Committed sea-level rise for the next century from Greenland ice sheet dynamics during the past decade

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We use a three-dimensional, higher-order ice flow model and a realistic initial condition to simulate dynamic perturbations to the Greenland ice sheet during the last decade and to assess their contribution to sea level by 2100. Starting from our initial condition, we apply a time series of observationally constrained dynamic perturbations at the marine termini of Greenland’s three largest outlet glaciers, Jakobshavn Isbøe, Helheim Glacier, and Kangerdlugssuak Glacier. The initial and long-term diffusive thinning within each glacier catchment is then integrated spatially and temporally to calculate a minimum sea-level contribution of approximately \(1 \pm 0.4\) mm from these three glaciers by 2100. Based on scaling arguments, we extend our modeling to all of Greenland and estimate a minimum dynamic sea-level contribution of approximately \(6 \pm 2\) mm by 2100. This estimate of committed sea-level rise is a minimum because it ignores mass loss due to future changes in ice sheet dynamics or surface mass balance. Importantly, \(>75\%\) of this value is from the long-term, diffusive response of the ice sheet, suggesting that the majority of sea-level rise from Greenland dynamics during the past decade is yet to come. Assuming similar and recurring forcing in future decades and a self-similar ice dynamical response, we estimate an upper bound of 45 mm of sea-level rise from Greenland dynamics by 2100. These estimates are constrained by recent observations of dynamic mass loss in Greenland and by realistic model behavior that accounts for both the long-term cumulative mass loss and its decay following episodic boundary forcing.

Ice sheet modeling | Climate change

In its Fourth Assessment Report (AR4), the Intergovernmental Panel on Climate Change (IPCC) estimated a rise in global sea levels of 0.18–0.59 m by 2100, but the report explicitly avoided accounting for ice dynamical effects because of a perceived limit in understanding at that time (1). This failing was, in part, due to oversimplified numerical models of ice flow and the glaciological and climate modeling communities have responded with efforts toward developing “next generation” numerical ice sheet models with improved predictive skill (2, 3). However, the list of necessary model improvements is nontrivial (3) and more reliable sea-level projections from next generation models are still likely to be several years away.

Since AR4, there have been several efforts to make a reasonable assessment of the effects of ice dynamics on future sea-level rise (SLR). These efforts largely fall into two categories, those that extrapolate current trends in dynamic mass loss into the future (e.g., 4, 5) and those based on semiempirical models (e.g., 6, 7). For the latter, although there is no explicit accounting for ice dynamics, the historically constrained response of ice sheets to temperature forcing is assumed to include both surface mass balance (SMB) and ice dynamical effects. Both methods lead to estimates for total SLR (SMB plus dynamics) by 2100 that are significantly larger than the IPCC AR4 estimates, which account primarily for changes in SMB. However, both of these methods have their inherent limitations. Extrapolation of current trends in dynamic mass loss is problematic because periods of significant dynamic mass loss may be episodic in nature, making it unclear if current trends are representative of the future. Similarly, semiempirical models rely on historical relationships between globally averaged temperatures and ice sheet volumes and it is unclear if these relationships will hold in the future, particularly if geographically local phenomena are important components or drivers of dynamic mass loss.

Indeed, the latter appears to be the case for the majority of dynamic mass loss from the Greenland Ice Sheet (GIS) during the past decade.\textsuperscript{1} Whole ice sheet and basin scale studies indicate that the GIS was loosing mass at an accelerating rate during the late 1990s and early to mid-2000s (8, 9). Between 2000 and 2008, this imbalance was split approximately equally between losses as a result of surface melting and dynamic thinning (8, 10). The majority of the dynamic thinning is thought to be due to outlet glacier acceleration during the late 1990s and early to mid-2000s (11–20). Although the cause for this thinning has been debated (17, 18, 21), there is strong evidence that changes in the balance of forces at the glaciers’ marine termini are responsible (13, 14, 17–20, 22) and that these changes were indirectly triggered by relatively warm ocean waters (15, 20, 22). In most cases speed-up events have been short-lived (13, 16, 20) and their long-term impact on ice sheet mass balance and SLR remains unclear.

Here, we make a first effort toward assessing the impact of GIS dynamics on future SLR using an appropriate thermomechanical, three-dimensional, higher-order ice flow model and a realistic initial condition of the GIS. Because the higher-order momentum formulation accounts for horizontal stress gradients, we are able to simulate the flow of outlet glaciers in a realistic way, and following perturbations at the ice sheet margin, account for the transfer of those perturbations into the ice sheet interior. We apply the model to perturbation experiments that mimic the dynamic changes observed on Greenland’s three largest outlet glaciers, Jakobshavn Isbøe (JI), Helheim Glacier (HG), and Kangerdlugssuak Glacier (KG) (Fig. 1) during the last decade. Following the perturbations, we run the model forward in time to estimate the cumulative dynamic thinning in these basins from 2000–2100.

An ice sheet continues responding to a dynamic perturbation long after that perturbation ceases and our model is well-suited to capture this behavior. By applying only well-constrained dynamic perturbations from the past decade in Greenland, we obtain mini-
mum estimates for future mass loss due to dynamics, and thus minimum estimates for Greenland’s dynamic contribution to future SLR. Following the concept of committed climate change (23), we refer to this minimum estimate as the committed sea-level rise (i.e., the long-term mass loss from an ice sheet in response to past perturbations). We show how recent observations and the modeling conducted here can be combined to estimate the committed future SLR from all dynamic thinning in Greenland during the past decade. Finally, we use our model results and some reasonable assumptions about future dynamic forcing and response in Greenland to make an upper-bound estimate for Greenland’s dynamic contribution to SLR during the next century.

Results

Flow Modeling. Our three-dimensional ice sheet model solves the first-order momentum balance equations and the advective-diffusive temperature equation for ice sheets (24, 25). We use our model, 5-km resolution ice sheet geometry (26) and balance-velocity fields (27), and a tuning procedure to derive a realistic, steady-state initial condition for the GIS. At the fronts of JI, HG, and KG, we apply a stress boundary condition appropriate for ice in contact with seawater adjusted so that steady-state model fluxes agree with flux estimates from the mid-1990s (8), prior to the initiation of large perturbations at the termini of these glaciers (Table 1). The balance velocities and steady-state, depth-averaged model velocities are shown in Fig. 1. Root-mean-square differences for the entire GIS are 38 m yr⁻¹ and 3–8 m yr⁻¹ for the JI, HG, and KG basins (Fig. S1). The flow model, boundary conditions, tuning procedure, and applied perturbations are discussed in further detail in Methods, SI Text, and Fig. S2.

We perturb the initial, steady-state ice sheet by altering the stress boundary condition over time at the fronts of JI, HG, and KG, so that modeled and observed flux changes are in agreement. Physically, the perturbations can be interpreted as a reduction in resistive stress near the glacier termini resulting from, e.g., the loss of a floating ice tongue (12, 15, 19, 22, 28), or from an increase in the longitudinal strain-rate near the glacier front following a retreat into deeper water (13, 14). The resulting instantaneous acceleration of ice at the glacier front propagates inland as a diffusive wave of thinning. When integrated in time and space, this thinning represents the committed future SLR from GIS outlet glacier dynamics during the past decade. For HG and KG, we continue to alter the applied perturbation for 2–3 y after the initial perturbation to match observed discharge changes (13). The modeled and observed discharge anomalies for HG and KG are shown in Fig. 2. For JI, the applied perturbation is simpler for two reasons. First, a published record of its discharge since it began accelerating in the mid to late 1990s (12) does not exist. Second, and more importantly, JI has undergone near continuous acceleration and retreat (29) since that time for reasons that are poorly understood. Because the resolution, physics, and simplified perturbations applied in the current model cannot recreate this complex behavior we apply a one-time, step-function to the stress boundary condition for JI, which approximates the initial doubling in discharge (see Table 1 and Fig. 2) observed on JI from the mid to late 1990s (8, 12, 29). Although the model cannot fully mimic the observed behavior of JI with this

Table 1. Initial, modeled ice flux for the JI, HG, and KG drainage basins compared with observations (8)

<table>
<thead>
<tr>
<th>Drainage Basin</th>
<th>Model flux, km³ yr⁻¹</th>
<th>Observed flux, km³ yr⁻¹</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jakobshavn Isbrae (JI)</td>
<td>23.5 [-1]</td>
<td>23.6</td>
</tr>
<tr>
<td>Helheim Glacier (HG)</td>
<td>25.8 [-2]</td>
<td>26.3</td>
</tr>
<tr>
<td>Kangerdlugssuak Glacier (KG)</td>
<td>27.8 [0]</td>
<td>27.8</td>
</tr>
</tbody>
</table>

For the modeled fluxes, the number in square brackets is the percent difference when compared to the observed flux.

‡Plotted discharge anomalies for the model are relative to the steady-state initial condition. For the HG and KG observations, they are relative to the year 2000 (13), and for JI the initial anomaly is relative to the mid to late 1990s (8, 12), at which time we assume that JI was near equilibrium.
simplified perturbation, the simulations of JI do provide a useful comparison between the model behavior and the observations, as discussed further below.

Comparison with Observations. To demonstrate that the perturbed model response is reasonable we compare modeled and observed profiles of the thinning rate in the JI, HG, and KG drainage basins. Our goal is not for the model to accurately match the observed profiles at any particular point in time, as such a comparison is beyond the model resolution and the simplifying assumptions made here. Rather, we aim to demonstrate that the perturbed model response—the magnitude and spatial pattern of thinning—is reasonable when compared to the observed response. Fig. 3 shows observed thinning rates calculated from differencing satellite laser altimetry- and satellite imagery-derived digital elevation models (30). From observations (13, 15, 20), we assume the trigger for acceleration and thinning of HG and KG occurred between 2003 and 2005, and thus we compare our modeled thinning rates to observations from 2004 to 2005. For both HG and KG, the modeled magnitude and spatial pattern of thinning are similar to the observations except that the observations show larger thinning rates farther inland. We attribute this difference primarily to the fact that deep bedrock troughs, which focus and channel the thinning inland (13, 30), are absent from the model bedrock topography. Thus, the pattern of modeled thinning is more radial than observed. Importantly, Fig. 3 shows that the model does not overestimate the amount of thinning relative to observations. As such, estimates of dynamic thinning will be biased toward zero relative to the observations.

For reasons discussed above, a comparison between the model and observations for JI is more difficult. In particular, we expect to significantly underestimate the observed thinning because of our simplified, one-time perturbation. Further, the initial perturbation applied to JI is based on discrete observations from 1996 to 2000 (8), but the earliest profile of inland thinning for JI is from 2003 to 2005, 6–8 y after the model perturbation is applied. Over this time period, the modeled perturbation has decayed significantly relative to reality and the match between modeled and observed thinning is poor; the maximum observed thinning on JI from 2003 to 2005 is approximately an order-of-magnitude larger than that predicted by the model. However, if we scale the modeled thinning rates by the observed, maximum thinning rate, we find that the observed and (scaled) modeled thinning profiles are remarkably similar (Fig. 3). We interpret this similarity as indicating that our model captures the essential long-term, spatial response of thinning on JI, while underpredicting its magnitude because of the simplified perturbation.

Committed Sea-Level Rise from JI, HG, and KG. Following the initial perturbations to JI, HG, and KG, we step the model forward in time for 100 y. At each timestep we spatially integrate modeled thinning rates within each basin, which we assume contribute directly to the rate of SLR. The discharge anomaly (the cause for the thinning) and the corresponding rate of SLR for the first 10 y of the simulation are shown in Fig. 2. From the modeled rate of SLR curves we calculate cumulative SLR curves (Fig. 4A). By 2100 the combined, cumulative SLR from JI, HG, and KG is 1.1 ± 0.35 mm (the uncertainty estimate is discussed further below). We stress that this estimate is a conservative, minimum estimate for two reasons. First, we have not fully accounted for the continued acceleration and thinning of JI, in which case the flux anomalies for JI in Fig. 2 are too small. Second, we have only considered dynamic thinning resulting from perturbations on these glaciers during the past decade. No accounting is made for perturbations on other outlet glaciers or for future perturbations that might result in additional thinning. Thus, these simulations provide estimates for the sea-level contributions from JI, HG, and KG over the next century in response to forcing during the past decade.

Committed Sea-Level Rise from all of Greenland. From the rate of SLR and cumulative SLR curves in Fig. 2B and Fig. 4A we make two observations. First, if we attribute all of the SLR for the three years3 following the perturbation to the perturbation itself

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3From observations (13) and modeling (Fig. 2), we estimate 3 y as an approximate e-folding timescale for decay of the initial perturbations in outlet glacier flux.
A function of time is then given by

$$SLR(t) = \int rF(t)dt = r\bar{F}(t),$$

where $\bar{F}(t)$ gives the normalized, cumulative SLR (in units of years); multiplying $\bar{F}(t)$ by an initial rate of SLR $r$ gives the cumulative SLR as a function of time. To extend Eq. 1 to multiple outlet glaciers, recall that the long-term, cumulative SLR is insensitive to small offsets in the timing of the initial outlet glacier perturbations (e.g., Fig. 4A) and assume that dynamic thinning was triggered simultaneously on $n$ of Greenland’s outlet glaciers. At some later time, the total, cumulative SLR from those $n$ perturbations is estimated by

$$SLR(t)_{\text{total}} = r_1F_1(t) + r_2F_2(t) + r_3F_3(t) + \ldots + r_nF_n(t).$$

If $\bar{F}(t)$ is a representative mean for the individual $F(t)$ in Eq. 2, such that $F_1(t) \approx \ldots \approx F_n(t) \approx \bar{F}(t)$, then

$$SLR(t)_{\text{total}} \approx (r_1 + r_2 + r_3 + \ldots + r_n)\bar{F}(t) = \left(\sum_{n}^{} r_n\right)\bar{F}(t).$$

Assumptions (i) and (ii) from this conceptual model are inherent in the summation term on the right-hand side of Eq. 3 and in the simplification $F_1(t) \approx \ldots \approx F_n(t) \approx \bar{F}(t)$. Assumption (iii) from our conceptual model is that $\bar{F}(t)$ can be estimated from the modeling discussed above. We take $\bar{F}(t)$ to be the mean of the normalized, cumulative SLR curves modeled here noting that, over a 100 y time period, $F(t)$ for the individual outlet glaciers vary from the mean $\bar{F}(t)$ by no more than $\pm 35\%$. We take this variation with respect to the mean as an estimate for the uncertainty introduced by assuming that $F_1(t) \approx \ldots \approx F_n(t) \approx \bar{F}(t)$ (see additional discussion in SI Text and Fig. S3).

Although we do not have a direct way to estimate the first term on the right-hand side of Eq. 3 we can estimate its mean value over the time period of interest here from observations. Recent estimates (10) give a mean SLR of 0.46 mm y$^{-1}$ from the GIS from 2000 to 2008, split equally between surface mass balance and ice dynamics (8, 10). We constrain the first term on the right-hand side of Eq. 3 by requiring that the mean rate of SLR predicted from the model for that same time period is equal to 0.23 mm y$^{-1}$. Conceptually, our modeled, cumulative mass discharge for Greenland over the last decade is scaled so that it agrees (on average) with that from (10) (see figure 2A in ref. 10). In future decades this scaled discharge decays according to our model. The resulting cumulative SLR estimate is shown in Fig. 4B. By 2100, the lower-bound estimate for SLR from GIS dynamics during the past decade is 5.8 ± 2.1 mm.

### Discussion and Conclusions

From perturbation experiments with a higher-order ice flow model we estimate a minimum, committed dynamic sea-level contribution from JI, HG, and KG of 1.1 ± 0.35 mm by 2100. We stress the minimum nature of this estimate, which accounts only for ice sheet thinning on these outlets as a result of known and well quantified past perturbations and their resulting long-term, diffusive response.

Based on support from numerous observational and modeling studies, we scale these results to the entire GIS for a minimum SLR of 5.8 ± 2.1 mm (from dynamics) by 2100. When added to an estimate for the cumulative SLR (~40 mm) from changes in SMB following a midrange future emissions scenario, the total cumulative SLR increases to approximately 46 mm, with 13% resulting from dynamic perturbations during the past decade.

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8We take this approach rather than attempting to use our model because of the lack of necessary data (e.g., ice thickness and outlet glacier flux time series) available for other outlet glaciers in Greenland.

9See ref. 31 and Methods.
Of the total estimated dynamic SLR by 2100, ≥75% is due to the long-term, diffusive response of the ice sheet. Thus, the majority of the SLR from perturbations to Greenland’s outlet glaciers during the past decade is yet to come. Put another way, the hundred-year mean rate of SLR from dynamic perturbations amounts to approximately 5% of the initial rate; an initial perturbation giving 1 mm y⁻¹ of SLR will contribute a hundred-year mean rate of 0.05 mm y⁻¹. Given some initial rate of SLR associated with future dynamic perturbations, this order-of-magnitude estimate provides a means by which the additional, diffusive contribution to SLR may be estimated.

When compared to other recent estimates of SLR from GIS dynamics by 2100 we find that our estimate is approximately 2–12% of that from semi-empirical models (6, 7) and approximately 5–25% of that based on the extrapolation of currently observed trends in dynamics (4, 5) (see Methods and SI Text for more discussion). This comparison is not straightforward to interpret; our method explicitly accounts for long-term SLR from dynamical changes that have already taken place but makes no attempt to account for future dynamical changes, whereas the methods we are comparing to do exactly the opposite. Nevertheless, our estimate is not an insignificant fraction of these other estimates and, assuming the methodologies behind these other estimates are sound, we conclude that they may underestimate future SLR from GIS dynamics by as much as 25%. This additional SLR is the result of perturbations that have already taken place and are thus already locked into the long-term evolution of the present-day ice sheet.

The modeling conducted here and some reasonable assumptions can be used to make approximate upper-bound estimates for future SLR from GIS dynamics, without accounting for future dynamical changes explicitly. As discussed above, numerous observations indicate that the trigger for the majority of dynamic thinning in Greenland during the last decade was episodic in nature, as the result of incursions of relatively warm ocean waters. By assuming that similar perturbations occur at regular intervals over the next century and that the ice sheet responds in a similar manner, we can repeatedly combine (sum) the cumulative SLR curve from Fig. 4B to arrive at additional estimates for SLR by 2100. For example, if perturbations like those during the last decade recur every 50, 20, or 10 y during the next 100 y, we estimate a cumulative SLR from GIS dynamics by 2100 of approximately 10, 25, and 45 mm, respectively (see Methods and Figs. S4–S6). This range of periodic forcing is reasonable with respect to natural modes of climate variability (e.g., the Atlantic multidecadal oscillation and the North Atlantic oscillation) that have been implicated in causing the warm ocean water incursions responsible for outlet glacier acceleration during the past decade (e.g., 15, 22). The upper-bound estimate of approximately 45 mm allows for the maximum influence of GIS dynamics on SLR while also accounting for the episodic and decaying nature of the causal perturbations (Methods). Although this upper-bound estimate relies on extrapolation of current observations of dynamic imbalance (as in refs. 4 and 5) it does so in a way that takes advantage of realistic, prognostic model behavior. Thus, if we assume that the previous decade of dynamic mass loss in Greenland is representative of the future, then we should expect no more than 45 mm of SLR from dynamics by 2100. For comparison, this estimate is approximately one half of the upper-bound estimate for the contribution from GIS dynamics from ref. 5. Addition of the estimated 40 mm of SLR from changes in SMB by 2100 (31) would result in a total SLR from Greenland of 85 mm by 2100.

Lastly, we emphasize that our modeling implicitly assumes no change in the magnitude or spatial distribution of basal sliding parameters in the future. For example, if basal sliding becomes more widespread on the GIS in the future, then both the lower- and upper-bound SLR estimates presented here are biased low.
C1 over time. At any timestep, the applied value of C1 is that required to enforce agreement between modeled and observed outlet glacier flux anomalies (e.g., Fig. 2A). See SI Text for additional discussion.

**SMB Contribution to SLR by 2100.** The estimate for Greenland's contribution to SLR by 2100 as a result of changes in SMB is taken from ref. 31, in which an SMB model over Greenland was calibrated to precipitation and temperature anomaly data for the period 1970–1999. The calibrated SMB model was then applied to temperature and precipitation anomaly data from 23 atmospheric and oceanic general circulation models participating in the IPCC AR4 (1) under the midrange, A1B future emissions scenario. The median SMB prediction was used for the value of the cumulative SLR by 2100 quoted here (for the 25 and 75% quartiles, the predicted cumulative SLR by 2100 is approximately 32 and 45 mm).

**Comparison to Other Estimates of SLR by 2100.** We compare our estimate for the committed SLR from ice dynamics by 2100 to other existing estimates in the literature. In most cases, a direct comparison is difficult because (i) existing estimates make assumptions about the future influence of ice sheet dynamics, and (ii) existing estimates are presented in terms of global SLR, rather than just that from GIS dynamics. In the latter case, we make reasonable estimates for how to partition the total SLR based on values in the published literature. This partitioning suggests that the total SLR contribution from the GIS can be split equally between that due to surface mass balance and that due to ice dynamics and that between 10 and 15% of the global SLR can be assigned to GIS dynamics. For estimates of dynamic SLR from the GIS by the year 2100, we find that our estimate is approximately 2–12% of that estimated using empirical models (6, 7) and approximately 5–25% of that estimated by extrapolation of current observations and trends (4, 5). Our partitioning of published SLR estimates and the comparison between those estimates and ours are discussed in more detail in SI Text.

**Upper-Bound Estimates for SLR by 2100.** To make a reasonable upper-bound estimate for future SLR from GIS dynamics by 2100, we make the following simplifying assumptions. First, we assume that the dominant mechanism for dynamic mass loss in the future will be the same as that explored here (i.e., acceleration of marine terminating outlet glaciers following stress-balance perturbations at their termini). Second, we assume that similar perturbations during the past decade may occur at regular intervals during the next century. Third, we assume that the ice sheet responds to each of those future perturbations in a similar manner to that shown in Fig. 48. For some given recurrence interval, the total SLR by 2100 is given by the cumulative sum of curves like that shown in Fig. 48 that have had their origins shifted forward along the time axis (and for which the portion of the shifted curve extending beyond 2100 has been removed). For example, for a single additional perturbation occurring at 2050, the total SLR by 2100 is given by summing the curve shown in Fig. 48 with the same curve when its origin has been shifted to a start date of 2050. The individual (colored) and cumulative (black) SLR curves for this example are shown in Fig. S4. Additional examples for recurrence intervals of 20 and 10 y are shown in Figs. S5 and S6. Here, we take the estimate of approximately 45 mm SLR using a 10 y recurrence interval as an upper bound for dynamic SLR from the GIS by 2100. A more frequent recurrence interval cannot be ruled out but is not supported by oceanographic and glaciological observations over the past decade (the only time period for which we have reasonably good observations). Over this time period, observations (15, 17, 18, 20, 22) suggest essentially one period of oceanographic forcing and dynamic ice sheet response in Greenland.

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Supporting Information

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SI Text

Tuning Procedure for Initial Condition. We use our three-dimen-
sional, thermomechanical, prognostic ice sheet model and mod-
ern-day observations of ice sheet geometry, surface temperature,
and balance velocity to derive a realistic, steady-state initial con-
dition for the Greenland ice sheet. The initial ice sheet geometry
(1) is discretized on a 5-km resolution horizontal grid with 11
vertical sigma-coordinate levels. Balance velocities are derived
using the same geometry data (see Methods). Boundary condi-
tions for the heat balance model are as described in Methods.

We use the model as follows to derive an initial, thermomecha-
nical steady state that is consistent with the specified geometry,
balance velocity, surface temperature, and geothermal flux:

i. With no-slip and zero-flux boundary conditions applied at
the ice sheet bed and margins, respectively, we step the model
forward in time to allow the ice temperature, velocity, and
effective viscosity to equilibrate (defined as the point at which
there are no further significant changes in ice temperature and
velocity, which occurs within approximately 40 thousand years.
Note that in step (iii) of the tuning process we allow for an
additional 40 thousand years of thermomechanical equilibrja-
(ice sheet geometry, surface temperature, and geothermal flux are held steady.

ii. Using the model velocity field from step (i), we calculate two-
dimensional (x, y plane) fields for the depth-averaged speed
and the basal traction magnitude. From the former, we calcu-
late the sliding speed required to bring the modeled depth-
averaged speed into agreement with the balance velocity. The
sliding speed and basal traction magnitude are then applied to
a frictional sliding law \( |\vec{u}_b| = \beta |\vec{u}_v| \), where \(|\vec{u}_v|\) is the basal
traction magnitude, \(|\vec{u}_b|\) the desired sliding speed, and \( \beta (x,y) \)
the frictional sliding coefficient, to derive a map of the sliding
coefficient: \( \beta (x,y) = |\vec{u}_b|/|\vec{u}_v| \).

iii. The map of \( \beta \) derived from step (ii) is applied so that sliding is
allowed wherever it is required to improve the match between
the model and the balance velocities. Where sliding is deter-
mined to be approximately 0, \( \beta \) is set to an arbitrarily large value
(e.g., 10^{10} Pa m^{-1}), effectively enforcing a no-slip con-
tion at the bed. A floating ice boundary condition is applied
as the lateral boundary condition for specific outlet glaciers
(as discussed in Methods). The model is again stepped forward
in time and allowed to come into thermomechanical equili
rium while holding the updated basal and lateral boundary
conditions and geometry steady [as in step (i)]. Steps (ii)
and (iii) are iterated on until the mismatch between the model
and target (balance) velocity fields becomes small.

iv. For additional fine-tuning of the \( \beta \) field, the ratio \( R = \frac{\bar{U}_\text{model}}{\bar{U}_\text{obs}} \)
calculated at each (x,y) point, where \( \bar{U}_\text{model} \) is the final depth-
averaged model velocity from steps (i)–(iii) above and \( \bar{U}_\text{obs} \)
is the target, balance velocity at the same point. Where \( |R - 1| > \gamma \)
for some 0 \( \leq \gamma \ll 1 \) we adjust \( \beta \) by \( \bar{\rho}_\text{sw} = \bar{\rho}_\text{old} + R \). The float-
ing ice boundary condition may also be further tuned at this
point, so that depth-averaged model velocities at the fronts of
J1, HG, and KG match their desired, initial values (as dis-
cussed in Methods). After this fine-tuning step, the tempera-
ture (and flow-law rate factor) distribution within the ice is
held fixed. This constraint ensures that minor transient
changes in ice sheet temperature do not contribute to changes
in the ice sheet volume when allowing for geometry evolution
during time dependent, forward model simulations, which
allows us to attribute all ice volume changes following outlet
 glacier perturbations directly to the dynamic response of the
model.

The tuning procedure leads to a quantifiable match between
the model depth-averaged velocity field and the target velocity
field; for the entire ice sheet the rms difference between the mod-
elled depth-averaged velocities and the balance velocities is
\( \sim 38 \text{ m yr}^{-1} \) and for the Jakobshavn (J1), Helheim (HG), and Kan-
gerdlugssuaq (KG) drainage basins, the rms differences are 8, 3,
and 7 m yr^{-1}, respectively (Fig. S1). Note that we anticipate anom-
ously large rms differences in some regions of the ice sheet
and these are removed when calculating the statistics cited above.
These regions include the majority of the area within two grid
cells from the margin (excluding the outlet glaciers where we ap-
ply a boundary condition appropriate for floating ice), for which
speeds are biased toward zero in the model because of a zero flux
boundary condition applied at the lateral margin. In addition,
some portions of the downstream region of the Northeast Green-
land Ice Stream flow anomalously slow in the model. We remove
these regions from the comparison because they have no rele-
vance to the outlet glacier basins that are of primary interest in
the modeling conducted here (with all velocities included in the
comparison, the rms difference between the modeled, depth-
averaged velocity field and the target, balance velocity field is
\( \sim 119 \text{ m yr}^{-1} \)).

We use the velocity, temperature, and effective viscosity fields
at the end of this tuning procedure as initial conditions for the
perturbation experiments discussed below. The derived map of
the basal sliding coefficient is held constant and steady for the
duration of the simulations; we implicitly assume that the mag-
nitude and/or spatial pattern of sliding parameters does not
change in the future (see additional discussion below in the para-
graph Outlet Glacier Boundary Conditions).

We adopt the tuning procedure described above in place of
a more traditional model spin up for several reasons. First,
the purpose of a long-term spin up (e.g., from the last to the present
interglacial) is primarily to incorporate a more realistic tempera-
ture history into the ice sheet, which would have some affect on
the ice sheet viscosities (relative to an ice sheet for which tem-
peratures were assumed to be in steady state with present-day
forcing). The ice sheet geometry and viscosity are both likely
to be of first-order importance with respect to ice sheet dynamics
but the former is much better constrained with respect to current
observations. Put another way, constraining the model to fit
the known present-day geometry makes more sense than forcing the
model to follow a climate history that is poorly constrained in
both space and time and ending with a present-day geometry that
does not agree with observations. Second, any spin-up procedure
involving assumed past climate and/or ice dynamics histories
will result in model transients that will exert an influence on the
systems evolution over timescales longer than those of inter-
store here (i.e., the next 100 y). By studying a system that is initially
in a steady state, we are better able to isolate and focus on the
dynamic influence of the marine margin perturbations of interest
here, rather than trying to isolate their effects relative to longer
timescale transients introduced by a spin-up procedure.

Target Velocity Field: Balance Velocities Versus Observed Velocities.
We choose balance velocities as the target velocity field in our
tuning procedure for a number of reasons. First, the map of
balance velocities used in our study offers complete coverage of
the ice sheet, whereas other maps based on, e.g., observations cover only a fraction of the ice sheet (and importantly, offer incomplete coverage over HG and KG) and suffer from large uncertainties in some regions. Second, deficiencies in the current knowledge of ice thickness mean that even a perfect flow model could not recreate observed velocity fields (i.e., the geometry forcing the model is not the same as that forcing the observed flow). Balance velocities are, by definition, in equilibrium with the geometry used to derive them, regardless of the deficiencies in that data. Here, the model and the balance velocities use the same datasets of ice sheet geometry, and are thus internally consistent with respect to geometric forcing. Third, observed velocities represent a snapshot in time of the velocity field. They may (and in the case of Greenland over the last decade most certainly do) contain short-term, dynamic variability not representative of steady-state flow. Clearly, sliding in the model should not be tuned to match a short-term, transient acceleration or deceleration, which would bias the modeled sliding speeds too, unreasonably fast or slow. The conservative choice in the present case, where we know that measured velocities record outlet glacier acceleration over the past decade, is to bias modeled sliding toward a relatively slower (i.e., unaccelerated) initial condition. A similar argument applies to whole ice sheet maps of velocity that are actually composites of observed velocities from a number of different time periods. In this case, not only is the velocity field biased by the fact that it is a snapshot of a known transient, but in different regions of the ice sheet the snapshots may represent different points in time.

Surface Mass Balance. The steady-state velocity field that coincides with the tuned, initial condition implies a particular surface-mass balance (SMB) field that must be applied to maintain a steady state when evolving the model geometry forward in time. We calculate this SMB field according to the steady-state continuity equation, \( \nabla \cdot (U H) = \text{SMB} \), where \( U = [u, v] \) is the depth-averaged, vector-valued velocity in the \((x, y)\) plane, \( H = H(x, y) \) is the ice thickness, and \( \text{SMB} = \text{SMB}(x, y) \) (in \( \text{my}^{-1} \) ice equivalent) is the SMB field required to balance the horizontal flux divergence. When stepping the model forward in time with no additional perturbations, this derived SMB field maintains a steady state with respect to the rate of ice thickness change. This SMB field is held steady during the perturbation experiments discussed above. For each basin of interest and at each timestep, the dynamic sea-level contribution is then given by the difference between this SMB field, integrated over the catchment area, and the rate of thickness change integrated over the catchment area.

When compared to a real SMB field for Greenland (e.g., that used in ref. 2, which is derived from a regional climate model constrained by sparse in situ observations) our SMB field exhibits many of the same regional scale features, e.g., positive values in the ice sheet interior becoming more negative toward the coasts and relatively large ablation (accumulation) rates in southwest (southeast) Greenland. On the local scale, our SMB field can exhibit relatively large excursions from the regional SMB (the standard deviation in our implied SMB field is approximately three times larger than that used to derive the balance velocities). These differences are largely an artifact of (i) the model versus target velocity field mismatch at the end of the tuning procedure, and (ii) the assumptions made in balance velocity calculations. If the modeled and target (balance) velocity fields matched each other perfectly, then the SMB field used to generate the balance velocities would be recovered perfectly when calculating the flux divergence from the model velocities. However, small mismatches (see Fig. S1) lead to noise in the SMB field calculated from the flux divergence of the model velocity field. A perfect match between the modeled velocities and the balance velocities used here is not possible, as the latter are based on the assumption that ice always flows downhill (i.e., the balance velocity calculation assumes zero-order accurate shallow ice approximation dynamics), whereas the flow model used here is of first-order accuracy, and as such, includes horizontal stress gradients that may violate that simple assumption. Because of this difference, some amount of noise in the SMB field necessary to maintain a model steady state is inevitable, at least for the tuning procedure described above. An approach to improve on this method would employ the higher-order momentum equations as constraints during the balance velocity calculation, or include the flux divergence expression above as a constraint in the higher-order momentum balance solved by the prognostic ice flow model.

Outlet-Glacier Lateral Boundary Conditions and Applied Perturbations. As discussed in Methods, the stress boundary condition applied at the marine termini of outlet glaciers in our model is adjusted for both initial tuning of the model and for introducing perturbations into the model. Changes to the boundary condition are made through the tuning parameter \( C = C_0 C_1 (t) \) (see also the boundary condition discussion in Methods). First, we alter the value of \( C_0 \) to obtain the steady-state initial fluxes given in Table 1, while holding \( C_1 \) constant at a value of 1. For JI, HG, and KG, these initial values of \( C_0 \) are 14.5, 3.7, and 7.6, respectively. For time-dependent experiments, we apply a perturbation to the model by further adjusting the stress boundary condition; the values of \( C_1 \) are altered while holding the values of \( C_0 \) fixed at their initial values. Changes in \( C_1 \) occur as step functions in time; once a perturbation is applied (once \( C_1 \) is changed), the boundary condition is left at its perturbed value indefinitely (as in the case of JI) or changed at later times as needed to match the observed time series of flux anomalies (in the case of HG and KG). A time series for the values of \( C_1 \) is shown in Fig. S2. The physical interpretation of these boundary condition changes is, (i) the permanent loss of a floating ice tongue that previously provided resistance to flow (through longitudinal-stress gradients) and is now absent (e.g., 3), or (ii) retreat of an outlet glacier front away from a previously stable geometric position (e.g., 4).

Because of the coarse model resolution, we cannot accurately account for motion of the grounding line (5, 6), retreat of the outlet glacier calving fronts, or the resulting effects on ice dynamics. This omission is equivalent to assuming that the length of grounding line and/or calving front retreat is less than the length of a single 5-km grid cell. For JI, HG, and KG observed calving front retreat during the time period simulated here were all \( \leq 1 \) grid cell (3, 4). This lack of retreat in the model is likely to be responsible for at least a fraction of the underestimation of thinning in the ice sheet interior (e.g., as observed for HG and KG in Fig. 3); previous work (7) demonstrates that a downstream perturbation moving inland over time results in more inland thinning occurring sooner, relative to a stationary downstream perturbation.

For JI, the observed, continued acceleration and thinning (3) that our modeling does not capture may also be due in part to processes not included in the modeling conducted here (e.g., weakening of the lateral margins or the ice-bed contact over time) but for which there are few or no constraining observations. Lastly, we reiterate that a number of previous observations strongly support the contention that outlet glacier acceleration in Greenland during the past decade was not related to changes in surface meltwater induced lubrication of the ice-bed interface, and thus not directly related to changes in basal conditions (i.e., sliding) as a function of subglacial hydrology. If sliding were to become easier or more widespread over significant portions of the Greenland Ice Sheet (GIS) in the future, then it follows that both the lower- and upper-bound sea-level rise (SLR) estimates discussed here will be biased low.
Estimate for the Uncertainty in the Modeled, Cumulative SLR. We estimate an uncertainty for the rates of SLR given by the model by comparing the normalized cumulative SLR curves for the individual glaciers modeled here to their normalized total value (that is, the individual normalized curves versus their mean value). This estimate assumes that the behavior of the sample of outlet glaciers modeled here is representative of the behavior of the larger sample of outlet glaciers that contributed to Greenland’s dynamic imbalance over the last decade. Over a 100 y time period, the normalized, cumulative SLR curves for the individual glaciers vary from the mean by at most \( \pm 35\% \) (Fig. S3). We take this value as an estimate for the uncertainty in the predicted, cumulative SLR.

Comparison to Other Published Estimates for Cumulative SLR by 2100. We partition other recent estimates of the cumulative SLR by 2100 into their component parts to calculate the SLR contribution from those estimates that is due to GIS dynamics alone (e.g., for cases where the estimates are for the global, total SLR). Based on values published in the recent literature, we make the following assumptions:

1. The ice sheet, versus glaciers and ice caps, versus thermal expansion, versus terrestrial waters contributions to the total SLR are partitioned according to recent SLR budget calculations from satellite altimetry and gravimetry and float measurements (8). This partitioning gives contributions of approximately 1, 1, 0.3, and \( \sim 0.2 \) mm y\(^{-1}\), respectively, for the ice sheet, glaciers and ice caps, thermal expansion, and terrestrial waters components of the total SLR. The total ice sheet fraction of SLR is then \( 1/(1 + 1 + 0.3 + 0.2) = 0.4 \).

2. Based on numbers given in ref. 9, we assign 50–75% of the SLR contribution from ice sheets to Greenland (the remaining is assumed to come from Antarctica).

3. Based on ref. 2, we split the SLR contribution from Greenland equally between surface mass balance (melting and sublimation) and ice dynamics.

Combining these numbers into a single scaling factor, \( \phi \), we have \( 0.10 \leq \phi \leq 0.15 \). Multiplying any global SLR estimate by \( \phi \) then gives an estimate for the fraction of that estimate that is due to GIS dynamics alone. We apply \( \phi \) to several estimates for the global, total SLR by 2100, as discussed below.

From ref. 10, the low range estimate for the upper bound on dynamic, cumulative SLR from the GIS by 2100 is 93 mm (obtained by doubling flow rates of all current outlet glaciers from 2000–2010 and then holding those flow rates constant until 2100). Our committed SLR estimate of 6 mm is 6% of the estimate based on ref. 10.

From ref. 9, the total GIS contribution to SLR by 2100, assuming (i) no acceleration of current mass imbalance, and (ii) acceleration at the current rate, is estimated at approximately 47 and 245 mm, respectively. We partition these estimates according to ref. 2 to get 23 and 123 mm, respectively, as a result of GIS dynamics. Our estimate of 6 mm is 5–26% of these estimates.

From ref. 11, the estimate for the total SLR by 2100 is 0.5–1.4 m, over the period from 1990–2100. The time period modeled here is 2000–2100, so we compare our SLR estimate to \((2100 – 2000)/(2100 – 1990) = 91\%\) of these estimates. Multiplying the lower bound of 0.5 m by 0.91 + \( \phi \) then gives 45–68 mm of SLR from GIS dynamics by 2100. Our estimate of 6 mm is 9–13% of this value. Multiplying the upper bound of 1.4 m by 0.91 + \( \phi \) gives 127–191 mm of SLR. Our estimate is 3–4% of this value. A lowest possible lower bound of 3.5 mm\,y\(^{-1}\) is also given in ref. 11, for a total of 350 mm of total SLR by 2100. Multiplying this number by \( \phi \) gives the fraction from GIS dynamics, which is 35–53 mm. Our estimate of 6 mm is 11–17% of this value.

From ref. 12 the estimated range for the total, global SLR by 2100 is 0.6–1.6 m. Following the same approach as above for ref. 11, we have 60–90 mm and 160–240 mm for the lower- and upper-bound estimates of SLR from GIS dynamics by 2100. Our estimate of 6 mm is 7–10% and 2–4% of these estimates, respectively.

Finally, we note that there are other studies based on semiempirical models that we could also compare our results to. However, the estimates for the total SLR by 2100 from those studies falls within the range of those discussed above.

Fig. S1. Modeled versus target velocities (speed) and rms differences for (A) the entire GIS, (B) JI, (C) HG, and (D) KG.

Fig. S2. Time series of stress boundary condition forcing factor, $C_1$. For JI, a single value is applied initially and held for the duration of the experiment, as discussed in the text. For HG and KG, $C_1$, is altered in time as needed so that modeled and observed flux changes are in agreement. Solid curves with circles denote the applied value of $C_1$ over time (left-hand axis). The dashed red and blue curves denote flux anomalies for HG and KG, respectively (right-hand axis), and the black-dashed curve shows their ratio (left-hand axis); the larger flux anomaly for KG relative to HG requires a relatively larger value of the boundary forcing.
**Fig. S3.** Normalized, cumulative rate of SLR curves for JI, HG, KG, and the normalized total of those three. The net SLR as a function of time is obtained by multiplying these normalized curves by an initial rate of SLR. Gray shaded curves marked \( \sigma \) denote an estimate for the uncertainty, which is \( \pm 35\% \) of the mean (black curve).

**Fig. S4.** Cumulative SLR (solid black line) as in Fig. 4B but assuming a perturbation recurrence interval of 50 y (colored lines, the sum of which equal the solid black line). For reference, the black-dashed line represents the cumulative SLR by 2100 under the assumption that the dynamic imbalance estimate from ref. 10 persists out to 2100.
Fig. S5. As in Fig. S4 but for a perturbation recurrence interval of 20 y.

Fig. S6. As in Fig. S4 but for a perturbation recurrence interval of 10 y.