

1 **Improved Understanding of Multicentury Greenland Ice**  
2 **Sheet Response to Strong Warming in the Coupled**  
3 **CESM2-CISM2 with Regional Grid Refinement**

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9 **Key Points:**

- 10 • For the first time, a variable-resolution atmosphere is coupled with the ocean and  
11 sea ice components in CESM with a dynamic GrIS  
12 • Slower Greenland surface melt increase is detected in the Arctic-refined simula-  
13 tion compared with simulations using a conventional 1° grid  
14 • The steeper VR GrIS surface topography favors slower ablation zone expansion,  
15 leading to weaker albedo feedback and slower melt increase

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## 16 Abstract

17 The simulation of ice sheet-climate interaction such as surface mass balance fluxes  
 18 are sensitive to model grid resolution. Here we simulate the multicentury evolution of  
 19 the Greenland Ice Sheet (GrIS) and its interaction with the climate using the Commu-  
 20 nity Earth System Model version 2.2 (CESM2.2) including an interactive GrIS compo-  
 21 nent (the Community Ice Sheet Model v2.1 [CISM2.1]) under an idealized warming sce-  
 22 nario (atmospheric CO<sub>2</sub> increases by 1% yr<sup>-1</sup> until quadrupling the pre-industrial level  
 23 and then is held fixed). A variable-resolution (VR) grid with 1/4° regional refinement  
 24 over broader Arctic and 1° resolution elsewhere is applied to the atmosphere and land  
 25 components, and the results are compared to conventional 1° lat-lon grid simulations to  
 26 investigate the impact of grid refinement. An acceleration of GrIS mass loss is found at  
 27 around year 110, caused by rapidly increasing surface melt as the ablation area expands  
 28 with associated albedo feedback and increased turbulent fluxes. Compared to the 1° runs,  
 29 the VR run features slower melt increase, especially over Western and Northern Green-  
 30 land, which slope gently towards the peripheries. This difference pattern originates pri-  
 31 marily from the weaker albedo feedback in the VR run, complemented by its smaller cloud  
 32 longwave radiation. The steeper VR Greenland surface topography favors slower abla-  
 33 tion zone expansion, thus leading to its weaker albedo feedback. The sea level rise con-  
 34 tribution from the GrIS in the VR run is 53 mm by year 150 and 831 mm by year 350,  
 35 approximately 40% and 20% smaller than the 1° runs, respectively.

## 36 Plain Language Summary

37 As one of the main contributors to global sea level rise, the Greenland Ice Sheet  
 38 (GrIS) has been losing mass as an accelerating rate during the recent decades. Better  
 39 understanding the interactions between the GrIS and the climate can help us make more  
 40 reliable future projections of GrIS mass loss. To simulate these interactions, a fully cou-  
 41 pled model infrastructure is necessary. Additionally, the model resolution needs to be  
 42 higher enough to resolve the surface topography and processes like orographic precipi-  
 43 tation. This study applies a 1/4° Arctic refined grid to an Earth System Model which  
 44 includes an interactive GrIS model to simulate multicentury GrIS evolution under an ide-  
 45 alized warming scenario, and compares the results with simulations using a lower res-  
 46 olution grid. We show that the simulation with the grid refinement has a slower increase  
 47 in melt, thus contributing less to global sea level rise. This difference mainly results from  
 48 the slower ablation zone expansion and thus weaker albedo feedback in the Arctic re-  
 49 fined grid simulation, with the smaller cloud longwave radiation playing a supporting  
 50 role.

## 51 1 Introduction

52 Recent data reveals an acceleration in the mass loss from the Greenland Ice Sheet  
 53 (GrIS), averaging 257 Gt yr<sup>-1</sup> between 2017 and 2020, a sevenfold increase compared  
 54 to the early 1990s (Otosaka et al., 2023). GrIS mass loss is driven both by atmospheric  
 55 warming (Hanna et al., 2021), which increases surface melt and meltwater runoff (Trusel  
 56 et al., 2018), and oceanic warming, which has caused glacier acceleration and enhanced  
 57 ice discharge (Straneo & Heimbach, 2013). Though the ice discharge increase played a  
 58 stronger role in GrIS mass loss between 1992 and 2018 (66 ± 8%), during the last two  
 59 decades, surface mass balance (SMB) decrease has become the dominant contributor due  
 60 to increased surface melt (Enderlin et al., 2014; Mouginit et al., 2019). The exceptional  
 61 summer surface melting causes a maximum mass loss of 444 Gt yr<sup>-1</sup> in 2019 (Tedesco  
 62 & Fettweis, 2020). The interactions between the ice sheet, the atmosphere, and the ocean  
 63 can initiate feedback effects, further amplifying or dampening the mass imbalance sig-  
 64 nals. One important positive feedback is the albedo/melt feedback. As snow or ice melts,  
 65 the surface with lower albedo, e.g., warmer snow/firn/bare ice/ground, is exposed, lead-

66 ing to increased absorption of shortwave radiation and thus further promoting melt of  
67 the original and nearby regions. Many other feedbacks can enhance or restrain GrIS mass  
68 loss, such as geometry/SMB feedbacks (Fyke et al., 2018). So, to better model the evo-  
69 lution of the GrIS, a coupled model that can represent these bidirectional interactions/feedbacks  
70 is necessary.

71 The accuracy of simulated SMB is sensitive to model grid resolution, especially in  
72 regions with steep and complex terrains. The mountainous GrIS margins, also where steep  
73 topographic gradients are located, are too smooth in conventional  $1^\circ$  to  $2^\circ$  global climate  
74 models (GCMs). Such models fail to resolve processes like orographic precipitation and  
75 allow too much moisture to penetrate into the ice sheet interior, causing positive pre-  
76 cipitation biases (Pollard & Groups, 2000). Research has shown that, with a higher hor-  
77 izontal resolution, the orographic precipitation can be better resolved and thus the posi-  
78 tive precipitation biases can be reduced (van Kampenhout et al., 2019; Herrington et  
79 al., 2022). In addition, the ablation zone around the GrIS margins where the majority  
80 of summer melt occurs can be as narrow as tens of kilometers, which cannot be resolved  
81 by  $1^\circ$  to  $2^\circ$  grids. It is therefore important to have a finer resolution for more accurate  
82 representation of GrIS SMB processes.

83 Modeling with a variable-resolution grid has several advantages. Though rapidly  
84 developing, widespread use of global-uniform high-resolution climate models (e.g., mod-  
85 els participated in the High-Resolution Model Intercomparison Project (HighResMIP;  
86 Haarsma et al., 2016)) is still impractical due to current limits in computational resources.  
87 Regional climate models (RCMs), usually in one-way nesting mode, offer regional high  
88 resolution with a lower computational cost. However, they need boundary conditions from  
89 GCMs or reanalysis, thus not allowing two-way interactions across the boundaries. More-  
90 over, the boundary conditions derived from a separate host model can introduce inconsis-  
91 tencies between the host model and the RCM. Variable-resolution modeling overcomes  
92 some of these challenges using a unified modeling framework, which can model the two-  
93 way interactions between the regional and large scales and is more computationally ef-  
94 ficient.

95 The application of regional grid refinement in GCMs can be dated back to the early  
96 use of stretched grids in late 1970s (Schmidt, 1977; Staniforth & Mitchell, 1978) and now  
97 it has been developed in many state-of-the-art GCMs (Harris et al., 2016; Zängl et al.,  
98 2022; Sakaguchi et al., 2023; Tang et al., 2023). In the Community Earth System Model,  
99 version2 (CESM2; Danabasoglu et al., 2020), regional grid refinement is supported by  
100 the spectral-element (SE; Lauritzen et al., 2018) dynamical core of the atmospheric com-  
101 ponent. Studies have proven its consistency in modeling global circulation and clima-  
102 tology (Zarzycki et al., 2015; Gettelman et al., 2018), fidelity in representing tropical and  
103 extra-tropical cyclones (Zarzycki & Jablonowski, 2014; Zarzycki et al., 2014; Zarzycki,  
104 2016) and regional climate, especially at regions with mountains or steep terrain (Rhoades  
105 et al., 2016; Huang et al., 2016; Huang & Ullrich, 2017; Wu et al., 2017; Rhoades et al.,  
106 2018; Rahimi et al., 2019; Bambach et al., 2022; Wijngaard et al., 2023). The variable-  
107 resolution CESM2 (VR-CESM2) has also been applied to the polar regions. van Kam-  
108 penhout et al. (2019) shows that the simulation of GrIS SMB in the accumulation zone  
109 is significantly improved by using two regionally refined grids over the GrIS at  $1/2^\circ$  and  
110  $1/4^\circ$ . In addition to improvements where a refined resolution is applied to the GrIS, the  
111 simulated clouds and precipitation in the Arctic is also substantially improved with two  
112 Arctic-refined meshes, one at  $1/4^\circ$  and another with an additional  $1/8^\circ$  patch of refine-  
113 ment over Greenland (Herrington et al., 2022). For the Antarctic,  $1/4^\circ$  regional refine-  
114 ment over the Antarctic Ice Sheet and the surrounding Southern Ocean indicates both  
115 improvements, mainly in temperature and wind fields, and degradations, primarily re-  
116 lated to surface melt, over the Antarctic Ice Sheet compared to  $1^\circ$  CESM2 (Datta et al.,  
117 2023). The VR-CESM2 in the above-mentioned studies are run in the coupled land-atmosphere

118 mode following the Atmospheric Model Intercomparison Project protocols (AMIP; Gates,  
119 1992).

120 This study analyzes the results from a set of simulations using the fully coupled  
121 configuration of CESM2 with a dynamic GrIS under an idealized strong warming sce-  
122 nario. A variable-resolution grid, which has  $1/4^\circ$  regional refinement over the broader  
123 Arctic region and  $1^\circ$  horizontal resolution elsewhere, is applied to the atmosphere and  
124 land components of CESM2. Unlike prior VR-CESM2 studies, we include coupling to  
125 a dynamic ocean model, similar to Tang et al. (2023). This work aims to: first, investi-  
126 gate multicentury future GrIS evolution, and second, compare the results of the variable-  
127 resolution run with those of global  $1^\circ$  resolution runs, to see where the regional refine-  
128 ment provides added value. Section 2 documents the model, grids, and experiment de-  
129 sign information. Section 3 presents the results and the comparison of the simulations.  
130 Finally, in Section 4, a discussion and conclusions are provided.

## 131 2 Methods

### 132 2.1 Model Description

133 CESM2 is an Earth System Model (ESM) maintained by the National Center for  
134 Atmosphere Research, which consists of atmosphere, ocean, land, sea ice, and land ice  
135 components and can be run in configurations with different levels of complexity. The ocean  
136 component, Parallel Ocean Program version 2 (POP2; Smith et al., 2010), runs on a nom-  
137 inal  $1^\circ$  displaced-pole grid with 60 vertical levels. Sea ice is represented by the Commu-  
138 nity Ice CodE for sea ice version 5 (CICE5; Hunke et al., 2015), using the same horizon-  
139 tal grid as POP2. Land processes are simulated by the Community Land Model version  
140 5 (CLM5; Lawrence et al., 2019), using the same horizontal grid as the atmosphere model.  
141 CLM5 also embeds the Model for Scale Adaptive River Transport (MOSART; Li et al.,  
142 2013) to handle land surface runoff based on topographic gradients.

143 The GrIS is simulated using the Community Ice Sheet Model, version 2.1 (CISM2.1;  
144 Lipscomb et al., 2019), using a 4-km rectangular grid with 11 terrain-following vertical  
145 levels. To simulate ice flow, a depth-integrated higher-order approximation (Goldberg,  
146 2011) of the Stokes equations is employed in the velocity solver. The parameterization  
147 of basal sliding utilizes a pseudo-plastic sliding law and a simple basal hydrology model,  
148 following the approach described by Aschwanden et al. (2016). In this parameterization,  
149 the yield stress is determined by the till friction angle and the effective pressure, with  
150 the former being influenced by the bedrock elevation through a fixed piecewise linear re-  
151 lationship. The bedrock evolution due to Glacial Isostatic Adjustment (GIA) effect is  
152 governed by an Elastic plate Lithosphere plus Relaxing Asthenosphere (ELRA) model  
153 (see for example Rutt et al., 2009). This study accounts for calving processes through  
154 a flotation criterion, where floating ice is instantaneously discharged to the ocean.

155 Two versions of CESM2 are used for the simulations in this study: CESM2.1 and  
156 CESM2.2. In CESM2.1, the atmosphere is simulated with the Community Atmosphere  
157 Model version 6 (CAM6; Gettelman et al., 2019), using the Finite-Volume (FV; Lin, 2004)  
158 dynamical core, with 32 vertical hybrid pressure-sigma levels. The CAM6 physical pa-  
159 rameterization package is described in detail in Gettelman et al. (2019). CESM2.1 is one  
160 of the ESMs to contribute to the Coupled Model Intercomparison Project Phase 6 (CMIP6;  
161 Eyring et al., 2016) and the Ice Sheet Model Intercomparison Project for CMIP6 (ISMIP6;  
162 Nowicki et al., 2016).

163 CESM2.2 uses the same CAM physics parameterizations and vertical grid, but con-  
164 tains enhanced functionality for the SE dynamical core, including the capability for run-  
165 ning VR grids (Herrington et al., 2022). The CMIP6 CESM2.1 simulations using CAM-  
166 FV are not reproducible in CESM2.2, and therefore two versions of the model are re-  
167 quired to compare the VR grids to the CMIP6 workhorse configuration.

## 168 2.2 Surface Mass Balance

169 The GrIS SMB simulated in CESM2 is the sum of ice accumulation and ice abla-  
 170 tion. The SMB processes of the GrIS are aggregated in CLM5, which includes up to 10  
 171 vertical snowpack layers with a maximum total depth of 10-m water equivalent. Only  
 172 snow accumulated over the 10-m threshold contributes to ice accumulation. Ice ablation  
 173 incorporates surface ice melt as well as sublimation. Part of rain and melt water pen-  
 174 etrates into the snow layers and refreezes, as another source of ice, while the rest runs  
 175 off to the ocean. Melt energy is calculated from the sum of net surface radiation, latent  
 176 and sensible turbulent heat fluxes, and ground heat fluxes at the atmosphere-snow in-  
 177 terface. To account for sub-grid variability, each glaciated grid cell in CLM5 is subdiv-  
 178 ided into 10 elevation classes (ECs) with fixed elevation ranges (Lipscomb et al., 2013;  
 179 Sellevold et al., 2019). The area fractions of the ECs are calculated from the higher-resolution  
 180 CISM2 topography. For each EC, the surface energy fluxes and SMB are calculated in-  
 181 dependently by downscaling atmospheric variables. Near-surface temperature and down-  
 182 ward longwave radiation are downscaled with fixed lapse rates ( $6 \text{ K km}^{-1}$  and  $32 \text{ W m}^{-2}$   
 183  $\text{km}^{-1}$ ). Relative humidity is assumed uniform vertically. Due to a CAM6 model bias lead-  
 184 ing to excessive rainfall over the GrIS, precipitation is repartitioned based on near-surface  
 185 temperature thresholds: precipitation falls as snow when the temperature is below  $-2^\circ\text{C}$ ,  
 186 as rain when the temperature is above  $0^\circ\text{C}$ , and as a linear combination of snow and rain  
 187 for temperatures between  $-2^\circ\text{C}$  and  $0^\circ\text{C}$ .

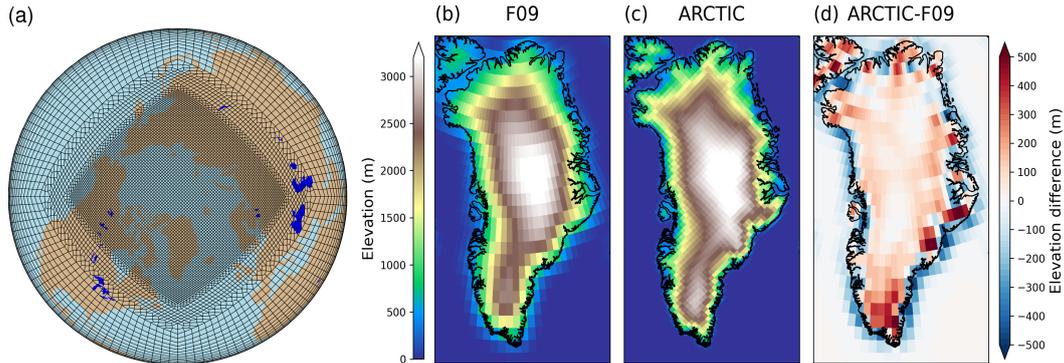
188 Though with biases such as overestimated precipitation, CESM2.1 at  $1^\circ$  with a fixed  
 189 GrIS geometry simulates a realistic historical Greenland SMB (van Kampenhout et al.,  
 190 2020). When coupled to CISM2, higher Greenland SMB and interannual variability is  
 191 simulated, partly due to the different ice sheet topography (Muntjewerf, Petrini, et al.,  
 192 2020).

## 193 2.3 Coupling Scheme

194 In the model framework, the GrIS is interactively coupled to the other Earth sys-  
 195 tem components. CISM2 receives the CLM5 SMB for each EC, and is then downscaled  
 196 by the coupler using a trilinear remapping scheme (bilinear horizontally and linear ver-  
 197 tically), with corrections of accumulation and ablation to conserve global water mass.  
 198 As the ice sheet evolves, the coupler updates the EC fractional glacier coverage in each  
 199 CLM5 grid cell based on the CISM2 ice sheet extent at annual frequency. The mean sur-  
 200 face elevation in CAM6 is manually updated based on the CISM2 topography every 20  
 201 model years, using the CESM topography software (Lauritzen et al., 2015). Surface runoff  
 202 from CLM5, basal melt and ice discharge from CISM2 constitute the freshwater fluxes  
 203 inputted into the ocean, which are supplied as salinity anomalies. More detailed descrip-  
 204 tion of the coupling scheme can be found in Muntjewerf et al. (2021).

## 205 2.4 Grids

206 The variable-resolution grid we use here - the *Arctic* grid (Fig.1a), is a  $1^\circ$  SE grid  
 207 with  $1/4^\circ$  regional refinement over the broader Arctic region (Herrington et al., 2022).  
 208 It is generated by using the software package SQuadgen ([https://github.com/ClimateGlobalChange/  
 209 squadgen](https://github.com/ClimateGlobalChange/squadgen)). The global  $1^\circ$  resolution runs use the latitude-longitude  $1^\circ$  grid, referred to  
 210 as *f09*, supported by the FV dynamical core. Figure 1b and 1c show a snapshot of the  
 211 surface topography over the GrIS before the start of the warming scenario for the *f09*  
 212 grid and the *Arctic* grid, respectively. The *Arctic* grid represents more detailed Green-  
 213 land topography (e.g., the south dome and ice sheet periphery) with a more accurate ice  
 214 sheet mask. The surface elevation differences between two grids can range up over 700  
 215 m (Fig.1d), which is partly caused by the initial ice sheet volume difference. The physics  
 216 time step of the *Arctic* simulations is 450 s, which is a  $4\times$  reduction relative to the de-  
 217 fault 1800 s time step used in the  $1^\circ$  grid.



**Figure 1.** (a) The Arctic grid (Herrington et al., 2022). Note what is shown is the element grid; the computational grid has  $3 \times 3$  independent grid points per element. Surface topography (m) of the Greenland Ice Sheet represented by (b) the f09 grid, (c) the Arctic grid, and (d) their difference before the start of the warming scenario.

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## 2.5 Experiment Design

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First, a pre-industrial simulation was branched off from the last leg of the spun-up pre-industrial Earth system/ice sheet state (Lofverstrom et al., 2020). A series of experiments were ran to bring the top of atmosphere radiative forcing into balance using common tuning parameters such `clubb_gamma` (Guo et al., 2015). The tuned pre-industrial control was then run for 180 years until the GrIS achieved near equilibrium state. Then an idealized warming scenario was started, in which the atmospheric  $\text{CO}_2$  concentration increased by 1% per year until reaching  $4\times$  the pre-industrial value after 140 years, followed by a 210-year simulation with the fixed  $4\times$  pre-industrial  $\text{CO}_2$  concentration (Fig.2a). This set of simulations using the Arctic grid (hereafter ARCTIC) was compared to two sets of simulations under the same forcing but using the f09 grid, in the older CESM2.1 code base. One is from Muntjewerf, Sellevold, et al. (2020) (hereafter F09M). In this code base, to reduce the too-high SMB over portions of the GrIS in CESM2 coupled runs, the cold rain ( $< -2^\circ\text{C}$ ) produced by CAM immediately runs off to the ocean instead of being converted to snow by CLM. Also, to facilitate low level convergence and precipitation near the coasts, the sub-grid roughness over Greenland is artificially increased. To limit the differences between models, we ran another set of f09 simulation using CESM2.1 but without these two adjustments (hereafter F09).

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## 2.6 Analysis

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### 2.6.1 Atmospheric and Oceanic Circulation Metrics

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The Greenland blocking index (GBI) uses the 500-hPa geopotential height ( $Z_{500}$ ) to estimate blocking over the Greenland region (Fang, 2004). Strong and persistent blocking can result in extreme summer melt at the ice sheet surface (Hanna et al., 2014). The revised GBI from Hanna et al. (2018) is used, which is calculated by subtracting the area-averaged  $Z_{500}$  over the Arctic region ( $60^\circ\text{N}$  to  $80^\circ\text{N}$ ) from the area-averaged  $Z_{500}$  over the Greenland region ( $60^\circ\text{N}$  to  $80^\circ\text{N}$ , and  $80^\circ\text{W}$  to  $20^\circ\text{W}$ ). Then the resulting time series is standardized with respect to the last 80 years of the pre-industrial period. Here only the JJA mean GBI is considered.

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The North Atlantic Meridional Overturning Circulation (NAMOC) index measures the strength of the North Atlantic Meridional Overturning Circulation, which is predicted to be weakened with the addition of GrIS meltwater to the ocean (Vizcaino et al., 2010;

249 Muntjewerf, Sellevold, et al., 2020). The NAMOC index is defined as the maximum of  
 250 the overturning stream function north of 28°N and below 500-m depth.

### 251 **2.6.2 Melt/albedo Feedback**

252 We use the melt/albedo feedback (or albedo feedback) calculations following Box  
 253 et al. (2012). The albedo feedback ( $\alpha_{feedback,a}$ ) is quantified by regression between 20  
 254 annual samples of detrended anomalies of summer (JJA) average net shortwave radia-  
 255 tion ( $SW_{net}$ ) and near-surface air temperature ( $T_{air}$ ) in units of  $\text{W m}^{-2} \text{K}^{-1}$ , with anoma-  
 256 lies indicated by the ' character in  $\Delta'$ :

$$\alpha_{feedback,a} = \Delta' SW_{net} / \Delta' T_{air} \quad (1)$$

257 The regression uses annual pairs of anomalies instead of successive values in the  
 258 time series. This pairing is illustrated in Fig.S1. Since this definition of albedo feedback  
 259 does not include lags that may cause albedo change (e.g., a low albedo year pre-conditions  
 260 the next year for low albedo), an alternative formulation of albedo feedback (referred to  
 261 as the bulk albedo feedback ( $\alpha_{feedback,b}$ ) is considered, which is the change in  $SW_{net}$  over  
 262 the change of  $T_{air}$ :

$$\alpha_{feedback,b} = \Delta SW_{net} / \Delta T_{air} \quad (2)$$

### 263 **2.6.3 Equilibrium Line Altitude (ELA)**

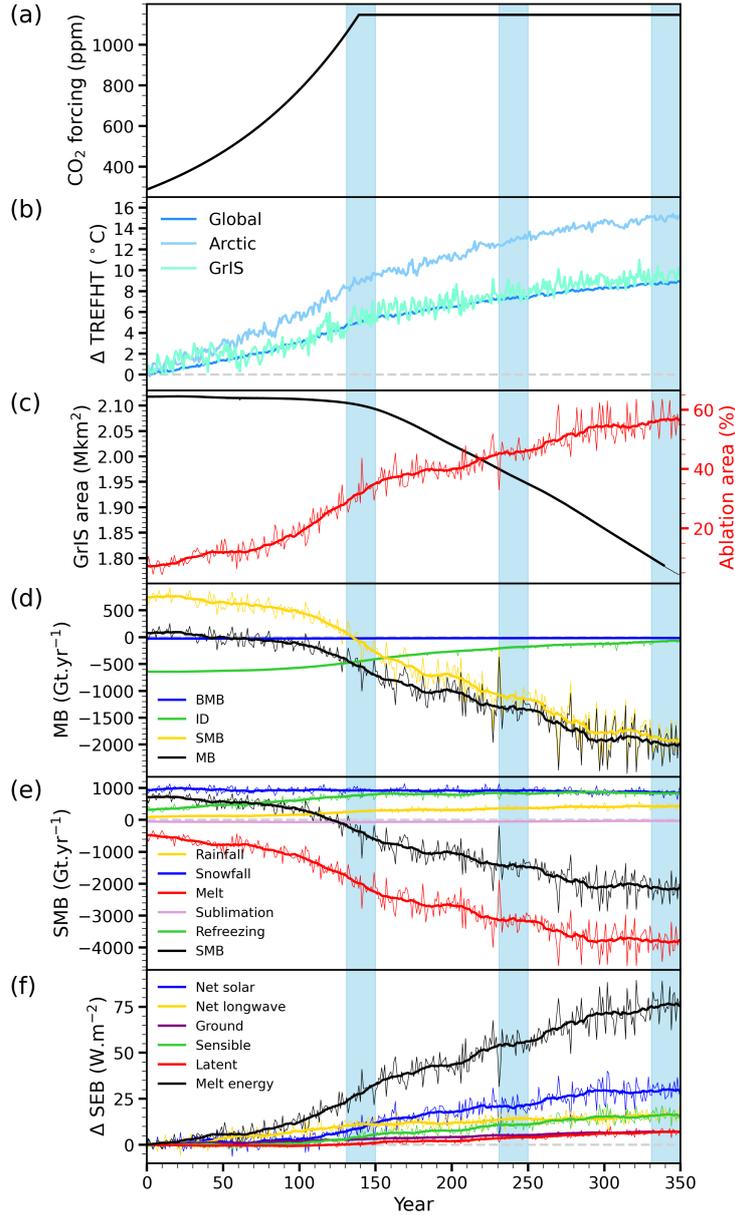
264 Equilibrium line has zero annual mean SMB, separating the ice sheet accumula-  
 265 tion zone and the ablation zone. To calculate the average ELA of the GrIS in our model,  
 266 we define the following algorithm: Loop through the ablation grids using the annual mean  
 267 SMB field. If any of the neighboring grids has positive SMB, compute the elevation where  
 268 SMB equals zero using the two SMB and elevation values of this positive SMB grid (with  
 269 grid area  $a_p$ ) and the ablation grid (with grid area  $a_n$ ). Save this computed elevation  
 270 as one ELA value, with an approximate length of the shared edge of the two grids as the  
 271 weight. The approximate edge length is calculated by  $(\sqrt{a_n} + \sqrt{a_p})/2$ . The final aver-  
 272 age ELA is the length-weighted average of all the saved ELA values.

## 273 **3 Results**

### 274 **3.1 Response of the GrIS in the VR Run**

275 To help visualize the spatial change patterns through time and later analyze the  
 276 differences between F09M, F09, and ARCTIC, we will focus on three time periods: year 131-  
 277 150, year 231-250, and year 331-350. Year 131-150 represents the CO<sub>2</sub> stabilization pe-  
 278 riod. Year 231-250 and year 331-350 are one and two centuries after, respectively, with  
 279 the latter also representing the end of our simulation.

280 Since atmospheric CO<sub>2</sub> radiative forcing is nearly logarithmic in concentration, the  
 281 1% yr<sup>-1</sup> increase of CO<sub>2</sub> causes a nearly linear rise in the annual average near-surface  
 282 temperature (0.3 K per decade; Fig.2b). By CO<sub>2</sub> stabilization, the temperature increase  
 283 is 5.0 K over the GrIS. Polar amplification (the ratio of Arctic and global temperature  
 284 increase) is 1.8. GrIS amplification (the ratio of GrIS and global temperature increase)  
 285 is much smaller (1.1) due to its perennial ice and snow cover. After CO<sub>2</sub> stops increas-  
 286 ing, the annual average near-surface temperature still rises, albeit more slowly (0.2 K per  
 287 decade) due to oceanic warming, adding another 3.8 K by the end of the simulation. The  
 288 relatively small area of the GrIS results in a larger variability of annual average near-  
 289 surface temperature compared to the Arctic and global mean.



**Figure 2.** Evolution of (a) atmospheric CO<sub>2</sub> concentration (ppm), (b) global, GrIS, and Arctic region (north of 60°N) annual near-surface temperature anomaly with respect to the end of pre-industrial period (°C), (c) GrIS total area (left vertical axis, million km<sup>2</sup>) and ablation area (%; as percentage of total ice sheet area), (d) mass balance (MB, black) and components (Gt yr<sup>-1</sup>), (e) SMB (black) and components (Gt yr<sup>-1</sup>), and (f) JJA anomaly of surface energy balance components compared to the end of pre-industrial period (W m<sup>-2</sup>). The thick lines in (c)-(f) show the 20-year running means. The blue shaded periods are used in subsequent analysis.

290 Figure 2d shows the evolution of the GrIS-integrated mass balance and components.  
 291 A similar pattern is found in Muntjewerf, Sellevold, et al. (2020). The mass loss accel-  
 292 erates after about 110 years, rising from 2.4 Gt yr<sup>-2</sup> before year 110 to 13.0 Gt yr<sup>-2</sup> be-  
 293 tween years 111 to 150. Then, after CO<sub>2</sub> stabilization, the GrIS mass loss decelerates

294 gradually, combined with larger oscillations and variability. With the mass loss, the ice  
 295 sheet area shrinks, decreasing from  $2.1 \times 10^6 \text{ km}^2$  to  $1.8 \times 10^6 \text{ km}^2$  (Fig.2c). The cu-  
 296 mulative contribution to global mean sea level rise (GMSLR) is 53 mm by year 150 and  
 297 831 mm by year 350 (Table 1). The decreasing SMB dominates the mass balance trend  
 298 (Fig.2d), which becomes negative at around year 130. Ice discharge gradually decreases  
 299 as marine-terminating outlet glaciers thin, decelerate, and even transition to land-terminating  
 300 glaciers.

**Table 1.** Annual Rate of Mass Loss ( $\text{mm yr}^{-1}$ ), Cumulative GrIS Mass Loss (mm), Mass Balance Components ( $\text{Gt yr}^{-1}$ ), GrIS Area ( $10^6 \text{ km}^2$ ), and GrIS Volume ( $10^6 \text{ km}^3$ ).

	Last 20 yrs of CTRL			Years 131-150			Years 231-250			Years 331-350		
	F09M	F09	ARCTIC	F09M	F09	ARCTIC	F09M	F09	ARCTIC	F09M	F09	ARCTIC
Annual mass loss	-0.04	-0.10	-0.05	2.08	2.06	1.48	5.28	4.49	3.50	6.36	5.93	5.40
Cumulative mass loss	-0.8	-2.0	-1.1	97	84	53	501	447	344	1,098	976	831
MB	19	41	23	-776	-761	-542	-1,974	-1,669	-1,285	-2,376	-2,195	-2,001
SMB	616	723	685	-380	-319	-72	-1,797	-1,463	-1,081	-2,284	-2,097	-1,909
ID	573	654	636	376	420	448	161	187	187	78	81	77
BMB	-24	-27	-25	-20	-22	-22	-16	-18	-18	-14	-16	-16
GrIS area	1.97	2.00	2.02	1.92	1.96	1.99	1.77	1.80	1.83	1.60	1.64	1.66
GrIS volume	3.23	3.27	3.25	3.20	3.24	3.23	3.05	3.10	3.12	2.81	2.89	2.93

Notes: Mass Balance (MB) = Surface Mass Balance (SMB) - Ice Discharge (ID) + Basal Melt Balance (BMB). Variables in this table are calculated using CISM2 outputs.

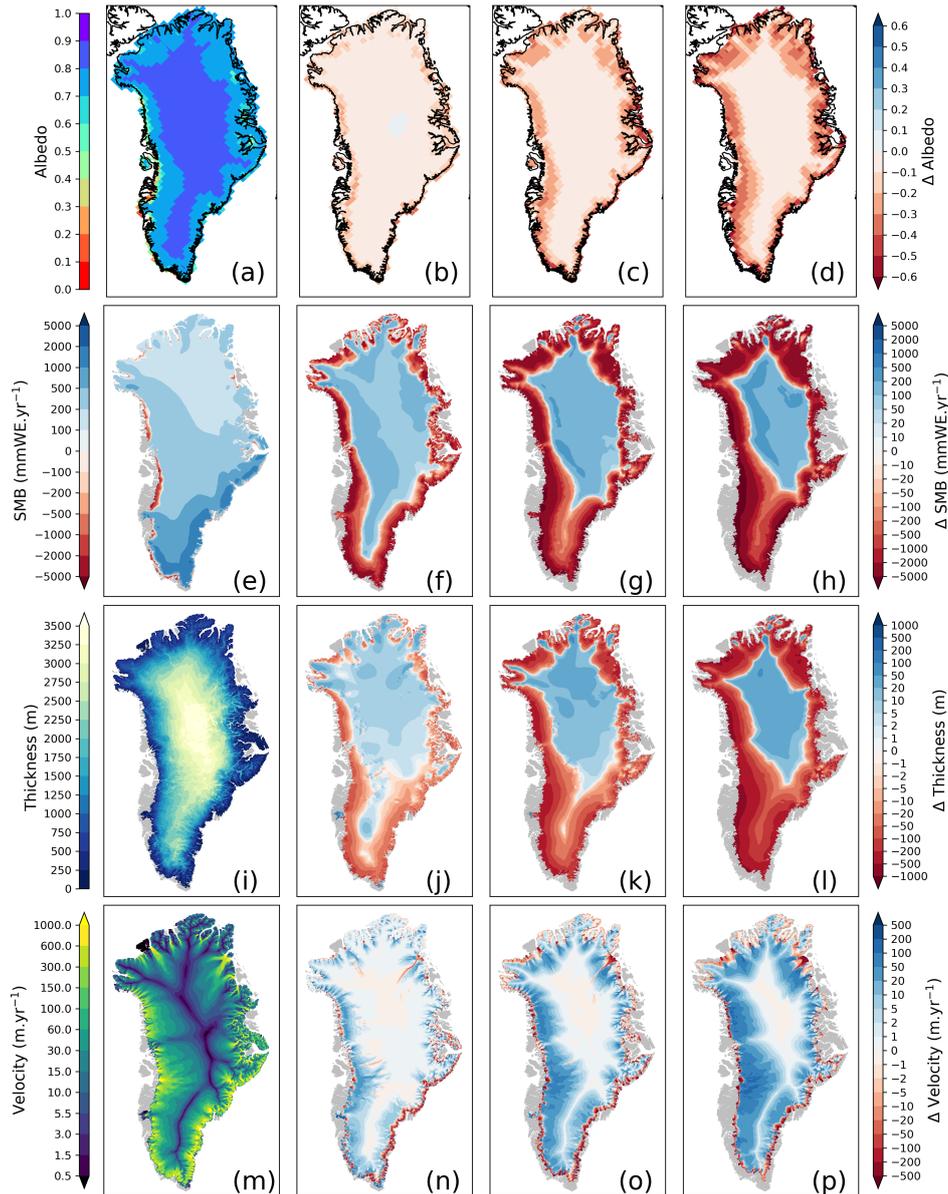
301 The accelerating SMB trend (decreasing) is dominated by the surface melt trend.  
 302 Figure 2e shows the evolution of SMB components. Surface melt increases profoundly:  
 303 the annual surface melt during year 131-150 and year 331-350 are more than four times  
 304 and eight times larger than the pre-industrial value, respectively (Table 2). The total  
 305 precipitation increases from the start due to increased rainfall but decreases during the  
 306 last century as the decreasing trend of snowfall dominates. By the end of the simulation,  
 307 the precipitation is about a quarter higher than the pre-industrial value (Table 2). Though  
 308 snowfall remains the major precipitation type, the proportion of liquid phase precipita-  
 309 tion gradually grows through time, making up nearly one-third of the total precipita-  
 310 tion by the end of the simulation (Table 2). Refreezing increases before  $\text{CO}_2$  stabiliza-  
 311 tion as more liquid water is available from both increased rainfall and surface melt, and  
 312 then becomes relatively stable due to saturated snow cover. This is also reflected by the  
 313 refreezing capacity - defined as the fraction of refreezing to available liquid water - de-  
 314 creasing from 55.4% pre-industrial to 19.7% by the end of the simulation (Table 2). Sub-  
 315 limation is relatively small throughout the simulation, so is not discussed further.

**Table 2.** Annual Ice Sheet-Integrated Surface Mass Balance and Components Mean ( $\text{Gt yr}^{-1}$ ).

	Last 20 yrs of CTRL			Years 131-150			Years 231-250			Years 331-350		
	F09M	F09	ARCTIC	F09M	F09	ARCTIC	F09M	F09	ARCTIC	F09M	F09	ARCTIC
SMB (4km)	616	723	685	-380	-319	-72	-1,797	-1,463	-1,081	-2,284	-2,097	-1,909
SMB	-	701	651	-745	-620	-369	-2,213	-1,752	-1,398	-2,552	-2,254	-2,159
Precipitation	942	1,026	955	1,047	1,273	1,200	1,106	1,374	1,277	1,156	1,308	1,265
Snowfall	850	934	869	782	961	930	699	932	913	695	819	849
Rain	92	92	86	265	312	270	406	443	364	461	488	416
Refreezing	142	295	307	680	872	758	784	956	824	781	1,019	830
Melt	485	476	468	2,147	2,400	1,986	3,662	3,606	3,083	4,009	4,073	3,806
Sublimation	-	52	57	60	53	70	34	34	52	19	19	32
Rain (%)	9.8	9.0	9.0	25.4	24.5	22.5	36.8	32.2	28.5	39.9	37.4	32.9
Refreezing (%)	24.6	51.9	55.4	28.2	32.2	33.6	19.3	23.6	23.9	17.5	22.3	19.7

Notes: SMB (4km) is calculated using CISM2 outputs. The other variables are calculated using CLM5 outputs. SMB = Snowfall + Refreezing - Melt - Sublimation. Rain (%) = Rain \* 100 / (Snowfall + Rain). Refreezing (%) = Refreezing \* 100 / (Rain + Melt).

316 The enhanced surface melt greatly reduces the SMB over the periphery of the GrIS.  
317 Fig.3f-h show the spatial distribution of SMB difference compared to the pre-industrial  
318 period, indicating the largest SMB decrease occurs along the southeast and west mar-  
319 gins ( $> 5,000 \text{ mm yr}^{-1}$  by the end of the simulation). In the ice sheet interior, SMB in-  
320 creases due to the locally enhanced hydrological cycle. This spatial pattern of SMB changes  
321 results in a similar spatial pattern of the ice sheet thickness changes, with slight thicken-  
322 ing over the high interior and significant thinning towards the margins (Fig.3j-l). The  
323 largest thinning along the western margins can exceed 1000 m by the end of the simu-  
324 lation. The surface ice velocity decreases at the ice sheet margins due to the thinning  
325 of outlet glaciers, though generally, the surface ice velocity increases due to the steeper  
326 surface slopes caused by the thickness changes (Fig.3n-p).



**Figure 3.** Spatial distribution over the GrIS for pre-industrial (first column) and differences with respect to the former by model years 131-150 (second column), 231-250 (third column) and 331-350 (fourth column). (a-d) JJA mean albedo, (e-h) annual mean surface mass balance ( $\text{mmWE yr}^{-1}$ ) with accumulation zones ( $\text{SMB} > 0$ ) and ablation zones ( $\text{SMB} < 0$ ), (i-l) ice sheet thickness (m), and (m-p) surface velocity ( $\text{m yr}^{-1}$ ).

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### 3.2 Drivers for melt changes

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The spatiotemporal evolution of surface energy balance (SEB) components over the GrIS provides an explanation for the main physical drivers of accelerating surface melt. Figure 2f shows the summer (JJA) average anomalies of surface energy terms compared to the pre-industrial period. During the first century, the enhancement in net longwave radiation (less longwave cooling) from increased atmospheric temperature and cloudiness (Fig.S2a) provides most of the additional melting energy. The increased net short-

334 wave radiation is dampened by reduced incoming shortwave radiation (Fig.S2b) result-  
 335 ing from enhanced cloudiness (Fig.S2a). At around year 110, the contribution of net short-  
 336 wave radiation to the melt energy increases rapidly, surpassing net longwave radiation  
 337 at around year 148, and becoming the dominant melt energy contributor. At the end of  
 338 the simulation, net shortwave radiation provides 37.3% of the total additional melt en-  
 339 ergy (Table 3). The accelerated net shortwave radiation increase at about year 110 is  
 340 speculated to be due to the activation of the melt/albedo feedback. We find that the ac-  
 341 celerated net shortwave radiation and melt increase coincides with a faster increase of  
 342 ablation area (Fig.2c). As the ablation zone expands, darker surfaces with lower albedo  
 343 are exposed, which absorbs more shortwave radiation and further enhances surface melt-  
 344 ing. The decrease of surface albedo mainly occurs within the ablation zones, with lit-  
 345 tle change in the ice sheet interior (Fig.3b-d). The ice sheet hypsometry also contributes  
 346 to the accelerated expansion of the ablation zone - the quasi-parabolic shape favors faster  
 347 ablation area expansion as it approaches the interior plateau. The turbulent fluxes also  
 348 increase at around year 110, although at slower speeds. As near-surface air temperature  
 349 continues to rise, the surface temperature inversion becomes stronger, thus enhancing  
 350 the turbulent fluxes. The contribution of sensible heat flux to total additional melt en-  
 351 ergy is comparable to the contribution of net longwave radiation by the end of the sim-  
 352 ulation (22.5% and 21.5%, respectively; Table 3). The latent heat flux becomes less neg-  
 353 ative through the simulation, which is potentially due to the decrease of sublimation as  
 354 well as an increase in deposition and condensation. At the end of the simulation, 10.1%  
 355 of the additional melt energy comes from latent heat flux (Table 3). The ground heat  
 356 flux rises from negative to positive, which is potentially a result of more refreezing in the  
 357 snow layer, and contributes 8.6% of the additional melt energy at the end of the sim-  
 358 ulation (Table 3).

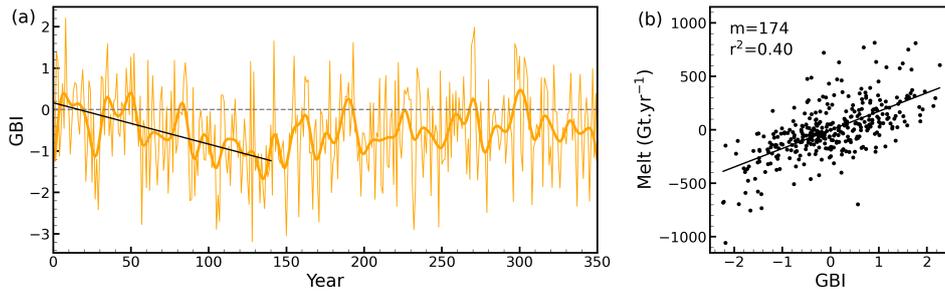
**Table 3.** Summer GrIS-averaged Albedo, Near-Surface Temperature and Skin Temperature (°C), Incoming Shortwave Radiation, Incoming Longwave Radiation at the Surface, and Surface Energy Balance Components Mean ( $W m^{-2}$ ).

	Last 20 yrs of CTRL			Years 131-150			Years 231-250			Years 331-350		
	F09M	F09	ARCTIC	F09M	F09	ARCTIC	F09M	F09	ARCTIC	F09M	F09	ARCTIC
Albedo	0.78	0.78	0.78	0.71	0.72	0.73	0.64	0.66	0.68	0.61	0.62	0.65
$T_{2m}$	-7.00	-7.05	-7.34	-1.22	-1.11	-1.88	0.50	0.43	-0.22	1.03	1.21	0.79
$T_{skin}$	-7.83	-7.83	-8.42	-2.14	-2.15	-3.10	-0.74	-0.85	-1.74	-0.31	-0.20	-0.96
$SW_{in}$	288.3	286.4	294.3	262.0	261.7	272.6	255.9	254.9	266.6	250.9	253.4	264.0
$LW_{in}$	231.0	232.6	224.1	267.7	268.5	257.8	276.1	276.8	266.7	280.3	280.7	271.6
Melt energy	8.8	8.5	8.4	40.3	43.4	35.7	71.0	68.3	59.7	84.6	83.6	80.0
$SW_{net}$	64.0	62.1	64.9	73.5	72.8	73.5	89.5	84.9	83.2	94.8	92.8	91.6
$LW_{net}$	-50.6	-49.0	-55.3	-38.2	-37.3	-44.0	-36.1	-34.9	-41.2	-34.1	-34.1	-39.9
SHF	4.8	4.5	7.1	8.4	10.4	11.2	14.2	14.7	17.6	16.1	16.7	23.3
LHF	-8.3	-8.0	-7.5	-6.9	-5.4	-7.1	-2.7	-2.3	-3.6	-0.6	0.1	-0.3
GHF	-1.2	-1.1	-0.8	3.6	2.9	2.1	6.1	5.9	3.8	8.4	8.2	5.3

Notes: Melt energy = net shortwave radiation  $SW_{net}$  + net longwave radiation  $LW_{net}$  + sensible heat flux SHF + latent heat flux LHF + ground heat flux GHF.

359 To examine the role of large scale circulation on summer GrIS surface melt, we cal-  
 360 culated the GBI. Figure 4a shows the evolution of the GBI for ARCTIC. There is a neg-  
 361 ative trend of the GBI before CO<sub>2</sub> stabilization, which indicates weakening summer block-  
 362 ing over the Greenland region. This agrees with the result of Sellevold and Vizcaíno (2020),  
 363 which uses an AMIP style configuration of CESM2.1 under the same 1% yr<sup>-1</sup> CO<sub>2</sub> warm-  
 364 ing scenario. After CO<sub>2</sub> stabilization, there is no significant trend for GBI. Figure 4b shows  
 365 the linear regression between GrIS-integrated JJA melt, and the JJA GBI. Both vari-  
 366 ables are filtered with a 10-year high-pass filter, thus representing the sub-decadal timescale.  
 367 On sub-decadal timescales, the GBI explains 40% of the annual variability of summer  
 368 surface melt. A more negative GBI results in less surface melt. Therefore, this indicates  
 369 the GBI (more general the atmospheric circulation pattern) is not a driver of the melt

370 acceleration at about year 110, but actually counteracts part of the effect of global warm-  
 371 ing on the surface melt before CO<sub>2</sub> stabilization.

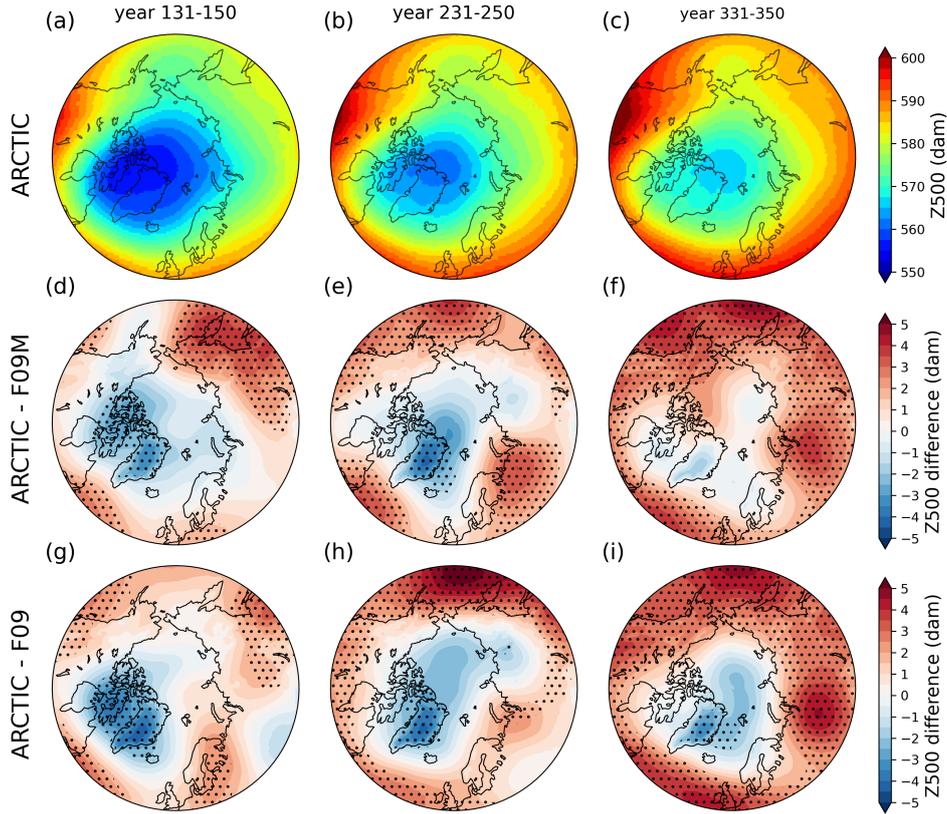


**Figure 4.** Evolution of the Greenland Blocking Index for ARCTIC (a), and the regression of JJA GrIS-integrated filtered surface melt ( $\text{Gt yr}^{-1}$ ) onto JJA filtered Greenland Blocking Index (b). The thick orange line in (a) shows the 10-year low-pass filtered time series. The timescale of the filtered quantities effectively removes both the mean and the trend of each time series. Black line is drawn where the regression is significant, with an annotated  $m$  (slope), and  $r^2$  for the explained variance for (b).

### 372 3.3 Impacts of Enhanced Resolution

#### 373 3.3.1 Large-scale Climate

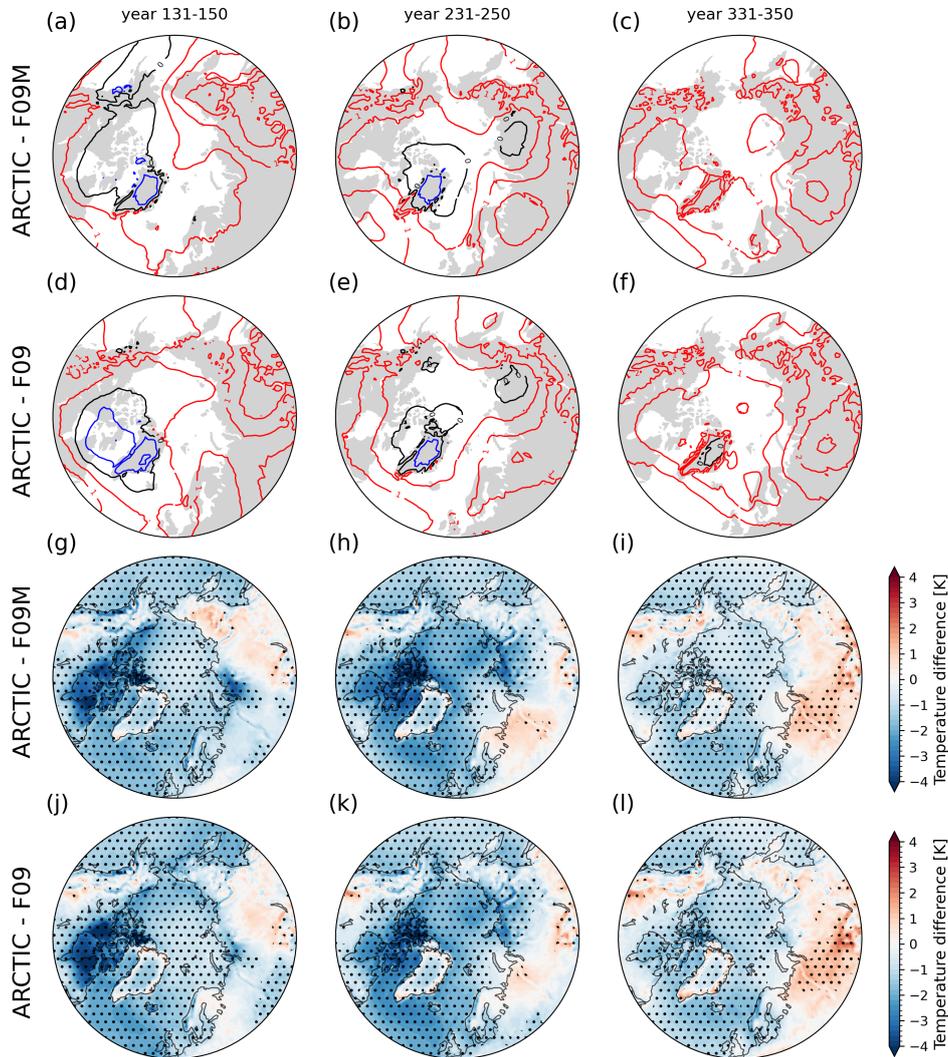
374 Before delving into the simulated GrIS responses, we first compare the represen-  
 375 tations of large-scale climate conditions between the regionally-refined ARCTIC run and  
 376 the  $1^\circ$  runs. Figure 5a-c show the summer mean 500 hPa geopotential height for ARCTIC  
 377 during the three time periods - year 131-150, year 231-250, and year 331-350. As the at-  
 378 mosphere warms, the 500 hPa geopotential height increases over the northern high lat-  
 379 itudes. The differences between ARCTIC and the  $1^\circ$  runs show a consistent pattern dur-  
 380 ing each period (Fig.5d-i). Compared to F09M and F09, ARCTIC has significantly lower  
 381 500 hPa geopotential height over Greenland except for the period year 331-350 for ARCTIC  
 382 and F09M, which possibly indicates weaker blocking and less anticyclonic flows over the  
 383 GrIS in ARCTIC. This lower geopotential region in ARCTIC extends over the Canadian Archipelago  
 384 during year 131-150. There is also significant higher geopotential over subpolar regions  
 385 in ARCTIC compared to F09M and F09, especially during year 331-350.



**Figure 5.** Northern hemisphere summer 500 hPa geopotential height (dam) of ARCTIC (a-c), and the difference between ARCTIC and F09M (d-f), ARCTIC and F09 (g-i). The three columns from left to right represent averaged periods year 131-150, year 231-250, and year 331-350, respectively. Dotted regions are where the two simulations are significantly different ( $p < 0.05$ ) by student t test.

386 Figure 6 compares the summer temperature over the northern hemisphere between  
 387 ARCTIC and the 1° runs. The lower-troposphere summer virtual temperature, computed  
 388 by equating a layer mean virtual temperature with the 500–1,000 hPa geopotential thick-  
 389 ness, is higher in ARCTIC over much of the northern hemisphere (Fig.6a-f). This agrees  
 390 with the common response to increasing horizontal resolution (also reducing physics time  
 391 step) in general circulations models (GCMs) (Pope & Stratton, 2002; Roeckner et al.,  
 392 2006) including CAM (Herrington & Reed, 2020): increasing the horizontal resolution  
 393 warms the climate, since higher resolved vertical velocities generate more condensational  
 394 heating. However, F09M and F09 have warmer lower troposphere than ARCTIC centered  
 395 over the GrIS, which also extends to Canadian Archipelago and Alaska during year 131-  
 396 150 (Fig.6a,d) and to East Siberia during year 231-250 (Fig.6b,e). This spatial pattern  
 397 of lower-troposphere virtual temperature differences agrees with that of the 500 hPa geo-  
 398 potential height difference, indicating that probably the stronger summer blocking over Green-  
 399 land regions in F09M and F09 causes the higher lower-troposphere virtual temperature  
 400 there. Compared to the lower troposphere, the near-surface temperature difference shows  
 401 a distinct spatial pattern with larger magnitudes (Fig.6g-l). Except for some terrestrial  
 402 regions (e.g., parts of Siberia and Eurasia), ARCTIC is significantly cooler than F09M and  
 403 F09 at near-surface level. This results from the different pre-industrial climate, in which  
 404 ARCTIC is cooler, likely due to increasing the albedo of snow over sea ice to tune this run.  
 405 The regions where the lower troposphere is cooler in ARCTIC during year 131-150 and year

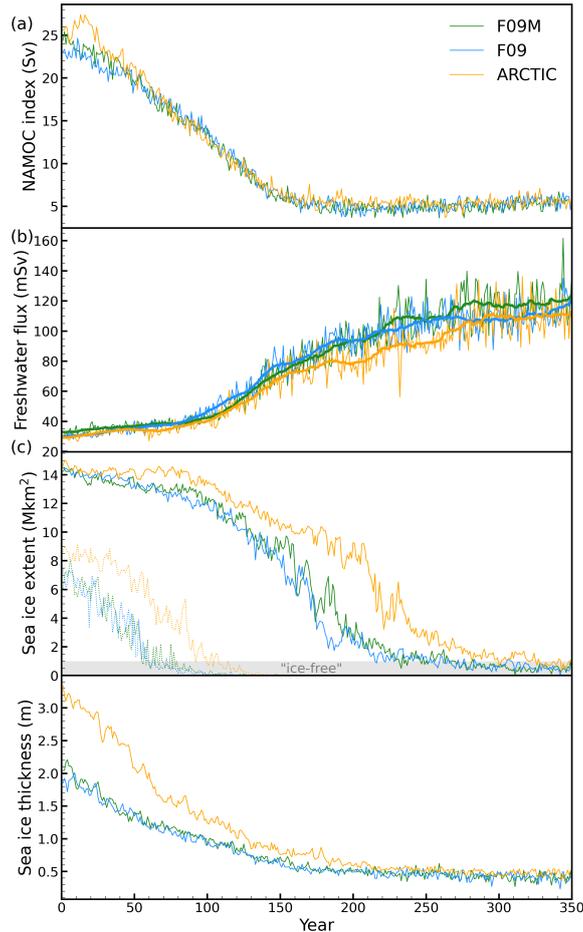
406 231-250 also have much lower near-surface temperature. However, the near-surface tem-  
 407 perature differences over the GrIS are much smaller due to its perennial ice and snow  
 408 cover. There are also regions at the ice sheet periphery that are warmer in ARCTIC. This  
 409 is caused by the difference of cloud conditions, which is discussed in Section 3.3.3.



**Figure 6.** Northern hemisphere summer lower troposphere virtual temperature differences (K) between ARCTIC and F09M(a-c), ARCTIC and F09(d-f). Lower troposphere layer mean virtual temperature is derived from the 1,000–500 hPa geopotential thickness, using the hypsometric equation. Northern hemisphere summer near-surface air temperature differences (K) between ARCTIC and F09M(g-i), ARCTIC and F09(j-l). The three columns from left to right represent averaged periods year 131-150, year 231-250, and year 331-350, respectively. Dotted regions are where the two simulations are significantly different ( $p < 0.05$ ) by student t test.

410 The NAMOC evolution is insensitive to the enhanced resolution in the atmosphere.  
 411 The three simulations have very similar NAMOC index, and even their initial differences  
 412 are diminished. During the period of CO<sub>2</sub> increase, the NAMOC weakens significantly  
 413 (Fig.7a, with the NAMOC index decreasing by 0.14 Sv yr<sup>-1</sup>, 0.13 Sv yr<sup>-1</sup>, and 0.12 Sv

414  $\text{yr}^{-1}$  in ARCTIC, F09M, and F09, respectively. Then the NAMOC index gradually stabi-  
 415 lizes and remains at about 5 Sv for almost two centuries. We also compare the evolu-  
 416 tion of the NAMOC index with the total freshwater flux from the GrIS into the ocean  
 417 (Fig.7b). The total freshwater flux is calculated by adding up surface runoff, basal melt,  
 418 and solid ice discharge. The NAMOC index declines at the start of  $\text{CO}_2$  increase, which  
 419 is much earlier than the rapid increase freshwater flux at around year 110, and remains  
 420 stable when the freshwater flux keeps increasing. This relationship is also found in simu-  
 421 lations under SSP5-8.5 forcing (Muntjewerf, Petrini, et al., 2020), which suggests the  
 422 relatively limited role of additional freshwater input from the GrIS on NAMOC weak-  
 423 ening compared to global warming in CESM2.



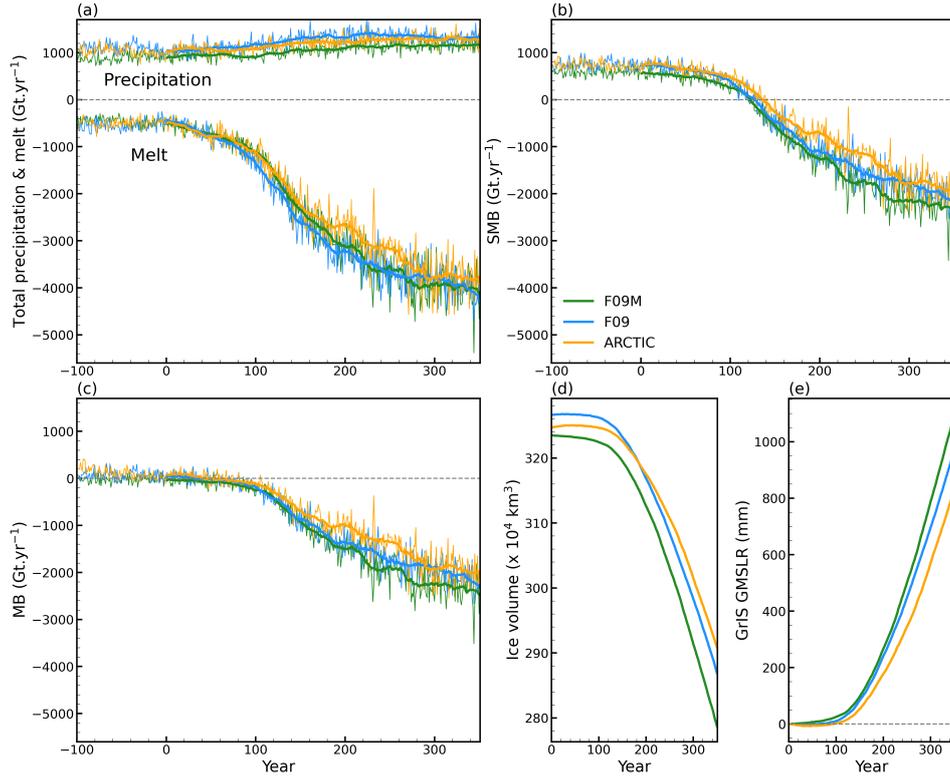
**Figure 7.** Evolution of the annual mean (a) NAMOC index (Sv), (b) total freshwater flux into the ocean (mSv), (c) Northern Hemisphere sea ice extent (million  $\text{km}^2$ ; ice concentration  $> 15\%$ ), and (d) averaged sea ice thickness (m). The thick lines in (b) represent 20-year running means. The dotted lines in (c) represent September mean sea ice extent and the gray shaded range represents the "ice-free" condition.

424 Slower decline of Northern Hemisphere sea ice extent is found in ARCTIC compared  
 425 to the  $1^\circ$  runs (Fig.7c). In the  $1^\circ$  runs, the Arctic becomes ice-free (sea ice extent  $<$  one  
 426 million  $\text{km}^2$ ) in September after about 60 years, while this happens about five decades  
 427 later in ARCTIC. There is an acceleration of annual sea ice decline ((Fig.7c)) at around

428 year 90 for all the simulations. This is possibly because the ice-free Arctic ocean absorbs  
 429 more radiation, which stores more heat in the ocean and slows down sea ice formation  
 430 in colder seasons, but the reason for the timing needs further investigation. Before the  
 431 Arctic becomes ice-free all year round by the end of the simulation, ARCTIC has larger  
 432 sea ice extent than the  $1^\circ$  runs due to its slower sea ice decline during the first two cen-  
 433 turies. This is mostly driven by the larger initial sea ice thickness in ARCTIC of the warm-  
 434 ing scenario (Fig.7d), which is a result of increasing the albedo of snow over sea ice in  
 435 ARCTIC during the tuning process.

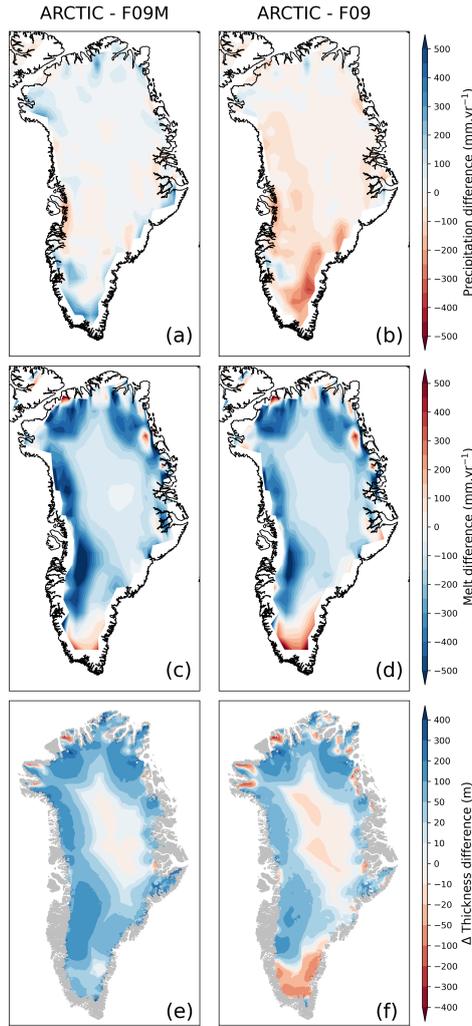
### 436 *3.3.2 SMB Evolution and Ice Sheet Changes*

437 The evolution of ice sheet-integrated quantities in the three simulations follows a  
 438 similar pattern but with different magnitudes and timing. Figure 8b compares the evo-  
 439 lution of the GrIS-integrated SMB of the three simulations. Compared to the  $1^\circ$  runs,  
 440 the SMB in ARCTIC decreases more slowly, such that by the end of the simulation, the  
 441 drop of annual SMB in ARCTIC is  $306 \text{ Gt yr}^{-1}$  and  $226 \text{ Gt yr}^{-1}$  ( $\sim 10\%$ ) smaller than F09M  
 442 and F09, respectively (Table 1). This SMB difference mainly results from the difference  
 443 in melt, with ARCTIC generally producing less surface melt than F09M and F09, especially  
 444 around the period year 180-260 (Fig.8a). During year 231-250, the GrIS-integrated melt  
 445 in F09M and F09 is  $579 \text{ Gt yr}^{-1}$  and  $523 \text{ Gt yr}^{-1}$  larger than ARCTIC, respectively (Ta-  
 446 ble 1). The difference in precipitation is relatively small (Fig.8a). F09 has larger precip-  
 447 itation than ARCTIC, which compensates some of its larger melt and results in smaller  
 448 SMB difference. In F09M, the adjustments that direct cold-rain to surface runoff and in-  
 449 crease sub-grid roughness over Greenland greatly reduce precipitation. Compared to F09,  
 450 it causes a  $204 \text{ Gt yr}^{-1}$  reduction of annual total precipitation averaged over the 350 years.  
 451 The slower SMB decrease of ARCTIC causes slower decrease of MB (Fig.8c), and thus slower  
 452 ice volume decrease (Fig.8d) and smaller global mean sea level rise (GMSLR) contribu-  
 453 tion (Fig.8e). Over the whole 350 years, F09M and F09 contribute to 267 mm and 145  
 454 mm ( $\sim 20\%$ ) more GMSLR than ARCTIC, respectively (Table 1). We note the different  
 455 initial ice volume and initial mass balance of the three simulations, which will be dis-  
 456 cussed in Section 4.



**Figure 8.** Evolution of GrIS-integrated (a) total precipitation and melt ( $\text{Gt yr}^{-1}$ ), (b) SMB ( $\text{Gt yr}^{-1}$ ), (c) MB ( $\text{Gt yr}^{-1}$ ), (d) ice volume ( $\times 10^4 \text{ km}^3$ ), and (e) accumulated contribution to global mean sea level rise (mm) for the three simulations. The thick lines in (a)-(c) represent 20-year running means.

457 Averaged over the whole 350 years, the smaller melt in ARCTIC is most significant  
 458 along the ice sheet periphery in western and northern GrIS, where F09M and F09 melt  
 459 over 300 mm more ice each year (Fig.9c,d). The south tip shows the reverse of the larger  
 460 spatial pattern, with more melt in ARCTIC. This is due to the weaker cloud shortwave  
 461 cooling there in ARCTIC (discussed in Section 3.3.3). The spatial pattern of melt differ-  
 462 ence is consistent for ARCTIC compared to F09M and F09 despite their different initial con-  
 463 ditions (Fig.8d). The spatial pattern of total precipitation difference is almost the op-  
 464 posite for ARCTIC compared to F09M and F09, with the former having lower precipita-  
 465 tion and the latter having larger precipitation centered over the south and southeast coasts  
 466 (Fig.9a,b). Due to the smoother topography in the  $1^\circ$  grid, there is more moisture pen-  
 467 etration into the ice sheet from southeast. For F09M, the effect of directing cold-rain to  
 468 surface runoff and increasing sub-grid roughness is evident, which greatly reduces the  
 469 total precipitation. Over the 350 years, weaker ice sheet thinning (difference  $> 100 \text{ m}$ )  
 470 is found in ARCTIC especially along the western and northern peripheries compared to  
 471 the  $1^\circ$  runs (Fig.9e,f). The larger precipitation in F09 reinforces its smaller melt in the  
 472 south tip, resulting in less ice sheet thinning there.



**Figure 9.** Maps of the difference between ARCTIC and F09M (left column), ARCTIC and F09 (right column): (a-b) annual mean precipitation ( $\text{mm yr}^{-1}$ ), (c-d) surface melt ( $\text{mm yr}^{-1}$ ) averaged over the whole 350 years, and (e-f) ice thickness change (m) between the end of the simulation and the end of pre-industrial period. Blue color indicates more precipitation, less melt, or less thinning. For precipitation and melt, ARCTIC results are remapped to the f09 grid for the comparison.

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### 3.3.3 Surface Energy Balance

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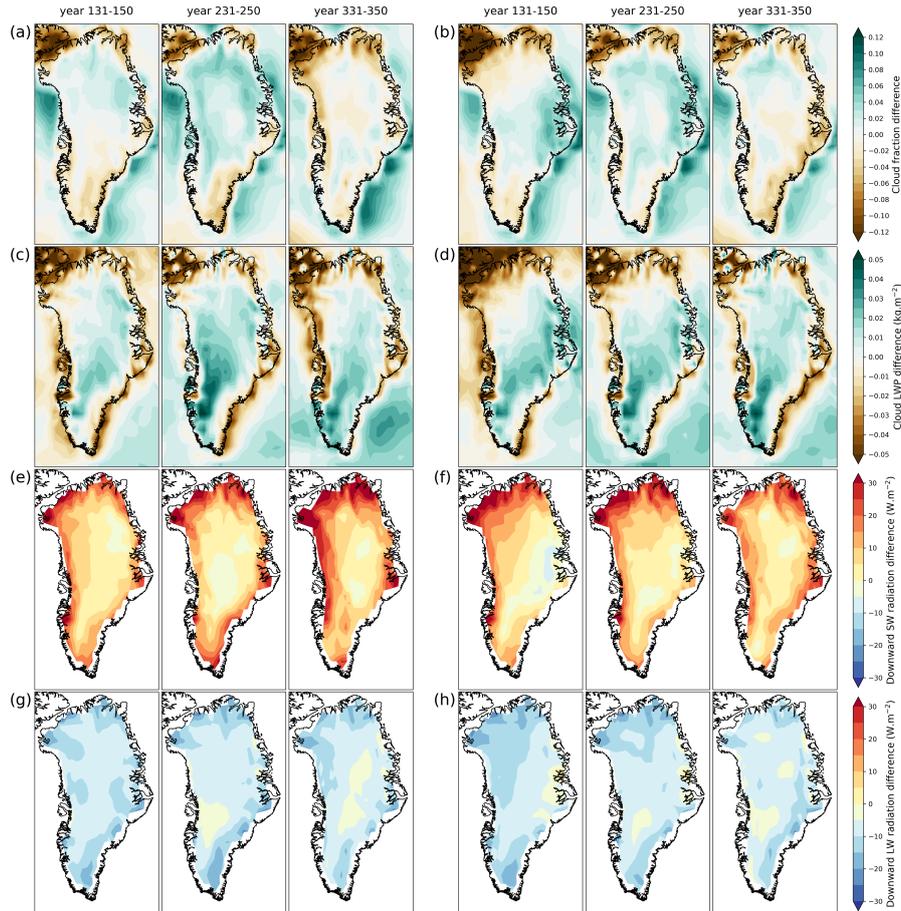
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To explain the differences in meltwater production between ARCTIC and the  $1^\circ$  runs, we examine how incident radiation is impacted by cloud conditions. Figure 10a,b illustrate the summer mean cloud fraction difference between ARCTIC and the  $1^\circ$  runs. There is a ring of cloud fraction surplus around the ocean perimeter of Greenland in ARCTIC, which is similar to previous findings in (Herrington et al., 2022). Lower cloud fraction in ARCTIC is found in the northern and southern peripheries, while in between the cloud fraction can be higher compared to the  $1^\circ$  runs, especially during year 231-250. This pattern is different to that in Herrington et al. (2022), where the smoother topography in lower resolution runs causes more moisture intrusion thus larger cloud fraction in the ice sheet interior. The higher pressure in the  $1^\circ$  runs over the GrIS (Fig.5) may contribute

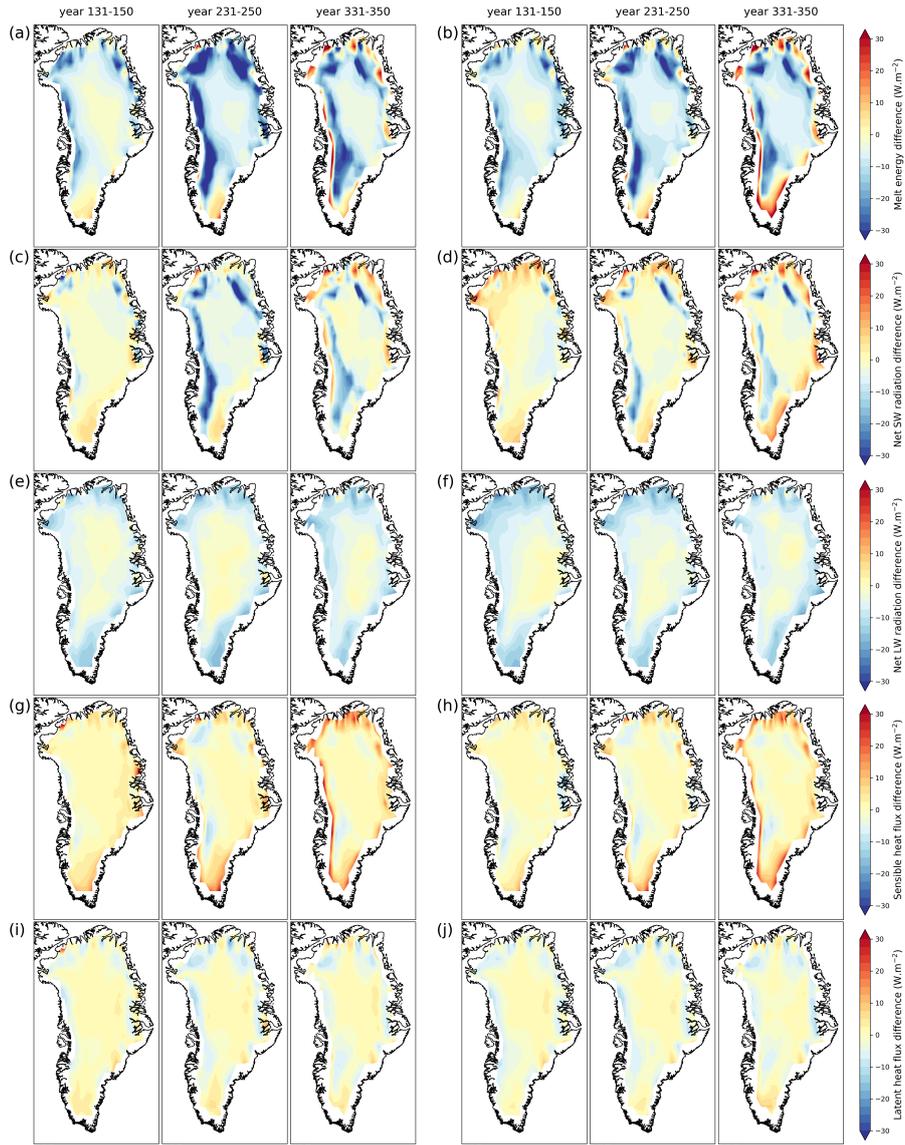
484 to their lower cloud fraction by generating stronger large-scale subsidence and cloud dis-  
 485 sipation. The spatial pattern of total cloud liquid water path (LWP) difference over the  
 486 GrIS generally follows that of cloud fraction difference, except for the much larger LWP  
 487 in the southern interior in ARCTIC (Fig.5c,d). The pattern of cloud difference impacts  
 488 the incident radiation. In ARCTIC, the incident shortwave radiation is generally larger  
 489 due to less cloud cover or thinner clouds, especially in northern Greenland and along the  
 490 ice sheet margins (Fig.10e,f). At the same time, smaller incident longwave radiation is  
 491 found all over the ice sheet for ARCTIC (Fig.10g,h), with the largest differences along the  
 492 north and southeast margins. This is a combined effect of the cloud conditions and lower  
 493 free atmosphere temperature over Greenland in ARCTIC.



**Figure 10.** Maps of the difference between ARCTIC and F09M (left three columns), ARCTIC and F09 (right three columns): (a-b) JJA mean cloud fraction, (c-d) total cloud liquid water path ( $\text{kg m}^{-2}$ ), (e-f) downward shortwave radiation ( $\text{W m}^{-2}$ ), and (g-h) downward longwave radiation ( $\text{W m}^{-2}$ ) averaged over three periods: year 131-150, year 231-250, and year 331-350.

494 Differences in the representation of clouds and the resultant incident radiation have  
 495 a direct impact on SEB terms and thus total energy available for melt. Despite the over-  
 496 all larger incident shortwave radiation in ARCTIC, especially along the ice sheet margins,  
 497 it has smaller net shortwave radiation in the interior compared to the  $1^\circ$  runs, peaking  
 498 at the gently sloping ice surface in the western and northern basins (or transitional area)  
 499 (Fig.11a,b). Therefore, the albedo in these regions of ARCTIC must drop more slowly than

500 those in the  $1^\circ$  runs. The net shortwave radiation difference during year 131-150 is still  
501 small, and it becomes dominant in the SEB during year 231-250 and year 331-350. The  
502 spatial pattern of net longwave radiation difference (Fig.11c,d) is consistent with the pat-  
503 tern of incident longwave radiation (Fig.10g,h), suggesting the weaker cloud longwave  
504 warming effect in ARCTIC. The spatial pattern of differences in net longwave radiation  
505 is consistent through time. Compared to the radiative fluxes, the difference of turbulent  
506 fluxes between ARCTIC and the  $1^\circ$  runs is relatively small (Fig.11e-h). The above men-  
507 tioned gently sloping ice surface in the western and northern basins also favors smaller  
508 sensible and latent heat fluxes in ARCTIC. Larger sensible heat flux is found in ARCTIC  
509 along the ice sheet margins especially during year 331-351, which originates from the higher  
510 near-surface temperature there (Fig.6g-l). The difference of ground heat flux is small be-  
511 tween ARCTIC and the  $1^\circ$  runs (not shown). As the sum of the above SEB terms, the to-  
512 tal melt energy difference has a spatial pattern that agrees with the averaged melt dif-  
513 ference (Fig.9c,d). In the aggregate, the spatial distribution of the lower melt energy in  
514 ARCTIC (Fig.11a,b) mainly results from the combination of smaller net shortwave radi-  
515 ation due to higher surface albedo and smaller net longwave radiation due to lower cloud  
516 fraction.



**Figure 11.** Maps of the difference between ARCTIC and F09M (left three columns), ARCTIC and F09 (right three columns): (a-b) JJA mean total melt energy, (c-d) net shortwave radiation, (e-f) net longwave radiation, (g-h) sensible heat flux, and (i-j) latent heat flux averaged over three periods: year 131-150, year 231-250, and year 331-350 in the unit of  $\text{W m}^{-2}$ . Grids that do not have 100 percent ice fraction were masked out to avoid bias caused by comparing grids with different ice fraction.

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### 3.3.4 Melt/albedo Feedback and the Impact of Ice Sheet Hypsometry

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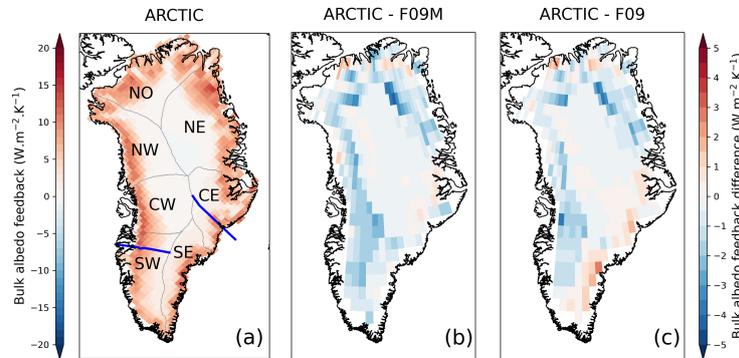
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We speculate that the slower decrease in albedo in ARCTIC is caused by a weaker albedo feedback. We first examine the bulk albedo feedback (Eq.2), which captures the long-term change in net shortwave radiation versus near-surface temperature when time-lags are considered. Over the whole 350 years, the bulk albedo feedback of ARCTIC (Fig.12a) and the  $1^\circ$  runs (not shown) is all positive over the GrIS, meaning more shortwave radiation is absorbed as near-surface temperature rises. The positive bulk albedo feedback is concentrated in the ablation zones, peaking in the lower elevations in the gently slop-

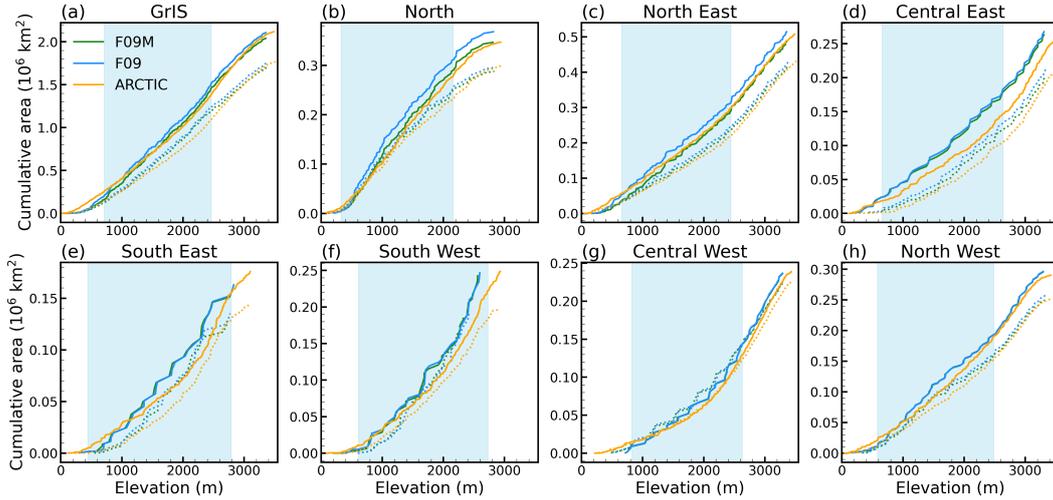
ing western and northern basins ( $> 15 \text{ W m}^{-2} \text{ K}^{-1}$ ). Figure 12b and 12c show the difference of the bulk albedo feedback between ARCTIC and F09M, ARCTIC and F09, respectively. In general, ARCTIC has weaker bulk albedo feedback compared to the  $1^\circ$  runs. A weaker bulk albedo feedback means less shortwave radiation is absorbed given the same amount of near-surface temperature increase. The spatial patterns of the bulk albedo feedback difference agree well with those of the net shortwave radiation difference (Fig.11c,d), indicating a limited role of near-surface temperature increase on causing this pattern. Unlike the bulk albedo feedback that peaks in the lower elevations, the differences are larger over the higher ablation zones. We also examine the albedo feedback (Eq.1) using annual paired detrended anomalies of net shortwave radiation and near-surface temperature over the three 20-year periods. By definition, the albedo feedback reveals an interplay of physical mechanisms, but it is more sensitive to interannual variabilities (e.g., of snowfall), which may make it harder to interpret. The albedo feedback differences show a similar but more variable spatial pattern compared to those of the bulk albedo feedback during year 231-250 and year 331-350 (Fig.S3e-f,h-i) when the albedo feedback is strong (Fig.S3b-c). The difference between bulk albedo feedback and albedo feedback also stresses the importance of melt preconditioning effect on following melt seasons.



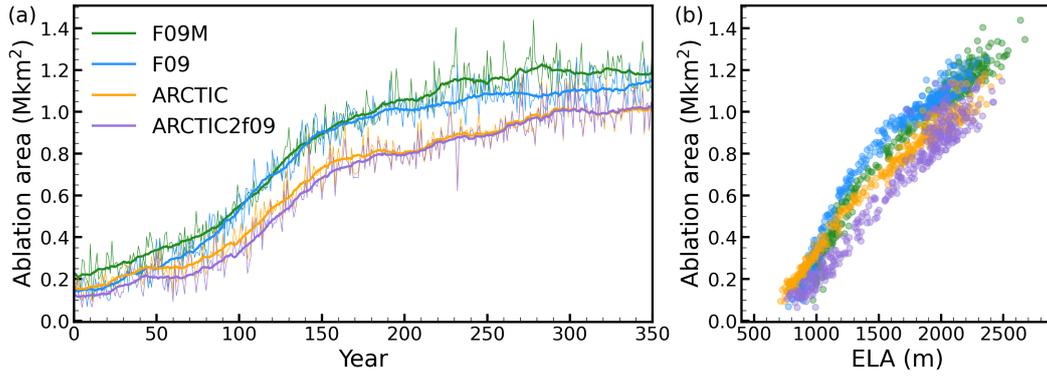
**Figure 12.** Maps of the bulk albedo feedback ( $\text{W m}^{-2} \text{ K}^{-1}$ ) defined by  $\Delta SW_{net}/\Delta T_{air}$  of ARCTIC (a), and the difference between ARCTIC and F09M (b), ARCTIC and F09 (c). Time changes from the end of pre-industrial period to the end of the simulation. Gray lines in (a) separate the seven drainage basins defined by Rignot and Mougnot (2012), and blue lines show the location of the two transects plotted in Figure 15.

The difference in albedo feedback between ARCTIC and the  $1^\circ$  runs can be largely explained by the different representations of surface topography in different resolution. Figure 13 shows the surface elevation-cumulative area relationships in the three simulations for the whole ice sheet and the individual basins defined by Rignot and Mougnot (2012). A steeper slope in the elevation-cumulative area relationship indicates a larger area increase per meter of elevation rise, thus representing a flatter topography. Ryan et al. (2019) demonstrates the dominant role of Greenland’s seasonally fluctuating snowline for reducing ice sheet albedo compared to bare ice albedo reduction caused by melt processes. Here, instead of the end-of-summer snowline elevation, we use the ELA to avoid the possible difficulties of snowline classification. Though different due to processes like superimposed ice formation (Cogley et al., 2011), a significant correlation is found between the ELA and the end-of-summer snowline elevation (Fausto & the PROMICE team\*, 2018). For the whole ice sheet, the average ELA rises from around 600 m to 2,500 m over the 350 years. Within this range, ARCTIC has a less steep slope of the elevation-cumulative area relationship than the  $1^\circ$  runs (Fig.13a), indicating a smaller area increase with the

557 same amount of elevation rise. The slope difference between ARCTIC and the  $1^\circ$  runs be-  
 558 comes even larger at the end of the simulation compared to the start. Figure 14b shows  
 559 the relationships between annual mean ELA and ablation area of the three simulations.  
 560 The match between the simulated ELA-ablation area relationship and the elevation-cumulative  
 561 area relationship demonstrates the role of topography on regulating the speed of ablation  
 562 zone expansion. With the same amount of ELA increase, the ablation area incre-  
 563 ment in ARCTIC is smaller due to its steeper topography, which results in a slower ablation  
 564 zone expansion (Fig.14a) and thus smaller albedo feedback of ARCTIC (Fig.12b,c).  
 565 To further verify the result, the annual mean ablation area and ELA are also computed  
 566 based on the ARCTIC outputs but remapped to the f09 grid (ARCTIC2f09) using the Earth  
 567 System Modeling Framework (ESMF) first-order conservative remapping algorithm (ESMF  
 568 Joint Specification Team, 2021). A similar slower ablation area expansion and smaller  
 569 slope of the ELA-ablation area relationship is found for ARCTIC2f09 (purple line and dots  
 570 in Fig.14) compared to the  $1^\circ$  runs, which attests our theory.



**Figure 13.** Hypsometric surface elevation-area relationships for the whole GrIS (a) and for different basins defined by Rignot and Mouginot (2012) (b-h). For the colored lines, the solid lines represent year 0, and the dotted lines represent year 350. The blue shaded elevation range indicates the annual GrIS- or basin-mean ELA variability in ARCTIC during the 350 years.

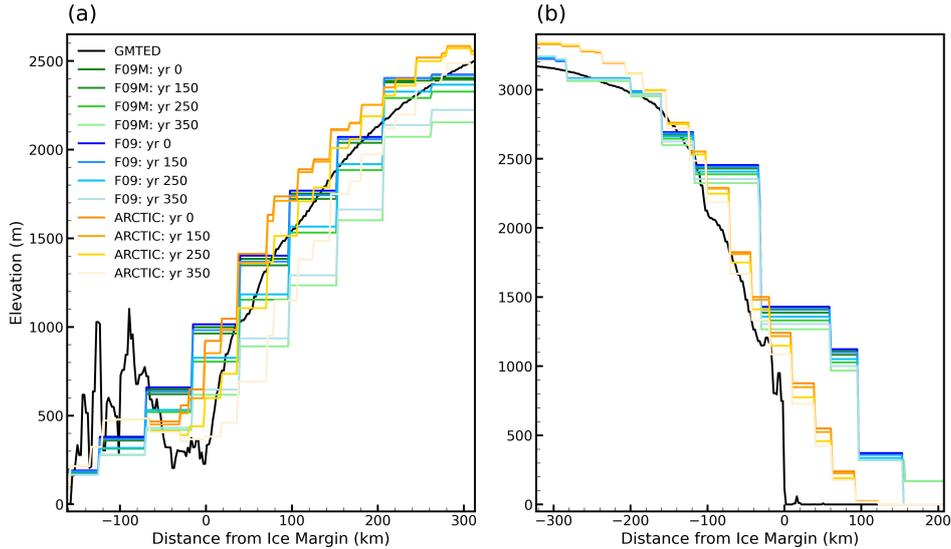


**Figure 14.** Evolution of the annual mean ablation area ( $\text{Mkm}^2$ ) (a) and the relationship between ablation area and mean equilibrium line altitude over the Greenland Ice Sheet (b).

571 Aside from topography, other factors can also affect albedo feedback. Along the  
 572 southeast coast, the larger precipitation in F09 (Fig.9b) potentially slows down the albedo  
 573 reduction more effectively, leading to its smaller bulk albedo feedback there compared  
 574 to ARCTIC (Fig.12c). The impact of clouds on net shortwave radiation also cannot be elim-  
 575 inated from the bulk albedo feedback calculation. However, since the cloud pattern differ-  
 576 ences that cause more downward shortwave radiation in ARCTIC (Fig.10e,f) would play  
 577 en opposite role, it is not considered as a contributor to the major albedo feedback dif-  
 578 ferences.

579 Due to the regional dependence of ELA and surface topography, we also check the  
 580 elevation-cumulative area relationship within separate GrIS drainage basins. The slope  
 581 difference between ARCTIC and the  $1^\circ$  runs is largest within the steepest Central East  
 582 and South East Basins (Fig.13d,e), but due to the narrow ablation zones and large pre-  
 583 cipitation in these basins, it does not result in large difference of the albedo feedback.  
 584 For all the other basins where the albedo feedback is smaller in ARCTIC such as the South  
 585 West, we also find that it has less steep slopes of the elevation-cumulative area rela-  
 586 tionship compared to the  $1^\circ$  runs (Fig.13f). This consists with our hypothesis that the to-  
 587 pography represented in different resolution grids causes the albedo feedback differences.

588 The different representations of surface topography originate from the grid reso-  
 589 lution differences and also the related numerical modeling requirements. Figure 15 shows  
 590 the representation of the ice sheet surface along two transects (positions shown in Fig-  
 591 ure 12a) for the three simulations at different times. One is the east-west "K-transect"  
 592 in southwest Greenland and the other is a transect extending from the central dome down  
 593 to the Kangerlussuaq glacier on the southeast coast. Compared to ARCTIC, the coarser  
 594  $1^\circ$  grids have fewer grid cells along the transects, which may result in flatter surfaces lo-  
 595 cally. Moreover, the smoothing and flattening of the raw topography, necessary to pre-  
 596 vent the model from exciting grid-scale numerical modes (Herrington et al., 2022), causes  
 597 the lower-elevation ablation zones to extend beyond the true ice sheet margin more in  
 598 the  $1^\circ$  runs. It results in flatter slopes along the transects in F09M and F09, which favor  
 599 faster ablation zone expansion. For the more gently sloping K-transect (Fig.15a) with  
 600 a relatively wide ablation zone (can climbs up the whole transect), this impact of sur-  
 601 face slope is more significant as reflected in the bulk albedo feedback (Fig.12b,c). How-  
 602 ever, for the steeper southeast transect (Fig.15b) where the ablation zone below  $\sim 2600$   
 603 m is relatively narrow, the impact is limited.



**Figure 15.** Model surface elevation (m) along the (a) K-transect, and (b) a transect spanning the central dome down to the Kangerlussuaq glacier in southeast Greenland, for F09M, F09, and ARCTIC at different times. The GMTED reference surface is a 1 km surface elevation data set Danielson and Gesch (2011) used for generating the CAM topographic boundary conditions.

#### 604 4 Summary and Discussion

605 In this study, we applied an Arctic-refined variable-resolution grid to a coupled Earth  
 606 system/ice sheet model, CESM2.2-CISM2.1, to investigate future, multicentury, climate  
 607 and GrIS evolution. The variable-resolution grid has a horizontal resolution of  $1/4^\circ$  over  
 608 the broader Arctic region and  $1^\circ$  elsewhere. The simulation was run under a multicen-  
 609 tury idealized  $4\times\text{CO}_2$  scenario and is compared with two other reference simulations under  
 610 the same forcing scenario but using a lower resolution grid ( $1^\circ$  lat-lon grid) to ex-  
 611 plore the impact of enhanced horizontal resolution.

612 Though with different magnitudes and timing, the response of the GrIS to the warm-  
 613 ing climate in the variable-resolution run is similar to what was found in Muntjewerf,  
 614 Sellevold, et al. (2020). In the variable-resolution run, the global annual average near-  
 615 surface temperature rises approximately linearly at a speed of 0.33 K per decade during  
 616  $1\% \text{ yr}^{-1} \text{ CO}_2$  increment. Polar amplification is 1.8, while GrIS amplification is much  
 617 smaller (1.1). After  $\text{CO}_2$  stabilization, the speed of global annual average near-surface  
 618 temperature rise decreases more than 40 percents compared to the previous stage (0.19  
 619 K per decade). An acceleration of the GrIS mass loss is found at around year 110 (2.4  
 620  $\text{Gt yr}^{-2}$  before year 110 to 13.0  $\text{Gt yr}^{-2}$  during year 111-150), which is driven by faster  
 621 SMB decrease. At this time, the expansion of the ablation area is large enough to trig-  
 622 ger a large melt/albedo feedback, which reinforces the absorption of shortwave radiation  
 623 and accelerates the melt. The increased sensible and latent heat fluxes also contribute  
 624 to this process due to expanded and strengthened near-surface temperature inversion.  
 625 By the end of the simulation, the ablation zone covers 57% of the ice sheet surface. The  
 626 cumulative contribution from the GrIS to global mean sea level rise is 53 mm by year  
 627 150 and 831 mm by year 350, which is about 40% and 20% smaller compared to the  $1^\circ$   
 628 runs. The sea level rise contribution from our variable-resolution run by  $\text{CO}_2$  stabiliza-  
 629 tion is also small compared to other CESM simulations under the CMIP RCP8.5 and  
 630 SSP5-8.5 scenarios by year 2100 (109 mm in Muntjewerf, Petrini, et al. (2020) and 76  
 631 mm in Lipscomb et al. (2013)), when the  $\text{CO}_2$  concentration is close to  $4\times\text{CO}_2$ . If com-

632 pared to projections using GrIS models forced by outputs from CMIP5 GCMs ( $90\pm 50$   
633 mm during 2015-2100; Goelzer et al., 2020), our estimation approaches the lower bound.

634 Compared to the  $1^\circ$  runs, the variable-resolution run has relatively slower MB and  
635 SMB decrease. This mainly originates from the smaller surface melt during summer in  
636 the variable-resolution run. The SMB difference between the simulations can be smaller  
637 than their difference in melt due to compensating terms, which stresses the importance  
638 of correctly modeling individual SMB components for making reliable projections of GrIS  
639 mass loss. The excessive melt in the  $1^\circ$  runs is concentrated in the western and north-  
640 ern transitional zones towards the margins, produced by the combined effect of greater  
641 net longwave and net shortwave radiation, with the latter playing a primary role. The  
642 larger net longwave radiation over the ice sheet peripheries in the  $1^\circ$  runs results from  
643 their larger and thicker cloud cover over these regions. The net shortwave radiation dif-  
644 ferences are a product of the stronger albedo feedback over the western and northern basins  
645 in the  $1^\circ$  runs. In coarser grids, stronger smoothing and flattening of topography due  
646 to grid resolution and necessity to prevent model from exciting grid-scale numerical modes  
647 leads to faster ablation zone expansion, thus stronger albedo feedback. Therefore, future  
648 sea level projections based on models with a coarse resolution may be biased high due  
649 to their inability of adequately resolving the Greenland topography.

650 Comparisons between these simulations are complicated by differences in grid res-  
651 olution, physics time step and dynamical core (dycore). Similar to the effect of increas-  
652 ing resolution, reducing physics time step can also increase resolved vertical velocities  
653 and thus condensational heating. By comparing two AMIP style CESM2.2 simulations  
654 using the same quasi-uniform  $1^\circ$  SE grid but different time steps, Herrington et al. (2022)  
655 showed that the simulation with a reduced time step has a warmer troposphere over nearly  
656 all latitudes. Then by comparing two simulations using the same time step but differ-  
657 ent grids - the Arctic grid and the quasi-uniform  $1^\circ$  grid, Herrington et al. (2022) showed  
658 that the warmer temperature caused by enhanced resolution is confined to the refined  
659 Arctic region with a larger magnitude. Therefore, the differences of the simulated cli-  
660 mate within the refined Arctic region in this study is more likely a result of changing the  
661 horizontal resolution though the impact from changing the physics time step cannot be  
662 eliminated. Since the SE dycore is less diffusive than the FV dycore, the resolved ver-  
663 tical velocities are larger in the SE dycore, which can also warm up the atmosphere. It  
664 is impossible to fully disentangle the impacts of changing resolution and dynamical core  
665 in the current simulations since these factors change simultaneously.

666 Another uncertainty in this study is the initial condition. The pre-industrial sim-  
667 ulations of ARCTIC was branched off from an initial condition similar to the initial con-  
668 dition of the F09M 1% CO<sub>2</sub> simulation. The aim of this is to achieve a near equilibrium  
669 state of the GrIS after changing the grid. As a more direct comparison, F09 went through  
670 a similar spin-up process as ARCTIC. The resulted initial ice sheet conditions before the  
671 start of the idealized warming period are different among the three simulations: F09 has  
672 a larger initial ice volume than ARCTIC, while F09M has a smaller initial ice volume than  
673 ARCTIC (Fig.8d). We found that ARCTIC has slower SMB decrease and lower melt com-  
674 pared to the  $1^\circ$  runs in multicentury scale, no matter whether their initial ice volume  
675 is larger or smaller. Also, ARCTIC has a cooler pre-industrial climate than F09M and F09.  
676 The impact of this cooler climate is kept through the whole idealized warming simula-  
677 tion, which is reflected in Figure 6g-l. Moreover, we note that all the three simulations  
678 have small positive drifts of GrIS MB before the start of the warming scenario (Table  
679 1). The cooler initial climate in ARCTIC results in a longer adjusting time for the GrIS  
680 to shift to mass loss, which degrades the specific projections of GrIS sea level rise con-  
681 tribution. However, our result suggests that the near-surface temperature is not the main  
682 driver for the different responses of the GrIS among the simulations. Considering all these,  
683 we can conclude that our finding is robust.

684 One limitation of the current model configuration lies in the ice-ocean interface.  
 685 The direct impact of oceanic thermal forcing on ocean-terminating ice fronts is not in-  
 686 cluded, and the floating criterion used in the calving parameterization is highly ideal-  
 687 ized. The limited understanding and implementation of processes such as calving and  
 688 submarine melting in ice sheet models has been identified as a major source of uncer-  
 689 tainties for future projections of the GrIS (Goelzer et al., 2020). Oceanic forcing can en-  
 690 hance solid ice discharge (Holland et al., 2008; Wood et al., 2018). Using another cou-  
 691 pled Earth system/ice sheet model, EC-Earth-PISM, under the same  $1\% \text{ yr}^{-1}$   $\text{CO}_2$  warm-  
 692 ing scenario, Madsen et al. (2022) showed that even by embedding a constant oceanic  
 693 thermal forcing and a simple geometric calving criterion, the ice discharge decrease is  
 694 much smaller after 350 years. Most modeling studies that include oceanic forcing esti-  
 695 mate a second order future sea level rise contribution from ice dynamics compared to  
 696 SMB for the entire ice sheet (Price et al., 2011; Fürst et al., 2015; Aschwanden et al.,  
 697 2019), but with improved bathymetry and bed topography mapping, Choi et al. (2021)  
 698 showed that ice dynamics could contribute comparably as SMB or more to GrIS mass  
 699 loss over this century. These studies illustrate the importance of properly including ice-  
 700 ocean interactions for future GrIS projections at century to multicentury scale. Currently,  
 701 the functionality of ice sheet-ocean interaction is under study in CESM and will be in-  
 702 cluded in the future versions. The Arctic grid or grids with even higher-resolution re-  
 703 finement will also be helpful by providing a better resolved atmospheric forcing for mod-  
 704 eling the ice dynamics in narrow fjords.

705 In the aggregate, our study demonstrates the value of employing variable-resolution  
 706 grids for coupled climate-GrIS modeling, providing valuable insights into ice sheet-climate  
 707 interactions. It underscores the critical role of grid resolution in modeling the evolution  
 708 of the GrIS on multicentury time scales, particularly in capturing topography-related pro-  
 709 cesses and feedbacks, and thus advances the projection of the GrIS' future sea level rise  
 710 contribution.

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## 720 Open Research

721 The data presented in this manuscript are stored in [https://doi.org/10.5281/  
 722 zenodo.10685261](https://doi.org/10.5281/zenodo.10685261), and the code for generating the plots are available at [https://github  
 723 .com/IceZYin/2022-VR-dynamic-GrIS.git](https://github.com/IceZYin/2022-VR-dynamic-GrIS.git).

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