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Delaying future sea-level rise by storing water on Antarctica

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Abstract

Even if greenhouse gas emissions were stopped today sea level would continue to rise for centuries with the long-term sea-level commitment of a two-degree-warmer world significantly exceeding 2 m. In view of the potential implications for coastal populations and ecosystems worldwide we investigate, from an ice-dynamic perspective, the possibility to delay sea-level rise by pumping ocean water onto the surface of the Antarctic Ice Sheet. We find that due to wave propagation ice is discharged much faster back into the ocean than would be expected from a pure advection with surface velocities. The delay time depends strongly on the distance from the coastline at which the additional mass is placed and less strongly on the rate of sea-level rise that is mitigated. A millennium-scale storage of at least 80 % of the additional ice requires placing it at a distance of at least 700 km from the coast line. The pumping energy required to elevate the potential energy of ocean water to mitigate the currently observed 3 mm yr^{-1} will exceed 7 % of the current global primary energy supply. At the same time the approach may be the only way to protect entire coastlines or specific regions that cannot be protected by dikes.

1 Introduction

Anthropogenic emissions of carbon into the atmosphere have increased global temperatures by almost one degree during the past two centuries (IPCC, 2013). Even after a complete cessation of carbon emissions temperatures are not expected to drop significantly for several centuries. Although the reduction of short-lived forcing agents has the potential to reduce the near-term warming by about 0.5°, global mean temperatures are not projected to decline significantly even under the strongest mitigation scenario, RCP2.6, which is already accounting for reductions in black carbon, methane, and other short-lived forcers (van Vuuren et al., 2011). As a consequence of the climate system's inertia and the ice sheets' response, mitigating

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greenhouse-gas emissions in the future can reduce but will not stop sea-level rise for centuries to come (Gillett et al., 2011; IPCC, 2013; Meehl et al., 2005; Solomon et al., 2009; Wigley, 2005). Conservative estimates for this so-called sea-level commitment are of the order of $2\,\mathrm{m\,degree}^{-1}$ of global warming above pre-industrial temperatures over a time period of two millennia (Levermann et al., 2013); estimates of sea-level sensitivity to warming from paleo-records of earlier warm periods are even higher (Rohling et al., 2008, 2013). In addition, there is recent evidence from observations (Rignot et al., 2014) and numerical models (Favier et al., 2014; Joughin et al., 2014) that the West Antarctic Ice Sheet has entered a state of irreversible ice discharge that would cause a sea-level contribution even without any additional warming. The associated long-term sea-level rise is estimated to be 1.1 m from the Amundsen-Sea sector or 3.3 m if the entire marine part of West Antarctica is affected (Bamber et al., 2009).

As a consequence global coastal adaptation to an ongoing sea-level rise will be required unless water is taken back out of the ocean. Such local protection may not be physically possible or economically feasible everywhere. In South Florida in the USA, for example, the base rock is limestone which makes the construction of levees very difficult (Strauss et al., 2014). In addition, dams may not be acceptable for some regions that live on tourism attracted by natural beaches. Local protection will most likely only be done for areas with high assets at risk and will not cover entire coast lines including poor areas and ecosystems.

Here, we evaluate, from an ice-dynamic perspective, the option of delaying global sea-level rise for three idealized scenarios of linear increases over the 21st century. Mitigating the currently observed rate (Cazenave and Dieng, 2014) of $\sim 3\,\mathrm{mm\,yr}^{-1}$ is considered in comparison to the mitigation of $1\,\mathrm{mm\,yr}^{-1}$ and a rate of $10\,\mathrm{mm\,yr}^{-1}$ within the "likely range" for the end of the century under the high emission scenario RCP8.5 (IPCC, 2013). As it might be difficult to store such large water masses in liquid form on land due to adverse effects on population, regional ecosystems, and expected changes in the hydrological cycle, we explore here whether it can be stored as ice on Antarctica.

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The Antarctic ice sheet is situated on the coldest continent on Earth with most of its surface temperatures far below the freezing point of ocean water throughout the year. The water volume that is equivalent to one meter of global sea-level rise would elevate the Antarctic ice sheet by ~ 25 m if distributed uniformly. The currently observed ~ 3 mm yr^{-1} of global average sea-level rise due to thermal expansion, additional water added from glaciers and ice sheets, and changes in land water storage corresponds to about 10^{12} m^3 yr^{-1} of ocean-water volume. Solely in terms of throughput mitigating a sea-level rise of 3 mm yr^{-1} would require 90 of the largest pump stations currently under construction in New Orleans each assumed to pump ~ 360 m^3 s^{-1} which corresponds to ~ 11×10^9 m^3 yr^{-1} (Alyeska Pipeline Service Company, 2013). The height of the ice sheet of about 4000 m means that it would require a constant power of 1275 GW to elevate the potential energy of the associated ocean water. This is equivalent to ~ 7 % of the global primary energy supply of the year 2012 (International Energy Agency, 2014). The power required for the actual pumping may even be higher and reach 2300 GW under optimistic assumptions (see Supplement). It will have to be generated by renewable resources to avoid the additional climate change and sea-level rise associated with fossil fuels. The Antarctic continent is windy enough to support such pumping through wind energy, with around 16.7 TW available in a 200 km wide band alone the coast of East Antarctica (Archer and Jacobson, 2005) (see Supplement). Around 8 % of that energy would need to be extracted to compensate the potential energy increase of the pumped water alone, which is equivalent to 850 000 wind-energy plants of 1.5 MW running on full capacity. Although the approach may be the only way to protect entire coast lines it will not be feasible without major technical innovations solving the fundamental energy problem. In the following we explore the option from an ice dynamical point of view.

Antarctica's currently observed ice loss occurs near the coast (Shepherd et al., 2012) while the surface in its interior is moving at a speed of less than 0.1 m yr^{-1} (Rignot et al., 2011). Because the ice is continually moving, ocean water put on the ice sheet will only

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delay sea-level rise. Here we estimate the associated delay time and its dependence on the distance from the coast and the application rate.

2 Ice sheet simulations

Using the Parallel Ice Sheet Model (PISM) (Bueler and Brown, 2009; Winkelmann et al., 2011) we estimate the ice sheet's response to different scenarios of ice addition. To this end we ran the model to equilibrium in a 100 kyr spin-up under constant atmospheric and oceanic boundary conditions. Surface air temperature and mass balance are taken from observations (Comiso, 1999). We apply sub-shelf basal melting and refreezing rates from a 20th century simulation of the Finite Element southern ocean model (FESOM) (Timmermann et al., 2012).

In our forcing simulations we disturb this equilibrium state with 100 year-long pulses of increased surface mass balance (SMB) in selected bands (see Fig. 1). The added surface mass compensates a 1, 3 and 10 $\mathrm{mm\,yr^{-1}}$ sea level rise (ΔSMB = rate of sea level rise × global ocean area/area of mass addition). The maximum sea level drop is therefore 1 m. We construct the bands of surface mass addition by drawing lines with distances of 200, 300, 400, 500, 600, 700, and 800 km from the coast and use these as the center lines of 200 km wide bands. The bands are limited to longitudes between 20° W and 165° E to only cover the East Antarctic ice sheet because of the currently observed imbalance in West Antarctica. The rate is applied for 100 years and then set to zero. Whether the pumping should be time-limited needs to be decided by society if such measure is ever to be implemented.

By adding the ice to the surface of an equilibrium simulation we do not account for any drift that might have been caused by previous variations in the boundary conditions such as the last deglaciation or the medieval warm period or anthropogenic warming. Although a possible drift within the present-day ice sheet could potentially alter the ice export as reported here, it can be assumed that the drift is negligible at distances of several hundred kilometres away from the coast.

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To represent the large-scale dynamics reasonably well, we use a 12 km horizontal resolution for the ice sheet simulations. Our hybrid shallow approximation ensures stress transmission across the grounding line and a smooth transition between regimes of fast flowing, sliding ice and slowly deforming, bedrock-frozen ice. The grounding line can freely evolve also under lower resolution due to a local interpolation of the grounding-line position that affects the basal friction and a new driving stress scheme at the grounding line. The interpolation leads to a reversible grounding-line dynamics that is consistent with Full-Stokes simulations in high resolution (Feldmann et al., 2014). Although the model is capable of simulating the coastal dynamics of the ice sheet within limitations, it is important to note that the results obtained here are predominantly dependent on the ice flow representation in the interior of the ice sheet for that large-scale continental ice-sheet models like PISM and others (Bindschadler et al., 2013; Calov et al., 2010; Greve et al., 2011; Huybrechts and Wolde, 1999; Pollard and Deconto, 2009; Swingedouw et al., 2008) have been designed.

In the standard simulation we do not alter the surface air temperature during the mass addition. To estimate the effect of surface warming due to the latent heat release of the sea water we conduct a second set of simulations that keeps the surface temperature at the freezing point of sea water during the surface mass addition.

3 Results

The ratio of the volume added during the first 100 years and the volume that is lost again after 1000 years depends strongly on the distance from the coast (Figs. 2 and 3). Consistent with earlier studies (Huybrechts and Wolde, 1999; Winkelmann et al., 2012) an ice volume equivalent to 10–15 % of the added ice is already lost at the end of the forcing period when the ice is added at a distance of 200 km from the coastline, while the sea-level contribution is strongly delayed at a distance above 500 km from the coast (Fig. 2). In order to minimize the return flow of the ice into the ocean, the specific positioning of the ice addition could be varied spatially making use of slow moving ice

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regions. Here we apply a simplified spatial distribution in order to demonstrate the main process and allow for a conceptual analysis of the simulations.

The time after which a certain equivalent of the added ice has been discharged into the ocean, is much shorter than would be expected from a mere advection of the added ice mass with the surface velocities of the ice sheet (Fig. S2 in the Supplement). That is because the ice thickness anomaly creates an imbalance between the driving stress and the viscous ice flow. As a consequence the ice transport occurs in waves from the strip of perturbation to the coast (Winkelmann et al., 2012) (Fig. 1) with the ice being discharged to the ocean not being the same ice that was added to the ice sheet earlier. Even though the ice wave also travels partially inland, it is possible that more ice is transported out of the continent than was initially added. This ice-loss exceedance, however, occurs only several millennia after the perturbation (Fig. S2). Whether it is directly related to the perturbation or a manifestation of a localized multistability of the ice dynamics is difficult to identify, because the differences between the initial and final ice topography are within the uncertainty range of the model performance.

The freezing of ocean water that is pumped onto the ice surface will release latent heat that will heat up both the atmosphere and the upper ice layer. The maximum injection of latent heat occurs for the $10\,\mathrm{mm\,yr^{-1}}$ sea-level-mitigation scenario and the 800 km band, which has the smallest area. The corresponding addition of $3.2\,\mathrm{m\,yr^{-1}}$ liquid sea water is equivalent to a latent heat injection of about $35\,\mathrm{W\,m^{-2}}$. In order to estimate the effect of the surface warming on the ice dynamics, we conduct a second set of simulations in which the surface temperature is held at the freezing point of sea water of $-1.9\,^{\circ}\mathrm{C}$ during the forcing. This imitates the situation in which the surface of the ice remains in a mixed state of ice and water. As Antarctica's inland-surface temperatures are far below zero, this constitutes a strong warming signal that diffuses down into the ice body and causes ice to soften and flow faster.

A warming from -20 to $-1.9\,^{\circ}\mathrm{C}$ would increase the long-wave radiative loss to the atmosphere by $70\,\mathrm{W\,m^{-2}}$ according to the Stefan–Boltzmann law assuming an emissivity of 0.95. If open water areas are sustained on the ice sheet, a sensible-

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heat-dominated loss can remove heat at a rate of $100\,\mathrm{W\,m^{-2}}$ or more as observed in ocean polynyas (Launiainen and Vihma, 1994). The maximum rate of latent heat injection of $35\,\mathrm{W\,m^{-2}}$ is much smaller than the potential of the atmosphere to remove the heat. Thus keeping surface ice temperatures at freezing point underestimates the atmospheric heat loss so that the simulations provide an upper bound for the induced warming of ice. For perturbation areas with large distance from the coast, the discharge rate of the ice sheet (Fig. 4, thin lines) is nearly identical to the response without warming (Fig. 4, thick lines) on the millennial time scale considered. Within the first millennium, the latent heat release of freezing sea water only alters the discharge when placed near the coast. After 2000 years, the additional warming can induce a discharge exceeding 100 % of the added ice in the 200 km simulations (Fig. S1).

It is possible that ice dynamic effects which are not included in these simulations (such as ice fractures or basal sliding conditions) alter the results quantitatively, but the shallow ice approximation which dominates the ice dynamics in the model in the interior of Antarctica has been shown to represent the interior ice sheet flow on multi-centennial and longer time scales (Greve and Blatter, 2009).

4 Discussion

All scenarios considered here assume that the only perturbation of the ice sheet is the addition of ice mass in bands of the interior of East Antarctica. At the same time Antarctica's coastal regions are out of balance in a number of regions predominantly in West but also in East Antarctica. In this study it is assumed that the addition of ice in the interior will not interfere with the imbalance at the coast. This might be an over-simplification, but currently available modeling studies (Favier et al., 2014; Joughin et al., 2014; Mengel and Levermann, 2014) indicate that perturbations near the coast will not reach as far inland over time periods of several centuries. A possible interference between the interior of the ice sheet and its coastal regions, however, needs further investigation, possibly with higher-resolution regional ice sheet models.

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We assume that it would be best to add the additional ice in form of snow as opposed to adding it as water which will then freeze. In this context it has to be noted that the additional ice that is added to the ice sheet is made of sea water and thereby will have salinity. The rheological effects of a "salt-ice" layer within an ice sheet are currently unknown and need further investigation. While initially the "salt-ice" will be at the surface of the ice sheet the dynamic effect will be dominated by its gravitational effect which is covered by the ice dynamics modelled. However, while snow is falling onto the "salt-ice" layer this layer's rheology will become relevant for the ice dynamics. At current and future snowfall rates (Frieler et al., 2015) this will take several centuries.

The conducted simulations suggest that pumping ocean water onto the interior of the Antarctic ice sheet can impose a significant delay of future sea-level rise. However, as an option to mitigate the sea-level rise we are already committed to a substantial energy problem has to be overcome. It goes far beyond the scope of existing projects and will require major technical innovations if at all possible. Therefore costs cannot be reliably estimated. Based on simple upscaling of the costs of the Trans Alaska Pipeline with height, length and throughput (see Supplement) the costs will be orders of magnitude higher than the costs associated with local adaptation measures (Hinkel et al., 2014). However, it is important to note that in current studies on sea-level adaptation, protection is only installed if considered economically favorable. In contrast, storing water on the Antarctic Ice Sheet would offer a general protection for entire coast lines and poor regions that would otherwise left unprotected. The associated investment could change the mitigation costs by increasing the demand (thereby technical progress) of renewable energies significantly. By creating an additional demand for renewable energies of the order of 10 % of the global energy supply, this measure thereby constitutes a link between the mitigation and adaptation problem of climate change.

This study has to be complemented by investigations on possible consequences of the procedure. To name just a few: it is likely that construction of the pipelines, pump stations, the energy generation, and the water extraction will induce disturbances in the

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coastal ecosystems. It has to be investigated how the water extraction will influence the small- and large-scale ocean circulation. The ice-rheological changes that are induced by the addition of salt water have to be investigated together with potential effects on the basal conditions of the ice.

The heat released from freezing and the pumping process itself is in the order of 10 TW (latent heat) + 1 TW (heat released from pumping). This corresponds to about 10 % of the maximum increase in latent heat transport in high northern latitudes under an SRES A1B transient simulation (Held and Soden, 2006). The latent heat release is considered as a main contribution to the Arctic amplification of global warming. Put into this perspective the pumping-induced energy over Antarctica is not negligible but significantly smaller than the warming induced latent heat released in northern high latitudes. Potential consequences for the atmospheric and oceanic circulation have to be further explored.

When the pumping is stopped the additional discharge from Antarctica will increase the rate of sea-level rise even beyond the warming-induced rate. In this sense the presented approach means raising a loan on Antarctica that future generations will have to pay back. In all simulations considered here pumping is stopped after 100 years, i.e. it is investigated as an option to delay part of the sea-level rise we are already committed to but not as a permanent measure that may induce further responses of the ice sheet not captured here.

If at all feasible, the considered scenarios do not at all represent an alternative to the mitigation of carbon emissions because the method does not address any other climate-change impact than sea-level rise. Furthermore unmitigated emission might induce a sea-level rise of 10 $mm\,yr^{-1}$ and beyond, which increases the impacts on Antarctica and the burden for future generations when mitigated by pumping of ocean waters. Thirdly, after the pumping is stopped, the sea level will accelerate quickly towards the rate that corresponds to the warming level plus that induced by the addition of ice onto Antarctica.

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Although a potential way to delay the committed sea-level rise from an ice dynamical perspective, it represents an open engineering problem whether it is at all possible to locally generate the required energy.

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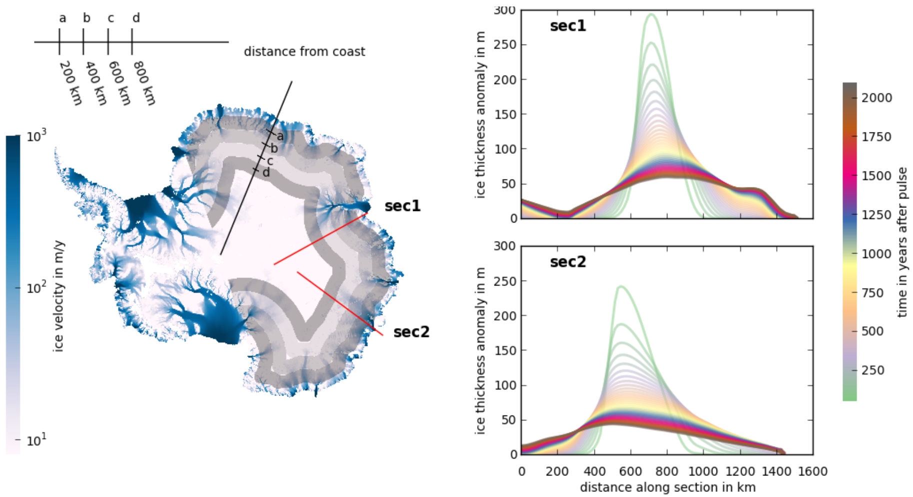

Figure 1. Bands of ice mass addition on East Antarctica and ice thickness relaxation for the 800 km band. Left panel: surface velocities of the ice flow of the Antarctic ice sheet (blue shading). Grey strips indicate where ice mass was added to East Antarctica in order to delay future sea-level rise in the different simulations. The ice was added in strips of 200 km width for 100 years. The right panels show the ice thickness relaxation after the end of the mass addition to the 800 km band in time steps of 50 years for two representative sections (left panel, red lines) as anomaly to the equilibrium simulation.

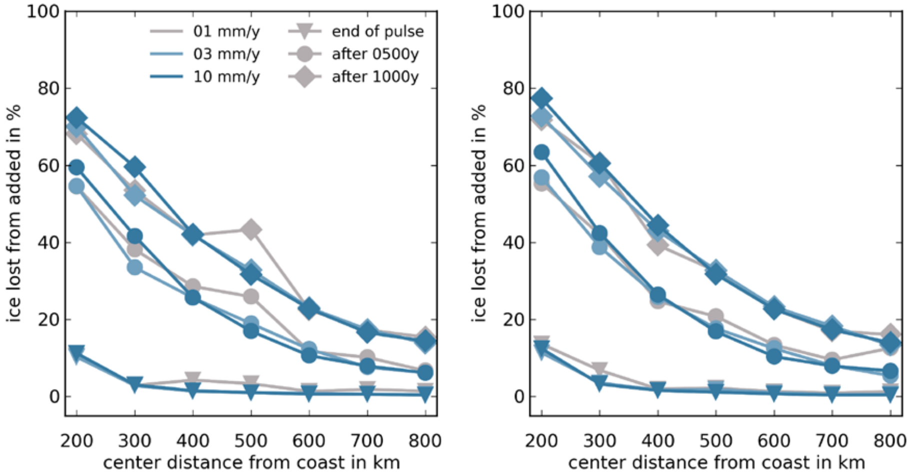

Figure 2. Additional ice discharge compared to mass addition for different distances from the coast. The fraction of the added ice that is lost again to the ocean as a function of distance from the coast at which the additional ice was placed for the simulations without (left panel) and with (right panel) surface warming. Colours indicate the magnitude of the mass addition equivalent to 1, 3 and 10 $\mathrm{mm\,yr^{-1}}$ sea level rise mitigation. Markers correspond to the different relaxation times: end of pulse, 500 and 1000 years after the pulse of mass addition ended.

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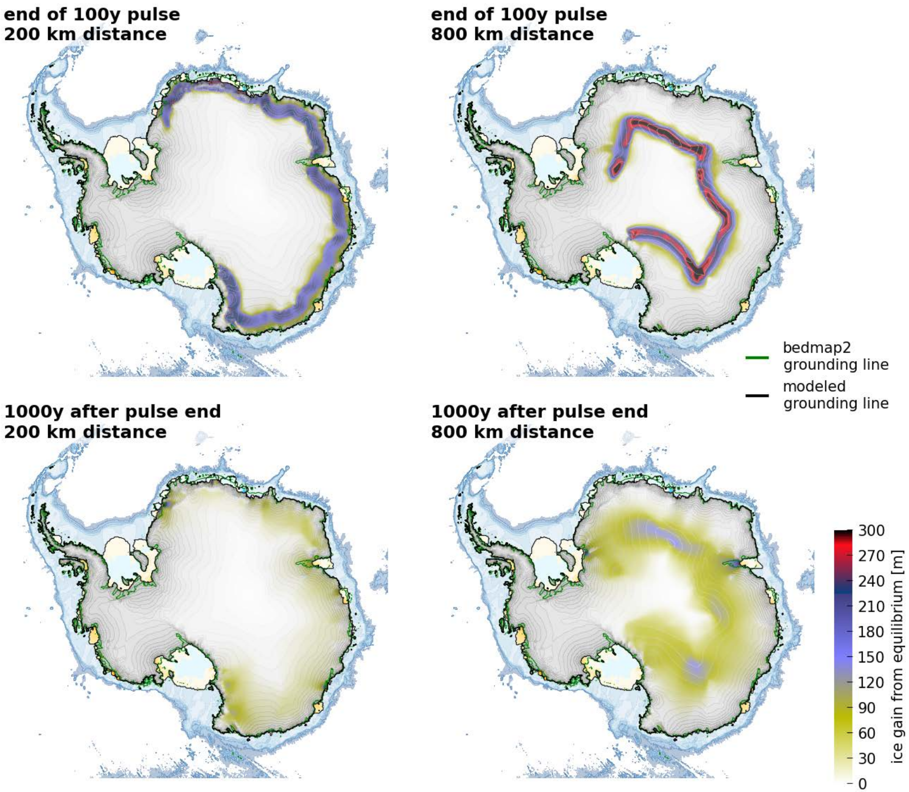

Figure 3. Difference in ice thickness compared to the initial state. Ice thickness gain at the end of the 100-year-long mass addition (upper panels) and 1000 years after the forcing ended (lower panels). The close-to-coast simulation (left panels) has lost most of the added ice to the ocean after 1000 years while there is a broad ice gain in the 800 km simulation (right panels). Figures are shown for the strongest scenario of 10 mm yr^{-1} of sea-level mitigation and without accounting for latent-heat release.

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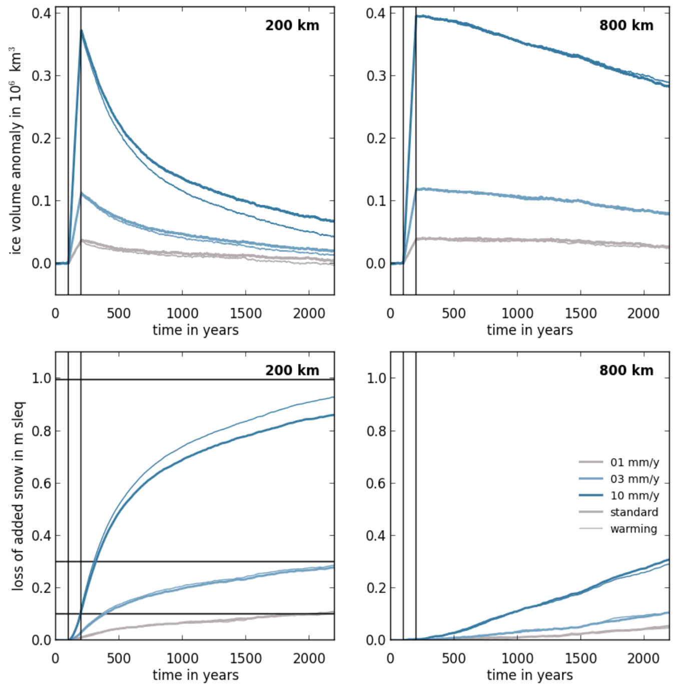

Figure 4. Response of the Antarctic Ice Sheet to a 100-year-surface-mass addition. Ice volume gain and relaxation (upper panels) and equivalent loss of the added snow (lower panels) for the close-to-coast (200 km, left panels) and farthest-inland (800 km, right panels) simulations for the 1, 3 and 10 mm yr^{-1} mitigation cases (grey, light blue and dark blue colours). Thick lines indicate simulations without surface warming, thin lines with surface temperature held at sea water freezing point during the forcing (upper bound). Vertical lines indicate the pulse interval, horizontal lines in lower left panel indicate the total ice volume added.

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