



Nudging an inversion-based ice sheet model towards observed mass change during model relaxation: application to the C.H. Ostenfeld Glacier.

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Abstract.

Snapshot inversions are widely used to initialize numerical ice sheet models, but they do not necessarily produce a dynamically
15 consistent state for transient simulations. The transition from an inverted snapshot to a prognostic simulation can generate a
substantial initial model shock in ice thickness, velocity, and potentially unrealistic transitions between grounded and floating
ice. To reduce these effects, modellers commonly use a relaxation period during which the ice sheet model is allowed to adjust
towards a dynamically consistent state, often at the expense of increasing errors relative to the observed present-day geometry
and velocity. Here, we introduce a nudging method for that relaxation period that reduces the modelled ice geometry drift
20 while allowing the model to adjust dynamically. The method is designed to (i) reduce model drift, (ii) limit the increase of
errors in ice geometry with respect to observations during the relaxation period, and (iii) enable transient simulations to be
consistent with observed present day thinning rates. We apply the method to the C.H. Ostenfeld Glacier in northern Greenland.
We find that nudging substantially reduces the drift in ice geometry and grounding-line position during the relaxation period
and that it enables the model to capture the observed present-day thickness change rate. The method provides a viable, simpler,
25 and cheaper alternative to a transient calibration, commonly used in spin-up based ice sheet models. The method is not ice
sheet model-specific and will, when implemented widely across models, contribute to reducing the uncertainty in sea level rise
projections by better constraining model initial dynamic states.

1 Introduction

Today, the Greenland Ice Sheet (GrIS) is the largest contributor to global mean sea level rise of all cryospheric contributions
30 and second to thermal expansion (Fox-Kemper et al., 2021). Recently, its contribution to sea level rise has accelerated (The



Imbie Team, 2020; Poinar et al., 2024), likely because of the near-universal retreat of the marine outlet glaciers (Greene et al., 2024). In the period 1972–2018, the ice mass loss of the GrIS was almost equally divided between losses through atmospheric warming and ice discharge through increased ice velocities and increased grounding line fluxes (Mouginot et al., 2019). The projected sea level rise contribution from the GrIS, simulated with four state of the art ice sheet models, ranges from 16–
 35 76 mm (SSP1-2.6), to 22–163 mm (SSP2-4.5) and to 27–354 mm (SSP5-8.5) in the year 2100 (Goelzer et al., 2025).

An important shortcoming in present-day ice sheet models is their struggle to capture contemporary mass loss and to simulate observed present day mass change rates (Briner et al., 2020; Aschwanden et al., 2021; Pattyn et al., 2018). The observed present-day thinning rates of the GrIS range up to 10 m/yr at the fast flowing outlet glaciers Sermeq Kujalleq (Jakobshavn
 40 Glacier), Kangerlussuaq Glacier and Helheim glacier (Smith et al., 2020). Ice sheet models typically fail to capture those rates, which calls into question their ability to capture future changes (Aschwanden et al., 2021).

There are broadly two ways to initialize an ice sheet model to present-day observations of ice thickness, ice surface velocities, and thinning rates: through a long-term spin-up or by using data-assimilation. The spin-up method, used for example by
 45 Winkelmann et al. (2011); Pollard and Deconto (2012); Greve and Blatter (2016); Berends et al. (2022); Lipscomb et al. (2019) runs a simulation forward with chosen present-day or historical forcing while optionally nudging uncertain parameters. Data assimilation is used for example by Macayeal (1993); Larour et al. (2012); Åkesson et al. (2021); Barnes and Gudmundsson (2022) and is solving the momentum balance of the ice sheet model and adjusting uncertain parameters to match the modelled ice surface velocities to observations.

Many studies that employ data assimilation acknowledge the transient adjustment, or initialization shock, that often occurs at the start of a forward simulation initialized from an inverted state and therefore include a relaxation period prior to the analysis period (e.g Bett et al. (2023); Rosier et al. (2024); Trevers et al. (2024); Williams et al. (2026); Holmes et al. (2025); Åkesson et al. (2022)). During this relaxation phase, the model is allowed to evolve toward a more dynamically consistent state, reducing drift associated with the inversion. Some studies have used the relaxation period to calibrate the surface mass
 55 balance forcing so as to reduce model drift and improve agreement with thinning rates (Trevers et al., 2024). Others have incorporated observations of mass change directly into the inversion cost function, with the aim of obtaining an initial state that is more consistent with both the observed velocity field and the observed mass evolution, thereby reducing the magnitude of subsequent transient adjustments (Gudmundsson et al., 2023).

In this study, we present a new approach for using the relaxation phase of a data-assimilation-based ice-sheet initialization to
 60 reduce model drift and to include another observational dataset, the thinning rates, to constrain unknown parameters. The method incorporates a nudging procedure, similar to those commonly used in spin-up approaches, into the post-inversion relaxation period. During this phase, nudging constrains the modelled ice thickness toward present-day observations while



still allowing the model to evolve toward a dynamically consistent state. The strength of the nudging can be adjusted to control the trade-off between maintaining agreement with the observed ice geometry and preserving the velocity field obtained through the inversion. In this way, the method combines the ability of data assimilation to reproduce observed velocities with the ability of spin-up techniques to maintain consistency with observed geometry and mass evolution. We demonstrate the approach using a realistic model setup of the present-day C.H. Ostenfeld Glacier in northern Greenland. Our overarching aim of developing this method is to constrain the modelled present-day state of this glacier better and to increase our confidence in future sea level rise projections, as well as to apply it to modelling scenarios encompassing the entire GrIS in the future.

2 Methods

We employ the thermomechanically coupled Ice-sheet and Sea-level System Model (ISSM; Larour et al. (2012)) which has been demonstrated to work successfully at GrIS outlet glaciers (Åkesson et al., 2022; Holmes et al., 2025; Barnett et al., 2026b).

2.1 Ice sheet model setup

We use the Blatter-Pattyn (BP) approximation (Blatter, 1995; Pattyn, 2003) to the momentum balance in ISSM:

$$\nabla_h \cdot \boldsymbol{\sigma}_{bp} + \frac{\partial}{\partial z} \left(\eta \frac{\partial \mathbf{u}}{\partial z} \right) = \rho g \nabla_h s, \quad (1.1)$$

where $\boldsymbol{\sigma}_{bp}$ is the horizontal deviatoric stress tensor, η the viscosity of the ice, $\mathbf{u}$ the velocity vector, ρ the density of ice, g the gravitational constant, s the surface height, ∇_h the horizontal gradient operator and $\frac{\partial}{\partial z}$ the partial derivative with respect to the vertical coordinate z . We assume the surface is stress free, and for the basal boundary condition under grounded ice, i.e. basal friction, we employ the Weertman sliding law:

$$\boldsymbol{\tau}_b = -C(\mathbf{x}, y) |\mathbf{u}_b|^{m-1} \mathbf{u}_b \quad (1.2)$$

where $\boldsymbol{\tau}_b$ is the basal friction, $C(\mathbf{x}, y)$ is a spatially varying tuneable parameter accounting for missing processes such as effective pressure, basal hydrology and basal roughness, and $\mathbf{u}_b$ is the basal ice velocity. The exponent m is chosen to be $1/3$ following Åkesson et al. (2022) and Rückamp et al. (2026).



Floating ice shelves are assumed to have zero basal friction and, unlike grounded ice, experience basal melting due to contact with ocean water. Basal melt (m_b) under the floating ice shelves is calculated linearly as follows:

$$m_b(z) = m_{\text{upper}} + \frac{m_{\text{deep}} - m_{\text{upper}}}{z_{\text{deep}} - z_{\text{upper}}} (z - z_{\text{upper}}) + P(x, y), \quad (1.3)$$

where the upper water melt rates (m_{upper}) at 10 meters below the sea surface are set to be 30 m yr⁻¹, while the deepwater melt rates m_{deep} at 500 meters below the surface are set to be 200 m yr⁻¹. The melt rates below and above the lower and upper limits z_{deep} and z_{upper} are capped at respectively the deepwater and upperwater melt rates. The spatially varying melt rate perturbation $P(x, y)$ can be tuned to match local ice thickness and/or thinning rates with respect to the observations.

For the surface mass balance, we use the statistically downscaled RACMO2.3p2 product at 1 km spatial and monthly temporal resolution, based on the ERA-40 (1958–1978), ERA-Interim (1979–1989), and ERA5 (1990–2024) reanalyses (Noël et al., 2019). To represent present-day conditions, we use the mean surface mass balance over the period 2000–2014 of this product.

We apply a non-moving ice front, since there is no agreed best approach to model calving rates of the GrIS (Choi et al., 2018; Barnett et al., 2026a). Our proposed nudging scheme described in Sect 2.3 is designed to target the basal melt for floating ice and friction for grounded ice, and cannot change the calving rates based on ice thickness, velocity or thickness change rate misfits. The ice front can in theory move inland, as the ice thins below the ice thickness threshold for which ice is considered to be dynamically active. In this study, we use a limit of 10 m.

The modelling domain consists of the drainage basin of C.H. Ostenfeld glacier and its tributary Harder glacier and Brikkerne glacier in Northern Greenland and the Victoria Fjord those glaciers drain in. The horizontal mesh resolution ranges from 1000 meters in the slow moving interior to 250 meters in areas where the ice is flowing faster than 200 m yr⁻¹. Five vertical layers are employed.

2.2 Inversion

The spatially variable parameter in the friction law in Eq. (1.2), $C(x, y)$ is found by iterating the momentum balance using observed ice thickness from Bedmachine v5 (Morlighem et al., 2022). To do so, we minimize a function J composed of the logarithmic misfit between observed and modelled ice surface velocities and a regularization term, as a function of the unknown friction parameter, over the part of the model domain that contains grounded ice:

$$J(C) = \gamma_1 \int_{\Gamma} \left[\ln \left(\frac{\sqrt{u_m^2 + \epsilon^2}}{\sqrt{u_o^2 + \epsilon^2}} \right) \right]^2 d\Gamma + \gamma_2 \int_{\Gamma} |\nabla C|^2 d\Gamma. \quad (1.4)$$



In Eq (1.4), $\mathbf{u}_m$ is the modelled ice surface velocity, $\mathbf{u}_o$ the observed ice surface velocities from the MEaSUREs program (Joughin et al., 2016), ϵ is a small number to prevent divisions by zero or taking the logarithm of zero. The coefficients γ_1 and γ_2 are weighting factors that control the relative importance of the velocity misfit and the regularization term in the cost function. These weights determine whether the inversion prioritizes minimizing velocity errors or enforcing spatial smoothness in $C(x, y)$. To find an optimal balance between low ice surface velocity errors and overfitting, we perform an L-curve analysis (Wolovick et al. (2023), Höyns et al. (2025)).

After the inversion and L-curve analysis to determine the most optimal trade off between smoothing and velocity errors, the ice sheet model is brought into thermal steady state using the surface temperatures and geothermal heat flux of the SeaRISE project (Bindshadler et al. (2013)), after which a future simulation can start. All forward simulations during the relaxation phase and the transient forward runs are ran with a timestep of half a month ($1/24^{\text{th}}$ year).

2.3 Relaxation period and nudging

Typically, after an inversion and thermal steady state (Section 2.2), the modeled ice sheet reproduces the observed ice thickness by construction and yields a close fit to the observed surface velocity field. When the surface mass balance and basal mass balance are imposed, this results in a spatial pattern of ice thickness change rates that are however not consistent with observations. As the model is integrated forward in time, these rates typically relax toward a (quasi-)equilibrium state, often desirable for future projections. Typical relaxation periods of GrIS outlet glacier ice modelling studies using ISSM are 5 – 50 years (Barnett et al., 2026b; Åkesson et al., 2022; Rückamp et al., 2026; Holmes et al., 2025; Jager et al., 2024). Other modelling studies on Greenland and Antarctica show that the relaxation period can be up to 250 years (Williams et al., 2026; Bradley and Hewitt, 2024; Trevers et al., 2024). Using such a relaxation period comes at the cost of a less accurate ice thickness and ice surface velocities.

To increase the control over the evolving ice sheet, we introduce the nudging procedure typically used in spin-up ice sheet models (Pollard et al., 2015; Lipscomb et al., 2019; Van Den Akker et al., 2026b) as a way to improve the relaxation procedure of data assimilation models. In this procedure, uncertain parameters such as $C(x, y)$ in Eq. (1.2) are nudged during a forward transient run based on the misfit between modelled and observed ice sheet thickness, and based on the thickness change rate of the modelled glacier. For basal friction nudging, we use the following equation, adapted from Van Den Akker et al. (2026b):

$$\frac{\Delta C_1}{\Delta t} = -\frac{H - H_{obs}}{\tau H_0} - \frac{1}{H_0} \frac{\Delta H}{\Delta t} - \frac{r}{\tau} (C_1 - C_i) \quad (1.5)$$



$$C_1 = \log_{10} C(x, y). \quad (1.6)$$

140 In Eqs. (1.5) – (1.6), Δt is a model timestep, ΔC_1 the change in C_1 during a model timestep, H the modelled ice thickness, H_{obs} the observed ice thickness, H_0 the thickness scale, ΔH the ice thickness change during a model timestep, τ the timescale, r the relaxation strength and C_i the relaxation target. The three terms represent nudging based on ice thickness, thickness change rate, and a prescribed relaxation target. The first term adjusts the friction coefficient in response to deviations between modeled and observed ice thickness. For example, if the modeled ice is too thick ($H - H_{obs} > 0$), the

145 first term is negative and the friction coefficient is reduced, promoting faster ice flow and enhanced ice discharge. The second term responds to the rate of thickness change: when the modeled ice thickness is increasing ($\Delta H / \Delta t > 0$), the friction coefficient is likewise reduced to counteract the positive mass imbalance and limit further thickening. The third term acts as a regularization or relaxation constraint, pulling the evolving friction field toward a target field C_i . In Van Den Akker et al. (2026a), this target field was introduced to prevent overfitting and was parameterized as a function of bedrock elevation.

150 Here, however, a spatially varying friction field is already available from a snapshot inversion procedure that includes smoothing and regularization. Consequently, the relaxation term can be used to constrain the updates to the friction field toward the inversion-derived solution. A similar approach is taken for the linear basal melt scheme in Eq. (1.3), targeting the melt rate perturbations $P(x, y)$:

$$\frac{\Delta P(x, y)}{\Delta t} = m_{m,0} \left(\frac{H - H_{obs}}{\tau_m H_{0,m}} + \frac{1}{H_{0,m}} \frac{\Delta H}{\Delta t} - \frac{r_m}{\tau_m} (P(x, y) - P_i) \right) \quad (1.7)$$

In Eq. (1.7), $m_{m,0}$ is a scaling constant, $H_{0,m}$ is the thickness scale, τ_m is the timescale and r_m the relaxation strength and P_i

155 the relaxation target. Ideally, there is no need for a melt rate perturbation (i.e. the basal melt rates fit together perfectly with the ice flux and surface mass balance rates to produce an ice sheet in its desired dynamic state) so the relaxation target P_i is chosen to be zero.

A correction to the mass transport equation during the relaxation period can be applied in the form of the mass change rates by Smith et al. (2020) to be able to start a continuation run from the present day observed mass change rates, as was

160 described for the Antarctic ice sheet by Van Den Akker et al. (2025). It is based on a correction to the mass transport equation of the form:

$$\frac{\Delta H}{\Delta t} = -\nabla \cdot (\bar{\mathbf{u}}H) + B - C. \quad (1.8)$$



In Eq. (1.8), $\bar{\mathbf{u}}$ is the vertical averaged ice velocity vector B represents the combined surface and basal mass balance, and C is the correction term applied during relaxation, taken here as the observed present-day thickness change rates from Smith et al. (2020). The purpose of C is to compensate for imbalances between the modeled ice fluxes and the observed mass budget, thereby maintaining the ice sheet close to its observed geometry throughout the relaxation period. Ideally, the relaxation period converges to a state in which the modeled thickness tendency approaches zero, i.e., $\partial H / \partial t \approx 0$. At this point, the ice flux divergence, surface mass balance forcing, and correction term are in approximate balance. When the correction term C is subsequently removed at the start of the forward simulation, the initial thickness change rates becomes $\partial H / \partial t \approx -C$. Consequently, the model begins its transient evolution with thickness change rates that are approximately equal to the observed present-day thickness change rates, which is the intended outcome of the relaxation period in this study.

2.4. Ensemble design

To showcase the effect of the different terms in the nudging procedure, we vary the thickness scale H_0 , the timescale τ and the relaxation strength r . For the thickness scale we test 75, 150, 300 and 500 meters, initial experiments showed that going below 50 m increases the model's instability significantly. For the timescale we choose 500 and 5000 years. For the relaxation strength we use 0 (no relaxation), 0.5 (medium relaxation) and 1 (full relaxation). Our nudging ensemble runs for 350 years, to give the nudging approach ample time to adjust the basal melt rate perturbation and the friction coefficient, and to give the non-nudging comparison simulation the opportunity to find a new stable equilibrium with little model drift. Runs are named using the parameter values for H_0 , τ and r e.g. H150_T500_R0.500.

We then choose a 'best-case' simulation from the nudging ensemble and run that simulation forward for 100 years into the future, while keeping atmospheric forcing constant at their present-day conditions and the friction coefficient and basal melt perturbations constant at their nudged values. This is to show what the model drift left in the ice sheet model leads to when not using nudging or when using the nudging without incorporating the present day thickness change rates. These simulations are also used to assess what the present day thickness change rates under constant present-day forcing will eventually lead to for the C.H. Ostensfeld glacier.

3. Model domain: the C.H. Ostensfeld Glacier

C.H. Ostensfeld Glacier is an outlet glacier draining the north-western GrIS, and terminates in Victoria Fjord. In contrast to neighbouring Ryder and Petermann glaciers, C.H. Ostensfeld Glacier experienced an abrupt loss of its approximately 20 km long floating ice tongue during the winter of 2002–2003 (Hill et al., 2018). This collapse has been attributed to the intrusion of relatively warm Atlantic Water into Victoria Fjord, which may have enhanced basal melting and destabilised the ice tongue (Jakobsson et al., 2026). In contrast, Ryder and Petermann glaciers are protected by submarine bedrock sills that restrict the

access of Atlantic Water to their respective fjords, potentially reducing their exposure to ocean-driven basal melting and saving their floating ice shelves (Jakobsson et al., 2020).

C.H. Ostenfeld Glacier, together with its tributary glaciers Harder and Brikkerne, contains a potential global mean sea-level equivalent of approximately 3.9 cm if the entire drainage basin were to be depleted (Millan et al., 2023). Ice thicknesses exceed 2500 m in the interior of the basin (Morlighem et al., 2022) (Fig. 1a), while ice velocities at the grounding line reach roughly 900 m yr⁻¹ near the grounding line (Fig. 1b) (Joughin et al., 2016). Thickening rates range between ~0.1 m yr⁻¹ in the interior to ~-1 m yr⁻¹ close to the grounding line (Fig. 1c). Integrating these mass change rates indicates that the C.H. Ostenfeld glacier is slightly out of balance, having a net mass change of -0.17 Gt yr⁻¹ (Smith et al., 2020).

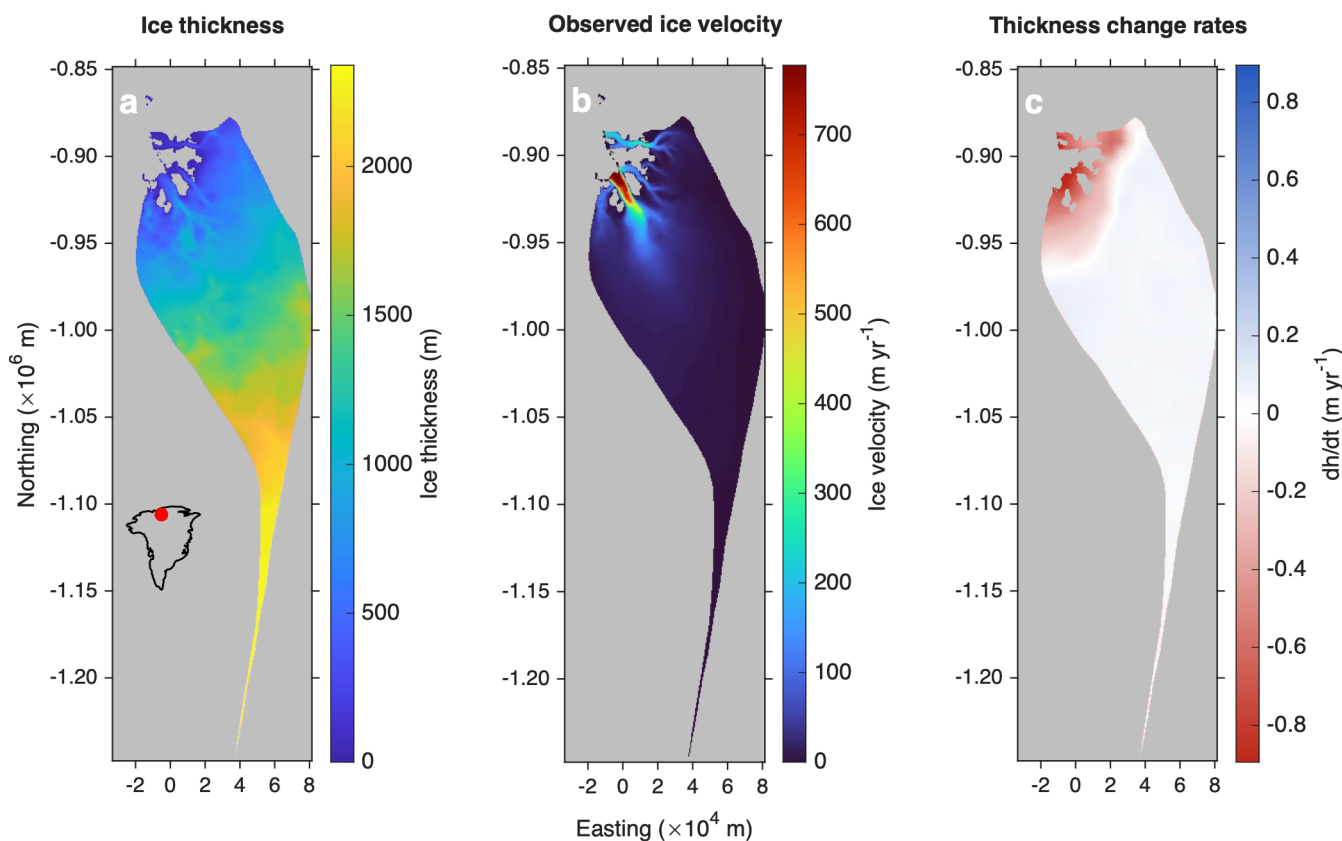


Figure 1. Overview of the C.H. Ostenfeld Glacier drainage basin showing (a) ice thickness and its position on the GrIS, (b) observed surface ice velocities, and (c) mass change rates of Smith et al. (2020), interpolated to the model grid. Areas with ice thickness below 10 m are masked and treated as inactive ice (and discarded from further analysis).



205 4 Results

4.1 C.H. Ostenfeld inversion and initialized state

The resulting spatial patterns in the modelled ice velocity and inverted friction coefficient for the selected optimal inversion setup are shown in Fig. S1. The inversion reproduces the main characteristics of the observed ice-surface velocity field of C.H. Ostenfeld Glacier reasonably well, capturing both the spatial extent of the main flow features and the transition towards slower-
 210 moving ice further upstream of the grounding line. However, the modelled velocities along the fast flowing main trunk of the glacier remain somewhat lower than the observations, indicating a residual velocity misfit despite an overall good agreement between the model and observations.

4.2 C.H. Ostenfeld relaxation period nudging

215 The results of the nudging ensemble in terms of ice thickness Root Mean Square Error (RMSE) compared to the observed ice thickness of Morlighem et al. (2022) are shown in Fig 2. All simulations are initialized with the observed ice thickness and the inverted quantities as described in Sect. 2. During the first decade of the relaxation period, the modelled C.H. Ostenfeld Glacier undergoes substantial model drift as the inverted parameters, surface mass balance, and basal melt scheme are introduced. In the absence of nudging (the non-nudging baseline, grey dashed lines in Fig. 2), the simulation continues to drift away from the
 220 observed ice thickness throughout the 350 year period, after which the RMSE exceeds 25 m. In contrast, all simulations employing nudging and the mass transport correction from Eq. (1.8) rapidly approach a quasi-stable state, with ice-thickness RMSE values of at most 17 m after 350 years. Note that a low RMSE in the simulation that have $\partial H/\partial t$ forcing does not mean that the model will be in steady state in a future simulation, since the observed thinning rates are accounted for via the correction term C in Eq. (1.8). A low RMSE characterizes models that are consistent with the patterns of ice thickness and additionally,
 225 when ran forward, are matching observations of ice thickness change.

For most configurations, the drift in ice-thickness RMSE is largely stabilised within the first 100 years of simulation. After 350 years, however, the ice-thickness RMSE is still continuing to increase in the non-nudged simulation, indicating a persistent departure from the observed geometry. Interestingly, not applying the mass-change-rate correction described by Eq. (1.8) results in a somewhat less stable model configuration. For most configurations not using the mass transport
 230 correction, the ice-thickness RMSE remains slightly positive in its temporal trend at the end of the 350-year simulation period, see Fig. 2a, suggesting that the correction does not fully prevent long-term drift in the simulated ice geometry.

Clearly, the strongest control on the RMSE thickness is the value of H_0 (Fig. 2a) Decreasing this value increases the relative importance of the first terms in Eqs. (1.5) and (1.7), causing the model to prioritize reducing these terms at the expense of the tendency and relaxation terms. In general, increasing the strength of the relaxation term slightly increases the thickness error.
 235 This suggests that stronger relaxation causes the model to prioritize maintaining the friction coefficient field close to that obtained from the inversion, rather than minimizing small thickness errors or model tendencies. Nevertheless, varying the



relaxation parameter across its full range (0–1) produces substantially less uncertainty than doubling the value of H_0 . Increasing the timescale τ from 500 to 5000 creates a similar simulation as when using a relaxation weight of 1.0 for simulations with τ being 500, and using τ of 5000 makes the model simulations on this timescale, 350 years, almost
 240 insensitive to changes in relaxation weight.

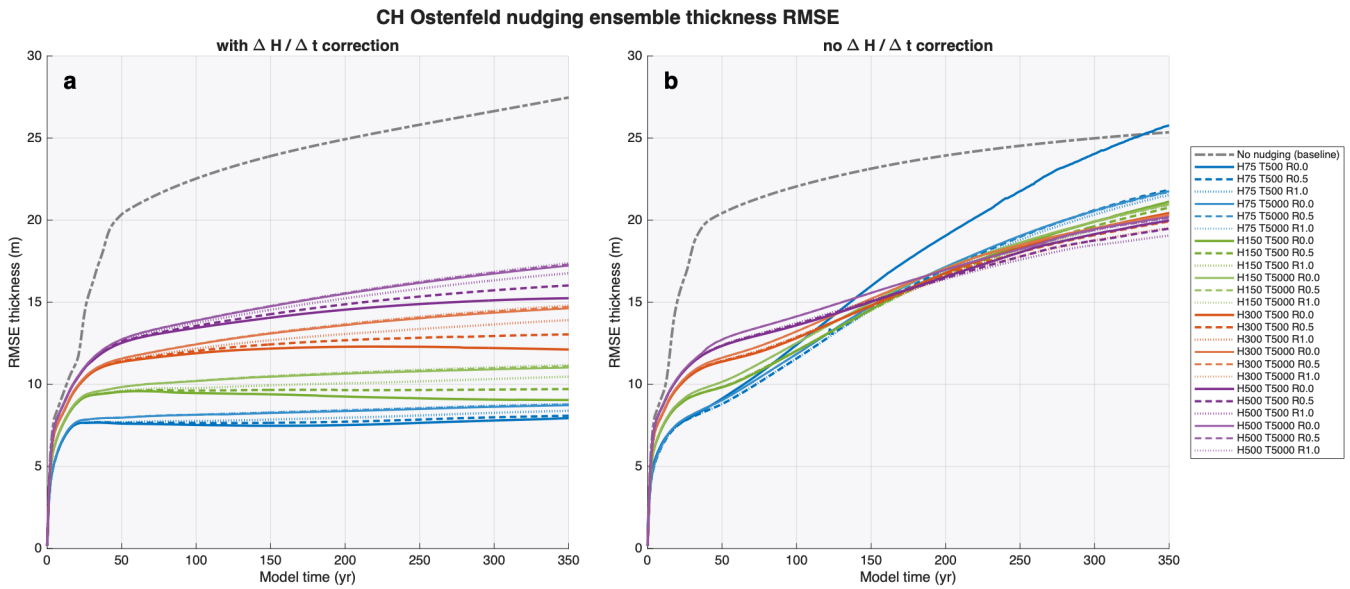


Figure 2. Ensemble of nudging parameters in terms of thickness RMSE compared to observations when using Eq. (1.8) with C being set to the observed mass change rates (a) and $C=0$ (b), compared to a relaxation period without nudging (grey dashed line in both plots, also with and without the mass transport correction). Blue lines indicate simulations with $H_0 = 75$, green lines with $H_0 = 100$, orange lines
 245 with $H_0 = 300$, and purple lines with $H_0 = 500$. Solid lines have $r = 0$ in Eq. (1.7). Dashed lines have $r = 0.5$, dotted lines have $r = 1$. Light colors have a longer time scale τ in Eq. (1.7) of 5000 years.

In Fig. 3, the RMSE in terms of ice velocity is shown. The opposite pattern is visible compared to the thickness RMSE in Fig. 2: increasing H_0 decreases the velocity error, and increasing the relaxation weight also decreases the velocity error. Using $H_0 = 75$ now brings the ice surface velocity RMSE in the simulations with the mass change rates (Fig. 2a) from approximately
 250 23 m yr^{-1} to 27 m yr^{-1} , compared to using $H_0 = 500$. Both are expected: the starting value and relaxation target of the friction coefficient are inverted given the observed surface velocities. Drifting away from this field will cause the ice surface velocities to drift away further from the observed target too. Making the relaxation term stronger by increasing its weight or by decreasing the weights of the thickness and tendency term by increasing H_0 will keep the friction coefficient field closer to its original value, thereby decreasing the ice surface velocity RMSE. The velocity RMSEs are comparable between the nudging and the
 255 no nudging baseline case. This is desirable, since the nudging procedure decreases then the ice thickness errors and introduces the thickness change rates at the cost of almost no increase, or in some cases even a drop, in the velocity errors.

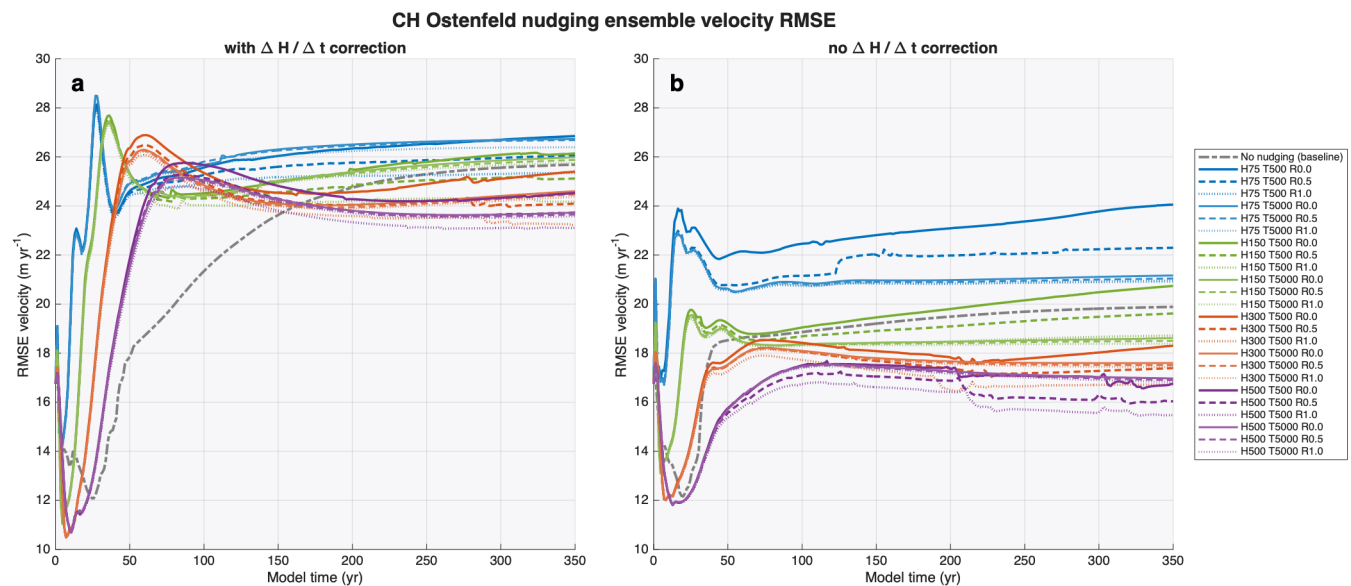
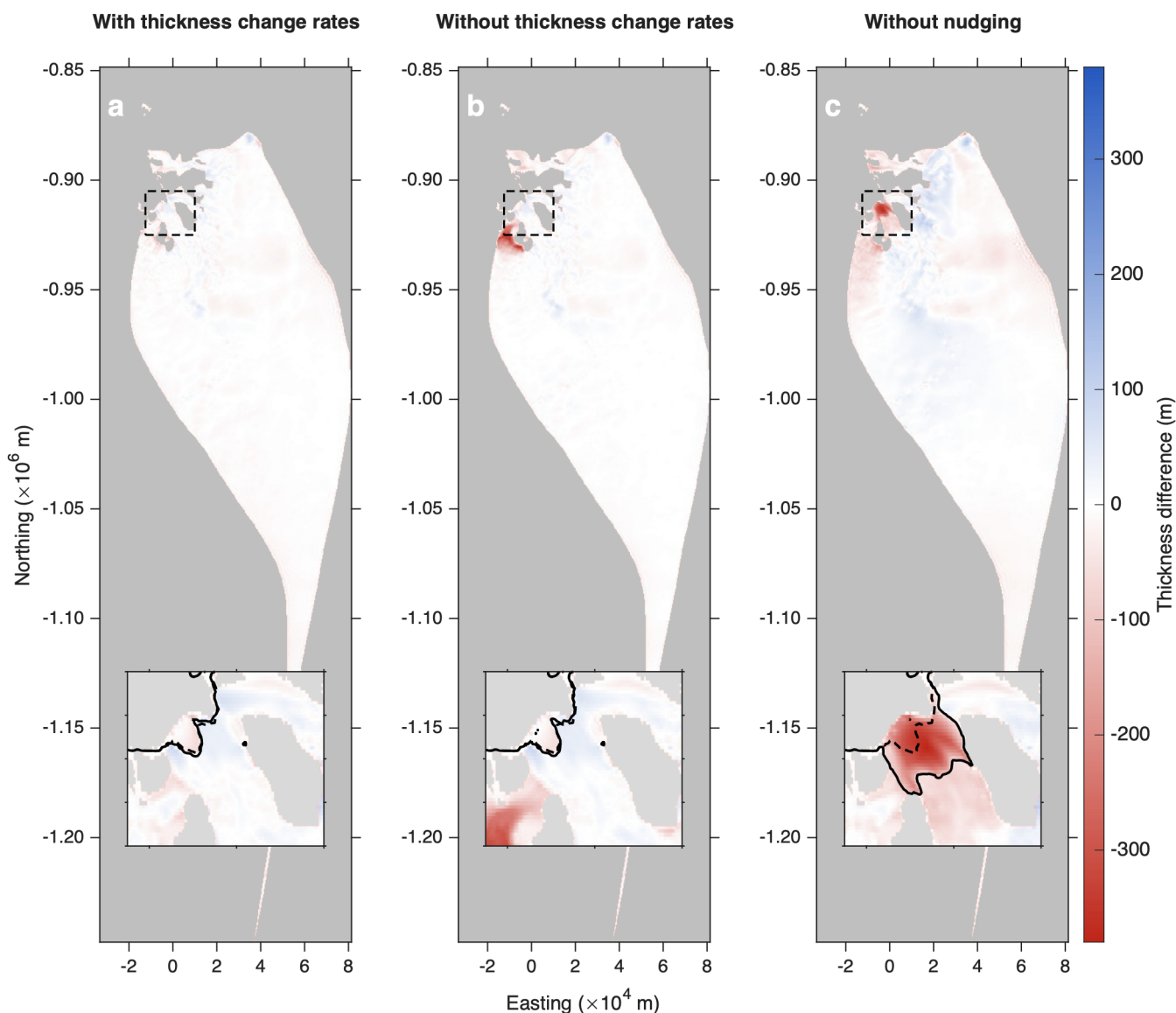


Figure 3. Ensemble of nudging parameters in terms of velocity RMSE compared to observations when using Eq. (1.8) with C being set to the observed mass change rates (a) and $C = 0$ (b), compared to a relaxation period without nudging (grey dashed line in both plots, also with and without the mass transport correction). Blue lines indicate simulations with $H_0 = 75$, green lines with $H_0 = 100$, orange lines with $H_0 = 300$, and purple lines with $H_0 = 500$. Solid lines have $r = 0$ in Eq. (1.7). Dashed lines have $r = 0.5$, dotted lines have $r = 1$. Light colors have a longer time scale τ in Eq. (1.7) of 5000 years.

The simulation configuration H150_T500_R0.500 exhibits one of the lowest stabilized thickness RMSE among the tested ensemble members, together with a comparatively low velocity RMSE. This configuration is therefore selected as the reference nudging simulation for future projections. Fig. 4 shows the modelled thickness anomaly relative to the BedMachine v5 geometry Morlighem et al. (2022), for the full model domain and for the grounding-line region, respectively, at the end of the 350 year relaxation period. Here it can clearly be seen that applying the nudging decreases the modelled ice thickness and grounding line position error during the relaxation period. Applying the observed ice thickness change rates also reduces the modelled ice thickness misfit at a nameless stagnant western tributary glacier. Here, the surface mass balance and the ice thickness rates are negative and the ice velocities are low. Through Eq. (1.8), $B - C$ becomes less negative when C are the observed thinning rates, making it easier for the ice sheet model to compensate for a negative surface mass balance through an increased ice flux.



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Figure 4. Ice thickness error (modelled – observed) with respect to the observations of Morlighem et al. (2022) at the end of the relaxation period (350 years) with nudging simulations H150_T500_R0.500. Using the mass change rates as described by Eq 1.8 and Van Den Akker et al. (2025) is shown in panel (a), without the present day mass change rates included (b) and when not using the nudging procedure i.e. a classical relaxation period simulation (c). (**Insets**) Ice thickness error zoomed in at the grounding line zone. The observed grounding line is shown with a dashed line, the modelled grounding line of the respective simulation is shown with the solid black line.

Fig. 5 presents the ice-surface velocity error across the full model domain and, in greater detail, within a zoomed region around the grounding line. These results underscore a key strength of the nudging approach: it introduces no significant trade-off in ice surface velocity performance. Although the ice-thickness errors decrease substantially once nudging is applied (Fig. 4), the



285 velocity errors remain comparable across all three simulations. In each case, the modelled ice sheet remains too slow along the
 main flowline of C.H. Ostenfeld glacier and slightly too fast along the Harder and Brikkerne glaciers.

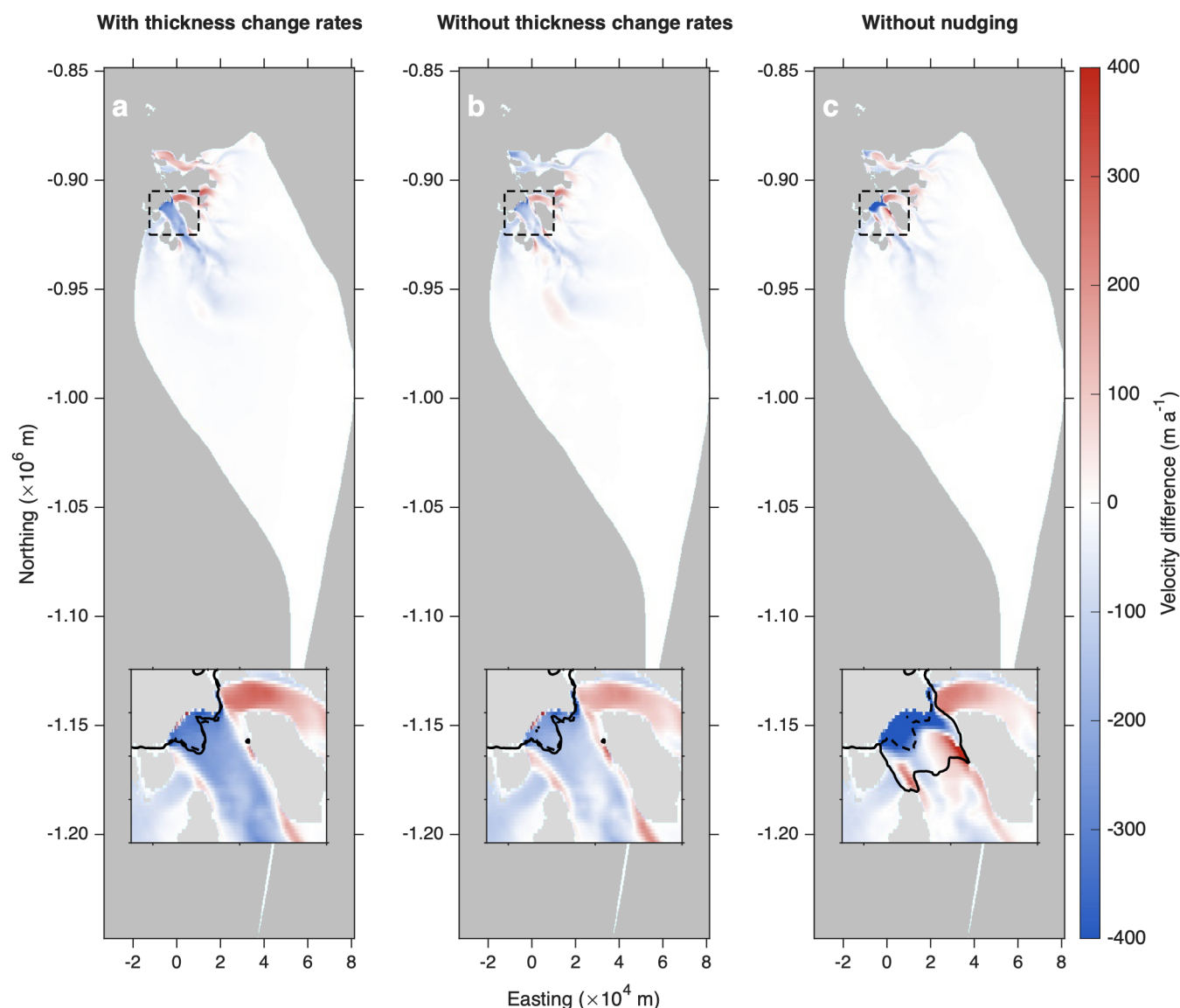


Figure 5. Ice surface velocity error (modelled – observed) with respect to the observations of Joughin et al. (2016) at the end of the relaxation
 period (350 years) with nudging simulations H150_T500_R0.500. Using the mass change rates as described by Eq 1.8 and Van Den Akker
 et al. (2025) is shown in panel (a), without the present day mass change rates included (b) and when not using the nudging procedure i.e. a
 classical relaxation period simulation (c). (**Insets**) Ice surface velocity error (modelled – observed) with respect to the observations in the
 grounding line zone, with the observed grounding line shown by a dashed line, the modelled grounding line of the respective simulation is
 shown with a solid black line.



4.4 C.H. Ostenfeld forward simulations

We then run the chosen best-performing configuration, H150_T500_R0.500, forward for a single transient time step to assess its instantaneous mass change rates when starting a future simulation. The resulting thickness change rates are shown in Fig. S2. The simulation without nudging (Fig. S2c) exhibits a very weak but spatially uniform thinning in the interior of the ice sheet. Notably, the spatial pattern of simulated mass change from this simulation differs substantially from the observed mass change rates (Fig. 1c). This indicates that the integrated mass loss rate might be negative and might even be close to the observed value, but that the model does not reproduce the present-day pattern of thickness evolution. Using a relaxation period in the classical way (i.e. running forward freely until the drift decreases) does in this case not lead to a stable ice sheet, or one that is thinning according to the mass change rates observations. When using nudging without the observed thickness change rates (Fig. S2b), the nudging induces only visible thinning at the nameless stagnant western tributary glacier where the surface mass balance is negative. The ice velocities (Fig. 1b) are already low there, so the nudging cannot counteract that. The simulation with the observed thickness change rates (Fig. S2c) reproduces those observed rates (Fig. 1c): the difference between modelled and observed rates are small. This demonstrates that the nudging approach can substantially improve the transient mass-balance behaviour of the model, enabling a forward simulation with mass change rates that closely match observations while maintaining a geometry and velocity close to the present-day state.

Then, we ran the three simulations shown in Figs. 4, 5 and S2 forward for 100 years under constant oceanic and atmospheric forcing, i.e., linear basal melt rates based on draft depth with the nudged perturbation rates and fixed SMB, respectively. The resulting spatial patterns of ice thickness change are shown in in Fig. 6. They further confirm that nudging without the mass-change rates removes almost all of the model drift except for the nameless stagnant western tributary glacier, and the modelled grounding line is after 100 years close to its observed present-day location (Fig. 6b). In the simulation including the mass-change rates (Fig. 6a), the spatial pattern of ice thickness change over the 100-year period largely follows the observed present-day mass-change pattern, with a retreated grounding line. This suggests that the imposed mass-change rates represent a robust signal that can persist for several decades after their initialization in ISSM. There is no indication of an irreversible grounding line retreat in the C.H. Ostenfeld Glacier under a 100-year simulation with constant climate forcing, but also not for a stabilization under current climate conditions.

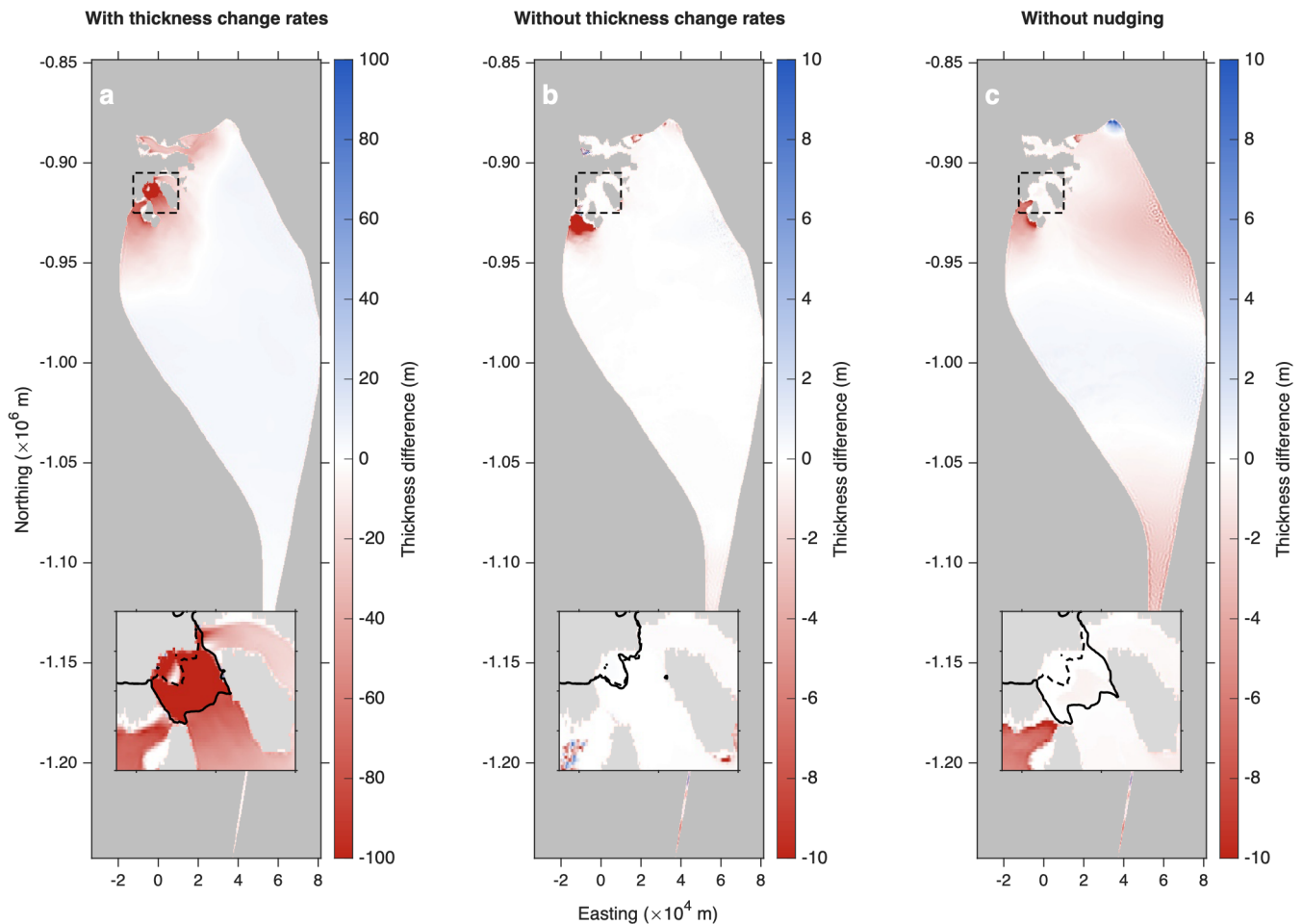


Figure 6. Spatial patterns of ice thickness change and grounding line movement for the three simulations after 100 years transient simulations with constant oceanic and atmospheric forcing, initialized using nudging and the observed thickness change rates (a), without the observed thickness change rates (b) and without nudging and without the observed thickness change rates (c). Note the different scales of the colorbars for (a) and (b,c).

5 Discussion

Snapshot inversions such as the one used here reproduce selected present-day observations, typically surface velocity, but do not guarantee a dynamically consistent state once integrated forward, particularly when independent surface and basal mass-balance products are introduced (Gillet-Chaulet et al., 2012; Larour et al., 2012; Åkesson et al., 2021; Holmes et al., 2025; Jager et al., 2026). Relaxation is therefore an almost near-universal requirement after inversion-based initialization. Here we show that nudging techniques more commonly used during long ice-sheet spin-ups (Deconto et al., 2021; Lipscomb et al.,



2021; Van Den Akker et al., 2026b) can instead be applied specifically during this post-inversion relaxation phase, giving explicit control over the departure of geometry and velocity from observations rather than letting the model drift freely, and at the same time add the possibility to add another dataset, the observed ice thickness change rates, to constrain uncertain parameters.

This differs fundamentally from nudging's usual role in a (long) spin-up. There, nudging is typically the primary mechanism to bring an unconstrained model, starting from an unknown friction field, toward an observed geometry over very long timescales. These timescales can go up to 10,000 or even 100,000 model years. Here, relaxation instead starts from a state already produced by inversion, which supplies a spatially varying friction field and a good initial fit to observed velocity. Nudging therefore acts as a final step to make the ice sheet modelled thickness, velocities, SMB and thickness change rate consistent, rather than the primary means of reaching an observationally consistent state. This, and the fact that the C.H. Osterfeld is a relatively small domain with low mass change rates, likely explains why most of the initial drift at C.H. Osterfeld Glacier is removed within roughly 100 years of nudging, two to three orders of magnitude faster than typically required in nudging-based spin-up (Deconto et al., 2021; Lipscomb et al., 2021; Van Den Akker et al., 2026b), thereby saving computational expenses.

An unconstrained relaxation of the C.H. Osterfeld glacier in this study proved to be an undesirable starting point for projections, even though some drift (including grounding-line retreat) is physically expected given the glacier's documented mass loss. Without nudging, that drift reflects a tangled mix of true dynamic imbalance and numerical or initialization artefacts (see the larger thickness error and greater grounding-line migration in the unnudged simulation in Fig. 4). The simulations including the observed thickness change rates additionally lets the initial mass trend of a projection run be prescribed consistently with observations rather than emerging as an uncontrolled by-product of applying the mass transport equation. Because our 100-year runs are initialized to the observed present-day mass-change rate and forced with fixed present-day conditions, their evolution can also be read as a form of committed ice loss, the glacier's response to its current state in the absence of further climate or ocean change, consistent with and complementary to the work of Badgeley et al. (2025). In their study, they transiently calibrate basal friction against a 2008–2020 velocity time series rather than relaxing from a single snapshot inversion, finding that ISSM simulations constrained to match ice thickness change observations project a larger sea-level contribution from northwestern Greenland. Together with the transient calibration method developed by Badgeley et al. (2025), ISSM now contains two ways of initializing a modelled ice sheet with the present-day observed thickness change rates, which can be used depending on the datasets available for the calibration: if timeseries of the ice velocities are available, one could use the method from Badgeley et al. (2025), if only snapshots are available, the method presented in this study can be used.



365 We refrain from presenting the 100-year constant-climate simulations results as a committed-sea-level-rise estimate for the following reasons: (i) the nudged friction and melt-perturbation fields are not extrapolated across the grounded/floating transition, so the model cannot yet consistently represent a migrating grounding line over longer integrations i.e. newly floating cells will not have a nudged basal melt perturbation, (ii) basal melt is a simple linear function of draft depth because the nudging ideally works on a cell based uncertain parameter and therefore acts on a perturbation to this rate, although
 370 future work could explore the potential of the nudging method on basin-averaged values, as was done for example by Lipscomb et al. (2021). More physically based schemes such as PICO or plume models (Reese et al., 2018; Pelle et al., 2019) could be incorporated without changing the underlying method, (iii) the calving front position is fixed (no-advance), a conservative choice excluding terminus retreat and its feedbacks, for which a physically based calving law such as a von Mises stress or crevasse-depth criterion would be a natural extension following the work of Barnett et al. (2026a), but also a
 375 challenge since the nudging as presented in this paper does not target uncertain parameters related to calving. It could therefore very well be that introducing calving during or after the nudging will lead to additional model shocks.

Other ice sheet models address the same initialization problem by constraining the relaxation itself, each in a different way. The ice sheet model WAVI uses a 250-year relaxation with accumulation adjusted from observed thickness-change rates, conceptually similar to our approach, while holding ice-shelf thickness fixed to suppress grounding-line migration and
 380 without applying nudging (Arthern et al., 2015; Bradley et al., 2024). BISICLES combines present-day accumulation with observed thickness change rates over a 50-year relaxation, likewise fixing ice shelf thickness (and hence the grounding line), but periodically re-inverts basal friction against observed velocities as geometry evolves (Cornford et al., 2013; Trevers et al., 2024). Úa, as used by Williams et al. (2026), has no dedicated relaxation phase described, but comparable spin-up procedures are common elsewhere in the Úa literature (Hill et al., 2024; Morlighem et al., 2024). Across all of these, one or
 385 more geometric constraints (ice front, grounding line position and/or shelf thickness) are imposed during the relaxation phases to suppress transient effects. However, the initial spatial pattern of modelled thickness-change rate after relaxation is generally not reported, making it hard to judge the size of the initial imbalance, how effectively relaxation removes it, or how the model would behave without the imposed constraints.

The nudged simulations of C.H. Ostfeld presented here produce an ice stream that is both slower and slightly thicker than
 390 observed, a combination needed to keep the ice shelf and grounding line near their present-day position, since ice flux scales with the product of thickness and velocity, and the two partly offset one another here. A likely driver is the model's representation of C.H. Ostfeld's floating terminus: BedMachine, used here, is timestamped at 2007 and features a floating ice shelf, whereas little to none remains today (Jakobsson et al., 2026). This gives the model an extra pathway for mass loss, through both basal melting and advection past the calving front, so modelled flux to the ice shelf likely exceeds the true
 395 present-day flux. During nudged relaxation the model compensates by reducing flux toward the calving front, at the cost of velocity, to limit upstream thinning and hold the grounding line in place. Without nudging, the model is instead free to



respond to this imbalance, and the grounding line retreats substantially, which is undesirable as an initialization target even where it may be a dynamically plausible response.

A further limitation concerns the initial fit to surface velocity. The snapshot inversion and L-curve analysis fit observations well overall but not perfectly, and the main flowline slows during relaxation as indicated by the lower velocities in the main flowline compared to the observations after the relaxation period. A likely explanation for this could relate to temporal and spatial inconsistency between the input datasets: the BedMachine v5 thickness field, the MEaSUREs velocity observations, and the mass-change-rate product used here do not represent exactly the same period, which can introduce inconsistencies in the inferred flux divergence (Goelzer et al., 2018; Aschwanden et al., 2021)

Looking ahead, the present-day mass-change-rate initialization presented here opens a route to testing whether currently observed mass loss alone is enough to commit a glacier to substantial retreat or acceleration, complementing studies of topographic and elevation-feedback controls on the stability of the GrIS (Feenstra et al., 2025; Petrini et al., 2025), while avoiding a potentially costly and/or uncertain simulation under pre-industrial climate conditions and glacier geometries. A natural next step is to apply the method to other rapidly changing marine-terminating glaciers, such as Jakobshavn Isbræ and Petermann Glacier. Such experiments could be designed around the Marine Ice Sheet Instability (Schoof, 2007, 2012) or the melt-elevation feedback, and would ask whether a glacier's present-day imbalance is by itself sufficient to initiate irreversible retreat. This would help separate future ice loss that follows from the continued present-day imbalance from loss that requires additional climatic or oceanic forcing.

6 Conclusion

In this study, we present a solution to explicitly control model drift during the relaxation method commonly used in inversion-based ice sheet models, by adapting a nudging approach typically applied during spin-up to the post-inversion relaxation period instead. It also allows an inversion based ice sheet model to start a simulation from the observed thinning rates. Uncertain parameters here are the basal friction coefficient and basal melt perturbations, and they are tuned against both the observed ice thickness and, critically, the observed present-day rate of ice thickness change. In addition to constraining the modelled ice geometry, this lets the relaxation largely remove model drift while explicitly incorporating present-day observed mass-change rates into the initial state of a forward simulation. We demonstrate the method by modelling the present-day state of C.H. Ostenfeld Glacier in North Greenland. We find that for a selected parameter setting, an unforced future simulation initialized without model drift and with present-day mass-change rates incorporated shows the glacier's simulated ice thickness change follows the present-day trend. Follow-up studies could extend this method to glaciers experiencing large present-day mass change or considered vulnerable to marine ice-sheet instability or elevation-feedback tipping, such as Petermann Glacier or Jakobshavn Isbræ, incorporating a more physically based basal melt product and a suitable calving law.



430 **Code and data availability**

ISSM is open source and freely available at <https://issm.jpl.nasa.gov/download/> (last access: 2 October 2026). Code for model simulations and data export will be uploaded to a Zenodo repository after revision and are freely available on request during revision from the main author.

435 **Author contributions**

TvdA designed and tested the nudging method outside of ISSM, MM incorporated the method into ISSM, NK provided guidance and feedback. TvdA wrote the manuscript with contributions from all authors.

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Competing interests

450 The contact author has declared that none of the authors has any competing interests.

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