Pre-calibration of a simple Greenland Ice Sheet model for use in integrated assessment studies

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Abstract

The Greenland Ice Sheet is vulnerable to climate warming, possibly resulting in substantial future sea level rise. Integrated assessment models combine treatments of the global economy with simplified treatments of Earth system processes. Such models are used to assess economic impacts of climate change and to identify optimal strategies for responding to climate change (for example). However, many integrated assessment models lack interactive treatments of Greenland Ice Sheet behavior. Here, we adapt a previously-published, simple model of the Greenland Ice Sheet for use in integrated assessment models. The expanded model includes improved treatments of the surface mass balance, heat transport through the ice body, and climate-enhanced basal sliding. We calibrate the model against 1) an ice volume curve from a more-complex model, and 2) data on the ice sheet's past behavior (sea level contributions in the geologic past and historical mass balance estimates). The tuned model successfully matches these data sets. Our results suggest that the expanded model can be a valuable tool for integrated assessment models and sea level studies in general. We also report implications of our study for the tuning of more-complex ice sheet models.

1 Introduction

The Greenland Ice Sheet (GIS) is a major feature of the Arctic and may make an important contribution to future sea level rise. The ice sheet covers an area of $1.7* 10^6$ km^2 (Bamber et al., 2001), and has a maximum elevation of 3.3 km above sea level (Ekholm, 1996). Its reflective surface and height exert an important control over middle to high northern latitude climates (Roe and Lindzen, 2001; Toniazzo et al., 2004; Lunt et al., 2004). If the GIS were to melt completely, global sea level would rise by an average of approximately seven meters (Bamber et al., 2001; Lemke et al., 2007). Although Antarctica holds more ice (~60 m sea level equivalent; Lythe et al., 2001), Greenland is often considered a more immediate concern because large parts of its surface experience melt conditions during the summer (Mote et al., 2007). In contrast, surface melting in Antarctica is largely restricted to the Antarctic Peninsula (Torinesi et al., 2003). Satellite measurements suggest that the GIS mass balance is already negative, and this negative trend may be accelerating (Velicogna, 2009; Alley et al., 2010, and references therein).

Many computer models describe the Greenland ice sheet's behavior (e.g., SICOPOLIS, Greve, 1997; PISM, Bueler and Brown, 2009; Glimmer-CISM, Rutt et al., 2010), and the state of the art has become much more sophisticated since the Intergovernmental Panel on Climate Change's Fourth Assessment Report (Solomon et al., 2007). Much post-2007 model development effort has gone into implementing higher-order treatments of ice flow. Prior to 2007, standard ice sheet models represented ice flow using the shallow-ice approximation (Hutter, 1983). This simplification applies over the bulk of the Greenland ice sheet, where the ice is grounded and flow is relatively slow (see Joughin et al., 2010, for surface velocity maps). However, it fails in ice streams and ice shelves (Kirchner et al., 2011), which are likely the most sensitive parts of the ice sheet. Improved models (e.g., Price et al., 2011; ISSM, Larour et al., 2012; Elmer/Ice, Seddik et al., 2012) provide better representations of ice flow. Other studies improve their models' surface mass balance treatments relative to standard methods (e.g., Otto-Bliesner et al., 2006; Robinson et al., 2010; Fyke et al., 2011).

Integrated assessment models (IAMs) represent the coupled economic-climate system, which could be strongly affected by sea level rise from enhanced Greenland Ice Sheet mass loss. Such models often include a relatively sophisticated economic model attached to simplified treatments of Earth system components (Sarofim and Reilly, 2010). These simplified treatments are tuned to match results from global climate models. Some studies use IAMs to identify optimal (utility-maximizing) balances between economic growth due to fossil fuel consumption and avoidance of negative impacts through emissions mitigation or adaptation to climate change (e.g., Nordhaus, 2008). Other uses include estimation of the social cost of carbon (the cost that should be imposed on carbon dioxide emissions in order to promote economically efficient decision-making; cf. Johnson and Hope, 2012), and construction of future greenhouse gas atmospheric concentration scenarios (e.g., the Representative Concentration Pathways; Moss et al., 2010).

Despite the likely importance of the Greenland Ice Sheet to future sea level change, many integrated assessment models either lack any representation of the GIS, or use treatments that are perhaps oversimplified. For example, the US Interagency Working Group on the Social Cost of Carbon (2010) considered three popular IAMs (DICE, PAGE, and FUND; Nordhaus, 2008; Hope, 2011; Anthoff et al., 2010; Johnson and Hope, 2012). Of these models, only the regionalized version of DICE calculates enhanced mass loss from the Greenland Ice Sheet internally. In that model, yearly sea level rise due to GIS melting is a linear function of the global mean surface air temperature anomaly, and this mass loss decreases exponentially with the stock of ice remaining in the ice sheet (Nordhaus, 2010).

The lack of Greenland Ice Sheet representations in many integrated assessment models suggests a need for a low-order model that captures relevant feedbacks but is quick to run (Fig. 1). Experience suggests that many model evaluations are required to satisfactorily explore parameter space in IAMs (Urban and Keller, 2010; McInerney et al., 2011), given the large number of unknowns associated with projecting future climate and economic development. For example, the Interagency Working Group on Social Cost of Carbon (2010) used 10^4 runs per emissions scenario, and Moles et al. (2004) give a figure of 3.5* 10^5 runs. Assuming a reasonable computing time of six months, these figures suggest a maximum ice sheet model execution time between 0.75 and 26 minutes, multiplied by the number of available computer processors. These times assume that the Greenland component dominates the overall computational cost of the model, but integrated assessment models include many other components that must be evaluated each time the model is run.

Carrying out 10^4-10^5 model runs with a three-dimensional ice sheet model would be prohibitively expensive (Fig. 1). For example, spinning up the three-dimensional, shallow-ice-approximation model SICOPOLIS (Greve, 1997; sicopolis.greveweb.net) requires ~1.5

days on one computer processor, with each spinup run covering 125,000 years (q.v. Applegate et al., 2012). This long spinup allows the modeled ice sheet to achieve a thermal state that is consistent with climate history (Rogozhina et al., 2011; Bindschadler et al., in review 2011), and the spinup must be repeated for each new parameter combination that is investigated. Thus, SICOPOLIS is ~80-3,000 times more expensive to run than the permissible upper limit for integrated assessment modeling. SICOPOLIS is a shallow-ice model, meaning that it achieves speed through the neglect of important stresses within the ice body (Kirchner et al., 2011). More-complex models that represent these stresses (e.g., Price et al., 2011) would presumably require even more computing time.

The problem is simplified somewhat because integrated assessment models only require estimates of ice volume change over time; accurate simulation of the geographic distribution of ice is not needed. As they are represented in IAMs, most of the impacts associated with enhanced mass loss from ice sheets are caused by sea level rise (e.g., Nicholls et al., 2008). Direct effects from global sea level rise in Greenland itself would be limited; large-scale mass loss from an ice sheet causes local sea level fall, due to gravitational effects (e.g., Mitrovica et al., 2009; Gomez et al., 2010).

Here, we show that the GLISTEN (GreenLand Ice Sheet ENhanced) model meets the speed criterion outlined above (Fig. 1), and reproduces ice volume trajectories from the three-dimensional ice sheet model SICOPOLIS. We also calibrate the model using geological data and modern observations, including 1) estimates of the ice sheet's contribution to sea level change at different times in the past (Alley et al., 2010), 2) historical mass balance estimates (Rignot et al., 2008), and 3) the modern ice sheet profile (Letreguilly et al., 1991).

GLISTEN is a Fortran port of an Excel spreadsheet model intended for classroom use (GRANTISM, the GReenland and ANTarctic Ice Sheet Model; Pattyn, 2006). We change the name of the port because GLISTEN does not treat the behavior of the East Antarctic ice sheet, which the predecessor model GRANTISM does. Beyond porting the model, we add improved treatments of the ice sheet's surface mass balance, heat transport, and climate-induced enhanced flow.

The paper proceeds as follows. Section 2 provides a brief description of the predecessor model GRANTISM (Pattyn, 2006) and indicates how the GLISTEN port treats various processes that are important to the real ice sheet. Section 3 describes a precalibration exercise in which we match GLISTEN to an ice volume curve from a three-dimensional model

(SICOPOLIS; Greve, 1997) and to various observational data sets. Finally, Section 4 places these results in a wider context and concludes the paper.

2 Model description

2.1 The predecessor model GRANTISM

GRANTISM, the GReenland and ANTarctic Ice Sheet Model, describes the response of the Greenland and Antarctic ice sheets to climate change (Pattyn, 2006). The model treats cross-sections through both ice sheets; for the Greenland domain, this transect follows the 72nd parallel. The model is easy to use and is implemented in Microsoft's widely-available Excel (tm) spreadsheet software.

To start GRANTISM (Pattyn, 2006), the user specifies the ice sheet's initial state (either the present-day ice geometry or an ice-free, relaxed-bedrock state), and the surface air temperature anomaly relative to the present-day. The user can also disable many processes that are normally active in the model, such as basal sliding, isostatic adjustment to ice loading changes, and changes in background sea level. The user then advances the model in time, with one time step elapsing for each keystroke. As model time advances, four panels on the model's graphical user interface show changes in ice thickness, bedrock elevation, velocity (total and basal sliding-only), surface mass balance (total mass balance, accumulation, and ablation), and surface air temperatures (mean annual and mean summer). The model responds in a reasonable way to user choices; for example, setting the surface temperature anomaly greater than zero causes the ice sheet to shrink.

GRANTISM (Pattyn, 2006) contains many features of research-grade ice sheet models. In particular, the model solves the equations describing ice flow using finite-difference methods (Pattyn, 2006; see also Hindmarsh, 2001; Greve and Calov, 2002; Greve and Blatter, 2009) much like those employed in more-complex ice sheet models (e.g., SICOPOLIS; Greve, 1997). Besides ice flow, GRANTISM captures the key insight that ice sheet changes depend primarily on surface air temperatures, snowfall, and the instantaneous state of the ice sheet. The model's treatment of a profile through the ice sheet is not necessarily a fatal oversimplification; see Parizek and Alley (2004) and Parizek et al. (2005) for an example of a research-grade profile model.

GRANTISM (Pattyn, 2006) is a user-friendly tool for teaching about the behavior of ice sheets. However, some modifications are needed before GRANTISM can be incorporated into integrated assessment studies. The model's implementation in Excel makes it difficult to couple to other Earth system components, which are typically written in high-level programming languages such as Fortran. Because it accepts just one surface air temperature anomaly value at a time, GRANTISM is best suited for examining the steady-state, equilibrium characteristics of the ice sheet under different climate states (e.g., Pattyn, 2006, his Fig. 4). Using GRANTISM to determine time-dependent changes in the ice sheet requires the model to accept input files describing surface air temperature and sea level anomalies over time.

2.2 The updated GLISTEN model

As noted above, GLISTEN is primarily a Fortran port of the predecessor model GRANTISM (Pattyn, 2006). Many process descriptions in GLISTEN are closely similar or identical to those in GRANTISM; for example, we retain GRANTISM's semi-implicit finite-difference methods for solving the ice flow equations. However, our use of Fortran, with an R wrapper for handling input and output tasks, improves the functionality of the model and makes it possible to couple GLISTEN to other models. We also update the model's surface mass balance treatment and add a parameterization of climate-induced enhanced basal slip (described below).

In the remainder of Section 2, we show how GLISTEN handles various ice sheet processes, being careful to point out similarities and differences between GLISTEN and GRANTISM (Pattyn, 2006). Each subsection begins with a brief description of how the process being discussed works on the real ice sheet, for the benefit of scientists from outside the cryosphere community. Hooke (2005), Greve and Blatter (2009), and Rutt et al. (2009) give more complete descriptions of ice sheet processes and model treatments.

2.2.1 Surface mass balance

At any given point on the Greenland Ice Sheet, the surface mass balance is the difference between the rate of mass addition by snowfall and the rate of mass loss from melting, sublimation, and wind erosion. The surface mass balance is positive on the central parts of the ice sheet, where low surface air temperatures prevent melting, and negative around the margins. Integrating surface mass balance and calving (see below) over the ice sheet's area gives the ice sheet's instantaneous total mass balance, or its mass change per unit time.

2.2.1.1 Accumulation

The amount of yearly snowfall on the Greenland ice sheet is known approximately from field measurements (e.g., Ohmura and Reeh, 1991; Bales et al., 2006), and likely changes with background surface air temperature (see review in van der Veen, 2002). The thicknesses of ice layers in ice cores, and the oxygen isotope values in the same ice layers, provide paired estimates of the amount of accumulation as a function of surface air temperature anomaly. In such records, there is a clear contrast between the cold, low-accumulation times of the last glacial period and the Younger Dryas, and the warmer, higher-accumulation Holocene. However, the surface air temperature-accumulation relationship breaks down during the Holocene itself (Cuffey and Clow, 1997). Climate models also give widely diverging estimates of how much accumulation on Greenland should change with surface air temperature (van der Veen, 2002; cf. Gregory et al., 2006).

GLISTEN treats accumulation a_{acc} (m yr^{-1}) as a function of the modern-day annual precipitation averaged over the model profile $\bar{a}_0$ (m yr^{-1}) and the instantaneous surface air temperature anomaly T_f ($^{\mathrm{O}}$C),

$$a_{acc} = \bar{a}_0 \cdot s^{T_f} \text{ for } T_f < 0, \qquad (1)$$

$$a_{acc} = \bar{a}_0 \text{ for } T_f \geq 0$$

(cf. Pattyn, 2006, his Eqn. 13). Both $\bar{a}_0$ and s are tuneable parameters with default values of 0.41 m yr^{-1} and 1.0533 (unitless; Clausen et al., 1988; Huybrechts and de Wolde, 1999). The modern-day precipitation values come from previously-published experiments with the regional climate model RACMO (Ettema et al., 2009) compiled by the seaRISE project (Bindschadler et al., in review 2011).

This approach closely imitates GRANTISM's (Pattyn, 2006), except that the predecessor model uses a second-order polynomial fit to data from Ohmura and Reeh (1991) instead of our spatially-constant prefactor.

Because Equation 1 use the profile-averaged modern accumulation as a prefactor, GLISTEN sets accumulation to a constant value everywhere over the model profile. On the real ice sheet, accumulation is greater around the ice sheet's margins than on the central parts of the ice sheet (Bales et al., 2006; Ettema et al., 2009). The original GRANTISM second-order polynomial fit shares this problem; its second-order polynomial fit has a maximum in the center of the ice sheet and declines to zero near the ice margin (Pattyn, 2006, his Fig. 3). Our approach also presumes that all precipitation falls as snow (see Stone et al., 2010, for a parallel example). In practice, ~40% of all precipitation over the Greenland Ice Sheet falls as rain (Bales et al., 2006), which may or may not refreeze in the snow pack (Marsiat et al., 1994; Reijmer et al., 2012). This treatment also prevents accumulation from going above present-day values (Pattyn, 2006; cf. Greve et al., 2011), which may be correct; we have little basis for estimating precipitation change for warmer-than-present climate states.

These simplifications likely bias the model's surface mass balance toward more positive values, other factors being equal. The spatially-constant accumulation field likely increases the amount of precipitation that falls on the ice sheet, instead of on unglaciated land or the open ocean. Assuming that all precipitation falls as snow overestimates accumulation over the whole ice sheet by a factor of ~1.7 (Bales et al., 2006). However, tuning of the model should allow reasonable representation of the average contribution of these processes to the ice sheet's mass balance.

2.2.1.2 Ablation

As noted above, local ablation on the Greenland Ice Sheet is the sum of mass losses from melting, sublimation, and wind erosion. We are unaware of any systematic estimates of wind erosion, which is sensitive to small-scale topography. Sublimation is usually associated with high-altitude, low-latitude glaciers that receive little precipitation (e.g., Rupper and Roe, 2008), but also happens on the Greenland ice sheet (Box et al., 2001). However, melting dominates sublimation when the whole ice sheet is considered (Ettema et al., 2009).

GLISTEN calculates ablation using the positive degree-day method, which is common in modeling of ice sheets (e.g., Greve et al., 2011) and small, alpine glaciers (e.g., Anderson and Mackintosh, 2006). The positive degree-day approach involves determining the integral of surface air temperature deviation above 0 ^{O}C over a time period of interest (usually a year) and multiplying by a constant, the positive degree-day factor f_{PDD}. Specifically, we apply the method of Calov and Greve (2005), which relates mean annual and mean July surface air temperatures to the number of positive degree-days. To obtain these surface air temperature estimates, we add the background temperature anomaly to the parameterizations of Fausto et al. (2009). We treat the positive degree-day factor as a tuneable parameter in GLISTEN.

This method assumes that the positive degree-day factor is constant over the whole ice sheet and through time. In reality, the positive degree-day factor varies with latitude (Braithwaite, 1995; Tarasov and Peltier, 2002) and surface character; snow has a lower positive degree-day factor than ice. We also assume that all meltwater runs off immediately, instead of refreezing in the snowpack (Reeh, 1991; Reijmer et al., 2012).

The use of a positive degree-day method for calculating surface ablation represents a small improvement over the predecessor model GRANTISM (Pattyn, 2006). In particular, the Calov and Greve (2005) method allows for variation among summer-month surface air temperatures and for daily variability, meaning that some ablation will occur on the modeled ice sheet even when surface air temperature anomalies are quite negative. In contrast, there is a definite surface air temperature anomaly threshold in the GRANTISM model below which no ablation happens at present sea level. This cutoff occurs at a surface air temperature anomaly of -7.29 ^{O}C (Pattyn, 2006, his Eqns. 10 and 15), well within the range of surface air temperature anomalies experienced by the Greenland ice sheet over the last glacial-interglacial cycle (Cuffey and Clow, 1997).

Despite our use of positive degree-days to calculate melting, the ablation treatment used in GLISTEN is still highly simplified compared to those used in many other ice sheet models. These simplifications probably lead to overestimates of ablation, other factors being equal. The real ice sheet begins the ablation season covered with snow, which has a higher albedo and a lower positive degree-day factor than ice. Thus, using a constant, ice-appropriate positive degree-day factor will overestimate ablation during the early part of the melt season. Moreover, the model does not track refreezing in the snowpack, meaning that water that would normally refreeze runs off instead. Again, model tuning should allow us to compensate for these simplifications.

2.2.2 Heat transport

Heat is transported through the real ice sheet by both diffusion and advection. Sources of heat include the atmosphere, geothermal heating, and mechanical sources such as deformation of the ice and its substrate and the passage of water through englacial tunnels and vertical moulins.

GLISTEN uses separate treatments to calculate temperatures within and beneath the ice sheet. Temperatures within the ice sheet T_i (K) are calculated from the surface air temperature anomaly T_f (^OC) in the same way as in GRANTISM,

$$T_i = T_f + 263.15 \text{ for } T_f < 0, \qquad (2)$$

$$T_i = 0.5T_f + 263.15 \text{ for } T_f \geq 0$$

(Pattyn, 2006, his Eqn. 6). This expression implies that temperatures within the ice body vary only with time, through variations in the surface air temperature anomaly T_f. Because this treatment contains no time-dependent component, changes in the surface air temperature anomaly are immediately reflected in the ice's resistance to flow (see below).

GLISTEN represents temperatures at the base of the ice sheet T_b (^OC) as

$$T_b = T_{ma}\operatorname{erfc}\left(\frac{H}{2\sqrt{kt}}\right) + q_G, \qquad (3)$$

where κ, the diffusion coefficient of ice, is given by

$$\kappa = \frac{k_i}{\rho_i C_i}. \qquad (4)$$

Here, T_{ma} is the mean annual surface air temperature (^OC), H is the ice thickness (m), t is time (yr), q_G is a tuneable constant term that represents geothermal heating (^OC), k_i is the thermal conductivity of ice (2.2 W m^{-1} K^{-1}), ρ_i is the density of ice (917 kg m^{-3}), and C_i is the heat capacity of ice (2000 J kg^{-1} K^{-1}). T_b, T_{ma}, and H are all functions of distance along the profile x.

This treatment assumes that all heat transport takes place by diffusion from the surface of the ice sheet, neglecting advection of heat due to ice flow. Given an arbitrarily long time and constant surface air temperatures, bedrock surface elevations, and ice thicknesses, this expression yields a decrease in basal temperatures from the ice margin to the center of the ice sheet. The constant term q_G adjusts where this curve intersects the zero-degree line along the model transect, and thus the fraction of the bed over which sliding is permitted to take place (no sliding occurs where the bed is frozen).

This treatment of basal temperatures improves on GRANTISM (Pattyn, 2006), which handles heat transport implicitly. However, the treatment is still highly simplified relative to that used in three-dimensional ice sheet models, which treat both diffusion and advection.

2.2.3 Ice flow and basal sliding

Ice deforms in response to applied stresses, but ice temperature, water content (e.g., Greve, 1997; Aschwanden et al., 2012), the orientation of crystal axes, and the presence or absence of impurities, also affect ice flow. Ice deformation is proportional to the third power of the applied stress (Glen, 1955). This driving stress is large where the ice is thick and/or surface slopes are steep, and lower elsewhere (Alley et al., 2010). Warm ice deforms more readily than cold ice. Where ice crystals have a preferred orientation, flow occurs more readily along that direction than predicted by the "normal" equations describing ice flow. Finally, impurities generally soften the ice relative to the pure material studied in the laboratory.

Where grounded ice is not frozen to its bed, it can slide. The effectiveness of this process depends on the areal concentration and size of asperities (e.g., Weertman, 1957), as well as basal water pressure and sediment availability. The correct form of the basal sliding law is under discussion (for a review, see Alley, 2000).

Real ice sheets can be divided into three distinct flow domains, based on their velocities and whether the ice is in contact with a solid substrate (Kirchner et al., 2011; see Joughin et al., 2010, for surface velocity maps of the Greenland Ice Sheet). Normal, grounded ice moves slowly, often a few meters per year or less. Ice streams (e.g., the Northeast Greenland Ice Stream; Fahnestock et al., 2001) have higher surface velocities, up to many kilometers per year, and rest on a slippery till substrate. Finally, ice shelves consist of formerly-grounded ice that is now afloat, although they are still connected to their parent land ice bodies. Because ice streams and ice shelves lack a resistant substrate, their stress balances are closely similar to one another and different from that of grounded ice (Kirchner et al., 2011; see also Bueler and Brown, 2009).

GLISTEN’s treatment of ice flow is nearly identical to GRANTISM’s (Pattyn, 2006). The conservation of matter and the stress-strain relationship for ice yield the vertically-integrated velocity due to deformation within the ice body, accounting for temperature-based differences in ice viscosity (Eqn. 2, above). We multiply the ice deformation velocity with a tuneable, dimensionless factor d that is analogous to the ice flow enhancement factor used in three-dimensional ice sheet models (e.g., Rutt et al., 2009). We then find the change in ice thickness in each model grid cell per time step using the total horizontal velocity, including basal sliding. The translation of velocities into thickness changes is accomplished using a semi-implicit finite-difference technique (Pattyn, 2006; see also Hindmarsh, 2001; Greve and Calov, 2002; Greve and Blatter, 2009).

GLISTEN’s method for calculating the basal sliding velocity u_b (m yr^{-1}) comes from Hindmarsh and le Meur (2001; see also Greve, 2005),

$$u_b = be^{(T_b/\gamma)}\left[\frac{\tau_d^p}{(\rho_i gH)^q}\right], \tag{5}$$

where b is a dimensionless tuning factor, T_b is the basal temperature from eqn. 3 (^{O}C), γ is a sub-melt sliding parameter (1 ^{O}C), g is the gravitational acceleration (9.81 m s^{-2}), and p and q are sliding law exponents (3 and 2, respectively). The driving stress τ_d is defined as

$$\tau_d = -\rho_i gH\nabla h \tag{6}$$

(Pattyn, 2006, his Eqn. 1), where h is the elevation of the ice surface.

Possible objections to this ice flow-basal sliding treatment are that it is calculated on a relatively coarse, one-dimensional grid (Δx = 36 km; Pattyn, 2006), and that it depends on the shallow-ice approximation (Hutter, 1983; Pattyn, 2006; Kirchner et al., 2011). The model assumes that all flow follows the defined transect, thereby neglecting branches in the flow field (see Parizek and Alley, 2004, for another example of this approach). The shallow-ice approximation is appropriate for grounded ice, but cannot capture the larger ice velocities associated with ice streams. Much effort is presently being devoted to developing ice sheet models that do not have this limitation (e.g., Pollard and DeConto, 2009, 2012; Price et al., 2011; Larour et al., 2012; Leng et al., 2012). Both branching of the flow field and enhanced velocities due to ice streams will likely become more pronounced if the ice sheet begins to decay. Finally, some parts of the ice sheet may be more vulnerable to mass loss than the single transect we have chosen (e.g., Born and Nisancioglu, 2012). Thus, GLISTEN's treatment of ice flow and basal sliding likely underestimates ice transport from the accumulation area to the marginal ablation zones without appropriate tuning.

2.2.4 Climate-enhanced ice transport

Given that our ice flow-basal sliding treatment likely underestimates future increases in ice delivery to the margins, we incorporate a parameterization into GLISTEN that allows ice fluxes to increase with climate warming. This parameterization is inspired by the so-called "Zwally effect" (Zwally et al., 2002) and recent model treatments of it (Parizek and Alley, 2004; Greve and Otsu, 2007). More generally, this parameterization is a qualitative representation of the possible "future... dynamical changes in ice flow" identified by the Intergovernmental Panel on Climate Change's Working Group 1 (2007, their table SPM3).

Specifically, we multiply the basal velocity u_b (Eqn. 5) by a tuneable prefactor Z_f wherever surface ablation exceeds accumulation (Section 2.2.1). Thus, as surface air temperatures rise and the fraction of the ice sheet surface that is in the ablation zone increases, the zone of enhanced marginal flow also grows (Parizek and Alley, 2004).

Considered solely as a representation of the Zwally effect, this parameterization neglects much of what is known about subglacial hydrology. Conceptually, the Zwally effect involves the penetration of surface meltwater to the bed, lubricating it and resulting in larger annually-integrated ice velocities. Theoretical and modeling work show that distributed basal hydrologic networks, which enhance ice flow, collapse readily to dendritic networks that do not contribute to ice speedup (e.g., Röthlisberger et al., 1972; Bartholomew et al., 2010; Schoof, 2010; cf. Gulley et al., 2012). Thus, the Zwally effect itself likely affects ice flow only during the beginning of the melt season, and may have little effect on annually-integrated ice fluxes. However, this simple parameterization compensates, to some extent, for the lack of higher-order ice flow dynamics in GLISTEN.

2.2.5 Other processes

GLISTEN handles isostatic adjustment of the bedrock surface in the same way as GRANTISM (Pattyn, 2006). Given a change in the thickness of ice in a given model grid cell, the bedrock surface in that grid cell relaxes toward its new elevation with a characteristic time scale θ. This time scale is a tuneable parameter in GLISTEN, and has a default value of 3,000 years. This treatment neglects changes in ice thickness in adjacent grid cells, as considered by elastic-lithosphere methods (e.g., Greve and Blatter, 2009).

In GLISTEN, ice that advances into a grid cell with a nonzero water depth simply calves. This treatment clearly neglects the possibility of ice shelf formation (although ice shelves make up a small fraction of the modern Greenland Ice Sheet's area), the penetration of warm ocean water into fjords around the ice sheet margin (e.g., Straneo et al., 2010), and the complexities associated with grounding line migration (e.g., Alley et al., 2007).

2.2.6 Conversion of simulated area to Greenland ice volume

Sea level rise studies require the time evolution of ice volume on Greenland. However, neither GRANTISM (Pattyn, 2006) nor GLISTEN gives this information directly; instead, they yield the cross-sectional area of ice over the modeled transect at any instant in time. To convert this area to ice volume, we multiply by the ratio of the total modern ice volume (7.3 m sle; Lemke et al., 2007) to the modern ice area of the transect (see Parizek and Alley, 2004, for another example of this approach). This conversion is likely most accurate for small volume changes, relative to the present day.

3 Precalibration of the updated GLISTEN model

3.1 Motivation and methods

Our goal in porting and updating GRANTISM (Pattyn, 2006) is to create a representation of the Greenland Ice Sheet for integrated assessment models. To be useful in this context, GLISTEN must

1) reproduce a curve of ice volume as a function of time from a more-complex ice sheet model, and

2) match data on the ice sheet's past behavior and present geometry.

Although these criteria may appear redundant, one does not imply the other. Criterion #1 is based on the needs of integrated assessment models and the standards used in the integrated assessment literature for evaluating different model components. As noted in the Introduction, integrated assessment models require estimates of ice volume change over time, not the spatial distribution of ice. Individual components of integrated assessment models are typically calibrated against more-complex models.

Criterion #2 acknowledges that most Greenland Ice Sheet models are tuned solely against the shape of the modern ice sheet, and occasionally its surface velocity field (Aschwanden et al., 2009; Bindschadler et al., in review 2011; for exceptions, see Tarasov and Peltier, 2003; Lhomme et al., 2005; Simpson et al., 2009). Reproducing a static "snapshot" of the modern ice sheet raises questions about whether the tuned models will behave appropriately when forced from this estimated modern state (Oreskes, 1994). Thus, we tune GLISTEN separately using time-distributed data.

To address the two criteria given above, we use a search algorithm to adjust GLISTEN's eight tuneable parameters (Table 1) until a good match is found between our selected tuning data sets and the model output. Based on the generally good matches that we identify (Section 3.2, below), we conclude that GLISTEN is a promising tool for incorporating insights on Greenland Ice Sheet behavior into integrated assessment models. In the remainder of Section

3.1, we describe the data sets that we tune the model against, our search algorithm, and our objective function for determining the quality of model fits to data.

3.1.1 Tuning data sets

To tune GLISTEN against a more-complex ice sheet model (criterion #1, above), we use the ice volume(time) curve from run #29 of a 100-member perturbed-parameter ensemble (Applegate et al., 2012) produced with the SICOPOLIS ice sheet model (Greve, 1997; Greve et al., 2011; sicopolis.greveweb.net). This ensemble member provided the best agreement with the modern ice volume (~7.2 m sea level equivalent, integrated over SICOPOLIS' 10-km grid; Bamber et al., 2001; Greve et al., 2011; cf. Lemke et al., 2007). As recommended by the seaRISE project (Bindschadler et al., in review 2011; http://websrv.cs.umt.edu/isis/index.php/SeaRISE_Assessment), the SICOPOLIS ensemble was forced by surface air temperature anomalies derived from oxygen isotopes in the GRIP ice core (Dansgaard et al., 1993) from 125 ka to preindustrial times (1840). The sequence of oxygen isotope measurements in this ice core is disturbed by flow before ~90 ka (e.g., Chappellaz et al., 1997). After 1840, the ensemble was driven by observed surface air temperatures from Vinther et al. (2006). The ensemble was also forced by background sea levels estimated from oxygen isotope values measured in planktonic foraminifera from ocean sediment cores (Imbrie et al., 1984).

To independently match GLISTEN to data on the ice sheet's past behavior and present shape, we use 1) assessed ice volume changes, relative to the present, during key periods in the last glacial-interglacial cycle (Alley et al., 2010, their Fig. 13), and the modern ice volume (Bamber et al., 2001); 2) estimates of the ice sheet's total mass balance in five individual years during the last six decades (Rignot et al., 2008); and 3) ice thicknesses along the model transect (Letreguilly et al., 1991). The Letreguilly et al. (1991) data set has been superseded by subsequent compilations (Bamber et al., 2001) and additional data collection; we use it here because it provides the basal boundary and initial condition for GRANTISM (Pattyn, 2006) and GLISTEN.

3.1.2 Search algorithm and objective function

We use Differential Evolution (Storn and Price, 1997; Price et al., 2007) to identify parameter combinations that produce good matches between model output and the data sets described above. Differential Evolution is a genetic algorithm that is widely used in optimization problems. It generates successive generations of model parameter combinations, testing each combination for "fitness" according to a user-defined objective function. Differential Evolution requires few evaluations to identify an optimal solution, and is less likely to become "stuck" in a local minimum of the response surface than gradient descent methods. We found that 8,000-11,000 model realizations per optimization were required to achieve good results with Differential Evolution for our problem. Given that each model run covers about 125,000 model years, this number of evaluations implies at least one billion (10^9) model years per calibration experiment.

Our objective function uses the product of Gaussian likelihoods, with a correction for autocorrelated residuals where appropriate (see discussion in Olson et al., 2012). Each individual "data point" i has a central estimate μ_i and an uncertainty σ_i, and these two parameters define a normal distribution for that data point. A model-produced value for the same quantity v_i will fall some distance from the best-estimate value μ_i, and the "correctness," or likelihood, of the model realization predicated on just data point i is calculated from the offset between the central estimate and the model prediction. Larger offsets, which indicate a worse fit to the observations, receive a smaller likelihood. The product of the likelihoods for all data points $i = 1, 2, \ldots n$ is then an estimate of the "correctness" of the model run as a whole, given the available data. In practice, we sum the logarithms of our Gaussian likelihoods, to avoid computer underflow errors.

We assign central estimates and uncertainties to our tuning data sets as follows.

SICOPOLIS emulation: We extract central estimates from SICOPOLIS' hindcast ice volumes (run #29 from Applegate et al., 2012) over two periods in the geologic past, plus the simulated modern ice volume. For the Eemian and the Last Glacial Maximum, we average these simulated ice volumes over the periods 118.5-115 ka and 20-19 ka. The actual Eemian warm period is somewhat older, with maximum ice loss from Greenland occurring ~125 ka (Kopp et al., 2009). However, our SICOPOLIS run is driven by the GRIP oxygen isotope curve, which has a quasi-Eemian warm period at ~118.5-115 ka. We assign an uncertainty of one meter sea level equivalent to all three "data points" from the SICOPOLIS ice volume curve.

Assessed ice volume changes: This set of constraints is similar to that described under "SICOPOLIS emulation," above, but uses ice volume changes relative to the present day from Alley et al. (2010, their Fig. 13) for the Eemian and Last Glacial Maximum.

Historical mass balance: Rignot et al. (2008) provide estimates of the Greenland Ice Sheet's total mass balance, with uncertainties, during six years covered by our model runs (1958, 1964, 1996, 2000, 2004, and 2005). Our runs do not include the years 2006 and 2008, also covered by Rignot et al. (2008), because the Vinther et al. (2006) surface air temperature record that drives the model ends in 2005.

Ice thicknesses: The data set of Letreguilly et al. (1991) provides the basal boundary condition and initial ice thicknesses for both GRANTISM (Pattyn, 2006) and GLISTEN (see Bamber et al., 2001, for an updated data set). We evaluate the likelihood of model parameter combinations for this data set using the method of Olson et al. (2012), which accounts for autocorrelated residuals.

3.1.3 Initial conditions, forcing functions, and time steps

For the pre-calibration experiments, we ran GLISTEN over two periods, 125 ka to 1840 and 1840 to 2005. The initial condition for the paleo-spinup (125 ka-1840) was the modern ice thicknesses and bedrock topography, as given by Letreguilly et al. (1991) and projected onto the model transect by Pattyn (2006). For each model run, the final state from the paleo-spinup provided the initial state for the historical part of the run (1840-2005).

As in Applegate et al. (2012; see above), the forcing functions for the paleo-spinup period were surface air temperature anomalies derived from the GRIP ice core (Dansgaard et al., 1993) and sea level anomalies based on ocean cores used in the SPECMAP project (Imbrie et al., 1984). After 1840, we used surface air temperature anomalies from Vinther et al. (2006) to drive the model. The sea level anomaly was held constant at 0 over the historical period, but this simplification should have little or no effect on our results (Applegate et al., 2012).

The time step over the paleo-spinup was 20 years, and this time step was shortened to 1 yr for the historical period.

3.2 Precalibration results

As noted above, we perform two precalibration experiments with GLISTEN. The first of these experiments matches GLISTEN to an ice volume curve from the SICOPOLIS ice sheet model (run #29 from Applegate et al., 2012). The second experiment matches GLISTEN to assessed ice volume changes, historical mass balance data, and modern ice thicknesses.

GLISTEN reproduces the data reasonably well, given appropriate tuning (Figs. 2-5). For example, GLISTEN matches the overall shape of the ice volume curve from SICOPOLIS when the two models are forced by the same surface air temperature and sea level anomaly curves (Fig. 2). Similarly, GLISTEN agrees well with assessed ice volume changes over geologic time (Fig. 3), mass balance estimates covering the last few decades (Fig. 4), and the observed ice profile (Fig. 5).

The best-fit values from the two experiments diverge widely for many parameters (Table 1). Given the general resemblance of the ice volume curves resulting from these experiments (compare Figs. 2 and 3), these differences may seem surprising. However, ice sheet models can yield similar results for key model outputs, even given very different input parameter combinations (Applegate et al., 2012, their Figs. 1, 3, and 9). Some parameters trade off against one another; for example, similar velocities can arise from either a high value of the ice deformation parameter *d* and a low value of the basal sliding parameter *b*, or vice versa. Further work is needed to characterize the likelihood surfaces near our best-fitting parameter combinations and potential tradeoffs among parameters.

4 Discussion

We have shown that the one-dimensional ice sheet model GLISTEN reproduces 1) an ice volume curve from a more-complex model, and 2) data on the ice sheet's past behavior and present geometry, after appropriate tuning. GLISTEN is a Fortran port of an earlier, spreadsheet-based model developed for classroom use (GRANTISM; Pattyn, 2006). We improved the ported model's treatments of surface mass balance, heat transport within the ice body, and climate-enhanced ice flux.

GLISTEN's success in meeting these tests suggest that it may be a useful tool for integrated assessment modeling and sea level studies in general. As noted in the Introduction, most integrated assessment models lack an intrinsic treatment of Greenland Ice Sheet melt, despite its potential importance for future sea level rise. GLISTEN runs quickly and successfully imitates other models, thereby satisfying two important criteria for new integrated assessment model components. In terms of sea level studies, GLISTEN may occupy an important niche between semi-empirical projections of future sea level rise (e.g., Rahmstorf, 2007; Grinsted et al., 2009) and three-dimensional, higher-order ice sheet models (e.g., Price et al., 2011; Bindschadler et al., in review 2011; Seddik et al., 2011; Larour et al., 2012).

GLISTEN's speed proved vital for calibrating the model against observations of the ice sheet's past behavior, raising implications for the calibration of other ice sheet models. Each of our two calibration experiments required 8,000-11,000 model evaluations covering the period 125 ka to present. These simulations took a few hours with GLISTEN, but a three-dimensional model like SICOPOLIS would require decades to carry out the same experiment on a single processor (Fig. 1). SICOPOLIS is a shallow-ice approximation model that trades mechanistic complexity for speed; higher-order ice sheet models, with more parameters and increased computational cost, simply cannot be calibrated in this way. Instead, such models are tuned primarily against the modern state of the ice sheet (e.g., Bindschadler et al., in review 2011). Tuning a model solely against the ice sheet's present state raises questions about whether the model behavior is reasonable when forced away from the present climate (e.g., Oreskes et al., 1994). Our tuning exercises with GLISTEN demonstrate that the model reproduces the expected amplitude of ice volume change between the Eemian, Last Glacial Maximum, and the present day, and simulates recent mass balance changes.

As we emphasize in the model description (Section 2), GLISTEN lacks many processes that may be important to the real ice sheet. For example, it does not have any treatment of advanced dynamics beyond a qualitative parameterization of climate-induced enhanced flow (Section 2.2.4), and the profile treatment means that any complexities in the flow field are not captured (see Sergienko et al., 2012, for a discussion of profile models and their limitations). GLISTEN best describes feedbacks associated with the ice sheet's surface mass balance, although many important details of these processes are also not treated by the model.

Future work with GLISTEN will involve futher testing and calibration, and incorporation into integrated assessment models. At present, the model uses the outdated Letreguilly et al. (1991) bedrock topography and ice thickness data set as its basal boundary and initial condition; this data will be replaced by the Bamber et al. (2001) compilation, and the effects of different horizontal spacings on the model results will be investigated. We will investigate how using transects other than the one chosen by Pattyn (2006; 72^{O} N) affects GLISTEN's ability to reflect the whole ice sheet's behavior. Finally, our present calibration identifies only best-guess parameter values, hindcasts, and projections. Probabilistic calibration with Markov chain Monte Carlo (e.g., Olson et al., 2012) will allow characterization of uncertainties associated with these quantities.

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Author contributions

BCT ported the original GRANTISM model to Fortran and wrote a preliminary R wrapper to control model input and output. JDHM modified BCT's Fortran code to include additional processes, developed R code to match the model to data, and wrote the initially-submitted version of the manuscript. PJA provided information on ice sheet processes and revised the paper text according to the reviewers' suggestions. REN produced Figure 1. KK designed the study and provided insight into integrated assessment modeling and sea level rise studies. All authors participated in discussions.

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Table 1. Tuneable parameters in the updated GLISTEN model and their best-estimate values in our two Differential Evolution tuning experiments (Section 3).

Parameter	Description	Search range and units	First occurs	Tuning to SICOPOLIS	Tuning to observations
s	Factor controlling accumulation change with temperature	0.1 to 10 (unitless)	2.2.1.1	1.1693	1.0841
a_0	Profile-averaged yearly precipitation	0.205 to 0.82 m yr^{-1}	2.2.1.1	0.6161	0.4119
f_{PDD}	Positive degree-day factor	-2.5* 10^{-3} to -1.0* 10^{-2} m day^{-1} $^{\mathrm{O}}$C^{-1}	2.2.1.2	-2.9* 10^{-3}	-2.6* 10^{-3}
q_{G}	Geothermal heating term	-10 to 10 $^{\mathrm{O}}$C	2.2.2	-7.5545	2.0214
d	Ice flow factor	0.1 to 10 (unitless)	2.2.3	6.9947	1.5355
b	Basal sliding factor	0.1 to 10 (unitless)	2.2.3	5.6669	1.0718
Z_{f}	Scaling factor for climate-enhanced ice flow	0.55 to 2.2 (unitless)	2.2.4	1.9624	0.8849
θ	Time scale of bedrock elevation adjustment	1500 to 6000 yr	2.2.5	5126	3928

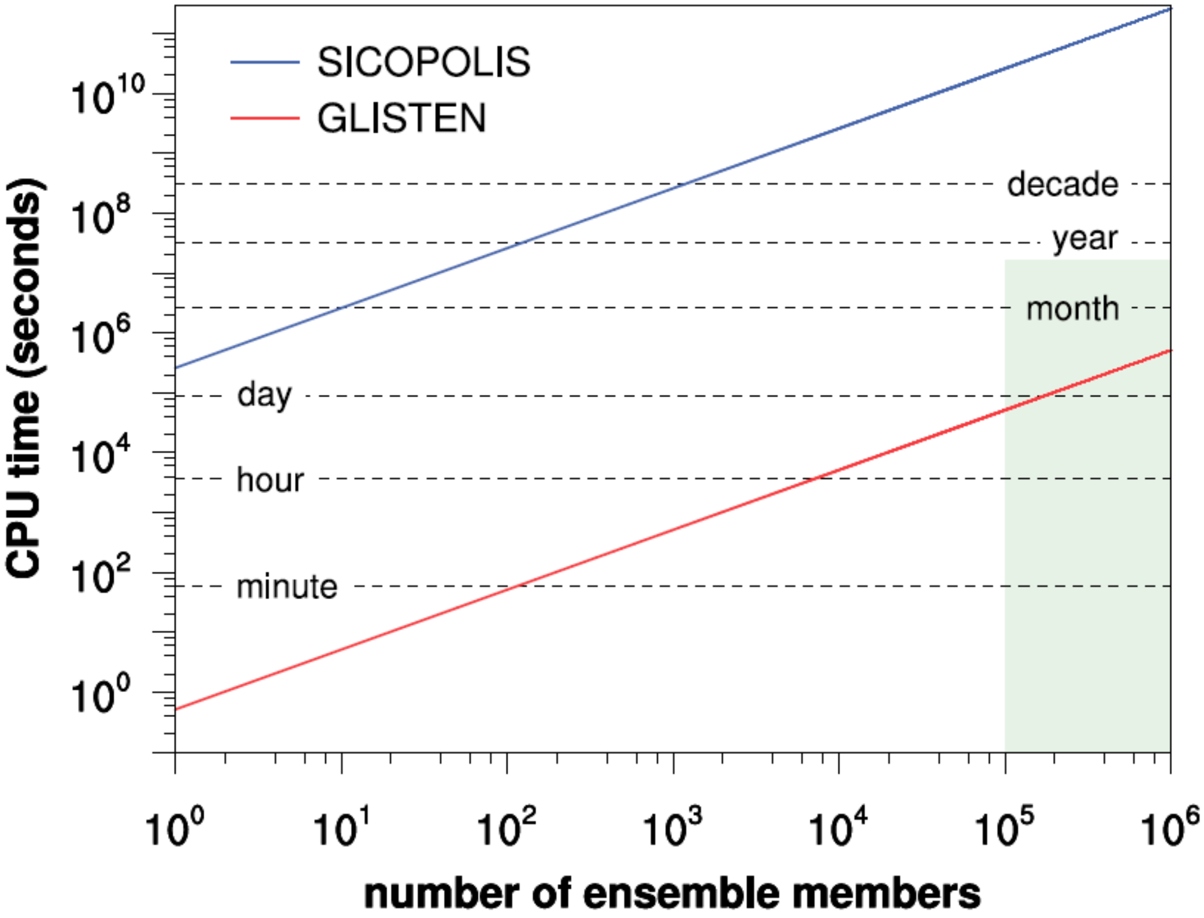

Figure 1. Computing time required for 125,000-yr ice sheet calculations using GLISTEN (red line) and the three-dimensional ice sheet model SICOPOLIS (blue line; Greve et al., 2011), as a function of the desired number of model runs, assuming that only one processor is available to perform the calculations. The green rectangle indicates an "acceptable zone" for integrated assessment studies, which require ~10^5 model evaluations (Moles et al., 2004) in six months or less. GLISTEN can complete this number of model evaluations in weeks to months, whereas SICOPOLIS would require decades.

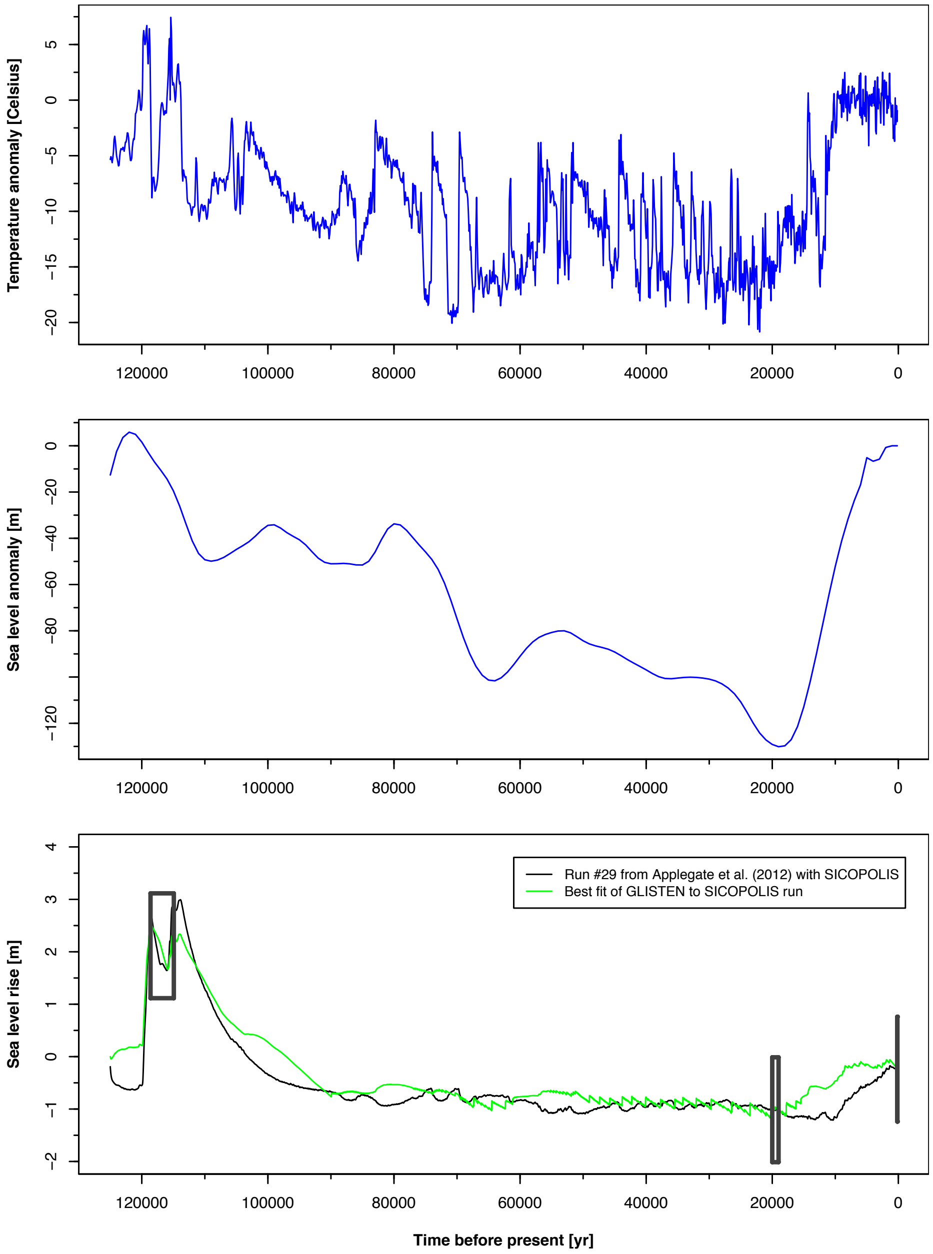

Figure 2. Forcings used to model ice sheet evolution over the last 125,000 yr, and demonstration that GLISTEN is able to match results from a three-dimensional ice sheet model (SICOPOLIS; Greve, 1997; Greve et al., 2011). Top: Greenland annual mean surface air temperature anomaly reconstructed from oxygen isotopes in the GRIP ice core (Dansgaard

et al., 1993; Johnsen et al., 1997). Middle: Sea level anomaly reconstructed from ocean sediment core oxygen isotopes (Imbrie et al., 1984). Bottom: Best fit of GLISTEN (green line) to run #29 from Applegate et al. (2012) using the SICOPOLIS ice sheet model (black line). The agreement between the model curves was tested for each calibration run during the periods indicated by the gray boxes. Both SICOPOLIS and GLISTEN were driven using the forcing curves in the top two panels.

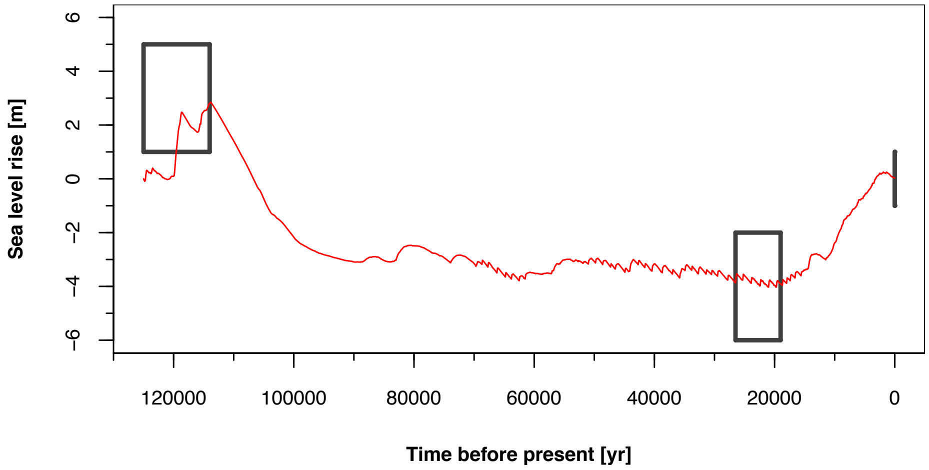

Figure 3. Sea level rise hindcast from the GLISTEN model over the last 125,000 years (red curve), after tuning to assessed past ice volume changes (gray boxes; Alley et al., 2010, their Fig. 13), the modern ice volume (Bamber et al., 2001), historical mass balance values (Fig. 4), and modern ice thicknesses (Fig. 5).

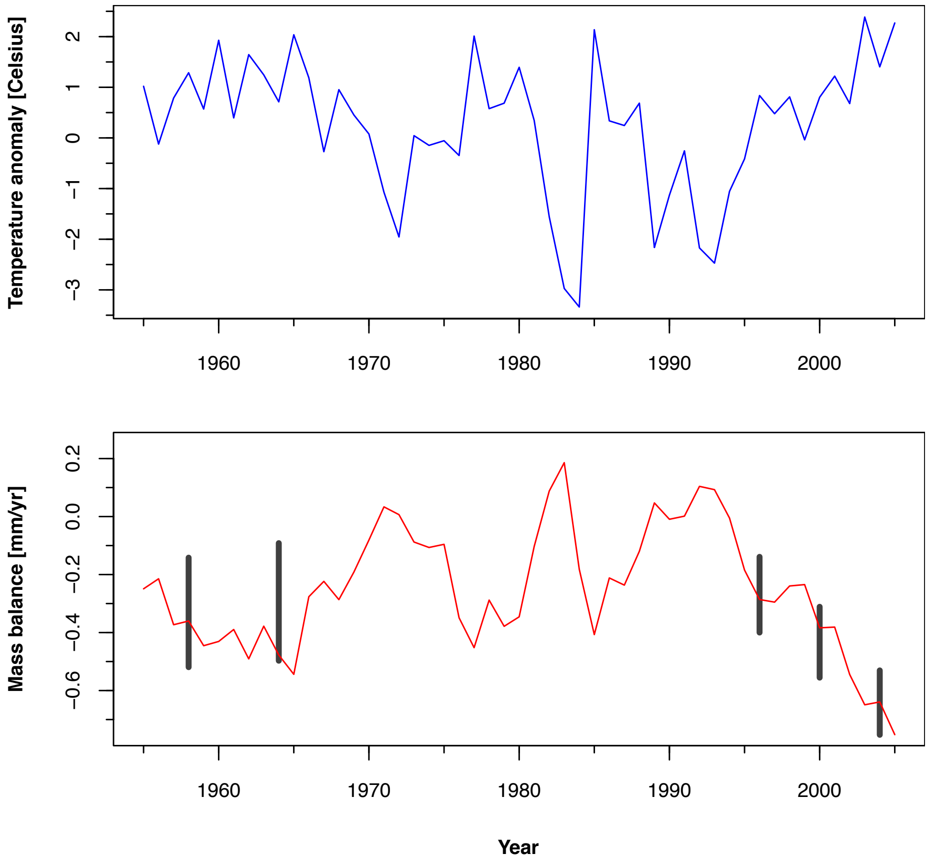

Figure 4. Surface air temperature anomalies used to force GLISTEN over the period 1955-2005 (top panel; Vinther et al., 2006), and GLISTEN's mass balance hindcast (red line) after tuning to historical mass balance values (gray bars; Rignot et al., 2008), assessed past ice volume changes and the modern ice volume (Fig. 3), and modern ice thicknesses (Fig. 5).

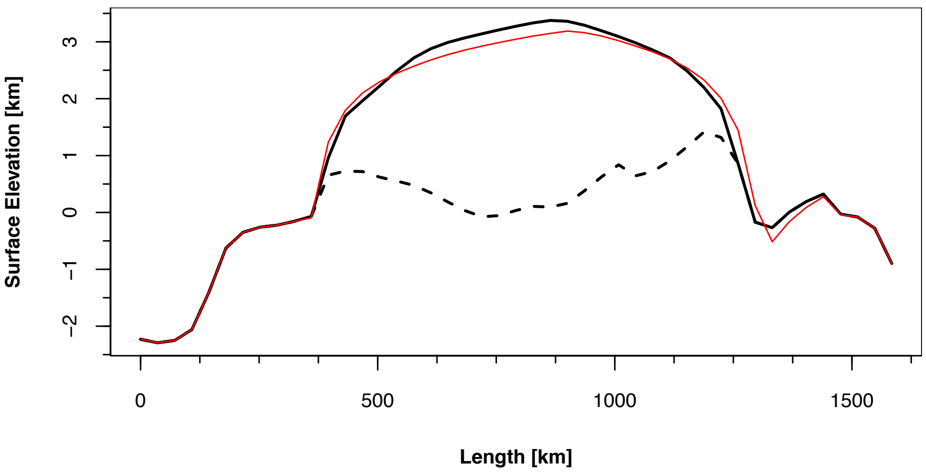

Figure 5. Modern Greenland Ice Sheet profile as estimated by GLISTEN (red line), after tuning to modern ice thicknesses, assessed past ice volume changes and the modern ice volume (Fig. 3) and historical mass balance values (Fig. 4). The observed modern ice surface and bedrock surface are shown for comparison (black lines; Letreguilly et al., 1991).