intensifies despite relatively small changes in global mean temperature, with a  $\sim 2.3$  regional amplification factor over Greenland by the end of the simulations (Table T1). After this transition, surface melt accelerates abruptly in southwest Greenland in both overshoot and non-overshoot pathways. (Fig. 4a-c). In the reference simulation, where warming stabilizes near  $1.7^\circ\text{C}$  above preindustrial without overshoot, southwest Greenland begins to experience renewed surface melt once the AMOC recovers. A similar delayed acceleration emerges in the overshoot simulations, particularly in the long overshoot pathway. The onset of renewed melt acceleration occurs under relatively low additional

global warming above 1.5 °C. Differences of  $\sim 0.1\text{--}0.3$  °C in stabilized global temperature are sufficient to shift SMB in southwest Greenland from above 100 Gt/yr to almost negative values once the AMOC resumes. These results suggest that transient AMOC weakening masks an underlying vulnerability of southwest Greenland under sustained warming above approximately 1.5°C. The delayed nature of this response has important implications for interpreting overshoot pathways. During periods of weakened AMOC, melt rates in southwest Greenland may appear partially stabilized despite continued climatic imbalance. Yet this apparent stabilization does not necessarily imply long-term resilience. Instead, AMOC recovery can unmask committed mass loss that only emerges centuries after peak warming. To assess how much the melt-elevation feedback contributes to accelerate surface melt in southwest Greenland after AMOC recovery, we repeated simulations with prescribed, fixed-elevation SMB (Fig. S4). The abrupt SMB decline coinciding with AMOC recovery occurs at the same timing and warming threshold in these simulations as in the full ensemble, indicating that the onset of accelerated melt is controlled primarily by the resumption of northward ocean heat transport rather than by melt-elevation-driven amplification. The melt-elevation feedback does, however, change the magnitude of surface melt during and after the overshoots (Fig. 4a-c vs Fig. S4), accounting for 25-30% more mass loss over four centuries (Fig. S3).

### Long-term implications for Greenland stability

To evaluate the long-term consequences of the surface melt acceleration in southwest Greenland following the AMOC recovery, we extended the GrIS simulations for 10,000 years using climatological forcing derived from the last three decades of the each NorESM-LM experiment. By this stage, the AMOC had fully recovered and the climate system had approximately stabilized to global mean warming levels between 1.7 and 2 °C above preindustrial (Fig. 1b, Table T1). These global mean warming values are above most recent assessments of a potential threshold for long-term GrIS instability [1, 5, 15, 30]. Under the reference scenario (landing on a global mean temperature of +1.7 °C above preindustrial), long-term Greenland evolution ranges from limited retreat to multi-meter sea-level contribution over millennial timescales (0.1-2.5 m SLE, Fig. 4d). The largest retreat consistently occurs in southwest Greenland (Fig. 4e), where the melt acceleration occurring after AMOC recovery is positively reinforced by the melt-elevation feedback as the ice sheet gets thinner. The short overshoot pathway (+1.8 °C above preindustrial) produces long-term outcomes broadly comparable to the reference case, although with an enhanced retreat in southwest Greenland in some ensemble members (Fig. 4h, S5). The long overshoot extension simulations (+2 °C above preindustrial) show widespread retreat of the western margin leading to ice loss in almost all central Greenland (Fig. 4g, S5) and a sea-level contribution between 3.6 and 4.7 m SLE (Table T1). This abrupt change in the GrIS long-term response occurs for a relatively small difference in the final mitigated global mean temperature compared to the reference and short overshoot scenarios (+0.2 and +0.3 °C, respectively). This change can be explained by the fact that under longer overshoot warming (a) southwest Greenland experiences a prolonged interval of lower SMB (although

mitigated by AMOC decline) before AMOC recovery occurs, thus becoming increasingly preconditioned for accelerated retreat once regional warming intensifies again, (b) after recovery, AMOC is slightly overshooting (Fig. 3a), thus convecting more heat towards southern Greenland. This shows that Greenland stability therefore depends not only on peak warming, but also on the transient evolution of ocean circulation and regional climate feedbacks. Overall, our long-term simulations suggests that southwest Greenland has a limited resilience to warming levels above 1.5 °C, and every decimal point of warming matters even within the Paris agreement limits.

## Summary

Our simulations reveal a strong coupling between ice melt in southern Greenland and transient AMOC changes during temperature overshoots. Importantly, AMOC weakening suppresses surface melt in southwest Greenland by reducing regional atmospheric and oceanic heat transport. However, in the Earth system model used here (NorESM2), this suppression is temporary and disappears once the AMOC recovers. Companion simulations without the melt-elevation feedback confirm that this surface melt acceleration is primarily driven by AMOC recovery rather than a consequence of melt-elevation feedback amplification, although the feedback increases the magnitude of committed GrIS loss. These results challenge a simplified interpretation of overshoot pathways in which returning global temperatures toward Paris-compatible levels stabilizes GrIS melt. Instead, vulnerable sectors of Greenland may experience delayed destabilization centuries after peak warming owing to the interaction between persistent background warming and recovering ocean circulation. The strong sensitivity of southwest Greenland to AMOC recovery further suggests that this region may already lie close to a critical transition [31] under warming levels only slightly above 1.5 °C. Once North Atlantic heat transport resumes, relatively modest additional warming is sufficient to shift southwest Greenland toward an unstable retreat state, with important implications for the ice sheet long-term (in)stability. This delayed emergence of the GrIS instability highlights the importance of regional climate feedbacks in determining the ice sheet resilience. Our findings also imply that future Greenland projections depend strongly on the representation of AMOC changes in Earth system models that provide its climate forcing. NorESM2-LM is among the CMIP models showing a strong AMOC weakening and a pronounced North Atlantic cooling that partially offsets atmospheric warming over Greenland. Models exhibiting weaker AMOC decline or higher transient climate sensitivity to greenhouse gas forcing [32] may therefore simulate earlier or stronger GrIS retreat during overshoot pathways. Better constraining future AMOC evolution therefore remains essential for assessing long-term Greenland stability, as well as exploring feedbacks between GrIS freshwater and the AMOC which this study is lacking and could potentially change the degree and duration of AMOC weakening. Overall, our results indicate that transient ocean-circulation changes can temporarily mask committed ice-sheet retreat under sustained warming. Such delayed responses complicate the interpretation of climate overshoot pathways because apparent stabilization on centennial timescales may conceal longer-term destabilization processes that emerge only after circulation recovery.

## Figures

## Methods

### Climate forcing

The NorESM2-LM [21, 22] simulations used to provide climate forcing over Greenland are described in [18, 23]. NorESM2 [21, 22] is based on the Community Earth System Model (CESM2.1, [33]) and includes atmospheric (CAM6-Nor), oceanic (BLOM), sea ice (CICE), and land surface (CLM5) components. NorESM2 can calculate SMB forcing over Greenland, regardless if the model setup includes an interactive ice sheet component or not. The SMB is calculated in CLM5 as the difference between accumulation (snowfall and refreezing of rainfall/melted snow within the snowpack) and ice loss from surface melt and sublimation at ten virtual Elevation Classes (ECs, 0, 200, 400, 700, 1000, 1300, 1600, 2000, 2500, 3000, and 10000 m), see details in [25]. The ECs calculation is key to resolve steep SMB gradients around the ice sheet margins when downscaling the SMB to higher-resolution ice sheet model domains [34]. The NorESM2 simulations used here have 1 degree resolution in the ocean/sea-ice components and 2 degree resolution in the atmosphere/land components (NorESM2-LM), and are driven by positive and negative bell-shaped CO<sub>2</sub> emission curves [23] similar to those developed for the Zero-Emission Commitment Model Intercomparison Project (ZECMIP) [35]. Here we focus on four main idealized emission pathways: (a) a reference simulation (Ref) where cumulative emissions reach 1500 PgC in the first 100 years and zero emissions afterwards, with no CDR being applied; (b) a high-emission (HiForc) simulation with cumulative emissions of 2500 PgC by year 100 and zero emissions afterwards (same as in Ref, but with higher emission levels); (c) two temperature overshoot simulations branching from the HiForc simulation where cumulative carbon emissions are brought down to the level of the Ref simulation (1500 PgC). The CDR deployment is idealized (no specific method is assumed) and is applied either at year 100, immediately after the peak in cumulative emission (high-short overshoot, HiShOs) or with a hundred years delay at year 200 (high-long overshoot, HiLoOs). For each simulation, we use two ensemble members to force our Greenland ice sheet model (note that in [18, 23] the ensemble members are three, but only for two members forcing data were available).

### Ice sheet model

NorESM2 is equipped with a dynamic Greenland ice sheet component using CISM2 [24]. The coupling between NorESM2 and CISM2 is described in [25] and is similar to those in CESM2.1 [34]. CISM2 is a 3-D thermomechanical ice sheet model which solves a depth-integrated viscosity approximation of the Stokes equations for incompressible viscous flow. Here we use CISM2 at a horizontal resolution of 4 km, using the same model parameters as in ISMIP6 stand-alone projections [28]. Glacial Isostatic Adjustment (GIA) is accounted for in our simulations using the elastic lithosphere-relaxing asthenosphere (ELRA) model [36], which allows for taking account of a regional bedrock response to changes in ice load. Here we run CISM2 within the

NorESM2 architecture, using the same setup as in [13] where SMB forcing at multiple ECs is downscaled onto the 4 km ice sheet model grid during runtime through the NorESM2 coupler (note that CISM2, CLM5 and the coupler are identical in NorESM2 and CESM2.1). This means that while our simulations are not bi-directionally coupled (*i.e.*, ice sheet changes do not feed back onto the climate system), we are able to account for local changes in SMB due ice sheet thinning (melt-elevation feedback). In our simulations, ocean-induced thinning at marine-terminating margins is accounted for using the parametrization proposed in [26], which is forced using NorESM2-LM surface runoff over Greenland (calculated in CLM5) and ocean temperature and salinity around Greenland.

## Simulations setup

**Initialisation.** In our ice sheet model initialisation, we calibrate the spatially varying basal friction coefficient to best match the present-day observed ice sheet elevation, while keeping the outside ice sheet margins to the present day extent (similar as in [25, 37]). We run the ice sheet model spin-up for total 5400 years (5000 years basal friction calibration, 400 years to relax the ice sheet dynamic) forced with SMB from the pre-industrial control NorESM2-LM simulation (Ctrl) in [18, 23]. At the end of our initialisation procedure, we obtain a total GrIS volume of 7.45 m SLE and local ice thickness differences to the present-day observed GrIS not exceeding  $\pm 5$  m (Fig. S7).

**400 years-long simulations.** We force these short-term simulations with year-by-year SMB and ocean-induced thinning from the NorESM2-LM Ref, HiForc, HiShOS, and HiLoOS simulations. For each simulation, we run six different ice sheet model configurations where we systematically vary (a) NorESM2-LM ensemble member [23] and (b) the sensitivity of the ocean forcing parametrization (low-mid-high) [26]. In these simulations, GIA is turned on assuming equilibrium at pre-industrial conditions.

**10,000 years-long extension simulations.** At the end of each 400 years-long simulation, we carry a long-term extension simulation by imposing random sequences of the last 30 years of SMB forcing for a total of 10,000 years. In these extension runs, ocean-induced thinning is no longer applied as the parametrization used here [26] heavily depends on present-day geometry which is no longer preserved in long-term simulations. Since the NorESM2 SMB forcing is provided at multiple ECs, while the forcing files don't change throughout these simulations the SMB varies as the ice sheet topography evolves (melt-elevation feedback).

## Partitioning mass loss

To assess the individual impact of SMB changes, ocean-induced thinning and melt-elevation feedback on the GrIS mass loss we perform three set of experiment for each 400 years-long CISM2 simulation: a first set where we apply only a fixed-elevation SMB forcing (no ocean forcing, no melt-elevation feedback), a second set where we apply both fixed-elevation SMB forcing and ocean forcing, and a third set where we apply SMB forcing at multiple ECs (thus including melt-elevation feedback) and ocean forcing. In Fig. S3, the sea-level contribution of (a) SMB is the mass loss calculated in the first set of experiments, (b) ocean forcing is the mass loss calculated in the

second set of experiments minus those in (a), (c) melt-elevation feedback is the mass loss calculated in the third set of experiments minus those in (b).

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**Data availability.** The NorESM2-LM simulations used as forcing are described in [23] and are available through the Norwegian Research Data Archive/Bjerknes Climate Data Centre and can be accessed under <https://doi.org/10.11582/2022.00012>. The CISM2 forcing files and output supporting the findings of this study, including the data underlying Figs. 1–4 and S1–S7, are stored on the Sigma2 National e-Infrastructure for Research Data (NIRD) and are available on request. The dataset will be made publicly available through the Norwegian Research Data Archive upon publication of the study.

**Code availability.** NorESM2 and CISM2 are open source and available at <https://github.com/NorESMhub/NorESM> and <https://github.com/ESCOMP/CISM>. Scripts used for analysis and figures are stored on the Sigma2 National e-Infrastructure for Research Data (NIRD) and will be made publicly available through the Norwegian Research Data Archive upon publication of the study.

**Author contributions.** M.P., designed the study, performed the ice sheet simulations and analysis, and wrote the manuscript. J.S. performed the NorESM2-LM simulations. H.G. developed the NorESM2–CISM2 coupling and contributed to study design and setup. P.M.L. contributed to study design. All authors discussed the results and edited the manuscript.

**Competing interests.** The authors declare no competing interests.

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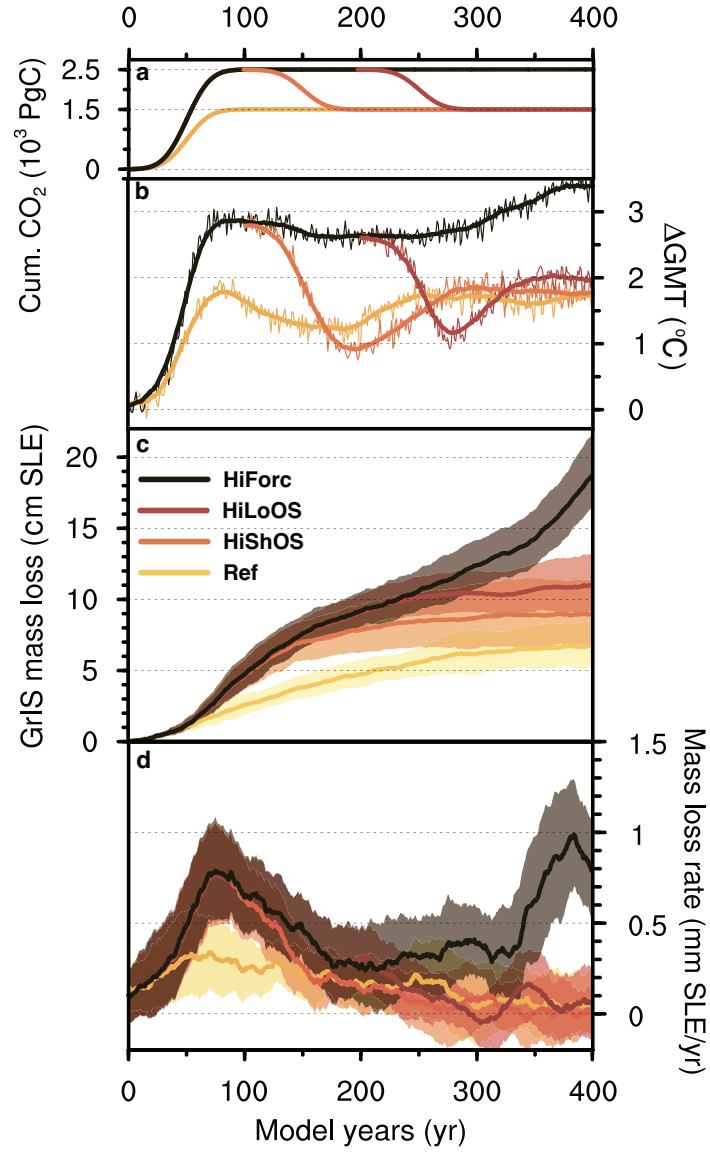


**Fig. 1** (a) Cumulative carbon emissions, (b) global mean temperature anomaly w.r.t. preindustrial (°C), (c) GrIS mass loss (cm SLE), (d) rates of GrIS mass loss (mm/yr SLE) under the reference (Ref), high-emission (HiForc), high-short overshoot (HiShOs), and high-long overshoot (HiLoOs) simulations. Thick lines in (b) show 30 years running mean. In (c) and (d) solid line shows ensemble mean, shading shows ensemble min/max.

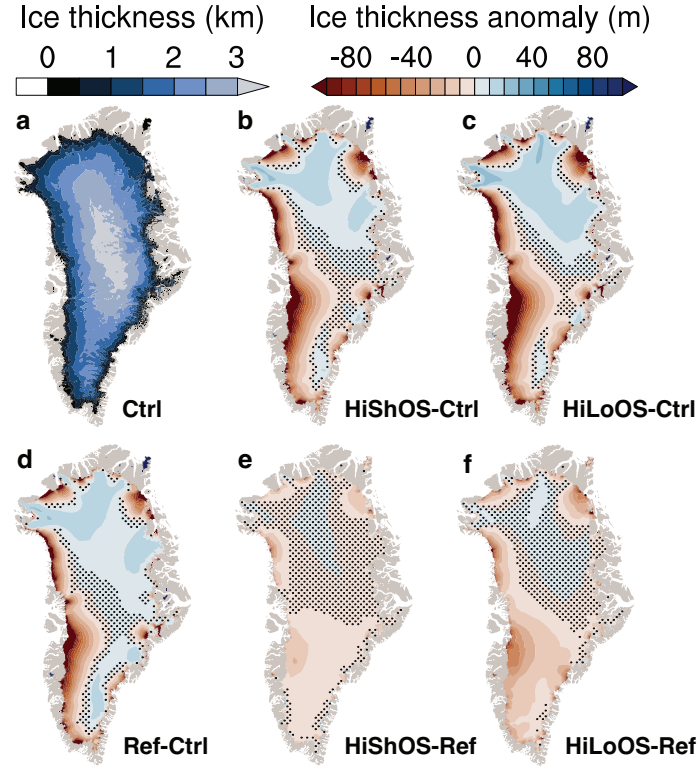


**Fig. 2** Simulated GrIS thickness (km) at the end of the preindustrial control simulation (Ctrl). GrIS thickness anomaly (m) at the end of the reference (Ref), high-short overshoot (HiShOS), and high-long overshoot (HiLoOs) simulations with respect to the (d-f) Ctrl and (g-h) Ref simulations. Grey stippling in (d-h) indicates ice thickness differences lower than 10 m.

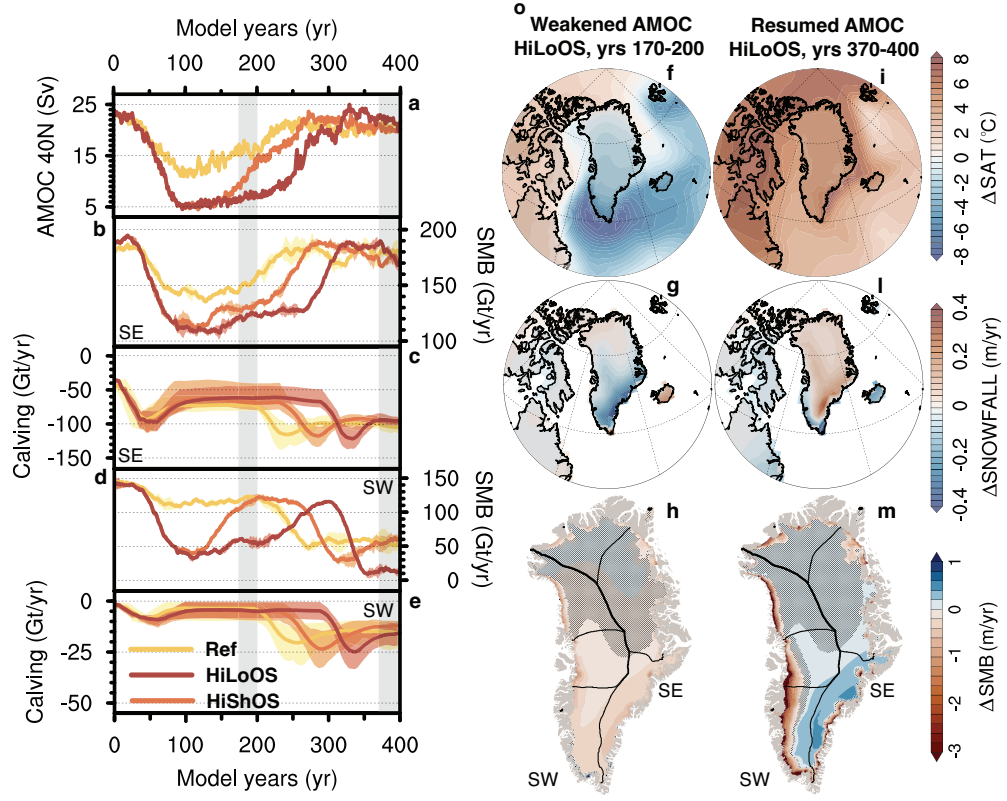


**Fig. 3** (a) AMOC strength at 40°N (Sv), (b) southeast Greenland SMB (Gt/yr), (c) southeast Greenland calving (Gt/yr), (d) southwest Greenland SMB (Gt/yr), (e) southwest Greenland calving (Gt/yr) under the reference (Ref), high-short (HiShOs) and high-long (HiLoOs) overshoot simulations. (f-m) Anomaly to preindustrial of (top row) surface air temperature (°C), (central row) snowfall (m/yr), (bottom row) SMB (m/yr) for two representative climatologies with (left column) weakened AMOC and (right column) resumed AMOC in the HiLoOS overshoot simulation.

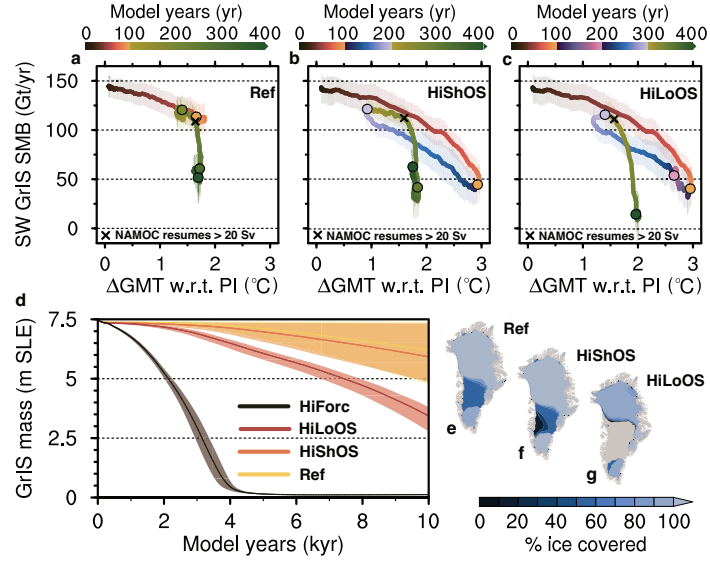


**Fig. 4** (a–c) Southwest Greenland SMB as a function of global mean warming for the (a) reference (Ref), (b) high-short (HiShOs) and (c) high-long (HiLoOs) overshoot simulations. Solid lines show the ensemble mean, shading indicates the ensemble min/max. Colours indicate the temporal evolution of the simulations, progressing through the emission ramp-up phase (red), stable-high phase (purple, only in (c)), ramp-down phase (blue, in (b,c)), and stable-low phase (green). Circles show years 100, 200, 300 and 400 of each simulation, while crosses mark the timing of AMOC recovery. (d) Long-term total GrIS evolution over 10,000 years using climatological forcing derived from the final decades of each simulation. Solid lines show the ensemble mean, shading indicates the ensemble min/max. (e–g) GrIS extent at the end of the 10,000 years long simulations. Colors indicate the percentage of simulations across each ensemble which are ice covered.

## Supplementary Files

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