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A rapidly converging spin-up method for the present-day Greenland ice sheet using the GRISLI ice-sheet model

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Abstract. Providing reliable projections of the ice-sheet contribution to future sea-level rise has become one of the main challenges of the ice-sheet modelling community. To increase confidence in future projections, a good knowledge of the present-day state of the ice flow dynamics, which is critically dependent on basal conditions, is strongly needed. The main difficulty is tied to the scarcity of observations at the ice-bed interface at the scale of the whole ice sheet, resulting in poorly constrained parameterisations in ice-sheet models. To circumvent this drawback, inverse modelling approaches can be developed and validated against available data to infer reliable initial conditions of the ice sheet. Here, we present a spin-up method for the Greenland ice sheet using the thermo-mechanical hybrid GRISLI ice-sheet model. Our approach is based on the adjustment of the basal drag coefficient that relates the sliding velocities at the ice-bed interface to basal shear stress in unfrozen bed areas. This method relies on an iterative process in which the basal drag is periodically adjusted in such as way that the simulated ice thickness matches the observed one. The process depends on three parameters controlling the duration and the number of iterations. The best spin-up parameters are chosen according to two criteria to minimize errors in sea-level projections: the final difference between the simulated and the observed Greenland ice volume as well as the final ice volume trend which must both be as low as possible. To increase confidence in the inferred parameters, we also make sure that the final ice thickness root mean square error from the observations is not greater than a few tens of meters. Our best results are obtained after only 420 years of simulation, highlighting a rapid convergence and demonstrating that our method can be used for computationally expensive ice-sheet models.

1 Introduction

Recent observations provide evidence that the rate of mass loss of the Greenland ice sheet (GrIS) is continuously increasing (Rignot et al., 2015; Mouginot et al., 2015). Simulating the GrIS response under future warm periods is therefore crucial to establish reliable projections of future sea-level rise at decade to century time scales (Bindschadler et al., 2013; Edwards et al., 2014), but also to investigate the effects of ice-sheet changes on the climate system Hansen et al. (2016); Defrance et al. (2017); Swingedouw et al. (2013); Böning et al. (2016). As a result, better constraining the GrIS evolution has become a key objective of the climate and ice-sheet modelling communities.

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Reliable simulations of the GrIS require a proper initialisation (i.e. spin-up) procedure to avoid an unrealistic evolution of the ice sheet caused by inconsistencies between the ice-sheet model initial conditions and the external forcing fields. For short-term projections (next decades to next centuries) starting from the present-day ice-sheet configuration, recent observations, such as surface and bedrock topographies (Bamber et al., 2013) and horizontal surface velocity (Joughin et al., 2010) offer only a partial description of the GrIS characteristics and the major source of uncertainty lies in the poor knowledge of the basal properties, such as the water content in the sediment and basal sliding, and of the vertical temperature profile. Indeed, the basal conditions have a strong impact on the ice motion (Boulton and Hindmarsh, 1987; Weertman, 1957; Kulessa et al., 2017). Optimizing the initial conditions of ice-sheet models is therefore an active area of research and a multidisciplinary effort. The initMIP project (Goelzer et al., 2017) gives a recent example of this effort. Its goal is to compare different initialisation techniques and to assess their impact on the dynamic responses of the models. Three main classes of initialisation techniques have been developed:

1.The free spin-up method allows the ice-sheet topography to evolve freely over a long enough time. This approach has long been the most commonly used technique to initialise ice-sheet models (Charbit et al., 2007; Huybrechts and de Wolde, 1999; Huybrechts et al., 2002, and other refecerence in (Rogozhina et al., 2011)). It consists in simulating the ice sheets during one or more glacial-interglacial cycles to account for the long-term ice-sheet history and thereby to obtain internal consistency between the simulated ice sheet and the climate forcing derived from ice core records. Since the ice-sheet topography evolves freely during the entire spin-up experiment, this may lead to a significant mismatch between modelled and observed present-day ice-sheet topography. Such spin-up methods can only be used with low computational cost models, which are often unable to properly capture fast ice flow processes.

2. The fixed topography spin-up method is similar to the free spin-up method except that during all the simulation the ice-sheet topography is kept constant and equal to its present-day observed value, while vertical temperature is allowed to freely evolve (e.g. Seddik et al., 2012; Sato and Greve, 2012). The disadvantage of this method is that an artificial drift may arise when free evolving topography is restored due to inconsistencies between internal and surface ice sheet fields.

3. The third kind of spin-up technique is based on an **inverse method** of the poorly known basal conditions in such a way that simulated surface velocities match the observed surface velocities (e.g. Gillet-Chaulet et al., 2012; Arthern and Gudmundsson, 2010; Morlighem et al., 2010; Gudmundsson and Raymond, 2008). However this approach may lead to internal inconsistencies between the simulated internal conditions (temperature and velocities) and the actual ones. The inconsistencies within the different observational datasets (surface and bedrock topography, velocities) can also have an impact on the results. An alternative approach, which avoids the previously mentioned shortcomings, consists in considering the observed ice-sheet geometry as the final target by finding appropriate basal conditions that minimise the differences between observed and simulated ice thickness (Pollard and DeConto, 2012).

Here, we present a spin-up approach that relies on the same basic principles as those developed by Pollard and DeConto (2012) (referred to as PDC12 in the following) for the Antarctic ice sheet. Similarly to PDC12, we compute the basal drag coefficient that minimises the error in the simulated ice thickness and relates basal stresses to basal velocities. However, while PDC12 requires long (multi-millennial) integrations for the method to converge, we suggest instead an iterative method of short (decadal to centennial) integrations starting from the observed ice thickness. Our method ensures a more rapid convergence

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and is thus suitable for computationally expensive models. The paper is organised as follows. In section 2 we present the main characteristics of the GRISLI ice-sheet model used in this study. Section 3 describes the spin-up method in detail. The main results are presented in section 4 and sensitivity experiments in section 5. These sections are followed by a discussion and the main conclusions of the present study.

2 The ice-sheet model GRISLI

The GRISLI ice-sheet model was first designed to describe the Antarctic ice sheet (Ritz et al., 2001) and further adapted to the northern hemisphere ice sheets (Peyaud et al., 2007). The version used in this study has been specifically developed for Greenland (Quiquet, 2012) with an horizontal resolution of 5 km x 5 km (301 x 561 grid points) and 21 vertical unevenly spaced levels, with the smallest grid spacing near the ice-bedrock interface to better resolve the basal motion. GRISLI is a hybrid model accounting for the coupled behaviour of temperature and velocity fields. It relies on basic principles of mass, heat and momentum conservation. The evolution of ice-sheet geometry is a function of surface mass balance, velocity fields and bedrock altitude. Since this study only deals with present-day steady-state simulations, the module describing the isostatic adjustment is not activated here. The evolution of the ice thickness is governed by the mass balance equation:

$$\frac{\partial H}{\partial t} = -\nabla(\overline{U}H) + M - b_{melt} \tag{1}$$

where H is the ice thickness, $\overline{U}$ is the depth-averaged velocity, M is the surface mass balance and b_{melt} is the basal melting.

The ice flow velocity is derived from a simplified formulation of the Stokes equations (i.e. the stress balance) using the shallow-ice (Hutter, 1983) and shallow-shelf (MacAyeal, 1989) approximations. The shallow-ice approximation (SIA) assumes that, owing to the small ratio of vertical to horizontal dimensions of the ice sheet, longitudinal stresses can be neglected with respect to vertical shearing along the steepest slope. Conversely, in the shallow-shelf approximation (SSA), the horizontal strain rates become dominant and the horizontal velocities do not vary with depth. In the model, the velocities are computed as the sum of the SSA and the SIA components, as in Bueler and Brown (2009), with the SSA velocity used as the sliding velocity. We assume no-slip conditions for a frozen bed (i.e. basal temperature below the melting point), and in these conditions, the SSA velocity is set to 0. In the model version used in this study, we assume a linear till, in which the basal shear stress (τ_b) and basal velocity (u_b) are related via the following expression:

$$\tau_b = -\beta u_b \tag{2}$$

where β is the basal drag coefficient and varies with space.

To describe the effect of ice rheology, the deformation rate and stresses are related via the Glen's flow law (Glen et al., 1957). As in other large scale ice-sheet models, GRISLI uses a flow enhancement factor in the Glen's flow law to artificially account for the impact of ice anisotropy on the deformation rate. This enhancement factor (Ef) typically ranges from 1 to 5. The grounding line position is defined according to a flotation criterion and floating points are treated following the SSA assumption only. Calving is not explicitly computed, but the ice-shelf front position is determined for a cut-off criterion of 250 m (Peyaud et al., 2007). The amount of ice obeying this criterion (ice thickness > 250 m) is computed as the calving flux.

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Since GRISLI is thermo-mechanically coupled, the ice temperature influences the ice velocity via the viscosity. The temperature is computed both in the ice and in the bedrock by solving a time-dependent heat equation. The temperature signal itself depends on ice deformation, surface conditions, and on basal temperatures, hence on the geothermal heat flux.

3 The spin-up method

The basic principle of inverse modelling approaches for ice-sheet spin-up is to adjust the basal sliding coefficient (β) which varies spatially, in order to reduce the mismatch between either the simulated surface ice velocities or the ice-sheet geometry and the observed ones.

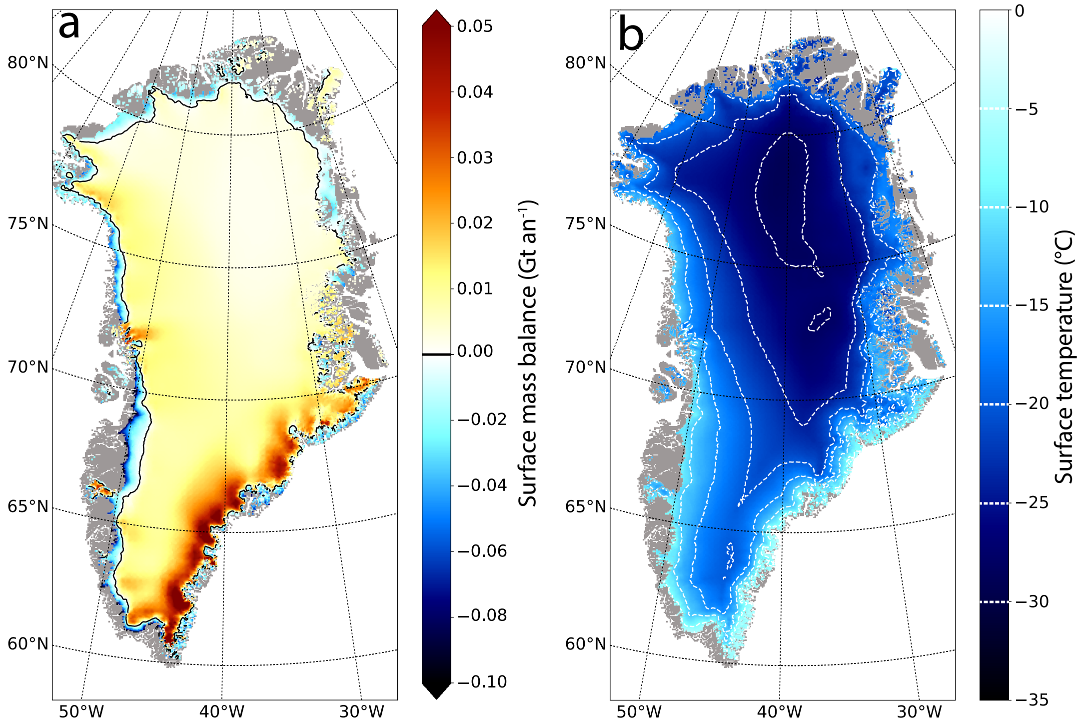

Figure 1. Climate forcing averaged over the 1979-2014 period simulated by the atmospheric regional model MAR (Fettweis et al., 2013) and interpolated on the GRISLI ice-sheet model grid (5 km x 5 km). a/ Mean surface mass balance (in Gt yr^{-1}). The black line represents the equilibrium line indicating the frontier between accumulation and ablation areas. b/ Mean annual surface temperature (in $^\circ$ C). The white dashed lines represent the 5° C isocontours.

Numerous studies are based on fitting the modelled ice velocities (e.g. Gillet-Chaulet et al., 2012; Arthern and Gudmundsson, 2010; Morlighem et al., 2010; Gudmundsson and Raymond, 2008), while Pollard and DeConto (2012) opted for fitting ice surface elevation. Here, we decided to adjust the basal sliding velocities via the adjustment of the β coefficient to fit the GrIS

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ice thickness to the observed one. Our choice is motivated by the need to refine the estimates of GrIS contribution to future sea-level rise.

The GRISLI climate forcing is provided by the surface mass balance and the surface air temperature simulated by the state of the art regional atmospheric model MAR (Fettweis et al., 2013) forced at its boundary by the ERA-Interim reanalyses (Berrisford et al., 2011). Both fields are averaged over the 1979-2014 period (Fig. 1). They are interpolated on the GRISLI grid (5 km x 5 km) and corrected for surface elevation differences between MAR and GRISLI by applying the method developed by Franco et al. (2012). We use the reconstruction from Maule et al. (2005) for the geothermal heat flux. Using these boundary conditions, GRISLI is run forward starting from the present-day observed ice thickness (Fig. 2a), from which the ice volume is inferred Bamber et al. (2013), and from the bedrock elevation. Initial vertical temperature and velocity profiles as well as the initial map of the basal sliding coefficient (Fig. 2a) are derived from previous GRISLI simulations carried out with boundary conditions close to those of the present study, and performed within the Ice2Sea project, which aimed at reducing the uncertainties on future sea-level rise projections Edwards et al. (2014). In order to avoid large inconsistencies between

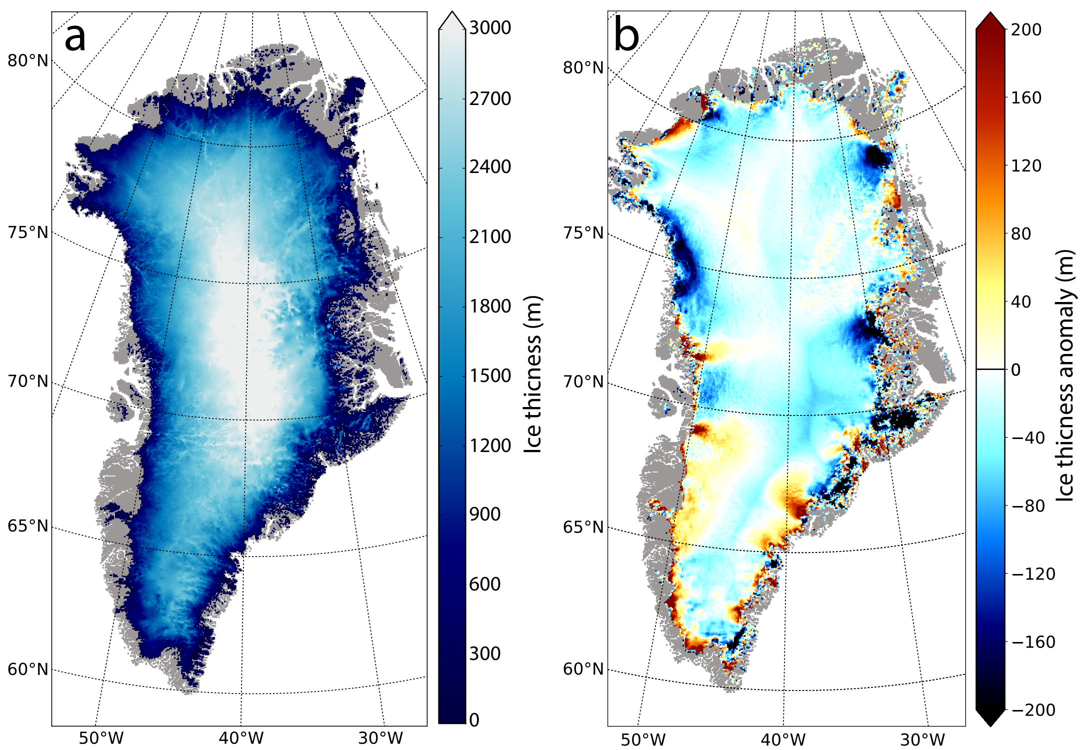

Figure 2. a/ Observed Greenland ice thickness (in m) from Bamber et al. (2013) interpolated on the GRISLI grid. Grey areas represent non ice-covered areas. b/ Difference between the simulated and the observed ice thickness (in m) obtained at the end of a 200-year-long simulation without spin-up procedure. The simulation has been carried out using the Ice2Sea initial conditions (see main text and Edwards et al. (2014)) and the climate forcing simulated by MAR.

the different datasets used as boundary and initial conditions, GRISLI is first run for 5 years. After this relaxation period, we start the spin-up procedure. This procedure is based on an iterative process set up to adjust the basal drag coefficient in such a way that the mismatch between observed and simulated ice thickness is reduced. At the end of the iterative process, we allow GRISLI to evolve freely for 200 years in order to assess the model performance in terms of trend and error in simulated ice volume compared to observations. The iterative process itself is divided in two main steps (Fig. 4).

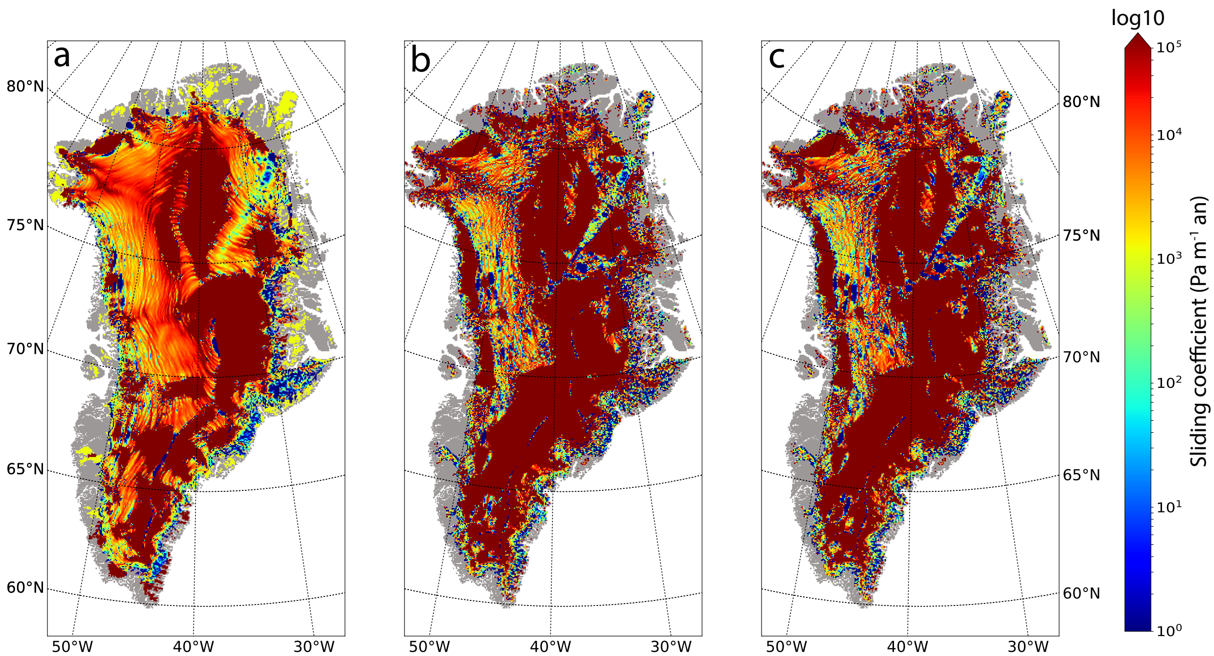

Figure 3. Spatial distribution of the basal drag coefficient (in log10 Pa m^{-1} an) a/ for the initial condition as used in the GRISLI ice2Sea simulations, b/ obtained for the best fit Nb$^{20}_{iter}$-Nb$^{50}_{year}$-Nb$^{6}_{cycle}$ and c/ obtained at the end of a spin-up procedure using the same spin-up parameters as those inferred from the best fit but starting from a uniform spatial distribution of the basal drag coefficient (β = 1).

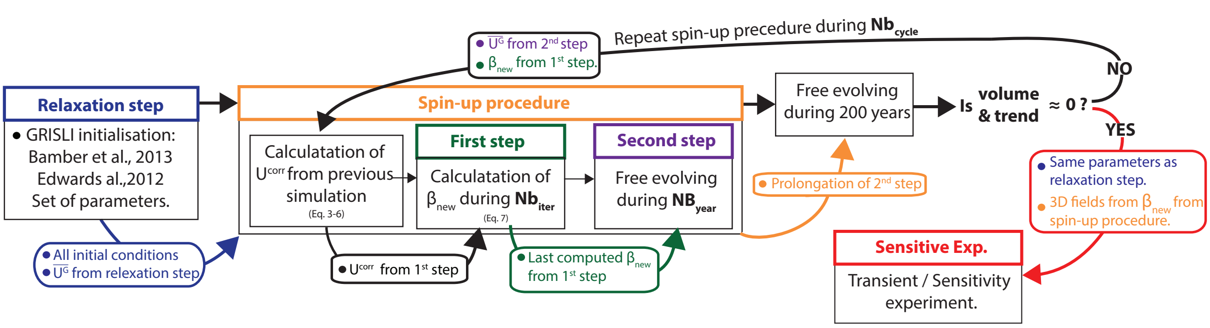

Figure 4. Schematic representation of the spin-up method.

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1st step: By using the vertically-averaged velocity $\overline{U}^G$ computed from the previous time step (or from the values obtained after the relaxation for the first iteration), we calculate a corrected vertically-averaged velocity field (U^corr) as a function of the computed (H^G) and observed ice thickness (H^obs) deduced from Bamber et al. (2013):

$$U^{Corr} = \frac{\overline{U^G} \times H^G}{H^{obs}} \tag{3}$$

U^{corr} can be seen as a the vertically-averaged velocity field corrected by a factor representing the difference between the observed and the simulated ice thicknesses. As seen before (section 2), the mean velocity field $\overline{U}^G$ in GRISLI is the sum of two velocity components: the sliding velocity U^{sli} and the velocity U^{def} due to vertical ice deformation:

$$\overline{U^G} = U^{sli} + U^{def} \tag{4}$$

Considering that the differences of velocity between $\overline{U}^G$ and U^{corr} are only due changes of the sliding velocity U^{sli}, we can also write:

$$U_{corr} = U^{sli}_{corr} + U^{sli} \tag{5}$$

Following Eqs. (4) and (5), we can deduce the corrected sliding velocity (U$^{corr}_{sli}$) needed to reduce the difference between H^G and H^{obs}:

$$U^{sli}_{corr} = U_{corr} - \overline{U^G} + U^{sli} \tag{6}$$

The new value of the basal drag coefficient allowing to reduce the gap between H^G and H^{obs} is deduced from the β_{old} value, inferred from the previous iteration and from the ratio between uncorrected and corrected sliding velocities:

$$\beta_{new} = \beta_{old} \times \frac{U^{sli}}{U^{sli}_{corr}} \tag{7}$$

with β_{new} calculated at each GRISLI grid point. H^G, U^G, U$^{sli}_{corr}$ andβ_{new} are updated during Nb$_{iter}$ time steps.

2nd step: With this new basal drag coefficient we let the model to freely evolve. After Nb$_{year}$ of the free-evolving simulation, we obtain a new GrIS topography and new corrected velocity fields computed from the mismatch between the simulated ice thickness after Nb$_{year}$ and the observations. With this, we can start a new cycle in which the 1st and 2nd steps are repeated. This new cycle uses the same set of spin-up parameters (Nb$_{iter}$ and Nb$_{year}$) and an initial guess of β coming from the previous iteration. All the iterations use the same initial conditions presented previously. The number of cycles carried out in this way is noted Nb$_{cycle}$. For all the experiments presented in the following, we performed a maximum of nine cycles.

To assess the spin-up performance and the quality of the inferred β coefficient, we perform, at the end of each cycle, a free-evolving simulation of 200 years with a β coefficient fixed to the values computed during the last cycle. The best Nb$_{cycle}$ for a given set of Nb$_{iter}$ and Nb$_{year}$ will be the one that provides a final volume as close as possible to the observation and a minimal trend over the last ten years of this free-evolving simulation. In addition to these two criteria, we also take into account the ice thickness root mean square error from the observations. Once the best fit is obtained, steady-state or transient GrIS simulations can be performed with reliable initial conditions, as done in Le clec'h et al. (2017) and Goelzer et al. (2017).

Four values of Nb_{iter} (20, 40, 80, 160 years) and of Nb_{year} (50, 100, 200 and 400 model years) have been tested with Nb_{cycle} ranging from 1 to 9, giving a total of 144 combinations of the spin-up parameters. The corresponding simulations are referred to as Nb^{X}_{iter}-Nb^{Y}_{year}-Nb^{Z}_{cycle} where X, Y and Z stand respectively for the Nb_{iter}, Nb_{year} and Nb_{cycle} values.

4 Results

4.1 Is the spin-up needed?

The annual mean climatological SMB for the 1979-2014 period integrated over the whole GrIS is 381 Gt yr^{-1} (Fig. 1a) with strongly positive values in southeastern Greenland (up to 0.04 Gt yr^{-1}), and largely negative ones over the ablation zone at the edges of the ice sheet, with values reaching -0.10 Gt yr^{-1} in the western area (Fig. 1a). The annual mean surface temperature is negative over the whole Greenland ice sheet, ranging from -29 ° C in the highest altitude regions to -0.5 ° C near the coast (Fig. 1b). To illustrate the need for a spin-up procedure, we performed a 200-year-long free-evolving simulation with this mean climatic forcing and with the initial conditions obtained after the 5-year-long relaxation (see Sect. 3), without any spin-up procedure. The simulated GrIS ice volume obtained in this experiment is smaller, by 20000 Gt, than the one estimated by Bamber et al. (2013) from observations (i.e. 2.71 10^{6} Gt) with negative ice thickness anomalies in the interior of the ice sheet, which are even stronger in the northwestern, northeastern and central eastern parts (Fig. 2b). Moreover, the ice volume decrease contributes by 0.6 mm yr^{-1} to the global sea level rise. Thus, despite an overall positive SMB, the model drift and the lack of spin-up procedure result in a decrease of the GrIS volume as large as the present melting due to the global warming Church et al. (2013). Therefore the use of a spin-up method to minimise the model drift is not avoidable if the goal is to produce reliable sea-level projections.

4.2 Spin-up performance

To assess the spin-up performance, we first examined the ice thickness root mean square error (RMSE), the ice volume anomaly (computed – observed) and ice volume trend for each combination of the spin-up parameters. An illustrative example is given here for Nb_{iter} = 20 and Nb_{year} ranging from 50 to 400 years, with Nb_{cycle} varying from 1 to 9 (Figs 5 and 6). For Nb_{cycle} = 1, all the tests corresponding to different Nb_{year} values start from the same initial conditions and the basal sliding velocity has not yet been updated with the new β coefficient.

4.2.1 Root mean square error

The RMSE behavior is approximately the same for all the Nb_{year} values (Fig. 5). A strong decrease between the first two cycles is obtained meaning that the departure between simulated and observed ice thickness is rapidly reduced. This decrease is then followed by a stabilisation occurring for Nb_{cycle} between 4 and 6 depending on the Nb_{year} value. Increasing Nb_{year} results in lower RMSE values. For example, for Nb_{cycle} = 6, the RMSE decreases from 84.8 m (Nb_{year} = 50) to 57.4 m (Nb_{year} = 400). This can be explained by the fact that for longer free-evolving simulations, the basal velocity (computed through the previously

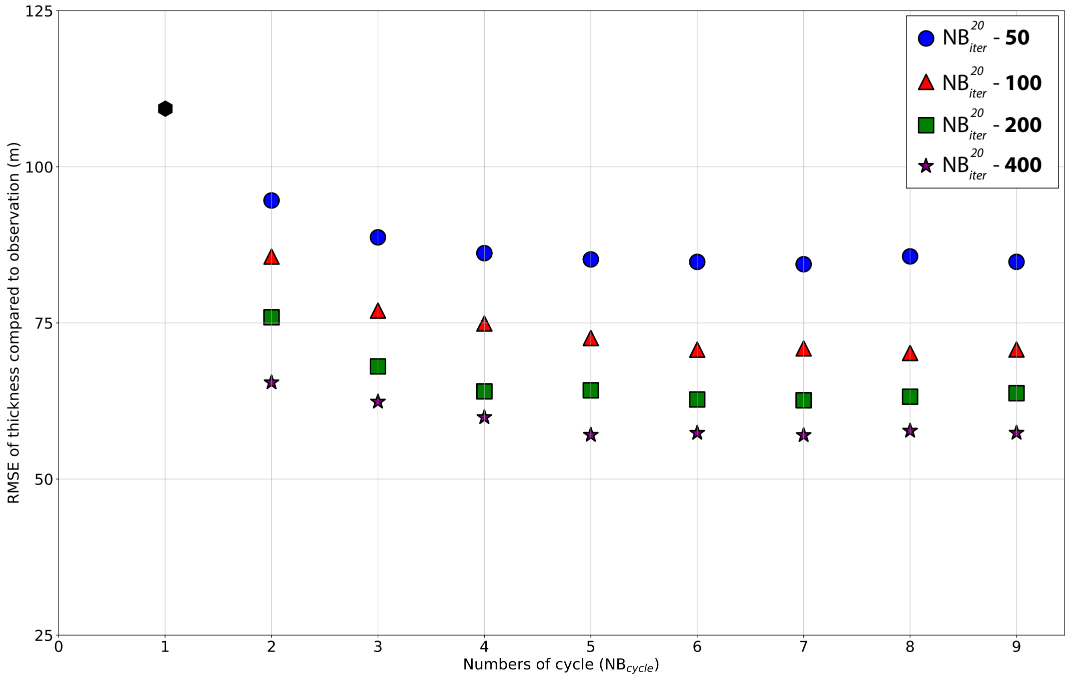

Figure 5. Ice thickness root mean square error w.r.t. observations from Bamber et al. (2013), in meters for Nb$_{iter}$ = 20 and the four Nb$_{year}$ values (50, 100, 200, 400) as a function of the number of iterations (Nb$_{cycle}$).

determined β coefficient) exerts a longer influence on the vertically averaged velocity, which in turn impacts the simulated ice thickness. This results in larger differences between simulated and observed ice thickness. This implies that the corrections of the β coefficient are more significant for the following cycle, and finally, that the method is more efficient to correct for the differences of the ice thickness with respect to the observed one.

4.2.2 Volume

The ice volume anomalies (ΔVol) obtained for the same set of spin-up parameters are displayed in Figure 6a. A strong ΔVol decrease is obtained, starting from a highly positive value (ΔV ∼28 000 Gt for Nb$_{cycle}$ = 1) and reaching negative values when Nb$_{cycle}$ increases. This illustrates that our method tends to underestimate the ice volume with respect to observations Bamber et al. (2013). This underestimation is also more pronounced for higher Nb$_{year}$ values. As a consequence, the combination of spin-up parameters providing the lowest RMSE values are also those for which the ice volume anomalies are the largest ones. For example, for Nb$_{cycle}$ = 6, ΔVol equals ∼10 000 Gt for Nb$_{year}$ = 400, while it is two order of magnitude below (ΔVol = 107 Gt) with Nb$_{year}$ = 50.

This behaviour can be explained when examining the ice thickness anomaly (ΔH, Fig. 7a). This anomaly tends to be negative in the central part and positive in coastal regions with the exception of the northwestern area which remains controlled by the

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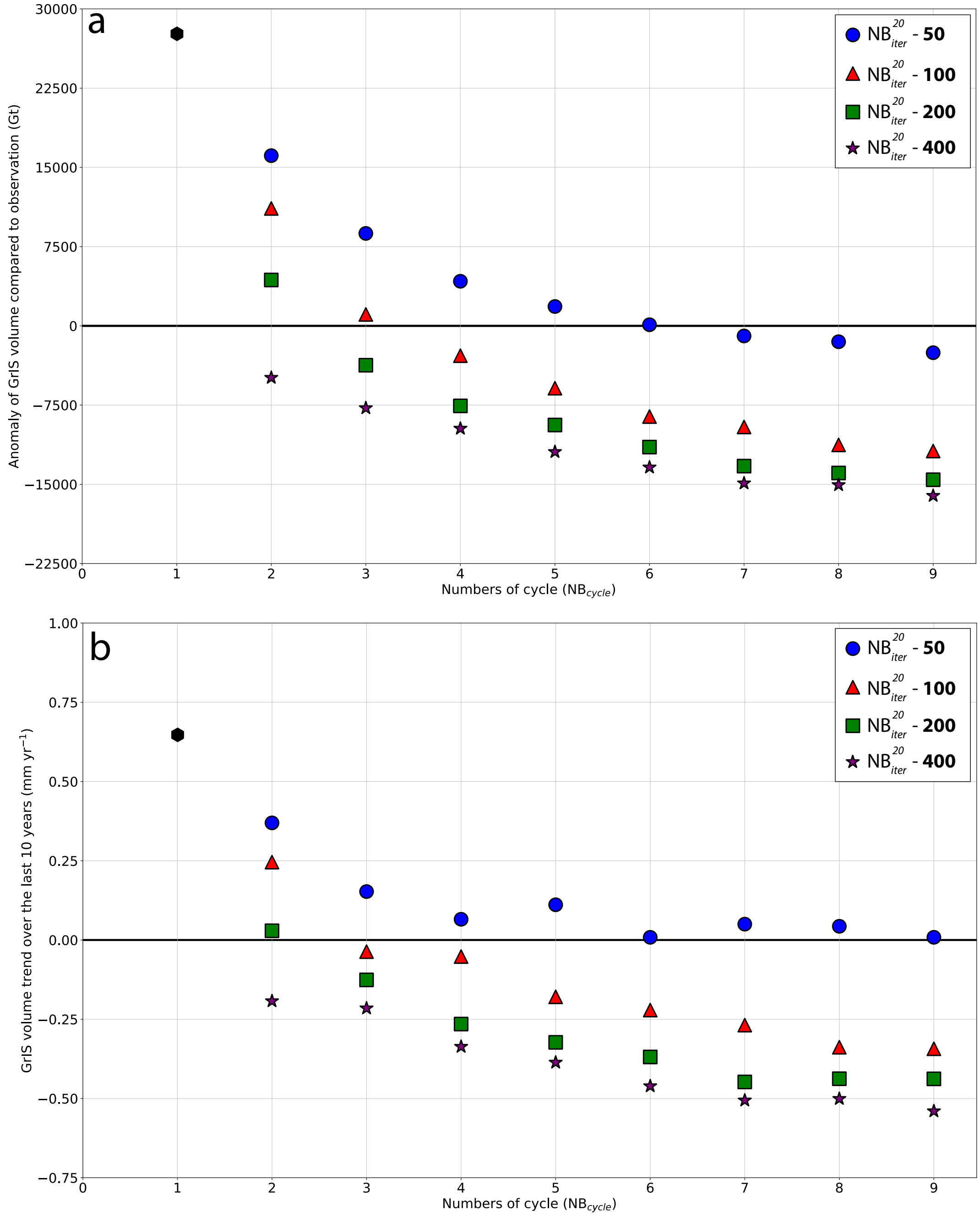

Figure 6. Same as figure 5 for the GrIS ice volume anomaly (a) and the ice volume trend (b) represented in Gt and mm yr^{-1} respectively.

SIA due to the high value of the basal sliding coefficient (Fig. 3b). Since our method is based on the adjustment of the basal

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sliding coefficient, it only operates over non-frozen bed where the SSA is activated. This occurs mainly in the peripheral regions of the ice sheet or in ice-stream areas. In the central part, where the basal temperature is most often below the melting point, ice velocities are mainly governed by the SIA and are thus not corrected.

Thus, during the first iterations, the ice velocities (and therefore the ice topography) have not been corrected so much and the regions where $\Delta H > 0$ are balanced by the central regions where $\Delta H < 0$, which are not impacted by the corrections. This compensating effect acts to reduce the ice volume anomaly. However, as Nb_{cycle} increases, corrections of ΔH become more efficient in the peripheral areas. In these regions, the simulated ice thickness is improved (Fig. 7b) with respect to observations and the RMSE is lowered, but the compensating effect is reduced and the ice volume anomaly increases.

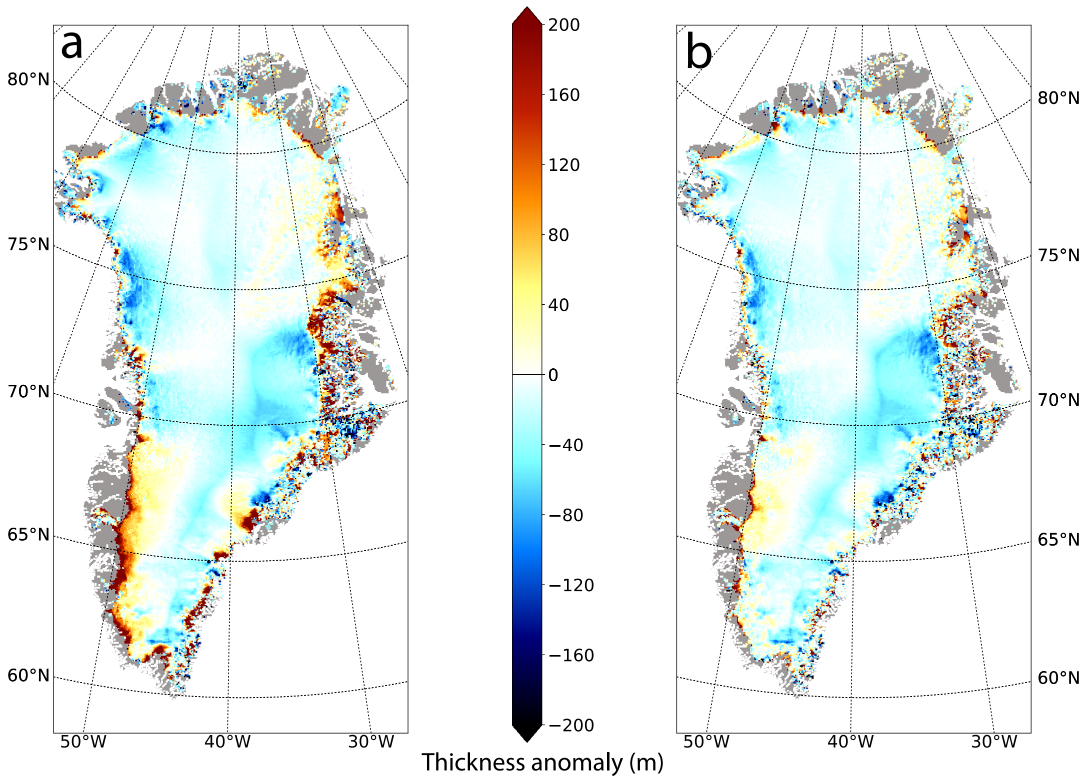

Figure 7. Difference between the simulated and the observed (Bamber et al., 2013) ice thickness (in meters) obtained for the spin-up parameters providing a/ the lowest ice volume anomaly and the lowest ice volume trend (Nb^{20}_{iter}-Nb^{50}_{year}-Nb^{6}_{cycle}) and b/ the lowest ice thickness RMSE (Nb^{20}_{iter}-Nb^{400}_{year}-Nb^{4}_{cycle}).

4.2.3 Ice volume trend

This analysis demonstrates that satisfying the ice volume anomaly criterion is in direct competition with the RMSE minimisation, and therefore that a compromise needs to be found. Since our main objective is to obtain a simulated ice volume as close as possible to the observed one (Bamber et al., 2013), the criterion related to the ice volume trend (IVT) must be also examined. Figure 6b shows that this trend follows a behavior similar to the ice volume anomaly with increasing values of Nb_{year} and Nb_{cycle}. The key feature appearing in this figure is the strong decrease towards negative values (down to -0.5 mm yr^{-1}) for most of the spin-up parameter combinations. However small IVT values are obtained for three set of parameters: Nb_{year} = 100 and Nb_{cycle} = 3 (IVT = -0.04 mm yr^{-1}), Nb_{year} = 200 and Nb_{cycle} = 2 (IVT = +0.03 mm yr^{-1}) and Nb_{year} = 50 and Nb_{cycle} = 6 (IVT = 8.5 10^{-3} mm yr^{-1}). While these two first combinations provide reasonable RMSE values (see Table 1), the ice volume anomalies are respectively 1079 and 4343 Gt. This must be compared to ΔVol = 107 Gt obtained with Nb_{year} = 50 and Nb_{cycle} = 6 which provide the smallest ΔVol and IVT values. Despite the corresponding RMSE being the highest one among all the tests which have been performed, it is less than 20 m higher than the lowest RMSE value. Thus the experiment Nb^{20}_{iter}-Nb^{50}_{year}-Nb^{6}_{cycle} matches our two main criteria (i.e. minimum ΔVol and IVT values) and the corresponding spin-up parameters appear as good candidates for the overall spin-up procedure.

Even if our approach is based on minimising the volume trend and fitting ice volume and ice thickness to the observed ones, the reliability of the method also depends of its capability to simulate ice velocities in good agreement with observations. We therefore compare our results to the surface ice velocity dataset provided by Joughin et al. (2010) (Fig. 8). The simulated results are slightly different from the observations especially in the central plateau where the region of low ice velocities is less extended in the simulations than in the observations. However, the overall patterns are in a good agreement, especially in regions of fast ice flow, providing confidence in our method.

4.2.4 Sensitivity to Nb_{iter} values

Results obtained for other Nb_{iter} values (40, 80 and 160) are reported in Table 1. None of these experiments fulfill both the ΔVol and IVT criteria. The Nb^{40}_{iter}-Nb^{100}_{year}-Nb^{3}_{cycle} provides the lowest ice volume trend (5.4 10-3 mm yr^{-1}) but the simulated ice volume anomaly is more than 20 times larger than for Nb^{20}_{iter}-Nb^{50}_{year}-Nb^{6}_{cycle}. Conversely, Nb^{160}_{iter}-Nb^{200}_{year}-Nb^{2}_{cycle} simulates a GrIS ice volume anomaly only 2.5 as large as for Nb^{20}_{iter}-Nb^{50}_{year}-Nb^{6}_{cycle} but the ice volume trend (0.041 mm yr^{-1}) is four times larger. Moreover, the duration of the spin-up procedure for Nb^{160}_{iter}-Nb^{200}_{year}-Nb^{2}_{cycle} is 1080 model years while it is only 420 years for Nb^{20}_{iter}-Nb^{50}_{year}-Nb^{6}_{cycle}. This confirms that Nb^{20}_{iter}-Nb^{50}_{year}-Nb^{6}_{cycle} appears as the best protocol in terms ΔVol and IVT criteria and that it is also designed to ensure a more rapid convergence.

5 Sensitivity to initial conditions and model parameters

5.1 Temperature equilibrium

In the work presented in section 4, the vertical temperature and ice velocities profiles taken as initial conditions came from previous experiments carried out with GRISLI in the framework of the Ice2Sea project (Edwards et al., 2014). These profiles have been chosen because they were assumed to reflect the present-day conditions. However, they are not necessarily in equilibrium with the climatic forcing taken from the MAR simulations (Fig. 1). Indeed, at the ice-sheet surface, the temperature obtained

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Table 1. Ice volume trend (in mm yr^{-1}) and ice volume anomalies (simulated – observed) obtained for the 16 combinations of the NB$_{iter}$ and NB$_{year}$ parameters. The values of NB$_{cycle}$ correspond to the number of iterations providing the lowest IVT and ΔVol values. Correpondiing ice thikness RMSE (w.r.t observations, Bamber et al. (2013) are also indicated.

NB$_{iter}$	NB$_{year}$	NB$_{cycle}$	**Final GrIS volume trend** (mm yr^{-1})	**Final volume difference from Obs.** (Gt)	**RMSE of thickness compared to Obs.** (m)
20	50	6	0,0085	107	84,8
	100	3	-0,0368	1079	77,0
	200	2	0,0291	4343	75,9
	400	3	-0,1927	-4908	65,5
40	50	6	0,0313	-405	77,3
	100	3	0,0054	2431	75,1
	200	2	0,0272	4430	75,2
	400	2	-0,1322	-4102	67,0
80	50	5	0,0110	554	75,3
	100	3	0,0272	2928	74,3
	200	2	0,0121	3370	74,7
	400	2	-0,0906	-2888	69,6
160	50	4	-0,0682	-1845	70,0
	100	3	-0,0163	-853	70,1
	200	2	-0,0405	-268	73,7
	400	2	-0,1225	-4503	71,5

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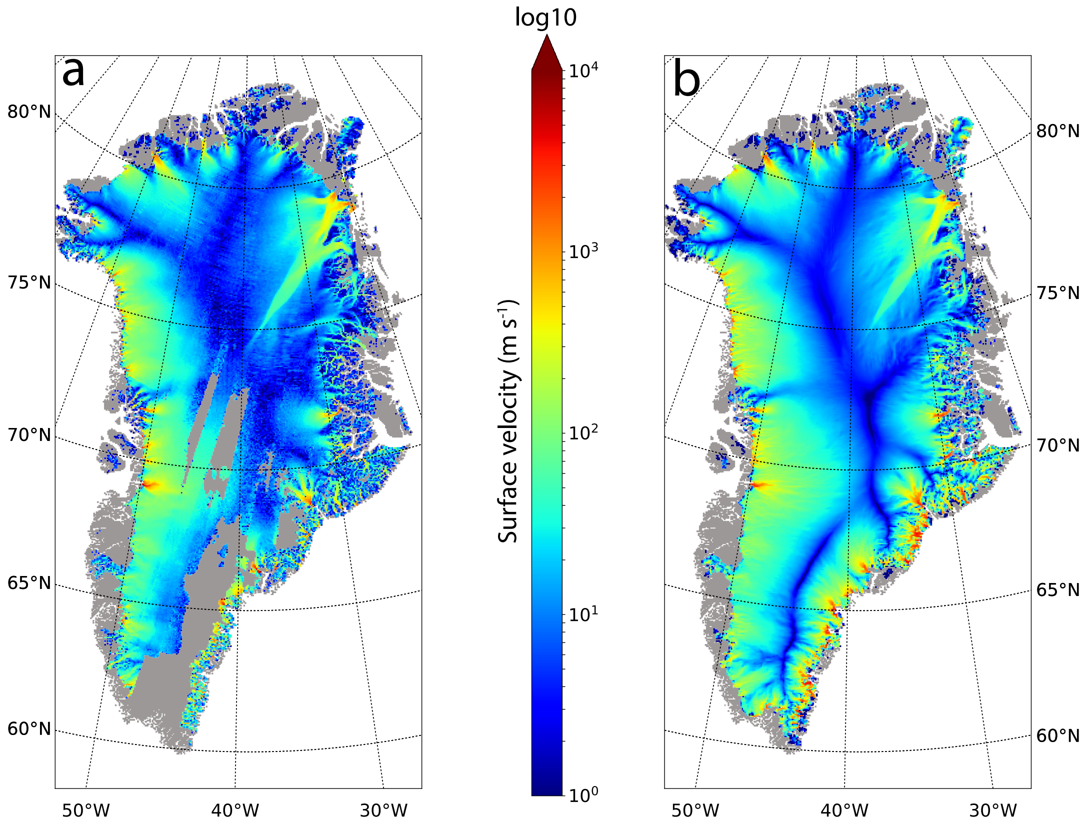

Figure 8. a/ Observed surface ice velocities coming from a compilation of interferometric synthetic aperture radar measurements obtained from RADARSAT data at different periods of the 2000s. b/ Simulated surface ice velocities obtained for our best fit (Nb$^{20}_{iter}$-Nb$^{50}_{year}$-Nb$^{6}_{cycle}$). Values are given in log10 m s^{-1}.

from MAR is about 5 ° C warmer than the Ice2Sea one (Fig. 9a). Therefore, we performed a 30 000-year-long simulation to make the vertical temperature and velocity profiles consistent with the surface climate forcing. This new experiment has been carried out with the ice topography fixed to the observed one (Bamber et al., 2013) and the basal sliding coefficient deduced from the spin-up procedure (i.e. Nb$^{20}_{iter}$-Nb$^{50}_{year}$-Nb$^{6}_{cycle}$). As illustrated in figure. 9a for a region located in the central part of the ice sheet (73-74.5 ° N, 40-43 ° W)), the ice sheet becomes progressively warmer as the result of inconsistencies between the initial vertical temperature profile and the surface climate.

New spin-up procedures have been undertaken with this temperature equilibrium and with Nb$_{iter}$ = 20 and 40 and the same Nb$_{year}$ (50, 100, 200, 400) and Nb$_{cycle}$ (1 to 9) as previously. These new spin-up tests reveal that the ice volume anomalies, the ice volume trends and the RMSE values are degraded compared to the results presented in section 4. For example, the lowest ΔVol and IVT values are obtained for Nb$^{40}_{iter}$-Nb$^{200}_{year}$-Nb$^{2}_{cycle}$ with ΔVol = -867 Gt and IVT = -0.90 mm yr^{-1} but the ice

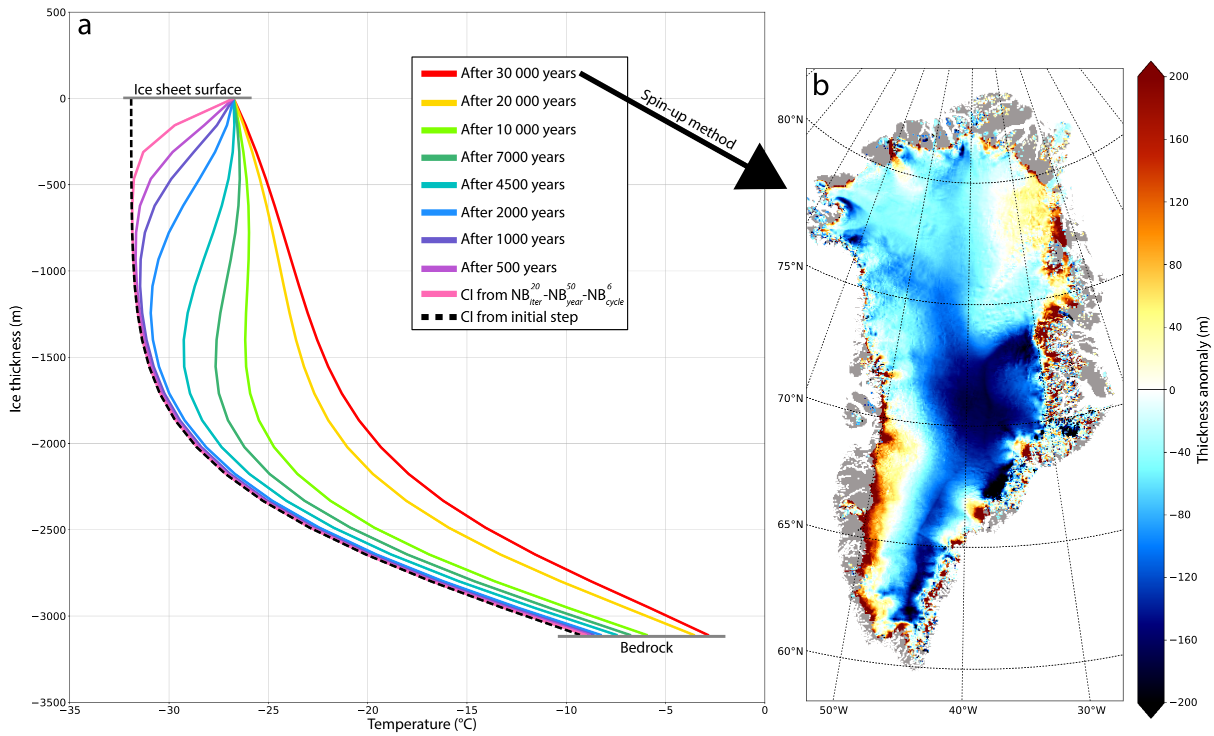

Figure 9. a/ Vertical profile of temperature (in ° C) in a central region of Greenland (73-74.5 ° N, 40-43 ° W) taken as initial condition from the Ice2Sea project (Edwards et al., 2014, see black dashed line) at the end of the spin-up procedure (Nb$^{20}_{iter}$-Nb$^{50}_{year}$-Nb$^{6}_{cycle}$) and at different periods of the temperature equilibrium experiment. b/ Difference between simulated and observed Ice thickness (in m) obtained after the temperature equilibrium and a new spin-up procedure performed with the spin-up parameters providing the best fit fit in terms of ice volume and ice volume trend.

thickness RMSE is 214.0 m. Conversely a lower RMSE value is reached for Nb$^{20}_{iter}$-Nb$^{400}_{year}$-Nb$^{8}_{cycle}$, but for these parameters ΔVol = -56 601 Gt.

These results illustrate the limitations of our spin-up method, as also shown in Figure 9b which displays the ice thickness anomaly for Nb$^{20}_{iter}$-Nb$^{50}_{year}$-Nb$^{6}_{cycle}$ (ΔVol = -56102 Gt, IVT = -1.19 mm yr^{-1}, RMSE = 118.3 m) after a 200-year free-evolving simulation. Actually, the warmer temperatures obtained after the 30 000 year-long simulation induce higher ice velocities due to the thermo-mechanical coupling. In the ice-sheet interior (SIA areas), these velocities are not corrected by our spin-up approach as shown by the β coefficient which reaches maximum values in these regions (Fig. 3b). Thus, an increased ice flux takes place from the central part to the peripheral regions leading to amplified negative ice thickness anomalies (Fig. 9b).

5.2 Sensitivity to the enhancement factor

In the ice-sheet interior, the ice flow is mainly due to internal ice deformation which is controlled by the temperature and thus by the viscosity. A possibility to reduce errors in ice surface elevation in these locations is to adjust the enhancement factor of the Glen's flow law, which relates viscosity to deformation rates. Lowering the Ef value allows to decrease the deformation and thus to slow down the ice flow velocities. Therefore we performed new sensitivity tests with Ef = 1 (instead of Ef = 3, as in the experiments presented in section 4) with the same spin-up parameters used in the previous section.

As expected, for given Nb_{iter} and Nb_{year} values, the ice thickness RMSE is improved when the number of iterations (Nb_{cycle}) increases. Contrary to previous tests (section 4), the parameters (Nb_{iter} =20, Nb_{year} = 400, Nb_{cycle} =8) providing the lowest RMSE value (55.9 m) are also those providing the lowest ice volume anomaly (ΔVol = 5694 Gt) and ice volume trend (IVT = 0.03 mm yr^{-1}). While the RMSE value is lower than the one obtained with Ef = 3 (Nb^{20}_{iter}-Nb^{50}_{year}-Nb^{6}_{cycle}), the ΔVol and IVT values are about 50 and 4 times higher. Indeed, a lower enhancement factor reduces the mismatch between observed and simulated ice thickness in central areas and hence the compensating effects between ΔH < 0 and ΔH > 0 regions (Fig. 10). Increasing Nb_{iter} does not improve the ΔVol and IVT results.

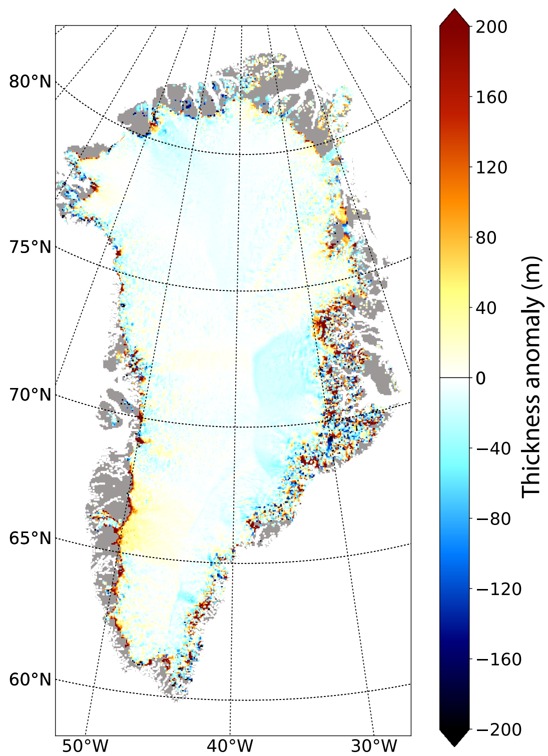

Figure 10. Difference between simulated and observed Ice thickness (in m) obtained for the spin-up parameters (Nb^{20}_{iter}-Nb^{400}_{year}-Nb^{8}_{cycle}) providing the best fit when Ef = 1.

5.3 Sensitivity to the basal drag coefficient

To evaluate the sensitivity of our spin-up approach to the initial distribution of the basal drag coefficient, we performed a new series of experiments starting from a uniform β equal to 1 instead of the one from Edwards et al. (2014). In terms of ice volume anomaly and ice volume trend, the parameters providing the best fit are exactly the same as for the experiments presented in the section 4 (i.e. Nb^{20}_{iter}-Nb^{50}_{year}-Nb^{6}_{cycle}), with ΔVol = -583 Gt and IVT = 0.018 mm yr^{-1}. Although these values are not as low as those obtained in our reference experiment (Nb^{20}_{iter}-Nb^{50}_{year}-Nb^{6}_{cycle}), they are still satisfactory, as the final ΔVol value is only 0.02 % that of the present-day ice volume. After Nb_{cycle} = 6, the new spatial distribution of the basal drag coefficient is very similar to that obtained in Nb^{20}_{iter}-Nb^{50}_{year}-Nb^{6}_{cycle} (Fig. 3c). This illustrates the robustness of the method and shows that it does not depend on the chosen initial distribution of the basal drag coefficient.

6 Summary and discussion

In order to improve the reliability of Greenland ice-sheet simulations in a future transient climate, an accurate evaluation of the present-day trend of ice flow dynamics is required. One the major difficulties in addressing this need lies in the poorly constrained observational data of the basal conditions that strongly control the ice motion in the entire ice sheet. Here, we present an inverse method to infer the spatial distribution of the basal drag coefficient in such a way that the mismatch between simulated and observed GrIS ice thickness is minimized. The best fit is defined for the sets of parameters providing minimum values of ice volume trend and difference between simulated and observed ice volume. This choice was motivated by the need to refine the projections of GrIS contribution to global sea-level rise. The great advantage of the method is its rapid convergence (a few hundred years) making it suitable for more computationally expensive models. Moreover, we have also shown that it only poorly depends on the initial guess of the spatial distribution of the basal drag coefficient.

However, choosing the ice volume anomaly as the main criterion to assess the performance of the spin-up method may lead to misinterpretations of the quality of the fitting procedure. As illustrated in Section 5, compensating effects may arise between regions of positive and negative ice thickness anomalies (w.r.t. observations). It is thus highly recommended to choose the best compromise between the minimisation of errors in ice volume on one hand and a low ice thickness root mean square error on the other hand. In this study we focused only on the results leading to RMSE values not greater than a few tens of meters. This remains in the range of PDC12 results who used the minimisation of ice thickness errors as the main target criterion. Because the basal sliding velocities are not computed in frozen bed areas in our hybrid model, reducing further the RMSE through inverse techniques of basal conditions, and thereby the compensating effects, is not an easy task. However, an appropriate tuning of the enhancement factor (Sect. 5.2) allows the adjustment of ice flow velocities in regions only governed by the shallow-ice approximation and may improve the final GrIS topography.

Another limitation of the method may come from the model resolution. The succession of higher/lower ice thickness due to the succession of valleys/ridges in mountain areas may be poorly resolved. Owing to the insulation effect of the ice, this may lead to an erroneous representation of the basal temperature patterns, and SSA regions may be erroneously interpreted as frozen bed regions and vice versa (Pattyn, 2010). Higher resolution models can also better account for the dynamics of small-scale

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outlet glaciers and for their interactions with floating ice that strongly influence the ice-sheet mass balance (e.g. Aschwanden et al., 2016). While this effect is less crucial for Greenland than for Antarctica, recent observations have highlighted increasing thinning rates in most coastal regions (Thomas et al., 2009; Rignot et al., 2015) causing grounding line retreat and significant destabilization of grounded glaciers.

The reliability of the method also depends on the quality of observations data. Errors in observed surface or bedrock topography would give rise to errors in the present-day estimated ice thickness and thus to an erroneous choice of the best spin-up parameters. In the same way, large uncertainties remain in the reconstruction of the geothermal heat flux that strongly impacts the basal temperature. Finally, we would like to stress that in our simulations, the spatial distribution of the basal drag coefficient does not change through time. However, changes in basal hydrologic conditions along with changes in ice surface elevation and ice extent are likely to occur in a changing climate. While a constant spatial distribution of the β coefficient may seem reasonable for short-term projections, it is more questionable at the century time scale, and future modelling efforts should therefore be undertaken to compute interactively the basal drag coefficient as a function of changes in basal conditions.

7 Code and data availability

The developments of the GRISLI source code are hosted at https://forge.ipsl.jussieu.fr/mailman/listinfo.cgi/grisli. Access to those who conduct research in collaboration with the GRISLI users group can be granted upon request to C. Ritz (catherine.ritz@univ-grenoble-alpes.fr). The model outputs from the simulations described in this paper are freely available from the authors upon request.

Author contributions. The implementation of the iterative process in the GRSILI model was initially done by C. Ritz and further optimised by S. Le clec'h, A. Quiquet and C. Dumas. Analyses of the experiments were performed by S. Le clec'h and discussed with the co-authors. The paper was written by S. Charbit and S. Le clec'h with contributions from A. Quiquet and M. Kageyama.

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