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Effect of uncertainty in surface mass balance elevation feedback on projections of the future sea level contribution of the Greenland ice sheet – Part 2: Projections

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Abstract

We apply a new parameterisation of the Greenland ice sheet (GrIS) feedback between surface mass balance (SMB: the sum of surface accumulation and surface ablation) and surface elevation in the MAR regional climate model (Edwards et al., 2013) to projections of future climate change using five ice sheet models (ISMs). The MAR climate projections are for 2000–2199, forced by the ECHAM5 and HadCM3 global climate models (GCMs) under the SRES A1B emissions scenario.

The additional sea level contribution due to the SMB-elevation feedback averaged over five ISM projections for ECHAM5 and three for HadCM3 is 4.3 % (best estimate; 95 % credibility interval 1.8–6.9 %) at 2100, and 9.6 % (best estimate; 95 % credibility interval 3.6–16.0 %) at 2200. In all results the elevation feedback is significantly positive, amplifying the GrIS sea level contribution relative to the MAR projections in which the ice sheet topography is fixed: the lower bounds of our 95 % credibility intervals (CIs) for sea level contributions are larger than the “no feedback” case for all ISMs and GCMs.

Our method is novel in sea level projections because we propagate three types of modelling uncertainty – GCM and ISM structural uncertainties, and elevation feedback parameterisation uncertainty – along the causal chain, from SRES scenario to sea level, within a coherent experimental design and statistical framework. The relative contributions to uncertainty depend on the timescale of interest. At 2100, the GCM uncertainty is largest, but by 2200 both the ISM and parameterisation uncertainties are larger. We also perform a perturbed parameter ensemble with one ISM to estimate the shape of the projected sea level probability distribution; our results indicates that the probability density is slightly skewed towards higher sea level contributions.

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1 Introduction

The Greenland ice sheet (GrIS) response to climate change has two parts: surface mass balance (SMB), which is the sum of surface accumulation and surface ablation (broadly speaking, the balance of snowfall versus meltwater runoff); and dynamic, the changes in ice flow and discharge from the ice sheet. Various approaches to simulating these have been taken for making projections of the GrIS contribution to sea level. As with all simulation problems, there is a trade-off between representing more processes, with the aim of increasing the physical realism of the simulations, and technical and computational resources, which limit implementation and the number of simulations.

The dynamic response is simulated with ice sheet models (ISMs), which solve the Stokes equations in complete or approximate form. SMB can be simulated with sophisticated, physically-based energy balance schemes in regional climate models (RCMs) such as MAR (Modèle Atmosphérique Régional: Fettweis, 2007) and RACMO2/GR (e.g. Ettema et al., 2009). The two aspects of the GrIS response can thus be combined by forcing an ISM with SMB simulated by an RCM. But RCMs usually use a fixed surface topography, neglecting the important effects of ice sheet surface elevation changes on the atmosphere (Edwards et al., 2013) by omitting the SMB-elevation feedback. An alternative to using SMB from an RCM is to simulate it within the ISM, so that the evolving ice sheet topography can dynamically alter the SMB. The most common approach to this is with a positive degree day (PDD) scheme, which parameterises SMB as a function of temperature and precipitation (supplied from observations for the present day, or a regional or global climate model for future projections) and possibly applies a simple snow pack model (e.g. Janssens and Huybrechts, 2000). PDD models incorporate the temperature aspect of the SMB-elevation feedback through a lapse rate correction, but not the precipitation aspect (except, in some cases, through a scaling factor for temperature), and also represent SMB much more simply than RCMs such as MAR and RACMO2/GR (Edwards et al., 2013).

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To make the best use of physically-based simulations of the GrIS response to climate change – simulating ice flow with an ISM and SMB with an RCM, while also including the SMB-elevation feedback – one can couple an ISM to an RCM. However, this is technically challenging and computationally expensive, effectively precluding exploration of climate and ice sheet modelling uncertainties.

The only way, therefore, to incorporate physical modelling of ice flow, SMB processes, and the SMB-elevation feedback while also exploring model uncertainties is with a parameterisation such as the one we present in a companion paper (Edwards et al., 2013), where we characterise the SMB response to elevation in MAR using a suite of simulations in which the MAR GrIS surface height is altered. The parameterisation is a set of four gradients, or “SMB lapse rates”, that relate SMB changes to height changes below and above the ELA and for regions north and south of 77° N. Here we apply the parameterisation in five ISMs to adjust MAR projections of SMB under the SRES A1B scenario (Nakićenović et al., 2000) as the ice sheet geometry evolves.

The climate community have been attempting to quantify uncertainty in global climate model (GCM) predictions for some time, focusing on uncertainty in their parameter values with perturbed parameter ensembles (PPEs) but also attempting to estimate structural uncertainty, which is uncertainty about the remaining discrepancy between a model and reality at the model’s most successful parameter values (e.g. Sexton et al., 2011; Sexton and Murphy, 2011). But probabilistic quantification of uncertainties in ice sheet model predictions has hardly yet been attempted. There is also an urgent need to propagate uncertainties along the causal chain from greenhouse gas forcing scenarios to the impacts of climate change. ISM projections are only beginning to tackle these challenges. Multi-model comparisons such as MISMIP (Pattyn et al., 2012) and SeaRISE (Bindschadler et al., 2013) compare ISMs with each other using standardised experiments, but the ice2sea project (<http://www.ice2sea.eu>) is among the first to systematically explore emissions scenarios and climate and ice sheet model uncertainties within a coherent framework (e.g. Rae et al., 2012; Fettweis et al., 2012).

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In Edwards et al. (2013), we make uncertainty assessment an integral part of the parameterisation by estimating probability distributions for the four elevation feedback gradients. Here we propagate these uncertainties to future projections by sampling values from the four distributions. We also use five different ISMs to assess the effect of ISM structural uncertainty that arises from different representations of ice flow and initialisation procedures, and explore GCM structural uncertainty by forcing MAR with two different GCMs. Our coherent experimental design and statistical framework, unusual in sea level projections, allow us: to propagate the three types of model uncertainty along the causal chain from SRES scenario to sea level contribution; to assess the relative importance of these through time; and to present probabilistic assessments of the effect of elevation feedback parameterisation uncertainty on the projected GrIS sea level contribution under the A1B scenario.

2 Method

2.1 Climate projections

The regional climate model MAR (Fettweis, 2007) has been adapted for simulating the climate over Greenland, with full coupling to a complex snow-ice energy balance model and relatively high horizontal resolution (25 km). Unlike most RCMs, MAR includes the positive feedback between ice surface albedo and melting (Franco et al., 2013). We discuss the processes and responses of MAR in more detail in Edwards et al. (2013).

We use five climate simulations performed for the ice2sea project. The first is a twenty year simulation of 1989–2008 in which MAR is forced at the boundaries by the ERA-INTERIM reanalysis (Dee et al., 2011), which we use as a baseline for initialisation and projections. The next two are twenty year simulations of 1980–1999 in which MAR is forced by two GCMs, ECHAM5 (Roeckner et al., 2003) and HadCM3 (Gordon et al., 2000), under 20th century climate forcings, which we use to calculate projection anomalies. The final two are one hundred year projections (2000–2099) forced by the

two GCMs under the SRES A1B emissions scenario (Nakićenović et al., 2000). We wish to make GrIS projections for two hundred years (2000–2199), so we extend the MAR simulations by repeating the final decade (2090–2099) ten times.

MAR simulates SMB values only over grid cells categorised as permanent ice, so the low spatial resolution (relative to ISMs) leads to missing values for parts of the GrIS margin. We generate SMB values for each Greenland land grid cell by using a linear fit of SMB versus surface height from ice grid cells within 100 km.

2.2 Ice sheet models

We implement the parameterisation in five ISMs with varying structure and complexity: Elmer/Ice, which solves the full Stokes equations; GISM, MPAS and CISM, which use a higher order approximation that reduces computational expense (HO: e.g. Pattyn, 2003); and GRISLI, which uses a hybrid of the first order shallow ice and shallow shelf approximations (SIA and SSA: Bueler and Brown, 2009) to further reduce computational expense. We also use a SIA version of GISM, and call the two versions GISM-HO and GISM-SIA. Elmer/Ice and MPAS use finite element numerical methods on unstructured grids, while the others use finite difference methods on regular grids at 5 km resolution.

Elmer/Ice builds on Elmer, the open-source parallel code mainly developed by the CSC-IT Center for Science Ltd in Finland. The unstructured mesh allows a variable grid resolution to focus computational resources at the ice sheet margin; here we use a minimum horizontal grid size of less than 1 km. GISM-SIA is a thermomechanical ISM (Huybrechts and de Wolde, 1999), which has been modified and extended for projections on centennial timescales using a new higher-order approximation of the force balance (GISM-HO: Fürst et al., 2011, 2013). The MPAS-Land Ice model is based on the MPAS (Model for Prediction Across Scales) climate modelling framework of Ringler et al. (2008); here we use a regular 5 km resolution hexagonal mesh. GRISLI (GRenoble Ice Shelves and Land Ice model) is a thermomechanically coupled ISM; it is a hybrid model that for grounded ice uses the SIA for vertical shearing and the SSA as a

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sliding law (Ritz et al., 2001; Bueler and Brown, 2009). The Community Ice Sheet Model (CISM) version 2.0 includes improvements to all components of the Glimmer-CISM SIA model (Rutt et al., 2009). More detailed information is given elsewhere (Elmer/Ice: Gillet-Chaulet et al., 2012; GISM-HO: Goelzer et al., 2013; MPAS: Perego et al., 2012; CISM: Price et al., 2011; Lemieux et al., 2011; Evans et al., 2012; GRISLI: Quiquet et al., 2012).

2.3 Initialisation

Determining initial conditions for ISMs involves finding a balance between observations of the present day ice sheet, reconstructions of past climate changes (to which the ice sheet is still responding), and the physical laws and parameterised processes incorporated in the ISM, while accounting for uncertainties and limitations in all of these. ISM initialisation mostly uses ad-hoc tuning methods rather than formal data assimilation as in numerical weather forecasting. Our initialisation procedures use observations of present day ice sheet geometry (surface elevation, bedrock elevation and ice thickness: e.g. Griggs et al., 2012), ice velocities (Joughin et al., 2010; Bamber et al., 2000), and geothermal heat fluxes (e.g. Shapiro and Ritzwoller, 2004).

Methods vary between ISMs but in general there are three stages. During the first two we fix the shape of the ice sheet to the observed geometry. First, we estimate the 3-D ice temperature field by solving the heat equation, in some cases accounting for the response to atmospheric temperature changes over one or more glacial-interglacial cycles. Second, we infer the basal drag coefficient (single-valued or a spatial pattern) that leads to the best agreement with observed ice velocities, given observational uncertainties and model limitations. Third, we allow the ice sheet geometry to evolve or “relax” so that it is internally consistent with the ice temperature and flow fields.

We derive ice temperatures for Elmer/Ice using the computationally cheaper SIA model SICOPOLIS (Greve, 1997); for CISM, MPAS and GRISLI from a quasi-steady state CISM simulation using the surface temperature and geothermal heat flux datasets from the SeaRISE project

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(http://websrv.cs.umt.edu/isis/index.php/Present_Day_Greenland); and for the two GISM models from a GISM-SIA simulation of several glacial-interglacial cycles, rescaling the ice temperature field to the observed ice thickness. We infer the basal drag coefficient from observed velocities using the control method (Elmer/Ice: Morlighem et al., 2010), an iterative inverse method (GRISLI), or tuning (CISM, MPAS). We obtain the initial ice sheet geometry by forcing the ISM with a present day mean SMB field. For most models this is the 1989–2008 mean (the reference period for the ice2sea project) of the ERA-INTERIM forced MAR simulation; for GISM we use the 1960–1990 mean SMB calculated with the GISM PDD scheme from ERA-40 data (Uppala et al., 2005) and observed surface elevation (similar to Hanna et al., 2011). We obtain the initial geometry for Elmer/Ice by allowing the upper surface to evolve for 55 yr under the present day SMB forcing, and use the same initial geometry for GRISLI, allowing it to evolve for a further 145 yr to obtain a coherent initial state for this model. We initialise the GISM models by allowing the geometry to evolve for 1000 yr while the ice sheet margin is fixed to observed values under present day SMB forcing and ice thickness changes are limited to 0.2 myr^{-1} . We do not relax the geometry for CISM and MPAS.

We use two methods to correct any remaining drift that arises from model imbalances. For Elmer/Ice and GRISLI we perform a control simulation and subtract this from the results (Sect. 2.5). For the others we diagnose and apply a “synthetic SMB” correction, which is the additional SMB required to keep the ice sheet close to the present day observed geometry under present day SMB forcing. The synthetic SMB correction is applied unaltered in perturbed simulations, similar in principle to the flux corrections that were formerly in common use in atmosphere–ocean GCMs. In CISM and MPAS there is no relaxation in the initialisation procedure, so the synthetic SMB accounts for the entire initial mass imbalance generated by the initial geometry.

We exclude Greenland’s glaciers and ice caps from the projection results using a mask produced for ice2sea (Rastner et al., 2012). More details on initialisation procedures can be found in the references given in the previous section.

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2.4 Boundary conditions

The boundary conditions are ice temperature, bedrock elevation, and projections of SMB. We keep the ice temperature fields determined during initialisation fixed, because we expect negligible ice temperature response to changing atmospheric forcing over two centuries. We use bedrock elevations from the ice2sea dataset (Griggs et al., 2012) and hold these fixed because we also expect negligible isostatic adjustment on this timescale.

The SMB forcings are therefore the only time-varying boundary conditions. We force the ISMs with “anomaly-corrected” SMB projections, $S^{RCM'}$, to remove the mean discrepancy in MAR between the GCM-forced and ERA-INTERIM reanalysis-forced simulations (Fettweis et al., 2012). For this we calculate SMB anomalies with respect to the present day by subtracting the 1989–2008 mean SMB (1989–1999: 20th century climate forcings; 2000–2008: A1B scenario) from the A1B projections. We add these anomalies to the present day SMB used for initialisation, to give:

$$S_{2000-2199}^{RCM'} = S_{2000-2199}^{A1B} - S_{1989-2008}^{20C,A1B} + S^{init},$$

where S^{init} is generally $S_{1989-2008}^{ERA1}$, the 1989–2008 mean of the ERA-INTERIM forced simulation (except GISM, $S_{1960-1990}^{ERA40}$). If used, the synthetic SMB field is also applied. Our initialisation and anomaly-correction methods use the approximation that mean SMB changes during the period 1989–2008 are small relative to the A1B projections (Rae et al., 2012).

2.5 Parameterisation

We add the SMB-elevation feedback parameterisation to each ISM, so that the anomaly-corrected SMB forcing can be adjusted as the GrIS surface height evolves. The adjustment is made each year, for each grid cell, using one of four gradients

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selected according to the current “reference” SMB and the region of the grid cell (Edwards et al., 2013).

For a given ISM grid cell in a given year t , a gradient b_t ($\text{kg m}^{-3} \text{a}^{-1}$) is used to adjust the SMB forcing ($\text{kg m}^{-2} \text{a}^{-1}$) using the height difference (m) between the previous year and the start of the simulation:

$$S_t^{\text{adj}} = S_t^{\text{RCM}'} + b_t(h_{t-1}^{\text{ISM}} - h_0^{\text{ISM}}),$$

where S_t^{adj} is the adjusted SMB, $S_t^{\text{RCM}'}$ is the anomaly-corrected RCM SMB, and h_t^{ISM} is the ISM height. The gradient b_t is selected from one of four values according to the reference SMB ($S^{\text{ref}} < 0$ or $S^{\text{ref}} \geq 0$) and region (north or south of 77°N) of the grid cell, where S_t^{ref} is the mean S^{adj} of the previous decade or, for the first decade, all available years.

We use each ISM to generate a set of five simulations:

- Control: forced with S^{init} (Elmer/Ice, GRISLI) or $S^{\text{init}} + S^{\text{syn}}$ (other models), to check for or subtract model drift from the projections;
- No feedback: forced with $S^{\text{RCM}'}$ with no adjustment, to estimate the response of the GrIS without elevation feedback;
- Best estimate: forced with S^{adj} , using the “best estimate” gradient set (see Edwards et al., 2013);
- 2.5th and 97.5th percentiles: as for best estimate, but using the gradient sets that correspond to the bounds of the 95 % “credibility interval” (CI: see Edwards et al., 2013).

We perform these for the two MAR simulations forced by ECHAM5 (all ISMs) and HadCM3 (Elmer/Ice, GISM-HO, GISM-SIA and GRISLI) under the A1B scenario over 2000–2200.

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We also sample the entire parameter distributions with GISM-SIA, performing a 99 member PPE of every percentile estimate of the gradient set (1st, 2nd, . . . , 99th percentiles, rather than only the 2.5th and 97.5th) for the ECHAM5 projection.

The drift in the control simulations is very small (0.03%, or less, of the cumulative projected sea level contribution at 2200) for all models except Elmer/Ice, which has a drift of 2–2.5% (–4 mm). For this model, drift is not constrained during the initialisation procedure and the free-surface elevation has been allowed to diverge from observations for a relaxation period of only 55 yr. The applied SMB is not corrected by a synthetic SMB, and the remaining drift shows the drift of the model when directly applying the 1989–2008 mean SMB given by MAR forced under ERA-INTERIM. This drift is corrected by subtracting the control simulation from the projections.

3 Results

The additional cumulative sea level contribution due to the SMB-elevation feedback, under the A1B scenario and averaged over five ISM projections for ECHAM5 and three for HadCM3, is 4.3% (best estimate; 95% credibility interval 1.8–6.9%) at 2100, and 9.6% (best estimate; 95% credibility interval 3.6–16.0%) at 2200 (Figs. 1 and 2; Tables 1 and 2). We exclude GISM-SIA from all summary statistics presented because it is so similar in structure, and identical in set-up, to GISM-HO that it would effectively give double weighting to GISM. In all ISM predictions the elevation feedback is positive: the lower bounds of the 95% CIs (the 2.5th percentile simulations) give larger sea level contributions than the “no feedback” case. This is expected because the mean GrIS SMB becomes negative at around 2050 (Rae et al., 2012) and five of the six gradients (best estimate and 2.5th and 97.5th percentiles, for north and south) for negative SMB are positive (Edwards et al., 2013) thus adjusting the SMB in the same direction: more negative, giving greater sea level contribution. However, this was not guaranteed, because there are both positive and negative values in the mean SMB forcing, the SMB forcing changes from 2000–2199, and the gradient sets (Edwards et al., 2013). The

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HadCM3-forced projection gives consistently higher sea level contributions than the ECHAM5-forced, because the HadCM3-forced MAR simulation projects greater melt-water runoff (Rae et al., 2012).

The Elmer/Ice and GRISLI total sea level contributions are virtually identical. This is a coincidence: the changes in ice discharge and SMB are different, but the differences just compensate each other so that the sea level contributions are the same. The two models have similar initialisation procedures, but the differing dynamic and SMB responses lead us to believe this is not the reason for the similar results. CISM seems to be an outlier in the ISM ensemble, particularly at 2200 (Fig. 2; Table 2), in both the no feedback result (low) and the magnitude of the feedback (small), though the latter may be a result of the former. The GISM-SIA results are very similar to those of GISM-HO. This can be explained by the identical model setup and their similarity in response to perturbations; the only difference between the two models is in the dynamic response, which is a much smaller contribution than the SMB response (Fürst et al., 2013).

We can compare the projections with and without SMB-elevation feedback from the two GCMs for the three ISMs that used both projections: Elmer/Ice, GISM-HO and GRISLI, from here on called the “starred” models (again, GISM-SIA is excluded here). Without elevation feedback, the mean GrIS sea level contributions at 2100 are 57 mm (ECHAM5) and 64 mm (HadCM3); with feedback, these increase by 3 mm (5%). At 2200, the mean GrIS sea level contributions without feedback are 164 mm (ECHAM5) and 176 mm (HadCM3); the feedback contributes an additional 16 mm (10%) and 19 mm (11%) respectively. The larger feedback adjustment at 2200 arises from the negative SMB forcing at the end of the first century which is sustained (by repetition of the decade 2090–2099 from the MAR simulation: Sect. 2.1) during the second century.

The results show that not only does the contribution due to the feedback increase with time, but so does the uncertainty. Our experimental design allows us to understand the changes and relative importance of the different modelling uncertainties. At 2100 (Fig. 1; Table 1), the GCM differences are, respectively, 2.6 and 1.7 times those of the elevation feedback parameterisation and ISM uncertainties: the mean difference

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between the two GCMs (averaged over the no feedback and three feedback estimates) for the three starred models is 7.8 mm; the mean spread of all five ISMs for ECHAM5 is 4.5 mm; and the mean 95 % CI (over all ISMs and GCMs) is 3.0 mm. But by 2200 (Fig. 2; Table 2), the relative importance has changed. The ISM spread and elevation feedback parameterisation uncertainty overtake the difference between the two GCMs: the mean difference between the two GCMs is 13.7 mm; the mean spread of ISMs for ECHAM5 is 25.8 mm; and the mean 95 % CI (over all ISMs and GCMs) is 20.8 mm. In other words, the parameterisation and ISM uncertainties increase by a factor of seven and six respectively between 2100 and 2200, while the difference between the two GCMs barely doubles. The relative contributions to our uncertainty thus depend on the timescale of interest.

This can be seen clearly in Fig. 3, which shows the fractional uncertainties as a function of time with a 15 yr running mean applied. At the start of the projections, the ISM structural uncertainty explored (the range of five ISM best estimates divided by their mean, for the ECHAM5 projections) is a large fraction because the signal is small, but the fraction decreases (the signal increases faster than the range) during the first half of the twentieth century. After this the fraction increases again (the range increases faster than the signal). The time-dependence of the GCM structural uncertainty explored (the difference between ECHAM5 and HadCM3 results divided by their mean, averaged over the best estimates for the three starred ISMs) appears at first glance to be similar, but the underlying reason is different: it rapidly falls to a minimum when the two GCMs projections happen to coincide for a short period (see also Rae et al., 2012). After a short increase due to the GCMs diverging again, the fractional uncertainty slowly falls because the signal increases faster than the range. The fractional uncertainty due to the parameterisation (95 % CI divided by the best estimate, averaged over all ISMs and GCMs) is initially very small, but increases fairly linearly with time: the width of the CI scales with the signal. In the latter part of the second century the parameterisation uncertainty overtakes that of the GCMs. These results are discussed further in the next section.

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The GISM-SIA PPE gives the entire sea level probability distribution (Fig. 4) rather than just the 95 % CIs. We can inspect any other interval, such as the 90 % CI (177.3–194.7 mm) or 50 % CI (182.3–190.6 mm). We can also test whether the projected distribution of sea level contributions is symmetric about the best estimate (186.1 mm). The 95 % CI, is symmetric (± 10.4 mm), and the 90 % CI nearly so (-8.8 mm, $+8.6$ mm), but the 50 % interval (-3.8 mm, $+3.5$ mm) indicates the probability density leans slightly towards the higher sea level contributions.

The unimodal and (broadly) symmetric shape of the PPE sea level distribution likely reflects the unimodal and (in most cases) symmetric shapes of the four posterior probability distributions of the gradient values (Edwards et al., 2013), combined with the linear model of adjustment. Figure 5 shows the dependence of the sea level contribution on the gradient sets. This cannot show which of the four gradients is most important, because they are varied simultaneously. But it illustrates the following: the relationship between sea level contribution and gradient set is mostly linear; the two gradients for below the ELA are the most uncertain (largest range in $\text{kg m}^{-3} \text{a}^{-1}$); and the gradient for above the ELA, south of 77°N , has the least symmetric probability distribution (equally-spaced samples in probability become increasingly far apart in $\text{kg m}^{-3} \text{a}^{-1}$).

4 Discussion

4.1 Elevation feedback

These results are the first application of an SMB-elevation parameterisation to RCM projections. Our method allows ISMs to be forced with the more physically realistic simulations of SMB from RCMs, rather than their own simpler PDD schemes, while incorporating the elevation feedback and exploring model uncertainties. Another use would be to correct low resolution MAR simulations onto a high resolution digital elevation map; the difference in elevation between the two topographies is analogous to the geometry evolution in ISM simulations. This correction might improve the realism

of MAR SMB simulations, which would be useful not only for forcing ISMs but also comparing with observations. The advantages of our parameterisation method over previous studies are described in the companion paper (Edwards et al., 2013).

It is important to try and quantify the magnitude of the SMB-elevation feedback relative to projections without it (such as Rae et al., 2012). Despite the negative gradient values we present in Edwards et al. (2013), the lower bounds of our 95 % CIs with feedback are always greater than the projections without feedback. But our results indicate that the additional contribution is relatively small (best estimates 5 % at 2100, 10–11 % at 2200), so century-long simulations with coupled RCM-ISM models may not be justified for the additional computational expense and technical difficulty of coupling. On the other hand, A1B is a medium-high emission scenario. When forced by the GCM MIROC5 under the high-end Representative Concentration Pathway RCP8.5 (Moss et al., 2010), MAR projections using perturbed ice sheet topographies consistent with SMB changes show an additional mass loss of 5–15 % over just one century (Fettweis et al., 2012). However, this estimate is with respect to projected SMB changes, while our results are evaluated with respect to the total mass change (SMB plus dynamic changes). In addition, the SMB-elevation feedback estimate from Fettweis et al. (2012) is made by perturbing the MAR ice sheet topography with corrections equivalent to the cumulative projected SMB changes over 2000–2079 without coupling MAR to an ice sheet model. In reality the ice sheet dynamics should partly compensate the surface height decreases from the melt increase in the ablation zone, by redistributing the additional mass gained at the top of the ice sheet, so the Fettweis et al. (2012) estimates likely overestimate this contribution. Only a coupled RCM-ISM would allow full simulation of the SMB-elevation feedback.

Previously some have suggested that PDD descriptions of ice sheet response are too sensitive to climate change (van de Wal, 1996; van de Berg, 2011), but comparisons by Goelzer et al. (2013), Helsen et al. (2011), Vernon et al. (2012) and Hanna et al. (2011) find they are less sensitive than the more physically-realistic representations of MAR and RACMO/GR. Goelzer et al. (2013) compare projections made with our

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parameterisation with the GISM PDD scheme, using the same ice sheet extent and RCM. Helsen et al. (2011) also compare their parameterisation of the RACMO2/GR SMB-elevation feedback with PDD results. In both cases, the PDD schemes predict a smaller sea level contribution in a warmer climate than the RCMs: for example, Goelzer et al. (2013) find the PDD model gives a 14–33 % lower sea level contribution than MAR. There are many differences between physically-based energy balance RCM schemes and empirical PDD schemes, including the presence of albedo feedback in the RCMs, higher spatial resolution in the PDD schemes, and different treatment of re-freezing (Goelzer et al., 2013). Our parameterisation of the MAR SMB-elevation feedback as four probability distributions is a more complex representation of the feedback than PDD schemes, which use a fixed temperature lapse rate: it incorporates the precipitation aspect, and the variation of the feedback with climate, topography, and region. However, the response of SMB to a modified topography is complex and nonlinear. Even with full inclusion of uncertainties, a gradient-based (i.e. linear) parameterisation cannot fully represent the SMB changes that would be modelled by a fully coupled RCM-ISM.

4.2 Modelling uncertainties

Our method is novel because we propagate three types of model uncertainty – GCM structural uncertainty, ISM structural uncertainty, and elevation feedback parameterisation uncertainty – along the causal chain from SRES scenario to sea level within a coherent experimental design and statistical framework. In particular, parameter uncertainty is estimated with a probabilistic method, which gives well-defined credibility intervals (CIs) rather than simple sensitivity tests; all ISMs use the same parameterisation and parameter sampling, and are forced with the same GCMs and RCM, enabling comparisons across ISMs; and MAR is forced with two GCM forcings under the same SRES scenario, enabling comparisons between ECHAM5 and HadCM3. A similar approach is taken by Shannon et al. (2013) for a new parameterisation of GrIS basal lubrication. This experimental design gives us the opportunity to assess the relative

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importance of a range of parametric and structural uncertainties and how these vary with time. We refer to Fettweis et al. (2012) for a discussion of the effects of changing the MAR tundra/ice mask and ice sheet topography, and using a range of GCMs for forcing, on simulations of current climate and projections of future climate change.

5 Parametric uncertainties result in a unimodal and fairly symmetric probability distribution for sea level contribution (Fig. 4), because of the linear SMB adjustment and the sampling from unimodal and (in most cases) symmetric probability distributions for the gradients (Edwards et al., 2013). If the gradient distributions derived from MAR were strongly asymmetric or multimodal, or the SMB adjustment nonlinear, the sea level
10 distribution might be a different shape. Here is the real value of Bayesian estimation of parameters (Edwards et al., 2013) and sampling these uncertainties with perturbed parameter ensembles, because such things are generally not known in advance. The slight asymmetry in the sea level probability distribution described in the previous section might become more pronounced for timescales longer than two centuries.

15 If substantial computational resources and time are available, a more thorough exploration of parametric uncertainty would perturb other parameters of the ISMs and parameters of the climate models. Climate model PPEs are used extensively by Murphy et al. (2009). Two ISM PPEs are presented by Stone et al. (2010) and Applegate et al. (2011), who vary multiple parameters simultaneously in order to explore their interactions. For example, surface lowering driven by parameterised dynamic processes
20 (such as basal lubrication: Shannon et al., 2013) could interact differently with the elevation feedback depending on the values of the relevant parameters. Performing multiple parameter perturbations might affect the shape of the distribution of projected sea level contributions.

25 Structural uncertainties are much more challenging to assess, because there is no well-defined space from which to sample (as there is for continuous or discrete parameters). Using a coherent experimental design for models, parameterisation, and parameter values is the key to systematic exploration, and therefore better understanding, of structural uncertainties. But there is currently no consensus about how to assess the

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extent to which an ensemble of models represents our uncertainty about the system they simulate. Nonetheless, it is clearly better to sample from this space, to explore it partially, than to ignore it.

Here the structural uncertainty explored most is in the ISMs (an ensemble of five).

5 Structural differences in ice sheet modelling includes representation of the Stokes equations, numerical methods, grid resolution, initialisation procedures, and drift correction. Goelzer et al. (2013) compare GrIS sea level projections for different model formulations. We explore some GCM structural uncertainty by using two GCMs. We emphasise that we do not claim to quantify ISM and GCM structural uncertainties, only
10 to sample from them. Our multi-model ensembles are small, and models have common components and errors that are likely to reduce the ensemble range relative to our actual uncertainty about the climate and GrIS. However, using multi-model ensembles is a necessary first step along that path.

We do not explore RCM structural uncertainty. An ensemble of RCMs would require
15 a new parameterisation of the SMB-elevation feedback for each. We do incorporate uncertainty in the structure of our parameterisation of MAR by including a discrepancy variance term when estimating the gradients (Edwards et al., 2013). We emphasise that this discrepancy variance refers to our parameterisation of the MAR elevation feedback, not the MAR simulation of the feedback in the real world. The latter is the RCM
20 structural uncertainty. We could sample from this with multiple RCMs, but quantifying it is very challenging due to their computational expense; such estimates are relatively rare (e.g. Sexton and Murphy, 2011).

We note that a small spread in model results does not correspond to certainty about the future: for example, models may not be independent (e.g. share code), or might co-
25 incidentally give similar results for different reasons (such as Elmer/Ice and GRISLI). It is therefore a research priority to make progress in quantifying model structural uncertainties. The ice sheet modelling community lags the climate community in this regard, but might use this to their advantage by learning from past discussions.

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Our quantification of the relative magnitude of modelling uncertainties through time (Fig. 3) is therefore specific to our particular choices and opportunities: the RCM, GCMs and ISMs; the emissions scenario; our method of extending the scenario from 2100 to 2199; and (trivially) our chosen CI width. If we used a scenario in which the SMB projections were more negative than for A1B, such as RCP8.5, or less negative, such as E1 (Rae et al., 2012), the elevation feedback would be correspondingly larger or smaller and so would the uncertainty. Similarly, if we were to use 200 yr GCM and RCM simulations, i.e. drive MAR with GCM simulations that were forced in the second century with the A1B scenario at 2100, rather than repeating the decade 2090–2099 from the MAR simulations ten times, the SMB forcing would likely continue to decrease rather than remain fixed. This would lead to a more rapid increase in the sea level projections with elevation feedback. The ISM fractional uncertainty might then not increase in the second century, or not as much (Fig. 3); this would depend on the relative responses of the ISMs. The GCM range might increase more quickly than the mean, giving a time-dependent fractional uncertainty that is more similar in shape to the ISM fractional uncertainty (Fig. 3). Such choices are unavoidable given the substantial computational expense of a three stage model chain in which two stages are climate models. Nevertheless, this work begins the process of determining where to focus computational resources and model development for different decision-making timescales.

5 Summary

We present the first application of a parameterisation of the GrIS SMB-elevation feedback to projections of future climate change. This is currently the only practicable method for making projections for the GrIS contribution to sea level that can incorporate physical modelling of ice flow, SMB processes, and the SMB-elevation feedback while also exploring modelling uncertainties. We estimate parameterisation uncertainty and explore GCM and ISM structural uncertainties within a coherent experimental design

and statistical framework. Projections are made for the SRES A1B emissions scenario with the MAR RCM, ECHAM5 and HadCM3 GCMs, and five ISMs. The experimental design allows us to estimate the relative importance of these model uncertainties, and how they vary over time, for the GrIS contribution to sea level over the next two centuries.

The additional sea level contribution due to the SMB-elevation feedback, averaged over five ISM projections for ECHAM5 and three for HadCM3 is 4.4 % (best estimate; 95 % credibility interval 1.8–6.9 %) at 2100, and 9.6 % (best estimate; 95 % credibility interval 3.6–16.0 %) at 2200. In all results, the lower bounds of our 95 % credibility intervals for sea level contributions with the SMB-elevation feedback are larger than for the “no feedback” case.

The relative contributions to uncertainty vary with time. In 2100, the GCM differences are larger than the ISM range and the parameterisation uncertainty, but by 2200, the ISM and parameterisation uncertainties are greater. This work indicates that the areas where future research should be directed – such as understanding climate or ice sheet model differences, or performing perturbed parameter ensembles – depends on the time scale of interest. For sufficiently large ensemble sizes, this method can be used to give an indication where future sea level research should be directed.

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Table 1. Projected sea level contribution (mm) at 2100 for the ECHAM5 (top) and HadCM3 (bottom) A1B projections: no feedback, 2.5th percentile, best, and 97.5th percentile estimates.

Model	No feedback	2.5th	Best estimate	97.5th
ECHAM5				
CISM	53.7	53.8	54.7	56.9
Elmer/Ice*	57.0	58.2	59.8	60.9
GISM-HO*	55.7	56.8	58.3	59.8
GISM-SIA	56.6	57.6	59.2	60.6
GRISLI*	57.1	58.1	59.6	61.0
MPAS	57.4	58.4	60.0	61.4
Mean (*)	56.6	57.7	59.2	60.5
HadCM3				
Elmer/Ice*	64.8	66.1	67.9	69.2
GISM-HO*	62.9	64.3	66.0	67.7
GISM-SIA	63.5	64.9	66.7	68.3
GRISLI *	64.7	66.0	67.7	69.3
Mean (*)	64.1	65.4	67.2	68.7

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Table 2. As for Table 1, but at 2200.

Model	No feedback	2.5th	Best estimate	97.5th
ECHAM5				
CISM	150.8	151.3	158.1	173.1
Elmer/Ice*	164.9	171.3	181.0	189.5
GISM-HO*	160.3	166.9	176.7	186.6
GISM-SIA	160.4	166.9	176.8	186.6
GRISLI*	165.7	171.6	181.2	190.8
MPAS	170.5	177.7	188.7	199.6
Mean (*)	163.6	169.9	179.6	189.0
HadCM3				
Elmer/Ice*	177.5	184.7	195.8	205.7
GISM-HO*	171.9	179.6	191.1	202.3
GISM-SIA	171.7	179.3	190.7	201.9
GRISLI*	177.2	184.1	195.0	206.0
Mean (*)	175.5	182.8	194.0	204.6

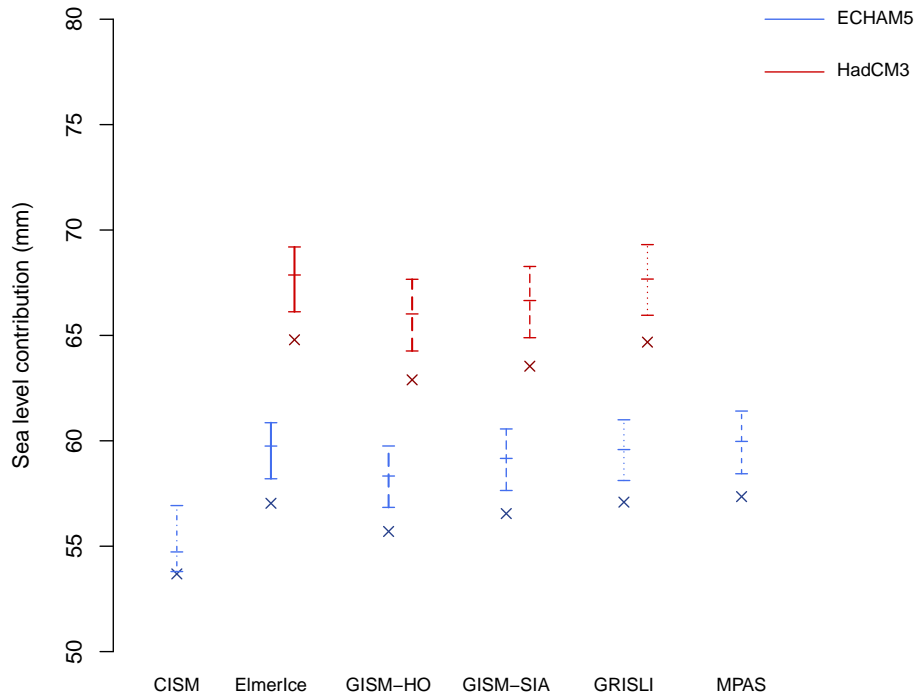


Fig. 1. Projected cumulative GrIS sea level contributions for the ECHAM5 and HadCM3 A1B projections at 2100. Crosses mark simulations with no elevation feedback. Bars mark the estimated 95% credibility intervals and (central tick) best estimates of the elevation feedback parameterisation.

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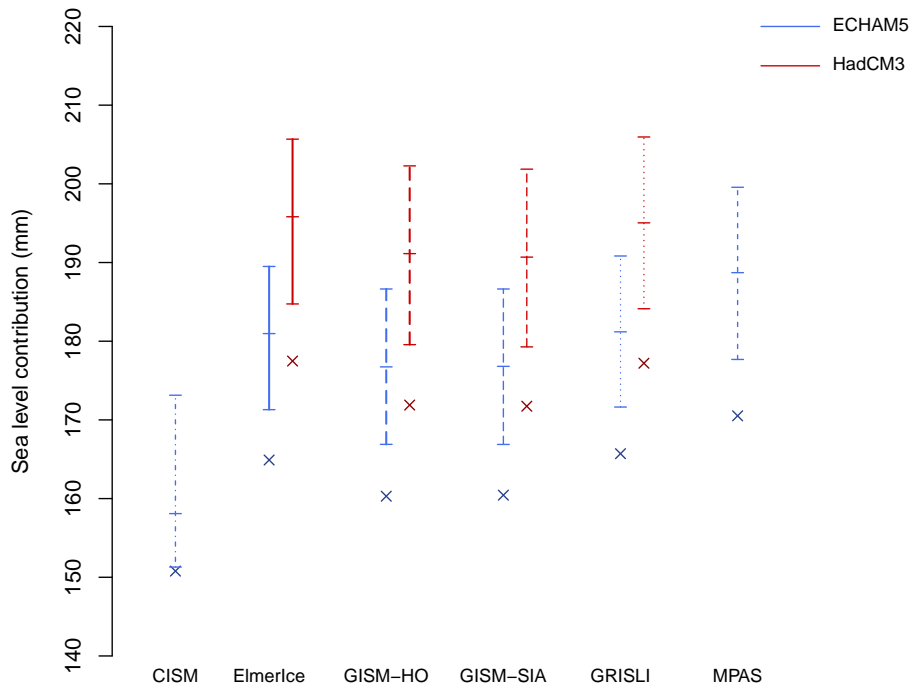


Fig. 2. As for Fig. 1, but at 2200.

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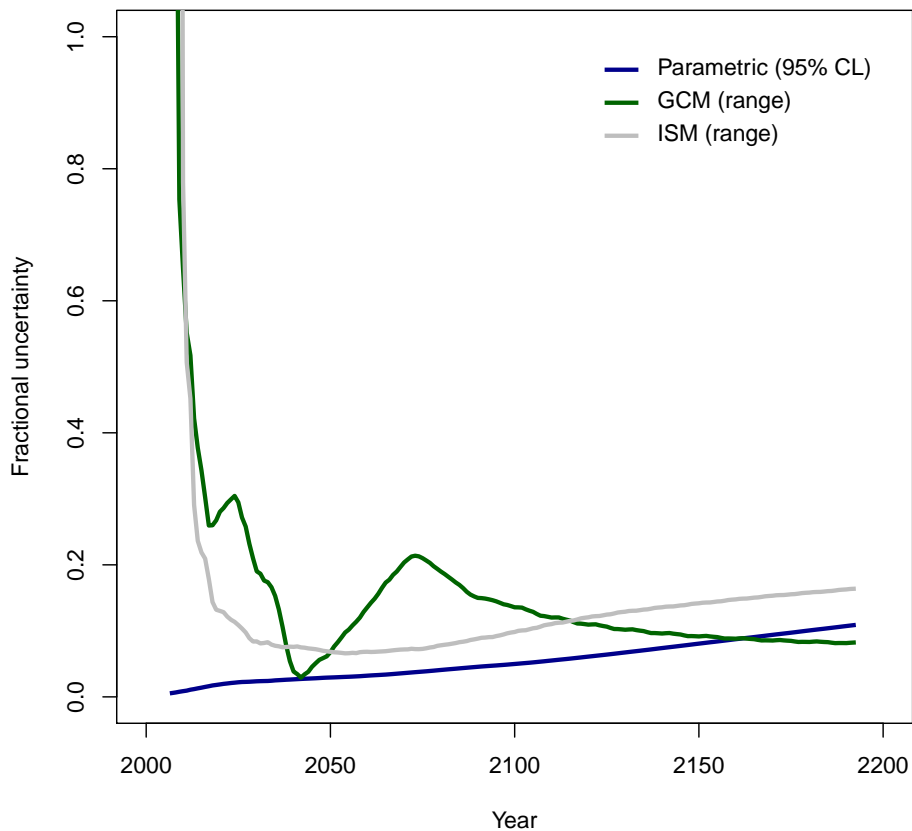


Fig. 3. Fractional uncertainties in projected cumulative GrIS sea level contributions for: elevation feedback parameterisation (95 % credibility interval divided by best estimate, averaged over ISMs and GCMs; GCMs (difference between ECHAM5 and HadCM3 divided by their mean, averaged over ISM best estimates; and ISMs (range of ISM best estimates divided by their mean, for ECHAM5 projections). A 15 yr running mean is applied.

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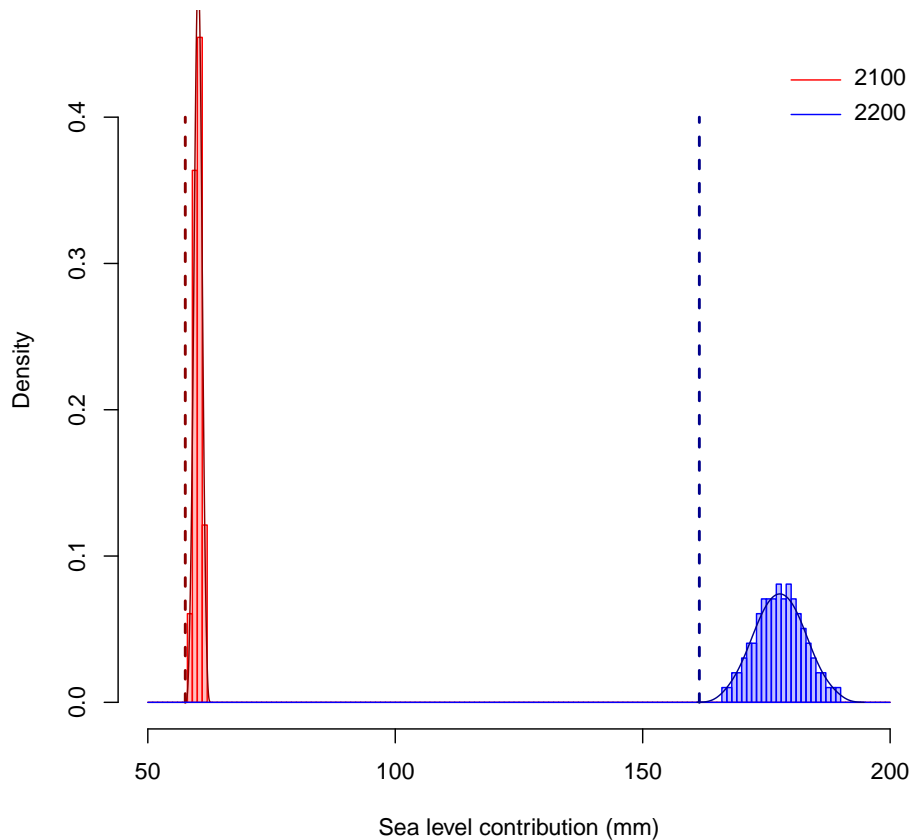


Fig. 4. GISM-SIA projections for sea level contribution in 2100 (red) and 2200 (blue): no elevation feedback (dashed vertical line) and with feedback (histograms and density estimates).

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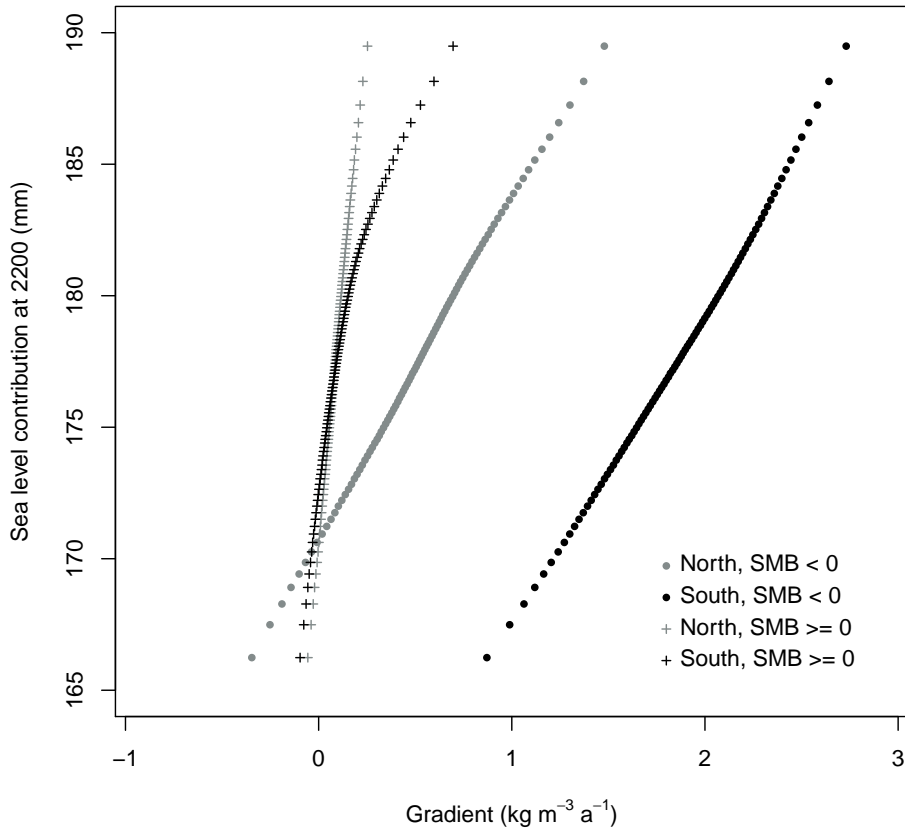


Fig. 5. GISM-SIA projections for sea level contribution in 2200 as a function of elevation feedback gradient values.

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