The Cryosphere Discuss., 5, 3175–??, 2011
www.the-cryosphere-discuss.net/5/3175/2011/
doi:10.5194/tcd-5-3175-2011

The Cryosphere
Discussions

Preliminary assessment of model parametric uncertainty in projections of Greenland Ice Sheet behavior

P. J. Applegate[1], N. Kirchner[1], E. J. Stone[2], K. Keller[3], and R. Greve[4]

[1]Physical Geography and Quaternary Geology/Bert Bolin Climate Centre,
Stockholm University, Sweden
[2]BRIDGE, School of Geographical Sciences, University of Bristol, Bristol, UK
[3]Department of Geosciences/Earth and Environmental Sciences Institute,
Pennsylvania State University, University Park, Pennsylvania, USA
[4]Institute of Low Temperature Science, Hokkaido University, Sapporo, Japan

Received: 19 October 2011 – Accepted: 6 November 2011 – Published: 15 November 2011

Correspondence to: P. J. Applegate (patrick.applegate@natgeo.su.se)

Published by Copernicus Publications on behalf of the European Geosciences Union.

Abstract

Lack of knowledge about the values of ice sheet model input parameters introduces substantial uncertainty into projections of Greenland Ice Sheet contributions to future sea level rise. Computer models of ice sheet behavior provide one of several means of estimating future sea level rise due to mass loss from ice sheets. Such models have many input parameters whose values are not well known. Recent studies have investigated the effects of these parameters on model output, but the range of potential future sea level increases due to model parametric uncertainty has not been characterized. Here, we demonstrate that this range is large, using a 100-member perturbed-physics ensemble with the SICOPOLIS ice sheet model. Each model run is spun up over 125 000 yr using geological forcings, and subsequently driven into the future using an asymptotically increasing air temperature anomaly curve. All modeled ice sheets lose mass after 2005 AD. After culling the ensemble to include only members that give reasonable ice volumes in 2005 AD, the range of projected sea level rise values in 2100 AD is 30 % or more of the median. Data on past ice sheet behavior can help reduce this uncertainty, but none of our ensemble members produces a reasonable ice volume change during the mid-Holocene, relative to the present. This problem suggests that the model's exponential relation between temperature and precipitation does not hold during the Holocene, or that the central-Greenland temperature forcing curve used to drive the model is not representative of conditions around the ice margin at this time (among other possibilities). Our simulations also lack certain observed physical processes that may tend to enhance the real ice sheet's response. Regardless, this work has implications for other studies that use ice sheet models to project or hindcast the behavior of the Greenland ice sheet.

1 Introduction

The Greenland Ice Sheet is projected to contribute to sea level change by 2100 AD (Meehl et al., 2007) and beyond, but both the rate of ice mass loss and its eventual magnitude are unknown. The ice sheet contains enough ice to raise mean sea level worldwide by about 7.2 m (Bamber et al., 2001), if totally melted. Satellite measurements suggest that the ice sheet's mass balance is negative and perhaps becoming more so with time (Velicogna, 2009; Alley et al., 2010, and references therein; Zwally et al., 2011). Scaling arguments suggest ~0.1–0.5 m of mean sea level rise due to Greenland ice loss by 2100 AD (Pfeffer et al., 2008). Pfeffer et al. (2008) argue that the lower end of this range is more plausible than the higher end, because the larger number requires a rapid factor-of-10 increase in ice velocities. However, the difference between the ends of this range is economically important (e.g., Sugiyama et al., 2008), indicating a need for further investigation.

Ice sheet models provide an additional way of assessing future sea level change due to Greenland ice loss (e.g., Huybrechts and de Wolde, 1999; Greve, 2000; Gregory and Huybrechts, 2006; Price et al., 2011). These models typically include simplified treatments of ice flow, basal sliding, snowfall, and surface melting. The ice sheet modeling community has developed advanced treatments of all these processes, plus enhanced basal flow due to surface melting, ice shelf growth, calving, and sub-shelf melt (e.g., Parizek and Alley, 2004; Alley et al., 2008; Pollard and DeConto, 2009; Walker et al., 2009; Bueler and Brown, 2009; Robinson et al., 2010; Price et al., 2011). These new treatments are not implemented in all models at the present time. However, the models show remarkable success in simulating many aspects of ice sheet behavior over millennial time scales and longer (e.g., van Tatenhove et al., 1995, 1996; Greve, 1997; Simpson et al., 2009; Pollard and DeConto, 2009).

Remaining challenges in assessing future Greenland Ice Sheet changes include (1) characterizing model response to parameter choices, (2) establishing an initial state for prognostic simulations, and (3) matching data on the ice sheet's past behavior (van

der Veen, 2002; Heimbach et al., 2008; Aschwanden et al., 2009; Stone et al., 2010; Greve et al., 2011). Ice sheet models have many uncertain parameters, and the choice of parameter values has a strong influence on modeled behavior (Stone et al., 2010; Greve et al., 2011). Because the thermal field within the ice sheet is mostly unknown (cf. Greve, 2005), ice sheet models are "spun up" to the present using reconstructed former surface temperatures and sea levels. Achieving a good match between the modeled and observed ice thickness distributions at the end of this spinup is challenging (Aschwanden et al., 2009; Greve et al., 2011). In general, simulated ice volumes at the ends of spin-up runs are larger than expected (e.g., Heimbach et al., 2008; Stone et al., 2010; Robinson et al., 2010; Vizcaino et al., 2010; Greve et al., 2011; cf. Bamber et al., 2001). Finally, data on past ice sheet variations (e.g., Alley et al., 2010, and references therein) provide a check on ice sheet models: if a model reproduces past changes well, then we can have more confidence in its projections of future changes (cf. Oreskes et al., 1994).

Perturbed physics tuning exercises may help address these challenges. In a perturbed physics ensemble, the model is run many times with different parameter combinations to identify a group of runs that provide a reasonable fit to observations, usually the modern geometry of the ice sheet (e.g., Ritz et al., 1997; Stone et al., 2010; Greve et al., 2011; for fits to paleo-data, see Tarasov and Peltier, 2003; Lhomme et al., 2005; Simpson et al., 2009). These "good" ensemble members are likely more reliable estimators of future behavior than the ensemble as a whole (cf. Weigel et al., 2010). This approach is well established in climate modeling (e.g., climateprediction.net; Stainforth et al., 2005; Piani et al., 2005), and a small but growing number of ice sheet modeling studies use ensemble methods (e.g., Tarasov and Peltier, 2004; Napieralski et al., 2007; Hebeler et al., 2008; Stone et al., 2010).

In this paper, we present results from a small perturbed-physics ensemble with the ice sheet model SICOPOLIS (Greve, 1997; Greve et al., 2011; sicopolis.greveweb.net). Our approach builds on existing work by Stone et al. (2010) by using a spinup procedure that takes past climate variability into account and is agreed upon by the ice sheet modeling community (seaRISE partners, 2008; Greve et al., 2011). The results indicate that our present uncertainty about the best values of model parameters translates to a large spread among model-based projections of future Greenland Ice Sheet behavior.

The paper proceeds as follows. We describe the ice sheet model, ensemble design, climate forcing time series, and ensemble culling in Sect. 2, Methods. Section 3, Results, describes similarities and differences among the ensemble members during different parts of the model spinup period, then discusses the effects of ensemble culling on the range of model-projected future sea level increases from the Greenland ice sheet. Section 4, Discussion, treats the success of our ensemble in addressing the three challenges in ice sheet modeling identified above. Section 5, Conclusions, emphasizes the main outcomes of the study and provides some caveats that should be borne in mind when interpreting our model output.

2 Methods

Briefly, we applied the Latin hypercube ensemble methods of Stone et al. (2010) to the SICOPOLIS ice sheet model, as set up by Greve et al. (2011). The Latin hypercube method provides a quasi-random sampling of parameter space that is more even than that produced by Monte Carlo methods (Bevington and Robinson, 2003; Saltelli et al., 2008) and avoids wasting model evaluations on uninfluential parameters, as can happen with a grid design (Urban and Fricker, 2009). We can thus make a reasonable exploration of parameter space with a relatively small number of model evaluations.

2.1 Model description

As noted above, we carried out our simulations with the SICOPOLIS ice sheet model. SICOPOLIS has been previously described by Greve (1997) and Greve et al. (2011), and we refer interested readers to those papers for more information. The model is broadly comparable to most other large-scale ice sheet models, such as Glimmer (Rutt et al., 2009).

The model setup that we use is specifically intended for the problem of projecting future sea level change (Greve et al., 2011). It includes a horizontal resolution of 10 km, with 81 grid points in the vertical direction. These points are concentrated near the base, where the bulk of ice deformation occurs. The model time step is 1 yr.

Use of the SICOPOLIS model allows us to incorporate many thousands of years of geological data into model spinup (Sect. 2.3, below). As with most Greenland ice sheet models, SICOPOLIS calculates stresses within the ice body using the shallow-ice approximation (e.g., Hutter, 1983; Greve and Blatter, 2009). This approximation is reasonable over the bulk of the ice sheet (perhaps 70–80 % by area, and a greater percentage by volume), but fails in areas where the real ice sheet exhibits fast flow, such as in ice streams (see Joughin et al., 2010, for surface velocity maps). Higher-order models provide improved representations of ice flow where the shallow-ice approximation fails (Pattyn et al., 2008; Hindmarsh, 2009), but typically require more computing time than shallow-ice models. To our knowledge, higher-order models have not yet been used to carry out long integrations like those we undertake here with SICOPOLIS.

2.2 Ensemble design

We vary five model parameters among 100 ensemble members (Fig. 1; Supplementary Material). The number of model runs was chosen to achieve a reasonable tradeoff between covering parameter space and minimizing computation time. For comparison, our total number of evaluated model time steps (12.65×10^6) is slightly larger than that of Stone et al. (2010), who performed a larger number of shorter model runs.

The free parameters (and their ranges) include the ice flow enhancement factor (1–5, dimensionless), ice and snow positive degree-day factors ("PDD factors"; 5–20 mm day^{-1} °C and 1–5 mm day^{-1} °C, respectively); the geothermal heat flux (30–70 mW m^{-2}), and the basal sliding factor (0–20 m yr^{-1}/Pa). The ice flow enhancement factor corrects for differences between the rheology of ice as it is measured in the laboratory and that observed on an ice sheet scale; these differences are likely due to impurities and anisotropic fabric within the ice (Greve, 1997, his Eqs. 3 and 4; see also

Rutt et al., 2009, their Eq. 9). The positive degree-day factors describe a statistical relationship between surface temperatures and the rate of surface lowering (Braithwaite, 1995; Calov and Greve, 2005). The geothermal heat flux varies over the Earth's surface, but is difficult to measure under the ice sheet (see discussion in Stone et al., 2010); thus, it is often taken to be constant for purposes of ice sheet modeling (e.g., Ritz, 1997). Finally, the basal sliding factor determines how rapidly the ice slides over its bed where the interface is not frozen (Greve and Otsu, 2007, their Eq. 2).

For the first four parameters, the ranges are roughly the same as those investigated by Stone et al. (2010), who based their ranges on data-based studies from the literature (e.g., Dahl-Jensen and Gundestrup, 1987; Braithwaite, 1995). We expanded the ranges for the positive degree-day factors so that the EISMINT-3 preferred values (Huybrechts et al., 1998) lie well within the investigated range, instead of at one end, as in Stone et al. (2010). We also expanded the range of the geothermal heat flux parameter; Stone et al. (2010) found that this parameter was relatively uninfluential, and we hypothesized that a larger range might show an effect. The range we investigate is still within previous estimates (Greve, 2005; Buchardt and Dahl-Jensen, 2007; Stone, 2010). The basal sliding parameter ranges from 0 to about double the best value identified by Greve and Otsu (2007).

This list of free parameters is somewhat different from that used by Stone et al. (2010), but consistent with Ritz et al. (1997), in that we fix the atmospheric temperature lapse rates (Fausto et al., 2009) and include the basal sliding factor as a free parameter. In effect, we take the surface temperature and precipitation as given, even though these data sets contribute additional uncertainty to projected ice volumes (van der Veen, 2002; Stone et al., 2010).

2.3 Initial condition and climate forcing time series

All runs were driven by the same surface temperature, precipitation, and sea level forcings. The paleoclimate spinup (Fig. 2) closely resembles that of Greve et al. (2011). We began with the observed modern ice thickness and bedrock elevation grid (Bamber

et al., 2001) at −125 ka, during the Eemian interglacial. This initial condition is not ideal; much work shows that the ice sheet contained ~30–85 % of its present volume during the Eemian (Alley et al., 2010, and references therein). However, the errors in the initial condition should average out over the spinup period (see discussion in Rogozhina et al., 2011).

From 125 ka onward, we drove the model using a temperature anomaly curve based on the GRIP oxygen isotope record (Dansgaard et al., 1993; Johnsen et al., 1997) and background sea levels from the SPECMAP compilation of oxygen isotope measurements in deep-sea sediment cores (Imbrie et al., 1984). Precipitation changes by ~7 % for each degree of temperature change relative to the present (Greve et al., 2011, their Eq. 6; cf. van der Veen, 2002). The transfer functions for converting oxygen isotope measurements to surface temperatures and sea levels are given in Greve et al. (2011). Modern-day surface temperatures and precipitation values are from Fausto et al. (2009) and Ettema et al. (2009); these patterns are scaled in the model according to the calculated temperature and precipitation anomalies.

Near the end of the paleoclimate spinup, we substituted an instrumental record of southwestern Greenland mean annual temperature anomalies (Vinther et al., 2006) for the GRIP-based temperatures (Fig. 2). The Vinther et al. (2006) compilation covers the years 1784–2005 AD; we chose to begin the instrumental period in 1840, when the number of missing temperature records per year becomes noticeably smaller. Use of the Vinther et al. (2006) temperatures helps us to capture the interannual variability that could be important in explaining modern mass balance trends (Alley et al., 2010; Zwally et al., 2011).

After 2005, the surface temperature anomaly increases according to

$$T_\mathrm{f}(t) = \Delta T \times [1 - \exp(-\Delta t/\tau)]. \qquad (1)$$

In this expression, ΔT is the final temperature anomaly (6 °C) above mean annual 1840–1869 AD temperatures, less the mean 1976–2005 AD temperature anomaly (~1 °C); Δt is the year less 2005 AD; and τ is the time scale (100 yr). The form of this relation and the time scale come from Greve (2000). The final temperature rise is reasonable, given that Greenland is expected to warm ∼1.5–2 times as much as the global average (Church et al., 2001). We used this temperature forcing curve instead of one produced by a climate model because our goal is to highlight parametric uncertainties within the ice sheet model; differences among surface temperature change projections (Stainforth et al., 2005; Meehl et al., 2007) represent another layer of uncertainty.

We hold the sea level anomaly constant between 1840 AD and the end of the simulations, consistent with some earlier simulations of future Greenland evolution (Huybrechts, 1998). In effect, this assumption is conservative; allowing sea level to rise over the future would tend to increase mass loss from the ice sheet. However, the real ice sheet is likely much less sensitive to changes in sea level than temperature, even considering the large-amplitude sea level changes that take place over glacial-interglacial cycles (Alley et al., 2010; see also Gomez et al., 2010). In any case, the science of sea level rise is evolving rapidly, so we chose to neglect this effect for the present study.

2.4 Culling the ensemble

We evaluate the trustworthiness of each run by comparing the simulated total ice volume in 2005 AD to the modern ice volume (∼7.2 m sea level equivalent; Bamber et al., 2001), which is somewhat uncertain. The modern ice volume was estimated by Bamber et al. (2001) from kriging of geographically distributed ice thickness measurements, made from airborne radar units. Adjacent measurements do not always agree exactly, reflecting some aggregate of bed roughness and measurement uncertainty. Moreover, observation density varies over the ice sheet; flight lines lie close to one another near airfields and become sparse farther south. Thus, the true modern ice volume could be either larger or smaller than the central estimate, but how wide the range of possible values might be is difficult to estimate.

To account for uncertainty in the modern ice volume, we used a simple windowing approach. All runs that fell within a certain distance of the modern ice volume in 2005

AD (dashed line in Fig. 3) were kept, whereas the others were discarded. We investigated window widths of ±20 %, 10 %, 5 %, and 2.5 % of the modern ice volume. These estimates of uncertainty in modern ice volume are somewhat ad hoc, but as we show later, they have little influence on our projection uncertainty.

Total ice volume is a reasonable comparison metric because we are interested in future sea level change, and because inferences about the past state of the ice sheet are usually stated in terms of volume changes relative to the present (e.g., Alley et al., 2010, their Fig. 13). Other metrics for comparing simulated ice sheets to the observed one exist, including the ice-covered area, maximum ice thickness (Ritz et al., 1997; Stone et al., 2010), the root mean squared ice thickness (e.g., Greve and Otsu, 2007), and matching the distribution of ice surface velocities (e.g., Aschwanden et al., 2009). Using an aggregate measure such as total ice volume helps avoid nontrivial statistical issues with autocorrelation, in which adjacent residuals between observations and model predictions are not independent of one another (e.g., Bloomfield and Nychka, 1992).

As noted above, data on the ice sheet's past behavior (Alley et al., 2010, and references therein) also provide constraints on model behavior. Consistent with most earlier Greenland ice sheet modeling studies, we neglect this information in our model tuning; instead, we use it to evaluate the reasonableness of the culled ensemble (Sect. 4.1, below).

3 Results

In this section, we make some general observations on the behavior of the full ensemble (Fig. 3, gray and blue curves) before treating those runs that reproduce the modern ice volume well (Fig. 3, blue curves). Because the initial condition and forcings are identical for all model runs (Fig. 2), variability among runs (Figs. 3, 4) is due solely to parameter choice (Fig. 1).

3.1 Eemian through the early glacial period (−125 ka to −75 ka)

As noted above, all simulations start from the modern ice geometry, with an ice volume of 7.2 m sea level equivalent (Fig. 3). Temperatures begin a few degrees below modern values (Fig. 2), before rising twice to much warmer values at about −120 and −115 ka. Although apparently reasonable for the Eemian, the temperature values and structure of the GRIP record during this time is suspect (Chappellaz et al., 1997). From this maximum, temperature and sea level generally decline. Ice volumes increase over the same period, stabilizing sometime before −75 ka (Fig. 3).

Despite the general resemblance of the model curves to one another (Fig. 3), there is substantial divergence among ensemble members. The spread among model runs is most noticeable during the Eemian warmth; some model realizations produce a nearly ice-free Greenland, whereas in others the ice volume changes only slightly.

3.2 Early glacial period through the early Holocene (−75 ka to −10 ka)

Between −75 and −10 ka, simulated ice volumes are remarkably stable and consistent among runs (Fig. 3). This result seems counterintuitive given the large temperature fluctuations seen in the GRIP record (up to ∼15 °C; Fig. 2). However, these fluctuations are only a few ka long, and the resulting ice losses due to mass balance changes are small compared to simulated ice volumes.

3.3 Early Holocene to the beginning of the preindustrial period (−10 ka to −0.16 ka, or 1840 AD)

Despite relatively stable Holocene temperatures (Fig. 2), simulated ice volumes generally decrease between −10 ka and −0.16 ka (Fig. 3). Many runs show a very slight growth near the end of this period, reflecting the “Little Ice Age” as seen in the central Greenland ice cores (Alley et al., 2010, and references therein).

As for the Eemian, the Holocene warmth produces considerable spread among the individual model runs. One run even grows for a few ka before slowly shrinking. This realization (#95, Supplementary Material) has exceptionally low values of both the ice and snow positive degree-day factors, which allow increased precipitation (Sect. 2.3; Greve et al., 2011) to overtake ablation temporarily.

3.4 1840 AD to 3500 AD

The Vinther et al. (2006) instrumental temperatures begin the climb out of the “Little Ice Age”, and the assumed future climate trajectory builds off the end of this record. Ice volumes decline very slightly between 1840 AD and 2100 AD, with larger decreases thereafter. As during the Eemian and earlier in the Holocene, the spread among model realizations is large.

3.5 The culled ensemble

Culling the ensemble reduces the divergence among model runs during warm periods (Figs. 3, 4). The 26 ensemble members that lie within 10 % of the estimated modern ice volume in 2005 AD change by comparable amounts between warm and cold periods. Curiously, the spread among these runs during cold periods is almost as large as that of the full ensemble.

Depending on the culling window width (Sect. 2.2, above), we have different numbers of ensemble members remaining (Fig. 4, bottom). An assumed 20 % uncertainty in modern estimated ice volume leaves 52 ensemble members out a possible 100. For 10 %, 5 %, and 2.5 % uncertainty in modern ice volume, the ensemble culling leaves 26, 11, and 6 model runs, respectively.

3.6 Future sea level change in the culled ensemble

The spread in projected future sea level change among model runs is large, even after ensemble culling (Fig. 4). For 2100 AD and an assumed 10 % uncertainty in modern Greenland ice volume, the median change in global mean sea level due to Greenland mass loss is ~0.14 m, and the range among the 26 model runs that meet the modern ice volume criterion is ~0.1 m, for a fractional uncertainty of ~70 %. Much of this uncertainty persists even if we use stricter culling criteria – for example, assuming that we know the modern ice volume to within 2.5 % gives a fractional uncertainty of 30 % in 2100 AD.

4 Discussion

Our results highlight the challenge presented by parametric uncertainty for projections of future Greenland ice sheet behavior, building on previous work in this area (e.g., van der Veen, 2002; Stone et al., 2010). In the Introduction, we identified three additional modeling problems that we hoped to address with our perturbed-physics ensemble. These were (1) characterizing model response to parameter choice, (2) establishing an initial state for prognostic simulations, and (3) matching data on the ice sheet's past behavior. We address the ensemble's success in meeting these goals here.

4.1 Parameter choice, simulated modern ice volume, and ice sheet sensitivity

Given these model results (Figs. 3, 4), we might ask which parameter values are most consistent with the modern observed ice volume. If there are one or more parameter combinations that match the modern condition well, these values can be used in other modeling experiments.

This question was previously posed by Stone et al. (2010), who noted that high values of the ice flow and ice positive degree day factors yielded the best matches with observed total ice volume; the other parameters played smaller roles. Our results are consistent with Stone et al. (2010) in that the ice PDD factor has a dominant influence on simulated ice volumes at the end of the spinup (Fig. 5; cf. Fig. 7 in Stone et al.,

2010). However, we find that treating the basal sliding factor as a free parameter reduces the influence of the ice flow factor.

We can make few general statements about optimal parameter combinations for ice sheet spinup (Fig. 5). None of our "best" runs, those that fall within 10 % of the modern ice volume in 2005 AD, have an ice positive degree-day factor greater than ~15 mm day^{-1} °C. This value is well within the range identified by Braithwaite (1995). Otherwise, the "best" runs span the entire free range of the four remaining parameters, indicating that any value of these parameters is potentially consistent with the modern ice volume. There is some suggestion that large basal sliding factors are most compatible with smaller values of the ice flow factor and vice versa (Fig. 1), but this apparent tradeoff might disappear given more model runs.

In our ensemble, the ice positive degree-day factor largely determines the near-term future ice sheet response (Fig. 6). As for modern ice volumes (Fig. 5), there is a strong relation between the ice PDD factor and 2005–2100 AD ice volume change, but the other parameters appear to be relatively unimportant (Fig. 6).

4.2 Simulated modern ice thicknesses

By construction, our culled ensemble matches the modern ice volume. However, problems persist in the modeled ice thicknesses (Fig. 7). Consistent with earlier results using a similar model setup (Greve et al., 2011, their Fig. 2), the ice is generally too extensive in the south and has large gaps in the north. Our tuning exercise did cover part of the falsely ice-free area in northern Greenland noted by Greve et al. (2011). Ice thickness errors of 1.5 km or more are present around the edges of the ice sheet; in particular, the ice is too thin upflow from Jakobshavn, about a third of the way northward on the west coast.

It is possible that an appropriately-tuned higher-order model would produce a better fit to the observed ice thickness grid than we have yet achieved with SICOPOLIS. However, the large-scale shape of the ice sheet is more strongly controlled by surface mass balance than ice flow; for thin ice, changes in ice thickness due to flow are small compared to those caused by surface mass imbalances (Greve, 1997). We attribute the bulk of the remaining errors in geographically distributed ice thickness values to problems with the modeled mass balance.

4.3 Comparison of modeled results to assessed past volume changes

As noted in the Introduction, the ice sheet's past behavior provides a check on ice sheet model results. Alley et al. (2010) recently surveyed the literature and gave assessments of ice volume changes, relative to the present, for three time slices covered by our runs, the Eemian, Last Glacial Maximum, and mid-Holocene. These assessments are based on a combination of isostatic rebound (Peltier, 2004; Fleming and Lambeck, 2004) and data-constrained ice sheet modeling (Cuffey and Marshall, 2000; Lhomme et al., 2005) studies. Because some of these studies use ice sheet models to translate observed ice core oxygen isotope values into ice sheet changes, these assessed ice volume changes are not truly independent of our results. However, they do provide a first-order check on our model output.

The culled ensemble simulates assessed ice volume changes well during the Eemian (−115 ka; Fig. 4), but has problems during the mid-Holocene and Last Glacial Maximum (−5 ka and −20 ka, respectively). In particular, none of the model runs produce a smaller-than-modern ice sheet during the mid-Holocene (−5 ka), as expected from paleo-data. The overlap between our "good" model runs and the estimated Last Glacial Maximum (−20 ka) ice volume change is also minimal, but this discrepancy could be reduced by a different weighting function for evaluating the model runs. Moreover, the agreement between our simulated late-Eemian ice volume changes and previously estimated values may be fortuitous; the peak in the GRIP temperature record at −115 ka could be due to flow disturbances in the ice core (Chappellaz et al., 1997; Cuffey and Marshall, 2000).

The apparent disagreement between simulated and assessed ice volume changes in the mid-Holocene (Fig. 4) could be due to the temperature forcing curve used to drive the model, or to the assumed-constant exponential relation between

surfacetemperature anomaly and precipitation (among other possibilities). Air temperatures over some parts of the Greenland Ice Sheet were likely warmer during the mid-Holocene than geographic scaling of the GRIP oxygen isotope record would indicate (Y. Axford, personal communication, 2011; Dahl-Jensen et al., 1998; Young et al., 2011; see also Vinther et al., 2009); if accounted for in a model simulation, these warmer regional temperatures might bring modeled ice volumes closer to the estimated values. Further, our model runs assume a constant ~7 % increase in precipitation per degree Celsius temperature increase (Greve et al., 2011). However, this relation has long been controversial (see discussion in van der Veen, 2002), and it may be especially poor for the mid- to late-Holocene (Cuffey and Clow, 1997). Further research is needed to reduce the discrepancies between model- and data-based reconstructions of past Greenland Ice Sheet configurations.

5 Conclusions

Our results (Fig. 4) suggest that parametric uncertainty in ice sheet model-based projections of Greenland Ice Sheet behavior is on the order of 30–70 %, expressed as the range of plausible model outcomes divided by the median. This outcome is not sensitive to the parameter ranges we investigated, but rather depends on the uncertainty in modern ice volumes. Our results do not provide a probabilistic assessment of future ice sheet changes; we make no statement about the relative plausibility of different model runs within our culled ensemble.

Our analysis neglects several sources of uncertainty that will tend to increase the ranges shown in Figure 4. In particular, we assume that the climate and subglacial topography boundary conditions are well known. Estimating the effects of errors in these data sets on modeled ice volumes is complex, but Stone et al. (2010) found that updating older, EISMINT-3 data sets (Huybrechts, 1998) to their modern equivalents increased simulated equilibrium ice volumes by ~17 % if model input parameters were held constant. It is clear that the best-estimate model parameter values depend strongly on the input data sets. The ice core-inferred paleotemperatures used to spin up ice sheet models also contribute to projection uncertainty (Stone et al., 2010; see also Rogozhina et al., 2011). Finally, future ice volumes depend on uncertain emissions trajectories and the broader climate system's response (Meehl et al., 2007).

Even our most responsive model runs may underestimate mass loss from the real ice sheet. If precipitation remains constant in the future, instead of increasing at ~7 %/degree C of warming (Sect. 2.3; Greve, 2011), then the ice sheet will shrink more rapidly than we project. There are a number of mechanisms for rapid ice loss that are not included in this ensemble. For example, surface melting may lead to basal lubrication and enhanced transport of ice to the margin (Zwally et al., 2002; Parizek and Alley, 2004; Bartholomew et al., 2010). We neglect this possibility here; see Greve and Otsu (2007) for model runs with SICOPOLIS that include this effect. Additionally, ocean warming may contribute to mass loss where the ice is in contact with the water (Straneo et al., 2010), and the resulting rapid thinning of marine ice margins could then propagate up ice streams to the central parts of the ice sheet. This scenario cannot be captured by shallow-ice models like SICOPOLIS, but is expected to appear in higher-order models. Complex models typically have more parameters than simpler ones, so sensitivity experiments with higher-order models (e.g., Price et al., 2011) might lead to a wider range of future Greenland states (cf. Saltelli et al., 2008).

Given these problems, both our uncertainty estimates and our projections of future ice sheet mass loss may be too small. Despite the large variation among individual model runs, all of our modeled ice sheets lose mass from 2005 AD onwards. Thus, our work agrees with the scientific consensus, which says that sea level rise due to enhanced mass loss from the Greenland ice sheet in the face of surface temperature increases is very likely (Lemke et al., 2007).

Supplementary material related to this article is available online at: http://www.the-cryosphere-discuss.net/5/3175/2011/tcd-5-3175-2011-supplement.zip.

Acknowledgements. We thank the seaRISE project members for compiling the input data sets and making them available on the Web (http://websrv.cs.umt.edu/isis/index.php/Data). R. Alley, J. Seguinot, Y. Axford, and D. Pollard provided helpful discussions and comments on the manuscript. P. Applegate thanks the Wednesday morning discussion group at Stockholm University for much useful feedback. Support for P. Applegate was provided by the Bert Bolin Climate Centre at Stockholm University through a postdoctoral fellowship. R. Greve was supported by a Grant-in-Aid for Scientific Research A (No. 22244058) from the Japan Society for the Promotion of Science. Partial support from the US Department of Energy and the Penn State Center for Climate Risk Management is gratefully acknowledged. The data processing for this paper was done in Octave (http://www.octave.org), and the figures were made with Octave and Inkscape (http://www.inkscape.org). This paper is a contribution from the Bert Bolin Centre for Climate Research, Stockholm University.

References

Alley, R. B., Horgan, H. J., Joughin, I., Cuffey, K. M., Dupont, T. K., Parizek, B. R., Anandakrishnan, S., and Bassis, J.: A simple law for ice-shelf calving, Science, 322, 1344 pp., 2008.

Alley, R. B., Andrews, J. T., Brigham-Grette, J., Clarke, G. K. C., Cuffey, K. M., Fitzpatrick, J. J., Funder, S., Marshall, S. J., Miller, G. H., Mitrovica, J. X., Muhs, D. R., Otto-Bliesner, B. L., Polyak, L., and White, J. W. C.: History of the Greenland Ice Sheet: paleoclimate insights, Quat. Sci. Rev., 29, 1728–1756, 2010.

Aschwanden, A., Khroulev, C., and Bueler, E.: SeaRISE Greenland – on ”spin-up” procedures, EOS, 90, abstract C23B-0500, 2009.

Bamber, J. L., Layberry, R. L., and Gogineni, S. P.: A new ice thickness and bed data set for the Greenland ice sheet 1. Measurement, data reduction, and errors, J. Geophys. Res., 106, 33773–33780, 2001.

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

Bevington, P. R. and Robinson, D. K.: Data reduction and error analysis for the physical sciences (3rd Edn.), McGraw-Hill, 320 pp., 2003.

Bloomfield, P. and Nychka, D.: Climate spectra and detecting climate change, Climatic Change, 21, 275–287, 1992.

Braithwaite, R. J.: Positive degree-day factors for ablation on the Greenland ice sheet studied by energy-balance modelling, J. Glaciol., 41, 153–160, 1995.

Bueler, E. and Brown, J.: Shallow shelf approximation as a ”sliding law” in a thermomechanically coupled ice sheet model, J. Geophys. Res., 114, F03008, doi:10.1029/2008JF001179, 2009.

Buchardt, S. L. and Dahl-Jensen, D.: Estimating the basal melt rate at NorthGRIP using a Monte Carlo technique, Ann. Glaciol., 45, 137–142, 2007.

Calov, R. and Greve, R.: A semi-analytical solution for the positive degree-day model with stochastic temperature variations, J. Glaciol., 51, 173–175, 2005.

Chappellaz, J., Brook, E., Blunier, T., and Malaize, B.: CH4 and del18O of O2 records from Antarctic and Greenland ice: a clue for stratigraphic disturbance in the bottom part of the Greenland Ice Core Project and the Greenland Ice Sheet Project 2 ice cores, J. Geophys. Res., 102, 26547–26557, 1997.

Church, J. A., Gregory, J. M., Huybrechts, P., Kuhn, M., Lambeck, K., Nhuan, M. T., Qin, D., and Woodworth, P. L.: Changes in sea level, in: Climate change 2001: Working Group 1: the scientific basis, editd by: Houghton, J. T., et al., Cambridge, 2001.

Cuffey, K. M. and Clow, G. D.: Temperature, accumulation, and ice sheet elevation in central Greenland through the last deglacial transition, J. Geophys. Res., 102, 26383–26396, 1997.

Cuffey, K. M. and Marshall, S. J.: Substantial contribution to sea-level rise during the last interglacial from the Greenland ice sheet, Nature, 404, 591–594, 2000.

Dahl-Jensen, D. and Gundestrup, N. S.: Constitutive properties of ice at Dye-3, Greenland, in Waddington, E., and Walder, J.: The Physical Basis of Ice Sheet Modelling, IAHS Pub. 170, 31–43, 1987.

Dahl-Jensen, D., Mosegaard, K., Gundestrup, N., Clow, G. D., Johnsen, S. J., Hansen, A. W., and Balling, N.: Past temperatures directly from the Greenland ice sheet, Science, 282, 268–271, 1998.

Dansgaard, W., Johnsen, S. J., Clausen, H. B., Dahl-Jensen, D., Gundestrup, N. S., Hammer, C. U., Hvidberg, C. S., Steffensen, J. P., Sveinbjörnsdottir, A. E., Jouzel, J., and Bond, G.: Evidence for general instability of past climate from a 250-kyr ice core record, Nature, 364, 218–220, 1993.

Ettema, J., van den Broeke, M. R., van Meijgaard, E., van de Berg, W. J., Bamber, J. L., Box, J. E., and Bales, R. C.: Higher surface mass balance of the Greenland ice sheet revealed by high-resolution climate modeling, Geophys. Res. Lett., 36, L12501, doi:10.1029/2009GL038110, 2009.

Fausto, R. S., Ahlstrom, A. P., van As, D., Boggild, C. E., and Johnsen, S. J.: A new present-day temperature parameterization for Greenland, J. Glaciol., 55, 95–105, 2009.

Fleming, K. and Lambeck, K.: Constraints on the Greenland Ice Sheet since the Last Glacial Maximum from sea-level observations and glacial-rebound models, Quat. Sci. Rev., 23, 1053–1077, 2004.

Gomez, N., Mitrovica, J. X., Huybers, P., and Clark, P. U.: Sea level as a stabilizing factor for marine-ice-sheet grounding lines, Nat. Geosci., 3, 850–853, 2010.

Gregory, J. M. and Huybrechts, P.: Ice-sheet contributions to future sea level change, Phil. Trans. R. Soc. A, 364, 1709–1731, 2006.

Greve, R.: 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, 1997.

Greve, R.: On the response of the Greenland Ice Sheet to greenhouse climate change, Climatic Change, 46, 289–303, 2000.

Greve, R.: Relation of basal measured temperatures and the spatial distribution of the geothermal heat flux for the Greenland ice sheet, Ann. Glaciol., 42, 424–432, 2005.

Greve, R. and Otsu, S.: The effect of the north-east ice stream on the Greenland ice sheet in changing climates, The Cryosphere Discuss., 1, 41–76, doi:10.5194/tcd-1-41-2007, 2007.

Greve, R. and Blatter, H.: Dynamics of Ice Sheets and Glaciers, Monograph Series Advances in Geophysical and Environmental Mechanics and Mathematics, XIV, 287 pp., 2009.

Greve, R., Saito, F., and Abe-Ouchi, A.: Initial results of the SeaRISE numerical experiments with the models SICOPOLIS and IcIES for the Greenland ice sheet, Ann. Glaciol., 52, 23–30, 2011.

Hebeler, F., Purves, R. S., and Jamieson, S. S. R.: The impact of parametric uncertainty and topographic error in ice-sheet modelling, J. Glaciol., 54, 899–919, 2008.

Heimbach, P. and Bugnion, V.: Greenland ice-sheet volume sensitivity to basal, surface and initial conditions derived from an adjoint model, Ann. Glaciol., 50, 67–80, 2009.

Hindmarsh, R. C. A.: Consistent generation of ice-streams via thermo-viscous instabilities modulated by membrane stresses, Geophys. Res. Lett., 36, L06502, doi:10.1029/2008GL036877, 2009.

Hutter, K.: Theoretical Glaciology: Material Science of Ice and the Mechanics of Glaciers and Ice Sheets, D. Reidel, Dordrecht, 1983.

Huybrechts, P.: Report of the third EISMINT workshop on model intercomparison, 1998. Available online at http://homepages.vub.ac.be/$\sim$phuybrec/eismint.html, accessed 13 Oct 2011.

Huybrechts, P. and de Wolde, J.: The dynamic response of the Greenland and Antarctic ice sheets to multiple-century climatic warming: J. Climate, 12, 2169–2188, 1999.

Imbrie, J., Hays, J. D., Martinson, D. G., McIntyre, A., Mix, A. C., Morley, J. J., Pisias, N. G., Prell, W. L., Shackleton, N. J.: The orbital theory of Pleistocene climate: support from a revised chronology of the marine del18O record, in: Milankovitch and climate: understanding the response to astronomical forcing, Part 1, edited by: Berger, A. J., Imbrie, J., Hays, J., Kukla, G., and Saltzman, B., D. Reidel Publishing Co., Dordrecht, 269-305, 1984.

Johnsen, S. J., Clausen, H. B., Dansgaard, W., Gundestrup, N. S., Hammer, C. U., Andersen, U., Andersen, K. K., Hvidberg, C. S., Dahl-Jensen, D., Steffensen, J. P., Shoji, H., Sveinbjrnsdttir, . E., White, J. W., Jouzel, J., and Fisher, D.: The del18O record along the Greenland Ice Core project deep ice core and the problem of possible Eemian climatic instability, J. Geophys. Res., 102, 26397–26410, 1997.

Joughin, I., Smith, B. E., Howat, I. M., Scambos, T., and Moon, T.: Greenland flow variability from ice-sheet-wide velocity mapping, J. Glaciol., 56, 415–430, 2010.

Lemke, P., Ren, J., Alley, R. B., Allison, I., Carrasco, J., Flato, G., Fujii, Y, Kaser, G., Mote, P., Thomas, R. H., and Zhang, T.: Observations: changes in snow, ice, and frozen ground, edited by: Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., and three others, Cambridge University Press, Cambridge, 2007.

Lhomme, N., Clarke, G. K. C., and Marshall, S. J.: Tracer transport in the Greenland Ice Sheet: constraints on ice cores and glacial history, Quat. Sci. Rev., 24, 173–194, 2005.

Meehl, G. A., Stocker, T. F., Collins, W. D., Friedlingstein, P., Gaye, A. T., Gregory, J. M., Kitoh, A., Knutti, R., Murphy, J. M., Noda, A., Raper, S. C. B., Watterson, I. G., Weaver, A. J., and Zhao, Z-C.: Global climate projections, edited by: Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., and three others, Cambridge University Press, Cambridge, 2007.

Napieralski, J., Hubbard, A., Li, Y., Harbor, J., Stroeven, A. P., Kleman, J., Alm, G., and Jansson, K. N.: Towards a GIS assessment of numerical ice-sheet model performance using geomorphological data, J. Glaciol., 53, 71–83, 2007.

Oreskes, N., Shrader-Frechette, K., and Belitz, K.: Verification, validation, and confirmation of

numerical models in the Earth sciences, Science, 264, 641–646, 1994.

Pattyn, F., Perichon, L., Aschwanden, A., Breuer, B., de Smedt, B., Gagliardini, O., Gudmundsson, G. H., Hindmarsh, R. C. A., Hubbard, A., Johnson, J. V., Kleiner, T., Konovalov, Y., Martin, C., Payne, A. J., Pollard, D., Price, S., Rückamp, M., Saito, F., Souček, O., Sugiyama, S., and Zwinger, T.: Benchmark experiments for higher-order and full-Stokes ice sheet models (ISMIPHOM), The Cryosphere, 2, 95–108, doi:10.5194/tc-2-95-2008, 2008.

Parizek, B. R. and Alley, R. B.: Implications of increased Greenland surface melt under global-warming scenarios: ice sheet simulations, Quat. Sci. Rev., 23, 1013–1027, 2004.

Peltier, W. R.: Global glacial isostasy and the surface of the ice-age Earth: the ICE-5G (VM2) model and GRACE, Ann. Rev. Earth Planet. Sci., 32, 111–149, 2004.

Pfeffer, W. T., Harper, J. T., and O'Neel, S.: Kinematic constraints on glacier contributions to 21st-century sea-level rise, Nature, 321, 1340–1343, 2008.

Piani, C., Frame, D. J., Stainforth, D. A., and Allen, M. R.: Constraints on climate change from a multi-thousand member ensemble of simulations, Geophys. Res. Lett., 32, L23825, doi:10.1029/2005GL024452, 2005.

Pollard, D. and DeConto, R.: Modelling West Antarctic ice sheet growth and collapse through the past five million years, Nature, 458, 329–333, 2009.

Price, S. F., Payne, A. J., Howat, I. M., and Smith, B. E.: Committed sea-level rise for the next century from Greenland ice sheet dynamics during the past decade, Proc. Natl. Acad. Sci. USA, 108, 8978–8983, 2011.

Ritz, C., Fabre, A., and Letreguilly, A.: 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, 11–24, 1997.

Robinson, A., Calov, R., and Ganopolski, A.: An efficient regional energy-moisture balance model for simulation of the Greenland Ice Sheet response to climate change, The Cryosphere, 4, 129–144, doi:10.5194/tc-4-129-2010, 2010.

Rogozhina, I., Martinec, Z., Hagedoorn, J. M., Thomas, M., and Fleming, K.: On the long-term memory of the Greenland Ice Sheet, J. Geophys. Res., 116, F01011, doi:10.1029/2010JF001787, 2011.

Rutt, I. C., Hagdorn, M., Hulton, N. R. J., and Payne, A. J.: The Glimmer community ice-sheet model, J. Geophys. Res.-Earth, 114, F02004, doi:10.1029/2008JF001015, 2009.

Saltelli, A., Ratto, M., Andres, T., Campolongo, F., Cariboni, J., Gatelli, D., Saisana, M., and Tarantola, S.: Global sensitivity analysis: the primer, Wiley, New York, 292 pp., 2008.

seaRISE partners: Assessing ice sheet contributions to sea level through the 21st century, seaRISE White Paper, 2008. Available online at http://websrv.cs.umt.edu/isis/index.php/SeaRISE_White_Paper, last accessed: 18 October 2011.

Simpson, M. J. R., Milne, G. A., Huybrechts, P., and Long, A. J.: Calibrating a glaciological model of the Greenland ice sheet from the Last Glacial Maximum to present-day using field observations of relative sea level and ice extent, Quaternary Sci. Rev., 28, 1631–1657, 2009.

Stainforth, D. A., Aina, T., Christensen, C., Collins, M., Faull, N., Frame, D. J., Kettleborough, J. A., Knight, S., Martin, A., Murphy, J. M., Piani, C., Sexton, D., Smith, L. A., Spicer, R. A., Thorpe, A. J., and Allen, M. R.: Uncertainty in predictions of the climate response to rising levels of greenhouse gases, Nature, 433, 403–406, 2005.

Stone, E. J., Lunt, D. J., Rutt, I. C., and Hanna, E.: Investigating the sensitivity of numerical model simulations of the modern state of the Greenland ice-sheet and its future response to climate change, The Cryosphere, 4, 397–417, doi:10.5194/tc-4-397-2010, 2010.

Straneo, F., Hamiton, G. S., Sutherland, D. A., Stearns, L. A., Davidson, F., Hammill, M. O., Stenson, G. B., and Rosing-Asvid, A.: Rapid circulation of warm subtropical waters in a major glacial fjord in East Greenland, Nat. Geosci., 3, 182–186, 2010.

Sugiyama, M., Nicholls, R. J., and Vafeidis, A.: Estimating the economic cost of sea-level rise, MIT Joint Program on the Science and Policy of Global Change Report No. 156, 2008. Available online at http://dspace.mit.edu/handle/1721.1/41522, last accessed: 18 October 2011.

Tarasov, L. and Peltier, W. R.: Greenland glacial history, borehole constraints, and Eemian extent, J. Geophys. Res., 108, 2143, 2003.

Tarasov, L. and Peltier, W. R.: A geophysically constrained large ensemble analysis of the deglacial history of the North American ice-sheet complex, Quat. Sci. Rev., 23, 359–388, 2004.

Urban, N. M. and Fricker, T. E.: A comparison of Latin hypercube and grid ensemble designs for the multivariate emulation of an Earth system model, Computers and Geosciences, 36, 746–755, 2010.

van der Veen, C. J.: Polar ice sheets and global sea level: how well can we predict the future?, Glob. Planet. Change, 32, 165–194, 2002.

van Tatenhove, F. G. M., van der Meer, J. J. M., and Huybrechts, P.: Glacial-geological/geomorphological research in west Greenland used to test an ice-sheet model, Quat. Res., 44, 317–327, 1995.

van Tatenhove, F. G. M., Fabre, A., Greve, R., and Huybrechts, P.: Modelled ice-sheet margins of three Greenland ice-sheet models compared with a geological record from ice-marginal deposits in central west Greenland, Ann. Glaciol., 23, 52–58, 1996.

Vinther, B. M., Andersen, K. K., Jones, P. D., Briffa, K. R., and Cappelen, J.: Extending Greenland temperature records into the late eighteenth century, J. Geophys. Res., 111, D11105, doi:10.1029/2005JD006810, 2006.

Vinther, B. M., Buchardt, S. L., Clausen, H. B., Dahl-Jensen, D., Johnsen, S. J., Fisher, D. A., Koerner, R. M., Raynaud, D., Lipenkov, V., Andersen, K. K., Blunier, T., Rasmussen, S. O., Steffensen, J. P., and Svensson, A. M.: Holocene thinning of the Greenland ice sheet, Nature, 461, 385–388, 2009.

Vizcaino, M., Mikolajewicz, U., Jungclaus, J., and Schurgers, G.: Climate modification by future ice sheet changes and consequences for ice sheet mass balance, Clim. Dyn., 34, 301–324, 2010.

Walker, R. T., Dupont, T. K., Holland, D. M., Parizek, B. R., and Alley, R. B.: Initial effects of oceanic warming on a coupled ocean-ice shelf-ice stream system, Earth Planet. Sci. Lett., 287, 483–487, 2009.

Weigel, A. P., Knutti, R., Liniger, M. A., and Appenzeller, C.: Risks of model weighting in multimodel climate projections, J. Climate, 23, 4175–4191, 2010.

Young, N. E., Briner, J. P., Steward, H. A. M., Axford, Y., Csatho, B., Rood, D. H., and Finkel, R. C.: Response of Jakobshavn Isbrae, Greenland, to Holocene climate change, Geology, 39, 131–134, 2011.

Zwally, H. J., Abdalati, W., Herring, T., Larson, K., Saba, J., and Steffen, K.: Surface melt-induced acceleration of Greenland Ice-Sheet flow, Science, 297, 218–222, 2002.

Zwally, H. J., Li, J., Brenner, A. C., Beckley, M., Cornejo, H. G., diMarzio, J., Giovinetto, M. B., Neumann, T. A., Robbins, J., Saba, J. L., Yi, D., and Wang, W.: Greenland ice sheet mass balance: distribution of increased mass loss with climate warming; 2003-07 versus 1992-2002, J. Glaciol., 57, 88–102, 2011.

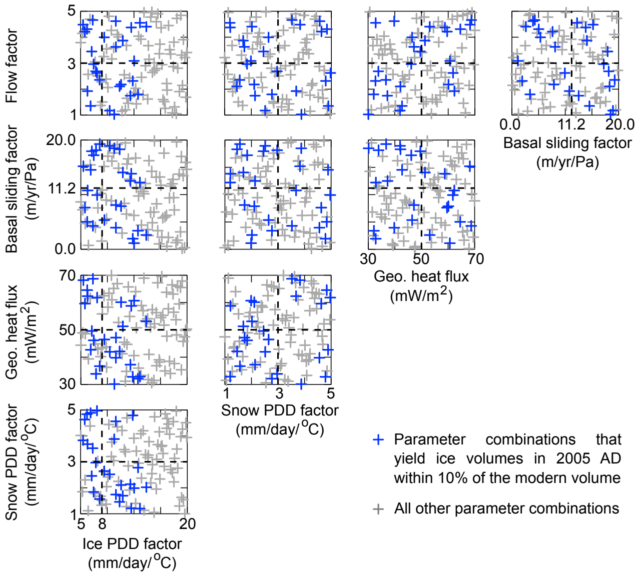

Fig. 1. Parameter combinations used in the perturbed-physics ensemble, as projected onto two-dimensional slices through the five-dimensional space. Dashed lines indicate EISMINT-3 best estimates for most model parameters (Huybrechts, 1998; Stone et al., 2010), except the basal sliding factor, which comes from Greve and Otsu (2007). Blue crosses indicate parameter combinations that are consistent with the modern ice volume after model spinup (within 10 % of the estimated modern ice volume in 2005 AD; Sect. 2.2).

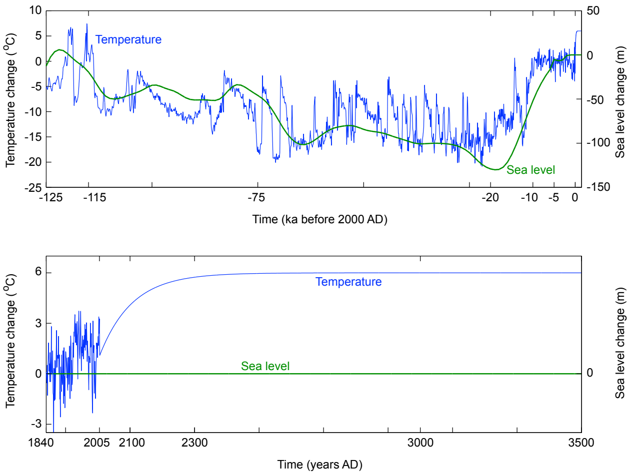

Fig. 2. Surface temperature (blue) and background sea level (green) curves used to drive the ice sheet model simulations. Top panel, full extent of runs (−125 ka to 3500 AD; 2000 AD is indicated by 0); bottom panel, 1840 AD to 3500 AD. Temperature and sea level curves for −125 ka through 1840 AD come from the seaRISE project (http://websrv.cs.umt.edu/isis/index.php/SeaRISE_Assessment; seaRISE partners, 2008; Greve et al., 2011), and are based on oxygen isotopes in ocean sediment cores (Imbrie et al., 1984) and in central-Greenland ice cores (Dansgaard et al., 1993; Johnsen et al., 1997). 1840–2005 AD temperatures come from southwestern Greenland observations (Vinther et al., 2006). Future temperatures assume an asymptotic increase to ~5 degrees C above 1976–2005 levels, with a time scale of 100 yr (Greve, 2000; see text). Background sea level is held constant from 1840–3500 AD. Labeled tick marks are those referred to in the text and figures; unlabeled tick marks are 25 ka apart in the top panel and 200 yr apart in the lower panel.

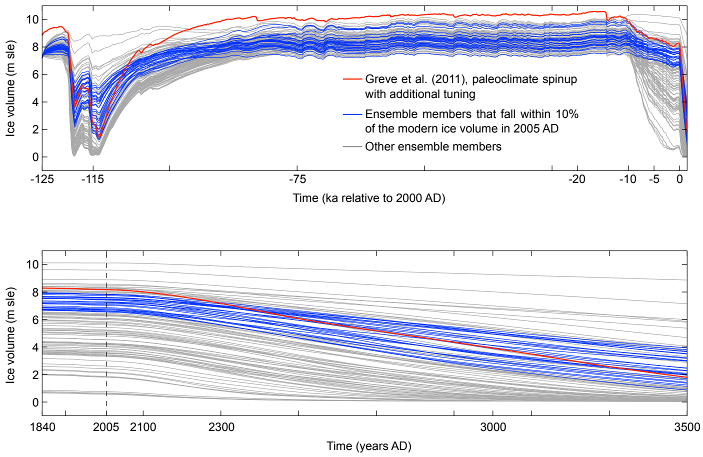

Fig. 3. Simulated ice volumes as a function of time for all 100 ensemble members, expressed in meters of sea level equivalent (m sle). All model runs begin at −125 ka before 2000 AD from the observed modern ice geometry, with an ice volume of ~7.2 m sea level equivalent (Bamber et al., 2001; see text). Blue curves, model runs that give ice volumes within 10 % of estimated modern ice volume in 2005 AD (dashed line). Red line, time evolution of the "paleoclimate spinup with additional tuning" model run described by Greve et al. (2011). Labeled tick marks are those referred to in the text and figures; unlabeled tick marks are 25 ka apart in the top panel and 200 yr apart in the lower panel.

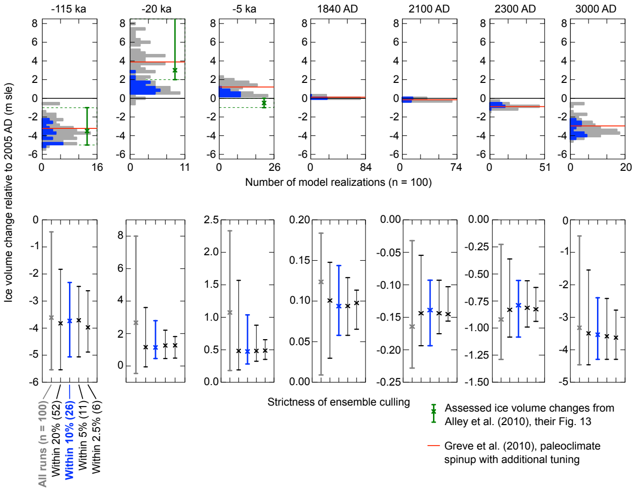

Fig. 4. Histograms of modeled ice volume change relative to 2005 AD (top) and the effects of different assumed uncertainties for the modern ice volume on the median and range of ice volume change projections and hindcasts (bottom). *Y*-axis scaling is the same for all panels in the top row, but differs among panels in the bottom row. Color coding is the same as in Fig. 2; gray, all model runs; dark blue, model runs that lie within 10 % of the estimated modern ice volume in 2005 AD; red line, "paleoclimate spinup with additional tuning" model run from Greve et al. (2011). The green points with error bars in the top panels indicate assessed changes in the ice sheet, relative to the modern, from Alley et al. (2010, their Fig. 13). None of our model runs produce a smaller-than-today ice volume during the mid-Holocene (−5 ka). For the future time slices (2100, 2300, and 3000 AD), the range of potential future ice volume changes is always at least 30 % of the median, regardless of how strictly the ensemble is culled.

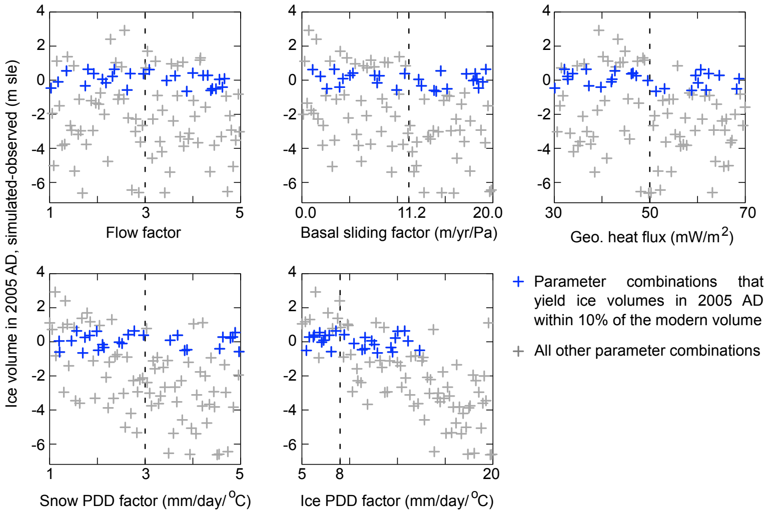

Fig. 5. Relation between input parameter values and simulated ice volumes in 2005 AD, for all ensemble members (gray crosses) and runs that lie within 10 % of the estimated modern ice volume (blue crosses). Vertical dashed lines indicate EISMINT-3 best estimates for most model parameters (Huybrechts, 1998; Stone et al., 2010), except the basal sliding factor, which comes from Greve and Otsu (2007). Influential parameters are indicated by points that are tightly arranged about a curve that dips steeply from one side of the plot to the other (Saltelli et al., 2008). In our ensemble, the ice positive degree-day factor has the greatest influence on simulated modern ice volume, with the snow PDD factor and the basal sliding factor taking second and third places. However, ice PDD factors greater than ∼15 mm day^{-1} °C appear inconsistent with the modern ice volume constraint.

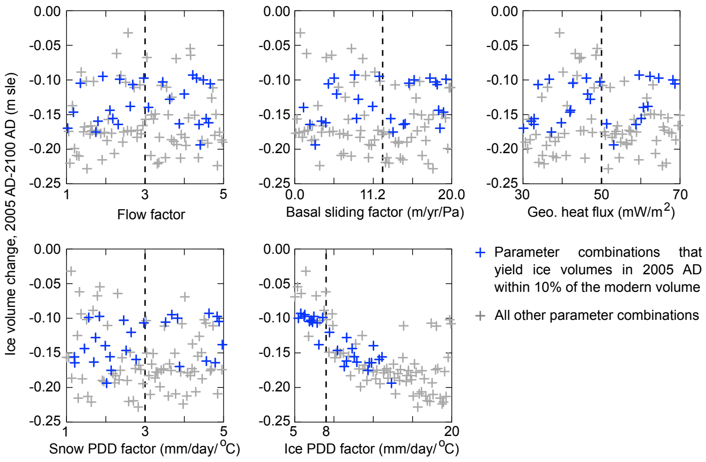

Fig. 6. Relation between input parameter values and simulated ice volume changes between 2005 and 2100 AD (Fig. 3). Color coding is the same as in Fig. 5. The ice positive degree-day factor appears to dominate the short-term future behavior of the model.

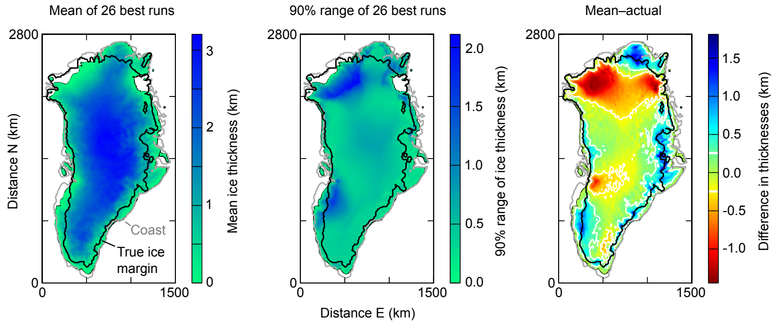

Fig. 7. Comparison of the spatial distribution of mean ice thickness among the 26 "best" model runs to the observed modern ice geometry. By "best" runs, we mean those that give ice volumes within 10 % of the estimated modern ice volume in 2005 AD (Bamber et al., 2001; Sect. 2.2, 3.5). 90 % range, difference between 95th and 5th percentiles of ice thickness values within each grid cell. Black line, modern ice margin; gray line, coast (Bamber et al., 2001, as gridded by Greve et al., 2011); white line, contour bounding areas where the differences between the observed and mean modeled ice thicknesses are less than 250 m. Simulated ice volumes are generally too large near the margins, but there are large areas of too-thin ice in the northern part of Greenland and upflow from Jakobshavn (about a third of the way northward on the western side).