

# The multimillennial sea-level commitment of global warming

Anders Levermann<sup>a,b,1</sup>, Peter U. Clark<sup>c</sup>, Ben Marzeion<sup>d</sup>, Glenn A. Milne<sup>e</sup>, David Pollard<sup>f</sup>, Valentina Radic<sup>g</sup>, and Alexander Robinson<sup>h,i</sup>

<sup>a</sup>Potsdam Institute for Climate Impact Research, 14473 Potsdam, Germany; <sup>b</sup>Institute of Physics, Potsdam University, 14476 Potsdam, Germany; <sup>c</sup>College of Earth, Ocean, and Atmospheric Sciences, Oregon State University, Corvallis, OR 97331; <sup>d</sup>Center for Climate and Cryosphere, Institute for Meteorology and Geophysics, University of Innsbruck, 6020 Innsbruck, Austria; <sup>e</sup>Department of Earth Sciences, University of Ottawa, Ottawa, ON, Canada K1N 6N5; <sup>f</sup>Earth and Environmental Systems Institute, Pennsylvania State University, University Park, PA 16802; <sup>g</sup>University of British Columbia, Vancouver, BC, Canada V6T 1Z4; <sup>h</sup>Universidad Complutense de Madrid, 28040 Madrid, Spain; and <sup>i</sup>Instituto de Geociencias, Universidad Complutense de Madrid-Consejo Superior de Investigaciones Científicas, 28040 Madrid, Spain

Edited by John C. Moore, College of Global Change and Earth System Science, Beijing, China, and accepted by the Editorial Board June 13, 2013 (received for review November 7, 2012)

Global mean sea level has been steadily rising over the last century, is projected to increase by the end of this century, and will continue to rise beyond the year 2100 unless the current global mean temperature trend is reversed. Inertia in the climate and global carbon system, however, causes the global mean temperature to decline slowly even after greenhouse gas emissions have ceased, raising the question of how much sea-level commitment is expected for different levels of global mean temperature increase above preindustrial levels. Although sea-level rise over the last century has been dominated by ocean warming and loss of glaciers, the sensitivity suggested from records of past sea levels indicates important contributions should also be expected from the Greenland and Antarctic Ice Sheets. Uncertainties in the paleo-reconstructions, however, necessitate additional strategies to better constrain the sea-level commitment. Here we combine paleo-evidence with simulations from physical models to estimate the future sea-level commitment on a multimillennial time scale and compute associated regional sea-level patterns. Oceanic thermal expansion and the Antarctic Ice Sheet contribute quasi-linearly, with  $0.4 \text{ m } ^\circ\text{C}^{-1}$  and  $1.2 \text{ m } ^\circ\text{C}^{-1}$  of warming, respectively. The saturation of the contribution from glaciers is overcompensated by the nonlinear response of the Greenland Ice Sheet. As a consequence we are committed to a sea-level rise of approximately  $2.3 \text{ m } ^\circ\text{C}^{-1}$  within the next 2,000 y. Considering the lifetime of anthropogenic greenhouse gases, this imposes the need for fundamental adaptation strategies on multicentennial time scales.

climate change | climate impacts | sea-level change

Sea-level projections show a robust, albeit highly uncertain, increase by the end of this century (1, 2), and there is strong evidence that sea level will continue to rise beyond the year 2100 unless the current global mean temperature trend is reversed (3–6). At the same time, inertia in the climate and global carbon system causes the global mean temperature to decline slowly even after greenhouse gas emissions have ceased (6), raising the question of how much sea-level rise we are committed to on a multimillennial time scale for different levels of global mean temperature increase. During the 20th century, sea level rose by approximately 0.2 m (7, 8), and it is estimated to rise by significantly less than 2 m by 2100, even for the strongest scenarios considered (9). At the same time, past climate records suggest a sea-level sensitivity of as much as several meters per degree of warming during previous intervals of Earth history when global temperatures were similar to or warmer than present (10, 11). Although sea-level rise over the last century has been dominated by ocean warming and loss of glaciers (7), the sensitivity suggested from records of past sea level indicates important contributions from the Greenland and Antarctic Ice Sheets. Because of the uncertainties in the paleo-reconstructions, however, additional strategies are required to better constrain the sea-

level commitment. Here we describe the models used and the resulting estimates of long-term sea-level rise from each component of the Earth system. We combine simulations from process-based physical models for the four main components that contribute to sea-level changes to give a robust estimate of the sea-level commitment on multimillennial time scales up to a global mean temperature increase of  $4^\circ\text{C}$ . Our results are then compared with paleo-evidence, with the good agreement providing an independent validation of our modeling results.

## Modeled Components of Sea Level

**Thermal Expansion.** The thermal expansion of the ocean has been investigated by a spectrum of climate models of different complexity, ranging from zero-dimensional diffusion models (12, 13) via Earth System Models of Intermediate Complexity (EMIC) (6, 14) to comprehensive general circulation models (15, 16). Although uncertainty remains, especially owing to uncertainty in the ocean circulation and thereby the distribution of heat within the ocean, the physical processes are relatively well understood even if not fully represented in all models. On multimillennial time scales as applied here, the application of comprehensive climate models is not feasible because of the required computational effort. Because the general processes responsible for oceanic expansion are, however, also integrated into lower resolution ocean models as used in EMICs, the range of long-term thermal expansion is likely to be covered by these models.

We take the thermal expansion of the ocean on multimillennial time scales from 10,000-y integrations with six coupled climate models. The results, which were used in the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (figure 10.34 in ref. 1), yield a rate of sea-level change in the range of  $0.20\text{--}0.63 \text{ m } ^\circ\text{C}^{-1}$  (Fig. 14). For reference, a homogeneous increase of ocean temperature by  $1^\circ\text{C}$  would yield a global mean sea-level rise of 0.38 m when added to reanalysis data (17). Uncertainty arises owing to the different spatial distribution of the warming in models and the dependence of the expansion on local temperature and salinity.

**Glaciers.** A number of different approaches have been used to estimate the contribution from glaciers to global sea level for the

Author contributions: A.L. designed research; A.L., P.U.C., B.M., G.A.M., D.P., V.R., and A.R. performed research; B.M., G.A.M., D.P., V.R., and A.R. contributed new reagents/analytic tools; and A.L. and P.U.C. wrote the paper.

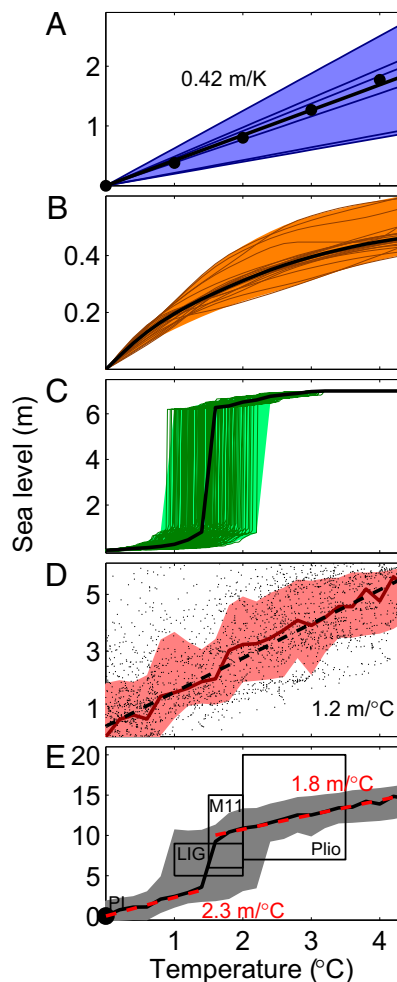
The authors declare no conflict of interest.

This article is a PNAS Direct Submission. J.C.M. is a guest editor invited by the Editorial Board.

Freely available online through the PNAS open access option.

See Commentary on page 13699.

<sup>1</sup>To whom correspondence should be addressed. E-mail: Anders.levermann@pik-potsdam.de.



**Fig. 1.** Sea-level commitment per degree of warming as obtained from physical model simulations of (A) ocean warming, (B) mountain glaciers and ice caps, and (C) the Greenland and (D) the Antarctic Ice Sheets. (E) The corresponding total sea-level commitment, which is consistent with paleo-estimates from past warm periods (PI, preindustrial, Plio, mid-Pliocene; see text for discussion). Temperatures are relative to preindustrial. Dashed lines and large dots provide linear approximations: (A) sea-level rise for a spatially homogeneous increase in ocean temperature; (A, D, E) constant slopes of 0.42, 1.2, and 1.8 and 2.3  $\text{m}/^\circ\text{C}$ . Shading as well as boxes represent the uncertainty range as discussed in the text. (A–C) Thin lines provide the individual simulation results from different models (A and B) or different parameter combinations (C). The small black dots in D represent 1,000-y averages of the 5-million-year simulation of Antarctica following ref. 36.

21st century, including extrapolations of either current rates of mass loss or current acceleration of rates of mass loss (18), estimates based on either constant or constantly declining accumulation area ratios in the future (19), and estimates based on explicit modeling of future glacier mass balances (20, 21). Although these approaches differ to a certain extent in the outcome for the 21st century, the total sea-level contribution from glaciers is limited by their total ice volume, which is small compared with that of the ice sheets on Greenland and Antarctica. Glaciers thus have a minor role on multimillennial time scales compared with the large contribution from oceanic expansion and the ice sheets. To our knowledge, there are no published global estimates of glacier contribution to sea-level change on time scales longer than a few centuries, or of the equilibrium response of glaciers to climate change.

The total possible contribution of glaciers (i.e., all of the land ice excluding the ice sheets) is limited to  $\sim 0.6$  m (22). We use two models (20, 21) to compute the long-term contribution for different levels of global mean temperature. Both models couple surface mass balance with simplified ice-dynamics models and are forced by temperature and precipitation scenarios for each glacier in the world. Radić and Hock's model (20) is forced by modified monthly temperature and precipitation series for 2001–2300 from four general circulation models (GCMs) of the coupled model intercomparison project CMIP-3 (U.K.MO-HadCM3, ECHAM5/MPI-OM, GFDL-CM2.0, and CSIRO-Mk3.0), using the A1B emission scenario, whereas the model by Marzeion et al. (21) is forced by temperature and precipitation anomalies from 15 GCMs of CMIP-5, using the representative concentration pathway (RCP)-8.5.

To obtain an estimate of the sea-level contribution on long time scales, temperature and precipitation patterns were kept constant at different levels of global warming. The uncertainty range is obtained as the model spread across the 19 different climate forcing and glacier model combinations. The resulting sensitivity of sea-level commitment decreases for increasing temperatures from  $0.21 \text{ m } ^\circ\text{C}^{-1}$  at preindustrial temperature levels to  $0.04 \text{ m } ^\circ\text{C}^{-1}$  at  $4^\circ\text{C}$  of warming (Fig. 1B). The decline in sensitivity with higher temperatures is to a large extent explained by loss of low-lying glacier surface area and to a lesser extent by increasing precipitation adding mass to high-elevation glaciers. Although glaciers and thermal expansion have contributed approximately equally to the sea-level increase of the last 40 y (7), the sea-level commitment from glaciers is relatively small compared with thermal expansion.

**The Greenland Ice Sheet.** Although it remains a challenge to simulate rapid ice discharge from the Greenland Ice Sheet in response to oceanic forcing (23), these fast ice fluxes are not crucial for a multimillennial estimate as attempted here. On a time scale of tens of thousands of years, the Greenland Ice Sheet shows threshold behavior with respect to the surrounding atmospheric temperature (24–27). Because summer temperatures around the ice sheet's margins are warm enough to produce melt over a large area of the ice sheet, perturbations in the climate strongly affect the surface mass balance of the ice sheet. Although the associated changes are most likely not abrupt in a temporal sense, they self-amplify owing to positive feedbacks, particularly that between surface elevation and temperature (28).

For the multimillennial contribution of the Greenland Ice Sheet, we apply the recent results from an ensemble of simulations from a regional energy–moisture balance climate model coupled to an ice-sheet model that accounts for the positive feedback between temperature and surface elevation (25). The model's parameters were constrained by comparison with surface mass balance estimates and topographical data for the present day and with estimated summit-elevation changes from ice-core records for the Last Interglacial period (LIG) (29), to ensure that the coupled model ensemble has a realistic sensitivity to climatic changes. In the transient response to global warming, simulated ice-sheet melting is comparable in timing and distribution to that of a GCM coupled to an ice sheet (24, 30).

The contribution to sea-level commitment from the Greenland Ice Sheet is relatively weak (on average  $0.18 \text{ m } ^\circ\text{C}^{-1}$  up to  $1^\circ\text{C}$  and  $0.34 \text{ m } ^\circ\text{C}^{-1}$  between 2 and  $4^\circ\text{C}$ ) apart from the abrupt threshold of ice loss between 0.8 and  $2.2^\circ\text{C}$  above preindustrial (90% credible interval) (Fig. 1C). This corresponds to a transition from a fully ice-covered Greenland to an essentially ice-free state (i.e., a reduction in ice volume of approximately 10% of the present-day volume, corresponding to a sea-level contribution of more than 6 m). Compared with previous studies (24, 30) the

model applied here shows a threshold at lower temperatures than would correspond to a negative surface mass balance of the entire ice sheet because of dynamic ice motion. This explains the lower threshold temperatures compared with earlier studies. However, the uncertainty range plotted here shows a significant overlap with these earlier estimates.

**The Antarctic Ice Sheet.** The total volume of the Antarctic Ice Sheet is equivalent to 55–60 m of global mean sea level (31). Most of the ice loss on Antarctica is due to solid-ice discharge, and thus the representation of the ice flow and sliding is particularly important. For continental-scale models, as required for estimates of the total sea-level contribution, a realistic representation of the grounding line position is a challenge at horizontal resolutions that allow for a simulation of whole Antarctica (32–34). The model used here applies a parameterization of the ice flux at the grounding line (35) that captures grounding line motion realistically on multimillennial time scales as suggested by comparison of the model's simulation of the last 5 million years (36) with local sediment records (37). Here we extract the sensitivity of the ice sheet from this model simulation by correlating the ice volume with the global mean temperature that forces the simulation. To this end we average the yearly temperature- and sea-level contribution data from the 5-million-year simulation over periods of 1,000 y (small dots in Fig. 1D). The sea-level contribution is then binned into temperature intervals of 0.2 °C. The median of the distribution within each temperature bin is provided as an estimate of the temperature sensitivity of the Antarctic ice sheet (thick red line in Fig. 1D). The uncertainty range is estimate as the 66% percentile of the distribution around the median (red shading). This uncertainty arises from uncertainty in the forcing data, the ice physics representation but also from the time-dependent nature of the simulation. For example, the existence of hysteresis behavior on the subcontinental scale can lead to different contributions for the same temperature increase.

The Antarctic Ice Sheet shows a relatively constant commitment of 1.2 m °C<sup>-1</sup> (dashed line in Fig. 1D). This mostly involves retreat of marine portions (grounded below sea level) of the ice sheet due to increased sub-ice shelf oceanic melt, predominantly in West Antarctica; it does not include future expansion of East Antarctic volume due to anthropogenic increases in snowfall, which is expected to be a minor effect on millennial scales (see below). Although threshold behavior can occur during marine retreat, the response shown is quite linear owing to temporal smoothing and the rapidly varying paleo-climatic forcing in the transient simulation. Future anthropogenic changes will occur on even faster time scales, so this approach is arguably more relevant than fully equilibrated sensitivity tests.

Similar to Greenland, the much larger East Antarctic Ice Sheet, which is predominantly grounded above sea level, has the potential of threshold behavior in response to increased surface melting (38). Current models (4, 5) suggest that the required warming is much larger than anticipated in the near future (equivalent to ~4 times preindustrial level of CO<sub>2</sub> or higher). Here we only discuss the commitment up to 4 °C of warming, neglecting the possible deglaciation of East Antarctica. However, there is an apparent conflict with geologic evidence of up to 20 m sea-level rise during the Miocene (39), which implies substantial contributions from East Antarctica at times when estimated CO<sub>2</sub> levels seem to have remained far below the model-derived thresholds. More work is needed to confidently rule out the possibility of East Antarctic threshold loss in the next few millennia.

## Paleo-Evidence

To compare the model results with past sea-level anomalies for the temperature range up to 4 °C, we focus on three previous

periods for which the geological record provides reasonable constraints on warmer climates and higher sea levels than pre-industrial: the middle Pliocene, marine isotope stage 11, and the LIG (Fig. 1E).

During the obliquity-paced warm intervals of the middle Pliocene (~3.3 to ~3 Ma), reconstructions of sea-surface temperatures (40) and climate model simulations (11, 41) suggest that peak global surface air temperatures were 1.8–3.5 °C warmer than preindustrial. Estimates of peak mid-Pliocene sea levels based on a variety of geological records are consistent in suggesting higher-than-present sea levels, but they range widely (5–40 m) and are each subject to large uncertainties. For example, coastal records (shorelines, continental margin sequences) are influenced by glacio-hydro-isostatic (42) or global mantle dynamic processes (43). Both signals are large (5–30 m) and uncertain, and significant differences in published predictions of dynamic topography suggest that the latter is particularly poorly constrained.

Benthic  $\delta^{18}\text{O}$  records are better dated than many coastal records and provide a continuous time series, but the  $\delta^{18}\text{O}$  signal reflects some unknown combination of ice volume and temperature, with some contribution from regional hydrographic variability also possible. During the mid-Pliocene warm intervals, the LR04 benthic  $\delta^{18}\text{O}$  stack (44) identifies low values that exceeded present values by 0.1–0.25 per mil. If the signal only records ice volume, then on the basis of the relationship of the  $\delta^{18}\text{O}$  of seawater to ice volume derived from pore water chemistry (~0.08 per mil per 10 m of sea-level equivalent) (45), and assuming this relationship still held during the Pliocene, the lowest  $\delta^{18}\text{O}$  values in the LR04 stack suggest peak sea levels ~12–31 m higher than present. Dowsett et al. (40), however, reconstructed Pliocene deepwater (>2,000 m) temperature anomalies at 20 sites that ranged from –0.9 °C to 4.2 °C, with an average of  $1 \pm 1.2$  °C (1 SD) warmer than present. A  $\delta^{18}\text{O}$ -temperature sensitivity of 0.28 per mil per °C suggests that, given these values, the lowest mid-Pliocene benthic  $\delta^{18}\text{O}$  values may be entirely explained by warmer deepwater temperature. Attempts to constrain the temperature component in benthic  $\delta^{18}\text{O}$  records indicate higher-than-present sea level during mid-Pliocene warm periods (46, 47), but these have large uncertainties ( $\pm 15$ –25 m) (48).

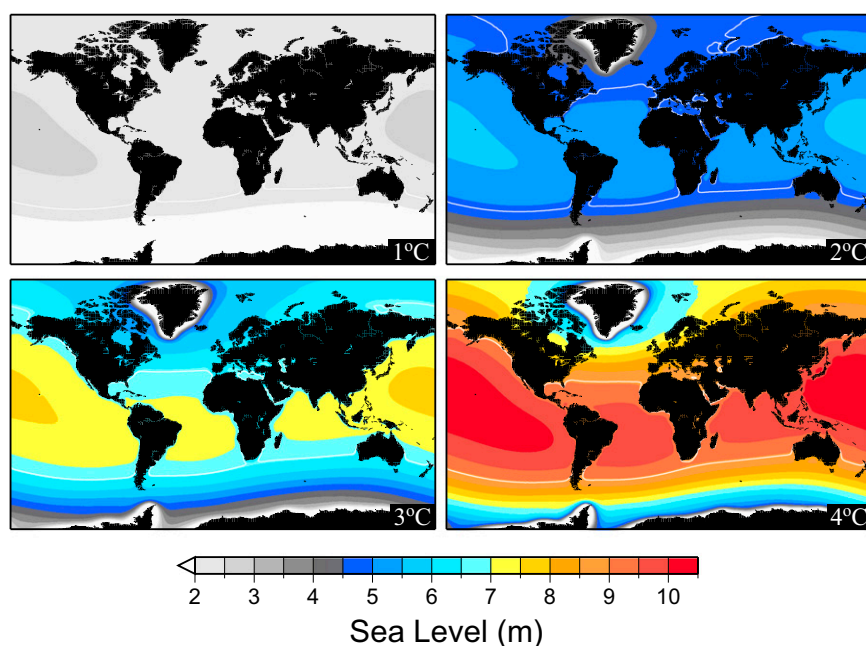
Records of ice-rafted debris off Prydz Bay, East Antarctica (49) and the sedimentary record from the Ross Embayment of the West Antarctic Ice Sheet (37) provide the most direct evidence that mid-Pliocene sea levels must have been higher than present. This is consistent with mid-Pliocene Ross Sea surface temperatures reconstructed from biological and geochemical proxy methods, which have a range from 2 to 8 °C warmer than present (50), with values of >5 °C being, according to one ice-sheet model, above the stability threshold for ice shelves and marine portions of the West Antarctic Ice Sheet (36). Finally, ice-sheet models forced by warm-interval mid-Pliocene temperature reconstructions and orbital forcing suggest near-complete deglaciation of Greenland (7 m sea-level equivalent) (51, 52).

We conclude that the balance of evidence indicates higher sea levels during the mid-Pliocene warm intervals, with the most robust lines of evidence coming from sedimentary records that suggest episodic deglaciation of the West Antarctic Ice Sheet and from ice-sheet models that suggest near-complete deglaciation of the Greenland Ice Sheet under warm Pliocene boundary conditions. We thus conclude that sea level was at least 7 m above present during mid-Pliocene warm periods, while, allowing for uncertainties in less direct sea-level proxies, did not exceed 20 m above present (39).

During marine isotope stage 11 (MIS 11; ~411–401 ka), Antarctic ice-core and tropical Pacific paleo-temperature estimates suggest that the global temperature was 1.5–2.0 °C warmer than preindustrial (53). Published studies of the magnitude of emergent shorelines attributed to MIS 11 have generated highly







**Fig. 3.** Regional patterns of sea-level change computed using an isostatic surface loading model (see text for details) for scenarios of 1, 2, 3, and 4 °C of warming. These results are based on the assumption that ice thinning was uniform over the ice sheets and progressed linearly for 2,000 y. The contribution of glaciers and ice caps and ocean thermal expansion to the spatial pattern is not included; however, their global mean sea-level contribution is included (Table 1). These results are based on a spherically symmetric Earth model with 1D, depth-dependent viscosity structure. This structure comprises an elastic outer layer of thickness 96 km (to simulate the lithosphere) and two sublithosphere layers with viscosity values  $5 \times 10^{20}$  Pas (96 km to 660 km depth) and  $10^{22}$  Pas (660 km to 2,900 km). These parameters are compatible with those inferred in a number of studies (e.g. refs 72 and 73). The white contour in each frame denotes the global mean sea-level value (Table 1).

after 2,000 y (Fig. 2C). The results are quantitatively consistent with previous estimates on a millennial time scale (3).

Mass loss from land ice results in a spatially variable sea-level change due to the resulting isostatic deformation and changes in gravity (63, 64). Using a model that simulates these processes (65–67), we computed global patterns of sea-level change associated with mass loss from the Greenland and Antarctic Ice Sheets based on the volume contributions in Table 1 (Fig. 3). Information about the spatial distribution of mass loss from within each ice sheet is not known, and so the model runs assumed a uniform thinning of ice over a 2,000-y period with no margin retreat. Although this simplification will affect the accuracy of the results (68, 69), the general characteristics of the pattern will not be affected. Over millennial time scales, the dominant component of solid Earth deformation is nonelastic, and so we used a viscoelastic Maxwell rheology (70). The patterns will be sensitive to the adopted Earth viscosity profile (Fig. 3 legend) and the time span of the simulation (68).

We note that the spatial contributions from oceanic warming and glaciers and ice caps are not considered in Fig. 3 (only the global mean of these contributions is shown). Compared with the global mean values (white contours in Fig. 3), the key characteristic of the patterns is a fall near the ice margins, a reduced sea-level rise (relative to the global mean) at intermediate to high latitudes, and an enhanced rise at intermediate to low latitudes. Note that the 1 °C scenario includes only a small amount of melting from the Greenland ice sheet (Table 1) and so sea-level rise in the Northern Hemisphere is relatively uniform.

## Discussion and Conclusion

Although the ice-sheet and climate models used here are not necessarily applicable to the fast time scales required to project

sea-level rise of this century, their ability to simulate the long-term signal has been validated against paleo-evidence on longer time scales and is consistent with results from other models on these time scales. In particular, model simulations providing the separate contributions from ocean warming, glaciers, and the Greenland and Antarctic Ice Sheets are consistent with paleo-estimates of total sea-level rise in warmer periods. Together with their underlying physical theories, this provides confidence in our understanding of long-term sea-level commitments. It should be noted that the transient response within the next century, especially of the Antarctic Ice Sheet, might be different even qualitatively, owing to increased snowfall in a warming environment (61), which may induce dynamic responses of the ice sheet (71). Consistent with paleo-records, however, Antarctica will contribute to global sea-level on a multimillennial time scale.

Although the Antarctic Ice Sheet and the thermal expansion of ocean water contribute quasi-linearly, with approximately  $1.2 \text{ m } ^\circ\text{C}^{-1}$  and  $0.4 \text{ m } ^\circ\text{C}^{-1}$ , respectively, Greenland shows a threshold behavior with a stepwise increase of approximately 6 m on a time scale on the order of several ten thousand years. After 2,000 y the model shows a superlinear response to the temperature increase. The contribution of glaciers declines at higher temperatures owing to the limited and decreasing ice stored there. As a consequence, the total sea-level commitment after 2,000 y is quasi-linear, with a sensitivity of  $2.3 \text{ m } ^\circ\text{C}^{-1}$ .

**ACKNOWLEDGMENTS.** This study was supported by the German Federal Ministry of Education and Research. P.U.C. and D.P. receive support from National Science Foundation Grants 1043517 and 1043018, respectively. B.M. was supported by Austrian Science Fund (FWF): P25362-N26. G.A.M. acknowledges support from the Canada Research Chairs program. A.R. was supported by the Gobierno de España, Ministerio de Economía y Competitividad under project CGL2011-29672-C02-01.

1. Solomon S, et al. (2007) *Climate Change 2007: The Physical Science Basis* (Cambridge Univ Press, Cambridge, UK).
2. Rahmstorf S (2007) A semi-empirical approach to projecting future sea-level rise. *Science* 315(5810):368–370.
3. Huybrechts P, et al. (2011) Response of the Greenland and Antarctic Ice Sheets to multi-millennial greenhouse warming in the Earth system model of intermediate complexity LOVECLIM. *Surv Geophys* 32(4):397–416.
4. Goelzer H, et al. (2012) Millennial total sea-level commitments projected with the Earth system model of intermediate complexity LOVECLIM. *Environ Res Lett* 7(4):045401.
5. Vizcaino M, Mikolajewicz U, Jungclaus J, Schurgers G (2010) Climate modification by future ice sheet changes and consequences for ice sheet mass balance. *Clim Dyn* 34:301–324.
6. Solomon S, Plattner GK, Knutti R, Friedlingstein P (2009) Irreversible climate change due to carbon dioxide emissions. *Proc Natl Acad Sci USA* 106(6):1704–1709.
7. Church JA, et al. (2011) Revisiting the Earth's sea-level and energy budgets from 1961 to 2008. *Geophys Res Lett* 38:L18601.
8. Church JA, White NJ (2006) A 20th century acceleration in global sea-level rise. *Geophys Res Lett* 33:L01602.
9. Pfeffer WT, Harper JT, O'Neal S (2008) Kinematic constraints on glacier contributions to 21st-century sea-level rise. *Science* 321(5894):1340–1343.
10. Kopp RE, Simons FJ, Mitrovica JX, Maloof AC, Oppenheimer M (2009) Probabilistic assessment of sea level during the last interglacial stage. *Nature* 462(7275):863–867.
11. Lunt D, et al. (2010) Earth system sensitivity inferred from Pliocene modelling and data. *Nat Geosci* 3(1):60–64.
12. Williams R, Goodwin P, Ridgwell A, Woodworth P (2012) How warming and steric sea level rise relate to cumulative carbon emissions. *Geophys Res Lett* 39:L19715.
13. Winkelmann R, Levermann A (2013) Linear response functions to project contributions to future sea level. *Clim Dyn* 40(11–12):2579–2588.
14. Petoukhov V, et al. (2005) EMIC Intercomparison Project (EMIP-CO2): Comparative analysis of EMIC simulations of climate, and of equilibrium and transient responses to atmospheric CO2 doubling. *Clim Dyn* 25(4):363–385.
15. Pardaens AK, Lowe JA, Brown S, Nicholls RJ, de Gusmão D (2011) Sea-level rise and impacts projections under a future scenario with large greenhouse gas emission reductions. *Geophys Res Lett* 38(12):L12604.
16. Yin J (2012) Century to multi-century sea level rise projections from CMIP5 models. *Geophys Res Lett* 39:L17709.
17. Levitus S, et al. (2009) Global ocean heat content 1955–2008 in light of recently revealed instrumentation problems. *Geophys Res Lett* 36:L07608.
18. Meier MF, et al. (2007) Glaciers dominate eustatic sea-level rise in the 21st century. *Science* 317(5841):1064–1067.
19. Bahr DB, Dyurgerov M, Meier MF (2009) Sea-level rise from glaciers and ice caps: A lower bound. *Grl* 36:3501.
20. Radić V, Hock R (2011) Regionally differentiated contribution of mountain glaciers and ice caps to future sea-level rise. *Nat Geosci* 4(2):91–94.
21. Marzeion B, Jarosch AH, Hofer M (2012) Past and future sea-level change from the surface mass balance of glaciers. *The Cryosphere* 6(6):1295–1322.
22. Radić V, Hock R (2010) Regional and global volumes of glaciers derived from statistical upscaling of glacier inventory data. *J Geophys Res Earth Surface* 115(F1):F01010.
23. Price SF, Payne AJ, Howat IM, Smith BE (2011) Committed sea-level rise for the next century from Greenland ice sheet dynamics during the past decade. *Proc Natl Acad Sci USA* 108(22):8978–8983.
24. Ridley J, Gregory JM, Huybrechts P, Lowe J (2010) Thresholds for irreversible decline of the Greenland ice sheet. *Clim Dyn* 35(6):1065–1073.
25. Robinson A, Calov R, Ganopolski A (2012) Multistability and critical thresholds of the Greenland ice sheet. *Nat Climate Change* 2(6):429–432.
26. Huybrechts P, Letreguilly A, Reeh N (1991) The (Greenland) ice sheet and greenhouse warming. *Palaeogeogr Palaeoclimatol Palaeoecol* 89:399–412.
27. Letreguilly A, Huybrechts P, Reeh N (1991) Steady-state characteristics of the Greenland Ice-Sheet under different climates. *J Glaciol* 37(125):149–157.
28. Levermann A, et al. (2012) Potential climatic transitions with profound impact on Europe Review of the current state of six 'tipping elements of the climate system'. *Clim Change* 110(3–4):845–878.
29. Robinson A, Calov R, Ganopolski A (2011) Greenland ice sheet model parameters constrained using simulations of the Eemian Interglacial. *Climate of the Past* 7(2):381–396.
30. Ridley J, Huybrechts P, Gregory JM, Lowe JA (2005) Elimination of the Greenland Ice Sheet in a high CO2 climate. *J Clim* 18(17):3409–3427.
31. Le Brocq AM, Payne AJ, Vieli A (2010) An improved Antarctic dataset for high resolution numerical ice sheet models (ALBMAP v1). *Earth Syst Sci Data* 2(2):247–260.
32. Pattyn F, et al. (2012) Results of the Marine Ice Sheet Model Intercomparison Project, MISMP. *Cryosphere* 6:267–308.
33. Pattyn F, et al. (2013) Grounding-line migration in plan-view marine ice-sheet models: Results of the ice2sea MISMP3d intercomparison. *J Glaciol* 59(215):410–422.
34. Gomez N, Mitrovica JX, Huybers P, Clark PU (2010) Sea level as a stabilizing factor for marine-ice-sheet grounding lines. *Nat Geosci* 3(1):850–853.
35. Schoof C (2007) Ice sheet grounding line dynamics: Steady states, stability, and hysteresis. *J Geophys Res* 112(F3):F03528.
36. Pollard D, DeConto RM (2009) Modelling West Antarctic ice sheet growth and collapse through the past five million years. *Nature* 458(7236):329–332.
37. Naish T, et al. (2009) Obliquity-paced Pliocene West Antarctic ice sheet oscillations. *Nature* 458(7236):322–328.
38. Langebroek P, Paul A, Schulz M (2009) Antarctic ice-sheet response to atmospheric CO2 and insolation in the Middle Miocene. *Climate Past* 5(4):633–646.
39. Foster GL, Rohling EJ (2013) Relationship between sea level and climate forcing by CO2 on geological timescales. *Proc Natl Acad Sci USA* 110(4):1209–1214.
40. Dowsett H, Robinson M, Foley K (2009) Pliocene three-dimensional global ocean temperature reconstruction. *Climate of the Past* 5(4):769–783.
41. Haywood AM, et al. (2013) Large-scale features of Pliocene climate: Results from the Pliocene Model Intercomparison Project. *Clim. Past* 9(1):191–209.
42. Raymo ME, Mitrovica JX, O'Leary MJ, DeConto RM, Hearty PL (2011) Departures from eustasy in Pliocene sea-level records. *Nat Geosci* 4(5):328–332.
43. Moucha R, et al. (2008) Dynamic topography and long-term sea-level variations: There is no such thing as a stable continental platform. *Earth Planet Sci Lett* 271(1):101–108.
44. Lisiecki LE, Raymo ME (2005) A Pliocene-Pleistocene stack of 57 globally distributed benthic delta O-18 records. *Paleoceanography* 20(1):PA1003.
45. Schrag D (2002) Control of atmospheric CO2 and climate through Earth history. *Geochim Cosmochim Acta* 66A688.
46. Dwyer G, Chandler M (2009) Mid-Pliocene sea level and continental ice volume based on coupled benthic Mg/Ca palaeotemperatures and oxygen isotopes. *Phil Trans R Soc A* 367(1886):157–168.
47. Sosdian S, Rosenthal Y (2009) Deep-sea temperature and ice volume changes across the Pliocene-Pleistocene climate transitions. *Science* 325(5938):306–310.
48. Miller KG, et al. (2012) The high tide of the warm Pliocene: Implications of global sea level for Antarctic deglaciation. *Geology* 40(5):407–410.
49. Passchier S (2011) Linkages between East Antarctic Ice Sheet extent and Southern Ocean temperatures based on a Pliocene high-resolution record of ice-rafted debris off Prydz Bay, East Antarctica. *Paleoceanography*, 10.1029/2010PA002061.
50. McKay NP, Overpeck JT, Otto-Bliessner BL (2011) The role of ocean thermal expansion in Last Interglacial sea level rise. *Geophys Res Lett* 38:4–9.
51. Dolan A, et al. (2011) Sensitivity of Pliocene ice sheets to orbital forcing. *Palaeogeogr Palaeoclimatol Palaeoecol* 309(1–2):98–110.
52. Hill D, Dolan A, Haywood A, Hunter S, Stoll D (2010) Sensitivity of the Greenland Ice Sheet to Pliocene sea surface temperatures. *Stratigraphy* 7(2–3):111–121.
53. Masson-Delmotte V, et al. (2010) EPICA Dome C record of glacial and interglacial intensities. *Quat Sci Rev* 29(1–2):113–128.
54. Hearty P, Kindler P, Cheng H, Edwards R (1999) A +20 m middle Pleistocene sea-level highstand (Bermuda and the Bahamas) due to partial collapse of Antarctic ice. *Geology* 27(4):375–378.
55. Bowen D (2010) Sea level similar to 400 000 years ago (MIS 11): analogue for present and future sea-level? *Climate Past* 6(1):19–29.
56. Raymo ME, Mitrovica JX (2012) Collapse of polar ice sheets during the stage 11 interglacial. *Nature* 483(7390):453–456.
57. Muhs D, Pandolfi J, Simmons K, Schumann R (2012) Sea-level history of past interglacial periods from uranium-series dating of corals, Curacao, Leeward Antilles islands. *Quat Res* 78(2):157–169.
58. Dutton A, Lambeck K (2012) Ice volume and sea level during the last interglacial. *Science* 337(6091):216–219.
59. de Vernal A, Hillaire-Marcel C (2008) Natural variability of Greenland climate, vegetation, and ice volume during the past million years. *Science* 320(5883):1622–1625.
60. Turney C, Jones R (2010) Does the Agulhas Current amplify global temperatures during super-interglacials? *J Quaternary Sci* 25(6):839–843.
61. Uotila P, Lynch AH, Cassano JJ, Cullather RI (2007) Changes in Antarctic net precipitation in the 21st century based on Intergovernmental Panel on Climate Change (IPCC) model scenarios. *J Geophys Res D Atmospheres* 112(D10):10107.
62. Schaeffer M, Hare W, Rahmstorf S, Vermeer M (2012) Long-term sea-level rise implied by 1.5 degrees C and 2 degrees C warming levels. *Nat Climate Change* 2(12):867–870.
63. Clark JA, Lingle CS (1977) Future sea-level changes due to West Antarctic ice sheet fluctuations. *Nature* 269:206–209.
64. Clark J, Farrell W, Peltier W (1978) Global changes in post-glacial sea-level—numerical calculation. *Quat Res* 9(3):265–287.
65. Mitrovica JX, Milne GA (2003) On post-glacial sea level: I. general theory. *Geophys J Int* 154(2):253–267.
66. Kendall RA, Mitrovica JX, Milne GA (2005) On post-glacial sea level—II. Numerical formulation and comparative results on spherically symmetric models. *Geophys J Int* 161(3):679–706.
67. Mitrovica JX, Wahr J, Matsuyama I, Paulson A (2005) The rotational stability of an ice-age earth. *Geophys J Int* 161(2):491–506.
68. Gomez N, Mitrovica JX, Tamisiea ME, Clark PU (2010) A new projection of sea level change in response to collapse of marine sectors of the Antarctic Ice Sheet. *Geophys J Int* 180(2):623–634.
69. Mitrovica JX, et al. (2011) On the robustness of predictions of sea-level fingerprints. *Geophys J Int* 187(2):729–742.
70. Peltier WR (1974) Impulse response of a Maxwell Earth. *Rev Geophys* 12(4):649–669.
71. Winkelmann R, Levermann A, Martin M, Frieler K (2012) Snowfall increases future ice discharge from Antarctica. *Nature* 492:239–242.
72. Kaufmann G, Lambeck K (2000) Mantle dynamics, postglacial rebound and the radial viscosity profile. *Phys Earth Planet Inter* 121(3–4):301–324.
73. Mitrovica J, Forte A (2004) A new inference of mantle viscosity based upon joint inversion of convection and glacial isostatic adjustment data. *Earth Planet Sci Lett* 225(1–2):177–189.