Delaying future sea-level rise by storing water in 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 2 °C 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 of delaying 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 coastline. 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 offers a comprehensive protection for entire coastlines particularly including regions that cannot be protected by dikes.

1 Introduction

Anthropogenic emissions of carbon into the atmosphere have increased global temperatures by almost 1 °C 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 °C, 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 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 m per 1 °C 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 were affected (Bamber et al., 2009).

As a consequence, global coastal adaptation to 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 southern 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 rely on tourism associated with natural beaches. Local protection will most likely only be done for areas where valuable assets are at risk and will not cover entire coastlines, including poor areas and ecosystems.

Here we evaluate 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 $\sim 3$ mm yr$^{-1}$ is considered in comparison to the mitigation of 1 mm yr$^{-1}$ and a rate of 10 mm yr$^{-1}$ within the “likely range” for the end of the century under the high-emission scenario RCP8.5 (IPCC, 2013). Since 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 whether it is possible to store it as ice in Antarctica from the perspective of ice dynamics.

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 equivalent to 1 m of global sea-level rise would elevate the Antarctic ice sheet by $\sim 25$ m if distributed uniformly. The currently observed $\sim 3$ 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. 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 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 ice addition scenarios. To this end we ran the model to equilibrium in a 100 ka spin-up under constant present-day atmospheric and oceanic boundary conditions. Surface air temperature (Comiso, 2000) and mass balance (Arthern et al., 2006) are taken from observations made available in the ALBMAP data set (Le Brocq et al., 2010). 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). The modeled equilibrium ice sheet state compares well to the currently observed ice sheet in terms of surface elevation and grounding line position (supplementary Fig. S2) and ice velocities (Supplement Figs. S3 and S4). Total modeled ice volume deviates less than 0.5 % from the observed state (Fretwell et al., 2013).

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 for a 1, 3, and 10 mm yr$^{-1}$ sea-level
rise \( \Delta \text{SMB} = \text{rate of sea-level rise} \cdot \text{global ocean area/area of mass addition} \). 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 we 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 in order to better estimate the sea-level delay time. Whether the pumping should be limited needs to be decided by society if such a 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, 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 kilometers away from the coast.

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, and slowly deforming bedrock-frozen ice. The grounding line can freely evolve even at lower resolution due to a local interpolation of the grounding-line position, which affects the basal friction and a new driving stress scheme at the grounding line. The interpolation leads to reversible grounding-line dynamics consistent with full-Stokes simulations at 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 which large-scale continental ice-sheet models like PISM and others (Bindschadler et al., 2013; Calov et al., 2010; Greve et al., 2011; Huybrechts and De 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 mass addition. To estimate the effect of surface warming due to the latent heat release of the seawater, we conduct a second set of simulations that keeps the surface air temperature at the freezing point underestimates the atmospheric heat loss so that the simulations provide an upper bound for the induced warming of ice.

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

The time after which the equivalent of 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. S5 of the Supplement). This 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), where the ice discharged to the ocean is not 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. However, this ice-loss exceedance occurs only several millennia after the perturbation (Fig. S5 in the Supplement). 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.

For perturbation areas far from the coast, the discharge rates in the sensitivity simulation accounting for latent heat release (Fig. 4, thin lines) are nearly identical to the response without warming (Fig. 4, thick lines) on the millennial timescale considered. Within the first millennium, the latent heat release of freezing seawater 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 in the Supplement).

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 Antarctica 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. However, a possible interference between the interior of the ice sheet and its coastal regions needs further investigation, possibly with higher-resolution regional ice sheet models.

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. However, the shallow-ice approximation that 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 timescales (Greve and Blatter, 2009).

We assume that it would be best to add the additional ice in the 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 seawater 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 modeled ice dynamics modeled. However, since 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 simulations conducted here 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 to which we are already committed, a substantial energy problem must be overcome. Solely in terms of throughput, mitigating 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 ~1 \times 10^{13} 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 Sect. S1 in the 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 using wind energy, with around 16.7 TW available in a 200 km wide band along the coast of East Antarctica (Archer and Jacobson, 2005) (see Sect. S2 in the Supplement). Around 8% of that energy would need to be extracted to compensate for 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.

The scope of such a project is unprecedented and would require major technical innovations, if possible at all. Therefore, costs cannot be reliably estimated. Based on simple up-scaling of the costs of the Trans-Alaska Pipeline with height, length, and throughput (see Sect. S1 in the 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 this study on sea-level adaptation, protection is only installed if considered economically favorable. In contrast, storing water on the Antarctic ice sheet would offer general protection for entire coastlines and poor regions that would otherwise be left unprotected. The associated investment could change the mitigation costs by significantly increasing the demand (thereby technical progress) for renewable energies. By generating an additional demand for renewable energies of the order of 10% of the global energy supply, this approach offers a link between the mitigation and adaptation problem of climate change.

This study must 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 coastal ecosystems. It should also be investigated how the water extraction will influence the small- and large-scale ocean circulation. The ice-rheological changes induced by the addition of salt water should be investigated together with potential effects on the basal conditions of the ice.

The heat released from freezing and the pumping process itself is of 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 a major contribution to the Arctic amplification of global warming. From 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 need to be further explored.

5 Ethical considerations

The Protocol on Environmental Protection to the Antarctic Treaty (Secretariat of the Antarctic Treaty, 1991) has declared a clear intention to minimize human influences on the Antarctic continent. The signing parties are “Convinced that the development of a comprehensive regime for the protection of the Antarctic environment and dependent and associated ecosystems is in the interest of mankind as a whole” and “commit themselves to the comprehensive protection of the Antarctic environment and dependent and associated ecosystems and hereby designate Antarctica as a natural reserve, devoted to peace and science.” The measures proposed here (if at all feasible) mean a major human intervention, putting the ecosystems of Antarctica and of the surrounding ocean at a high risk. Thus, the protection of global coastlines and associated natural and human would not only have to be weighted against the enormous efforts but also against the loss of Antarctica as a unique natural reserve.
Storing water on the Antarctic continent also raises questions of inter-generational justice. When pumping is stopped, the additional discharge from Antarctica will increase the rate of sea-level rise even beyond the warming-induced rate. In this way, the approach presented here means taking out a loan on Antarctica that future generations will have to pay back. In all simulations considered here, pumping ceases after 100 years: 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 climate-change impact other 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. And, after pumping is stopped, sea level will accelerate quickly towards the rate that corresponds to the warming level, plus that induced by the addition of ice to Antarctica.

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

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Supplement of

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S1 Efforts of delaying sea level rise in comparison to local adaptation

Storing ocean water 700km inland on the Antarctic Ice Sheet requires overcoming a height difference of about 4000m. Mitigating a sea-level rise of 3 mm yr\(^{-1}\) requires 1275 GW of power to increase the potential energy of the water accordingly. This corresponds to a power of 10 kW that is required to pump 1 m\(^3\)/s vertically to a height of 1m. The actual energy required for the pumping will be higher due to friction and inefficient pumping. Here we assume a required power of 18 kW to pump 1 m\(^3\)/s to a height of 1m. This value is based on the project Tagus-Segura Water Transfer (Melgarejo Moreno, Joaquín Sanz Montaño, 2009). In its first section (TRAMO I) a power of 135 MW is applied to achieve a throughput of 31.7 m\(^3\)/s over a height difference of 245 m, which corresponds to 17.3 kW/(m\(^3\)/s)/m. This would mean that the required power to mitigate a sea-level rise of 3 mm yr\(^{-1}\) would reach about 2300 GW. To avoid further greenhouse-gas emissions the power would have to be generated based on local renewable technologies.

That poses a fundamental engineering problem that goes far beyond the scope of existing projects. While wind power plants and pipeline technology readily exist, currently the required technology is not available for the very low temperatures of Antarctica, as far as we know. We cannot provide reliable estimates of the associated costs at this stage.

However, to put the engineering effort into perspective we here compare the technical requirements of the pumping and transporting system to that of the Trans Alaska Pipeline (TAP)(Alyeska Pipeline Service Company, 2013). Thereby we assume that the transport of the water is split up into 90 individual pipelines each ensuring a throughput of \(1.1 \times 10^9\) m\(^3\) yr\(^{-1}\) as provided by the largest New Orleans pumping station.
<table>
<thead>
<tr>
<th></th>
<th>Trans Alaska Pipeline</th>
<th>Individual pipeline of the potential Antarctic pumping system</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Length</strong></td>
<td>1287 km</td>
<td>700 km</td>
</tr>
<tr>
<td><strong>Maximum elevation</strong></td>
<td>1444 m</td>
<td>4000 m</td>
</tr>
<tr>
<td><strong>Air temperature along route</strong></td>
<td>-60°C to 35°C</td>
<td>far below 0°C</td>
</tr>
<tr>
<td><strong>Annual throughput</strong></td>
<td>(0.1245 \cdot 10^9) m(^3) (assuming a maximum daily throughput as reached in 1988)</td>
<td>(11 \cdot 10^9) m(^3)</td>
</tr>
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</table>

The costs of building the TAP amount to a total of 8 billion US$. On the one hand this includes costs not relevant for the Antarctic pipelines and pumps such as:

- costs of the Valdez oil terminal (US$1.4)
- launching/receiving facilities for cleaning pigs that sweep the pipe of built-up wax, water or other solids that precipitate out of the oil stream
- 124,300 heat pipes along the pipeline transferring ground heat into the air during cold periods to lower the ground temperature to ensure that soils remain frozen throughout the summer to steadily support the pipeline (to be replaced by alternatives adequate for Antarctica).

While on the other hand the Antarctic systems will

- have to provide an orders of magnitude higher throughput (assuming 90 individual lines starting at the ocean with an annual throughput of \(11 \cdot 10^9\) m\(^3\) as provided by the largest New Orleans pump station would require an upscaling of the maximum annual throughput of the TAP (calculated based on its maximum daily throughput reached in 1988) by a factor of 90 for each individual line).
- have to overcome a larger elevation difference (4000m in comparison to a maximum of 1444 m along the TAP)
- require heating along the pipelines
require a system to distribute the water on the ice sheet

need an adjustments of the pipeline due to moving ice
to provide a simplistic upscaling of the TAP costs excluding the terminal (6.6 US$) to the required length (L), height (H), and capacity (P) we assume that the overall costs C can be split up into components, a first one C\_1 that scales with P and H (mainly representing the costs of the pump stations) and a second one C\_2 that scales with L and P (mainly representing the costs of the pipeline itself). Assume that the two components of the overall costs of the TAB C\_TAB are given by C\_1\_TAB = \alpha C\_TAB and accordingly C\_2\_TAB = (1-\alpha) C\_TAB that would mean that

\[ C = C\_1\_TAB \times \frac{P}{P\_TAB} \times \frac{H}{H\_TAB} + C\_2\_TAB \times \frac{P}{P\_TAB} \times \frac{L}{L\_TAB} \]

\[ \approx \alpha \times 6.6 \times 10^{12} \text{ US$} \times 8015 \times 3 + (1 - \alpha) \times 6.6 \times 10^{12} \text{ US$} \times 8015 \times 50 \]

with C = 158,697 billion US$ for \( \alpha = 1 \) and C = 2,644,550 billion US$ for \( \alpha = 0 \). Both numbers do not include the costs of running and maintaining the system. Assuming they are minor compared to the cost of construction, construction costs could be divided by 100 to estimate the annual costs over the 21st century. In both cases they are orders of magnitude higher than the estimated annual investment and maintenance costs of local adaptation by dikes of US$ 12–71 billion in 2100 (Hinkel et al., 2014) which however only protects regions that are economically expendient to protect. This might exclude for example the AOSIS small island states and their culture as well as UNESCO cultural heritage sites (Marzeion and Levermann, 2014).

That means that such a project would only become competitive with major efficiency gains by technical innovations and learning from the experience of the project itself.
S2 Wind Energy potential in Antarctica

We follow the methodology of ref. (Archer and Jacobson, 2005) to estimate the wind energy that is available in a 200 km wide band along the East Antarctic coast (Fig. S1). We use the yearly mean wind data for 2014 at 10m above the surface from the Era-Interim dataset (ECMWF, n.d.). Following ref. (Archer and Jacobson, 2005), we use only regions with average wind speed above 6.5 m/s. The power of a single wind turbine can be estimated by (ref. (Archer and Jacobson, 2005), equ. 19):

\[ P = P_{\text{rated}} \times CF = P_{\text{rated}} \times (0.087V - \frac{P_{\text{rated}}}{D^2}) \]

with \( P \) as average power output, \( P_{\text{rated}} \) the rated power of a single wind turbine, \( CF \) the conversion factor, \( V \) the average yearly wind speed and \( D \) the rotor diameter of the wind turbine. Six wind turbines of the 1.5 MW class can be installed per square kilometre (Archer and Jacobson, 2005). Integrating over the East Antarctic band area with wind stronger than 6.5 m/s, we estimate a wind power potential of 16.7 TW. Regions difficult to access, such as mountain ranges, and regions of fast flowing ice have not been excluded from the total area. We do not apply a conversion of 10m above-surface winds to such at the height of the rotor (80m in ref. (Archer and Jacobson, 2005)). As winds at 80m are generally higher than at 10m above the surface, the real wind energy potential may be higher.

Wind power utilization in Antarctica will pose a number of major challenges that cannot be addressed here. These include building fundaments on moving and deformable ground, functioning under extremely cold temperatures in Antarctic winters and handling very high wind velocities.
Fig. S1: Wind energy potential in East Antarctica. Using ERA-Interim data (ECMWF, n.d.), we use the methodology of ref. (Archer and Jacobson, 2005) to estimate the wind power potential in a 200km-wide strip along the East Antarctic coast. High wind speeds above 6.5 m/s on yearly average in the major part of the strip allow for large-scale wind energy extraction.
**S3 Modelled initial equilibrium ice sheet state**

Fig. S2: Comparison of the currently observed surface elevation and grounding line position (left panel) to the simulated initial state of the ice sheet (right panel). The grounding line (red line, Bindschadler et al., 2011) is generally well reproduced for East Antarctica and most parts of West Antarctica. East Antarctica’s modelled surface elevation compares well to the observed state (Fretwell et al., 2013). The modelled surface elevation is slightly higher close to the South pole and slightly lower towards the East Antarctic coast as compared the observation.
Fig. S4: Comparison of the currently observed ice velocities at the surface (left panel, Rignot et al., 2011) to the velocities of the simulated initial state of the ice sheet (right panel). The model reproduces the flow features of the East Antarctic ice sheet.
Fig. S5: Observed and modelled velocities, point-to-point comparison on log-log scale.

Observed velocities (Rignot et al., 2011) are on the x-axis and modelled velocities on the y-axis.
S4 Long term response

Fig. S5: Delayed sea-level rise as fraction of the initial reduction. Colour coding indicates the application band where the ice is added (red: 800 km distance from the coast, violet: 500 km distance from the coast, and blue: 200 km distance from the coast). The colour intensity describes the application rate from 1 mm yr\(^{-1}\) (light), via 3 mm yr\(^{-1}\) (normal) to 10 mm yr\(^{-1}\) (dark). Thin lines show the results of the simulations accounting for latent-heat release while thick lines represent the results without the additional warming. Vertical lines mark the 5\(^{th}\) percentile of the distribution of delay times assuming an ice parcel traveling 800 km (red), 500 km (violet), and 200 km (blue) with a constant surface velocity drawn from the distribution of surface velocities observed within the area of less than 800 km, 500
149 km and 200 km distance from the coast. Real advection of ice particles to the coast by the
150 surface velocities will even take longer because generally the ice will not be flowing the
direct path assumed here.

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