CLIMATE CHANGE

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Combustion of available fossil fuel resources sufficient to eliminate the Antarctic Ice Sheet

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The Antarctic Ice Sheet stores water equivalent to 58 m in global sea-level rise. We show in simulations using the Parallel Ice Sheet Model that burning the currently attainable fossil fuel resources is sufficient to eliminate the ice sheet. With cumulative fossil fuel emissions of 10,000 gigatonnes of carbon (GtC), Antarctica is projected to become almost ice-free with an average contribution to sea-level rise exceeding 3 m per century during the first millennium. Consistent with recent observations and simulations, the West Antarctic Ice Sheet becomes unstable with 600 to 800 GtC of additional carbon emissions. Beyond this additional carbon release, the destabilization of ice basins in both West and East Antarctica results in a threshold increase in global sea level. Unabated carbon emissions thus threaten the Antarctic Ice Sheet in its entirety with associated sea-level rise that far exceeds that of all other possible sources.

How the Antarctic Ice Sheet evolves in response to future emissions of greenhouse gases is of primary importance for coastal populations (*1*) and ecosystems (*2*). Although Antarctica has already begun to lose ice (*3*), the consequences of combustion of the remaining fossil-fuel resources (*4*) to the ice sheet's future mass balance are still unknown. Antarctica's contribution to future sea level is determined by changes in its surface mass balance and dynamic discharge (*5*). Atmospheric and oceanic warming can lead to enhanced ice loss from the Antarctic Ice Sheet or even its disintegration, potentially sped up by positive feedback mechanisms such as the marine ice sheet instability (*6*, *7*) and the surface elevation feedback (*8*, *9*). Enhanced snowfall over Antarctica, on the other hand, might offset or exceed this ice loss (*10*). The interaction of these processes is still insufficiently understood so that the evolution of the Antarctic Ice Sheet remains unclear, particularly for the long term.

We examine the ice-sheet evolution over the next ten thousand years with the Parallel Ice Sheet Model (PISM) (*11*–*13*), taking all of these processes into account. PISM represents the ice flow in sheets, streams, and shelves in a consistent way through a hybrid scheme of shallow approximations (see Materials and Methods), and has been shown to give a good approximation of grounding-line motion even at lower resolutions (*14*). We force the ice-sheet model with scenarios releasing fossil fuel emissions from ca. 100 to 12,000 gigatonnes of carbon (GtC) to the atmosphere after year 2010, following a logistic equation in time (see Materials and Methods) (*15*). We hereby cover the full range of available carbon resources (*4*). For these emission scenarios (Fig. 1A), CO_2 concentrations and global mean temperature pathways are computed using the Earth system model GENIE (*16*, *17*) (see Materials and Methods). Although the interval of carbon release into the atmosphere does not last more than 500 years in any scenario, elevated CO_2 concentrations persist for millennia. Because the efficiency of carbon buffering by both the ocean and deep-sea sediments progressively declines at higher emissions (*17*), CO_2 remains above 50% of its peak value for more than ten thousand years under the higher emissions scenarios (Fig. 1B). We find that this long lifetime of perturbations to atmospheric CO_2 concentrations in conjunction with the logarithmic nature of the warming versus CO_2 relationship means that global mean temperatures decline by less than 5% per thousand years once more than about 5000 GtC have been emitted. Instead, temperatures remain close to the maximum level corresponding to the total amount of carbon released (*18*–*20*) (Fig. 1C).

The long-term global warming scenarios generated by the Earth system model are downscaled to surface and ocean temperature anomalies for Antarctica, using ratios that were derived from long-term simulations with ECHAM5/MPIOM (*21*) (fig. S1). These regional warming scenarios are then used to force PISM and result in long-term sea-level rise as depicted in Fig. 1D.

Over the next century, the projected sea-level contribution is consistently within the range of −6 to 14 cm given for century-scale IPCC-AR5 (Intergovernmental Panel on Climate Change Fifth Assessment Report) estimates (*5*) (fig. S2). Under low emission scenarios, the century-scale mass balance is dominated by the expected increase in snowfall: Because of the higher moisture holding capacity of the warmer air surrounding the Antarctic Ice Sheet, snowfall increases, especially near the ice-sheet margins (*22*, *23*). Under higher emission scenarios, however, the dynamic discharge overcompensates for the ice gain through enhanced snowfall, and Antarctica loses mass, leading to sea-level rise over the next century.

This qualitative difference vanishes when looking at the long-term evolution of the ice sheet. Over the entire 10,000 years of simulation, sea level keeps rising for all scenarios. This extends, by more than an order of magnitude, the IPCC's statement that sea level will continue to rise for centuries to come (*5*). Figure 2 shows the ice loss in terms of global sea-level rise after 1000, 3000, and 10,000 years (see also fig. S5). Ice loss is mainly driven by two self-reinforcing feedbacks: the marine ice-sheet instability and the surface elevation feedback. In the former, ocean warming leads to enhanced sub-shelf melting and, subsequently, to a retreat of the grounding line, which separates the grounded ice sheet from the floating ice shelves. In our simulations, the basal melt sensitivity, that is, the increase of large-scale sub-shelf melting per degree of warming, is generally within the observed (*24*, *25*) range of 7 to 16 m $year^{-1}$ $°C^{-1}$ (see fig. S3). When the sub-shelf melt rate is large enough to force the grounding line to retreat into an area where the ice is grounded below

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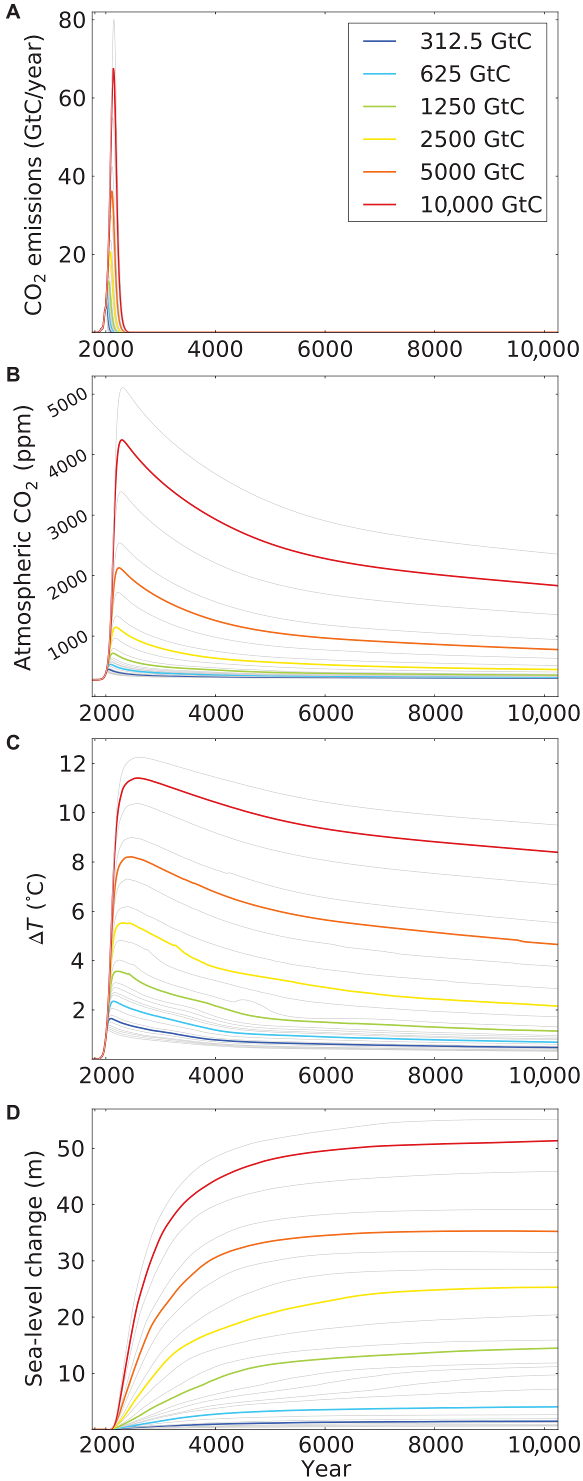

Fig. 1. Climate scenarios and time series of ice loss. (**A**) CO_2 emission scenarios releasing from 93 to 12,000 GtC to the atmosphere after year 2010, following a logistic equation (*15*). (**B**) Resulting CO_2 concentration pathways for the years 1750 to 12,000 as simulated with GENIE for the emission scenarios depicted in (A). (**C**) Global mean temperature anomaly with respect to the pre-industrial temperature, simulated with GENIE. (**D**) Ice loss from Antarctica (in meters sea-level equivalent) in response to warming of the atmosphere and the Southern Ocean, simulated with PISM.

sea level on an inward-sloping bedrock (*26*), it can become unstable. Several regions in Antarctica are subject to this so-called marine ice-sheet instability (*6*): Recent observations show that the grounding line has probably already transgressed into the unstable regime in the Amundsen Basin of West Antarctica because of warm water intrusion into the shelf cavities (though this might potentially be unrelated to anthropogenic warming) (*27*). Consistent with these observations (*27*) and with simulations performed with other ice-sheet models (*28*, *29*), in our simulations the West Antarctic Ice Sheet becomes unstable after cumulative carbon emissions reach 600 to 800 GtC.

A similar topographic configuration can be found in the Wilkes Basin, the largest marine-based drainage basin in East Antarctica. It has been shown to be subject to the marine ice-sheet instability (*30*), and, once triggered, would yield a global sea-level rise of 3 to 4 m. In our simulations, the Wilkes grounding line retreats significantly under fossil fuel carbon emissions of 1000 GtC or more. At this stage, corresponding to a peak warming of 2° to 3°C, both the West Antarctic Ice Sheet and the Wilkes Basin are experiencing rapid ice loss due to their instability, which manifests itself in a visible threshold in sea-level rise (see Fig. 2). Other marine-based drainage systems become unstable under higher emission scenarios, until most of the marine ice is eventually lost to the self-reinforcing feedback after about 2500 GtC of cumulative carbon release (Fig. 4).

Once a critical temperature is reached, a second self-reinforcing feedback kicks in and destabilizes the remaining ice: This so-called surface elevation feedback between the lowering surface elevation and the increasing surface mass loss has been mainly discussed with respect to the Greenland Ice Sheet (*8*, *9*). Numerical simulations suggest that a decline of the Greenland Ice Sheet is inevitable once the mean summer surface warming exceeds a threshold of 0.8° to 3.2°C (*9*). The long-term

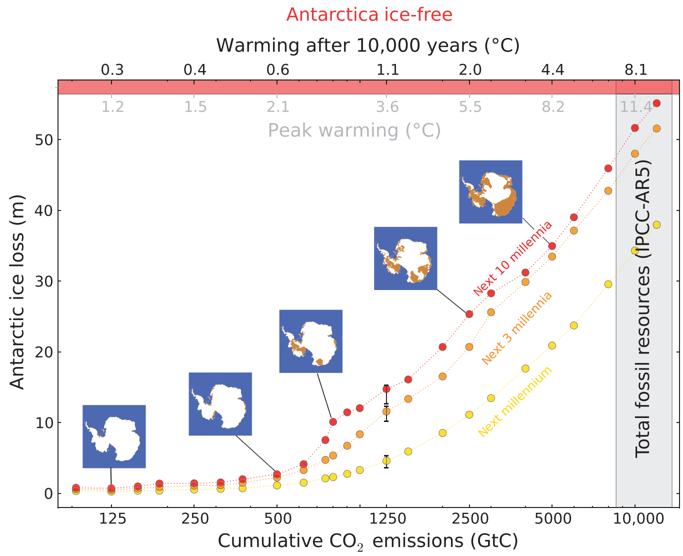

Fig. 2. Sea-level commitment from Antarctic ice loss. Given is the total sea-level change after 1000 years (yellow), 3000 years (orange), and 10,000 years (red) after year 2000 for each of the scenarios depicted in Fig. 1. The maximum temperature anomaly and the temperature anomaly after 10,000 years are given on the upper *x* axis. If all of the currently attainable carbon resources [estimated to be between 8500 and 13.600 GtC (*4*)] were burned, the Antarctic Ice Sheet would lose most of its mass, raising global sea level by more than 50 m. For the 125 GtC as well as the 500, 800, 2500, and 5000 GtC scenarios, the ice-covered area is depicted in white (ice-free bedrock in brown). For more details, see fig. S5.

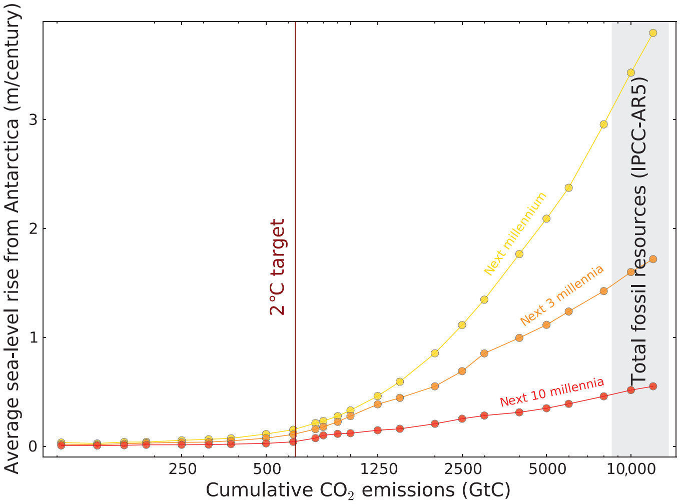

Fig. 3. Rate of sea-level rise. Given is the average rate of sea-level rise for the next 1000 years (yellow), 3000 years (orange), and 10,000 years (red). Between 2010 and 2014, there has been an increase in cumulative emissions of about 40 GtC. To avoid exceeding the 2°C limit, the cumulative emissions need to be restricted to another 600 GtC. For more details, see fig. S6.

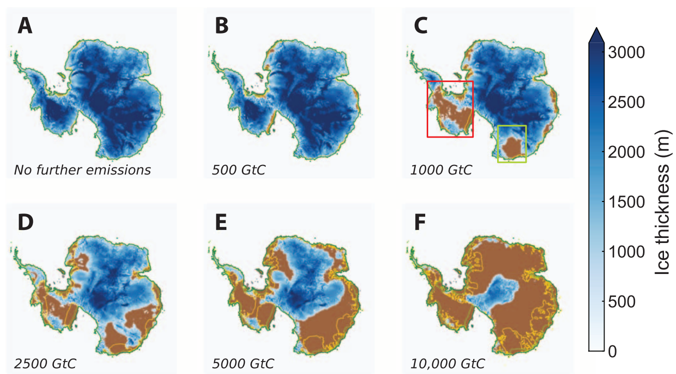

Fig. 4. States of the Antarctic Ice Sheet after 10,000 years. (A to F) Shown is the ice thickness for (A) present-day Antarctica and the states of the ice sheet after forcing it for 10,000 years with cumulative emissions of (B) 500 GtC, (C) 1000 GtC, (D) 2500 GtC, (E) 5000 GtC, and (F) 10,000 GtC, simulated with the ice-sheet model PISM. Ice-free bedrock is shown in brown. For each scenario, the grounding line position after 100, 300, and 1000 years is shown in green, light green, and yellow, respectively. In the 1000 GtC scenario, both the West Antarctic Ice Sheet (red rectangular) and the Wilkes Basin (green rectangular) become unstable. For the 10,000 GtC scenario, the Antarctic continent is almost ice-free.

evolution of the Greenland Ice Sheet therefore critically depends on current and future cumulative CO_2 emissions and the associated surface warming (*31*). The surface elevation feedback is also at work in Antarctica, only at higher warming levels because the ice-sheet surface is overall colder than that of the Greenland Ice Sheet. Surface melting in PISM is parameterized using a positive degree day (PDD) scheme (*32*) (see Materials and Methods). Our results are rather robust with respect to the choice of model parameters that yield a maximum deviation of 9% in sea-level rise from the mean value for the 1250 GtC scenario, as shown by the error bars in Fig. 2 (see also fig. S4).

The Antarctic Ice Sheet is severely affected by high carbon emissions through both the marine ice-sheet instability and surface elevation feedbacks. On the time scale of millennia, large parts of the ice sheet melt or drain into the ocean, raising global sea level by several tens of meters. Most of the ice loss occurs within the first millennium, leading to high rates of sea-level rise during this period (Fig. 3; for more details, see also fig. S6). Our simulations show that cumulative emissions of 500 GtC commit us to long-term sea-level rise from Antarctica of 1.15 m within the next millenium, which is consistent with the sensitivity of 1.2 m/°C derived with a different ice-sheet model (*33*, *34*). Paleo data suggest that similar rates of sea-level rise have occurred during past warm periods (*35*). If the 2°C target, corresponding to about 600 GtC of additional carbon release compared to year 2010, were attained, the millennial sea-level rise from Antarctica could likely be restricted to 2 m. In our simulations, this would keep the ice sheet below the threshold for the collapse of the Wilkes Basin. However, if that threshold is crossed, the Antarctic ice cover is significantly reduced in thickness and area (Fig. 4). If we were to release all currently attainable fossil fuel resources, Antarctica would become almost ice-free. It is unclear whether this dynamic discharge would be reversible and, if so, on which time scales.

Ice loss from Antarctica is sensitive to ocean temperatures around the ice shelves and, to a lesser extent, air temperatures over Antarctica. Estimates of emissions consistent with temperature increases are sensitive to the assumed climate sensitivity (*36*), which is about 3.1°C for the model configuration of GENIE used here. Higher climate sensitivity would cause these temperature thresholds to be reached at lower levels of fossil fuel use, and lower climate sensitivity would cause them to be reached with higher levels of use.

PISM captures the large-scale ice dynamics on long-time scales central to this study. There are possible effects that are not captured by the model, such as the influence of surface melt water on the basal conditions of the ice as well as the role of ice defects such as fractures. The basal sliding is parameterized. Its interaction with the ice dynamics and thermodynamics remains a challenge for glaciology and may not be captured well by the model. Furthermore, in our simulations, the ice sheet's feedback onto the oceanic and atmospheric conditions is limited to topographic changes, which is a clear caveat of the approach but allows the integration of the model on the considered long time scales.

While the resolution and process limitations mean that the ice-sheet model applied might have some deficiencies in representing the fast ice stream motion relevant for the short-term response of an ice sheet to external perturbation, it is well suited for the long-term projections of the continental-scale evolution of the ice sheet discussed here. Our results show that the currently attainable carbon fuel resources are sufficient to eliminate the Antarctic Ice Sheet and that large parts of the ice sheet are threatened at much lower amounts of cumulative emission. The successive sea-level rise far exceeds all other possible contributions from thermal expansion or ice loss from mountain glaciers or the Greenland Ice Sheet. Thus, if emissions of fossil-fuel carbon result in warming substantially beyond the 2°C target, millennial-scale rates of sea-level rise are likely to be dominated by ice loss from Antarctica. With unrestrained future CO_2 emissions, the amount of sea-level rise from Antarctica could exceed tens of meters over the next 1000 years and could ultimately lead to the loss of the entire ice sheet.

MATERIALS AND METHODS

Atmospheric CO_2 and climate scenarios

We estimate the long-term evolution of the concentration of CO_2 in the atmosphere as well as global mean temperature pathways using the

"GENIE" Earth system model (*16*, *17*). GENIE is based on an efficient, three-dimensional dynamic ocean with a simplified atmospheric energy balance climate model. It incorporates a representation of global carbon cycling that includes geochemical interactions with marine sediments (*17*) and a temperature-only representation of the weathering terrestrial carbonate and silicate rocks on land but excludes the terrestrial biosphere. This configuration of GENIE exhibits an observed (1994) anthropogenic CO_2 inventory close to observations and a millennial-scale CO_2 response comparable to other Earth system models (*16*). Following a 10,000-year spin-up, we drive GENIE with emission scenarios based on a logistic equation (*15*), releasing between 93 to 12,000 GtC into the atmosphere after year 2010.

Temperature downscaling

The ratios between the global mean temperature and the near-surface air temperature and ocean temperature at about 400-m depth in the region south of 66°S are almost constant on long time scales, as shown in a 6000-year simulation with the coupled climate model ECHAM5/MPIOM (*21*). They approach values of 1.8 and 0.7, respectively. The time evolution of these ratios is used to scale the global mean temperature pathways from GENIE to obtain near-surface air and ocean temperature anomalies to uniformly drive the ice-sheet simulations with PISM.

Ice sheet model

Simulations of the Antarctic Ice Sheet are carried out with the PISM (*11*–*13*), stable version 0.5, on a 15-km rectangular grid. PISM is based on a hybrid shallow approximation of ice flow, ensuring a smooth transition from the vertical shear–dominated flow in the interior of the ice sheet to the fast-flowing ice shelves. Both the grounding line and the calving front are simulated at subgrid scale and evolve according to the physical boundary conditions. Grounding-line motion is reversible and shown to be consistent with full-Stokes simulations for higher resolutions (*14*). On the time scales considered here, the bedrock deforms with changing load from the ice. This effect is included, following refs. Lingler *et. al* and Bueler *et. al* (*37*, *38*).

Surface mass balance

The Antarctic surface temperature is parameterized according to the ERA Interim data (*39*), as a function of latitude and surface elevation. Recent results from ice core data as well as regional and global climate models suggest that the increase in accumulation can be well approximated by assuming a linear relation to the temperature anomaly, with factors between 5 and 7% per degree of warming (*23*). Surface melting and runoff are computed through a PDD scheme, with melt coefficients between 3 and 8 mm/PDD (*32*). The sensitivity of our results to both the PDD and the accumulation coefficient is shown in fig. S4. The resulting changes in ice loss deviate by less than 9% from the values presented in the paper.

Sub-shelf melting

Sub-shelf melt rates are computed on the basis of the temperature and salinity data from the BRIOS model (*40*), solving a three-equation system for the boundary layer beneath the ice shelf. Southern Ocean temperature anomalies are uniformly applied to the BRIOS temperature field, resulting in increased sub-shelf melting. The respective basal melt sensitivity, that is, the increase in sub-shelf melting with respect to the temperature anomaly, is within the observed range of 7 to 16 m year^{-1}°C^{-1} (*24*, *25*).

SUPPLEMENTARY MATERIALS

Supplementary material for this article is available at http://advances.sciencemag.org/cgi/content/full/1/8/e1500589/DC1

Fig. S1. Surface and ocean warming in Antarctica.
Fig. S2. Sea-level change within the next century.
Fig. S3. Sub-shelf melt sensitivity.
Fig. S4. Sensitivity to climate parameters.
Fig. S5. Sea-level commitment from Antarctic ice loss.
Fig. S6. Rate of sea-level rise.

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