Clim Dyn (2015) 45:2001–2011
DOI 10.1007/s00382-014-2451-7

Increasing temperature forcing reduces the Greenland Ice Sheet's response time scale

**Patrick J. Applegate · Byron R. Parizek ·
Robert E. Nicholas · Richard B. Alley · Klaus Keller**

Received: 3 July 2014 / Accepted: 13 December 2014 / Published online: 23 December 2014

Abstract Damages from sea level rise, as well as strategies to manage the associated risk, hinge critically on the time scale and eventual magnitude of sea level rise. Satellite observations and paleo-data suggest that the Greenland Ice Sheet (GIS) loses mass in response to increased temperatures, and may thus contribute substantially to sea level rise as anthropogenic climate change progresses. The time scale of GIS mass loss and sea level rise are deeply uncertain, and are often assumed to be constant. However, previous ice sheet modeling studies have shown that the time scale of GIS response likely decreases strongly with increasing temperature anomaly. Here, we map the relationship between temperature anomaly and the time scale of GIS response, by perturbing a calibrated, three-dimensional model of GIS behavior. Additional simulations with a profile, higher-order, ice sheet model yield time scales that are broadly consistent with those obtained using the three-dimensional model, and shed light on the feedbacks in the ice sheet system that cause the time scale shortening. Semi-empirical modeling studies that assume a constant time scale of sea level adjustment, and are calibrated to small preanthropogenic temperature and sea level changes, may underestimate future sea level rise. Our analysis suggests that the benefits of reducing greenhouse gas emissions, in terms of avoided sea level rise from the GIS, may be greatest if emissions reductions begin before large temperature increases have been realized. Reducing anthropogenic climate change may also allow more time for design and deployment of risk management strategies by slowing sea level contributions from the GIS.

Keywords Greenland ice sheet · Glaciology · Ice sheet modeling · Semi-empirical · Sea level

Electronic supplementary material The online version of this article (doi:10.1007/s00382-014-2451-7) contains supplementary material, which is available to authorized users.

P. J. Applegate (✉) · R. E. Nicholas · K. Keller
Earth and Environmental Systems Institute, The Pennsylvania State University, University Park, PA, USA
e-mail: patrick.applegate@psu.edu

B. R. Parizek
Mathematics and Geoscience, The Pennsylvania State University, DuBois, PA, USA

R. B. Alley · K. Keller
Department of Geosciences, The Pennsylvania State University, University Park, PA, USA

K. Keller
Department of Engineering and Public Policy, Carnegie Mellon University, Pittsburgh, PA, USA

1 Introduction

The future behavior of the ice sheets represents an important unknown in estimating future sea level rise. Total sea level rise includes contributions from the ice sheets, small glaciers, and thermosteric expansion of ocean water, as well as other, smaller, sources. However, the maximum contributions from the ice sheets dwarf those of other sources. Small glaciers outside of Greenland and Antarctica contain enough ice to raise global mean sea level by $\leq$0.5 m (e.g. Dyurgerov and Meier 2005; Radic and Hock 2010), and thermosteric expansion will likely contribute $\leq$0.55 m to global mean sea level rise by 2100 (Sriver et al. 2012). By comparison, the Greenland Ice Sheet would contribute ~7.3 m to global mean sea level rise if it were to melt completely (Bamber et al. 2001, 2013); maximum contributions from the West Antarctic and East Antarctic ice sheets are

~4.5 and ~53.3 m, respectively (Lythe and Vaughan 2001; Fretwell et al. 2013), again assuming total melting. The ice sheets' ability to make substantial contributions to sea level change is confirmed by past rapid sea level rises like meltwater pulse 1A, in which global mean sea level rose by >10 m over a few centuries at the end of the last glacial period (Deschamps et al. 2012; cf. Gregoire et al. 2012).

The uncertainty in future ice sheet behavior directly affects present-day adaptation decisions, as well as the estimated economic costs associated with sea level rise (e.g., Yohe et al. 1996; Sugiyama et al. 2008; Anthoff et al. 2010; Nicholls et al. 2011). Nicholls et al. (2008a) estimate that ~10^8 people, or >1 % of the world's population, live within 1 m of mean sea level. The exposure of people and property in coastal cities to flooding risk will likely grow over time (Nicholls et al. 2008b; Hallegatte et al. 2013). Possible measures for adapting to future sea level rise include building seawalls, raising buildings above the expected height of flood waters, and moving people and infrastructure away from the coast. However, these measures are costly, and they compete with other funding priorities. In a recent analysis, whether or not hardening the Port of Los Angeles' facilities against sea level rise passes a cost-benefit test is strongly sensitive to ice sheet sea level contributions over the next century (Lempert et al. 2012).

Here, we focus on the Greenland Ice Sheet (GIS), which is (1) large enough to make a substantial contribution to future sea level, and (2) vulnerable to atmospheric warming. Of the terms in the sea level budget, the GIS has the second-largest maximum contribution after the relatively stable East Antarctic Ice Sheet (Sugden et al. 1993; cf. Mengel and Levermann 2014). A substantial fraction of the GIS' present-day mass loss is accomplished through direct melting (Alley et al. 2010; Robinson et al. 2011; Bamber et al. 2013). Thus, the GIS likely responds to Arctic air temperatures, which may lead the rise in global mean temperatures (Manabe and Stouffer 1980).

Total sea level rise, including contributions from the Greenland Ice Sheet, is often conceptualized as an asymptotic relaxation to a new equilibrium level that is a function of temperature change (e.g. Rahmstorf 2007; Grinsted et al. 2010). Such semi-empirical models are calibrated using observations of sea level and temperature covering the last few centuries (e.g. Rahmstorf 2007), sometimes extended to the last ~2,000 years using proxy information (e.g. Grinsted et al. 2010; Kemp et al. 2011). The calibrated models are then driven into the future using projected global mean temperatures (Rahmstorf 2007; Vermeer and Rahmstorf 2009; Grinsted et al. 2010) or radiative forcing changes (Jevrejeva et al. 2012). Although semi-empirical models have a number of limitations (e.g. Church et al. 2013), the papers describing these models are widely cited.

Despite its importance, the time scale of Greenland Ice Sheet mass loss in response to temperature increase remains deeply uncertain. Semi-empirical models typically assume that the time scale of total sea level adjustment is ≫100 years (e.g. Rahmstorf 2007; Rahmstorf et al. 2012). In principle, semi-empirical models could be built up of several terms, each of which would represent a different component of sea level rise; however, the relatively short record of temperature and sea level change does not allow confident separation of the different contributions (Grinsted et al. 2010). If this separation could be performed, however, the time scale of the Greenland component would presumably be longer than that of faster-acting components such as thermosteric expansion and glacier melt. Some studies (e.g. Vermeer and Rahmstorf 2009; Kemp et al. 2011) divide sea level response to temperature change into fast and slow components, but these formulations do not resolve ice sheet contributions explicitly. Similarly, a recent review of the literature (Lenton et al. 2008) suggests that achieving a "largely ice-free state" on Greenland would require >300 years. They further characterize this response as a "slow" transition (although loss of the ice sheet in 300 years implies a mean rate of sea level change greater than 2 m/century).

Thus, many semi-empirical modeling studies assume or imply that the time scale of Greenland Ice Sheet contribution to sea level rise is long and does not change as the forcing temperature increases. However, the ice sheet system contains positive feedbacks. For example, a temporary increase in surface melt within the ablation zone leads to additional and accelerating melt (Born and Nisancioglu 2012), because melting reduces surface elevation and temperatures are greater at lower elevations. These changes in surface mass balance may then cause dynamical changes in the ice sheet, further contributing to mass loss (Huybrechts and de Wolde 1999; Parizek and Alley 2004). Taken together, these nonlinearities may cause the time scale of GIS response to decrease as the forcing temperature rises (e.g. Fyke 2011; Robinson et al. 2012). This possibility was explicitly anticipated by the first semi-empirical modeling studies (Rahmstorf 2007; Grinsted et al. 2010).

Estimating the time scale of GIS response from computer models of ice sheet behavior is complicated by structural differences among models and uncertainties in model parameter values. Ice sheet models fall along a continuum with respect to their treatment of ice dynamics (e.g. Kirchner et al. 2011). Shallow-ice approximation models neglect selected stress components within the ice body in exchange for computational efficiency; full-Stokes models include a complete description of the physics of ice flow, but require more computing resources than do simpler models. Both classes of models suffer from large uncertainties in boundary conditions and poorly-understood processes (see Sect. 4 below). Different models often yield divergent estimates of future ice volume change for the same temperature forcing trajectory, particularly past the first century (e.g. Bindschadler et al. 2013; Nowicki et al. 2013). Perturbed-parameter ensembles (e.g. Ritz et al. 1996; Stone et al. 2010; Applegate et al. 2012) indicate that model parameter choice strongly affects the modeled ice sheet's response to temperature change. In particular, Applegate et al. (2012) noted that parameter combinations that match the observed ice sheet similarly well sometimes give widely divergent ice volume change projections. Improved methods for calibrating ice sheet models may reduce the divergence among projections from different models (Chang et al. 2014; see also Shannon et al. 2013; Edwards et al. 2014).

Previous ice sheet modeling studies that have noted a shortening of the time scale of GIS response with increased temperature anomaly include Fyke (2011) and Robinson et al. (2012). Fyke (2011) used an ice sheet model coupled to a simplified climate model to evaluate the response of the Greenland Ice Sheet under abrupt, sustained atmospheric carbon dioxide concentration increases, and specifically diagnosed the time scale of ice sheet response from the resulting ice volume curves. Robinson et al. (2012) investigated the equilibration properties of the GIS under abrupt, sustained temperature increases using a perturbed-parameter ensemble of an ice sheet model coupled to a regional energy and moisture balance model. Although Robinson et al. (2012) noted that the time scale of ice sheet response became shorter with increased temperature, they did not quantify this change in time scale.

Here, we quantify the time scale (specifically, the *e*-folding time) of contributions to sea level rise from the Greenland Ice Sheet, using two different ice sheet models and accounting for model parametric uncertainties. Our results suggest that this time scale is strongly dependent on the forcing temperature. As described above, these results contrast with the assumptions of semi-empirical modeling studies, but are in line with previous ice sheet modeling studies (Fyke 2011; Robinson et al. 2012).

2 Methods

2.1 Estimating the e-folding time τ from ice sheet model runs

Specifically, we estimate the *e*-folding time of Greenland Ice Sheet response to temperature change τ, and the equilibrium contribution of the ice sheet to sea level rise ΔV, for a range of Greenland temperature increases ΔT_{grl}. We conceptualize GIS response to increased temperature as a relaxation to a new equilibrium (Fyke 2011; Fig. 1a),

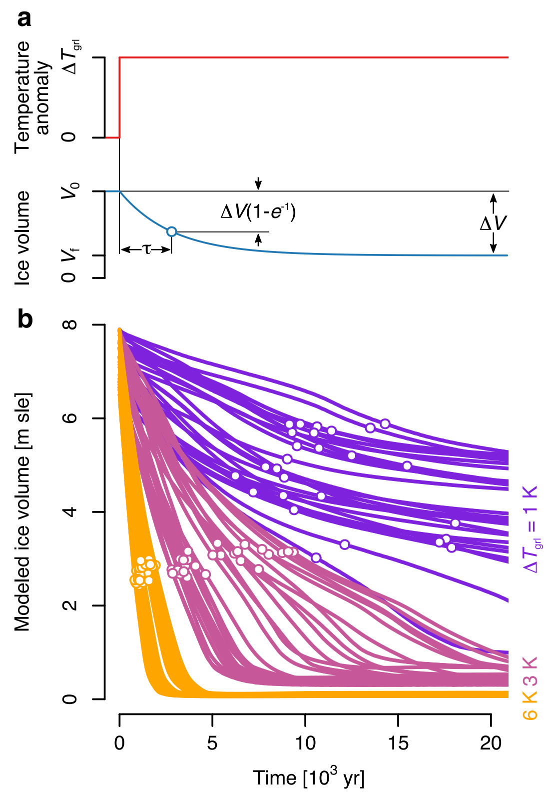

Fig. 1 **a** Conceptual model of Greenland Ice Sheet volume response to an instantaneous Greenland temperature change ΔT_{grl} (Eq. 1). **b** Modeled ice volume responses to selected Greenland temperature anomalies using the three-dimensional ice sheet model SICOPOLIS (Greve 1997; Greve et al. 2011; sicopolis.greveweb.net). The modeled $V(t)$ *curves* shown in **b** generally agree with the conceptual model shown in **a**. In **b**, differences among curves in each color group reflect model parametric uncertainty (Applegate et al. 2012). *Open circles* indicate the *e*-folding time τ, diagnosed as the time when each individual curve reaches $V_0 - \Delta V(1 - e^{-1})$ (Eq. 2). We investigated eight ΔT_{grl} values in total (Electronic Supplementary Material Figure 1). Compare to Fyke (2011, his Fig. 4.2) and Robinson et al. (2012, their Fig. 3b). m sle, meters of sea level equivalent

$$
\begin{aligned}
V(t) &= \Delta V e^{(-t/\tau)} + V_f \\
\Delta V &= V_0 - V_f,
\end{aligned}
\tag{1}
$$

where $V(t)$ is ice volume as a function of time, V_0 is the ice volume just before the temperature increase, and V_f is the final ice volume. Several existing ice sheet modeling studies present results from experiments in which temperatures over the ice sheet are instantaneously increased; curves of ice volume as a function of time from such experiments resemble an asymptotic adjustment to a new equilibrium (e.g. Letréguilly et al. 1991, their Fig. 10; Fyke 2011, his Fig. 4.2; Robinson et al. 2012, their Fig. 3b). Thus, ice sheet model simulations suggest that the Greenland Ice Sheet adjusts asymptotically to temperature change, consistent with Eq. 1.

Given a single ice sheet model run in which near-surface air temperatures suddenly increase by an amount ΔT_{grl}, we estimate the *e*-folding time τ by identifying the point along the $V(t)$ curve where the ice sheet's volume has declined by $1 - e^{-1}$ (about two-thirds) of its equilibrium change ΔV, relative to its starting volume V_0. Substituting for t and ΔV in Eq. 1 and rearranging, we obtain

$$V(t = \tau) = V_0 - \Delta V\left(1 - e^{-1}\right). \quad (2)$$

This definition of the time scale is consistent with that used in semi-empirical studies (e.g. Grinsted et al. 2010; Jevrejeva et al. 2012).

2.2 Ice sheet model simulations

To obtain these $V(t)$ curves, we performed a perturbed-parameter ensemble with the three-dimensional ice sheet model SICOPOLIS (Greve 1997; Greve et al. 2011; sicopolis.greveweb.net) and additional runs using a profile model of the Greenland Ice Sheet (Parizek and Alley 2004; Parizek et al. 2005).

SICOPOLIS is a shallow-ice approximation model, meaning that it achieves computational speed by neglecting selected stress components within the ice body (Kirchner et al. 2011). Speed of execution is key for estimating the time scale of Greenland Ice Sheet response while accounting for parametric uncertainty; the SICOPOLIS runs described here include a total of ~45 million model years (including spinup), well beyond what is practical with many higher-order ice sheet models.

The profile model describes the ice sheet's behavior along a transect across Greenland at ~72°N. Crucially, this model can be run with either a shallow-ice or a higher-order dynamical core, which accounts for more stress components than the shallow-ice core. The model is the same in both cases, except for the numerical solution method and some small differences in the surface mass balance scheme (see below). The shallow-ice dynamical core solves the diffusion formulation of continuity using the Galerkin method of weighted residuals (Parizek and Alley 2004), whereas the higher-order dynamical core solves the flux form of the continuity equation using the Petrov–Galerkin method of weighted residuals (Parizek et al. 2010) and solves the momentum balance on an adaptive mesh.

The two models use similar methods for calculating the ice sheet's surface mass balance. Precipitation increases exponentially with surface air temperature anomaly in both models. In SICOPOLIS, this increase in precipitation with temperature is ~7 %/K (Greve et al. 2011); in the profile model, it is somewhat smaller (~5 %/K; Parizek and Alley 2004). In SICOPOLIS, precipitation falls as liquid rain or solid snow depending on the monthly temperature. In contrast, the profile model assigns precipitation a "fate" depending on the presence or absence of snow and superimposed ice on the ice sheet surface and whether the surface is melting when the precipitation falls. Both models use the positive degree-day method for estimating ablation (e.g. Braithwaite 1995), with different melt factors for snow and exposed ice. The profile model specifically tracks the development of superimposed ice within the snowpack, and this superimposed ice must melt before ice sheet ice becomes available for melting. SICOPOLIS adds snowmelt and rainwater that freezes within the snowpack directly to the ice column in each grid cell. In both models, liquid water (either precipitation or meltwater) that cannot be accommodated in the snowpack is lost from the ice sheet. The profile model neglects the advection of snow when the higher-order dynamical core is in use; this process is included when the shallow-ice dynamical core is being used. These surface mass balance treatments are similar to, or more sophisticated than, those of many competing models (see Bindschadler et al. 2013, their Table 1).

The SICOPOLIS runs build on a recently-published, perturbed-parameter ensemble (Applegate et al. 2012). The existing ensemble (Applegate et al. 2012) includes 100 members, spun up from 125,000 years ago to the present. The spinup includes forcings from ice core-derived paleotemperatures and sea levels estimated from oxygen isotopes in ocean sediment cores, following the recommendations of the SeaRISE project (Bindschadler et al. 2013). Because the forcings applied to the simulations are time-dependent, the simulated modern ice sheets are not necessarily in equilibrium with late Holocene climates (e.g. Vinther et al. 2009). Twenty-seven ensemble members produced a simulated modern ice volume within 10 % of the estimated value (Bamber et al. 2001, 2013). We applied instantaneous, Greenland-specific, temperature increases $\Delta T_{grl} = 0$, 1, 2, 3, 4.5, 6, 9, and 12 K, relative to 1976–2005, to these "best" model runs. All model runs were equilibrated over ≥60,000 years.

The profile model was run for Greenland temperature changes $\Delta T_{grl} = 0$, 3, 4.5, 6, and 12 K, using both dynamical cores. The runs using the shallow-ice dynamical core were equilibrated over 10,000 years, whereas the simulations using the more-expensive higher-order dynamical core were run over 450–2,000 years.

As pointed out above, earlier studies (e.g. Huybrechts and de Wolde 1999; Parizek and Alley 2004; Born and Nisancioglu 2012) have identified both mass balance-related and dynamically-driven feedbacks in the ice sheet system, and these feedbacks may interact with one another in complex ways (e.g. Edwards et al. 2014 and references therein). To assess this possibility, we performed another set of runs with the profile model. In this set of experiments, ice flow and basal sliding were turned off, and ice thicknesses were allowed to change only through snowfall and melting (Huybrechts and de Wolde 1999). These runs were integrated over only 500 years, after which several of the modeled ice sheets began to grow upward indefinitely because snow continued to fall on the accumulation zone, whereas the ablation zone's width was substantially reduced.

2.3 Plausible future Greenland temperature changes

Both SICOPOLIS and the profile ice sheet model accept Greenland-specific temperature changes as input. However, it is unclear a priori how much temperatures might change over Greenland in the future, or what the relationship between Greenland temperature change and global mean temperature change is. To help answer these questions, we examined temperature changes in all climate model runs stored in the CMIP5 archive (http://cmip-pcmdi.llnl.gov/cmip5/) that follow the RCP8.5 radiative forcing trajectory (Meinshausen et al. 2011; Riahi et al. 2011). Eighy-eight such runs in the CMIP5 archive extend to the end of the century; 14 of these runs continue to 2300. We calculated Greenland-specific temperature anomalies by (1) interpolating all the climate model temperature fields to a consistent grid with nodes every 1°, (2) averaging over all grid cells that cover Greenland (both ice-covered and land), and (3) taking the differences between 30-year averages of temperature change over the grid representation of Greenland. The baseline period was chosen to be 1970–1999. Differences were calculated between this period and 2070–2099, as well as between the baseline period and 2270–2299. We also diagnosed the ratio of Greenland temperature change to global mean temperature change for both of these epochs.

3 Results

In our runs with the three-dimensional ice sheet model SICOPOLIS (Fig. 1b), the *e*-folding time of GIS response declines exponentially with increasing temperature forcing (Fig. 2a). For small temperature anomalies, this *e*-folding time is long (many thousands of years). However, large temperature anomalies ($\Delta T_{grl} \geq 12$ K) give short *e*-folding times of ~100 years. As expected, small temperature changes yield small ice sheet volume reductions (subject to large parametric uncertainties), whereas larger ones eventually remove the ice sheet (Robinson et al. 2012; Gregory and Huybrechts 2006; Fig. 2b).

The time scales inferred from the profile model runs (Fig. 3a) are similar to those from SICOPOLIS for $\Delta T_{grl} \geq 3$ K (Fig. 3b). Moreover, the $V(t)$ curves generated using the profile model's higher-order dynamical core agree closely with those from the shallow-ice dynamical core (Fig. 3a). The curves from the mass balance-only runs match those from the shallow-ice and higher-order dynamical cores reasonably well for large temperature forcings, but less well at lower ΔT_{grl} values (Fig. 3a).

In CMIP5 climate model runs that follow the RCP8.5 radiative forcing trajectory, near-surface air temperatures

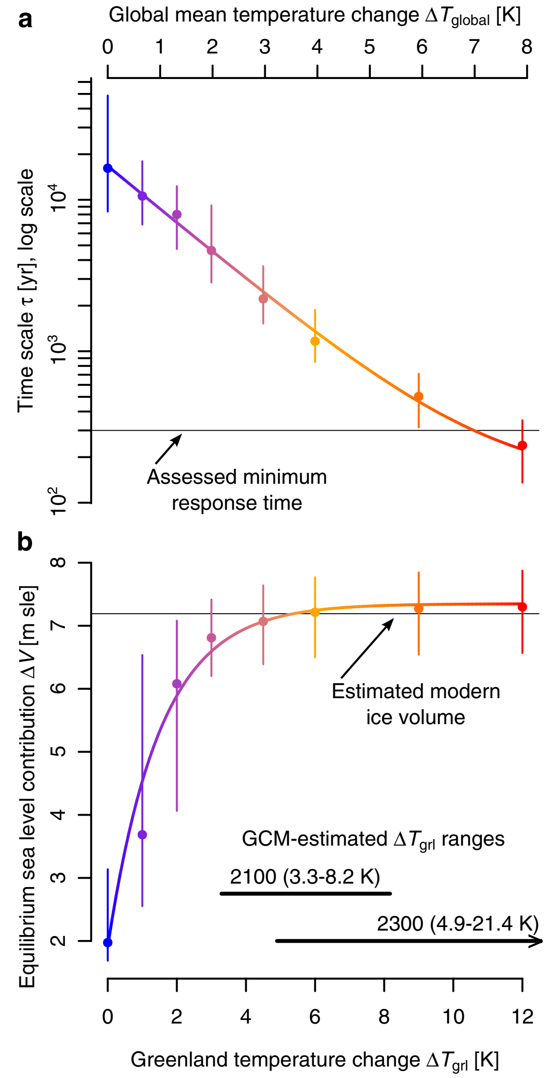

Fig. 2 **a** *e*-folding times of ice sheet response and **b**, equilibrium ice volume changes for different imposed temperature anomalies, as diagnosed from our modeled ice volume curves (Fig. 1b, ESM Fig. 1). The *e*-folding time of response is long (10^3–10^4 years) for small temperature increases, but becomes short (~100 years) for large forcings. *Dots*, median of the 27 model runs associated with each temperature change (Fig. 1b, ESM Fig. 1); *vertical lines*, 95 % ranges of model runs associated with each temperature change; *multicolored curves*, decaying exponentials fitted to the median values. The GCM-estimated ΔT_{grl} ranges, and the global mean temperature change axis, are based on climate model runs from the CMIP5 archive (http://cmip-pcmdi.llnl.gov/cmip5/); see Fig. 4 for details. Estimated modern ice volume (~7.3 m sle) and minimum time for GIS loss (300 years) from Bamber et al. (2001, 2013) and Lenton et al. (2008), respectively. m sle, meters of sea level equivalent

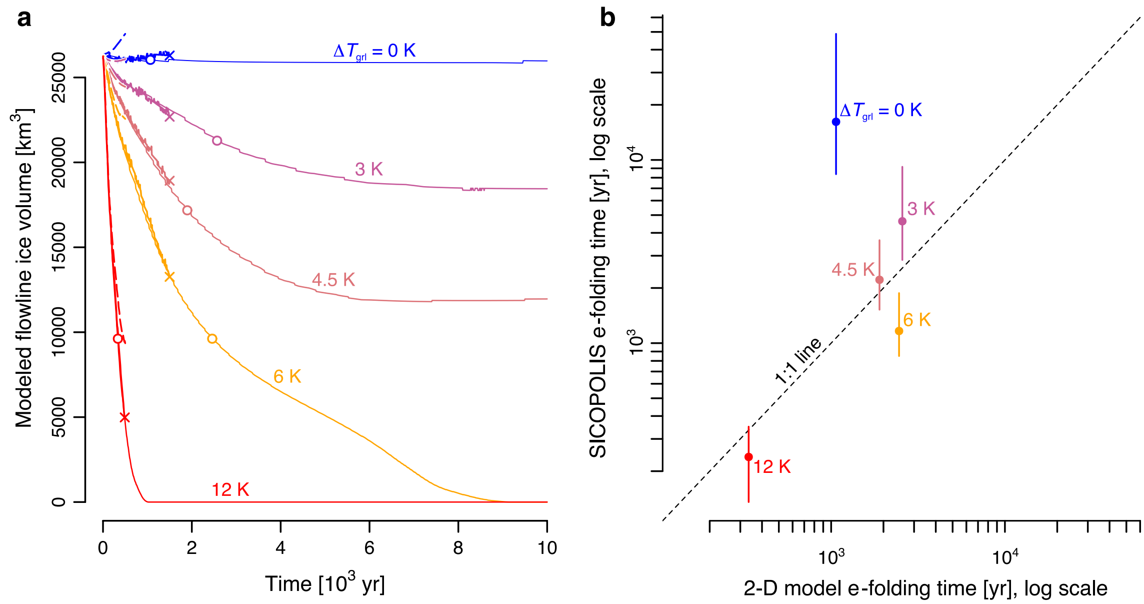

Fig. 3 **a** Modeled ice volume responses to selected Greenland temperature anomalies using a profile model (Parizek and Alley 2004; Parizek et al. 2005, 2010), and **b** a comparison of the *e*-folding times estimated using this model and the three-dimensional ice sheet model SICOPOLIS (Greve 1997; Greve et al. 2011; sicopolis.greveweb.net). In **a**, *thick, solid lines* indicate model results using the profile model's higher-order dynamical core; *thin, solid lines* indicate model results using the profile model's computationally-cheaper shallow-ice approximation dynamical core; and *thin, dashed lines* indicate model results in which ice volume changes only in response to surface mass balance (that is, ice flow is "turned off"). *Small crosses* indicate the ends of the higher-order model runs. The ice volume curves from the shallow-ice and higher-order dynamical cores lie on top of one another, suggesting that the lack of higher-order physics in SICOPOLIS does not affect our overall conclusions. *Open circles* indicate the point on each curve corresponding to the *e*-folding time τ (Fig. 1), diagnosed as the time when each individual curve reaches $V_0 - \Delta V(1 - e^{-1})$ (Eq. 2). The agreement between the surface mass balance-only model runs and runs that incorporate dynamics improves with increased temperature forcing. In **b**, the *e*-folding times deduced from the two models generally lie close to the 1:1 line, suggesting that our diagnosed *e*-folding times are not sensitive to the particular model used to derive them. The one exception is for $\Delta T_{grl} = 0$ K, where the behavior of the two models is qualitatively different. Compare to Fig. 1

over Greenland increase 3.3–8.2 K by the late 21st century, whereas this range grows to 4.9–21.4 K by the late 23rd century (Fig. 4). For the late 21st century, the average ratio of Greenland temperature change to global mean temperature change among the different model runs is 1.52, with a range of 1.06–2.02 (Fig. 4; cf. Gregory and Huybrechts 2006; Frieler et al. 2012; Fyke et al. 2014). This ratio is somewhat lower in the late 23rd century (mean 1.41, with a range of 0.96–2.01). However, the subset of model runs that make up this average also produces a slightly-smaller mean Greenland amplification factor than the full ensemble in the late 21st century (mean 1.49, with a range of 1.19–2.02).

4 Discussion

Our results show that the time scale of Greenland Ice Sheet contribution to sea level rise changes over two orders of magnitude (Fig. 2a) for temperature increases that might be achieved over the next few centuries (Fig. 4). This result appears to persist despite uncertainties in model parameter choice (Figs. 1b, 2a) and differences in ice sheet model structure (Fig. 3b).

In particular, the good agreement between the shallow-ice and higher-order ice volume curves from the profile model (Fig. 3a) suggests that an appropriately-tuned, three-dimensional, higher-order ice sheet model would yield time scales similar to those we obtain using SICOPOLIS (Fig. 2a). The close agreement between the results from the shallow-ice and higher-order dynamical cores in the profile model (Fig. 3a) is due to the inclusion of processes that dominate Greenland Ice Sheet evolution along the modeled transect, namely surface mass balance and basal sliding (e.g., the inclusion or exclusion of a migrating zone of sliding activation with surface warming; Parizek and Alley 2004). These findings suggest that the simulation of key processes can be just as important as the inclusion of higher-order stress terms in ice sheet models (e.g. Parizek et al. 2010, 2013; Christianson et al. 2013), and that simplified models are crucial for answering targeted (often process-oriented) questions.

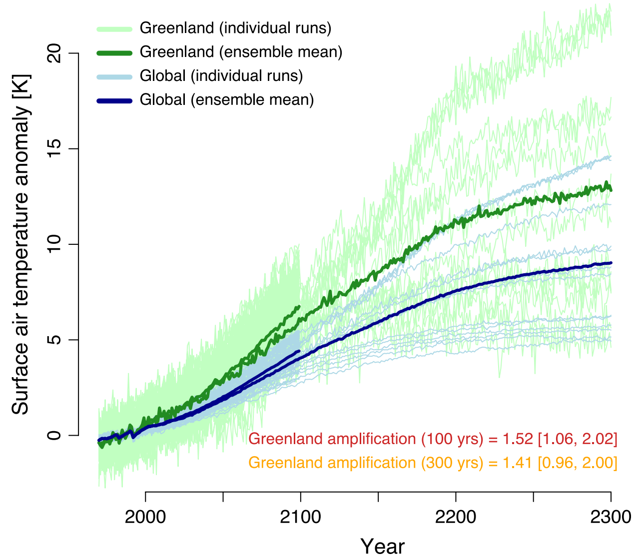

Fig. 4 Global mean (*blue*) and Greenland-specific (*green*) temperature anomaly trajectories derived from climate model runs following the RCP8.5 emissions scenario from the CMIP5 archive (http://cmip-pcmdi.llnl.gov/cmip5/). Eighty-eight runs extend to 2100; 14 of these runs continue to 2300. *Dark blue* and *green lines* indicate the mean of the model runs. Note that the 14 runs that extend to 2300 give a generally lower average than the full ensemble over the period of overlap (1950–2100). The amplification factors are calculated for the epochs 2070–2099 and 2270–2099; the *numbers outside the brackets* indicate the values for the ensemble-average curves, whereas the *numbers in the brackets* indicate the smallest and largest values from the ensemble for each epoch. The results from this analysis were used to generate the global mean temperature axis and the *black bars* in Fig. 2 of the main text

Our ice sheet model experiments represent crudely, or neglect, many processes that are important on the real ice sheet. In particular, both of the models we apply here use the positive degree-day method for calculating surface melt. Previous studies have shown that the positive degree-day method has shortcomings relative to more-sophisticated melt calculation schemes (e.g. Braithwaite 1995; van de Wal 1996; Bougamont et al. 2007; Robinson et al. 2010; van de Berg et al. 2011). We have not explored the effects of albedo feedbacks (Robinson et al. 2012), or changes in the distribution of temperature or precipitation over the year, on our model output. However, the ranges of the positive degree-day factors that we investigated in the SICOPOLIS ensemble are quite large (Applegate et al. 2012), suggesting that we have adequately explored possible variations in melt over the Greenland Ice Sheet's surface. Other processes that we parameterize or neglect include surface meltwater-driven lubrication of the ice-bed interface (this process is implicitly included in the profile model simulations; Zwally et al. 2002; Parizek and Alley 2004; Bartholomew et al. 2010; Shannon et al. 2013) and the penetration of warm ocean waters into fjords, accelerating the drawdown of ice through outlet glaciers (Joughin et al. 2008; Straneo et al. 2010). Including these processes in our simulations would likely shorten our estimated *e*-folding times, rather than lengthen them (e.g. Parizek and Alley 2004).

4.1 Differences between SICOPOLIS and the profile model

Although the results from SICOPOLIS and the profile model are generally similar to one another, there are two important differences. SICOPOLIS suggests that the ice sheet will largely disappear for $\Delta T_{\mathrm{grl}} \geq 3$ K (Fig. 2b; see also Gregory and Huybrechts 2006; Robinson et al. 2012), whereas the profile model shows a more moderate decline in equilibrium ice volumes with temperature increase (Fig. 3a). Also, our SICOPOLIS runs suggest that the ice sheet's volume will slowly decline by ~2 m for no additional temperature forcing ($\Delta T_{\mathrm{grl}} = 0$ K), whereas the profile model gives a near-zero ice sheet volume change under similar conditions. Thus, SICOPOLIS generally estimates larger mass losses than the profile model.

These discrepancies in model behavior may result from differences in the two models' domains. SICOPOLIS covers the whole of Greenland, whereas the profile model treats a single west-east transect across Greenland at 72°N (Parizek and Alley 2004). In our SICOPOLIS ensemble, the ice sheet preferentially loses mass in the north, consistent with other studies (Born and Nisancioglu 2012; Nowicki et al. 2013). Given that the primary center of mass loss lies outside the profile model's domain, we expect the profile model to predict smaller ice losses than a three-dimensional model like SICOPOLIS. Our simulations are consistent with this expectation.

If the unforced volume decline from SICOPOLIS is correct, it may reflect continuing adjustment of the ice sheet to the Holocene warm period (Vinther et al. 2009) or post-Little Ice Age warming (Fyke et al. 2011). Extrapolation of the best-fit exponential curve in Fig. 2b to its intersection with the ΔT_{grl} axis yields $\Delta T_{\mathrm{grl}}(\Delta V = 0) \sim -0.5$ K, relative to the 1976–2005 Greenland average temperature. This result includes large and unquantified uncertainties, but is broadly consistent with other studies; Greenland temperature anomalies during the period 1970–1995 were about 0.5 K cooler than the 1976–2005 average (Vinther et al. 2006), and the ice sheet was likely in balance during this time (e.g., Rignot et al. 2008; Alley et al. 2010).

4.2 Diagnosing the reasons for time scale shortening with the profile model

The profile model allows us to identify the feedbacks that cause the observed time scale shortening. Inspection of the output fields from the profile model (not shown) indicates that small to moderate increases in surface temperatures result in a progressive and accelerating surface lowering near the ice sheet margins (Born and Nisancioglu 2012), which steepens ice surface slopes near the margin of the ice sheet. This steepening increases the driving stress, which speeds ice flow into the ablation zone and generates a wave of thinning that propagates toward the central parts of the ice sheet (Huybrechts and de Wolde 1999; Parizek and Alley 2004). Sufficiently high ΔT_{grl} values cause the ice sheet's surface mass balance to quickly become negative (Gregory and Huybrechts 2006; Robinson et al. 2012). Consistent with the conclusions of these earlier studies, the agreement between the ice volume curves from the mass balance-only runs and the dynamic runs becomes progressively better as ΔT_{grl} increases (Fig. 3b; see also Huybrechts and de Wolde 1999). This result suggests that ice dynamics are an important determinant of the time scale of GIS response for small temperature anomalies, but the ice sheet's response is dominated by surface mass balance changes at high ΔT_{grl} values.

4.3 Incorporating our results into semi-empirical modeling studies

Our results (Fig. 2) might be incorporated in semi-empirical modeling studies in the following preliminary way. First, total sea level rise would be broken into two components, one reflecting the GIS and another describing all other contributions. The GIS term would have variable $\tau(\Delta T_{\mathrm{grl}})$ and $\Delta V(\Delta T_{\mathrm{grl}})$, described by our best-fit decaying exponential functions (Fig. 2). In other words, the GIS component would have a new *e*-folding time and equilibrium ice volume change each year, based on the forcing time series $\Delta T(t)$. Next, the non-GIS term would be probabilistically matched to the sea level data, less the estimated past contributions from the GIS. Finally, the two-term model would be run into the future with an appropriate temperature forcing trajectory. This preliminary approach neglects uncertainties in the derived $\tau(\Delta T_{\mathrm{grl}})$ and $\Delta V(\Delta T_{\mathrm{grl}})$ curves and the scaling between global mean temperature increases and Greenland temperature increases (Fig. 4); however, these problems might be addressed by treating the parameters of the fit and the temperature scaling factor as uncertain parameters.

4.4 Implications for other studies

The variability in climate model-estimated Greenland temperature trajectories (Fig. 4) introduces additional uncertainty into studies that estimate future sea level contributions from the Greenland Ice Sheet using complex ice sheet models. Both paleo-data and theoretical studies suggest that the GIS changes size in response to temperature (see review in Alley et al. 2010). For a single emissions trajectory, IPCC-class climate models project Greenland temperature changes by 2100 that range from 3.3 to 8.2 K (Fig. 4). Although GIS mass loss is likely regardless of the particular temperature trajectory, the rate and magnitude of sea level contributions will differ, depending on how the climate system reacts to a given change in radiative forcing. Thus, ice sheet modeling studies that provide probabilistic estimates of future sea level contributions will likely need to sample a range of temperature trajectories for each emissions scenario, as well as a range of possible emissions scenarios.

It might be argued that the plausible range of temperature changes is smaller than the one we report, because some models in the CMIP5 ensemble show less skill over Greenland than others (Belleflamme et al. 2013; Fettweis et al. 2013) and thus yield temperature changes that are too high or low. However, the CMIP5 ensemble provides only a limited sampling of uncertainties associated with model parameter values and initial conditions, which can be substantial (e.g. Stainforth et al. 2005; Deser et al. 2012; Olson et al. 2013). Thus, the range of plausible Greenland temperature changes that we derive from the CMIP5 ensemble might be too narrow, rather than too wide. Climate model calibration (e.g., Bhat et al. 2012) could help to reduce the range of plausible future temperature increases; such a calibration is beyond the scope of the present study.

Semi-empirical modeling studies of sea level rise neglect the time scale shortening that we observe, and this neglect may lead such studies to underpredict future sea level rise (Fig. 5). As noted in the Introduction, such models assume that the time scale of sea level rise is constant and long, and they are tuned to data covering the last 2,000 years at most. Over this period, Northern Hemisphere temperature anomalies have not gone above 0.5 K, relative to 1961–1990, for

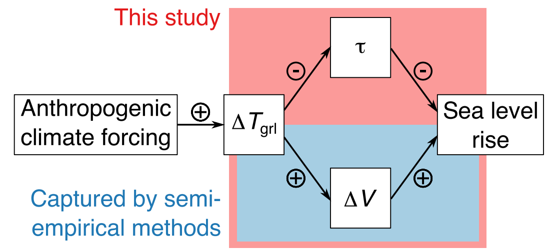

Fig. 5 Systems diagram (Kump et al. 2010) showing two ways in which anthropogenic climate forcing and concomitant Greenland temperature increase ΔT_{grl} drive sea level rise (cf. Fig. 2). Our study captures both the shortening of the GIS response time scale τ and the increase in equilibrium sea level contribution ΔV with temperature increase; semi-empirical modeling studies (e.g. Grinsted et al. 2010; Jevrejeva et al. 2012) neglect the shortening of the GIS response time scale τ identified by this work

any extended period of time (e.g. Moberg et al. 2005, their Fig. 2b). These small temperature anomalies imply that the GIS' response time scale has been long, and its equilibrium sea level contribution has been low, over the calibration period of semi-empirical models ($\tau \sim 10^4$ years, $\Delta V < 3$ m sle; Fig. 2b). In contrast, a Greenland temperature rise of 6 K over the 21st century appears plausible given climate model projections (Fig. 4). Such a temperature rise might result in an order-of-magnitude shortening of ice sheet response time in the near term, and loss of the ice sheet over the long term ($\tau \sim 10^3$ years, $\Delta V \sim 7.3$ m sle; Fig. 2b).

4.5 Policy implications

We speculate that near-term reductions in greenhouse gas emissions could pay large dividends in terms of avoided sea level rise. Our results suggest that the relationships between temperature change, GIS response time scale, and GIS equilibrium sea level contribution are approximately exponential (Fig. 2). Thus, the benefit, in terms of avoided sea level rise contributions from the GIS, of a unit of avoided emissions is greatest if emissions reductions are begun before much temperature change has already happened. Alternatively, one could say that mitigation becomes less effective in preventing or delaying sea level rise contributions from the Greenland Ice Sheet as temperature rises. Near-term reductions in greenhouse gas emissions may also buy time to design and implement improved strategies for adapting to sea level change.

Acknowledgments Jeremy Fyke, Aslak Grinsted, Ralf Greve, Ben Horton, Nina Kirchner, Johan Kleman, Chris Little, Michael Oppenheimer, Hans-Peter Plag, Dave Pollard, Alex Robinson, Kelsey Ruckert, and Gary Yohe provided discussions and comments on draft versions of this paper. We thank Ralf Greve for providing his ice sheet model, SICOPOLIS, freely on the Web (sicopolis.greveweb.net). Nina Kirchner, Philipp Hancke, Martin Jakobsson, and Björn Eriksson provided access to computing resources at Stockholm University, Sweden. We also thank the editor, Jian Lu, and two anonymous reviewers for their comments on the manuscript. This work was partly supported by the US Department of Energy, Office of Science, Biological and Environmental Research Program, Integrated Assessment Program, through grant DE-SC0005171; the National Science Foundation through the Network for Sustainable Climate Risk Mangement (SCRiM) under NSF cooperative agreement GEO-1240507; and the Penn State Center for Climate Risk Management. Support for BRP and RBA was provided by the Center for Remote Sensing of Ice Sheets (NSF OPP 0424589), NSF grant AGS 1338832, and NASA through Grant NNX10AI04G. We also acknowledge the climate modeling groups that participated in the Coupled Model Intercomparison Project Phase 5 (CMIP5; http://cmip-pcmdi.llnl.gov/cmip5/), which supplied the climate model output used in this paper; these groups are listed in Electronic Supplementary Materials Table 1. The US Department of Energy's Program for Climate Model Diagnosis and Intercomparison, in partnership with the Global Organization for Earth System Science Portals, provides coordinating support for CMIP5. Some figure colors were drawn from www.colorbrewer.org by Cynthia Brewer at Pennsylvania State University. All opinions and errors are ours.

References

Alley RB, Andrews JT, Brigham-Grette J, Clarke G, Cuffey K, Fitzpatrick J, Funder S, Marshall S, Miller G, Mitrovica J, Muhs DR, Otto-Bliesner BL, Polyak L, White JWC (2010) History of the Greenland Ice Sheet: paleoclimatic insights. Quatern Sci Rev 29:1728–1756

Anthoff D, Nicholls RJ, Tol RS (2010) The economic impact of substantial sea-level rise. Mitig Adapt Strat Glob Change 15:321–335

Applegate PJ, Kirchner N, Stone EJ, Keller K, Greve R (2012) An assessment of key model parametric uncertainties in projections of Greenland Ice Sheet behavior. Cryosphere 6:589–606

Bamber JL, Layberry RL, Gogineni S (2001) A new ice thickness and bed data set for the Greenland Ice Sheet: 1. Measurement, data reduction, and errors. J Geophys Res Atmos 106:33773–33780

Bamber J, Griggs J, Hurkmans R, Dowdeswell J, Gogineni S, Howat I, Mouginot J, Paden J, Palmer S, Rignot E, Steinhage D (2013) A new bed elevation dataset for Greenland. Cryosphere 7:499–510

Bartholomew I, Nienow P, Mair D, Hubbard A, King MA, Sole A (2010) Seasonal evolution of subglacial drainage and acceleration in a Greenland outlet glacier. Nat Geosci 3:408–411

Belleflamme A, Fettweis X, Lang C, Erpicum M (2013) Current and future atmospheric circulation at 500 hPa over Greenland simulated by the CMIP3 and CMIP5 global model. Clim Dyn 41:2061–2080

Bhat SK, Haran M, Olson R, Keller K (2012) Inferring likelihoods and climate system characteristics from climate models and multiple tracers. Environmetrics 23:345–362

Bindschadler RA, Nowicki S, Abe-Ouchi A, Aschwanden A, Choi H, Fastook J, Granzow G, Greve R, Gutowski G, Herzfeld U, Jackson C, Johnson J, Khroulev C, Levermann A, Lipscomb WH, Martin MA, Morlighem M, Parizek BR, Pollard D, Price SF, Ren D, Saito F, Sato T, Seddik H, Seroussi H, Takahashi K, Walker R, Wang WL (2013) Ice-sheet model sensitivities to environmental forcing and their use in projecting future sea level (the SeaRISE project). J Glaciol 59:195–224

Born A, Nisancioglu K (2012) Melting of Northern Greenland during the last interglaciation. Cryosphere 6:1239–1250

Bougamont M, Bamber JL, Ridley JK, Gladstone RM, Greuell W, Hanna E, Payne AJ, Rutt I (2007) Impact of model physics on estimating the surface mass balance of the Greenland Ice Sheet. Geophys Res Lett 34:L17501

Braithwaite RJ (1995) Positive degree-day factors for ablation on the Greenland Ice Sheet studied by energy-balance modelling. J Glaciol 41:153–160

Chang W, Applegate PJ, Haran M, Keller K (2014) Probabilistic calibration of a Greenland Ice Sheet model using spatially resolved synthetic observations: toward projections of ice mass loss with uncertainties. Geosci Model Dev 7:1933–1943

Christianson K, Parizek BR, Alley RB, Horgan HJ, Jacobel RW, Anandakrishnan S, Keisling BA, Craig BD, Muto A (2013) Ice sheet grounding zone stabilization due to till compaction. Geophys Res Lett 40:5406–5411

Church JA, Clark PU, Cazenave A, Gregory JM, Jevrejeva S, Levermann A, Merrifield MA, Milne GA, Nerem RS, Nunn PD, Payne AJ, Pfeffer WT, Stammer D, Unnikrishnan AS (2013) Sea level change. In Stocker TF, Qin D, Plattner G-K, Tignor M, Allen SK, Boschung J, Nauels A, Xia Y, Bex V and Midgley PM (eds) Climate change 2013: the physical science basis. Contribution of working group I to the fifth assessment report of the intergovernmental panel on climate change, Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA

Deschamps P, Durand N, Bard E, Hamelin B, Camoin G, Thomas AL, Henderson GM, Okuno J, Yokoyama Y (2012) Ice-sheet collapse

and sea-level rise at the Bolling warming 14,600 years ago. Nature 483:559–564
Deser C, Knutti R, Solomon S, Phillips AS (2012) Communication of the role of natural variability in future North American climate. Nat Clim Change 2:775–779
Dyurgerov MB, Meier MF (2005) Glaciers and the changing earth system: a 2004 snapshot. Institute of Arctic and Alpine Research Occasional Paper 58
Edwards TL, Fettweis X, Gagliardini O, Gillet-Chaulet F, Goelzer H, Gregory JM, Hoffman M, Huybrechts P, Payne AJ, Perego M, Price S, Quiquet A, Ritz C (2014) Effect of uncertainty in surface mass balance-elevation feedback on projections of the future sea level contribution of the Greenland Ice Sheet. Cryosphere 8:195–208
Fettweis X, Franco B, Tedesco M, van Angelen JH, Lenaerts JTM, van den Broeke MH, Gallee H (2013) Estimating the Greenland Ice Sheet surface mass balance contribution to future sea level rise using the regional atmospheric climate model MAR. Cryosphere 7:469–489
Fretwell P, Pritchard HD, Vaughan DG, Bamber J, Barrand N, Bell R, Bianchi C, Bingham R, Blankenship D, Casassa G, Catania G, Callens D, Conway H, Cook AJ, Corr HFJ, Damaske D, Damm V, Ferraccioli F, Forsberg R, Fujita S, Gim Y, Gogineni P, Griggs JA, Hindmarsh RCA, Holmlund P, Holt JW, Jacobel RW, Jenkins A, Jokat W, Jordan T, King EC, Kohler J, Krabill W, Riger-Kusk M, Langley KA, Leitchenkov G, Leuschen C, Luyendyk BP, Matsuoka K, Mouginot J, Nitsche FO, Nogi Y, Nost OA, Popov SV, Rignot E, Rippin DM, Rivera A, Roberts J, Ross N, Siegert MJ, Smith AM, Steinhage D, Studinger M, Sun B, Tinto BK, Welch BC, Wilson D, Young DA, Xiangbin C, Zirizzotti A (2013) Bedmap2: improved ice bed, surface and thickness datasets for Antarctica. Cryosphere 7:375–393
Frieler K, Meinshausen M, Mengel M, Braun N, Hare W (2012) A scaling approach to probabilistic assessment of regional climate change. J Clim 25:3117–3144
Fyke J (2011) Simulation of the global coupled climate/ice sheet system over millennial timescales. Unpublished Ph. D. dissertation, Victoria University of Wellington
Fyke J, Weaver A, Pollard D, Eby M, Carter L, Mackintosh A (2011) A new coupled ice sheet/climate model: description and sensitivity to model physics under Eemian, Last Glacial Maximum, late Holocene and modern climate conditions. Geosci Model Dev 4:117–136
Fyke J, Eby M, Mackintosh A, Weaver A (2014) Impact of climate sensitivity and polar amplification on projections of Greenland Ice Sheet loss. Clim Dyn 43:2249–2260
Gregoire LJ, Payne AJ, Valdes PJ (2012) Deglacial rapid sea level rises caused by ice-sheet saddle collapses. Nature 487:219–223
Gregory J, Huybrechts P (2006) Ice-sheet contributions to future sea-level change. Philos Trans R Soc A Math Phys Eng Sci 364:1709–1732
Greve R (1997) Application of a polythermal three-dimensional ice sheet model to the Greenland Ice Sheet: response to steady-state and transient climate scenarios. J Clim 10:901–918
Greve R, Saito F, Abe-Ouchi A (2011) Initial results of the SeaRISE numerical experiments with the models SICOPOLIS and IcIES for the Greenland Ice Sheet. Ann Glaciol 52:23–30
Grinsted A, Moore JC, Jevrejeva S (2010) Reconstructing sea level from paleo and projected temperatures 200 to 2100 AD. Clim Dyn 34:461–472
Hallegatte S, Green C, Nicholls RJ, Corfee-Morlot J (2013) Future flood losses in major coastal cities. Nat Clim Change 3:802–806
Huybrechts P, de Wolde J (1999) The dynamic response of the Greenland and Antarctic ice sheets to multiple-century climatic warming. J Clim 12:2169–2188
Jevrejeva S, Moore JC, Grinsted A (2012) Sea level projections to AD2500 with a new generation of climate change scenarios. Global Planet Change 80:14–20
Joughin I, Das SB, King MA, Smith BE, Howat IM, Moon T (2008) Seasonal speedup along the western flank of the Greenland Ice Sheet. Science 320:781–783
Kemp AC, Horton BP, Donnelly JP, Mann ME, Vermeer M, Rahmstorf S (2011) Climate related sea-level variations over the past two millennia. Proc Natl Acad Sci 108:11017–11022
Kirchner N, Hutter K, Jakobsson M, Gyllencreutz R (2011) Capabilities and limitations of numerical ice sheet models: a discussion for Earth-scientists and modelers. Quatern Sci Rev 30:3691–3704
Kump LR, Kasting JF, Crane RG (2010) The earth system, 3rd edn. Prentice-Hall, Englewood Cliffs
Lempert R, Sriver RL, Keller K (2012) Characterizing uncertain sea level rise projections to support investment decisions. California Energy Commission White Paper CEC-500-2012-056
Lenton TM, Held H, Kriegler E, Hall JW, Lucht W, Rahmstorf S, Schellnhuber HJ (2008) Tipping elements in the Earth's climate system. Proc Natl Acad Sci 105:1786–1793
Letréguilly A, Huybrechts P, Reeh N (1991) Steady-state characteristics of the Greenland Ice Sheet under different climates. J Glaciol 37:149–157
Lythe MB, Vaughan DG (2001) BEDMAP: a new ice thickness and subglacial topographic model of Antarctica. J Geophys Res 106:11335–11351
Manabe S, Stouffer RJ (1980) Sensitivity of a global climate model to an increase of CO_2 concentration in the atmosphere. J Geophys Res Oceans (1978–2012) 85:5529–5554
Meinshausen M, Smith SJ, Calvin K, Daniel JS, Kainuma M, Lamarque J, Matsumoto K, Montzka S, Raper S, Riahi K, Thomson A, Velders GJM, van Vuuren DPP (2011) The RCP greenhouse gas concentrations and their extensions from 1765 to 2300. Clim Change 109:213–241
Mengel M, Levermann A (2014) Ice plug prevents irreversible discharge from East Antarctica. Nat Clim Change 4:451–455
Moberg A, Sonechkin DM, Holmgren K, Datsenko NM, Karlén W (2005) Highly variable Northern Hemisphere temperatures reconstructed from low-and high-resolution proxy data. Nature 433:613–617
Nicholls RJ, Tol RS, Vafeidis AT (2008a) Global estimates of the impact of a collapse of the West Antarctic ice sheet: an application of FUND. Clim Change 91:171–191
Nicholls RJ, Hanson S, Herweijer C, Patmore N, Hallegatte S, Corfee-Morlot J, Chateau J, Muir-Wood R (2008b) Ranking port cities with high exposure and vulnerability to climate extremes: exposure estimates. OECD Environment working papers 1
Nicholls RJ, Marinova N, Lowe JA, Brown S, Vellinga P, de Gusmao D, Hinkel J, Tol RSJ (2011) Sea level rise and its possible impacts given a 'beyond 4 °C world' in the twenty-first century. Philos Trans R Soc A 369:161–181
Nowicki S, Bindschadler RA, Abe-Ouchi A, Aschwanden A, Bueler E, Choi H, Fastook J, Granzow G, Greve R, Gutowski G, Herzfeld U, Jackson C, Johnson J, Khroulev C, Larour E, Levermann A, Lipscomb WH, Martin MA, Morlighem M, Parizek BR, Pollard D, Price SF, Ren D, Rignot E, Saito F, Sato T, Seddik H, Seroussi H, Takahashi K, Walker R, Wang WL (2013) Insights into spatial sensitivities of ice mass response to environmental change from the SeaRISE ice sheet modeling project II: Greenland. J Geophys Res Earth Surf 118:1025–1044
Olson R, Sriver R, Haran M, Chang W, Urban NM, Keller K (2013) What is the effect of unresolved internal climate variability on climate sensitivity estimates? J Geophys Res 118:4348–4358
Parizek BR, Alley RB (2004) Implications of increased Greenland surface melt under global-warming scenarios: ice-sheet simulations. Quatern Sci Rev 23:1013–1027
Parizek BR, Alley RB, MacAyeal DR (2005) The PSU/UofC finite-element thermomechanical flowline model of ice-sheet evolution. Cold Reg Sci Technol 42:145–168

Parizek B, Alley R, Dupont T, Walker R, Anandakrishnan S (2010) Effect of orbital-scale climate cycling and meltwater drainage on ice sheet grounding line migration. J Geophys Res Earth Surf 115:F01011

Parizek B, Christianson K, Anandakrishnan S, Alley R, Walker R, Edwards R, Wolfe D, Bertini G, Rinehart S, Bindschadler R, Nowicki SMJ (2013) Dynamic (in)stability of Thwaites Glacier, West Antarctica. J Geophys Res Earth Surf 118:638–655

Radic V, Hock R (2010) Regional and global volumes of glaciers derived from statistical upscaling of glacier inventory data. J Geophys Res 115:F01010

Rahmstorf S (2007) A semi-empirical approach to projecting future sea-level rise. Science 315:368–370

Rahmstorf S, Perrette M, Vermeer M (2012) Testing the robustness of semi-empirical sea level projections. Clim Dyn 39:861–875

Riahi K, Rao S, Krey V, Cho C, Chirkov V, Fischer G, Kindermann G, Nakicenovic N, Rafaj P (2011) RCP 8.5—a scenario of comparatively high greenhouse gas emissions. Clim Change 109:33–57

Rignot E, Box JE, Burgess E, Hanna E (2008) Mass balance of the Greenland Ice Sheet from 1958 to 2007. Geophys Res Lett 35:L20502

Ritz C, Fabre A, Letréguilly A (1996) Sensitivity of a Greenland Ice Sheet model to ice flow and ablation parameters: consequences for the evolution through the last climatic cycle. Clim Dyn 13:11–23

Robinson A, Calov R, Ganopolski A (2010) An efficient regional energy–moisture balance model for simulation of the Greenland Ice Sheet response to climate change. Cryosphere 4:129–144

Robinson A, Calov R, Ganopolski A (2011) Greenland Ice Sheet model parameters constrained using simulations of the Eemian Interglacial. Clim Past 7:381–396

Robinson A, Calov R, Ganopolski A (2012) Multistability and critical thresholds of the Greenland Ice Sheet. Nat Clim Change 2:429–432

Shannon SR, Payne AJ, Bartholomew ID, van den Broeke MR, Edwards TL, Fettweis X, Gagliardini O, Gillet-Chaulet F, Goelzer H, Hoffman MJ, Huybrechts P, Mair DWF, Nienow PW, Perego M, Price SF, Paul Smeets CJP, Sole AJ, van de Wal RSW, Zwinger T (2013) Enhanced basal lubrication and the contribution of the Greenland Ice Sheet to future sea-level rise. Proc US Natl Acad Sci 110:14156–14161

Sriver RL, Urban NM, Olson R, Keller K (2012) Toward a physically plausible upper bound of sea-level rise projections. Clim Change 115:893–902

Stainforth DA, Aina T, Christensen C, Collins M, Faull N, Frame DJ, Kettleborough JA, Knight S, Martin A, Murphy JM, Piani C, Sexton D, Smith LA, Spicer RA, Thorpe AJ, Allen MR (2005) Uncertainty in predictions of the climate response to rising levels of greenhouse gases. Nature 433:403–406

Stone E, Lunt D, Rutt I, Hanna E (2010) Investigating the sensitivity of numerical model simulations of the modern state of the Greenland ice-sheet and its future response to climate change. Cryosphere 4(3):397–417

Straneo F, Hamilton GS, Sutherland DA, Stearns LA, Davidson F, Hammill MO, Stenson GB, Rosing-Asvid A (2010) Rapid circulation of warm subtropical waters in a major glacial fjord in East Greenland. Nat Geosci 3:182–186

Sugden DE, Marchant DR, Denton GH (1993) The case for a stable East Antarctic ice sheet: the background. Geogr Ann Ser A Phys Geogr 75:151–154

Sugiyama M, Nicholls RJ, Vafeidis A (2008) Estimating the economic cost of sea-level rise. MIT joint program on the science and policy of global change report 156

van de Berg WJ, van den Broeke M, Ettema J, van Meijgaard E, Kaspar F (2011) Significant contribution of insolation to Eemian melting of the Greenland Ice Sheet. Nat Geosci 4:679–682

van de Wal RSW (1996) Mass-balance modelling of the Greenland Ice Sheet: a comparison of an energy-balance and a degree-day model. Ann Glaciol 23:36–45

Vermeer M, Rahmstorf S (2009) Global sea level linked to global temperature. Proc Natl Acad Sci 106:21527–21532

Vinther BM, Andersen KK, Jones PD, Briffa KR, Cappelen J (2006) Extending Greenland temperature records into the late eighteenth century. J Geophys Res 111:D11105

Vinther BM, Buchardt SL, Clausen HB, Dahl-Jensen D, Johnsen SJ, Fisher DA, Koerner RM, Raynaud D, Lipenkov V, Andersen KK, Blunier T, Rasmussen SO, Steffensen JP, Svensson AM (2009) Holocene thinning of the Greenland Ice Sheet. Nature 461:385–388

Yohe G, Neumann J, Marshall P, Ameden H (1996) The economic cost of greenhouse-induced sea-level rise for developed property in the United States. Clim Change 32:387–410

Zwally HJ, Abdalati W, Herring T, Larson K, Saba J, Steffen K (2002) Surface melt-induced acceleration of Greenland ice-sheet flow. Science 297:218–222