



Increased future ice loss due to nested ice-climate feedbacks if Antarctic heatwaves intensify

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Abstract. Record-breaking extreme events, including heatwaves leading to surface melt and extreme snowfall, have been observed in Antarctica in recent years. Such extreme events are generally expected to become more frequent and intense with anthropogenic climate change. However, they are not typically considered in the climatic forcing used as the surface boundary condition for ice-sheet models projecting Antarctic mass changes into the future. With a suite of numerical ice-sheet simulations, we explore the additional Antarctic contribution to future sea-level rise when including atmospheric temperature extremes of varying frequency and intensity in projections. Our results suggest that by 2100, increasingly severe and frequent heatwaves have a limited effect on sea-level rise. However, by 2300, atmospheric extreme events could amplify sea-level rise by more than 2 m as atmosphere impacts on Antarctica contribute, rather than mitigate, ice loss. This study also highlights substantial uncertainties in ice-atmosphere interactions (e.g. the role of surface albedo) that amplify the sea-level rise response in an interlinked manner and play a key role in determining future sea-level rise.

1 Introduction

Globally, climate change leads to an increase in the intensity and frequency of extreme weather events (Masson-Delmotte et al., 2021). Heatwaves — prolonged periods of much warmer-than-average air temperatures — are expected to become more frequent, severe, and persistent (Domeisen et al., 2022), with an increased probability of "record-shattering extreme events" (Fischer et al., 2021), under global warming. However, Antarctica remains under-represented in global-scale analyses of the implications of future extreme events.

Despite its frequent absence from world maps projecting the future of extreme events, the last decade has seen several record-breaking extreme events in Antarctica (Lazzara et al., 2012). Perhaps the most prominent example is the March 2022 heatwave, which affected large parts of East Antarctica, and during which atmospheric temperatures temporarily exceeded 38.5°C above the local daily average (Wille et al., 2024a, b). The daily maximum temperature was up to six standard deviations away from the mean probability distribution (a "6- σ event"), and the temperature anomalies could only be reproduced by models in hindcast



(Blanchard-Wrigglesworth et al., 2023). More recently, winter temperatures near the Peninsula rose to +15°C in June 2026, as reported by scientists working in the region, which registered at 20°C above normal (Watts, 2026). While atmospheric internal variability generally drives extreme climate and weather anomalies in the interior of Antarctica (Clem et al., 2020; Wang et al., 2023), the underlying causes of this particular event remain to be analyzed. In general, global-scale general circulation models (GCMs) that participate in model intercomparisons, such as CMIP6, under-represent climate extremes globally (Duan et al., 2025) and in Antarctica (Bracegirdle et al., 2024; Blanchard-Wrigglesworth et al., 2023; Diamond et al., 2024).

While studies consistently predict atmospheric warming across Antarctica, little is known about the future evolution of temperature extremes in Antarctica. Projections assuming continued, very high, emissions in the future, show a 3–8°C increase in near-surface air temperatures by 2100 (Sun et al., 2025), and well over 10°C warmer than today by 2300 (Burgard et al., 2025). A recent study yielded no general significant trend (yet) in the frequency and intensity of temperature extremes averaged over Antarctica (Wei et al., 2019), although weather station data reveal that the Peninsula region experienced a statistically significant increase in the number of extreme high temperatures in the late 20th century (Turner et al., 2021). An analysis of climate model projections shows that an increase in global warming can modify the mean and extreme climate conditions in Antarctica, such as the number of warm days (Saurral et al., 2020). A global warming level of 2.0°C above pre-industrial would lead to an robust increase in the maximum temperature lying above its corresponding, present-day, 90th percentile, especially in the Amundsen–Bellingshausen Sea region. The number of warming events considered heatwaves is also projected to double in most of West Antarctica and could as much as triple in the interior of East Antarctica (Feron et al., 2021). However, what is uncontested is that extremes defined against a cooler climate baseline become less exceptional and new extremes can emerge. A recent review concluded that it is “virtually certain that future Antarctic extreme events will be more pronounced than those observed to date” (Siegert et al., 2023) and Blanchard-Wrigglesworth et al. (2023) conclude that the magnitude of heatwaves that are comparable to the March 2022 heatwave could be amplified beyond the warming expected from forced climate change in the future. The prevailing view in past analyses of climate model outputs (cf. e.g. Feron et al., 2021) is that these projections of future temperature extremes do not necessarily imply any long-term effect on mass balance — a view that this study examines.

At present, Antarctic mass loss is dominated by ocean-induced melting and calving, increasingly contributing to global sea level rise (Otosaka et al., 2023). Recent ice-sheet model projections show, however, that with a gradual warming in near-surface air temperature, the atmosphere becomes the main driver for ice loss when warming exceeds approximately +7.5°C above pre-industrial, which could occur by 2100 under a very high warming scenario (Coulon et al., 2024). Surface melt and runoff both increase with warming (Gilbert and Kittel, 2021), potentially outbalancing snow accumulation increases (Nicola et al., 2023). More frequent and extreme atmospheric temperatures could further amplify net mass loss at the surface, suggesting that atmospheric extremes could play a larger role in modulating the overall mass loss than assumed for present-day conditions. Such surface melt triggers structural changes (Fig. 1) through the *melt-elevation feedback* – where ice thinning lowers the surface into warmer altitudes triggering additional melt (Levermann and Winkelmann, 2016) – and the ice surface (*snowmelt-albedo feedback*), where surface melt amplifies absorption of insolation (Box et al., 2012; Ryan et al., 2023; Jakobs et al., 2021). Furthermore, episodic atmospheric extremes like atmospheric rivers can weaken ice-shelf buttressing through increased

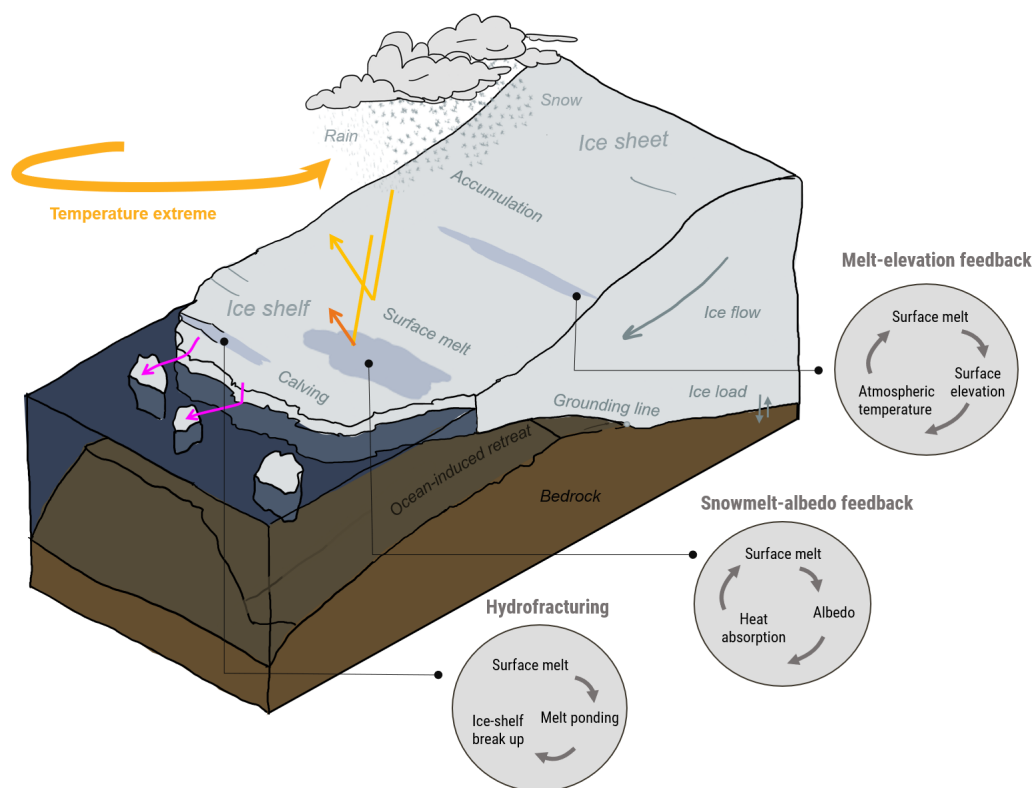


Figure 1. Overview of ice sheet processes and properties associated with temperature extremes. Key feedback mechanisms and non-linear processes — the melt-elevation feedback, snowmelt-albedo feedback, and hydrofracturing — are associated with surface melt.

hydrofracture and calving, as observed historically prior to the Larsen B collapse (van den Broeke, 2005) and at Amery Ice Shelf (Francis et al., 2021).

While studies have examined single temperature and surface melt extreme occurrences in Antarctica (Wille et al., 2024a, b; Tang et al., 2026; Gorodetskaya et al., 2023), the ice dynamic response to such temperature extremes has not yet been investigated. This is despite recent evidence that heatwaves can increase the sea-level contribution from Greenland by up to 0.5 m when atmospheric extremes are included in climate forcing, highlighting the potential non-linear impacts of increasingly extreme summers (Beckmann and Winkelmann, 2023). Furthermore, Antarctica may be currently undergoing Greenlandification (Mottram et al., 2025), a general trend whereby (surface) mass balance processes in Antarctica mirror those in Greenland (i.e. extensive surface melt) with progressing climate change. With this in mind, we explore how Antarctica's sea-level rise contribution could change when extreme temperatures are included in climate forcings. Specifically, we investigate Antarctica's response to extreme climate events, defined here as patterns in extreme heat that persist for an extended period of time (IPCC, 2018) (at least one month), using the continental ice-sheet models PISM (Parallel Ice Sheet Model) and ISSM (Ice-sheet and



Sea-level System Model), forced with output from two global climate models – CESM2-WACCM (Danabasoglu et al., 2020) and UKESM1-0-LL (Sellar et al., 2019) – to 2300.

We model regionally-varying monthly temperature extremes and examine the ice-sheet’s response to heatwaves that vary in intensity and frequency. We explicitly quantify the effect of different feedback mechanisms and processes i.e. the melt-elevation, snowmelt-albedo feedbacks and hydrofracture-induced calving (Fig. 1), on 2300 sea-level contributions under extreme events, showing that extremes produce more ice loss due to the interlinked nature of these feedbacks.

2 Methods

2.1 Ice-sheet model configurations

2.1.1 PISM model setup and initialization

To model ice flow, we use PISM version 2.2 (Bueler and Brown, 2009; Winkelmann et al., 2011). The model initialization uses a dynamic spin-up procedure to bring the ice sheet into a quasi-equilibrium state with pre-industrial climate conditions, i.e. atmosphere and ocean conditions representing the year 1850. The model is then run through the historical period from 1850 to 2015 forced by NorESM atmosphere and ocean anomalies, similar to Reese et al. (2023), including the baseline climatologies (see below for details on the climate forcing). Ice dynamics are computed using a hybrid stress balance that combines the Shallow Shelf Approximation (SSA), which captures membrane stresses relevant for fast-flowing and floating ice, with the Shallow Ice Approximation (SIA), which represents vertically sheared flow typical of grounded interior ice. This hybrid SSA–SIA approach allows the model to efficiently represent both slow inland ice flow and fast outlet glaciers (where the sliding law is represented in the shelfy-stream approximation, SSA; Bueler and Brown, 2009) within the same dynamical framework. Basal sliding, i.e. the velocity at the ice–bed interface, is represented using a generalized power-law formulation (Bueler and Brown, 2009; Schoof and Hindmarsh, 2010) that links the basal shear stress τ_b to the basal sliding velocity, $\mathbf{u}$:

$$\tau_b = -\tau_c \frac{\mathbf{u}}{u_{\text{threshold}}^q |\mathbf{u}|^{1-q}} \quad (1)$$

The yield stress τ_c is determined by the assumed properties of the underlying subglacial (glacial) till and the subsequent effective till pressure, i.e. the effective pressure of the overlying ice on the saturated till. In our simulations, $u_{\text{threshold}} = 100 \text{ m yr}^{-1}$ and $q = 0.75$. For modeling the ice flow, PISM uses an enthalpy formulation (Aschwanden et al., 2012) of the energy balance equations, and a Glen-Paterson-Budd-Lliboutry-Duval (Duval and Lliboutry, 1985) as the material constitutive relation (flow law) in the momentum balance equations, which relates the stress deviator tensor σ_{ij} to the strain rate tensor D_{ij} . The effective ice rheology, ν , is in turn dependent on the liquid water fraction ω , pressure P , grain size, d , and ice temperature T :

$$\sigma'_{ij} = 2\nu(D, T, \omega, P, d) D_{ij} \quad (2)$$

We employ a simple viscoelastic Earth deformation model (Lingle and Clark, 1985; Bueler et al., 2007) to account for the response of the bedrock to variations in ice loading. The model has two layers, an elastic lithosphere and a viscous mantle. All



PISM simulations are performed on an uniform horizontal grid with a spatial resolution of 8 km (and 13 m to 100 m in the vertical). Advancements in PISM development allows us to calculate the surface mass balance (SMB) from temperature and precipitation fields through two approaches: a positive degree day scheme (PDD) and the dEBM-simple model, as discussed in detail below.

2.1.2 ISSM model setup and initialization for independent validation of key results

We also run simulations using v4.24 of the Ice-sheet and Sea-level System Model (ISSM; Larour et al., 2012; ISSM Team, 2024), another, commonly used, finite-element ice-sheet model. The ice-sheet initialization uses snapshot inverse methods to generate an initial ice-sheet state representative of the period 1995 to 2015 (a common reference period used in intermodel comparison projects; Seroussi et al., 2024). Compared to a model using a spin-up procedure, this initialization technique yields a better representation of present-day velocity patterns, when compared to present-day observations. However, due to the mismatch of different input datasets the model might experience strong drift (Seroussi et al., 2011). With the adaptive meshing capabilities of ISSM (Hecht, 1998), we generate an irregular, triangular grid with a variable horizontal resolution. The mesh has a 1 km resolution within regions of highest observed ice velocity (Mouginot et al., 2019), a maximum of 3 km at Thwaites and Pine Island glaciers to capture potential future retreat in this area, and up to 50 km in the ice sheet interior. The final mesh consists of 199,226 elements (for comparison in PISM: $761 \times 761 = 579,121$ grid cells). To match present-day ice velocities (Mouginot et al., 2019), we invert for the basal friction coefficient across the grounded ice area using the Budd friction law (with $p = 1$, $q = 1$; Budd et al., 1979). In ice-shelf areas (where ice is floating and thus basal friction is negligible), we invert for the ice stiffness (or rigidity) parameter, B , that influences the ice viscosity following the generalized Glen's flow law (Glen, 1955). We use cost functions to minimize the misfit between modelled, $v = (vx, vy)$, and measured surface velocities $v^{obs} = (v_x^{obs}, v_y^{obs})$, accounting for the absolute misfit, the relative (log) misfit and a regularization term that penalizes strong gradients in the relevant field. For the cost functions coefficients for the absolute and relative misfit, we put special weights on regions with high present-day ice-thickness change rates based on Smith et al. (2020).

To generate a thermal solution, i.e. three-dimensional temperature and velocity fields, we extrude the 2D mesh to a 3D mesh with 10 almost evenly distributed vertical layers (extrusion exponent of 1.1), by projecting the SSA velocities onto this 3D mesh and resolve the thermal equation (Seroussi et al., 2013) using the enthalpy formulation from Aschwanden et al. (2012) and surface temperature and geothermal heat flux as boundary conditions. Solving an inversion solution for ice shelf viscosity and basal friction coefficients is conducted twice through this approach. Following this iterative approach, we collapse the 3D mesh into 2D, and run the mass and momentum balance solutions forward in time, with bi-weekly time steps and using the shelfy-stream approximation, for a 20 year relaxation period. This relaxation uses a mean SMB field from RACMO2.3p3 (mean 1995-2015; van Dalum et al., 2021) and the present-day basal mass balance field from ISMIP6 (see below for details). The SMB field may account for the considerable large PDD model drift in the future period, as discussed in section 3.1.

2.2 Parameterizations of ice-atmosphere interactions

In all model configurations, the surface mass balance is calculated as $SMB = \text{accumulation} - \text{melt} + \text{refreezing}$.



2.2.1 dEBM-simple (PISM)

In PISM, we use the “simplified” diurnal Energy Balance Model developed by Krebs-Kanzow et al. (2021) and implemented in PISM by Zeitz et al. (2021) to calculate melt. The scheme is tuned to present-day Antarctic surface conditions (Garbe et al., 2023). Surface melt is calculated by a temperature-dependent and insolation-dependent component, as follows:

$$M = \frac{\Delta t_{\Phi}}{\Delta t, \rho_w L_m} [(1 - \alpha)\tau SW_{\Phi} + c_1 T_{\text{eff}} + c_2], \quad (3)$$

where Δt_{Φ} is the melt period of a full day Δt , ρ_w is density of water and L_m is latent heat of fusion, here $3.34 \times 10^5 \text{ J kg}^{-1}$. α is the albedo, τ the atmospheric transmissivity, SW_{Φ} is the incoming solar shortwave radiation at the top of the atmosphere and T_{eff} is the cumulative temperature exceeding the melting point per month (similar to the estimate of Positive-Degree Days, PDD). The constants c_1 and c_2 result from tuning the surface module to present-day surface conditions (i.e. historical RACMO2.3p2 melt data, cf. Garbe et al., 2023).

2.2.2 PDD (PISM, ISSM)

With both PISM and ISSM, we also calculate surface melt using a standard Positive Degree Day (PDD) method. The PDD method estimates how many days above 0°C occur in a year by assuming that hourly temperatures, T , are normally distributed around the monthly mean, T_m . The computed positive degree days are used to estimate the amount of surface melt based on degree-day factors. For all simulations we use the default values of 3.29 mm PDD^{-1} for snow melt and 8.79 mm PDD^{-1} for ice. In both PISM and ISSM, the calculation of PDD follows (Huybrechts and De Wolde, 1999) (Eq. C15):

$$PDD = \frac{1}{\sigma_{PDD} \sqrt{2\pi}} \int_0^{1 \text{ year } T_m + 2.5\sigma_{PDD}} \int_{0^{\circ}\text{C}} T \exp\left[\frac{-(T - T_m)^2}{2\sigma_{PDD}^2}\right] dT dt. \quad (4)$$

We use $\sigma_{PDD} = 3.5^{\circ}\text{C}$ as the standard deviation of hourly temperatures around the monthly mean, to align with previous studies (Garbe et al., 2023). To achieve a working PDD scheme for Antarctic conditions, changes to the ISSM code were made (see the data repository for this paper). Most importantly, adjustments of the PDD factors based on the mean June-July-August temperature were disabled.

We use an atmospheric lapse rate of $\Gamma = 8.2^{\circ}\text{C km}^{-1}$ so that changes in the ice thickness feed back into the monthly mean temperature over the ice sheet.

2.2.3 Refreezing

In PISM, refreezing is represented as a fixed fraction of melted snow and ice (set at the default fixed value of 0.5), with the remaining meltwater assumed to run off. In ISSM x PDD, refrozen, superimposed ice is calculated considering the thermal refreezing capacity of the ice and the capillary retention in the remaining snow.



2.3 Parameterizations of ice-ocean interactions

160 In PISM, sub-shelf melting is parameterized with the Potsdam Ice-shelf Cavity mOdel, PICO (Reese et al., 2018), which represents the vertical overturning circulation present in ice-shelf cavities driven by the ice pump. PICO is a box-type model, where sub-shelf melting in box k is given by

$$m_k = -\frac{\gamma_T^*}{\frac{\rho_i}{\rho_w} \frac{L}{c_p}} (aS_k + b - cp_k - T_k). \quad (5)$$

Here, ρ_i is ice density (in PISM: 910 kg m^{-3}), ρ_w is density of sea water (here 1028 kg m^{-3}), c_p is the heat capacity of seawater (here $3974 \text{ J kg}^{-1} \text{ }^\circ\text{C}^{-1}$), L is the latent heat of fusion (here $= 3.34 \times 10^5 \text{ J kg}^{-1}$) and γ_T^* is the effective turbulent heat exchange coefficient (here $= 3 \times 10^5 \text{ m s}^{-1}$). T_k and S_k are the ambient ocean temperature and salinity of box B_k , and a and b represent coefficients in the implemented freezing equation (see Reese et al., 2018). Temperature and salinity are advected into the ice-shelf cavity from far-field ocean temperature and salinity. The input ocean temperature field is adjusted by a constant offset, $\delta T_{\text{PICO region}}$, which results from the tuning process of γ_T^* and another parameter that influences the strength of the vertical overturning circulation.

In ISSM, we use the ISMIP6 quadratic local basal melt parameterization (Jourdain et al., 2020), which reads

$$m(x, y) = \gamma_0 \times \left(\frac{\rho_w c_w}{\rho_i L} \right)^2 \times \{ \max[\text{TF}(x, y, z_{\text{draft}}) + \delta T_{\text{sector}}, 0] \}^2. \quad (6)$$

$\text{TF}(x, y, z_{\text{draft}})$ is the thermal forcing at the ice–ocean interface and γ_0 is similar to the PICO exchange velocity and those commonly used to calculate ice–ocean heat fluxes (Holland and Jenkins, 1999; Jenkins et al., 2010). δT_{sector} is the temperature correction for each sector and is limited to the range $[-2, 2]^\circ\text{C}$. We use the ISMIP6 basal melt parameterization that does not factor in a local slope dependency of the ice draft, as discussed in Jourdain et al. (2020). We tune γ_0 and δT_{sector} to match mean annual melt rates from 1997–2021 from (Davison et al., 2023) for 130 ice shelves (see Supplement Table S1 and S2 for details).

In all model simulations, we apply thickness calving so that all floating ice thinner than 50 m is calved off. In PISM, we use both thickness- and *eigen-calving*, with the latter being a 2D strain-rate based parametrization which is commonly used in combination with thickness calving (Garbe et al., 2020; Albrecht et al., 2020; Klose et al., 2024; Winkelmann et al., 2026), to improve realism and numerical stability.

Our standard configuration is PISM x CESM x dEBM-simple, meaning simulations run with climatic boundary conditions from CESM2-WACCM, where surface melt is calculated via dEBM-simple, to which we compare all other runs (Supplement Table S3). The difference between the two ice-sheet model setups is summarized in Table A1.

2.4 Design of extreme atmospheric forcing scenarios

All ice sheet model experiments are run from 2015 to 2300. To account for climate forcing uncertainty, spatially-varying monthly anomalies in temperature and precipitation relative to the present-day baseline (which we define as the mean over the period 1995–2015) are derived from two CMIP6 Earth System Models: CESM2-WACCM (Danabasoglu et al., 2020)



190 and UKESM1-0-LL (Sellar et al., 2019). We use historical simulations from each model over the period 1995–2015, then follow the SSP5-8.5 scenario (Meinshausen et al., 2020). We use the SSP5-8.5 scenario, a high-end member of possible future climate change, for which ice sheet model forcings are available until 2300, when changes in the ice sheets become more prominent. The realism of reaching a 8.5 W m^{-2} radiative forcing by 2100 is contested (Hausfather and Peters, 2020; Schwalm et al., 2020), but cumulative emissions have been tracking this high-end scenario (Schwalm et al., 2020) in the recent
195 past. Focusing on the upper end allows us to better assess worst-case sea-level rise, which is the primary aim of this study given that atmospheric extremes have not been included in forcing scenarios thus far.

To maintain consistency with our respective model initializations, in ISSM we use the RACMO2.3p3-ERA5 product (van Dalum et al., 2021, 2022) as the reference climatology for temperature and precipitation. For all PISM runs, we use temperature and precipitation fields from RACMO2.3p2 (van Wessem et al., 2022). The two products are very similar, but summer
200 temperatures are regionally $1\text{--}2^\circ\text{C}$ warmer in RACMO2.3p2 (van Dalum et al., 2022).

To investigate the sensitivity of the Antarctic Ice Sheet to extreme warming events, we impose synthetic heatwaves by amplifying mean monthly forcing anomalies. Specifically, we scale 10-year mean January surface air temperature anomalies by factors of $I = 1.25, 1.5$, and 2.0 , beginning in 2015, and applied at intervals of every $f = 5, 10$, and 20 years through 2300. The first extreme for each scenario is imposed in January 2015. Restricting the perturbation to January isolates the
205 peak austral summer period, when surface melt processes could be most prominent. Our approach is analogous to Beckmann and Winkelmann (2023) and allows us to compare the impacts of increasingly intense and frequent warming extremes on the Greenland and Antarctic Ice Sheets. To account for the heterogeneous warming trend in the Antarctic (Wang et al., 2025), we resolve the heatwaves spatially, i.e. we calculate the heatwave using the temperature fields to obtain varying anomalies across the ice sheet. That means that the pattern of the synthetic heatwave does not stay fixed through time. The magnitude of the
210 resulting January Antarctic mean surface air temperature anomalies are included in Supplement Fig. S1. For comparison, this approach results in local deviations from the monthly mean temperature similar to the observed March 2022 heatwave for the present-day forcing (Supplement Fig. S2). During March 2022, ERA5-Land monthly mean temperatures (calculated from daily data) were up to 3.6°C warmer than without the heatwave in parts of East Antarctica (Copernicus Climate Change Service, 2024). A similar magnitude is obtained in our idealized forcing for a synthetic heatwave in 2020 using a factor of $I = 2.0$, or in
215 2115 using a factor of $I = 1.25$.

To account for changes in the ocean, we use yearly ocean temperature and salinity data (for PISM) and thermal forcing (for ISSM) as used in the ISMIP6 exercise (Seroussi et al., 2024) of the matching CMIP6 model used for the atmosphere (CESM2-WACCM or UKESM1-0-LL). For PISM, the ocean forcing is bias-corrected for present-day ocean observations (by Reese et al., 2023).

220 2.5 Experimental design

Our core simulations include one control (with constant present-day climate conditions) for each model configuration (PISM x dEBM, PISM x PDD and ISSM x PDD respectively) and two baseline simulations (with SSP5-8.5 forcings, without added extremes).



We perform nine different heatwave scenarios for the two different climate forcings each. This yields a total of 63 core simulations.

225 We further perform sensitivity tests to examine the impact of several modeling choices on our results, comparing against our standard PISM x CESM x dEBM-simple set-up. We perform "NO-MASS" runs in which changes in the ice-sheet's mass through accumulation or ablation is not passed to the ice thickness variables, thereby mimicking the setup of a regional atmospheric model that calculates SMB under a fixed ice geometry. Using this setting signifies a violation of the mass continuity equation but allows us to determine the effect of SMB changes to the ice flow and isolate the dynamic component in the sea-level rise signal. To calculate the dynamic sea-level rise component (dynSLR), we subtract the sea-level rise potential of the simulated grounded surface mass balance (SMB_{gr}) from a run with fixed geometry from the total grounded mass, VAF_{gr} in the fully dynamic simulation:

$$dynSLR_{BSL, extremes} = \Delta VAF_{gr} - SMB_{gr, no mass}. \quad (7)$$

In addition to that, "NO-ALB" simulations have the same set-up as PISM x CESM x dEBM-simple, but no changes to the albedo as a function of the calculated melt are allowed, i.e. there is no snowmelt-albedo feedback to surface melt. In those runs, slope b_α of equation (8) in Garbe et al. (2023) is set to zero. To test whether temperature extremes can cause irreversible changes in ice-shelf integrity, we run an additional set of PISM simulations that incorporate the effect of increased surface melt on hydrofracturing ("HYDRO-FRACT"). This is implemented through an offline coupling approach, as follows. In general, frontal retreat can be prescribed in PISM providing a (time-evolving) mask that specifies the maximum ice extent at a given time, expressed as the fraction of each grid cell covered by ice (0=no ice, 1=ice). In our study, the goal was to derive an updated frontal-retreat mask based on simulated (extreme) surface melt. This approach is similar to that used in the ISMIP6 multi-model intercomparison for sea-level rise until 2300 (Seroussi et al., 2024). We analyze a simulation without frontal retreat and, starting in 2015, extract the first year a 10-year running mean at each grid point surpasses 725 mm yr^{-1} , a threshold commonly associated with calving triggered by melt water ponding (Trusel et al., 2015; Nowicki et al., 2020; Seroussi et al., 2024). When this threshold is reached over floating ice, the land ice area fraction is changed from 1 to 0, causing the affected grid cell to be classified as ocean. If this transition occurs away from the calving front, openings (holes) may develop within the ice shelf. To ensure consistency with previous (and potential forthcoming) model intercomparison exercises, we refrain from applying a different threshold or defining it via another variable, e.g. runoff (Jourdain et al., 2025). We further test the effect of lapse-rate correction i.e. the melt-elevation feedback ("NO-ELV"), bedrock adjustment ("NO-BED") and calving ("NO-CALV") on our results. For an overview of all sensitivity tests see Table 1.

3 Results

3.1 More frequent and extreme heatwaves increase ice loss

The baseline scenarios, without extremes, reflect a broad spread in sign and magnitude of sea-level rise (Fig. 2). The projected contribution to future sea-level change ranges from a slight lowering, to several meters of sea-level rise by 2300, which is



Table 1. List of sensitivity tests tested with PISM x dEBM x CESM

Sensitivity test	Identifier	With the effect	Without the effect
Dynamic mass loss	NO-MASS	Ice thickness is updated every time step based on the simulated SMB, BMB and ice flow	No evolving ice geometry (“no mass run”)
Melt-elevation feedback	NO-ELV	Temperature correction via lapse rate of $\Gamma = 8.2^{\circ}\text{C km}^{-1}$	$\Gamma = 0^{\circ}\text{C km}^{-1}$
Snowmelt-albedo feedback	NO-ALB	Slope in albedo parameterization (eq. 8 in Garbe et al 2023): $b_{\alpha} = -740.4 (\text{kg m}^{-2} \text{s}^{-1})^{-1}$	$b_{\alpha} = 0 (\text{kg m}^{-2} \text{s}^{-1})^{-1}$; alternatively: PDD scheme
Hydrofracturing	HYDRO-FRACT	Offline coupling approach of updating the frontal retreat mask using simulated surface melt threshold of 725 mm within a 10-year period	None (default)
Bedrock uplift	NO-BED	Lingle-Clark bedrock deformation model	None (default)
Calving	NO-CALV	Thickness calving (+ eigencalving with PISM)	No calving option set

255 expected given the uncertainty in model physics (PISM vs. ISSM), parameterization uncertainty (PDD vs. dEBM-simple), climate forcing (CESM vs. UKESM) and consideration of non-linear feedback processes (e.g. with or without hydrofracturing). For instance, ISSM x PDD x CESM produces sea-level lowering of around 95 cm by 2300; in contrast, the baseline scenario of PISM x dEBM x CESM shows sea-level rise of 1.08 m. When accounting for the non-linear disintegration of ice shelves due to surface melt, Antarctic melt could lead to 1.64 m of sea-level rise in the future (PISM x dEBM x CESM x HYDRO-FRACT).

260 Please note that to ensure that our analysis reflects the underlying forced climate evolution and long-term signal rather than biases introduced by the model initial state, we remove the model drift i.e. how the model would behave when present-day climate conditions in the atmosphere and ocean were to be held constant (see Appendix Fig. A1).

The sea-level rise contribution from the Antarctic Ice Sheet directly increases under re-occurring temperature extremes, which become more severe and intense with global warming. While the effect of atmospheric extremes is limited before

265 2100, they lead to substantial additional sea-level rise compared with the BSL by 2300 (Fig. 3). Ice loss is mainly driven by the negative surface mass balance simulated in heatwave years (Supplement Fig. S3). The broadest spread in sea-level rise compared to the baseline is simulated by the configuration PISM x dEBM x CESM x HYDRO-FRACT. With this set-up the most severe heatwave case (2.0 x more intense January temperature anomalies every 5 years, denoted “I=2.0, f=5”) causes additional sea-level by 2300 of +2.41 m (BSL: 1.64 m; I=2.0, f=5: 3.79 m). In the standard configuration, PISM x dEBM x

270 CESM, the extremes scenario produces an additional sea-level rise of 2.0 m. For model configurations projecting a sea-level lowering in the future, i.e. PISM X PDD X CESM or ISSM X PDD x CESM, heatwaves reverse the net sea-level contribution by generating sufficient surface melt to counteract gains in snowfall accumulation. The sea-level rise contributions using ISSM x PDD with the most severe extremes scenario (I=2.0, f=5: around 0.6 m) are similar by 2300, but ISSM x PDD x CESM diverges more from its baseline: the baseline sea-level rise with CESM forcing is -0.90 m (signifying a sea-level drop), while

275 the baseline for UKESM forcing is +0.05 m. By 2100, the largest sea-level rise contribution is +0.07 m by PISM x dEBM x CESM x HYDRO-FRACT, compared to -0.1 m sea-level lowering simulated by the baseline of PISM x PDD x CESM (Fig. 3 a).



As temperature anomalies amplified by the imposed heatwaves are small, temperature extremes in these scenarios change the signal by less than 0.05 m.

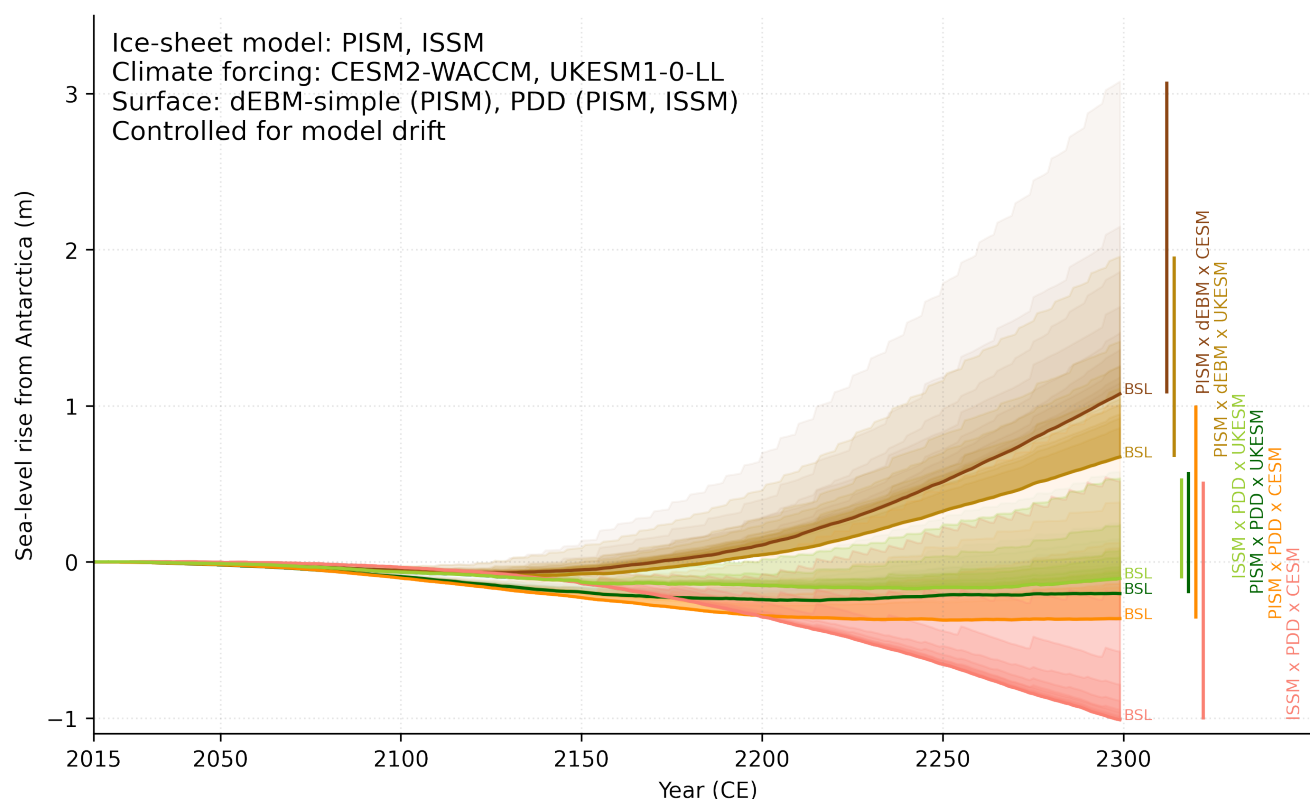


Figure 2. Antarctic sea-level rise contribution with extreme events included. BSL=baseline, i.e. simulations without extremes. For each simulation, the model drift is removed i.e. the simulation in which present-day climate conditions are held constant.

3.2 Ice dynamic response to extreme surface forcing

280 In our baseline scenarios without extremes, we find thinning of more than 500 m around the ice-sheet margin for the model configuration PISM x dEBM x CESM x BSL, with thickening in the Dronning Maud Land sector of East Antarctica and Marie Byrd Land in West Antarctica of 50 to 150 m (Fig. 4 a.). In these regions, the general increase in snowfall drives the positive mass change. In the most extreme heatwave scenario ($I=2.0$, $f=5$), thinning extends to the interior, where 40 to 70 m ice is lost by 2300 compared to the baseline.

285 Increased ocean-induced sub-shelf melting and surface melt thins the ice-shelves around Antarctica in all scenarios. Ice shelf break up and disintegration commences between 2050 and 2100 (Fig. A2), which is only marginally altered by atmospheric temperature extremes. The Filchner–Ronne Ice Shelf and Ross Ice Shelf approach near-complete collapse by 2200, whereas the Amery Ice Shelf disappears around the year 2100 in our simulations, roughly independent from the heatwave scenario.

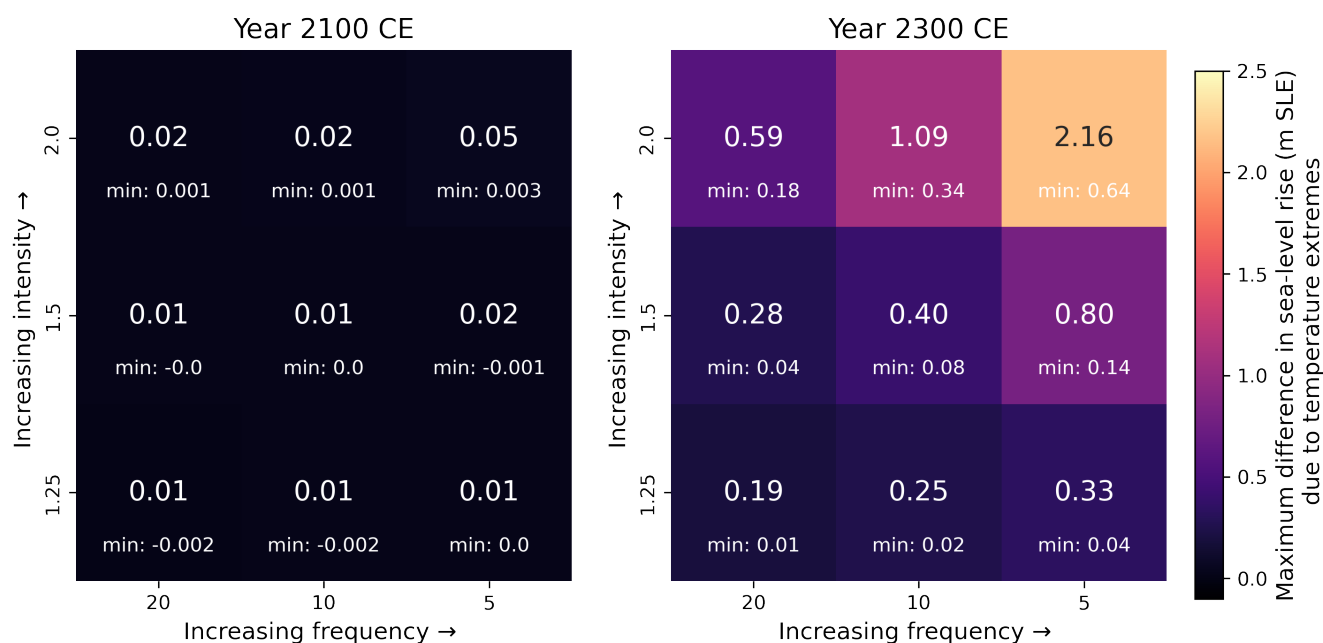


Figure 3. Overview of maximum additional sea-level rise from temperature extremes. The extremes are imposed at varying frequency (every $f=5, 10, 20$ years) and intensity ($I=1.25, 1.5, 2.0 \times$ January warming anomalies). The minimum change (min.) is given for each scenario as well.

These results are consistent with existing assessments of Antarctic ice-shelf vulnerability to future ocean warming (Burgard et al., 2025) and do not reveal any unexpected behaviour due to extremes. With UKESM, a sudden acceleration in basal melting after 2070, especially at the Ross Ice Shelf (Siahaan et al., 2022), drives earlier ice-shelf disintegration.

In the evolution of the individual ice-shelf area, we thus find a strong dependence on the chosen climate model. While Amery Ice Shelf varies relatively little for different climate forcings (Fig. A2), Filchner-Ronne Ice Shelf collapses 30 years earlier when forced by CESM than by UKESM (around 2090). At Ross Ice Shelf, the ice-shelf area declines rapidly after 2100 under UKESM forcing, but later for CESM forcing (around 2130-2150) independent of the chosen surface module – PDD or dEBM-simple. When accounting for hydrofracture, ice shelves are prone to collapse more than a century earlier in scenarios with extremes than those without (Supplement Fig. S4). In the most severe extreme scenario ($I=2.0, f=5$), Antarctic ice-shelf area declines earlier and more rapidly compared to its baseline (Fig. A2), so that most shelves are lost more than two decades earlier with the same climate forcing.

The grounding lines around Antarctica, where bedrock, ice and ocean meet, retreat in several outlet glaciers in our simulations. The most extensive grounding line retreat occurs within the ice streams draining into the Filchner-Ronne Ice Shelf, the Siple Coast at Ross Ice Shelf, at Amery Ice Shelf, at Totten Glacier, and the northern Antarctic Peninsula. The most significant



influence of heatwaves is in the Amery Ice Shelf region where heatwaves cause ice sheet retreat much further inland of the simulations without extremes (cf. Fig.4, Fig. A3).

305 By the year 2300, surface velocities are projected to increase across Antarctica in all PISM simulations (Fig. 4 a.). In some regions, the change is more than 2 km yr^{-1} . For example, mean grounded ice flow at Filchner-Ronne Ice Shelf increases from 20 to 50 m yr^{-1} and at Ross Ice Shelf from around 12 to more than 18 m yr^{-1} . In most ice-shelf regions, grounded ice flow increases the least for the most intense heatwave scenario, $I = 2.0$, $f=5$ (Supplement Fig. S5). In contrast, the baseline scenarios are not always those with the largest increases in ice flow speed. For the Amery Ice Shelf region, scenario $I=1.25$, $f=5$ increases
 310 most significantly (Fig. 4, Fig. A3.). In this scenario, the grounding line retreats furthest, which could be a result of the unique slope changes that depend on where thinning occurs. The slope becomes steepest because of limited melting at the plateau but heightened melting at lower altitudes (Fig. 4 b.). The Antarctic Peninsula is an exception to this general acceleration of grounded ice flow, instead showing substantial glacier retreat. Only a few major Peninsula ice streams remain by 2300 across all simulations, leading to an overall decrease in speed.

315 To understand the differences in the simulated acceleration, we examine the primary controls on ice flow. The driving stress is a result of gravitational acceleration (g), the surface slope (α) and ice thickness, h ($\tau_d = \rho g h \sin(\alpha)$, where ρ is the ice density, here 917 kg m^{-3}). This relationship highlights the fact that changes in the overall ice thickness distribution are key in determining ice flow. More frequent and intense extremes near the margins can lead to enhanced thinning, increasing upstream driving stress and triggering further dynamic thinning. If, however, the ice sheet thins significantly everywhere, the resulting
 320 relatively uniform reduction in thickness can lead to a decrease in overall driving stress (e.g. as in $I=2.0$, $f=5$, Fig. 4 a.).

We further find that heatwave signatures propagate as a change in ice sheet thermal and viscous properties: the effective ice viscosity is lower in the simulations with surface temperature extremes than in the baseline case (Supplement Fig. S6,a.). Temperature extremes could thus also lead to ice softening (i.e. reduction in viscosity due to the temperature dependence), which in turn can enhance ice flow. The signal extends into the upper 1000 m of the ice column where ice temperatures are
 325 also elevated. The ice viscosity responds immediately after a summer extreme event and then changes slowly over time, when plotted against time for a single location (Supplement Fig. S6 b).

These changes in effective viscosity are likely due to the choice of our flow law (Aschwanden et al., 2012), which explicitly represents the thermo-mechanical coupling in ice, allowing warmer ice to deform more readily through a reduction in viscosity (or an increase in ice softness). Despite this, it is difficult to quantify the strength of this coupling and its long-term impact
 330 on ice flow, as the ice flux is influenced by several interacting processes, including changes in surface slope, dynamic flow adjustments, and associated velocity increases. Isolating the effect and a more robust quantification of how single extreme events affect ice flow warrants further investigations.

With additional simulations, we isolate the direct effect of SMB changes on the overall mass balance to estimate dynamic sea-level rise driven by temperature extremes. We remove the effect of ocean-driven melt by further subtracting our baseline
 335 scenario. Overall, the change in the dynamic sea-level rise component differs significantly between model configurations (Fig. A4), but more frequent and severe heatwaves decrease this trend independent of the chosen model configuration.

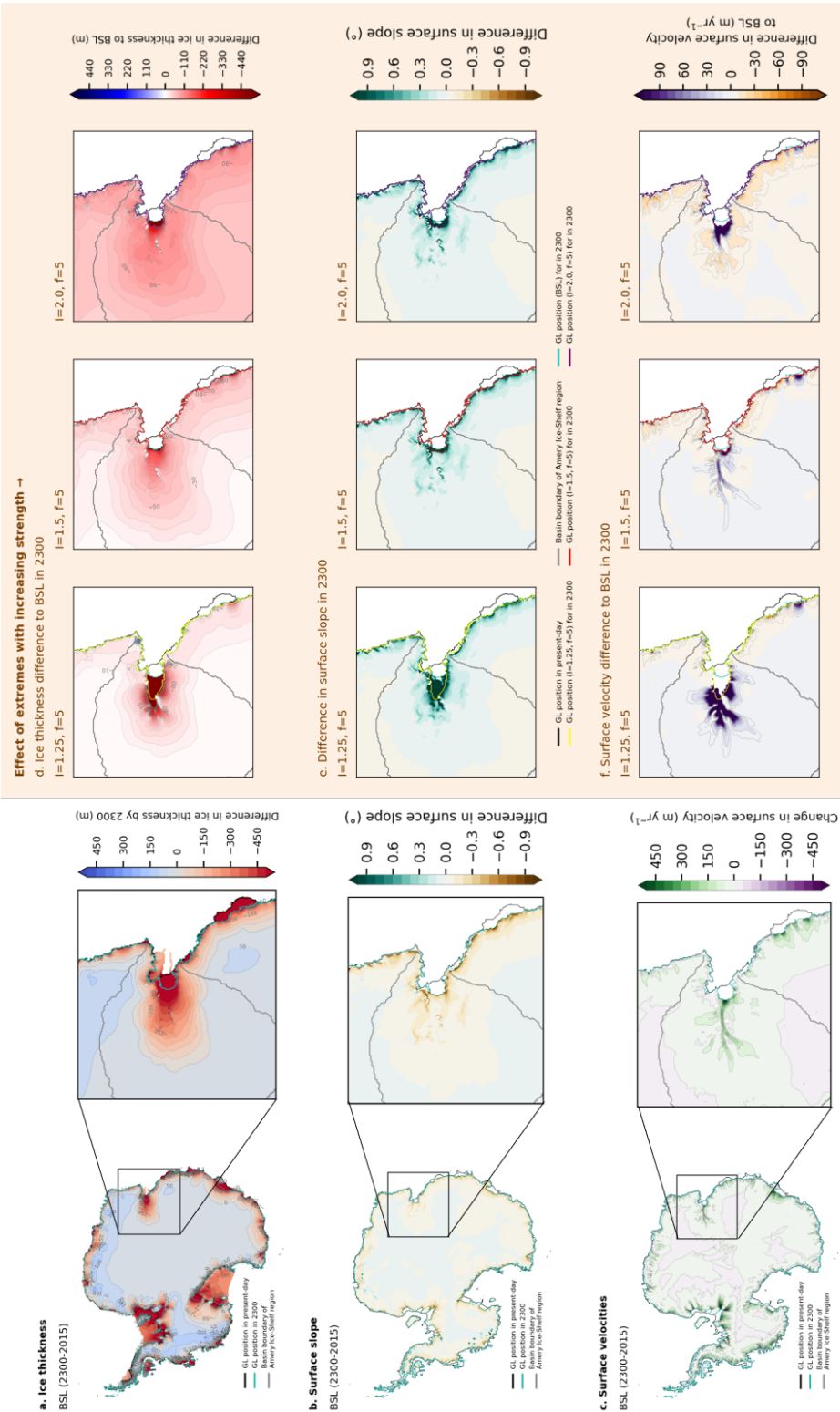


Figure 4. Overview of changes by 2300 and the effect of three selected extreme scenarios compared to the BSL. Antarctic-wide changes in ice thickness (a.) surface slope (b.) and surface velocity changes (c.) for the baseline scenario compared to present day, and the effect of re-occurring (here shown: every 5 years) heatwaves of intensity 1.25, 1.5 and 2.0 (d.-f.) by 2300 at Amery Ice Shelf (inlet).



More extreme events lead to greater thinning, reduced ice thickness, and consequently a smaller ice flux ($Q = H \times U$), particularly for the simulations forced by CESM. With this forcing, atmospheric warming without extremes already causes substantial thinning, such that additional thinning associated with extremes has little further effect on ice dynamics; instead, the thickness reduction (ΔH) dominates the flux response. With UKESM forcing, by contrast, atmospheric warming is less pronounced, so the additional thinning induced by extremes may have a larger impact on ice dynamics and flow velocities.

3.3 Growing importance of ice-climate feedback mechanisms

Our simulations highlight the growing importance of ice-atmosphere interactions that play only a minor role in shaping Antarctic ice dynamics in the absence of extreme events. With our model configuration PISM x dEBM x CESM, we analyze additional sea-level rise resulting from non-linear processes. In particular, we examine the effects of the melt-elevation feedback, which most ice-sheet models already include, the snowmelt-albedo feedback and hydrofracturing (Fig. 5). These feedback processes cause compounding ice loss; that is, beyond the additional sea-level rise caused by one process (melt-elevation feedback), further ice loss is simulated when accounting for the second (snowmelt-albedo) and third processes (hydrofracturing). The melt-elevation feedback scales with additional extremes and increases the ice loss in 2300 by around 50 cm compared to simulations in which this feedback is disabled. Using ISSM with similar experimental setups (in which we set the lapse rate to 0°C per km thickness difference), this effect is lower, with a maximum of 0.12 m by 2300 (Supplement Fig. S7).

We find that the snowmelt-albedo feedback is the largest contributor to sea-level rise when using the dEBM-simple parameterization that produces temperature-driven and insolation-driven components of the total surface melt. Here, surface melt depends on the albedo, which in turn reduces for increasing melt. The strength of this feedback depends on the magnitude of the temperature extremes: more severe extremes generate greater melt and these simulations subsequently gain additional melt through the feedback loop. Models that do not account for this feedback could underestimate future sea-level rise by up to 2.4 m (for the scenario $I=2.0$, $f=5$). Even with moderate extremes ($I=1.25$) every 20 years, the difference in sea-level rise by 2300 reaches 1.45 m. By contrast, for the baseline scenario sea-level rise is 1.37 m by 2100. When disabling the snowmelt-albedo feedback, the difference between the dEBM-simple surface module and the PDD becomes negligible in the first 150 years, before reaching a maximum of 20 cm difference by 2300.

The effect of hydrofracturing is weaker than the snowmelt-albedo effect, with 0.53 - 0.72 m additional sea-level rise to 2300. The difference between simulations that include all processes (melt-elevation feedback, ice-albedo feedback, and hydrofracturing) and those without is up to 2 m in the baseline projection, demonstrating that feedback-related modelling choices alone can have a substantial influence on projected outcomes.

In additional simulations, we also included bedrock uplift – a potential negative feedback mechanism that could slow ice loss. When including this process, due to an overall reduction in ice loss (thinning), the sea-level rise contribution is lower, but only by -9% ($I=1.25$, $f=5$) to -5% ($I=1.25$, $f=20$) (Supplement Fig. S8).

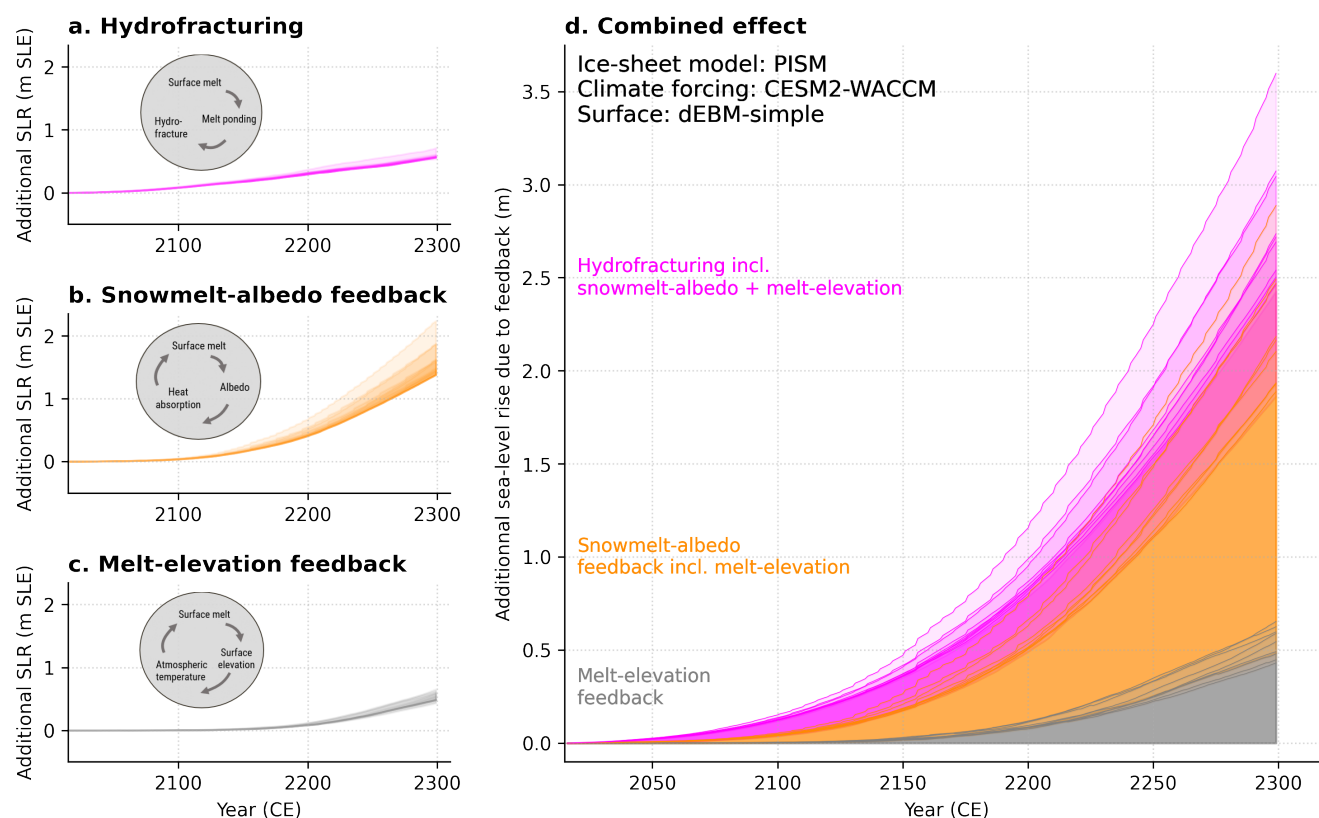


Figure 5. Assessed strengths of ice-atmosphere feedback mechanisms and processes. Shown is the additional sea-level rise when compared to simulations when these effects are not included.

4 Discussion

We have shown that Antarctica retains a “memory” of the imposed temperature extremes; the resulting shifts in its internal state, geometry and thermodynamics persist to drive dynamical feedbacks beyond the period of the forcing. Surface melt from temperature extremes is amplified by positive feedback and non-linear processes affecting the ice flow (e.g. calving). Overall, our simulations highlight the importance of the interlinked ice-atmosphere interactions that become increasingly important with ongoing warming. These are neglected when using an ice-sheet model that does not incorporate any of the melt-elevation and snowmelt-albedo feedbacks, or hydro-fracturing. Future sea-level rise could be underestimated by several metres as feedback mechanisms are, in our simulations, responsible for the majority of the sea-level rise signal. Model projections that do not account for these feedback mechanisms and processes systematically underestimate future ice loss. Because current ice-sheet models are forced by smoothed climate averages, the existing ISMIP6 projections (Seroussi et al., 2024) spread would likely be significantly amplified and shifted toward higher sea-level rise extremes if non-linear responses to episodic atmospheric events were incorporated.



380 Similar to Beckmann and Winkelmann (2023), we impose highly idealized heatwaves whose strengths are dependent on the internal climate sensitivity of the Earth System Models we used. CESM2-WACCM is known for producing a strong climate change signal in Antarctica (Coulon et al., 2025). This climate model does not have an evolving ice-sheet geometry coupled to its sub-components, thus introducing uncertainty in projections. Furthermore, we do not analyze the underlying drivers of the heat extremes. That is, we do not examine whether the extreme event is caused by climate change or if it is an expression
 385 of internal climate variability. Global circulation models have been shown to under-represent the magnitude and frequency of temperature extremes in Antarctica (Blanchard-Wrigglesworth et al., 2023), further motivating the idealized sensitivity framework employed here in which heatwaves are imposed at regular intervals to systematically assess the ice-sheet response independently of model-specific biases.

This study serves as a first-of-its-kind sensitivity study for Antarctica focusing on the isolated effect of temperature extremes
 390 on ice-sheet dynamics. We did not account for extreme snowfall events that have been associated with past heatwaves. Further, we did not tune the PDD scheme to historic or present-day observations and instead used standard literature values for parameters for the degree-day factors and melt thresholds (for snow: 3.29 mm PDD^{-1} , for ice: 8.79 mm PDD^{-1}). These parameters were thus not calibrated to our specific climate forcing or ice-sheet geometry and do not vary spatially. As a result, the scheme may introduce biases in surface melt (Zheng et al., 2023) and thus SMB, contributing to the inherent model drift (Fig. A1).

395 The simulations with the same ice-sheet model (PISM, using a PDD scheme and a very high-emission scenario) in Beckmann and Winkelmann (2023) show that the sea-level contribution from Greenland might increase by 2 to 45 cm by 2300, a 0.2% to 14% increase depending on the extreme event scenario. The results presented here for Antarctica are more extreme, with an increase of more than 2 m in the worst-case scenario. This larger increase is mainly due to the size difference of the two ice sheets, but, as shown, ice loss is amplified due to several feedback mechanisms that could play a large role in Antarctica in the
 400 future (note: the snowmelt-albedo feedback was not considered in Beckmann and Winkelmann, 2023).

As we do not observe extensive grounding line retreat by 2300, especially in the Amundsen Sea region as some studies suggest (Reese et al., 2023), we also tested a more sensitive grounding line scheme in PISM. In this, the surface gradient is calculated using centred differences of the ice thickness across the grounding line to improve the approximation of driving stress across the grounding line (instead of using only upstream grid cells). The alternative scheme does not yield significant
 405 differences by 2100 or 2300 (Supplement Fig. S9): for the baseline scenario, the difference is 0.03 m, for the most extreme heatwave scenario ($I=2.0$, $f=5$) 0.02 m, sea-level equivalent ice volume, both minor compared to the overall simulated ice loss.

To study the effect of calving on the sea-level rise contribution of Antarctica, we also disabled thickness and eigen-calving schemes. Unlike in our primary simulations, simulations with no calving lead to 0.3 m less ice loss in 2300 (Supplement Fig. S10). However, the implementation of the hydrofracturing process needs further consideration, in particular through in-
 410 vestigation of an appropriate coupling timescale. In this study, we use an offline-approach: we first evaluate the melt rates from the scenario without hydrofracturing and in a second simulation apply the updated frontal retreat mask. We anticipate that developing a full implementation of online-coupling, with yearly coupling timescale, would enhance the simulated melt-calving relationship, amplifying sea-level rise. This resolution was prohibitively expensive for our scenarios. We also did not include or test different sub-shelf melt schemes for the two models, a major source of uncertainty in recent multi-model sea-level rise



415 projections (Beckmann et al., 2025). However, here we concentrate on surface processes and their uncertainty space, inves-
tigating for the role of feedback processes that are typically neglected in intercomparison studies (cf. Seroussi et al., 2024).
With our simulations we demonstrated a growing importance of surface processes and feedback mechanisms towards 2300 that
represent a similar magnitude of uncertainty as the basal melt sensitivity of different models (Beckmann et al., 2025). Even
for models that simulate a sea-level lowering by 2300 through ice mass gain (e.g. as shown for ISSM), surface extremes can
420 change the sign of sea-level rise.

5 Conclusion

Our study supports previous findings that the primary source of uncertainty in sea level rise lies in the choice of the underlying
ice-sheet model, but that in the case of very strong warming, the external forcing has an increasing effect in the ice-sheet
response (Coulon et al., 2025). The parameters controlling ice–atmosphere interactions become a key source of uncertainty
425 by 2300, similar to the importance of feedback processes. Regardless of the forcing responsible for these heatwaves, we have
shown that feedback processes at the ice–atmosphere interface must be considered when assessing the response of the ice sheet
to future climate change, especially atmospheric extreme events. In particular, the snowmelt–albedo feedback significantly
influences the magnitude of the simulated ice-sheet changes in response to a warmer atmosphere and can increase Antarctica’s
future sea-level rise contribution significantly. The use of a traditional PDD scheme in ice-sheet models does not account for
430 this effect and thus might be limited in simulating Antarctic SMB changes in the future. While increasing snowfall alone might
act as mitigating sea-level rise in the future, atmospheric temperature extremes will counteract this and, through increasing
surface melt, can even shift the sign of the overall sea-level response. Without adequately representing the relevant feedback
processes at the ice–atmosphere interface, ice-sheet model projections may not accurately represent future change, independent
of how frequent and severe temperature extremes will occur in Antarctica. Monitoring and understanding the drivers of extreme
435 events in Antarctica will be vital in assessing a realistic upper range for the ice-sheet sea-level contribution and thus inform
forward-looking adaptation measures.



Code and data availability. The data and code necessary to reproduce the findings and figures of this study will be made publicly available in a data repository upon final publication. The PISM source code is available from <https://github.com/pism/pism> and the ISSM source code can be freely downloaded from <https://github.com/ISSMteam/ISSM>. For PISM version 2.2 see <https://github.com/pism/pism/releases/tag/v2.2.0> and for ISSM version v4.24 see <https://github.com/ISSMteam/ISSM/releases/tag/v4.24>.

Appendix A: Additional figures and tables

A1

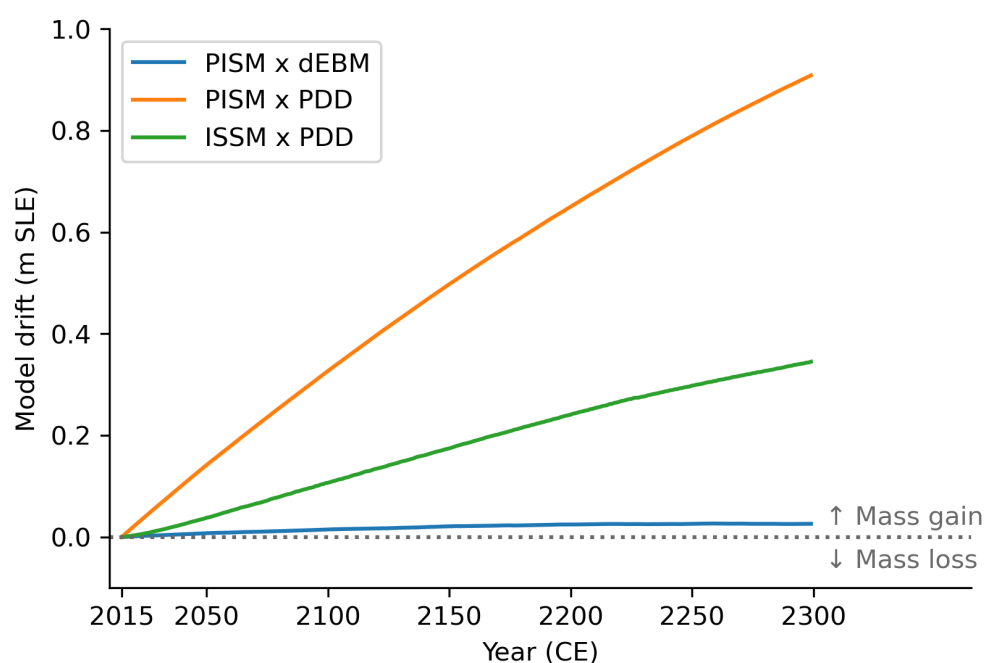


Figure A1. Control simulations (DRIFT) indicating underlying model drift. The change in ice mass when present-day climate conditions (in atmosphere and ocean) are held constant is subtracted from the sea-level rise projections in Fig. 2.

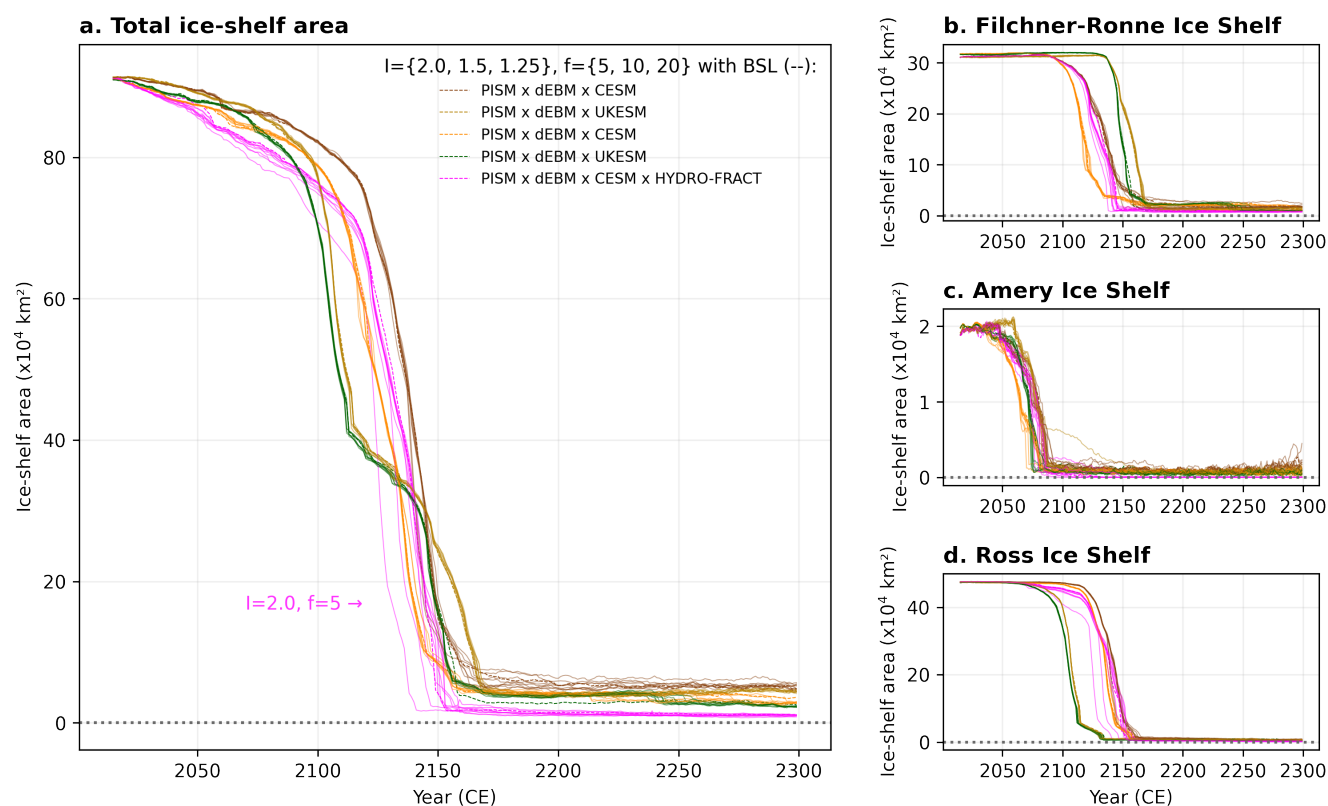


Figure A2. Antarctic ice-shelf area. a. Total ice-shelf area for the different model configurations and scenarios (BSL, heatwaves). b.-d. Same as a. but plotted only for Filchner-Ronne, Amery and Ross Ice Shelf respectively.

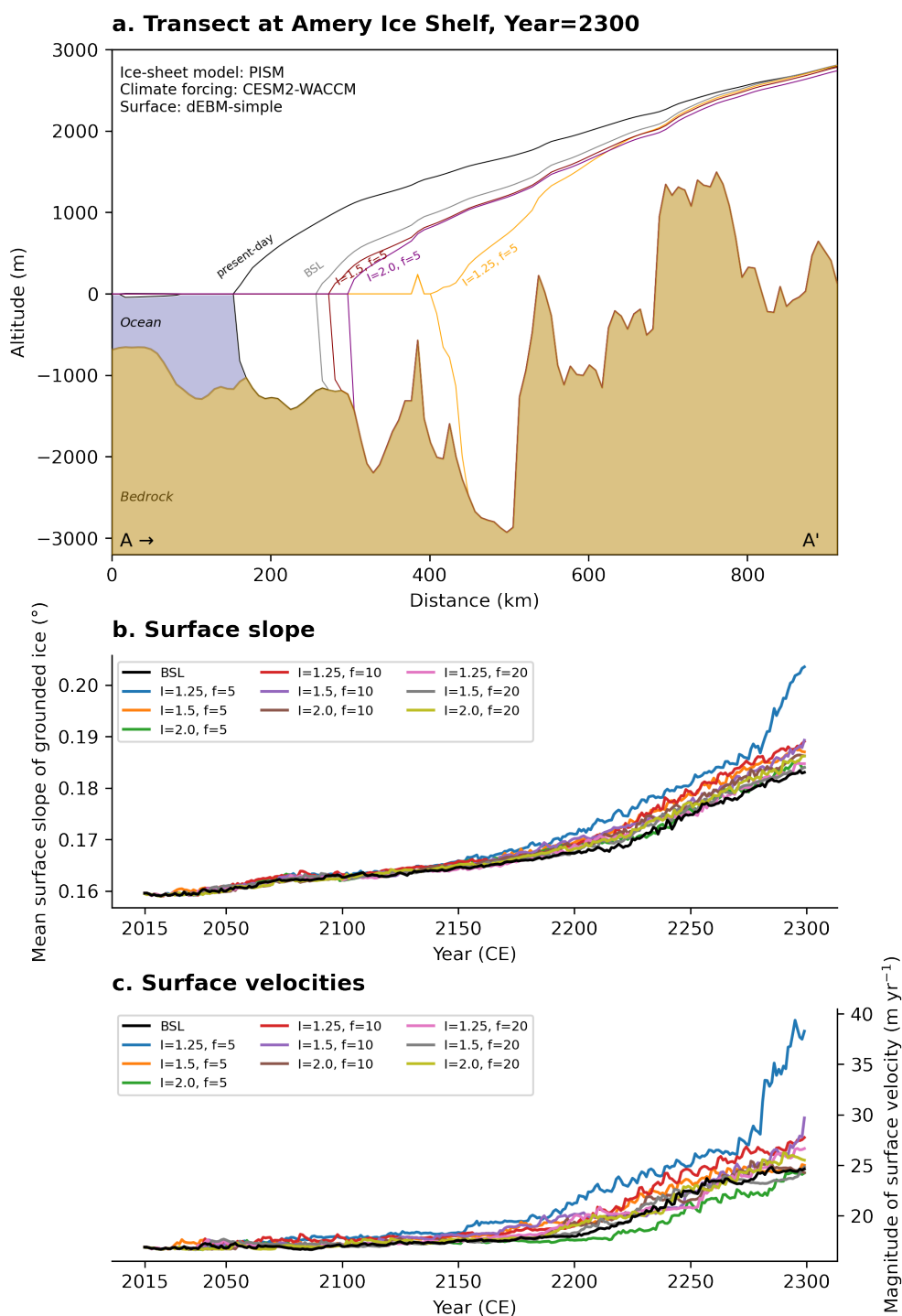


Figure A3. Changes at Amery Ice Shelf. a. Grounding line positions in Year 2300 for selected extreme event scenarios. b. Change in mean surface slope in region. c. Evolution of surface ice velocities of grounded ice in basin.

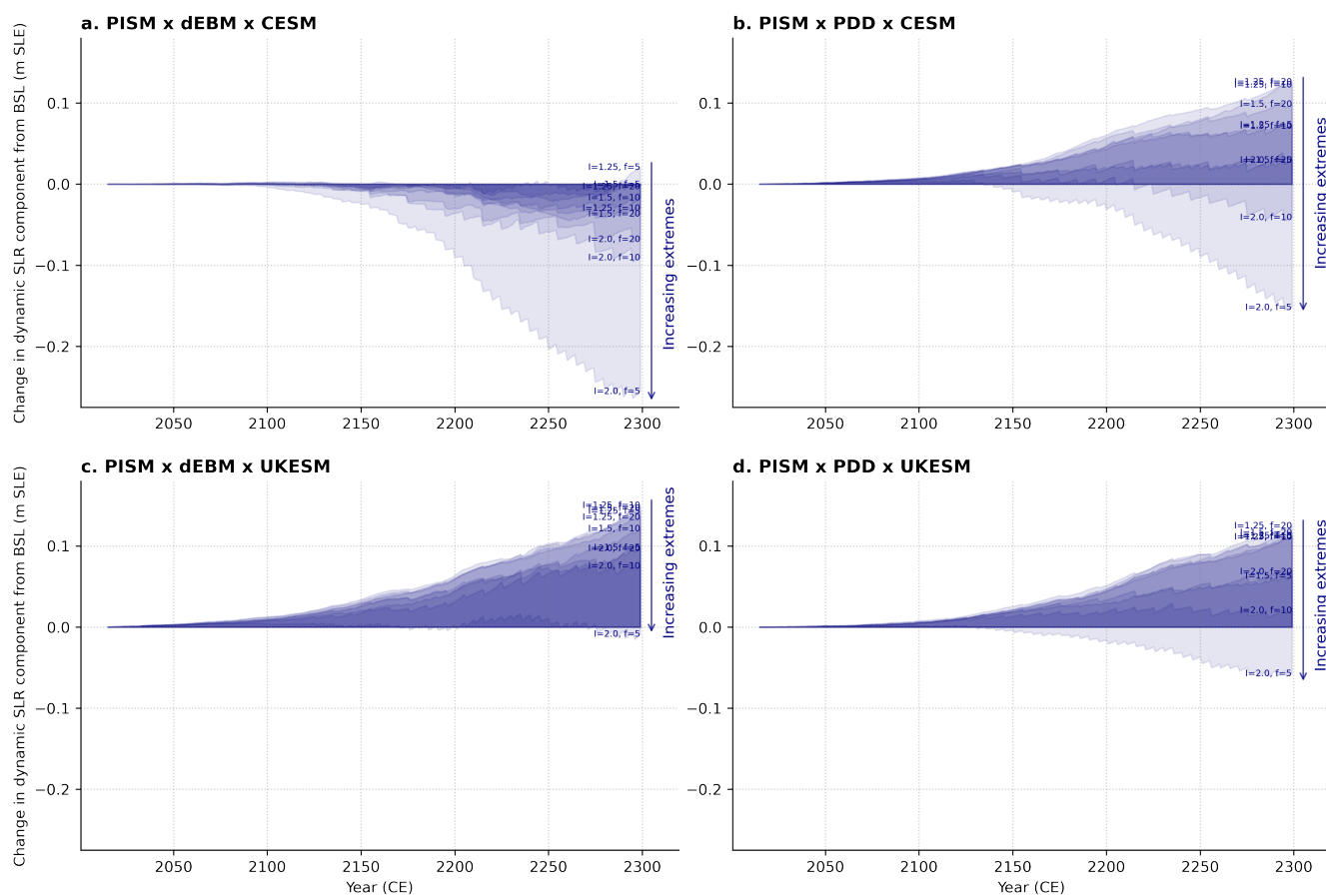


Figure A4. Change in the dynamic sea-level rise contribution. Shown is the dynamic ice loss as simulated by the ice-sheet model PISM using the dEBM and PDD scheme and two different climate model forcings relative to the respective baseline scenario.



Table A1. Summary of differences between model configurations.

	PISM	ISSM
Model initialization	Pre-industrial spin-up, historic period	Inversion in present-day
Grid	regular, 8 km (horizontal), 13 m-100 m (vertical)	adaptive, 1 - 50 km (horizontal only)
Stress balance, sliding, friction	SSA+SIA (hybrid), Glen-Paterson-Budd-Lliboutry-Duval flow law based on Glen-Paterson-Budd	SSA only, Budd friction law
Parametrizations		
Surface (SMB)	dEBM-simple, PDD	PDD
Refreezing	Fraction of meltwater (0.5)	based on capillary retention and latent heat
Sub-shelf melt	PICO	ISMIP6 Quadratic local (Jourdain et al., 2020)
Calving	Thickness (H=50 m), Eigen-calving	Thickness (H=50 m)
Frontal retreat	Growth limited to present-day ice extent, offline melt coupling (additional experiments)	-
Present-day climatology		
Surface Mass Balance	RACMO2.3p2-ERA5 product	RACMO2.3p3-ERA5 product
Ocean temperatures	bias-corrected potential temperature from ISMIP6	ISMIP6 thermal forcing



Author contributions. LN and RW conceived the study. LN and JB further designed the methodology, while LN curated the data and software, performed the formal analysis, and wrote the manuscript. All authors contributed to the writing of the manuscript.

445 *Competing interests.* One author (FSM) is an editor of TC. The authors declare no other competing interests or conflicts of interest.

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