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Projecting Antarctic ice discharge using response functions from SeaRISE ice-sheet models

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Abstract

The largest uncertainty in projections of future sea-level change still results from the potentially changing dynamical ice discharge from Antarctica. While ice discharge can alter through a number of processes, basal ice-shelf melting induced by a warming
 ocean has been identified as a major if not the major cause for possible additional ice flow across the grounding line. Here we derive dynamic ice-sheet response functions for basal ice-shelf melting using experiments carried out within the Sea-level Response to Ice Sheet Evolution (SeaRISE) intercomparison project with five different Antarctic ice-sheet models. As used here these response functions provide separate contributions for four different Antarctic drainage regions. Under the assumptions of linear-response theory we project future ice-discharge for each model, each region and each of the four Representative Concentration Pathways (RCP) using oceanic temperatures from 19 comprehensive climate models of the Coupled Model Intercomparison Project, CMIP-5, and two ocean models from the EU-project Ice2Sea. Uncertainty in the cli-

- ¹⁵ matic forcing, the oceanic response and the ice-model differences is combined into an uncertainty range of future Antarctic ice-discharge induced from basal ice-shelf melt. The additional ice-loss (Table 6) is clearly scenario-dependent and results in a median of 0.07 m (66 %-range: 0.04–0.10 m; 90 %-range: –0.01–0.26 m) of global sea-level equivalent for the low-emission RCP-2.6 scenario and yields 0.1 m (66 %-range: 0.06–
- 0.14 m; 90 %-range: -0.01-0.45 m) for the strongest RCP-8.5. If only models with an explicit representation of ice-shelves are taken into account the scenario dependence remains and the values change to: 0.05 m (66 %-range: 0.03-0.08 m) for RCP-2.6 and 0.07 m (66 %-range: 0.04-0.11 m) for RCP-8.5. These results were obtained using a time delay between the surface warming signal and the subsurface oceanic warming as
- observed in the CMIP-5 models. Without this time delay the ranges for all ice-models changes to 0.10 m (66 %-range: 0.07–0.12 m; 90 %-range: 0.01–0.28 m) for RCP-2.6 and 0.15 m (66 %-range: 0.10–0.21 m; 90 %-range: 0.02–0.53 m) for RCP-8.5. All probability distributions as provided in Fig. 10 are highly skewed towards high values.





1 Introduction

The future evolution of global mean and regional sea-level is important for coastal planning and associated adaptation measures. The Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change (IPCC) provided sea-level projec-

- tions explicitly excluding change in dynamic ice-discharge from both Greenland and Antarctica (Solomon et al., 2007). While the part of the ice-sheet directly susceptible to ocean water on Greenland is limited, marine ice sheets in West Antarctica alone have the potential to elevate sea level globally by 3.3 m (Bamber et al., 2009). Previous projections of the Antarctic ice-sheet mass-balance have used fully coupled climate-ice-
- sheet models (e.g. Huybrechts et al., 2011; Vizcaino et al., 2010). These simulations include feedbacks between the climate and the ice sheet and thereby provide very valuable information especially on multi-centennial time scale. On shorter, i.e. decadal to centennial, time scales the direct climatic forcing is likely to dominate the ice-sheet evolution. For 21st-century projections it might thus be more favorable to apply the out-
- ¹⁵ put of comprehensive climate models as external forcing to the ice sheet, neglecting feedbacks while possibly improving on the accuracy of the forcing anomalies. Here we follow this approach.

In order to meet the relatively high standards that are set by climate models for the oceanic thermal expansion and glacier- and ice-cap models which use the full range of state-of-the-art climate projections, it would be desirable to use different ice-sheet models for a robust projection of the sea-level contribution. While changes in basal lubrications, ice-softening from surface warming and changes in surface elevation through altered precipitation can affect dynamic ice-discharge from Antarctica, changes in dynamic ice loss is likely to be dominated by changes in basal melt underneath the ice

shelves. Here we combine the dynamic response of five ice-sheet models to changes in basal melt with the full uncertainty range of future climate change for each of the Representative Concentration Pathways (RCP, Moss et al., 2010; Meinshausen et al.,





2011b) using the current simulations from the Coupled Model Intercomparison Projection, CMIP-5 (Taylor et al., 2012).

The response functions for these five different Antarctic ice sheet models are derived from a standardized melting experiment (M2) from the Sea-level Response to Ice Sheet

- Evolution (SeaRISE) intercomparison project (Bindschadler et al., 2012). This community effort gathers a broad range of structurally different ice-sheet models to perform a climate-forcing sensitivity study for both Antarctica (Nowicki et al., 2012b) and Greenland (Nowicki et al., 2012a). A suite of prescribed numerical experiments on a common set of input data represents different types of climate input, namely enhanced sub-shelf
 melting, enhanced sliding and surface temperature increase combined with enhanced net accumulation.
 - The spread in the response of the participating models to these experiments results from differences in the stress-balance approximations, the treatment of grounding line motion, the implementation of ice-shelf dynamics, the computation of the surface-mass
- balance, and in the computational demand which sets strong limits on the spin-up procedure. This approach allows for the identification of the sensitivity of the response of state-of-the-art ice-sheet models to changes in different types of climate-related boundary conditions. An interpolation analysis of the results is performed in (Bindschadler et al., 2012) in order to provide a best-guess estimate of the future sea-level
- ²⁰ contribution from the ice sheets. Here we process the SeaRISE-Antarctica results of one of the experiments (M2: uniform, constant melt-rate of 20 m a⁻¹ applied to all ice shelves) in the framework of linear response theory in order to provide projections of ocean-warming-induced ice loss from different drainage basins for the different RCP scenarios. These methods have been used before, for example to generalize climatic
- ²⁵ response to greenhouse gas emissions (Good et al., 2011).





2 Brief description of the ice-sheet models

All models used here are described in more detail by Bindschadler et al. (2012) (Table 2). Here we provide a brief summary for the purpose of this paper referring to relevant publications from which even more detailed descriptions can be obtained.

5 2.1 AIF

The Anisotropic Ice-Flow model is a 3-D ice-sheet model (without ice shelves) incorporating anisotropic ice flow and fully coupling dynamics and thermodynamics (Wang et al., 2012). This is a higher-order model with longitudinal and vertical shear stresses. The model uses finite difference method to calculate ice-sheet geometry including isostatic bedrock adjustment, 3-D distributions of shear and longitudinal strain rates, enhancement factors which account for the effect of ice anisotropy, temperatures, horizontal and vertical velocities, shear and longitudinal stresses. The basal sliding is determined by Weertman sliding law based on a cubic power relation of the basal shear stress. Ice sheet margin moves freely and the rounding line is detected by the floating condition. As the model lacks of ice shelves, the prescribed basal melt rates are

¹⁵ Ing condition. As the model lacks of ice shelves, the prescribed basal melt rates are applied to the ice-sheet perimeter grid-points only with a bed below sea level. The icesheet margin, i.e. the grounding line, moves freely within the model grid-points and the grounding line is detected by the floating condition without sub-grid interpolation.

2.2 Penn-State-3D

The Penn State 3D ice sheet model uses a hybrid combination of the scaled Shallow Ice (SIA) and Shallow Shelf (SSA) equations for shearing and longitudinal stretching flow respectively. The location of the grounding line is determined by simple flotation, with sub-grid interpolation as in (Gladstone et al., 2010). A parameterization relating ice velocity across the grounding line to local ice thickness is imposed as an internal boundary-layer condition, so that grounding-line migration is simulated reasonably well





without the need for very high, i.e. of the order of 100 m, resolution (Schoof, 2007). Ocean melting below ice shelves and ice-shelf calving use simple parameterizations, along with a sub-grid parameterization at the floating-ice edge (Pollard and DeConto, 2009; Pollard and Deconto, 2012)

5 2.3 PISM

The Parallel Ice Sheet Model (www.pism-docs.org) used here is based on version stable 0.4, which incorporates the Potsdam Parallel Ice Sheet Model (PISM-PIK) (Winkelmann et al., 2011; Martin et al., 2011). Ice flow is approximated by a hybrid scheme incorporating both the SIA and SSA approximations (Bueler and Brown, 2009). An enthalpy formulation (Aschwanden et al., 2012) is used for thermodynamics, and the model employs a physical stress-boundary condition to the shelfy-stream approximation at ice fronts, in combination with a sub-grid interpolation (Albrecht et al., 2011) and a kinematic first-order calving law (Levermann et al., 2012) at ice-shelf fronts. In PISM-PIK, the grounding line is not subject to any boundary conditions or flux corrections.

¹⁵ Its position is determined from ice and bedrock topographies in each time step via the floatation criterion. The grounding line motion is thus influenced only indirectly by the velocities through the ice thickness evolution. Since the SSA (shallow shelf approximation) velocities are computed non-locally and simultaneously for the shelf and for the sheet, a continuous solution over the grounding line without singularities is ensured and buttressing effects are accounted for.

2.4 SICOPOLIS

The SImulation COde for POLythermal Ice Sheets is a three-dimensional, polythermal ice sheet model that was originally created by Greve (1995, 1997) in a version for the Greenland ice sheet, and has been developed continuously since then (Sato

²⁵ and Greve, 2012) (sicopolis.greveweb.net). It is based on finite-difference solutions of the shallow ice approximation for grounded ice (Hutter, 1983; Morland, 1984) and the





shallow shelf approximation for floating ice (Morland, 1987; MacAyeal, 1989). Special attention is paid to basal temperate layers (that is, regions with a temperature at the pressure melting point), which are positioned by fulfilling a Stefan-type jump condition at the interface to the cold ice regions. Basal sliding is parameterized by a Weertman-

- type sliding law with sub-melt sliding (that allows for a gradual onset of sliding as the basal temperature approaches the pressure melting point, Greve, 2005), and glacial isostasy is described by the elastic lithosphere/relaxing asthenosphere (ELRA) approach (Le Meur and Huybrechts, 1996). The position and evolution of the grounding line is determined by the floating condition. Between neighbouring grounded and float-
- ¹⁰ ing grid points, the ice thickness is interpolated linearly, and the half-integer auxiliary grid point in between (on which the horizontal velocity is defined, Arakawa C grid) is considered as either grounded or floating depending on whether the interpolated thickness leads to a positive thickness above floatation or not.

2.5 UMISM

- The University of Maine Ice Sheet Model consists of a time-dependent finite-element solution of the coupled mass, momentum, and energy conservation equations using the SIA (Fastook, 1990, 1993; Fastook and Chapman, 1989; Fastook and Hughes, 1990; Fastook and Prentice, 1994) with a broad range of applications (for example Fastook et al., 2012, 2011) The 3-D temperature field, on which the flow law ice hard-
- ness depends, is obtained from a 1-D finite-element solution of the energy conservation equation at each node. This thermodynamic calculation includes vertical diffusion and advection, but neglects horizontal movement of heat. Also included is internal heat generation produced by shear with depth and sliding at the bed. Boundary conditions consist of specified surface temperature and basal geothermal gradient. If the calcu-
- ²⁵ lated basal temperature exceeds the pressure melting point, the basal boundary condition is changed to a specified temperature, and a basal melt rate is calculated from the amount of latent heat of fusion that must be absorbed to maintain this specified temperature. Conversely, if the basal temperature drops below the pressure melting





point where water is already present at the bed, a similar treatment allows for the calculation of a rate of basal freezing. A map-plane solution for conservation of water at the bed, whose source is the basal melt or freeze-on rate provided by the temperature solution, allows for movement of the basal water down the hydrostatic pressure gradi-

- ent (Johnson and Fastook, 2002). Areas of basal sliding can be specified if known, or determined internally by the model as regions where lubricating basal water is present, produced either by melting in the thermodynamic calculation or by movement of water beneath the ice sheet down the hydrostatic gradient. Ice shelves are not modeled explicitly in UMISM. However, a thinning rate at the grounding line produced by longi tudinal stresses is calculated from a parameterization of the thinning of a floating slab
- (Weertman, 1957). No sub-grid grounding line interpolation is applied.

3 Deriving the response functions

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In order to use the sensitivity experiments carried out within the SeaRISE project (Bindschadler et al., 2012), we assume that for the 21st century the temporal evolution of the ice-discharge can be expressed as

$$S(t) = \int_{0}^{t} \mathrm{d}\tau \, R\left(t - \tau\right) m\left(\tau\right) \tag{1}$$

where S is the sea-level contribution from ice discharge, m is the forcing represented by the basal-melt rate and R is the ice-sheet response-function. t is time starting from a period prior to the beginning of a significant forcing. The responses function R can thus be understood as the response to a delta-peak forcing with magnitude one.

$$S_{\delta}(t) = \int_{0}^{t} \mathrm{d}\tau \, R \left(t - \tau \right) \delta \left(\tau \right) = R \left(t \right)$$





Please note that we express ice-discharge throughout the paper in units of global mean sea-level equivalent. That means that in deriving the response functions we only diagnose ice loss above flotation that is relevant for sea level. As a simple consequence the response function is unitless. Furthermore the basal-melt signal applied as well as the ice-discharge signal used to derive the response functions are anomalies with respect to a baseline simulation under present-day boundary conditions (Bindschadler et al.,

2012).
Linear response theory, as represented by Eq. (1), can only describe the response of a system up to a certain point in time; 100 yr is a relatively short period for the response
of an ice-sheet and the assumption of a linear response is thereby justified. There are a number of ways to obtain the system-specific response function *R* (e.g. Winkelmann and Levermann, 2012). Within the SeaRISE project the switch-on basal-melt experiments can be used conveniently since their response directly provides the time integral of the response function for each individual ice-sheet model. Assuming that over a forcing period of 100 yr the different topographic basins on Antarctica from which ice

¹⁵ lorcing period of 100 yr the different topographic basins on Antarctica from which ice is discharged, respond independently of each other, we diagnose the additional ice-flow from four basins separately (Fig. 1) and interpret them as the time integral of the response function for each separate basin. The response function for each basin is shown in Fig. 2. The aim of this study is specifically to capture differences between
 ²⁰ individual ice-sheet models. The fact that the models differ is nicely illustrated by their different response functions.

To obtain *R* we use the response to the temporal stepwise increase in basal melt by 20 m a^{-1} (denoted M2-experiment in Bindschadler et al., 2012). The ice-sheet response to a step forcing is equivalent to the temporal integral of the response function *R* with *t* = 0 being the time of the switch-on in forcing

$$S_{\rm sf}(t) = \int_{0}^{t} \mathrm{d}\tau \, R \left(t - \tau \right) \Delta m_0 \cdot \Theta \left(\tau \right) = \Delta m_0 \cdot \int_{0}^{t} \mathrm{d}\tau \, R \left(\tau \right)$$

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Discussion Paper TCD 6, 3447-3489, 2012 Antarctic ice discharge from SeaRISE response **Discussion** Paper functions A. Levermann et al. **Title Page** Introduction Abstract **Discussion** Paper Conclusions References **Tables Figures** Back Close **Discussion** Paper Full Screen / Esc Printer-friendly Version Interactive Discussion



where $\Theta(\tau)$ is the Heavyside function which is zero for negative τ and one otherwise. We thus obtain the response function from

$$R(t) = \frac{1}{\Delta m_0} \cdot \frac{\mathrm{d}S_{\mathrm{sf}}}{\mathrm{d}t}(t)$$

Only the M2-experiment will be used because the anomalies in basal-melt rates as obtained from the CMIP-5 simulations (compare Fig. 4 and the range of basal-melt sensitivities (7–16 m a⁻¹ K⁻¹) are limited by 20 m a⁻¹. Since a linear relation between response and forcing is assumed (Eq. 1) the forcing from which the response functions are derived should be similar to the forcing applied in the projections. Basal melt rates of the M1- (2 m a⁻¹) and M3- (200 m a⁻¹) experiments are either too low or too high and consequently yield slightly different results.

The spatial distribution of the ice loss after 100 yr of additional basal-ice shelf melting illustrates the different dynamics of the ice-sheet models resulting from different representations of ice dynamics, surface mass balance and basal sliding parameterizations (Fig. 3). Part of the individual responses result from the different representations of the basal ice-shelf melt. In the UMISM model basal melt was applied along the entire coastline which yields a particularly strong response in East Antarctica (Fig. 2). This is likely an overestimation of the ice-loss compared to models with a more realistic representation of ice shelves. On the other hand, coarse resolution ice-sheet models as used here cannot capture small ice shelves as they are present especially around

East Antarctica. These models thus have a tendency to underestimate the fraction of the coastal ice that is afloat and thus sensitive to changes in ocean temperature might be also underestimated (compare for example (Martin et al., 2011) for the PISM model). While we will also provide projections using only the three models with explicit representation of ice shelves (PennState-3D, PISM and SICOPOLIS), it is therefore worthwhile to consider the full spectrum of ice-sheet models in order to capture the full uncertainty range.

Discussion Paper TCD 6, 3447-3489, 2012 Antarctic ice discharge from SeaRISE response **Discussion** Paper functions A. Levermann et al. **Title Page** Abstract Introduction Discussion Paper Conclusions References **Figures** Tables Back Close **Discussion** Paper Full Screen / Esc Printer-friendly Version Interactive Discussion

(2)



4 Assembling the forcing

4.1 Global mean temperature range

In order to relate regional changes in ice-discharge to the global climate signal, we use the Representative Concentration Pathways (RCP) (Moss et al., 2010; Meinshausen et al., 2011b). The range of possible changes in global mean temperature that result from each RCP is obtained by constraining the response of the emulator model MAGICC 6.0 (Meinshausen et al., 2011a) with the observed temperature record. This procedure has been used in several studies and aims to cover the possible global climate response to specific greenhouse-gas emission pathways (e.g. Meinshausen et al., 2009). Here we use a set of 600 time series of global mean temperature from the year 1850 to 2100 for each RCP that cover the full range of future global temperature changes as detailed in (Schewe et al., 2011).

4.2 Subsurface oceanic temperatures from CMIP-5

In order to be applicable to the ice-sheet response-functions which translate changes in
 basal-melt rate into changes in ice discharge, the global mean temperature scenarios need to be downscaled to sub-shelf basal-melt rates. To this end we use the simulations of the recent Coupled Model Intercomparison (CMIP-5) to obtain a scaling relationship between the global mean temperature and the oceanic subsurface temperature for each model. This has been carried out for the CMIP-3 experiments by Winkelmann
 et al. (2012) and is repeated here for the more recent climate models of CMIP-5. For comparison, Yin et al. (2011) assessed output from 19 AOGCMs under scenario A1B to determine how subsurface temperatures are projected to evolve around the ice sheets. They showed decadal-mean warming of 0.4–0.7 °C and 0.4–0.9 °C around Antarctica (25th to 75th percentiles of ensemble, West and East respectively) between 1951–2000 and 2091–2100.





Please note that the underlying assumption of the scaling is not that the oceanic temperatures at different places in the Southern Ocean scale with the global mean temperature, but that the anomalies of the ocean temperatures that result from global warming scale with the respective anomalies in global mean temperature. This assumption is consistent with the linear-response assumption underlying Eq. (1). The oceanic temperatures used here are taken in the subsurface at the mean depth of the ice-shelf underside in each sector (Table 1), in order for them to be close to the entrance of the ice-shelf cavities.

Due to the fact that the surface warming signal needs to be transported to depth, the best linear regression is found with a time delay between global mean surface air temperature and subsurface oceanic temperatures. The scaling coefficients as well as the time delay are detailed in Tables 2–5. The r^2 values are high which supports the application of the linear regression except for the IPSL model where the slope between the two temperature signals is also very low. We explicitly keep this model in order to include the possibility that almost no warming occurs underneath the ice-shelves.

- Figure 4 show the median and the 30 % quantiles (denoted the likely range by the IPCC, Solomon et al., 2007) for the oceanic subsurface temperatures as obtained from a random selection of global mean temperature pathways combined with a randomly selected scaling coefficient and the associated time delay Δt from Tables 2–5. Note
- that there are physical reasons to assume a time delay between the surface and the subsurface temperatures, however, the correlation is already high without assuming such a time delay which is why we will also provide projections without time delay in Sect. 5. The oceanic temperature time-series without time delay are provided as inlays in Fig. 4 for comparison.
- ²⁵ Ocean temperature anomalies need then to be transformed into basal melt anomalies. If the temperature change would be transported undiluted into the cavity and through the turbulent mixed layer underneath the ice shelf the simple formula

$$m = \frac{\rho_{\rm O} c_{\rho_{\rm O}} \gamma_T}{\rho_i L_i} \cdot \delta T_{\rm O} \approx 42 \frac{m}{aK} \cdot \delta T_{\rm O}$$

CC ①

(3)

where $\rho_0 = 1028 \text{ kg m}^{-3}$ and $c_{\rho_0} = 3974 \text{ J kg}^{-1} \text{ K}^{-1}$ are density and heat capacity of ocean water. $\rho_i = 910 \text{ kg m}^{-3}$ and $L_i = 3.35 \times 10^5 \text{ J kg}^{-1}$ are ice density and latent heat of ice melt, $\gamma_T = 10^{-4}$ as adopted from Hellmer and Olbers (1989), would be applicable. This formula however, does not account for the complex oceanic circulation underneath

⁵ the ice shelf and the transport processes through the mixed layer as well as the properties of the mixed layer itself. Observations suggest an interval of $7 \text{ m a}^{-1} \text{ K}^{-1}$ (Jenkins, 1991) to 16 m a⁻¹ K⁻¹ (Payne et al., 2007). See Holland et al. (2008) for a detailed discussion and comparison to other observations. The coefficient used for each projection will be drawn uniformly from this interval.

10 4.3 Application of ice-sheet response functions to projections from regional ocean models

In a first step the ice-sheet response functions of the Weddel- and Ross-Sea sectors are applied to the output from the high-resolution global finite-element FESOM and the regional ocean model BRIOS.

15 4.3.1 BRIOS

The Bremerhaven Regional Ice Ocean Simulations is a coupled ice-ocean model which resolves the Southern Ocean south of 50° S zonally at 1.5° and meridionally at 1.5° × cosφ. The water column is variably divided into 24 terrain-following layers. The sea-ice component is a dynamic-thermodynamic snow/ice model with heat bud²⁰ gets for the upper and lower surface layers (Parkinson and Washington, 1979) and a viscous-plastic rheology (Hibler III, 1979). BRIOS considers the ocean-ice shelf interaction underneath ten Antarctic ice shelves (Beckmann et al., 1999; Hellmer, 2004) with time-invariant thicknesses, assuming flux divergence and mass balance to be in dynamical equilibrium. The model has been successfully validated by the comparison
²⁵ with mooring and buoy observations regarding, e.g., Weddell gyre transport (Beckmann et al., 1999), sea ice thickness distribution and drift in Weddell and Amundsen





seas (Timmermann et al., 2002a; Assmann et al., 2005) and sub-ice-shelf circulation (Timmermann et al., 2002b).

4.3.2 FESOM

The Finite-Element Southern Ocean Model is a hydrostatic, primitive-equation ocean model with an unstructured grid that consists of triangles at the surface and tetrahedra in the ocean interior. It is based on the Finite Element model of the North Atlantic (Danilov et al., 2004, 2005) coupled to a dynamic-thermodynamic sea-ice model with a viscous-plastic rheology and evaluated in a global setup (Timmermann et al., 2009; Sidorenko et al., 2011). An ice-shelf component with a three-equation system for the computation of temperature and salinity in the boundary layer between ice and ocean and the melt rate at the ice shelf base (Hellmer et al., 1998) has been implemented. Turbulent fluxes of heat and salt are computed with coefficients depending on the friction velocity following Holland and Jenkins (1999). The present setup uses a hybrid vertical coordinate and a global mesh with a horizontal resolution between 30 and 40 km in the offshore Southern Ocean, which is refined to 10 km along the Antarctic

coast, 7 km under the larger ice shelves in the Ross and Weddell Seas, and to 4 km under the small ice shelves in the Amundsen Sea.

Outside the Southern Ocean, resolution decreases to 50 km along the coasts and about 250–300 km in the vast basins of the Atlantic and Pacific Oceans, while on the other hand some of the narrow straits that are important to the global thermohaline

other hand some of the narrow straits that are important to the global thermohaline circulation (e.g., Fram- and Denmark Straits, and the region between Iceland and Scotland) are represented with high resolution (Timmermann et al., 2012). Ice shelf draft, cavity geometry, and global ocean bathymetry have been derived from the RTopo-1 dataset (Timmermann et al., 2010) and thus consider data from many of the most recent surveys of the Antarctic continental shelf.

Regional climate-change scenarios available from simulations with BRIOS and FE-SOM have been presented by Hellmer et al. (2012) and Timmermann and Hellmer (2012). We utilize data from the SRES A1B scenario, which represents greenhouse





gas forcing between the RCP-6.0 and RCP-8.5 and the E1-scenario of the IPCC-AR4, Solomon et al., 2007), which is comparable to RCP-2.6. Both models were forced with boundary conditions obtained from two global climate models: ECHAM-5 (full lines in Fig. 5) and HadCM-3 (dashed lines in Fig. 5). Note that negative sea-level contributions
 ⁵ as obtained for the Ross-Sea sector with ECHAM-5 forcing corresponds to declining temperatures and thereby declining basal melt rates. Since such declining melt rates or even refreezing corresponds to a different physical process it is unlikely that the linear response functions from the SeaRISE experiments are applicable in such a case.

As has already been discussed by Timmermann and Hellmer (2012), Fig. 5 shows that the role of the global climate model in projecting ice discharge is clearly the dominating uncertainty. In order to cover the relevant uncertainty range it is thus favorable to use the broadest possible spectrum of climatic forcing. In Sect. 5 we attempt to capture the climatic and oceanic uncertainty along with the ice-dynamical one with a number of caveats remaining.

15 5 Projecting Antarctic sea-level contribution

5.1 Probabilistic approach

In order to capture the climate uncertainty as well as the uncertainty in the oceanic response and the ice-sheet response, we follow a probabilistic approach. For each scenario, a climate forcing that is consistent with the observed climate change and the range of climate sensitivity of 2–4.5 degrees for a doubling of CO2 is randomly selected (i.e. picked from the 600 global-mean-temperature time series as described above). This evolution of global mean temperature is then translated into a time series of subsurface ocean temperature change by random selection of a set of scaling coefficients and the associated time delay that correspond to one of the 19 CMIP-5 model (Tables 2–5). This time series is then used to compute the ice loss from the associated ice-sheet basin. The procedure is repeated 50 000 times. The 66 %-range





(i.e. the central 66 % of the ice-discharge curves) is provided as the uncertainty range. In addition to this "likely" range as defined by the IPCC also the "very likely" range of the 90 %-range is provided in Figs. 7 and 10 as well as in Table 6.

5.2 Results for the different basins and different models

- ⁵ Figure 6 shows the uncertainty range of the projected contribution from the different oceanic sectors comprising uncertainty in climate and ocean circulation. While the individual time series will differ from the projections with the ocean models, FESOM and BRIOS, the order of magnitude of the range of the sea-level contribution covered is the same. FESOM and BRIOS yield a particularly strong response when forced with the HadCM3 model (dashed lines in Fig. 5) and a weaker response when forced with ECHAM-5. The response of the models from the downscaled global simulations covers this range. While the largest uncertainty is observed in the Amundsen-Sea sector which forces the Pine-Island-Thwaites glaciers, the contributions of all sectors are relatively similar with a scatter of the median from 0.01 to 0.03 m (Fig. 7). Note, however, that
- the contributions from the different regions are not independent and thus the median of the full ensemble cannot necessarily be obtained as the sum of the individual medians of the basins. The histogram of the ice-discharge contribution for the year 2100 further also shows the strongly skewed probability distribution. The long tail towards higher sea-level contributions makes the estimate of the 90 %-range of the distribution of the information of the information of the information of the information of the information.
- ²⁰ (thin horizontal lines at the top of each panel) very difficult, because it is based on few extreme combinations which might not be robust.

The total ice discharge varies strongly between the different ice-sheet models (Fig. 8) as can be expected from the differences in the response functions of Fig. 2. The weakest ice-loss is projected from the PISM model while the strongest signal is obtained

²⁵ from PennState-3D. Together with SICOPOLIS the three models with explicit representation of ice shelves span the full range of responses. The two models without explicit ice-shelf dynamics, AIF and UMISM, compute medium ranges. While there is a clear dependence on the climatic scenario, the uncertainty between different





ice-sheet models is comparable to the scenario spread. The strongest difference between models with and without explicit ice-shelf representation is observed in East Antarctic (dashed line in Fig. 6). The difference results mainly from the strong contribution from the UMISM model which assumes basal melt along the entire coastline (compare Fig. 3). While this assumption might be an overestimate, it can serve as an upper limit to the real response.

5.3 Scenario dependence

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The full uncertainty range including climate-, ocean- and ice-sheet-model spread shows large uncertainty increasing with time along the 21st-century projections (Fig. 9).
While model uncertainty is large, there is a clear scenario dependence in both the median, the 66% and the 90% percentile of the distribution (Table 6). This scenario-dependence is independent of the selection of the ice-sheet models or whether the time lag is included or the global mean temperature anomaly is scaled down to the oceanic subsurface without time delay (Fig. 10 and Table 6). All distributions are strongly
skewed towards high sea-level contributions. This skewness strongly influences the median of the distributions as well as the 66%- and 90%-ranges. Consequently the median is not the value with the highest probability. The large tails makes an estimate of the 90%-range, i.e. the "very likely" range as denoted by the IPCC, very uncertain.

6 Conclusion and discussions

The aim of this study is to estimate the full range of possible future ice-discharge from Antarctica that can be induced by ocean warming within the 21st century. To this end we include climatic models that yield practically no warming of the Southern Ocean subsurface (e.g. IPSL) but also take into account the relatively strong response of the UMISM ice-sheet model in East Antarctica which might overestimate the situation there.





The uncertainty ranges comprising climatic, oceanic and ice-dynamical uncertainty show a clear dependence on the global climate-change scenario (Table 6). For the RCP-2.6 which was designed to result in a median increase in global mean temperature below 2 K in most climate models, the 66%-range of ice-loss is equivalent to

5 0.04–0.10 m with a median of 0.07 m when all ice-sheet models are considered. This range increases to 0.06–0.14 m for the RCP-8.5 with a median contribution of 0.1 m. Excluding the two ice-sheet models without explicit representation of ice-shelf dynamics reduces the values to 0.03–0.08 m for RCP-2.6 and 0.04–0.11 m for RCP-8.5.

As discussed above, a time lag between the oceanic temperature change and the change in global mean temperature is physically reasonable and applied in these projections. However, the correlation between surface warming and subsurface temperature change improves only marginally when introducing the time lag and it is not clear whether small scale processes may accelerate the heat transport at finer resolution (Hellmer et al., 2012). It is thus worthwhile to consider the ice loss without a time lag (Fig. 9b). If the basal melt rates are applied immediately the 66 %-range of the sea level contribution increases from 0.06–0.14 m to 0.10–0.21 m for RCP-8.5.

The simulations with the high-resolution finite-element ocean model FESOM and the regional ocean model BRIOS (Fig. 5) illustrates that abrupt ocean circulation changes can have strong influence on the basal melt rates (Hellmer et al., 2012). The compara-

²⁰ bly coarse-resolution ocean components even of the CMIP-5 global climate models are unlikely to resolve such small scale changes. Estimates on their forcing as presented here will thus be dominated by basin-scale temperature changes of the interior ocean.

The computation of the basal-melt anomalies from the temperature anomalies is simplified in comparison to the real situation in which changes in ice thickness and salinity

²⁵ need to be accounted for. Due to the lack of ice-shelf models in the coupled global climate models the salinity and the ice-thickness cannot be reliably projected within the probabilistic approach taken here. The interval of 7–16 m a⁻¹ K⁻¹ applied to the temperature anomalies (not absolute temperatures) is the best estimate without taking salinity and ice-topographic changes into account that can be derived from observations.





The probabilistic approach applied here assumes a certain interdependence of the different uncertainties. The global climatic signal is selected independently from the oceanic scaling coefficient. However the range of scaling coefficients is derived from the correlation within the different CMIP-5 models. The ice-sheet uncertainty is again independent of the other two components.

It is important to note that all resulting probability distributions are strongly skewed towards higher sea-level contributions. This strongly influences the median as well as the 66%-range and is responsible for some of the scenario-dependence. While these values can be considered robust, the estimate of the "very likely" 90 %-range is rather uncertain and cannot be used with confidence. The 90%-range which is denoted by the IPCC as the "very likely"-range reaches up to 0.45 m for all models including the time-delay and even up to 0.53 m without time delay.

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ConclusionsReferencesTablesFiguresI▶II▶IBackCloseFull Screen / EscPrinter-friendly VersionInteractive DiscussionEsc

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6, 3447-3489, 2012

Antarctic ice

discharge from

SeaRISE response

functions

A. Levermann et al.

Title Page

Abstract

Introduction

Discussion Paper

Discussion Paper

Discussion Paper

Discussion Paper

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TCD

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Discussion Paper

Discussion

Paper

Discussion Paper

Discussion Paper





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Table 1. Mean depth of ice shelves in the different regions denoted in Fig. 1 as computed from (Le Brocq et al., 2010). Oceanic temperature anomalies were averaged vertically over a 100 m range around these depth.

Region	Depth [m]
Amundsen Sea	305
Ross Sea	312
Weddell Sea	420
East Antarctica	369





Model	Coeff	r^2	Λt	Coeff	r ²
Model	without Λt	,	[vr]	with At	,
	Without Δt		ניען		
ACCESS1-0	0.17	0.86	0	0.17	0.86
ACCESS1-3	0.30	0.94	0	0.30	0.94
BNU-ESM	0.37	0.88	30	0.56	0.92
CanESM2	0.15	0.83	30	0.24	0.88
CCSM4	0.22	0.89	0	0.22	0.89
CESM1-BGC	0.19	0.92	0	0.19	0.92
CESM1-CAM5	0.12	0.92	0	0.12	0.92
CSIRO-Mk3-6-0	0.16	0.79	30	0.28	0.83
FGOALS-s2	0.24	0.90	55	0.54	0.93
GFDL-CM3	0.26	0.81	35	0.49	0.85
HadGEM2-ES	0.23	0.70	0	0.23	0.70
INMCM4	0.67	0.90	0	0.67	0.90
IPSL-CM5A-MR	0.07	0.22	90	0.44	0.45
MIROC-ESM-CHEM	0.12	0.74	5	0.13	0.75
MIROC-ESM	0.11	0.55	60	0.35	0.61
MPI-ESM-LR	0.27	0.80	5	0.29	0.82
MRI-CGCM3	0.00	0.02	85	-0.07	0.04
NorESM1-M	0.30	0.94	0	0.30	0.94
NorESM1-ME	0.31	0.89	0	0.31	0.89

Table 2. Amundsen-Sea sector: scaling coefficients and time delay Δt between increase in global mean temperature and subsurface ocean temperature anomaly.



Model	Coeff. without Δ <i>t</i>	r ²	Δ <i>t</i> [yr]	Coeff. with Δt	r ²
ACCESS1-0	0.07	0.73	35	0.14	0.80
ACCESS1-3	0.07	0.73	35	0.15	0.81
BNU-ESM	0.37	0.89	0	0.37	0.89
CanESM2	0.11	0.82	55	0.31	0.91
CCSM4	0.37	0.95	20	0.49	0.96
CESM1-BGC	0.37	0.95	25	0.53	0.96
CESM1-CAM5	0.23	0.79	50	0.63	0.88
CSIRO-Mk3-6-0	0.19	0.80	55	0.60	0.90
FGOALS-s2	0.09	0.73	85	0.39	0.86
GFDL-CM3	0.11	0.55	60	0.31	0.62
HadGEM2-ES	0.31	0.92	0	0.31	0.92
INMCM4	0.26	0.83	10	0.30	0.83
IPSL-CM5A-MR	-0.02	0.00	85	-0.06	0.03
MIROC-ESM-CHEM	0.07	0.50	65	0.32	0.77
MIROC-ESM	0.03	0.27	65	0.18	0.59
MPI-ESM-LR	0.08	0.65	85	0.41	0.70
MRI-CGCM3	0.21	0.63	40	0.47	0.83
NorESM1-M	0.26	0.90	5	0.28	0.92
NorESM1-ME	0.25	0.85	50	0.64	0.92

Table 3. Weddell sector: scaling coefficients and time delay Δt between increase in global mean temperature and subsurface ocean temperature anomaly.



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Model	Coeff. without Δ <i>t</i>	r ²	Δ <i>t</i> [yr]	Coeff. with Δ <i>t</i>	r^2
ACCESS1-0	0.18	0.77	20	0.26	0.79
ACCESS1-3	0.09	0.76	15	0.12	0.77
BNU-ESM	0.28	0.83	20	0.36	0.84
CanESM2	0.14	0.74	45	0.32	0.80
CCSM4	0.14	0.91	5	0.15	0.92
CESM1-BGC	0.14	0.90	0	0.14	0.90
CESM1-CAM5	0.16	0.85	0	0.16	0.85
CSIRO-Mk3-6-0	-0.06	0.28	0	-0.06	0.28
FGOALS-s2	0.18	0.89	60	0.45	0.93
GFDL-CM3	0.23	0.85	25	0.37	0.89
HadGEM2-ES	0.25	0.62	0	0.25	0.62
INMCM4	0.59	0.83	0	0.59	0.83
IPSL-CM5A-MR	0.02	0.04	95	0.14	0.12
MIROC-ESM-CHEM	0.23	0.85	0	0.23	0.85
MIROC-ESM	0.23	0.78	0	0.23	0.78
MPI-ESM-LR	0.16	0.70	40	0.31	0.73
MRI-CGCM3	0.08	0.04	0	0.08	0.04
NorESM1-M	0.12	0.79	0	0.12	0.79
NorESM1-ME	0.12	0.68	20	0.16	0.73

Table 4. Ross-Sea sector: scaling coefficients and time delay Δt between increase in global mean temperature and subsurface ocean temperature anomaly.



Discussion Paper

Discussion Paper

Discussion Paper

Model	Coeff.	r^2	Δt	Coeff.	r^2
	without Δt		[yr]	with Δt	
ACCESS1-0	0.20	0.92	30	0.35	0.94
ACCESS1-3	0.27	0.92	0	0.27	0.92
BNU-ESM	0.35	0.92	0	0.35	0.92
CanESM2	0.21	0.96	0	0.21	0.96
CCSM4	0.13	0.96	5	0.13	0.97
CESM1-BGC	0.12	0.94	25	0.17	0.95
CESM1-CAM5	0.15	0.94	0	0.15	0.94
CSIRO-Mk3-6-0	0.22	0.93	15	0.28	0.94
FGOALS-s2	0.17	0.90	55	0.41	0.94
GFDL-CM3	0.21	0.89	35	0.39	0.93
HadGEM2-ES	0.23	0.95	0	0.23	0.95
INMCM4	0.55	0.97	0	0.55	0.97
IPSL-CM5A-MR	0.14	0.89	0	0.14	0.89
MIROC-ESM-CHEM	0.11	0.89	0	0.11	0.89
MIROC-ESM	0.09	0.85	50	0.24	0.88
MPI-ESM-LR	0.20	0.94	15	0.26	0.95
MRI-CGCM3	0.26	0.94	0	0.26	0.94
NorESM1-M	0.15	0.76	0	0.15	0.76
NorESM1-ME	0.15	0.74	60	0.49	0.85

Table 5. East-Antarctic-Sea sector: scaling coefficients and time delay Δt between increase in global mean temperature and subsurface ocean temperature anomaly.



Table 6. Projections of ice-discharge in 2100 according to Fig. 10. Numbers are in meters sea-level equivalent for the different global climate RCP-scenarios with and without time delay Δt . Models with explicit representation of ice-shelf dynamics are PennState-3D, PISM and SICOPOLIS.

Set-up	RCP	Median	33%	66 %	5%	95%
All models	2.6	0.07	0.04	0.10	-0.01	0.26
with Δt	4.5	0.08	0.05	0.11	-0.01	0.32
	6.0	0.08	0.05	0.11	-0.01	0.33
	8.5	0.1	0.06	0.14	-0.01	0.45
Only models with	2.6	0.05	0.03	0.08	-0.01	0.20
explicit ice-shelf	4.5	0.06	0.03	0.09	-0.01	0.23
representation	6.0	0.06	0.03	0.09	-0.01	0.24
with Δt	8.5	0.07	0.04	0.11	-0.02	0.31
All models	2.6	0.10	0.07	0.12	0.01	0.28
without Δ <i>t</i>	4.5	0.12	0.08	0.15	0.02	0.35
	6.0	0.12	0.08	0.16	0.02	0.37
	8.5	0.15	0.10	0.21	0.02	0.53
Only models with	2.6	0.08	0.05	0.10	0.01	0.22
explicit ice-shelf	4.5	0.09	0.06	0.12	0.01	0.27
representation	6.0	0.09	0.06	0.12	0.01	0.28
without Δt	8.5	0.11	0.08	0.15	0.02	0.38







Fig. 1. The four different basins for which ice-sheet response functions are derived from the SeaRISE M2-experiments. Green lines enclose the oceanic regions over which the subsurface oceanic temperatures were averaged. Vertical averaging was carried out over a 100 m depth range centered at the mean depth of the ice-shelves in the region taken from Le Brocq et al. (2010) as provided in Table 1.

Discussion Paper TCD 6, 3447-3489, 2012 Antarctic ice discharge from SeaRISE response **Discussion** Paper functions A. Levermann et al. Title Page Abstract Introduction **Discussion** Paper Conclusions References **Tables** Figures **|**◀ Back Close **Discussion** Paper Full Screen / Esc Printer-friendly Version Interactive Discussion









Fig. 2. Linear response functions for the five ice-sheet models of Antarctica for each region as defined by Eq. (2) and as obtained from the SeaRISE-M2-experiments. The projections up to the year 2100, as computed here, will be dominated by the response functions up to year 100 since this is the period of the dominant forcing. For completeness the inlay shows the response function for the full 500 yr, i.e. the period of the original SeaRISE experiments. As can be seen from Eq. (1), the response function is dimensionless.



Fig. 3. Ice-thickness change after 100 yr under the SeaRISE experiment with homogeneous increase in basal ice-shelf melting of 20 m a^{-1} (experiment M2 and Fig. 6 in Nowicki et al., 2012b). Due to their coarse resolution some models with explicit representation of ice shelves such as the PISM model tend to underestimate the length of the coastline to which an ice shelf is attached which might lead to an underestimation of the ice loss. The UMISM model assumes basal melting along the entire coastline which is likely to result an overestimation of the effect. Black contours represent the initial grounding line which moved to the green contour during the M2-experiment after 100 yr. Lines within the continent show the drainage basins as in Fig. 1.







Fig. 4. Oceanic subsurface-temperature anomalies as obtained from scaling the range of global mean temperature changes under the different RCP scenarios to the oceanic subsurface outside the ice-shelf cavities. For the down-scaling the oceanic temperatures were diagnosed off-shore of the ice-shelf cavities within the four regions defined in Fig. 1 at the depth of the mean ice-shelf thickness as defined in Table 1. These temperature anomalies were plotted against the global mean temperature increase for each of the 19 CMIP-5 climate models used here. The best scaling was obtained when using a time delay between global mean temperature and oceanic subsurface temperature anomalies. The scaling coefficients with the respective time delay are provided in Tables 2–5. The line corresponds to the median temperature evolution. The shading corresponds to the 30 % percentile. Inlays show the temperature anomalies without time delay.







Fig. 5. Ice loss as obtained from forcing the five response functions (Fig. 2) with the basal melt computed with the high-resolution global finite-element model FESOM (FES) and the regional ocean model BRIOS (BRIO). The full lines represent simulations in which BRIOS and FESOM were forced with the global climate model ECHAM-5; dashed lines correspond to a forcing with the HadCM-3 global climate model. Results are shown for the strong climate-change scenario A1B and the relatively low-emission scenario E1. A medium basal melt sensitivity of 11.5 m a⁻¹ K⁻¹ was applied. The results illustrate the important role of the global climatic forcing.







Fig. 6. Uncertainty range of contributions to global sea level from basal-melt induced ice discharge from Antarctica for the different basins as obtained from the procedure described in the text. Results shown here include all five ice-sheet models with the global climate forcing applied with a time delay as given in Tables 2–5. The full red curve is the median enclosed by the shaded 66 %-range of the distrubtion. The full distribution is given in Fig. 7. The strongest difference between models with and without explicit representation of ice-shelves can be seen in East Antarctica. The dashed black line envelopes the 66 %-range of the models with ice-shelf representation (the full black line is the median).







Fig. 7. Probability density function for the sea-level contribution from basal-melt-induced ice discharge for each region for the year 2100. Different colors represent the four RCP scenarios. Thick horizontal lines at the top of each panel provide the 66%-range of the distribution, the black dot is the median and the thin line the rather uncertain estimate of the 90%-range. Median contributions from each sector are relatively similar. A clear scenario-dependence of the ice-discharge can be observed even if the model-uncertainty is still larger than the scenario uncertainty. The distributions are highly skewed towards higher sea-level contributions.







Fig. 8. Uncertainty range of contributions to global sea level from basal-melt induced ice discharge from Antarctica for the different ice-sheet models. Lines, shading and color coding as in Fig. 6.







Fig. 9. Uncertainty range of contributions to global sea level from basal-melt induced ice discharge from Antarctica including climate-, ocean- and ice-model uncertainty. Lines, shading and color coding as in Fig. 6. While here all ice-sheet models have been used, distributions for the year 2100 omitting models without explicit representation of ice shelves are presented in Fig. 10. The upper panel provides estimates with the time delay between global mean surface air temperature and subsurface ocean temperature (Tables 2–5).







Fig. 10. Uncertainty range including climate, ocean and ice-sheet uncertainty. Different colors represent different set-ups for the total sea-level contribution from basal-melt induced ice-discharge for the year 2100. Different panels provide estimates for the different RCP scenarios. In each panel red are the curves using only the three models with explicit representation of ice-shelves (PennState-3D, PISM, SICOPOLIS). Blue are curves using all models. Dark colors represent simulation using the time delay of Tables 2–5. Light colored lines give distributions with this time lag omitted. All distributions are highly skewed towards high sea-level contributions which strongly influences the median, the 66 %-range (thick horizontal line at the top of the panel) and the 90 %-range (thin horizontal line at the top of the panel). The scenario-dependence of each of these estimates is clearly visible in the number provided in Table 6.



