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1 Introduction

The Arctic is the region of the Earth experiencing the largest increase in temperature since the pre-industrial era (Serreze and Barry, 2011), with consequences already perceptible on the mass evolution of the polar ice caps and the Greenland ice sheet (Rignot et al., 2011). The atmospheric conditions control the variability of the near-surface temperature (ST) and the surface mass balance (SMB) of the GrIS. The SMB represents the difference between snow accumulation, which is further transformed into ice, and the ablation, which are processes of ice loss. Accumulation and ablation are both sensitive to ST. Variations of ST and SMB directly affect the GrIS total ice mass by impacting its characteristics such as thickness, ice volume and ice extent. In turn, variations of the GrIS characteristics affect the ST and the zones of ablation (SMB < 0) and accumulation (SMB > 0). GrIS changes can also disrupt the atmospheric circulation over Greenland caused by changes in topography thermal contrast between ice sheet surface and atmosphere layers, surface albedo and ice sheet area, as shown by Vizcaino et al. (2015), Vizcaíno et al. (2008) and Lunt et al. (2004). Quantifying the balance between these different processes and feedbacks is required to understand and predict more confidently the evolution of the GrIS under current and future global warming.

Although numerous studies highlighted the importance of correctly representing the interactions between the GrIS topography changes and the atmosphere (Vizcaino et al., 2015; Edwards et al., 2014a; Alley and Joughin, 2012; Huybrechts et al., 2002), only few global or regional models have taken the GrIS topography changes into account to compute the future evolution of the SMB, ST and energy budget over the GrIS. The climate models usually represent the ice sheet component with a fixed and constant topography, even under a warm transient climate forcing. Recently, Vizcaino et al. (2015) used an atmosphere-ocean general circulation model (AOGCM) coupled with an ice sheet model (ISM) to explore topography feedbacks on ice mass loss. They found an ice mass loss amplification of 8–11 % (by 2100) and of 24–31 % (by 2300). Since both their ice sheet and climate models have a relatively coarse resolution (3.75° for the atmospheric component and 10 km for the ice sheet model), they focus on the added value of incorporating the coupled processes and less on exactly reproducing the observed GrIS, which would require more detailed physics in both models. They explain that their study must be regarded as a necessary first step towards more advanced coupling of ice sheet and climate models at higher resolution. Indeed, a higher resolution is necessary to represent correctly the steep slopes at the ice sheet margins (typically, the altitude varies by 2000 m over distances of the order 100 km). However, the computation of atmospheric fields at a resolution similar to the ISM one requires large computing resources. Franco et al. (2012) developped an interpolation method allowing to correct each SMB component (snowfall, rainfall, run-off, sublimation and evaporation) as a function of topography changes. They showed that their corrective method is able to significantly reduce the model output differences between a coarse spatial resolution model and a high resolution model. Edwards et al. (2014a) developped an alternative parametrisation of the interactions between the GrIS and the atmosphere to correct the SMB computed by a regional atmospheric model (RCM) by taking into account the GrIS topography changes computed by an ISM. This method only requires limited additional supercomputing resources. Under the SRES A1B emissions scenario, Edwards et al. (2014b) showed that a larger melting of the GrIS is obtained when the elevation feedback is taken into account (ranging from 4.4 % in 2100 to 9.6 % in 2200). However, in both parametrisations by Franco et al. (2012) and Edwards et al. (2014b), the authors only consider a strict linear relationship between topography

The Cryosphere Discuss., https://doi.org/10.5194/tc-2017-230
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Discussion started: 6 November 2017

The Cryosphere Discussions

the framework of an intercomparison project of initialisation methods (initMIP project). It turned out that GRISLI is one of the models that compares the best with observations (Goelzer at al., 2017).

3 Coupling methods and experiments

In this study, we compare three methods of different complexities to account for the atmosphere – GrIS interactions. We investigate how these methods affect the computed SMB, ST, SLR contribution and surface elevation changes. For the three methods described below, MAR is forced every 6 hours at its lateral boundaries by the MIROC5 fields and run at a 25 km resolution from 1976 to 2100. The forcing fields from MIROC5 evolve in response to the historical (until 2005) and the RCP8.5 scenarios (from 2006 to 2100).

Before forcing GRISLI with the 25 km resolution SMB and ST outputs from the MAR model, we need to spatially interpolate them onto the 5 km GRISLI grid. However, as SMB and ST are very sensitive to the Greenland topography we also need to correct them by the topography changes due to the difference of resolution between MAR (25 km) and GRISLI (5 km). Using a linear interpolation method if the steep topography at the GrIS margin and the complex orographic features in these areas are not taken into account could lead to important biases in SMB and ST in these regions (Franco et al., 2012).

In the present study, for all the experiments, we first interpolate the MAR outputs on the 5 km GRISLI grid using a simple bilinear interpolation. Then, the fields are corrected for the altitude difference induced by the difference of resolution between MAR and GRISLI. These topography corrections are based on the method developped by Franco et al. (2012) who derive a local vertical gradient of SMB (or ST) as a function of altitude for each GRISLI grid point. This gradient is then used to compute the correction due to the difference of altitude between the Bamber et al. (2013) topographies seen by MAR (resolution 25 km) and by GRISLI (resolution 5 km). This procedure can be followed at a daily time scale (Noël et al., 2016). For our purpose, we choose to average the daily vertical gradients at the annual time scale and to apply the altitude correction every year.

For all the experiments described below, the coupling between MAR and GRISLI starts in 2020 when the SMB anomalies are large enough to induce significant topography changes in GRISLI.

We investigate three levels of coupling between the ice sheet and the atmospheric models:

- The "No Coupling" method (hereafter referred to as NC). The present-day GrIS geometry (topography and ice extent) provided by Bamber et al. (2013) is prescribed to MAR as a boundary condition during the entire simulation duration. GRISLI is forced until 2150 (see Sect. 2.1.2) using the downscaled MAR outputs described above. The NC method does not allow to account for the ice sheet feedback on the climate. Rather, it provides the response of GrIS under a specific climate forcing.

We call this one-ways coupling

- The "One-way coupling" method (hereafter called 1-W). This method goes a step further. It is based on the same principle as the NC method, but the SMB and ST fields are corrected based on an updated altitude ($H(t)$) given by (Eq. 2):

I would call this "corrected 1w"

$$H(t) = H_{Bamber} + \Delta H_{GRISLI}(t) \quad (2)$$

The Cryosphere Discuss., https://doi.org/10.5194/tc-2017-230
Manuscript under review for journal The Cryosphere
Discussion started: 6 November 2017

The Cryosphere Discussions

where H_{Bamber} is the Bamber et al. (2013) topography at 5 km and ΔH_{GRISLI} is the topography anomaly simulated by GRISLI between the initial topography computed for year 2000 from the equilibrium state (t=0) and the ongoing time step (t). In doing so, this method artificially accounts for the elevation feedback because the SMB and ST are initially computed by MAR on a fixed ice sheet topography. With this method GRISLI is forced off-line by the MAR atmospheric conditions already computed in the NC run, therefore allowing sensitivity experiments in GRISLI with limited additional computer time. However, the changes in GRISLI topography are not taken into account by MAR.

- The Fully Coupled method (hereafter 2-W). This coupling method is the most accurate way to represent the interactions between the GrIS and the atmosphere but it is also more computationally expensive. At the end of a MAR simulated year, MAR is paused and GRISLI is forced by the 5 km interpolated SMB and ST just computed by MAR. GRISLI then computes a new GrIS topography and extent which are aggregated on to the 25 km MAR grid for the simulation of the next year of the MAR experiment. GRISLI and MAR are never stopped, just alternatively paused and resumed until 2150.

The differences between the GRISLI equilibrium state after the initialisation step and the observed topography (Bamber et al., 2013) (cf Sect. 2.2.2) could lead to inconsistencies between the results obtained by MAR under its usual setup, i.e. calibrated with the Bamber et al. (2013) topography, and the results that would be obtained by using directly the GRISLI topography. For this reason, in both the 2-W and the 1-W experiments, we use anomalies of GrIS topography applied on Bamber et al. (2013) topography rather than the absolute topography from GRISLI.

4 Results

4.1 The uncoupled simulation: the NC experiment

4.1.1 MAR

The mean ST over the first two decades (2000-2020) and averaged over the whole GrIS is equal to -18.7°C and the mean SMB to 434 Gt yr^{-1} (Fig. 1). After 2020, the ST increases by 0.065°C yr^{-1} until 2100 (Fig. 1). Over the same period (2020-2100), the averaged GrIS SMB decrease by 12.3 Gt yr^{-1} (Fig. 1). The surface temperature anomaly in 2100 compared to 2000 shows a warming ranging from +1.5 °C in the southern part of the GrIS to more than +8 ° C in the northern part (Fig. 2A). This warming in northern Greenland is a direct response to the MIROC5 forcing fields due to the polar temperature amplification. However, regional heterogeneities are observed in the annual mean GrIS SMB spatial distribution (Fig. 2B). Indeed, between 2000 and 2100 there is a positive SMB anomaly (i.e more ice accumulation) in a zone located along a South-North transect in the central part of the GrIS. This ice accumulation is mainly governed by the larger snowfall on the GrIS central part in winter and spring seasons (not shown). An opposite trend 10 times larger than ice accumulation, is simulated over the edges of the ice sheet, with a negative SMB (i.e ice ablation). In these regions, the summer season governs this negative SMB and is characterised by larger rainfall and melting ice (meltwater and runoff increase) than for other seasons (not shown). As a

The Cryosphere Discuss., https://doi.org/10.5194/tc-2017-230
Manuscript under review for journal The Cryosphere
Discussion started: 6 November 2017

The Cryosphere Discussions

consequence, the limit between the accumulation (SMB > 0) and ablation (SMB < 0) areas, also called the equilibrium line altitude (ELA, a line where SMB = 0), shifts inland through time (Fig. 3). This shift explains the mean decrease of SMB over the whole GrIS until 2100 seen in Fig. 1.

4.1.2 GRISLI

In 2100, a decrease of 15.8 m in mean ice sheet thickness is simulated (Fig. 2C) with a standard deviation of 32.7 m. In 2150, the mean ice thickness decrease reaches 38.6 ± 68.4 m. We can distinguish two types of regions: there is a thinning over the GrIS coastal regions and a thickening over the central GrIS regions (Fig. 2C). The thickness changes for the 2000-2150 period show the same patterns as the 2000-2100 period, but with a larger magnitude. Over the thinning regions, between 2000-2100 (resp. 2000-2150) the ice thickness decreases by 40 ± 32.5 m (resp. 78.5 ± 70 m). On the contrary, in central regions, the ice thickness increases with a median value of 4.4 ± 12 m (resp. 7.8 ± 13.4 m). The thickness anomaly is due to the complex combination of changes both in surface atmospheric conditions and ice dynamics conditions.

The ice dynamics is impacted by the warming climate. Generally, the simulated ice velocities increase from the central part of the ice sheet to the coastal regions (Fig. 4A). However, at the ice sheet margins, the ice velocities strongly decrease. In GRISLI, the fine scale structure of the Jakobshavn (western coast) and the Kangerlussuaq (eastern coast) glaciers is relatively well reproduced (Fig. 5A-B). For these glaciers and their associated ice-streams, within 100 years, the surface velocities slow down by more than 200 m yr^{-1} in the coastal regions, while they increase by more than 60 m yr^{-1} in the interior (Fig. 5C-D). Because the region feeding the ice stream has a frozen base, the SIA assumption is the predominant simplification used by GRISLI to compute the ice velocities. In this area, the ice velocity increase is due to the SIA velocity component (Fig. 5E-F). As mentioned previously, by 2100, the thickness decreases at the margins and increases in the interior (Fig. 5K-L), resulting in steeper slopes and thus in larger SIA velocities. However, the coastal regions have a temperate base and the SSA component of the velocity is predominant. Thus, the ice flow velocity is mainly controlled by the ice flux coming from the inland part. As the ice flux depends on the ice thickness which decreases over the coastal areas, the SSA velocity component decreases (Fig. 5G-H). This increase-decrease ice velocity pattern has been also reported by Peano et al. (2017), using GRISLI forced by CMIP5 models under the RCP8.5 scenario.

As a result of ice accumulation-ablation changes and ice velocity changes, the ice mask (numbers of grid points covered by permanent ice) decreases by 3.7 % in 2100 compared to the initial one (in 2000-2005). During the first 20 years (2000-2020), the total GrIS volume remains stable with no additional contribution to SLR compared to year 2000. After 2020, the GrIS volume decreases, resulting in a global contribution of +7.6 cm in 2100.

Extending the GRISLI NC experiment until 2150, forced by the same 2095 MAR forcing climate results in an amplification of all the changes observed in 2100 and discussed above: the extent of the ablation zone, the larger thinning and the slow down ice velocities in the coastal regions. As a consequence, the GrIS contribution to global SLR is amplified, reaching +18.7 cm in 2150.

The Cryosphere Discuss., https://doi.org/10.5194/tc-2017-230
Manuscript under review for journal The Cryosphere
Discussion started: 6 November 2017

The Cryosphere Discussions

4.2 Differences between the 2-W and the NC experiments.

4.2.1 Impact on SMB and ST

This is interesting :)

The near-surface temperature (ST) simulated for 2150 in the 2-W experiment over the whole GrIS is warmer than in the NC experiment, except in the region at the edge of the GrIS, which is strongly colder (until -10°C, Fig. 6A). The ST of this region is sensitive to the atmospheric circulation. At the edges of the ice sheet, there is an intensification of the strong and cold katabatic winds coming from the central part of the GrIS in 2-W compared to NC. The katabatic winds have a daily time scale resolution and are represented by the MAR model (Gallée and Pettré, 1998; Gallée et al., 1996). These stronger winds are due to the higher coastal surface slope simulated in 2-W than in NC (see Sect. 4.2.2). As a consequence, they prevent warmer and wetter air from the ice-free areas (i.e. covered by tundra) and ocean from reaching the GrIS margin regions (van den Broeke and Gallée, 1996). Thus, the ST decrease over the edges of the GrIS (Fig. 6A). The second consequence of the strengthening of the katabatic winds is to enhance the atmospheric exchange at the middle of the slope over the GrIS. Indeed, at the surface, the lower atmospheric pressure generated by the stronger katabatic winds is filled in by the warmer air coming from higher atmospheric levels in the boundary layer. Thus the warming of the upper part of the boundary layer combined with the lower surface elevation, explains the ST increase on the coastal regions inland from the very edge of the ice sheet. These two types of colder and warmer regions simulated in 2-W with respect to NC are already present after 100 years of experiment (Fig. S1A-B).

How can you test this hypothesis ?

In 2150, the SMB difference between the 2-W and NC experiments exhibits two distinct patterns. With the 2-W approach, the SMB increases by 0.6 m yr^{-1} over the eastern coast, the south central part and in some local regions in the northern part of the GrIS (Fig. 6B). These regions are characterized by a larger snowfall in winter season compared to the NC experiment. The processes explaining these increased SMB regions are probably linked to the strengthening of the atmospheric circulation along the Greenland eastern coast coming from northern latitudes, thus bring wetter and colder mass air. Despite these regions of positive SMB anomalies in 2150, the pattern in the SMB difference between 2-W and NC is generally a negative anomaly ranging from -2.3 m yr^{-1} to -0.4 cm yr^{-1} over the coastal areas of the GrIS. Following the decrease of the ST, the surface snow melting cumulated with less snowfall and more precipitation falling as rain instead of snow drive this SMB patterns. As a result, in 2150, there is a decrease of 112 Gt yr^{-1} of ice over the ablation area in 2-W with respect to NC, and, over the accumulation area, the simulated SMB in the 2-W experiment is 23 Gt yr^{-1} lower than the one simulated in the NC experiment. In 2100, the SMB anomaly shows the same patterns as the 2150 SMB pattern, but with lower magnitude (Fig. S1A). Over the ablation zone, the mean value of the SMB anomalies increases by a factor of 10 between 2050 and 2150 (see Table 1).

The SMB changes have an impact on the extent of the ablation zone. This area increases with time and, at the end of the simulation is 14 % larger in 2-W than in NC (Table 1). As a result, the ELA shifts more inland, in 2-W, by +12 km in the north eastern GrIS (wrt NC, Fig. 3).

Conclusion:
- Higher coastal slope
- More katabatic winds

The Cryosphere Discuss., https://doi.org/10.5194/tc-2017-230
Manuscript under review for journal The Cryosphere
Discussion started: 6 November 2017

The Cryosphere Discussions Open Access EGU

4.2.2 Impact on thickness and ice dynamics

The patterns of the surface elevation changes between the 2-W and the NC experiments (Fig. 6C) follow the SMB anomaly patterns. Over the eastern coast, southern central part and locally in some regions on the northern part of Greenland, the ice thickness increases, reaching more than 10 m in some locations such as East Greenland (Fig. 6C). This increasing ice thickness is explained by a larger positive SMB and a lower surface temperature over these regions. The second anomaly pattern is found all along the Greenland coast, where a decrease of the ice thickness is found in areas of lower ST and in the ablation zone. The main changes occur on the western edge of the GrIS where the thinning between 2-W and NC reaches more than -25 m (Fig. 6C). Further inland, there is a smaller thinning (-0.2 ± 3 cm in average after 150 years). As a result, averaged only over the entire ablation area, the thinning after 150 model years is equal to 9.0 ± 11.1 m (see also Table 2).These ice thickness anomaly patterns are observable, with lower magnitude, when comparing 2-W and NC after 100 years (Fig. S1C).

The main consequence of the increased thinning in coastal regions is the increase of the surface slope between the central part and the margin of the ice sheet. Increased surface slopes results in stronger katabatic winds. Furthermore, the thinnest parts of the GrIS become ice or snow-free or snow free, exhibiting bare ice and modifying albedo feedbacks, with a decrease of the surface albedo which amplifies the GrIS melting.

The ice dynamics computed by GRISLI are also impacted by the full representation of the interactions. Compared to the NC experiment, the ice velocities simulated with the 2-W experiment show a succession of positive and negative anomalies (Fig. 4B). The ice velocities increase from the central part of the GrIS to the coastal regions. The increase-decrease velocity pattern is amplified in the 2-W compared to the NC because of the larger thickness anomaly and follows the same processes than explain in Sect. 4.1.2.

4.2.3 Impact on SLR contribution and ice sheet area

After 150 model years, the melting contribution to global SLR reaches +20.4 cm in the 2-W experiment. In comparison, the melting obtained in the NC experiment is equivalent to a SLR of +18.5 cm . This difference (Fig. 7) is linked to the better representation of the interactions between the GrIS and the atmosphere. The coupling allows for a better representation of the processes occurring at the margin, and in particular the ice sheet margin retreat. As a result of ice melting, GrIS coastal grid points can become ice free. The GrIS extent in the 2-W experiment is reduced by 52 400 km^2 compared to the NC experiment and increases exponentially with time (Table 1). Thus, the ice sheet mask field, which represents the ice coverage percentage of each grid cell of the grid used by the models, and which is therefore dependent of the ice sheet extent, decreases with time (Fig S2B).

To evaluate the projected GrIS melting contribution to SLR, the SMB integrated over the ice-covered areas (i.e. the sheet mask field) is often used (Fettweis et al., 2013). However, this method could lead to strong uncertainties in the SLR contribution obtain. For example, using the NC result in 2150, if we integrate the SMB over the no updated ice sheet mask (as in the NC method), we calculate an integrated SMB 158 Gt yr^{-1} lower than using the updated ice sheet mask simulated by GRISLI (as

The Cryosphere Discuss., https://doi.org/10.5194/tc-2017-230
Manuscript under review for journal The Cryosphere
Discussion started: 6 November 2017

The Cryosphere Discussions
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in the 2-W method). This higher integrated SMB, obtained when using no updated ice sheet mask, is only explain by taking into account the GrIS regions becoming ice free compared to the updated ice sheet mask. We show that using a fixed ice sheet mask (as in NC) leads to a large overestimation of the contribution to SLR calculated from SMB.

Don't bury the lede.

4.3 Differences between 2-W and 1-W experiments.

Anomalies of ST, SMB and surface elevation for the averaged 2145-2150 period between the 2W and the 1W experiments present similar features (Fig. 8) than those obtained between the 2W and the NC experiments, but with lower magnitude. Over the coastal regions, a larger increase in ST is obtained with 2-W as well as a lower SMB and a larger decrease in surface elevation (Fig. 8), hence highlighting the role of the feedbacks between the ice sheet and the atmosphere that are taken into account in 2-W but not in NC. As an example, the katabatic wind feedback preventing the prenetration of warm air results in colder 2-W ST compared to 1-W.

In 2150, the GrIS SLR contribution obtained in the 1-W experiment reaches +19.9 cm, i.e 0.5 cm less than in the 2-W experiment (Fig. 7). Although this difference seems quite low, the local altitudes changes are larger in 2-W than in the 1-W experiment. Indeed, even if the median value of the ice thickness anomalies (2-W vs 1-W) between 2000 and 2150 are quite similar (respectively 73.3 m and 72.4 m), some regions show stronger surface anomalies (Fig. 8A). Scatter plots of surface elevation anomalies between 2-W and NC (red dots), 1-W and NC (green dots) and 2-W and 1-W (blue dots) as a function of the ice sheet altitude show the spatial variability of the ice thickness response to warming climate (Fig. 9A). The 2-W method yields negative and positive anomalies relative to NC, while the 1-W method mainly yield s negative values (relative to NC). For both experiments, the regions where the ice thickness is under 1000 m are the most impacted by the warming climate. For these lower to medium altitude points, there is a strong variability of surface elevation anomalies. Thus, for the 2-W experiment, the anomalies range between +16.4 m (98 % quantile value) and -43.1 m (2 % quantile value). For the 1-W experiment, the surface anomalies range between -1.5 m (98 % quantile value) and -45.2 m (2 % quantile value). Above 1000 m, the higher the altitude, the smaller the surface anomalies (Fig. 9A). The regions at low elevations are the most sensitive to the coupling method and to the warmer climate. This sensitivity to altitude increases with time, and is stronger for the 2-W experiment than for the 1-W experiment (Fig. 9B). High altitude regions are less sensitive to climate changes and to the coupling method used (Fig. 9B).

These thickness changes are correlated with changes in ice ablation and ice accumulation area. After 150 years, the ELA shifts more inland in 2-W than in 1-W (Fig. 3), and the ice ablation area is 11 % greater in 2-W than 1-W after 150 years.

What is the correlation factor?
They are correlated everywhere in this paper,
so why mention just now?

The Cryosphere Discuss., https://doi.org/10.5194/tc-2017-230
Manuscript under review for journal The Cryosphere
Discussion started: 6 November 2017

The Cryosphere Discussions
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5 Discussion

5.1 Limits of the models

The 5 km grid resolution of GRISLI does not allow represents the smallest peripheral GrIS glaciers to be finely represented. This could limit our results: as we have shown that the coastal GrIS regions are the most sensitive to climate forcing, the GrIS contribution to SLR could be enhanced by increasing the spatial resolution of the ISM in these regions. Furthermore, as we hypothetise an identical basal drag over time, we underestimate the acceleration of the ice flow of the glaciers due to the basal lubrification coming from meltwater or rainfall percolation at depth and reaching the bedrock (Kulessa et al., 2017). Anther limit of the GRISLI model is its simple representation of the grounding line position and thus of the buttressing effect which could impact the ice dynamics (Gagliardini et al., 2010). Except for these aspects, GRISLI is a good tool to be coupled in future ESMs in order to take the GrIS evolution into account with a combination of a good representation of the ice dynamics and a limited impact on the added computational resources. As for the ISM, increasing the grid resolution of MAR would allow to better represents atmospheric-topography feedbacks and more complex atmospheric processes which can have an impact on the SMB in the steep coastal regions. However, as for ice dynamic modeling, the higher is the resolution, the higher is the computational resources needed to produce results.

The absence of any representation of the GrIS-ocean feedbacks is also a limiting factor. Indeed, as the GrIS is an island, several glaciers are in direct contact with the ocean and feedbacks could take place between peripheral glaciers and the ocean. The warm ocean water could accelerate the melting of the glaciers and the added fresh water in the ocean could in turn, have an impact on sea surface temperatures, oceanic circulation and sea-ice cover. The GrIS and the atmosphere evolution could be both modified by this added fresh water flux in the ocean system.

Is this really a problem once tidewater glaciers have retreated?

5.2 Limits of the methods

The 1-W coupling method neglects the spatial variability of the thickness anomaly and underestimates regional feedbacks compared to the 2-W method. These differences are explained by the linear relationship used, in the 1-W coupling method, to correct the atmospheric fields (SMB and ST) as a function of the surface elevation anomaly, as developped by Franco et al. (2012). Indeed, the relationship between the atmosphere and the GrIS changes is nonlinear because surface elevation changes interact not only with both SMB and ST, but also with all the other atmospheric fields which influence the SMB or the ST directly or indirectly, as for example the winds, the humidity, and the albedo. The 2-W method appears to be the best way to simulated atmospheric-GrIS feedbacks.

If the main objective is to compute the SLR contribution from the entire GrIS without investigating atmospheric or GrIS changes at the regional scale, the use of the 2-W coupling method with a high resolution seems avoidable until 2100. However, over longer time scales, or to study more regional processes changes, the use of a the 2-W coupling method is necessary to represent the local feedbacks between the atmosphere and the GrIS fields and ensure that the SLR contribution is not underestimated by simulating. As an example, the changes in the GrIS extent and ice surface slope have a direct impact on

Beyond 2150?

The Cryosphere Discuss., https://doi.org/10.5194/tc-2017-230
Manuscript under review for journal The Cryosphere
Discussion started: 6 November 2017

The Cryosphere Discussions

surface albedo and strength of katabatic winds.

Although the difference in the GrIS melting contributions to SLR between 1-W and 2-W seems low, the use of the 2-W method to compute the ice sheet evolution for 50 additional years (from 2100 to 2150) with the constant forcing of the year 2095 contributes to increase the ice mass loss contribution to SLR by +0.5 cm compared to use of the 1-W method. This volume contribution ranges between 25 % to 100 % of the loss of peripheral Greenland glaciers volume in 100 years derived from the RCP8.5 scenario (Radić et al., 2014; Machguth et al., 2013).

6 Conclusions

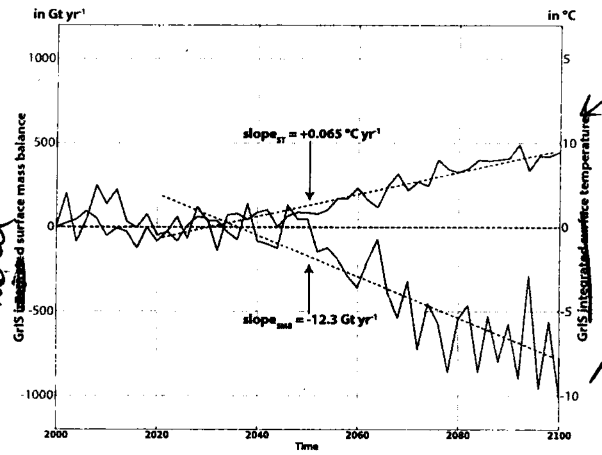
In this study, we have improved the representation of the interactions between the GrIS and the atmosphere by developing a full coupling between the Greenland ice sheet model GRISLI and the atmospheric model MAR (2-W experiment). To assess the importance of this improvement, we have investigated the atmosphere and ice sheet responses to the RCP 8.5 warming climate scenario, and we have compared the 150 years of our fully coupled experiment (2-W) with two other experiments using a less complex coupling method (1-W) and a no coupling at all (NC). The fully coupled approach under the RCP 8.5 scenario produces a GrIS melting contribution to SLR of +20.4 cm in 2150, while the 1-W and NC methods produce a GrIS contribution to SLR respectively of +19.9 cm and +18.4 cm, respectively. The difference of 0.5 cm between the 2-W and the 1-W methods represents at least 25 % of the contribution of the peripheral Greenland glaciers melting estimated for the next 100 years using the same RCP 8.5 scenario (Radić et al., 2014; Machguth et al., 2013). This difference, increasing with time, is mainly explained by representation of local interactions between the GrIs and the atmosphere, only possible with the 2-W method. Furthermore, even if the difference is not perceptible in 2100 and it is low in 2150, we have shown that the ice loss computed from the integration of the SMB over a fixed ice sheet mask is 21 % higher than that obtained with the use of an evolving ice sheet mask. This means that most of RCM-based studies have probably overestimated the ice loss computed from a change in SMB.

However, with the 5 km grid resolution of GRISLI, we cannot reproduce the fine-scale structure of the Greenland coast and glaciers. Using an ice sheet model with higher resolution and more complex physics (i.e. Full-Stokes models) and a fully coupled method would probably amplify the sensitivity of these coastal regions. This argument is also valid for the atmospheric model for which a higher resolution would be beneficial for the representation of fine scale atmospheric processes over the ice sheet. We showed that it is at small spatial scales that the coupling method makes most difference. It would therefore be very interesting to find the optimal resolution of the ice sheet and the atmospheric model, for ISM-RCM coupling. Furthermore, since the Greenland ice sheet and glaciers are in contact with the oceanic component, changes in oceanic characteristics, due to the input of freshwater from GrIS melting or due to the warming climate scenario, could in turn disrupt the GrIS and atmosphere evolution. The next step of this study will be therefore to improve the representation of the oceanic component by developping a fully coupled method between an ISM, an RCM and an oceanic model to evaluate the impacts on the Greenland polar region but also on remote regions.

The Cryosphere Discuss., https://doi.org/10.5194/tc-2017-230
Manuscript under review for journal The Cryosphere
Discussion started: 6 November 2017

The Cryosphere Discussions

Fig 2A implies this should be ~5C anomaly, not 10C



Need larger font

Figure 1. Results for the no coupling experiment. Total surface mass balance in Gt by year over GrIS (red line) and annual surface temperature in Celsius averaged over GrIS (green line) compared to the year 2000. The dashed lines represent the linear interpolation both for ST and SMB over 2020 to 2100.

Is this surface T or ΔT anomaly as compared w/ base?

The Cryosphere Discuss., https://doi.org/10.5194/tc-2017-230
Manuscript under review for journal The Cryosphere
Discussion started: 6 November 2017

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Put titles on top

Color scale 3 backwards

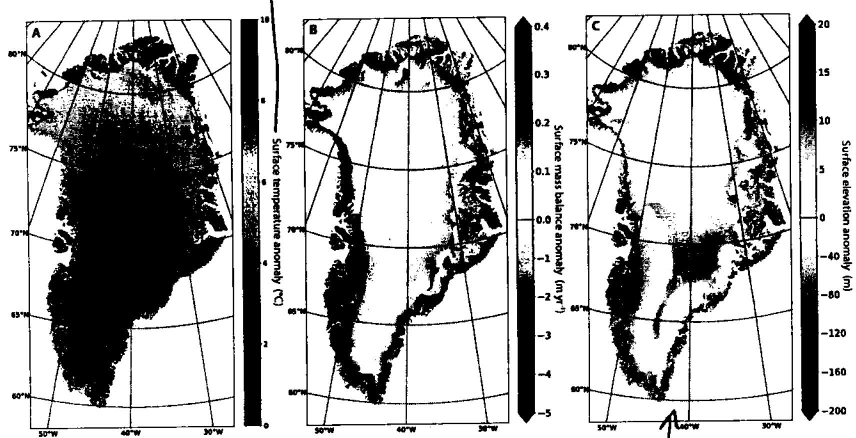

Figure 2. Mean anomalies between the last five years (2095-2100) and the first five years (2000-2005) of the No coupling experiment: (a) Annual surface temperature anomalies °C; (b) Annual surface mass balance anomalies in Gt yr^{-1}; (c) Surface elevation anomalies in m. The positive scales for the b and c are 10 times lower than the negative scales.

Are any of these anomalies greater than total ice in that grid cell?

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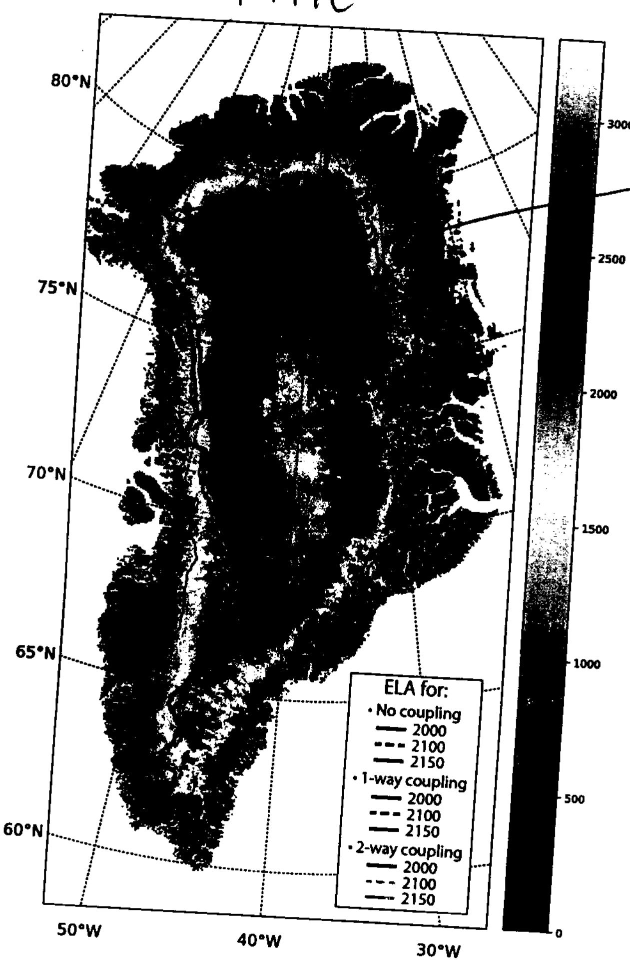
Title?



What is the significance of an ELA at 3000m?

Units?

Figure 3. Mean GrIS surface elevation for the last five years (2145-2150) of the 2-W experiment. The solid black line represents the equilibrium line altitude (limit between accumulation and ablation zone) in 2000 for the NC, 1-W and 2-W experiments. The straight and dashed colour lines represent respectively the ELA for the periods 2145-2150 and 2095-2100 for: in blue for the NC experiment ; in red for the 1-W experiment; in orange for the 2-W experiment. The ELA of the 2095-2100 (dashed lines) of NC, 1-W and generally 2-W are superimposed.

Conclusion: No=1W, 2W melts more.

Do we get something like Fig. 2 for 1w & 2w cases?

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Don't use green for >0 and <0. Try Red-Blue scale.

Don't use red @ both ends. Try: Green=fast Yellow=medium red=slow

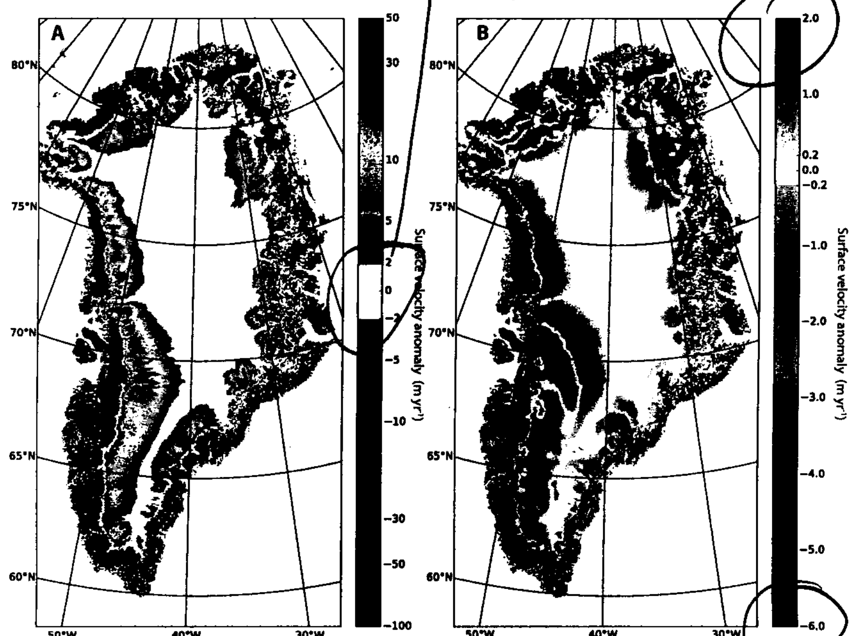

Figure 4. Mean surface velocity anomalies in m yr^{-1}: (a) between the last five years (2095-2100) and the first five year (2000-2005) of the NC experiment; (b) between the last five years (2145-2150) of the 2-W experiment and the last five years (2145-2150) of the NC experiment.

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(Title?) (Title?) Reverse color scale

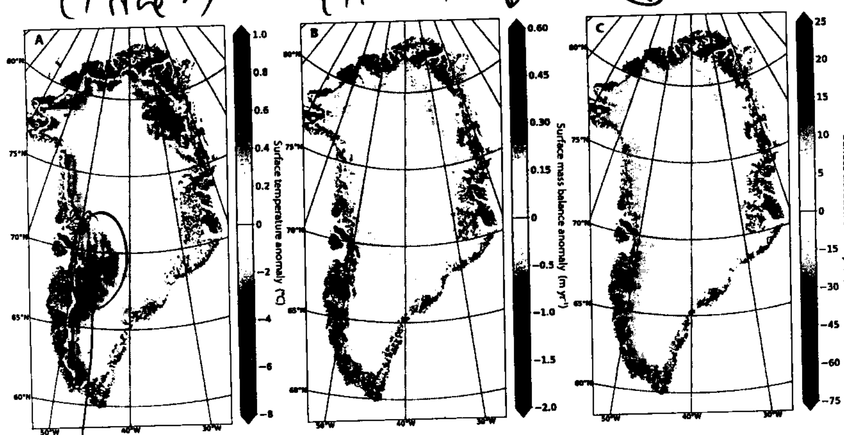

Figure 6. Mean anomalies between the last five years (2145-2150) of the 2-W experiments and the last five years (2145-2150) of the NC experiments: (a) Annual surface temperature anomalies in ° C; (b) Annual surface mass balance anomalies in Gt yr^{-1}; (c) Surface elevation anomalies in m.

temperature

Why the N-S striping artifacts?

Are those MAR or GRISLI plots?

Overall: what is correlation between SMB anomaly & elevation anomaly? It seems quite high.

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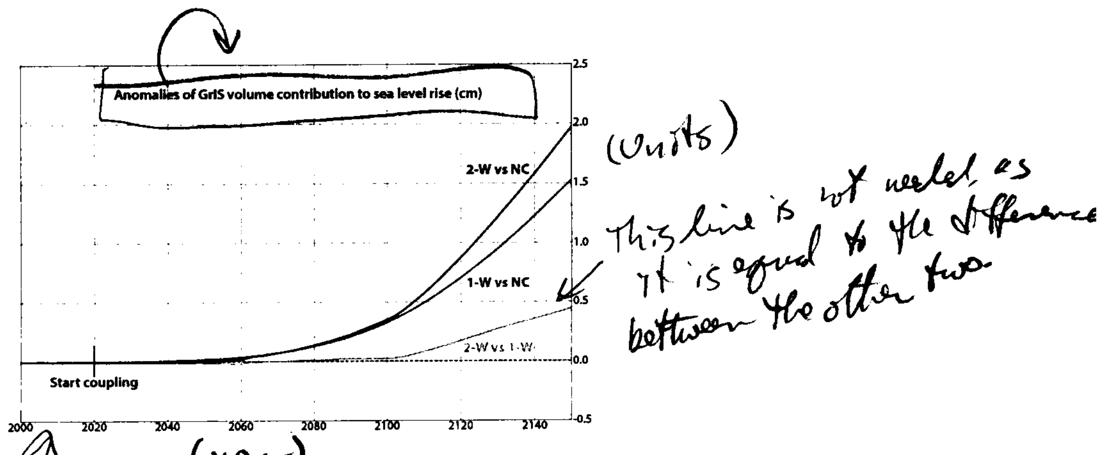

Figure 7. Contribution of the Greenland ice sheet volume contributions to sea level rise (cm) compared to the year 2000 for the three experiments : - Blue line: anomaly between 2-W experiment and no coupling experiment; - red line: anomaly between 1-W experiment and no coupling experiment (NC); - Orange line: anomaly between 2-W coupling experiment and 1-W coupling experiment.

Plot absolute NC, 1w, 2w.
That works better w/ Section 4.2.3.