Dear Editor,

As requested by the Editor Prof. Valentina Radic, we have revised the paper in time. We are very grateful to you for accepting the successive deadline extensions and we apologized to the time takes to submit this revised version. This time have been helpful for us to largely improve the manuscript following the numerous reviewers' comments and detailed concerns
during the peer-review process.

Following the J. Fike and the referee #3 reviews, we have even more clarified sections dealing with initialisation of the atmospheric model MAR (Section 2.1) and of the ice sheet model GRISLI (Section 2.2).

We also redid all the figures to fit with the revised result Section 4 and following the referee #3 comments.

Finally, we made our best to improve the English language.

Best regards,

Sébastien Le clec'h (on behalf of all co-authors)

We would like to thank the reviewer J. Fyke for the evaluation of our study. Please find below the reviewer's comments in black font and the author's response in blue font.

Responses to J. Fyke (Reviewer 1)

I find this paper to be much improved. Thanks to the authors for working to address reviewer comments. I have two remaining areas of concern that I think require greater 'caveating' in the final paper.

Thank you for this comment.

P6L4: "SISVAT takes only 7 years to reach equilibrium". I am slightly surprised that the full SMB field over GrIS takes 7 years to equilibrate (for example, there must be a few locations where longer equilibration is necessary to establish the simulated ELA..?). Perhaps in other words, how is the treatment of bare ice ablation zones treated and is snow (as represented in SISVAT) initialized with spatially varying thicknesses? I think readers (especially those familiar with the multi decade timescale of firn, for example) will be surprised by the 7 year spin-up, and will want further information on snow initialization procedure. Please include.

Thanks for the suggestion. In fact, the snow model SISVAT is first initialised with the averaged snowpack coming from previous MAR simulations carried out under present-day conditions (1960-1999). Using this snowpack equilibrium MAR is then initialised from 1970 to 1975 using MIROC5 forcing fields.

In the new version of the manuscript, we modified the paragraph:

"Because the snowpack in the land model requires generally longer time scale than MAR to reach an equilibrium with the atmospheric forcing, here MAR is spun-up for 6 years forced at its lateral boundaries by outputs from MIROC5 from 1970 until 1975 and by an initialised snowpack coming from a previous MAR simulation carried out under present-day conditions (1960-1999)."

P9L5: Thank you for noting the application of negative SMB values outside the present-day ice sheet mask. While I accept that this is the philosophy/methodology chosen for this study (which is mirrored elsewhere, e.g. ISMIP6 standalone experiments I believe) I think this needs a greater caveat in the context of a two-way coupling manuscript. This is because in GCMs, the use of flux corrections (which this functionally is, in the form of a flux-based bias correction) has been essentially entirely abandoned, because it technically violate true coupling (the conservative transfer of fields between components, as in the real world). This introduces hard-to-understand ambiguity into final results (e.g. impact of coupling-induced feedbacks).

We acknowledge that our methodology is not suited for a true coupling. In addition to the flux correction you rightly mention, we also follow an anomaly method between the ISM topography changes and the atmospheric model. Thus, in our model framework, we can't have mass and energy conservation between the ISM and the atmospheric model. This caveat has been further emphasized in the revised manuscript (Sec. 5):

"Moreover, the use of an anomaly method to account for the change in topography is incompatible with a conservative coupling between the ice-sheet model and the climate model. This is further amplified by the fact that we use a flux correction outside the present-day ice margin to force ice removal. This methodology has been followed to limit the impact of biases from the atmospheric model and from the initialization procedure, but the imposed ice removal outside the present-day ice mask may bias locally the model response towards increased ice thinning. Since our simulations are run under the RCP8.5 forcing scenario, this has probably a negligible impact. However further studies of future climate with alternative scenarios and/or GCM forcing, and even more paleoclimate studies, should ideally avoid using this kind of flux correction".

Finally, I'm not sure I agree with the final sentence of the paragraph referenced here: it could be that remaining SMB biases (which are probably mostly related to MIROC biases) would drive GrIS expansion, even during the warming future. From a paleoclimate perspective, this approach also limits the model configuration from expanding in colder climate states (though I recognize that's not the point here).

We fully agree with your comment and we do not recommend the use of an artificial SMB correction outside the present-day ice mask (i.e. strongly negative SMB) for paleoclimatic experiments. MAR uses MIROC outputs only at its lateral boundaries. Inside its domain, MAR generates its own boundary layers climatic fields and is less impacted by near-surface MIROC biases. In our study, and under the RCP8.5 scenario, the GrIS margin region is only marked by ice thinning. However, under a different RCP scenario and/or a different GCM forcing we could obtain locally an ice expansion.

In general, please more clearly describe the potential caveats of the imposition of an artificial SMB 'moat' around the ice sheet based on the observed ice sheet shape, so that readers are clearly aware of the possible implications to the main feedback-quantifying results.

Please see the text addition suggested in our response to your first comment. We hope this will help the reader to better understand the caveat of using such a flux correction.

We would like to thank the reviewer #3 for the evaluation of our study. Please find below the reviewer's comments in black font and the author's response in blue font.

Responses to Reviewer 3

Summary:

The authors have dramatically improved the quality of the paper since the last version and most of my comments have been sufficiently responded to. Since the paper has changed a lot, I still have a number of additional minor comments that the authors should consider before publication.

Thank you very much for this comment.

Because Figure 9 does not appear to be displayed in the manuscript, I feel I have to tick major revisions. If this problem should be resolved and the figure is checked by at least one of the reviewers, I don't need to see the manuscript again.

After reading your comments, we realized that Fig. 9 did not appear in the pdf file. In the new revised version, we will pay a close attention that the problem does not occur again.

General comments:

It is not clear to me why the sea-level contribution (and some other quantities like surface elevation, ice thickness, ice masks ...) should be averaged over 10 years periods in this reporting. In consequence, a forcing period of 100 years (2000-2100) is practically reported as a 90 year difference in the results, which is confusing. The sea-level contribution is physically a time integrated quantity and does not exhibit inter-annual variability that has to be averaged out. In most cases (instead where strong inter-annual variability exists (SMB, temperature) I would omit the averaging and calculate direct differences. If not, the labels in the table would have to be adjusted to the centre of the averaging period in all tables and plots instead.

Following the reviewer 1's suggestion, we averaged our model results over a 10-year period (in the first revised version) instead of a 5-year period (in the initial manuscript). This is important for climatic variables (e.g. SMB, temperature, winds) in order to build robust climatologies. However, we acknowledge that for integrated quantities such as sea level, ice volume and ice thickness, it is more relevant not to average the values. To account for your comment, we did the computations of a few diagnostics (e.g. ice thickness change) using the variable at 2150. As a result, Figures 3, 4, 8c, 10c and 11 have been modified accordingly. However, we verified that the new results are very close to those obtained from a 10-year mean (within 1 or 2%) showing that the message given in the previous version of the manuscript was not altered.

Concerning, the sea-level: We acknowledge that there was an error in the captions of Fig. 12 and of Table 1 caption (corrected in the new revised manuscript) and that the values reported in Table 1 correspond to those obtained in 2050, 2100 and 2150 (not averaged over a 10-year time period).The situation is different in Table 2 (now combined with Table 1 following your suggestion) since the sea-level is here inferred from SMB variations between 2100 and 2000 or between 2150 and 2000. Since SMB is a climatic variable (that may be subject to inter-annual variability) we think that the use of a 10-year mean is more appropriate.

In the reorganisation of the manuscript (that I appreciate), the 2W experiment is now discussed first as the reference and differences to NF and PF later. Since you introduce 2W as the standard experiment, I would plot and discuss the differences NF-2W and PF-2W (instead of e.g. 2W-NF). It just changes the sign, but seems more consistent with the rest of the document. I think it would also make sense to explain the model setup/coupling method in that order, starting with 2W.
Thank you for the proposed reorganization. We prefer to keep the presentation of the coupling methods in our initial order. As it is presented in the manuscript, the methods are shown from simple to more complex and we think that it is easier to follow since the reader has simply to understand the additions from one method to the other. However, it is true that it is somehow subjective and that we can change the order if you strongly suggest us to do so.

Minor comments

I believe the affiliation count is incorrect with Fettweis and Wyard linked to Brussels and Ritz linked to Liege.
This has been corrected in the new version.

P1 L9 Start a new sentence after MAR: "They are fed ..."
Modified

P1 L16 Remove "important". There is already an "important" in the line before.
Removed and replaced by “significant”.

P1 L18 Remove "tend to" before "favour" to make this statement less vague.
Removed

P1 L20 Does it "reduce the SMB signal" or rather re-distribute the additional mass. Clarify.
Replaced by “counteracts the SMB signal”

P1 L22 This should probably be "ice volume above floatation".
Thank you for this precision. This has been specified in the revised manuscript.

P2 L1 The results this conclusion is based on have not been described so far in the abstract. This must arise from a comparison to the uncoupled experiments. Suggest to mention those before this conclusion.
This has been clarified:
“*The comparison between the coupled and the two uncoupled experiments suggests that the effect of the different feedbacks is amplified over time with the most important feedbacks being the SMB-elevation feedbacks*”.

P2 L14 Replace "polar" by "Arctic".
Replaced.

P2 L18-21 Complicated sentence, consider splitting in two and revising.

We have splitted the sentence in two, as follows:

"However various feedbacks between the atmosphere and the GrIS impact the ice-sheet surface characteristics such as ice extent and thickness. This has potential consequences on ice dynamics (e.g., due to changes in surface slopes) and may lead to SMB variations that can therefore affect the total ice mass of Greenland".

P2 L23 Add "changes" before "in ice-covered area" to make the relation clearer.

Clarified as suggested.

P2 L23 I think with "albedo feedback" you mean here the change in land surface type from ice to tundra, which leads to warming and further ice sheet retreat. This is typically a slow process, I would call "planetary albedo feedback". There is also a more immanent "melt-albedo feedback" as melting snow at the surface absorbs more short-wave radiation, leading to more melting and increasing albedo. These feedbacks are quite different in nature and time scale and should ideally be distinguished.

Yes, we fully agree with you and replaced "albedo" by "planetary albedo"

P2 L27 Replace "predict" by "project".

Replaced.

P2 L29 In the manuscript you use different ways to order the references. Here it goes from newest to oldest. The TC guideline is open ("In terms of in-text citations, the order can be based on relevance, as well as chronological or alphabetical listing, depending on the author's preference."), but I would at least try to be consistent in all cases throughout the manuscript to not confuse the reader.

Thank you for this remark. We have chronologically ordered the references throughout the revised paper.

P3 L32 The approach of Edwards et al. was only used to correct for the SMB-height feedback. Here it sounds like it was also used for downscaling ("An alternative approach [to Franco]"). Please reformulate.

We agree with the possible confusion and have reformulated the sentence to clarify the approach of Edwards et al. (2012) as follows:

"An alternative approach to correct the SMB field from surface elevation changes is based on statistical relationships between altitude and SMB (Edwards et al., 2014b). Also been derived from MAR, this approach computes a SMB-elevation feedback gradient for regions below and above the equilibrium line altitude in the northern and southern parts of GrIS, with limited additional computing resources".

P4 L9 Replace "surface energy balance" by "surface mass balance".

A better representation of the surface mass balance requires necessarily a more complex representation of the surface energy balance. To clarify our idea we have changed the sentence as follows:

"Additionally, the use of a detailed snow model such as that implemented in MAR (Fettweis et al., 2017) or RACMO2 (Noël et al., 2015) allows a more accurate description of the surface

properties (e.g., snow cover, albedo, surface melting) and therefore a better representation of the surface energy balance and hence of surface mass balance".

P4 L27 Replace "second" by "third".
Replaced.

P4 L31 "to the other, uncoupled experiments".
Modified.

P5 L11-13 "and 24 vertical levels" misses a verb. Suggest "MAR has a horizontal resolution of ... and 24 vertical levels ..."
The sentence has been modified as follows:

"MAR has a horizontal resolution of 25 km x 25 km and 24 vertical levels to describe the atmospheric column in sigma-pressure coordinates (Gallée and Schayes, 1994). The MAR domain covers the Greenland region (6600 grid points), from 60°W to 20°W and from 58 °N to 81°N".

P5 L17 Maybe "covered by at least 0.001 % tundra and at least 0.001 % snow" to make clear both surface types are represented. Also add an explanation why that is done.
Thank you for the suggestion. In the new revised version we explained why MAR used a minimun percentage of tundra and snow in each grid cell:

"Each grid cell is assumed to be covered by at least 0.001 % of tundra and at least 0.001 % of snow so that the retreat or the expansion through time of snow (reciprocally tundra), especially at the margin, can be explicitly represented outside the original MAR ice-sheet mask".

This has been done with the aim of coupling MAR with an ice sheet model. Note also that when a grid cell is covered by 50 % of snow or ice it is considered as a permanent ice sheet grid cell, as specified in Fettweis et al. (2017).

P5 L21 Add why these parameter changes were needed? What was improved?
This point has been clarified in the revised paper:

"The differences with previous MAR versions (e.g., Fettweis et al. 2013) are only related to adjustments of some parameters in the representation of cloudiness and bare ice albedo. These new parameterisations allow to better account for the positive feedback that cloud cover exerts on surface melting (Van Tricht et al., 2016) and to represent the impact of melt ponds that strongly reduce surface albedo (Alexander et al., 2014)".

P6 L12 Regular grid? What projection? Refer to Bamber?
We apologize for the missing information. GRISLI uses a 5 x 5 km regular grid projected on a polar stereographic projection with a standard parallel at 71°N and a central meridian at 39°W. This has been added in the new version of the manuscript.

P6 b_melt is only defined for grounded ice, I suppose? What is done for shelf melting? Clarify.
This equation is valid for both grounded and floating ice, and so, b_{melt} is defined for both grounded ice and the ice shelves.

P6 L19 Remove "also" before "affects".
Removed.

P6 L27 Consider defining "The shallow ice approximation (SIA)" instead.
Modified as recommended for both SIA and SSA.

P7 L17 Section title should be "Initialisation procedure". "Spin-up" should be reserved for a consistent long-term transient run as often done with ice sheet models.
This suggestion has been followed.

P7 section 2.2.2 At the time of the first review of this manuscript, the GMDD paper of the same author describing the initialisation procedure was not available. It was therefore not possible to understand the procedure from the limited information given in this manuscript. I have therefore asked for more detail to be included. Now that the GMDD paper is published, it would be enough to give a broad overview of the method here and refer to the other paper. At the moment, a large part of 2.2.2 is a copy-and-paste from the other paper with a fair amount of technical detail. I don't think this is a problem in terms of plagiarism, but I would encourage the authors to rewrite and shorten this part to summarise the most important aspects.
Based on your suggestion, the presentation of the initial procedure has been entirely revised. We now only present the basic principles of the method without any equation, and we ask the reader to refer to the GMDD paper. Please see the revised version.

What I miss so far (also in the GMDD paper) is a clear idea of the basic principle of the method. While similar to PD2012, the main difference is that beta is modified in function of the thickness ratio, rather than the thickness difference. This has not been clearly state and is lost in the complicated formulation.
There are in fact two important differences with PD2012. On the one hand, as you rightly point out, the thickness ratio is used instead of the ice surface elevation difference. On the other hand, our procedure iterates from present-day ice thickness with multiple cycles short (i.e. decadal to centennial-scale) simulations. This drastically reduces the computation time relative to PD2012. We hope that the difference between the PD02012's method and ours is now better explained in the revised manuscript:

"*This method is based on the same basic principles as that of Pollard and DeConto (2012) except that their basal drag coefficient is adjusted as a function of the difference between modelled and observed ice surface elevation while we use the ice thickness ratio instead. Moreover, while the method suggested by Pollard and DeConto (2012) requires long (multi-millennial) integrations for the method to converge, we use an iterative method of short (decadal to centennial) integrations starting from the observed ice thickness allowing a more rapid convergence".*

*I also find the formulation through U^{corr} very confusing, since the modelled velocities are never directly corrected nor compared to observations. When the equations 4,5 and 6 are put together, one can arrive at a simple expression for the way beta is adjusted in the method (see below). I think it would add substantially to the process-understanding to include it here and/or of still possible in the GMD paper. In the method the basal drag coefficient beta is updated iteratively by multiplying the old beta with a factor rbeta = beta_{new}/beta_{old} (was U^{slid}/U^{corr}_{slid} in Eq 6 before). By combining Eq 6,5 and 4 one can show that the inverse of rbeta is beta_{old}/beta_{new} = rH + U_{deformation}/U_{sliding} * (rH-1), where rH = H^{G}/H^{obs}, is the ratio of modelled and observed ice thickness, and U_{deformation} and U_{sliding} are the modelled velocities due to deformation and sliding, respectively. This means that the adjustment of beta is in the end a function of the thickens ratio with stronger adjustment in regions dominated by deformation.*

Thank you for this suggestion. As mentioned above, we only provide the basic principles of the initialization procedure without giving any equation.

P8 L5-31 It seems to me that you start by skipping step 1 and then end by skipping step 2. If that is the case it would be clearer to exchange 1 and 2 and start the description with what is now called step 2.

We hope that the different steps of the procedure are more clearly presented in the new version.

P8 L20 Maybe add that beta_new = beta_old in the first iteration for clarity.

These notations are no longer used in the new version

P8 L28 I don't think NB_cycle is used afterwards. Maybe it is possible to avoid using the symbol all together.

Yes, you are right. Your suggestion has been followed.

P9 L7 I would say the "impact" of this condition is strong because it keeps ice from building up where there is none in reality. What you want to say is that it has a negligible impact on the projected SL contribution.

Applying a strong negative SMB outside the present-day observed ice sheet extent could have strong impact for colder climate projection. However, using RCP8.5 forcing scenario, the ice sheet expansion would be probably very limited, and we believe that it would have only a little impact at the global scale, and in particular on the GrIS contribution to sea-level rise. As also suggested by the Reviewer 1, we added a discussion of this caveat in Section 5. Moreover, the sentence you refer to has been changes in:

"This avoids ice growth where there is none in reality and allows to correct for both the potential atmospheric model biases (e.g., positive SMB values over tundra areas) and the initialisation procedure biases (i.e. too strong ice export towards the margins). However, for GrIS projections run the RCP8.5 forcing scenario, this condition has only a limited impact on GrIS contribution to sea-level rise since the ice extent will likely to keep on retreating over the next centuries".

P9 L8 Remove "quite" before "negligible".

Removed.

P9 L23 This sentence repeats most of what is already in the sentence line 20. Remove?

We agree with this suggestion. The sentence has been removed.

P9 L27 I don't think the SMB can be said to be consistent with the 5 km topography. Maybe "generate a 5 km resolution SMB for the Bamber et al. (2013) topography"?

By consistent we mean that the new 5 km resolution SMB inferred from the method of Franco et al. (2012) is now adapted to the fine scale features of the 5 km Bamber et al. (2013) topography. We have clarified this point in the revised paper:

"Thus, this method allows to generate a 5 km resolution SMB entirely adapted to the fine scale features of the 5 km Bamber et al. (2013) topography".

P9 section 3 You chose to show results for 2W first and discuss differences to the other experiments afterwards, I agree with that. But wouldn't it then make sense to also explain the model setup/coupling method in that order, starting with 2W?

See our response to your general comment. While we consider the 2W experiment as a reference, we think that the presentation of the different methods going from lower to higher complexity makes the understanding easier for the reader.

P10 L18 It is not clear what "aggregated" means. Some process to go from 5 km to 25 km, but what is it exactly? Please explain that better.

We acknowledge that this sentence was confusing. We tried to clarify this point in the revised version. The text has been changed in:

"Then, GRISLI computes a new GrIS topography and a new ice extent at a 5 km resolution. This new GrIS topography is then aggregated (i.e. geographically averaged) at the yearly time scale onto the 25 km MAR grid. The number of ice covered GRISLI grid points within a MAR grid cell relative to the number of ice-free GRISLI grid points is used to compute the new ice extent in MAR and to update the fraction of tundra relative to ice/snow covered surface type for the subsequent MAR run".

P10 L28 First sentence of 4.1.1 difficult to read. Move "in the 2W experiment" before (Eq. 1).

Modified as recommended.

P11 L5 You say there is no inter-annual variability anymore after 2100, but where does the +- 13 Gt/yr come from then? I suppose MAR still has inter-annual variability, but not the GCM boundary condition. This should be clarified.

The other numbers mentioned in this paragraph (280 Gt yr^{-1} and - 638 Gt yr^{-1}) are associated with mean standard deviations equal to 95 Gt yr^{-1} and 271 Gt yr^{-1}, representing errors of ~34 % and ~42%. This must be compared to the error of 13 Gt yr^{-1} which represents an error of only 1.6%. This much smaller uncertainty (compared to 34 and 42 %) illustrates the small inter-annual variability due to the repeated 2095-year MIROC5 as atmospheric forcing. Note also that the inter-annual variability simulated by MAR is driven by the inter-annual variability of the GCM (here MIROC5). However, MAR generates its own surface boundary layer fields which are not impacted by MIROC5 forcing.

P11 L8 I have problems with the formulation "two distinct patterns". You chose to analyse the results separated in two different regions. If you would have made four regions, you would probably observe four different distinct patterns. Can this be formulated better?

The point which is addressed here is that most grid points having surface elevation higher than 2000 m are characterized by a positive SMB anomaly, whereas lower altitude grid points have a negative SMB anomaly. This is the reason why we thought that "two distinct patterns" was relevant. However, we have reformulated the sentence in the revised version:

"The SMB anomaly between the beginning and the end of the 2W experiment is displayed in Fig. 2a. 65 % of the grid points having surface elevations higher than 2000 m are characterized by a positive SMB anomaly […]"

*P11 L26 There should also be a *shift* of the ELA, which should be important to note here.*

The ELA shift is discussed explicitly in the previous paragraph. Therefore, we did not find necessary to mention again this point in the paragraph dedicated to ST.

P11 L26-27 The relationship between the first and the second part of this sentence has escaped me.

We apologize for misunderstanding. We have modified the sentence to be clearer. This new sentence explains that even if the northern part of the GrIS is marked by the strong temperature increase between 2150 and 2000, the ablation processes over this region are counteracted by the increasing snowfall, as shown in Fig. 2b.

P11 L29 The ice thickness anomaly at what time? Clarify.

Clarified to specify that the the ice thickness anomaly we consider here is computed between the beginning and the end of the 2W experiment.

P11 L29 Reformulate "two distinct patterns".

Reformulated as recommended following your P11L8 comment.

P12 L8-9 Fig 5a and 5b look nearly identical, because you plot the integrated total SMB (5a) and integrated total ice flux divergence. It would be much clearer to compare integrated SMB anomalies (and integrated flux divergence anomalies) instead.

Here we disagree. The aim of this section is to evaluate the respective roles of atmospheric forcing (i.e. SMB) and ice dynamics (i.e. ice flux divergence) on the ice thickness variations between 2000 and 2150. These variations are directly given by the integral of dH/dt (i.e. left-hand term of equation 2), which also represents the ice thickness anomaly between 2000 and 2150. According to Equation 2 and assuming that the integrated basal melting is negligible, they can therefore be directly inferred from the integral of SMB and the integral of the ice flux divergence. Conversely, taking the integrated ice flux divergence and SMB anomalies cannot be directly compared to the ice thickness variations between 2000 and 2150 for the reasons mentioned above.

P12 L10 Replace "ice melting" by "runoff".

Replaced.

P12 L15 I think "ice dynamics" is a too broad term here, you may want to say that "ice flow" is impacted by changes in the geometry instead.

We replaced “ice dynamics” by “ice flow” or “ice flux” when we deal with ice flow but kept the term “ice dynamics” for mechanistic processes.

P12 L19 "by the combination" of what. There is only one thing mentioned in the following: larger surface slope. Clarify.

Yes, you are right. We clarified the sentence in the revised version and removed “by the combination of”.

P12 L23 What do you mean by "fully consistent with the decrease of ice thickness"? The velocities are dependent on both thickness and surface slope (and maybe on changes in basal sliding), as you state yourself in the next sentence. This is confusing.

We replaced the sentence you refer to with:

“Compared to the 2000—2010 period, this decrease ranges from -213 m yr^{-1} (5th percentile) to -0.2 m yr^{-1} (95th percentile), and favours the decrease in ice thickness”.

P12 L29 Is it really that clear cut? Changes in surface slope above 1500, ice thickness changes below? Maybe the wording should be modified to "dominated by" or "governed by", allowing for transitions between the two.

This has been reformulated in:

“For the Jakobshavn glacier, the ice-sheet areas located above 1500 m, are mainly characterised by an increase of more than 15 m yr^{-1} (i.e. 10 %) of the vertically-averaged ice velocities

as a result of increasing surface slopes (Fig. 7c). Conversely, areas below 1000 m are dominated by a slow down of the ice flow of more than 200 m yr^{-1} (i.e. 29 %) due to the decreasing ice thickness (Fig. 7c)".

P13 L30 "in the same region with full ice cover."
Modified.

P13 L31 Remove "artificially".
Removed.

P14 L7 "positive elevation-SMB feedback".
Modified as suggested.

P14 L8 "and thus increased runoff".
Modified.

P14 L21 You should explain why the SMB is lower in PF compared to 2W here. The SMB change in PF is only related to elevation change but in 2W wind changes are dominant.
SMB computed in both PF and 2W experiments accounts for the surface elevation changes simulated by GRISLI through parameterised SMB-elevation feedbacks in PF and explicitly computed feedbacks in 2W. The main difference between both experiments lies in the fact that, in 2W, simulated ice-sheet changes feedback onto the MAR simulated climate. This leads to a cold air convergence, as explained in Section 4.2.1, slowing down ice melt, hence a higher SMB in 2W compared to PF. This has been specified in Section 4.3:
"[...] the SMB simulated in PF has even become lower compared to 2W due to a much complex representation of the ice-sheet climate interactions. Indeed, as mentioned in section 4.2.1, in the 2W experiment, the GRISLI topography feedbacks onto the MAR simulated climate leads to a cold air convergence at the ice-sheet margins and thus to a higher simulated SMB".

P14 L29 Could you explain why there is mainly negative values in PF? It seems strange to me that there are not more positive values. Or are the positive changes so small that they are not visible on the graph. If so, it would be good to mention that.
First, we draw your attention that Fig.11a has been changed to represent the NF-2W and NF-PF ice thickness differences (instead of 2W-NF and PF-NF as in the previous version of the manuscript) as a function of the GrIS surface elevation. Thus, NF-PF values are now positive. The only difference between PF and NF relies on the parameterization of the SMB-surface elevation feedbacks taken into account in PF and ignored in NF: All the SMB components are identical in both experiments and the vertical interpolation inferred from Franco et al's method is only applied to SMB. As SMB (in PF) changes as a function of GRISLI simulated topography changes, the ice thickness decreases. Therefore the NF-PF differences are positive.

P15 L21 I am not sure the attribution to model resolution is as clear cut as it is written here, it seems like a speculation rather than a result for me. Using a different model as such could be an explanation as well I wouldn't exclude. Or is there any other information that convinces you about the resolution dependence, maybe your own experience running MAR at a lower resolution? Please add if that is the case.
Our aim was not to present the model resolution as a clear-cut statement to explain the differences between the Vizcaino et al's results and ours since we used the expression "may be explained". Moreover, Vizcaino et al. (2015) also mention the need to replicate their own study

with models having a finer resolution and a much complex representation of the model physics. As an example, SICOPOLIS3.0 is based on SIA only, suggesting that fast ice flow is not accurately represented. Moreover, the coarser resolution implies that ablation areas or processes such as katabatic winds, which are both strongly dependent on topography, and thus, on resolution are less well represented compared to models of finer resolution. Nevertheless, in the revised version, we tried to be less conclusive. The sentence has been reformulated in:

“While the importance of the SMB-elevation feedback may be dependent on the model itself, the larger contribution found in Vizcaíno et al. (2015), compared to our own study, could be explained by the coarser resolutions of ECHAM5.2 (~ 3.75) and of SICOPOLIS3.0 (10 km) with respect to MAR and GRISLI resolutions, implying for example that ablation areas or processes such as katabatic winds are less well represented”.

P15 L25 Albedo feedbacks have not been really discussed so far. Suggest to remove specific mention here.

Yes, you are right. “Albedo” has been removed.

*P15 L8 You have to make clear somewhere (earlier) that *the ice sheet mask (in the ISM) is free to evolve in all three experiments*, but that MAR does not see mask changes in PF and NF, because it is calculated on a fixed observed ice sheet geometry. I.e. make a clear distinction between the (modelled) ice mask in the ISM and in MAR.*

As recommended we added this information earlier, in the description of NF and PF experiments (see Sections 3.1 and 3.2).

P15 L31 It seems that at the very margin the albedo effect is over-compensated by the cooling due to wind anomalies. Make that clearer. Also note that the melt-albedo feedback (as distinguished above) should be active in the ablation at any case.

We think that “over-compensated” is too strong and we prefer the term “counteracted” since there is a clear negative ice thickness anomaly at the margins of the ice sheet at the end of the 2W experiment. We clarified as follows:

“As MAR sees the ice sheet retreating over time in 2W concomitantly with the increase in bare ground or tundra fractions (Fig. S5b), the albedo feedback takes place favouring further the ice melting, though counteracted by the katabatic wind anomalies (see Section 4.2)”.

P16 L1 "A widely used method to estimate the GrIS contribution to global sea-level rise".

Modified.

P16 L3 I would say you do something altogether different here by coupling two models. Reformulate.

Reformulated as follows:

“In the present study, we use a more complex method since the ice mass variations related to SMB changes are computed by MAR over a changing ice-sheet mask and topography as simulated by GRISLI”.

P16 L4 "by GRISLI, and MAR sees those changes in the topography", or similar.

Your suggestion has been taken into account (see our previous response).

P16 L9 A note on notation. These SMB symbols read like "MSK minus NF". Maybe use underscores instead.

We changed the notations in $SMB_{MSK_{NF}}$ and $SMB_{MSK_{2W}}$ respectively.

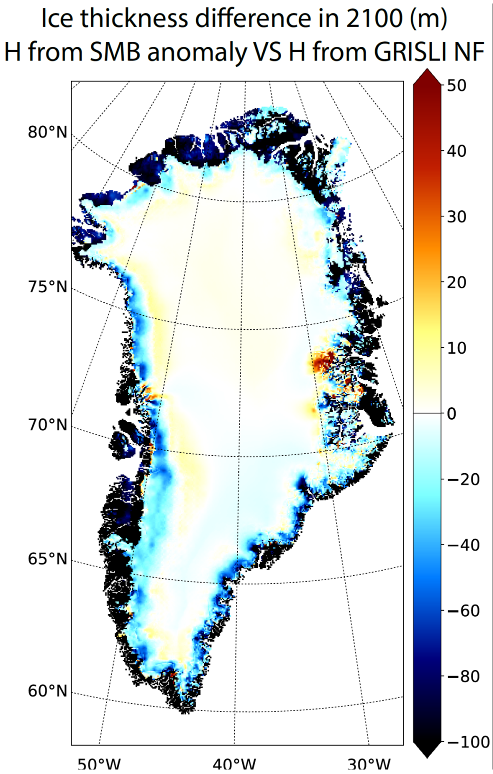
P16 L12 "this error has a similar magnitude compared to errors made when ...".
Modified.

P16 L14 I have mentioned this in my earlier comments and without insisting on it, I would like to reiterate it here: I think it would be interesting to see what mask changes would be produced by using the SMB changes in NF and PF to modify the ice sheet mask offline.
In order to answer your question, we computed the sum of the SMB anomalies obtained at the end of each year with respect to the mean SMB (i.e. SMB_{ref}) obtained during the reference period (1979—2005). This allows to evaluate the magnitude of the SMB variation relative to the reference period and to compute the corresponding variation of topography (obtained by neglecting the ice dynamics). For example, for year i the new topo is given by:

$$(topo)_i = \sum \Delta SMB_i + (topo)_{i-1}$$

With:

$$\Delta SMB_i = SMB_i - SMB_{i-1}$$



We then plotted the difference in ice thickness between the topography obtained in the way described above and the topography simulated by GRISLI in the NF experiment (in 2100). This figure (see above) shows a strong thinning at the ice-sheet margins. This result was expected because there is no ice flow from the ice-sheet interior to the margins since ice dynamics is not accounted for. To go a step further we could have determined the new ice mask associated with this new topography and carried out new MAR experiments to deduce diagnostics similarly to $SMB_{MSK_{NF}}$. However, we do think that this kind of experiments does not bring any added value to the manuscript. Indeed, the effect of the mask has been already extensively discussed. Moreover, this figure highlights the role of the ice dynamics which has been described in section

4.1.3. It seems therefore not necessary to add this new figure in the new version manuscript because, to some extent it would be to some extent redundant with what has previously been done.

P16 L18-20 This is already a long sentence and may be better to split, but I would suggest to add "two uncoupled experiments" before PF.

The sentence has been splitted in two and changes have been added as suggested.

P16 L21 remove "first" before step.

Removed.

P16 L27 I agree that extending your experiments by using additional and different GCMs would be an important next step in its own right. I would mention that first, not at the end of the paragraph. How the difference between 2W and PF would play out in those runs is certainly interesting, too. I would speculate that the stronger the forcing, the more important the differences between coupled and uncoupled experiments are, similar to the finding of stronger relative error for longer simulations (i.e. stronger forcing).

The sentence you refer to has been changed in:

"Whatever the experimental design, the large spread in SLR projections highlights the great uncertainty associated with the choice of the global climate model used to force MAR at its lateral boundaries. It raises the question to what extent the differences between 2W and PF or NF experiments would be amplified (resp. mitigated) with a stronger (resp. weaker) climate forcing than that simulated by MIROC5".

P16 L30 "Another question ..."

Modified.

P16 L32 "Another consequence is ..."

Modified.

P17 L5 I think repeating the exercise with a different RCM would also be important, since they may have different sensitivities.

The sentence has been changed in:

"There is therefore a strong need for iterating the present study with different global climate simulations run under an extended RCP8.5 scenario, but also with different regional climate models, that may have different sensitivities, to assess more accurately the impact of the different GrIS-atmosphere feedbacks and to better evaluate the uncertainty associated with the projected sea-level rise contribution from the GrIS."

P17 L6 "Another limitation is ..."

Modified.

P17 L9 Replace "we choose to" by "we have chosen in this study to".

Replaced.

P17 L14 I am not convinced it would be a good compromise to accept unrealistic model drift in the ISM in exchange of a better absolute SMB. I would say there is simply no way around further improving the initialisation, which may also require higher resolution of the ice sheet model.

There is apparently a misunderstanding. In this paragraph we discuss the limit of using an anomaly method to represent topography changes in the MAR model. We propose a solution to avoid the use of this anomaly method by using the GrIS topography simulated after the ISM initialisation as initial condition of the MAR model. Following this, MAR could then be initialised with the GRISLI simulated topography instead of the observed one. We do not propose a method having an unrealistic model drift (indeed, we propose to still use the Scrl topography that minimises the model drift. The key question here concerns the error made when using Sctrl as initial MAR topography rather than the present-day observed topography taken from Bamber et al. (2013). In this respect we have also to draw your attention on a the typographic error in the previous version (i.e. *GRISLI* was used instead of *MAR*. We have changed the sentence in: "*As an example, a reasonable compromise to avoid the use of anomaly method would be to use the topography obtained at the end of the spin-up iterative process (rather than S_{ctrl}) as initial ~~GRISLI~~ MAR topography to keep the mismatch with the observed topography as low as possible, and to initialise and perform MAR simulations with this spin-up topography*".

P17 L18 This is not a full sentence. Reformulate.

Reformulated:

"In addition, difference of resolution between MAR (25 km) and GRISLI (5 km) can cause artefacts in the results, especially at the edges of the ice sheet."

P17 L33 I agree, but aside from improving the model in this regard, you would also need to force the model with appropriate ocean forcing, which is so far excluded in your setup.

We fully agree with your comment and we do think that considering the oceanic forcing and or including an oceanic component in the modelling is a main step to improve the SLR projections. This is discussed two paragraphs later.

P18 L4 ""it seems to be preferable" is a bit weak a statement. I would say "it is essential".

Modified as suggested.

P18 L17 Add "taken" before "constant in time".

Added.

P18 L27 Can you find a better description than "GRISLI-like models"? Maybe "Large-scale GrIS models".

As we already used the term "Large-scale GrIS models" few lines over, we changed "GRISLI-like models" by the same used term.

P19 L4 I don't think the models used in Edwards et al were more sophisticated. In fact, GRISLI was one of them.

There was an error in the reference of Edwards et al in the previous version. In fact, instead of Edwards et al. (2014a), the appropriate reference is Edwards et al. (2014b). In this paper, besides using GRISLI, these authors also used Full Stokes models (Elmer/Ice) and High order approximation models (GISM, MPAS and CISM), which are more complex models than the hybrid GRISLI based on the SIA/SSA.

P19 L10 Apparently the use of AD is discouraged. Use CE instead or just write "year 2000 to 2150".

We thank you for the update. We remove *AD* in the entire manuscript.

P19 L16 I hope "In turn, changes in the shape of Greenland modify the ice velocities." Is not a conclusion of this study. That's ice sheet modelling.

We fully agree with you. However, what is obvious for the ice-sheet modelling community is not necessarily obvious for another community. Since this paper deals with RCM-ISM coupling, it may interest readers from climate modelling community who may be not familiar with the consequences of topography changes on ice dynamics. Moreover, our section 4.1.3 is devoted to changes in ice dynamics. It seems therefore justified to remind in the conclusion that changes in ice-sheet geometry influence the ice velocities. However, we slightly modified the sentence so as not to give the impression that this issue is totally novel:

"Accounting for the GrIS-atmosphere feedbacks amplifies the ice mass loss and changes in ice-sheet geometry with increased surface slopes from the central regions to margin areas and consequences on the Greenland ice velocities".

P19 L18 "accounting for the feedbacks".

Modified.

P19 L19 "sea-level rise of 20.4 cm in 2150 against 7.9 cm only in 2100" only illustrates that the forcing increases, but not that the feedback does. You could use changing differences between 2W and NF for that.

Thank you for this remark. We modified the sentence as follows:

"The effect of accounting for the feedbacks between GrIS and the atmosphere increases with time and becomes significant at the end of the 21st century, as illustrated by the 2W – NF difference in GrIS contribution to sea-level rise in 2150, i.e. 1.9 cm, against 0.3 cm in 2100."

P19 L25 Problematic conclusion, because RCM-only experiments (like you) also ignore other factors like ocean forcing. So they overestimate the SMB component, but that may be compensated by missing outlet-glacier changes. These details do matter.

Our sentence refers to the fact that RCM-based studies that do not account for changes in ice sheet extent (as in PF and NF) and/or ice-sheet topography (as in NF) overestimate the SLR contribution. This is somehow unrelated to the considerations of the oceanic feedbacks due to ice-sheet melting. We acknowledge that this source of uncertainty cannot be quantified with our setup that does not represent the oceanic feedback. The direction (over or underestimation) as well as the magnitude of these feedbacks have still to be quantified but are out of scope of our manuscript. The role of the oceanic feedbacks are nonetheless already extensively discussed in the discussion section. We also would like to draw your attention on the fact that 1/ a higher resolution of the ice-sheet model would allow to better resolve the outlet glaciers and to better represent the ice discharge to the ocean, but not the SMB which is primarily driven by the atmospheric forcing, 2/ capturing the changes in outlet glaciers (simulated by the ice-sheet model) has no effect on an experiment such NF since changes in ice-sheet geometry are not taken into account. In order not to present this last sentence as a universal truth, we have attenuated our statement by adding *"with our modelling setu*p". This obviously implies that the effect of the ocean component is not taken into account.

P19 L30 You don't mention the code for the GRISLI model. How can the readers obtain the code of the ice sheet model?

We apologized for the omission and have added this information.

P20 L6 It is not really my role to comment on the Acknowledgements, but in appreciation of your own work and that of the other reviewers, I think it is more than "the writing" that has improved in the manuscript. Suggest to write "helped to improve the manuscript".

We are of course very grateful for all the comments you provided

Tables:

P26 Table 1 No need to average (see general comment). I suggest to calculate 2050 2100 and 2150 differences to the year 2000 instead.

This has been corrected.

P27 Table 2 Same comment as for Table 1. No need to average. Then remove "Mean" from the caption.

Table 2 has been merged with table 1. However, the column referring to SMB (not SLR) is now included in the text Sect. 4.4. This has been clarified in the caption.

For a better overview these results (SLR) could be joined with table 1 and the two SMB values reported in the text.

Change label in first row to SMB_(MSK_NF).

Same comment as above: these SMB symbols read like "MSK minus NF". Maybe use under-scores instead.

As recommended, tables are now joined, symbols have been modified and SMB values in Gt/yr are only specified in the text as follows:

"Differences in SMB values exceed 23 % in 2150 (-842 Gt yr^{-1} for $SMB_{MSK_{NF}}$ against – 647 Gt yr-1 for $SMB_{MSK_{2W}}$)".

Figures:

Figure 1: Caption: What ice mask definition is used for the integral over the "entire GrIS", Clarify. Over which periods are the regression lines defined? Specify in the caption.

We used the ice-sheet mask taken from Bamber et al. (2013). The regression lines are computed from 2000 to 2039 and from 2041 to 2099. This information now appears in the caption. Note also that in the previous version, the different values reported in the figures were given in m/yr. For consistency with the main text, we have changed the figure and converted the SMB components in Gt yr^{-1}.

What does the black dashed line stand for? If it is for 0, a solid thin line may be better and should extend all the way to the left y-axis

We replaced the black dashed zero line by a light grey solid line and indicated in the caption what it corresponds to.

Figure 2: Why are b and c given on the MAR grid? Shouldn't they be given on the GRISLY grid for consistency?

Snowfall and rainfall amounts are not directly used by GRISLI and are therefore not downscaled at the ISM resolution.

In this figure as well as many other, anomalies are denoted x -- y. Not sure if the symbol "--" stands for minus. If so, why not use the minus sign "-" instead?

As recommended we replaced "--" by "-" for anomaly notation.

You should add a note that the colour key is non-linear.

The colour scale is linear but not symmetric between positive and negative values. This has been specified in the caption

The dashed grey surface elevation contours are not visible on a A4 printout of the paper (same for most of the other 2D plots). Consider using a darker colour (black). On the contrary, the lat/lon lines are much less important and could be less pronounced.

Your recommendations have been taken into account for all 2D plots.

Figure 3: Also here, it seems strange to report surface elevation as a time average, when you could as well just plot the 2150 result. Even stranger for the ELA! Just plot the 2000, 2100 and 2150 results in line with your legend in the figure. It looks like the 2000 and 2100 lines of the ELA are identical between experiments. You should mention that or that the lines are not (hardly) distinguished to avoid confusion. Use colours for the ELA lines that are not part of the colour key.

As recommended, we no longer use time averaging for the surface elevation and the colors for the ELA have also been changed. However, as ELA is inferred from climatic variables that are averaged on a 10-year period throughout the paper (see our response to your general comment), we used a 10-year time-averaging for the ELA representation. Following your recommendation, we also changed the colours of ELA.

Figure 4: Why not use the same colour scheme as in Figure 5a/b? That would make comparison easier.

We now use the same color scale as Figures 5a/b.

*Figure 5: It would help the interpretation a lot to compare integrated SMB *anomalies* (Fig 2a) and integrated ice flux divergence *anomalies* instead. Now the two plots are on first view identical and it is hard to visually extract the difference.*

See our response to your comment P12 L8-9

Explain in the text what the integrated ice flux divergence is when introducing EQU 2.

We added the definition of ice flux divergence in Section 2.2.1 when we present the different terms of Equation 2.

Add to caption that the grey shade is the land-sea mask.

Instead of the land-sea mask, we specified in the caption that the grey shade corresponds to non-ice-covered areas. .

Figure 6: Since you introduce 2W as the standard experiment, I would plot NF-2W and PF-2W instead (opposite sign). Add to caption that the grey shade is the land-sea mask. Remove the log10 label from the plot (applies also for other plots).

As recommended we plotted NF-2W and PF-2W instead of 2W-NF and 2W-PF. We also changed the caption. For grey shade, see response to your comment on Fig. 5.

Figure 7: Plot labels a,b,c, ... and refer to those in the caption.

Done

The Joughin et al (2010) data set has since been updated to full coverage over the GrIS. Consider using this improved data (ref below). The inset panels for observations are way to small. Include as additional full size panels on the top. You can gain extra space by removing redundant colour keys (the middle two are repeated). Remove "surface" before "velocity".

JOUGHIN, I., SMITH, B., & HOWAT, I. (2018). A complete map of Greenland ice velocity derived from satellite data collected over 20 years. Journal of Glaciology, 64(243), 1-11. doi:10.1017/jog.2017.73

We have modified the Figure 7 as recommended.

Figure 8: Same here, suggest to plot NF-2W differences instead.

Modified as suggested.

Figure 9: does not appear to be in the manuscript. I see only the caption!

This should be ok for any pdf viewer now.

Figure 10: Same comments as for Fig 8

Modified as suggested.

Figure 11: Shouldn't red in a) and blue in b) be identical? For some points that does not seem to be the case.

There was an error in the labels (not in the caption though). The red in a) is NF-2W while the blue in b) is PF-2W

Again, I doubt the averaging is really needed. If maintained, the labels should be adjusted to the mid-point of the averaging period.

The plot has been redone without time averaging.

Except for a few points, PF-NF is only negative. Why?

See our response to your comment P14-L29.

Figure 12: Please use a different set of colours in b and a to avoid confusion. The averaging is really not needed here.

The caption was wrong, the averaging is not used here. We have corrected the caption in the new version of the manuscript.

Supplement

Figure S1:Match figure title with caption (anomaly vs error).

Caption corrected as recommended

Figure S2: Mention that the scale is non-linear.

The colour scale is linear but not symmetric between positive and negative values. Caption modified accordingly..

Figure S3: Suggest to repeat and update the caption from S2 for convenience of the reader. Mention that the scale is non-linear.

Updated as recommended.

Figure S4: Mention that the scale is non-linear.

As previously mentioned, the colour scale is linear but not symmetric between positive and negative values. This had been added in the caption

Figure S5: Not clear what "loosing 100 % of the ice cover" means. I think you mean the pixels that have 0% ice cover in the end, even if they had less than 100 % to start with. Maybe "pixels loosing all of the initial ice cover"?

You are right, we clarified the caption.

Figure S6: Remove first part of the caption which is scrambled. Shouldn't the annual snowfall anomaly (a) be equal to the sum of the 4 seasons? Or are you plotting the snowfall anomaly in m/season instead of m/yr? Same problem may apply to S7 and S4, but I don't know if plotting ratios solves it for you in the latter.

With the seasonal values expressed in m/yr, the annual anomaly is not the sum of the 4 seasons but the mean of the 4 seasons.

Figure S7: Suggest to repeat and update the caption from S6 for convenience of the reader.

Done.

Figure S8 Mention that the scale is non-linear. The red solid line is not visible for me on this plot.

Once again, the colour scale is linear but not symmetric between positive and negative values. There was an error in the previous Fig. S8 caption (there no red line). We no longer refer to this line in the new version of the manuscript.

Figure S10: I would e.g. call this a difference not an anomaly. I think it should be a reference to Eq. 2 instead.

Plot and caption have been changed as suggested.

Generally throughout the manuscript. I would use "differences" between different experiments and "anomalies" within the same experiment or when a differences reveals an anomalous signal. "Errors" can be used when a clear target reference exists.

Thank you for the comment. This has been changed throughout the revised paper.

Figure S11: Consider plotting an updated velocity field with full coverage:

JOUGHIN, I., SMITH, B., & HOWAT, I. (2018). A complete map of Greenland ice velocity derived from satellite data collected over 20 years. Journal of Glaciology, 64(243), 1-11. doi:10.1017/jog.2017.73

Thank you for the new reference. We have updated the velocity field.

Assessment of the Greenland ice sheet - atmosphere feedbacks for the next century with a regional atmospheric model coupled to an ice sheet model

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Abstract. In the context of global warming, a growing attention is paid to the evolution of the Greenland ice sheet (GrIS) and its contribution to sea-level rise at the centennial time scale. Atmosphere-GrIS interactions, such as the temperature-elevation and the albedo feedbacks have the potential to modify the surface energy balance and thus to impact the GrIS surface mass balance (SMB). In turn, changes in the geometrical features of the ice sheet may alter both the climate and the ice dynamics governing the ice sheet evolution. However, changes in ice sheet geometry are generally not explicitly accounted for when simulating atmospheric changes over the Greenland ice sheet in the future. To account for ice sheet-climate interactions, we developed the first two-way synchronously coupled model between a regional atmospheric model (MAR) and a 3D ~~ice-sheet~~ ice sheet model (GRISLI). Using this novel model, we simulate the ice sheet evolution from 2000 to 2150 under a prolonged RCP8.5 scenario. Changes in surface elevation and ~~ice-sheet~~ ice sheet extent simulated by GRISLI have a direct impact on the climate simulated by MAR~~, and~~. They are fed to MAR from 2020 onwards, i.e. when changes in SMB produce significant topography changes in GRISLI. We further assess the importance of the atmosphere-ice sheet feedbacks through the comparison of the two-way coupled experiment with two other simulations based on simpler coupling strategies: i) a one-way coupling with no consideration of any change in ice sheet geometry; ii) an alternative one-way coupling in which the elevation changes feedbacks are parameterised in the ~~ice-sheet~~ ice sheet model (from 2020 onwards) without taking into account the changes in ~~ice-sheet~~ ice sheet topography in the atmospheric model. The two-way coupled experiment simulates an important increase in surface melt below 2000 m of elevation resulting in an important SMB reduction by 2150 and a shift of the equilibrium line towards elevations as high as 2500 m despite a slight increase in SMB over the central plateau due to enhanced snowfall. In relation with these SMB changes, modifications of ~~ice-sheet geometry tend to~~ ice sheet geometry favour ice flux convergence towards the margins, with an increase in ice velocities in the GrIS interior due to increased surface slopes and a decrease in ice velocities at the margins due to decreasing ice thickness. This convergence ~~tends to reduce~~ counteracts the SMB signal in these areas. In the two-way coupling, the SMB is also influenced by changes in fine scale atmospheric dynamical processes, such as the increase in katabatic winds from central to marginal regions induced by increased surface slopes. Altogether, the

GrIS contribution to ~~sea level riseinferred from ice volume variations~~ sea-level rise, inferred from variations in ice volume above floatation, is equal to 20.4 cm in 2150. ~~Our results suggest~~ The comparison between the coupled and the two uncoupled experiments suggests that the effect of the different feedbacks is amplified over time with the most important feedbacks being the SMB-elevation feedbacks. As a result, the experiment with parameterised SMB-elevation feedback provides a ~~sea-level~~ sea- level contribution from GrIS in 2150 only 2.5 % lower than the two-way coupled experiment, while the experiment with no feedback is 9.3 % lower. The change in the ablation area in the two-way coupled experiment is much larger than those provided by the two simplest methods, with an underestimation of 11.7 % (resp. 14 %) with parameterised feedbacks (resp. no feedback). In addition, we quantify that computing the GrIS contribution to sea level rise from SMB changes only over a fixed ~~ice-sheet~~ ice sheet mask leads to an overestimation of ice loss of at least 6 % compared to the use of a time variable ~~ice-sheet~~ ice sheet mask. Finally, our results suggest that ice loss estimations are diverging when using the different coupling strategies, with differences from the two-way method becoming significant at the end of the 21^{st} century. In particular, even if, averaged over the whole GrIS, the climatic and ~~ice-sheet~~ ice sheet fields are relatively similar, at the local and regional scale there are important differences, highlighting the importance of correctly representing the interactions when interested in basin scale changes.

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~~ Arctic ice caps and the Greenland ice sheet (Rignot et al., 2011). The evolution of the Greenland ice sheet (GrIS) is governed by variations of ice dynamics and surface mass balance (SMB), the latter being defined as the difference between snow accumulation, further transformed into ice, and ablation processes (i.e. surface melting and sublimation). While surface melting strongly depends on the surface energy balance, snowfall is primarily controlled by atmospheric conditions (wind, humidity content, cloudiness…). However, various feedbacks between the atmosphere and the GrIS ~~may lead to SMB variations that can therefore directly affect the GrIS total mass by impacting its surface characteristics ,~~ impact the surface characteristics such as ice extent and thickness~~, with~~ . This has potential consequences on ice dynamics (e.g., due to changes in surface slopes) and may lead to SMB variations that can therefore affect the total ice mass. These changes may in turn alter both local and global climate. As an example, changes in near-surface temperature and surface energy balance may occur in response to changes in orography (temperature-elevation feedback) or changes in ice-covered area (planetary albedo feedback; see ~~Vizcaíno et al. (2008, 2015) and Lunt et al. 2004~~ Lunt et al. 2004 and Vizcaíno et al. (2008, 2015)). On the other hand, topography changes may alter the atmospheric circulation patterns (Doyle and Shapiro, 1999; Petersen et al., 2003; Moore and Renfrew, 2005) causing changes in heat and humidity transports.

Quantifying the balance between these different processes and feedbacks that regulate transient ice sheet change is required to understand and ~~predict~~ project 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 ~~(Vizcaíno et al., 2015; Edwards et al., 2014a, b; Alley and Joughin, 2012; Huybrechts et al., 2002)~~(Huybrechts et al., 2 , only few global or regional models have taken the GrIS topography changes into account to compute the future evolution of the SMB and energy budget over the GrIS. For example, the CMIP5 climate models (Taylor et al., 2012) unanimously represent the ice sheet component with a fixed and constant topography, even under a warm transient climate forcing.

To explore the importance of SMB-elevation feedbacks for the future GrIS evolution, Vizcaíno et al. (2015) used a coarse resolution atmosphere-ocean general circulation model (AOGCM, ECHAM5.2) coupled to an ~~ice-sheet~~ ice sheet model (ISM, SICOPOLIS3.0) forced under different RCP scenarios (up to 2100) and their extensions (from 2100 to 2300). Compared to a control experiment in which the ISM is forced off-line by the atmospheric model run with the fixed present-day GrIS topography, they found an amplification of ice mass loss of 8–11 % and 24–31 % in 2100 and 2300 respectively, when the elevation feedbacks are taken into account (i.e. in the coupled experiment). This results from the combination of the positive elevation-SMB feedback in low lying areas, the negative feedback related to the elevation-desertification effect in accumulation areas, and the changes of surface slopes resulting from high mass loss in ablation areas and slight snowfall increase in the accumulation zone, enhancing the ice transport from the central regions to the ice margins. Their study is focused on the added value of incorporating the coupled processes. However, as specified in Vizcaíno et al. (2015), their model is not able to accurately reproduce the observed GrIS because 1/ the ~~ice-sheet~~ ice sheet model, based on the shallow-ice approximation (Hutter, 1983), is not designed to properly represent fast ice flows in outlet glaciers and 2/ the resolution of the AGCM ($\sim$3.75°) and the ~~ice-sheet~~ ice sheet model (10 km) are too coarse to correctly capture the steep slopes at the ice margins and the atmospheric processes acting on the SMB calculation.

Using the AGCM NCAR-CAM3 run at different spatial resolutions (T21 to T85) and coupled to the SICOPOLIS ~~ice-sheet~~ ice sheet model, Lofverstrom and Liakka (2018) investigated how the atmospheric model resolution influences the simulated ice sheets at the Last Glacial Maximum. They found that the North American and the Eurasian ice sheets were properly reproduced with the only T85 run. According to the authors, this is likely due to the inability of the atmospheric model to properly capture the temperature and precipitation fields (used to compute the SMB) at lower horizontal resolutions, as a consequence of the poorly resolved planetary waves and smooth topography. However, running high resolution atmospheric models at the global scale requires large computing resources. To circumvent the low resolution, some authors have used the method of elevation classes and are therefore able to offer high resolution in the direction of the slope gradient (e.g., Vizcaíno et al., 2013).

An alternative solution consists in using regional climate models (RCM) to produce high resolution atmospheric fields and much more robust energy balance and SMB calculations. A number of RCMs have been developed for the polar regions such as MAR (Fettweis et al., 2017), RACMO2 (Noël et al., 2015), Polar MM5 (Box, 2013) or HIRHAM5 (Langen et al., 2015). However, the highest resolution of the RCMs is limited by the use of the hydrostatic approximation and often remains below the resolution of Greenland ~~ice-sheet~~ ice sheet models which are generally running at a 5-10 km scale ~~(Bueler and Brown, 2009; Price et al., 2011; Greve et al., 2011)~~ (Bueler and Brown, 2009; Greve et al., 2011; Price et al., 2011) or even below (e.g., Gagliardini et al., 2013). This means that SMB fields must be corrected for resolution (and thus for elevation) differences between the RCM and the ISM. With the aim of investigating the influence of the MAR resolution on the computed SMB fields, Franco et al. (2012) developed a method to downscale each SMB MAR ~~component~~ components (snowfall, rainfall, runoff, sublimation and evaporation) onto a finer grid as a function of elevation changes. An alternative approach to correct the SMB field from surface elevation changes is based on statistical relationships between altitude and SMB ~~has also~~ (Edwards et al., 2014b). Also been derived with MAR~~(Edwards et al., 2014b)~~ , Edwards et al. (2014b) approach compute a SMB-elevation feedback gradient for regions below and above the equilibrium line altitude in the northern and southern parts of GrIS, with limited additional computing resources. However, in both parameterisations by Franco et al. (2012) and Edwards et al. (2014b), the authors only consider a strict linear relationship between topography and SMB changes. Although changes in temperature can be derived from a linear vertical lapse rate, other processes governing the SMB such as those related to energy balance, precipitation or atmospheric circulation do not follow a linear relationship with the altitude. While this approach may be valid at the local scale for small elevation changes, it may lead to a misrepresentation of the SMB-elevation feedbacks for substantial changes in altitude, especially at the ~~ice-sheet~~ ice sheet margins.

One of the first requirements to improve the representation of atmosphere-GrIS feedbacks is to use a high resolution atmospheric model ~~(Fettweis et al., 2017; Noël et al., 2015; Langen et al., 2015; Box, 2013)~~ (Box, 2013; Langen et al., 2015; Noël et al., 2015; to better represent the elevation gradients and therefore the steep topography near the ice margins in the ablation zone. Additionally, the use of a detailed snow model such as those implemented in MAR (Fettweis et al., 2017) or RACMO2 (Noël et al., 2015) allows a more accurate description of the surface properties (e.g., snow cover, albedo, surface melting) and therefore a better representation of the surface energy balance and hence of surface mass balance. RCMs developed for polar regions are also able to represent more atmospheric and land surface processes occurring in these regions such as bare ice albedo (Box et al., 2012) and katabatic winds (Ettema et al., 2010; Noël et al., 2014), being also strongly dependent on topography and thus on resolution.

The second fundamental requirement to describe the interactions between atmosphere and GrIS is to represent the ice sheet topography changes in the atmospheric model by using an ISM (instead of the fixed geometry typically used) to take into account the effects of ice dynamics on the ice sheet topography changes. This can be achieved through a numerical coupling between the RCM and the ISM. More than twenty ice sheet models exist ~~(e.g., Ritz et al., 2001; Larour et al., 2012; Fürst et al., 2013; Bueler a~~ (e.g., Ritz et al., 2001; Bueler and Brown, 2009; Larour et al., 2012; Fürst et al., 2013; de Boer et al., 2014; Pattyn, 2017), and are currently compared in the Ice Sheet Model Intercomparison Project (Nowicki et al., 2016; Goelzer et al., 2018). They represent thermo-dynamical and physical processes of the GrIS with different levels of complexities (Gagliardini et al., 2013; Saito et al., 2016). They all compute the dynamical response of the GrIS to a given climate forcing such as the SMB and the near surface temperature (ST) fields computed by RCMs (or global models). However, as SMB and ST from the climate models do not take into account the GrIS evolution, the climate forcing used by the ISM could be flawed.

In order to explicitly represent the feedbacks between the GrIS and the atmosphere and to evaluate their impacts on the ~~ice-sheet~~ ice sheet evolution, we coupled the polar regional climate model MAR (Fettweis et al., 2017) to the GRISLI ~~ice-sheet~~ ice sheet model (Ritz et al., 2001; Quiquet et al., 2012). To assess the importance of an explicit representation of the surface elevation and albedo feedbacks, this coupled experiment is then compared to a one-way coupling experiment, in which the GrIS-atmosphere interactions are not taken into account, and to a ~~second~~ third experiment where the effects of topography changes on the simulated SMB are parameterized.

A description of the atmospheric and the ~~ice-sheet~~ ice sheet models is given in Sect. 2. Sect. 3 describes the experimental setup of the three coupling methods considered in this study. The results are presented in Sect. 4. We first describe the coupled experiment in detail before comparing it to the other ~~coupling~~ uncoupled experiments. These sections are followed by a discussion related to the different coupling approaches (Sect. 5) and the conclusions of this study (Sect. 6).

2 Models

2.1 The MAR atmospheric model

MAR is a regional atmospheric model fully coupled with the land surface model SISVAT (Soil Ice Snow Vegetation Atmosphere Transfer model, see Gallée and Duynkerke 1997) which includes the detailed one-dimensional snow model Crocus (Brun et al., 1992) which simulates fluxes of mass and energy between snow layers and reproduces snow grain properties and their effect on surface albedo. MAR has been developed to simulate the GrIS SMB and has been extensively validated against in situ observations (Fettweis et al., 2017). In MAR the SMB is computed as follows:

$$SMB = SF + RF - SU - RU \tag{1}$$

Where SF, RF, SU and RU represent snowfall, rainfall, sublimation and runoff respectively. Note that RF contributes to the SMB since liquid precipitation may percolate and refreeze at depth either in the snowpack or in the ice column.

The MAR ~~horizontal resolution is~~ has a horizontal resolution of 25 km x 25 km ~~covering the~~ and 24 vertical levels to describe the atmospheric column in sigma-pressure coordinates (Gallée and Schayes, 1994). The MAR domain covers the Greenland region (6600 grid points), from 60 °W to 20 °W and from 58 °N to 81 °N~~, and 24 vertical levels to describe the atmospheric column in sigma-pressure coordinates (Gallée and Schayes, 1994)~~. SISVAT has 30 levels to represent the snowpack (with a depth of at least 20 m over the permanent ice area) and 7 levels for the soil in the tundra area.

MAR uses the solar radiation scheme of Morcrette et al. (2008). The representation of the atmospheric hydrological cycle (including a cloud microphysical model) is based on Lin et al. (1983) and Kessler (1969). ~~Each~~ To facilitate the coupling with an ice sheet model that has a higher resolution than MAR, each grid cell is assumed to be covered by at least 0.001 ~~%~~ % of tundra and ~~snow.~~ at least 0.001 % of permanent ice. This makes possible the explicit computation of ice sheet SMB outside the original ice-sheet mask. At each time step SISVAT computes the albedo of each surface type ~~and~~ as well as the characteristics of the snowpack which are weighted and averaged as a function of the snow and vegetation coverage in each grid point ~~, and then~~ before being exchanged with MAR. The present work uses MAR version 3.6. The differences with previous MAR versions ~~used in Fettweis et al. (2013, 2017)~~ (e.g., Fettweis et al., 2013) are only related to adjustments of some parameters in the representation of cloudiness ~~(i.e. cloud life time)~~ and bare ice albedo~~(i. e. parameterisation of the~~. These new parameterisations allow to better account for the positive feedback that cloud cover exerts on surface melting (Van Tricht et al., 2016) and to represent the impact of melt ponds ~~).~~ that strongly reduce surface albedo (Alexander et al., 2014)

At its lateral boundaries, MAR is forced with 6-hourly atmospheric fields (temperature, humidity, wind and surface pressure) and surface oceanic conditions (sea surface temperature and sea ice extent) provided either by reanalysis dataset (such as ERAinterim or NCEP) to reconstruct the recent GrIS climate ~~(1900-2015) (Fettweis et al., 2017)~~ (1900-2015, Fettweis et al., 2017) or by general circulation models (GCMs) to perform future projections such as those used for the last IPCC report (e.g., Fettweis et al., 2013). As a result, the atmospheric circulation simulated by MAR over the Greenland ice sheet is strongly dependent on the quality of the climatic fields computed by GCMs or reanalyses as an input to the model. Fettweis et al. (2013) have shown that GCMs which satisfactorily simulate the present-day free-atmosphere mean summer temperature at 700 hPa and the large-scale circulation over Greenland at 500 hPa are best suited to force MAR. For the present study we therefore choose to force MAR with the MIROC5 model outputs (Watanabe et al., 2010) because it has been shown to be the best GCM choice from the CMIP5 database to reproduce the present-day climate compared to the results of MAR forced by reanalyses (Fettweis et al., 2013). The greenhouse gas forcing used in MAR (scenario RCP8.5) is the same as that used in the MIROC5 simulation (Watanabe et al., 2010). Except for the experiment presented later in this study in which MAR is coupled to an ~~ice-sheet~~ ice sheet model, the topography of the GrIS as well as the surface types (ocean, tundra and permanent ice) are taken from the Bamber et al. (2013) dataset aggregated on the 25 km grid.

~~Before starting our experiments, MAR needs to be properly initialised to limit unwanted trends in the results. The snowpack model implemented in SISVAT requires less than 7 years~~ Because the snowpack in the land model requires generally longer time scale than MAR to reach an equilibrium with ~~atmospheric fields. Here, MAR is initialised with the atmospheric forcing fields~~ the atmospheric forcing, here MAR is spun-up for 6 years forced at its lateral boundaries by outputs from MIROC5 from 1970 until 1975 and ~~the MAR simulations start in 1976.~~ by an initialised snowpack coming from a previous MAR simulation carried out under present-day conditions (1960-1999). However, in this paper, the MAR results will be analysed for the period spanning from years 2000 to 2150.

2.2 The GRISLI Ice sheet model

2.2.1 Model description

The GRISLI (GRenoble Ice Shelf and Land Ice) ~~ice-sheet~~ ice sheet model was first developed to compute the dynamical evolution of the Antarctic ice sheet ~~(Ritz et al., 2001; Philippon et al., 2006; Alvarez-Solas et al., 2011a)~~(Ritz et al., 2001; Philippon et al., 2006; . It has then been successfully applied to the northern hemisphere ice sheets (Peyaud et al., 2007; Alvarez-Solas et al., 2011b; Charbit et al., 2013) and the Greenland ice sheet ~~(Quiquet et al., 2013, 2012)~~(Quiquet et al., 2012, 2013). In the present work, we use a 5 km resolution grid covering Greenland (301 x 561 grid points) and 21 evenly spaced vertical levels in the ice and 4 levels in the bedrock. The regular grid is projected on a polar stereographic grid with a standard parallel at 71 °N and a central meridian at 39 °W (same as in Bamber et al. (2013)). GRISLI is a three-dimensional thermo-mechanically coupled ISM computing the temporal evolution of the ice sheet, which is a function of ~~the~~ surface mass balance, ice flow and basal melting:

$$\frac{\partial H}{\partial t} = -\nabla(U^G H) + SMB - b_{melt} \quad (2)$$

where t is time, H the ice thickness, U^G the vertically-averaged velocity, $\nabla(U^G H)$ the ice flux divergence, SMB the surface mass balance and b_{melt} the basal melting. Basal melting occurs when the basal temperature is at the pressure melting point.

The ice temperature plays a crucial role in the dynamics of the ice sheet because it ~~also~~ affects the viscosity, and thus the ice flow in the entire ice column (Ritz et al., 1996, 2001). In turn, heat released by internal ice deformation and basal dragging over the bedrock modifies the temperature. The temperature field is computed by solving a time-dependent heat equation both in the ice and in the bedrock accounting for advection and vertical diffusion processes. At the surface, the boundary condition is provided by the prescribed surface temperature. At the base of the ice sheet, the boundary condition is given either by the geothermal heat flux or by the temperature melting point at the ice-bed interface.

The ice flow is computed using both the shallow ice (Hutter, 1983) and shallow shelf (MacAyeal, 1989) approximations to solve the Stokes equations (Ritz et al., 2001). The ~~SIA (~~shallow ice approximation (SIA) assumes that ice flow is caused only by vertical shear stress, neglecting the longitudinal stresses. This assumption is only valid for slow flowing ice. For fast flowing regions, vertical shearing becomes smaller than longitudinal shearing and the ~~SSA (~~shallow-shelf approximation (SSA), which neglects the vertical stresses, is used. The ice thickness and the ~~ice-sheet~~ ice sheet surface slopes control the SIA and the SSA velocity components, but the SSA is also governed by basal dragging. Using a hybrid model (i.e. based on both SIA and SSA approximations) allows to better represent the different deformation regimes found in an ice sheet. In GRISLI, the SSA velocity is used as a sliding velocity (Bueler and Brown, 2009) when the basal temperature is at the pressure melting point. In this case, we assume here a power-law basal friction (Weertman, 1957) and the presence of sediments allowing for viscous deformation. The relationship between the basal shear stress (τ_b) and the basal velocity (u_b) is expressed as:

$$\tau_b = -\beta u_b \tag{3}$$

where β is a time constant but spatially variable basal drag coefficient. For cold base conditions, the sliding velocity is set to zero.

The resulting velocity for every model grid point is the addition of the SIA and SSA components. For floating ice points (ice shelves), we assume no basal drag. In addition, if the ice thickness of the floating ice shelves is below 250 m and if no neighbouring points are grounded, the point is removed and the corresponding ice mass loss is considered as a calving flux. Determination of the grounding line position is based on a floatation criterion.

The isostatic adjustment in response to ice loading changes is governed by the relaxation of the asthenosphere with a characteristic time constant of 3000 years and by the deformation of an elastic lithosphere (Le Meur and Huybrechts, 1996).

The climatic forcing is given by the mean annual SMB and the mean annual ST. Because seasonal variations of surface temperature are rapidly dampened, ST is considered as a good approximation of the bottom snowpack temperature. The initial GrIS surface and bedrock topographies come from Bamber et al. (2013) and the geothermal heat flux is taken from Fox Maule et al. (2005).

2.2.2 ~~Spin-up~~ Initialisation procedure

Due to the long time scale response of the ice sheet to a given climate forcing, a proper initialisation of the model is required before performing forward experiments. For future sea-level projections, the aim of the initialisation is to start the simulations from a present-day equilibrated ice sheet geometry as close as possible to the observed one while ensuring consistency between internal properties of the ~~ice-sheet~~ ice sheet (e.g. basal sliding velocities and vertical profile of temperature) with the climate forcing. Here ~~,~~ we use an inverse method ~~, which is fully described in Le Clec'h et al. (2018)). Based on the same basic principles as that of Pollard and DeConto (2012), our method consists in the adjustment of the spatially-varying basal drag coefficient (and thus of the basal sliding velocities , see Eq. 3) so as~~ of the basal sliding velocities in order to reduce the ~~difference between the observed and the simulated ice thickness. However, while the study by Pollard and DeConto (2012) requires long (multi-millennial) integrations for the method to converge, we use instead an iterative method of short (decadal to centennial) integrations starting from the observed ice thickness.~~

~~The~~ mismatch between the simulated GRISLI ice thickness and the observed one (Bamber et al., 2013). The method is fully described in Le Clec'h et al. (2018). Below we only remind the basic principles of ~~the procedure are the following: Initial~~ our approach.

The procedure starts from initial conditions (i.e. vertical temperature and velocity profiles~~(Gillet-Chaulet et al., 2012) as well as the initial spatial distribution of the basal drag coefficientused for the first iteration are derived from~~ , first guess of the β coefficient) coming previous GRISLI simulations carried out within the framework of the Ice2Sea project (Edwards et al., 2014a) ~~. The climatological means of~~ and from the present-day observed topography (Bamber et al., 2013). The SMB and ST ~~(1976-2005)~~ climatological means computed by MAR-MIROC5 (Fettweis et al., 2013) are used as climate forcing.

~~A 5-yr relaxation run is first carried out in order to avoid large inconsistencies between the different datasets used as climate forcings and boundary and initial conditions that are not necessarily mutually consistent. After this relaxation period, the method consists in an~~ Our initialization procedure is based on an iterative process divided in two main steps:

1/ The first step consists in the iterative adjustment of the spatially-varying basal drag coefficient ~~(and thus of the basal sliding velocities) so as to reduce the mismatch between the observed and the simulated ice thickness. The iterative process is divided in two main steps:~~

~~1. In the first step, we use the vertically-averaged velocity (U^G) simulated by GRISLI and computed from the previous time step to calculate a corrected vertically-averaged velocity field (U^{corr}) as a function of a corrective factor H^G/H^{obs} that represents the mismatch between the simulated (H^G) and the observed (H^{obs}) ice thickness :~~

$$U^{corr} = \frac{U^G \times H^G}{H^{obs}}$$

~~The mean velocity field is the sum of the sliding velocity and the velocity due to ice deformation.Assuming that the deformation velocity remains unchanged, the difference between U^{corr} and U^G is only due to changes in sliding velocities . This can be expressed as:~~

$$U^{corr}_{slid} - U^{slid} = U^{corr} - U^G$$

~~The adjusted~~ related to sliding velocities through Eq. 3. At each model time step, the ratio of modelled to observed ice thickness (i.e. the data-model mismatch) is used to adjust the sliding velocities via the basal drag coefficient~~(β_{new}) allowing to reduce~~

~~the mismatch between H^G and H^{obs} is deduced from the (β_{old}) values inferred from the previous iteration (or from initial condition for the first iteration) and from the ratio between the uncorrected and the corrected sliding velocities:~~

$$\beta_{new} = \beta_{old} \times \frac{U^{slid}}{U^{corr}_{slid}}$$

~~β_{new} is calculated for each GRISLI grid point. H^G, U^G, U^{corr}_{slid}, and β_{new} are updated every year during 20 years.~~

~~For the first iteration (after the relaxation period), the first step is skipped because there is no difference between H^G and H^{obs} and the procedure starts with the 2^{nd} step (see below).~~, while the shearing velocities (SIA velocities) remain unchanged. This adjustment is typically performed for short time periods (a few decades).

~~2. Using the new~~ 2/ After this adjustment phase, the model is let to freely evolve (second step) for a few decades to a few centuries with the last inferred basal drag coefficient~~(β_{new}) computed during the 1^{st} step, we let the model to freely evolve during 200 years.~~.

At the end of this second step, we obtain a new GrIS topography and~~new corrected velocity fields (U^{corr}) computed from the mismatchbetween the simulated ice thickness (obtained at the end of the free-evolving 200-yr simulation) and the observed one. With these new U^{corr} values,~~, thus, a new data-model mismatch, which is used to start a new cycle ~~is started~~ in which the ~~1^{st} and 2^{nd}~~ first and the second steps are repeated. ~~Each cycle uses initial conditions from the 5-year relaxed ice sheet topography~~This new cycle starts with exactly the same initial and boundary conditions as those used for the first cycle, except for the basal drag coefficient and the data-model ice thickness mismatch which are inferred from the values computed at the previous cycle. The overall process is stopped when the ice thickness root mean square error is not significantly improved from one cycle to the other. This ensures a good compromise between the reduction of the mismatch between observed and simulated ice thickness and the rapidity of the convergence of the ~~spin-up method~~initialisation procedure. In the present paper, the ~~number of cycles that provides the~~ best fit with observations (RMSE = ~~+~~63 m) is ~~Nb_{cycle} = 8.~~ obtained for a first step duration of 20 years, a duration of 200 years for the relaxation GRISLI simulation (second step) and 8 iterative cycles.

This method is based on the same basic principles as that of Pollard and DeConto (2012) except that their basal drag coefficient is adjusted as a function of the difference between modelled and observed ice surface elevation while we use the ice thickness ratio instead. Moreover, while the method suggested by Pollard and DeConto (2012) requires long (multi-millennial) integrations for the method to converge, we use an iterative method of short (decadal to centennial) integrations starting from the observed ice thickness allowing a more rapid convergence.

To further reduce the model drift in terms of ice volume and before starting the forward GRISLI experiments, a 2000-yr GRISLI relaxation run is performed ~~after the last step 2 (~~as a continuation of the second step of the last iteration (i.e. after the end of the 8^{th} cycle)~~under boundary and initial conditions identical to those of the last step 2.~~. As such, the value of the basal drag coefficient is that obtained at the end of the ~~1^{st}~~ first step of the 8^{th} cycle. Over the last 150 years of this free-evolving simulation, the model drift is only $\pm$ 10^{-5} mm yr^{-1} sea-level equivalent. At the end of this 2000-yr simulation, the simulated GrIS topography is slightly different from the observations (RMSE = ~~+~~132 m, see Fig. S1). It will be referred hereafter to as S_{ctrl} and will be used as initial topography for the transient GRISLI simulations described in the following.

For all the simulations presented in this study, including those carried out within the ~~spin-up~~ iterative initialisation framework, we apply a strong negative SMB value outside the observed ~~ice-sheet~~ ice sheet extent (Bamber et al., 2013). This avoids ice growth ~~over actual present-day ice-free regions~~ where there is none in reality and allows to correct for both the potential atmospheric model biases (e.g., positive SMB values over tundra areas) and the ~~spin-up~~ initialisation procedure biases (i.e. too strong ice export towards the margins). ~~For GrIS projections , the impact of this condition is quite negligible since GrIS~~ However, for GrIS projections run the RCP8.5 forcing scenario, this condition has only a limited impact on GrIS contribution to sea-level rise since the ice extent will likely to keep on retreating over the next centuries.

3 Coupling methods

The aim of this study is to assess to what extent accounting for the atmosphere-GrIS interactions influences the GrIS evolution in terms of changes in SMB, ST, ice thickness and SLR. To achieve this goal, we designed three experiments based on coupling methods of different complexities to account for the interactions between MAR and GRISLI. For all the experiments described below, the climatic forcing is designed as follows: MAR is forced at its lateral boundaries by transient MIROC5 atmospheric fields from the CMIP5 historical run (~~1970-2005~~1970—2005) and RCP8.5 scenario (~~2006-2100~~2006—2100). In order to extend the MAR simulation until ~~AD~~ 2150 and in the absence of a MIROC5 simulation performed under a prolonged RCP8.5 scenario (i.e. after 2100), MAR is forced from 2101 to 2150 with the 2095-yr MIROC5 climate. We chose the year 2095 because, averaged over the entire GrIS, the 2095 mean climate is one of the closest to the decadal ~~2090-2100~~ 2090—2100 one. This implies that both climate changes and large-scale inter-annual variability are neglected beyond 2100.

3.1 The No Feedback experiment

The first one (referred hereafter to as NF) is based on a one-way coupling approach in which GRISLI is forced by the climatic outputs (SMB and ST) obtained from the MAR simulation spanning from 2000 to ~~AD~~ 2150. The aim of this experiment is to examine the ~~ice-sheet~~ ice sheet response to the climatic forcing without accounting for the feedbacks related to GrIS changes. ~~The SMB and ST time series simulated by MAR from 2000 to AD 2150 are used to force the GRISLI ice-sheet model~~This means that, while the ice sheet mask and topography evolve freely in the ISM, MAR is run throughout the simulation on a fixed observed ice sheet geometry and does not see any changes in the ice sheet mask. Using an inverse distance weighting method, ~~they~~ the SMB and the ST are first interpolated on the GRISLI grid to account for the difference of resolution between both models (25 km vs 5 km). To account for the differences in surface elevations between the 25 km and 5 km Bamber et al. (2013) topographies, we also apply a vertical correction following Franco et al. (2012) who derived a local vertical gradient ~~of each SMB component~~ for the SMB as a function of altitude. Thus, this method allows to generate a 5 km resolution SMB entirely ~~consistent with the~~ adapted to the finer scale features of the 5 km Bamber et al. (2013) topography. While this procedure can be followed at the daily time scale (Noël et al., 2016), in the present study, the vertical gradients are averaged at the annual time ~~sale~~ scale and used as corrective factors to downscale ~~at the end of each model year the~~ the SMB and ST fields onto the 5 km grid ~~.~~ at the end of each model year. Note that while the topography and the ice sheet mask evolve freely in the ISM, MAR uses a fixed ice mask deduced from Bamber et al. (2013) in the NF experiment.

3.2 The Parameterised Feedbacks experiment

In the second experiment (referred to as PF in the following), the SMB and ST fields simulated by MAR are corrected each year following the method of Franco et al. (2012) to account for the evolution of the simulated GRISLI topography. This correction is made from 2020 onwards, as changes in SMB through ~~2006-2020~~ 2006—2020 do not produce any significant topography changes in GRISLI. The new corrected SMB and ST values are computed at the altitude S(t) defined on the 5 km grid as:

$$S(t) = S_{Bamber} + \triangle S_{GRISLI}(t) \tag{4}$$

where S_{Bamber} is the present-day observed topography defined on the 5 km GRISLI grid and $\triangle S_{GRISLI}(t)$ the difference between the altitude simulated by GRISLI at time t and S_{ctrl}. Due to the topography differences between MAR and GRISLI, this approach has been chosen to avoid large inconsistencies between the SMB and ST fields computed by MAR and the ones corrected to account for the GRISLI topography.

This method offers the possibility to account artificially for the elevation feedbacks when using existing RCM simulations in which the topography ~~is~~ and the ice sheet mask are kept constant. As such, it is also transferable to any ice sheet model. However, ~~in this approach~~as in the NF experiment, the changes in GrIS geometry have no consequence on the climate as simulated by the atmospheric model.

3.3 The two-way coupling experiment

The third method (2W in the following) is based on a two-way coupling strategy between MAR and GRISLI. Both models ~~use~~ used the same boundary and initial conditions as those of the NF and PF experiments. At the end of a MAR model year, MAR is paused and GRISLI is forced by the downscaled SMB and ST fields with the method of Franco et al. (2012) as in PF (Eq. ~~??~~4). Then, GRISLI computes a new 5 km GrIS topography and a new ice extent at a 5 km ~~which are aggregated~~ resolution. This new GrIS topography is then aggregated (i.e. geographically averaged) at the yearly time scale onto the 25 km MAR grid. The ~~aggregated ice extent~~ number of ice covered GRISLI grid points within a MAR grid cell relative to the number of ice-free GRISLI grid points is used to compute the new ice extent in MAR and to update the fraction of tundra relative to ice/snow covered surface type for the subsequent MAR run. To account for the differences between MAR and GRISLI topographies, the surface elevation which is aggregated onto MAR is computed from GRISLI surface elevation anomalies added to the present-day observed topography (Eq. ~~??~~4). It is then used as the updated surface elevation~~in MAR~~. As previously mentioned, topography changes are negligible before 2020. Hence, changes in ~~ice-sheet~~ ice sheet geometry are fed to MAR only after this date. Compared to the NF and PF approaches, this two-way coupled method is the most accurate to represent the GrIS-atmosphere feedbacks.

4 Results

4.1 The Greenland ice sheet evolution in the 2W experiment

4.1.1 Changes in the forcing climate

The evolution of the SMB and of its different components simulated in the 2W experiment (Eq. 1) ~~,~~ and integrated over the entire GrIS ~~and simulated in the 2W experiment~~ is displayed in Fig. 1. During the ~~2000-2040~~ 2000—2040 period, the averaged SMB remains positive with a mean value equal to 280 $\pm$ 95 Gt yr^{-1} (where the notation $\pm$ represents the standard deviation computed from yearly values) but slightly decreases by 4 Gt yr^{-1}. This decrease becomes substantially stronger from 2040 to 2100 (-17 Gt yr^{-1} on average), and the mean SMB reaches strong negative values (-638 $\pm$ 271 Gt yr^{-1} over the ~~2090-2100~~ 2090—2100 period). As the same MIROC5 year 2095 is repeatedly used to force MAR after 2100, there is no longer inter-annual variability and the integrated SMB remains quite stable between 2100 and 2150 (-812 $\pm$ 13 Gt yr^{-1})~~, despite a slight~~ . Indeed, MAR generates its own surface boundary layer fields which are not impacted by the MIROC5 forcing, explaining a slight SMB increase of $\sim$1 Gt yr^{-1} over the last 50 years. Throughout the simulation, the evolution of the SMB signal is dominated by surface runoff whose increase rate (in absolute value) ranges from 5 Gt yr^{-1} (~~2000-2040~~2000—2040) to 19 Gt yr^{-1} (~~2040-2100~~2040—2100). After 2100, it slightly decreases ($\sim$2 Gt yr^{-1}), explaining the slight SMB increase.

The SMB anomaly between the beginning and the end of the 2W experiment ~~exhibits two distinct patterns (~~is displayed in Fig. 2a~~)~~. 65 % of the grid points having surface elevations higher than 2000 m are characterized by a positive SMB anomaly, ranging from 0.07 m yr^{-1} (5^{th} percentile) to 0.2 m yr^{-1} (95^{th} percentile) at the end of the simulation. This SMB increase is particularly pronounced in the eastern part of the GrIS between 67 and 70 °N and in the north central part. It is due to a strong increase in snowfall (> 0.5 m yr^{-1}, Fig. 2b) which occurs mainly during the winter season in the east and during autumn in the north (Fig. S2). On the other hand, 87 % of the GrIS grid points with surface elevation lower than 2000 m are dominated by an increase in surface runoff (Figs. 2c, S3) and by an increase in the fraction of rainfall over snowfall in summer and in autumn (Fig. S4). As a result, strong negative SMB anomalies are found in these regions ranging from -3.3 m yr^{-1} (5^{th} percentile) to -0.1 m yr^{-1} (95^{th} percentile) and reaching more than -6 m yr^{-1} along the western and the southeastern margins (Fig. 2a).

The equilibrium line altitude (ELA, i.e. altitude for which SMB = 0) increases significantly between the beginning and the end of the 2W experiment, as a consequence of increased runoff for areas below 2000 m. As an example, at around 73.5 °N, on the eastern side of the ice sheet, the ELA moves from $\sim$1000 m to $\sim$2500 m (Fig. 3). In other regions, at the end of the 2W experiment, the ELA is generally situated between 1500 and 2000 m high, except in the northern part where it is between 1000 and 1500 m. This shift of ELA towards higher altitudes represents an increase of 24 % of the ablation area between the beginning and the end of the experiment.

The ST anomaly (Fig. 2d) ranges from ~~+~~2.2 °C (~~5th~~ 5^{th} percentile) to ~~+~~6.5 °C (~~95th~~ 95^{th} percentile) and is characterized by a south-north gradient with the highest values found in the northern part. Beyond 78 °N the ST anomaly reaches locally values greater than ~~+~~11 °C. This temperature increase from 2000 to 2150 contributes to the amplification of the ablation processes below the ELA. However, while the stronger temperature anomaly is found in the northeastern part of the ice sheet, ~~ablation~~

~~processes are modulated~~ this region is aslo marked by the increasing snowfall in 2150 compared to ~~2000.~~ 2000 (Fig. 2b) which counteracts the ablation processes.

4.1.2 Changes in Greenland ~~ice-sheet~~ ice sheet geometry

~~The~~ Figure 4 displays the ice thickness anomaly ~~(Fig. 4) also presents two distinct patterns.~~ between the beginning and the end of the 2W experiment. For surface elevations higher than 2000 m in the northern part, and higher than 2500 m in the central and southern parts of the ice sheet, the ice thickness increases by 5 m on average, with the increase ranging from 1.5 m (5^{th} percentile) to 17 m (95^{th} percentile). On the other hand, in regions whose surface elevation is lower than 2000 m, the ice thickness decreases from -248 m (5^{th} percentile) to -3 m (95^{th} percentile) with a mean value equal to -100 m. As a result of these GrIS ice thickness changes, the surface slope between the central part of the ice sheet and the margins increases. On top of that the ice sheet mask (defined as the fraction of a MAR grid cell with permanent ice cover, ~~Fig. S5~~e.g. Fig. S5a) decreases by 2.8 $\pm$ 0.1 % (mean $\pm$ standard deviation computed from yearly values) over the ~~2140-2150~~ 2140—2150 mean period compared to the ~~2000-2010~~ 2000—2010 mean period, and some GrIS margin regions become ice free~~(red grid points in Fig. S5).~~ .

4.1.3 ~~Changes in~~ Impact of ice ~~dynamics~~flow changes

The ice thickness anomaly is due to the complex combination of changes in surface atmospheric conditions (SMB, Fig. 5a), ice ~~dynamics~~ flow (ice flux divergence, Fig. 5b) and basal melting (not shown), following the continuity equation (Eq. 2). To quantify the role of ice ~~dynamics~~ flow on the GrIS geometry (Fig. 4), we plotted the ice flux divergence integrated over 150 years (~~2000-2150~~2000—2150, see Fig. 5b). In particular, over the central plateau, the cumulated SMB (Fig. 5a) reaches about ~~+~~50 m, 40 m of which are transported away by the ice ~~dynamics~~ flow (Fig. 5b). As a result, the ice thickness anomaly is reduced to only ∼10 m in this region (Fig. 4). An opposite behaviour is found near the western coast, where the ~~ice melting~~ runoff is partly compensated by ice convergence, resulting in a less negative ice thickness anomaly than that related to the SMB forcing. This shows that ice ~~dynamics~~ flow act to counteract ice loss from surface melting, as previously noticed by several authors (Huybrechts and de Wolde, 1999; Goelzer et al., 2013; Edwards et al., 2014a). As a consequence, it appears to be essential to account for ice dynamics to estimate accurately the mass balance of the whole ice sheet.

In turn ice ~~dynamics~~ flow is impacted by changes in ice sheet geometry as illustrated by the mean surface velocity anomaly (Fig. 6a). For regions with surface altitudes between 2000 and 2500 m, the anomaly of the ice flow increases from the inner GrIS areas towards the edges of the ice sheet. The increase in the mean ice flow for the ~~2140-2150~~ 2140—2150 period compared to ~~2000-2010~~ 2000—2010 period, ranges from 0.08 m yr^{-1} (5^{th} percentile) to 17 m yr^{-1} (95^{th} percentile). These faster ice velocities at the end of the 2W experiment are mainly explained by ~~the combination of~~ a larger surface slope between the central and the margin regions of the ice sheet. This is consistent with information inferred from ice flux divergence as shown in Fig. 5b.

On the contrary, for the margin regions, with altitudes lower than 1500 m, the anomalies of surface ice velocities strongly decrease (Fig. 6a). Compared to the ~~2000-2010~~ 2000—2010 period, this decrease ranges from -213 m yr^{-1} (5^{th} percentile) to -0.2 m yr^{-1} (95^{th} percentile), and ~~is fully consistent~~ agrees with the decrease in ice thickness (Fig. 4).

The changes in local ice ~~dynamics~~ flow between the first and the last 10 years of the 2W experiment are also related to changes in surface slope and ice thickness, particularly at the margins. To investigate the ice ~~dynamics~~ flow changes at the local scale, we used the examples of the Jakobshavn (western coast) and the Kangerlussuaq (eastern coast) glaciers for which the fine scale structures of the ice velocity, obtained after the GRISLI initialisation procedure, are relatively well reproduced compared to the observations (Figs. 7a-b).

For the Jakobshavn glacier, ~~and for altitudes~~ the ice sheet areas located above 1500 m, ~~the vertically-averaged ice velocities increase by~~ are mainly characterised by an increase of more than 15 m yr^{-1} (i.e. 10 %) of the vertically-averaged velocity as a result of increasing surface slopes ~~, and slow down by~~ (Fig. 7c). Conversely, areas below 1000 m are dominated by a slow down of the ice flow of more than 200 m yr^{-1} (i.e. 29 %) ~~for altitudes below 1000 m~~ due to the decreasing ice thickness (Fig. 7~~e~~e). For altitudes above 500 m, the vertically-averaged velocity is mainly driven by the SIA velocity (Figs. 7~~c-e~~e-g). On the contrary, below 500 m, basal sliding velocities are large due to low basal drag coefficient (see Fig. 3 in Le Clec'h et al. 2018) and the SSA velocity component dominates the ice flow (Figs. 7~~e-g~~e-i). However, while basal drag is lower in locations below 500 m, the ice flow is limited by the strongly reduced ice thickness (Fig. 4).

The Kangerlussuaq glacier is located in regions where the bedrock is characterised by a succession of valleys surrounded by mountains merging in a canyon where the deepest part is located 100 km away from the coast (Morlighem et al., 2017). The ice flow of the Kangerlussuaq is therefore divided in different branches with increasing ice velocities towards the ice sheet margin and becoming even larger when merging in the canyon (Fig. 7~~b~~d). As for the Jakobshavn glacier, the ice flow accelerates at the end of the 2W experiment as a consequence of the increase in surface slope for high altitudes (∼~~2000-2500~~ 2000—2500 m, see Fig. 4). Conversely, a strong decrease of the ice flow is found in most of margin regions (Fig. 7~~d~~f) directly related to the ice thinning (Fig. 4). Contrary to the case of the Jakobshavn glacier that presents large basal sliding velocities only below 500 m, the Kangerlussuaq shows low basal drag coefficients in the entire glacier (see Fig. 3 in Le Clec'h et al. 2018) and thus the ice flow is mainly governed by the SSA component (Fig. 7~~h~~j).

These results are in line with Peano et al. (2017) who also found a decreasing ice flux at the end of the 21^{st} century (w.r.t 1970) in downstream regions of the Jakobshavn and Kangerlussuaq glaciers as a consequence of ~~ice-sheet~~ ice sheet thinning at the margins.

4.2 Differences between the ~~2W~~ NF and the ~~NF~~ 2W experiments.

4.2.1 Impact on SMB and ST

To assess the importance of the atmosphere-GrIS feedbacks, we now compare the ~~2W and the NF~~ NF and the 2W experiments. The main SMB differences between both experiments, averaged over the ~~2140-2150~~ 2140—2150 period, highlight ~~lower~~ higher SMB values in ~~2W compared to NF~~ NF compared to 2W for altitudes below 2000 m, with the exception of some margin locations in the eastern part (Fig. 8a). This SMB ~~anomaly behaviour~~ difference is driven by a snowfall ~~reduction~~ increase in low altitude areas (Fig. S6) and by the runoff ~~increase in 2W~~ decrease in NF with respect to ~~NF~~ 2W (Fig. S7). This ~~increased~~ decreased runoff results from ~~warmer~~ colder temperatures over the whole GrIS (up to ~~0.8~~ -1 °C in the western and northern parts, Fig. 8b), except in the ~~region at the~~ regions at the very edge of the GrIS, which sees a ~~strong cooling (as low as -10~~ significant warming (up to 8 °C, Fig. 8b) ~~. The warming~~ despite an increase in ice thickness (w.r.t 2W). The cooling can be explained by the absence of the temperature-altitude feedback ~~being active in~~ in the NF experiment. Indeed, taking this feedback into account in 2W, ~~resulting~~ results in lower altitudes as the ice thickness decreases (Sect. 4.1.2 and Fig. ~~8c~~4) and therefore ~~warmer temperatures. The cooling~~ in warmer 2W temperatures compared to NF. The warming simulated in NF (w.r.t 2W) over the very edge of the ice sheet ~~occurs despite the ice sheet thinning over these regions and~~ can be explained by changes in atmospheric circulation.

Indeed, unlike ~~NF,~~ 2W, NF allows for an explicit computation of changes in ice sheet surface slopes due to increased ~~melt~~ melting at the margin. This has important consequences on the atmospheric circulation and in particular on the katabatic winds (Fig. 9). Over the ice sheet, the ~~steeper~~ surface slopes simulated in ~~2W~~ NF in 2150 are less steep compared to those simulated in 2W (discussed in Sect. 4.1.2) lead to a slight increase in katabatic winds (Fig. 9). However, at the ice sheet margin, i.e. where the ice mask in MAR is below 100 %, there is a substantial decrease in surface winds. This ~~is because the change~~ stems from the fact that the changes in surface elevation ~~as~~ seen by the atmospheric model ~~is~~ are computed from the aggregated changes in GRISLI at 5 km. As such, a non-zero fraction of tundra, which presents no change in surface elevation, results in smaller elevation changes compared to grid ~~cell in the same region~~ cells with permanent ice cover only. This induces ~~artificially~~ in 2W lower surface slopes at the margin with respect to the interior and ~~a decrease in~~ thus weakened surface winds in these regions. Altogether, the ~~slight increase in~~ slightly stronger katabatic winds over the ice sheet and their ~~reduction~~ weakening at the margin lead to a cold air convergence towards the ice sheet edge, absent in NF (Figs. 8b, 9 and Figs. S8-S9). Another consequence of the ~~katabatic winds increase~~ stronger katabatic winds in 2W (w.r.t NF) due to increased surface slopes in the GrIS interior, is to enhance the atmospheric exchanges along the slope of the ice sheet. The area with 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. The warming of the upper part of the boundary layer in 2W combined with the lower surface elevation, explains the ST increases in the interior of the GrIS.

4.2.2 Impact on ice thickness and ice dynamics

The most important ice thickness difference between the last ten years of the ~~2W and the NF~~ NF and the 2W experiments is a ~~smaller thickness in 2W compared to NF. This~~ higher ice thickness in NF compared to 2W. As mentioned in the previous section, this is mainly explained by the positive ~~temperature-elevation~~ SMB-elevation feedback in 2W that results in increased surface temperatures compared to NF, and thus increased ~~melting~~runoff, when surface elevation decreases. Areas with this type of behaviour cover most of the Greenland ice sheet slopes and reach the interior of the ice sheet from the western or the northeastern margins. The largest changes occur over the western edge of the GrIS, where the ~~thinning between~~ thickening between NF and 2W ~~and NF~~ reaches more than ~~-25~~ 50 m (Fig. 8c). The ice thickness ~~anomaly~~ difference pattern is essentially mimicking the SMB differences between NF and 2W ~~and NF~~ (Fig. 8a), suggesting that the two-way coupling induces only a relatively limited change in ice ~~dynamics~~flow, as shown by the ice flux divergence anomaly (Fig. S10), although the surface velocities (Fig. 6b) are slightly ~~smaller in 2W due to smaller~~ higher in NF due to higher ice thickness (Fig. 8c).

4.3 The PF experiment

As previously described (see Sect. 3.2), the PF experiment is based on a parameterisation of the surface elevation feedbacks. In this section, we present the differences ~~(2W − PF)~~ PF-2W in SMB, ST and ice thickness averaged over the 2140-2150 period (Fig. 10) so as to examine the efficiency of this parameterisation. The first key feature is that the ~~2W − PF SMB anomaly~~ PF-2W SMB difference (Fig. 10a) is less ~~negative than the 2W − NF~~ positive than the NF-2W one (Fig. 8a). This results from the fact that the decreasing altitude is taken into account in PF through the altitude feedback parametersisation, leading to smaller differences with the 2W experiment. In most margin areas, the SMB simulated in PF has even become lower compared to 2W ~~.~~ due to a much complex representation of the ice sheet climate interactions. Indeed, as mentioned in see Sect. 4.2.1, in the 2W experiment, the GRISLI topography feedbacks onto the MAR simulated climate leads to a cold air convergence at the ice sheet margins and thus to a higher simulated SMB. Cumulated over the entire GrIS, the ~~2W − PF~~ PF-2W SMB difference is ~~-28~~ 28 Gt yr^{-1} (~~-149~~ 149 Gt yr^{-1} for ~~2W − NF~~NF-2W). In the same way, the ~~2W − PF~~ PF-2W differences in ST and ice thickness (Figs. 10b-c) are also less pronounced than the ~~2W − NF~~ NF-2W differences (Figs. 8b-c), highlighting the importance of the elevation feedbacks. These results show that over ∼150 years, the topography correction used in PF, ~~allows~~ makes possible to obtain from an uncoupled experiment ~~to obtain~~ simulated fields close to those of the 2W coupled experiment.

To illustrate the spatial variability of the ice thickness response to the different coupling methods, we plotted the ice thickness ~~anomalies between~~ differences between NF and 2W ~~and NF~~ (red dots, Fig. 11a), ~~PF and NF~~ NF and PF (green dots, Fig. 11a) and PF and 2W ~~and PF~~ (Fig. 11b) as a function of the ice sheet altitude. The ~~2W method yields negative and positiveanomalies relative to NF~~NF-PF ice thickness differences are mostly positive, while the ~~PF method mainly yields negative values~~ NF-2W (Fig. 11a) ~~,~~ and PF-2W (Fig. 11b) differences yield both positive and negative values, while the NF-PF differences are positive, illustrating the stronger ~~spatial variability in~~ variability in the 2W experiment. For both the 2W and PF experiments~~compared to NF~~, the regions at low to medium elevations are the most sensitive to the coupling approach with the stronger spatial variability of the ice thickness found for altitudes below 1000 m. For example, the ~~2W − NF ice thickness anomalies range between -31.9~~ NF-2W ice thickness difference ranges between 31.9 m (5th percentile) and ~~+6.5~~ -6.5 m (95th percentile), and between ~~-27.8~~ 27.8 m (5th percentile) and 0 m (95th percentile) for the ~~PF − NF~~ NF-PF case. Overall, the ice thickness ~~anomalies~~ differences decrease with increasing altitudes (Fig. 11a) and increase with time (Fig. 11b). ~~They are also stronger in the 2W experiment than in the PF and NF simulations.~~

4.4 Impact on GrIS contribution to sea-level rise and ice sheet mask

At the end of the simulation (i.e. after 150 model years), the GrIS contribution to sea-level rise (computed from the change in GRISLI ice volume), simulated in the 2W experiment, reaches 20.4 cm, against 18.5 cm and 19.9 cm in the NF and PF experiments respectively (Table 1 and Fig. 12). Owing to the negligible model drift (∼10^{-5} mm yr^{-1}, see Sect. 2.2.2), these differences only result from the better representation of the GrIS-atmosphere feedbacks in 2W leading to increased runoff due to warmer temperatures (see Sect. 4.2). In 2100 (Table **??**), the differences between the three experiments are smaller, with the NF and PF contributions being respectively 4.4 % and 0.4 % lower than the 2W contribution, against 9.3 % and 2.5 % in 2150. These results reflect several key aspects. First they show that the GrIS mass loss substantially accelerates from the second half of the 21st century onwards and that the effect of the different feedbacks, as simulated in 2W, is enhanced over time. Figure 12b displays the sea-level anomalies between 2000 and 2100 to better illustrate the divergence of the three experiments as soon as ~~2025-2030~~2025—2030. Secondly they illustrate the effect of the feedbacks ~~itself~~thenselves. As an example, accounting for the parameterised feedbacks (PF) leads to an additional SLR contribution (w.r.t NF) of 4.2 % in 2100 (7.6 % in 2150). This is smaller than that reported in Calov et al. (2018) who also used the MAR model to force the hybrid SICOPOLIS3.3 including a representation of subglacial hydrology. However, our estimate is comparable with the 4.3 % additional contribution found by Edwards et al. (2014a) in 2100 who used ECHAM5 and HadCM3 to ~~forced~~ force MAR simulations under the SRES A1B scenario and five ISM projections, and within the range of uncertainties of the 8 $\pm$ 5 % additional surface mass loss reported in Fettweis et al. (2013). As for Vizcaíno et al. (2015), they also conclude that the melt-elevation feedbacks, simulated with the ECHAM5.2-SICOPOLIS3.0 coupled model under the RCP8.5 scenario, contribute to 11 % to SMB changes and to 8 % to SLR. ~~Compared to our own study~~While the importance of the SMB-elevation feedback may be dependent on the model itself, the larger contribution found in Vizcaíno et al. (2015) ~~may~~ compared to our own study, could be explained by the coarser resolutions of ECHAM5.2 (~3.75°~~for the atmospheric component) and of~~) and SICOPOLIS3.0 (for the GrIS, 10 km) with respect to MAR and GRISLI resolutions, implying for example that ablation areas or processes such as katabatic winds are less well represented. Our results also suggest that, at the centennial time scale, the SMB-elevation feedback is the most important since its parametrization in PF allows to reduce the mismatch between the 2W and NF GrIS SLR contributions by 73.7 % (resp. 91.4 %) in 2150 (resp. 2100), the remaining contributions being attributed to ~~albedo and~~ atmospheric feedbacks. However, to assess more accurately the relative importance of the elevation feedbacks, a more appropriate procedure would be to cut off the elevation feedbacks in the 2W experiment.

Compared to the NF and the PF experiments for which the ~~ice-sheet~~ ice sheet mask is fixed to observations from 2000 to ~~AD~~ 2150, the 2W ice sheet extent is reduced by ~2.8 % in 2150 as a result of increased ablation. As MAR sees the ice sheet retreating over time in 2W concomitantly with the increase in bare ground or tundra fractions (Fig. S5b), the albedo feedback takes place favouring further the ice melting ~~.~~, though counteracted by the katabatic wind anomalies (see Sect. 4.2). Although the ice sheet retreats, the extent of the ablation zone increases with time. This process is faster in 2W than in NF and PF. In 2150, the ablation zone is 14 % (resp. 11.7 %) larger in 2W than in NF (resp. PF) causing 112 Gt yr^{-1} of extra ice ablation in 2W (w.r.t NF). As a consequence, the ELA is located further inland in 2W compared to NF with a maximum inland retreat of 120 km located in northeastern Greenland (Fig. 3).

A widely used method to estimate the ~~projected GrIS~~ GrIS contribution to global sea-level rise is to compute the GrIS mass loss as the time-integral of the SMB computed by an atmospheric model over a fixed ~~ice-sheet mask (Fettweis et al., 2013; Meyssignac et al.~~ ice sheet mask (Fettweis et al., 2013; Church et al., 2013; Meyssignac et al., 2017). In the present study, we ~~go a step further~~ use a more complex method since the ice mass variations related to SMB changes are computed by MAR over a changing

~~ice-sheet mask~~ ice sheet mask and topography as simulated by GRISLI. However, in both the NF and the PF experiments, the atmospheric model does not account for the variations in the ~~ice-sheet~~ ice sheet extent simulated in GRISLI and the ~~ice-sheet~~ ice sheet mask, taken from the observations (Bamber et al., 2013), is kept constant throughout the simulation. Taking the changes in ~~ice-sheet~~ ice sheet mask into account may have strong impacts on the computed GrIS contribution to sea-level rise. To illustrate the influence of the ice sheet mask, we used the SMB outputs from the NF experiment at the MAR resolution and applied the integrated SMB method over the fixed observed ~~ice-sheet~~ ice sheet mask (SMB$_{\cancel{MSK=NF}MSK_{NF}}$) and over the updated 2W mask (SMB$_{\cancel{MSK=2W}}$~~). Results reported in Table ?? indicate differences in SMB values exceeding~~ $_{MSK_{2W}}$). Differences in SMB values exceed 23 % in ~~2150.~~ 2150 (-842 Gt yr^{-1} for SMB$_{MSK_{NF}}$ against -647 Gt yr^{-1} for SMB$_{MSK_{2W}}$). In the same way, compared to a time variable ~~ice-sheet~~ ice sheet mask, the use of a fixed ~~ice-sheet~~ ice sheet mask overestimates the sea-level rise by ~6 % in 2150. Though a bit lower, this ~~number is far from being negligible compared to the~~ error has a similar magnitude compared to errors made when the SMB-elevation feedbacks are not taken into account (i.e. 7.6 %) and when all the feedbacks are ignored (i.e. 9.3 %). This strongly suggests that realistic SLR projections cannot neglect the evolution of the ~~ice-sheet~~ ice sheet extent, only accounted for through the use of an ~~ice-sheet~~ ice sheet model.

5 Discussion

The evolution of the GrIS and its contribution to sea-level rise presented in this study are the first ones inferred from a regional atmospheric model synchronously coupled to an ~~ice-sheet~~ ice sheet model, thus accounting for the GrIS-atmosphere feedbacks. To evaluate the added value of a coupled RCM-ISM model, we explored the importance of the GrIS-atmosphere feedbacks by comparing the results of the coupled experiment to those coming from two uncoupled experiments, PF (Parameterised elevation feedbacks) and NF (No feedback)~~and~~. We showed that the impact of taking the feedbacks into account increases over time. This study is therefore a necessary ~~first~~ step toward a more accurate assessment of the contribution of Greenland to future sea-level rise and of its impact on the climate system. However, future refinements could be envisaged.

One of the main uncertainty in assessing the GrIS contribution to future sea-level rise comes from the climate projections themselves. For example, using five different global climate models to force MAR at its lateral boundaries under RCP8.5 conditions, Fettweis et al. (2013) provide SMB-inferred estimates of this contribution ranging from 4.6 to 13.1 cm in 2100. This range is fully comparable to that reported by Calov et al. (2018) who used MAR simulations (forced by three GCMs chosen from the Fettweis et al. (2013) sample) to force the SICOPOLIS ~~ice-sheet~~ ice sheet model. Whatever the experimental design, the large spread in SLR projections highlights the great uncertainty associated with the choice of the global climate model used to force MAR at its lateral boundaries. It raises the question ~~as to whether the ice-sheet response simulated in our~~ to what extent the differences between 2W ~~experiment relative to that of the NF and PF~~ and PF or NF experiments would be ~~similar, amplified or mitigatedwith a different GCM climate forcing having a different sensitivity from~~ amplified (resp. mitigated) with a stronger (resp. weaker) climate forcing than that simulated by MIROC5. ~~A second~~ Another question concerns the impact of a constant MIROC5 climate used to force MAR beyond 2100. As outlined in Sect. 3, this results in discarding the continued change that the climate will likely undergo beyond 2100 suggesting that our SLR projections are underestimated. ~~The second~~

Another consequence is that inter-annual variability is neglected after 2100. This can lead to conservative estimates of the Greenland contribution to sea level rise in the future due to non-linearities of the SMB. On the other hand, the imprint of the 2095 MIROC5 climate may amplify regional changes of the GrIS response. There is therefore a strong need for iterating the present study with different global climate simulations run under an extended RCP8.5 scenario~~and used as a MAR forcing,~~ , but also with different regional climate models that may have different sensitivities, to assess more accurately the impact of the different GrIS-atmosphere feedbacks and to better evaluate the uncertainty associated with the projected sea-level rise contribution from the GrIS.

~~A second~~ Another limitation is related to the 2000-yr relaxation GRISLI experiment, run at the end of the ~~spin-up~~ initialisation procedure to reduce the model drift in terms of ice volume~~, that~~ . This produces residual differences with the observed topography (Bamber et al., 2013) used in the MAR simulations. This has important consequences on the MAR simulated climate. In particular, the steeper slopes existing in the GRISLI topography (i.e. S_{ctrl}) tend to produce unrealistic katabatic winds. Therefore, we ~~choose~~ have chosen in this study to use an anomaly method of the surface elevation onto which the SMB and ST fields are downscaled at the 5 km resolution grid (Eq. ~~??~~4). The objective of this approach was first to maintain the realism of the simulated present-day climate computed on the observed topography (Bamber et al., 2013) and, secondly, to avoid inconsistencies between the climate simulated by MAR and that used to force GRISLI. However, this implies that the forcing climate is not fully consistent with the GRISLI topography. This should be taken into consideration in a future work to improve the quality of our results. As an example, a reasonable compromise to avoid the use of anomaly method would be to use the topography obtained at the end of the ~~spin-up iterative~~ iterative initialisation process (rather than S_{ctrl}) as initial ~~GRISLI~~ MAR topography to keep the mismatch with the observed topography as low as possible, and to initialise and perform MAR simulations with this ~~spin-up topography.~~ new topography. Moreover, the use of an anomaly method to account for the change in topography is incompatible with a conservative coupling between the ice sheet model and the climate model. This is further amplified by the fact that we use a flux correction outside the present-day ice margin to force ice removal. This methodology has been followed to limit the impact of biases from the atmospheric model and from the initialization procedure, but the imposed ice removal outside the present-day ice mask may bias locally the model response towards increased ice thinning. Since our simulations are run under the RCP8.5 forcing scenario, this has probably a negligible impact because the GrIS is likely to experience a retreat from the present-day ice margin. However further studies with alternative scenarios and/or GCM forcing, and even more paleoclimate studies, should ideally avoid using this kind of flux correction.

In addition, difference of resolution between MAR (25 km) and GRISLI (5 km) ~~. This~~ can cause artefacts in the results, especially at the edges of the ice sheet. Indeed, in the corresponding MAR grid cells, a fraction of permanent ice cover may coexist with a non-zero fraction of tundra. Since the surface elevation changes computed in MAR from the aggregated GRISLI topography are weighted as a function of the fraction of the different surface areas, they may be underestimated as tundra soil type is not subject to any change in altitude. This artefact has been illustrated in Sect. 4.2.1 with the example of the behaviour of katabatic winds that are artificially reduced in our simulation at the ~~ice-sheet~~ ice sheet margin. Moreover, since the margin regions are those experiencing the strongest changes in altitude, they are also the most sensitive to climate change. As a consequence, an improper estimation of the topography changes may induce improper SMB changes. This underlines the need for increasing the atmospheric model resolution as far as possible to avoid such artefacts and to better represent the fine scale atmospheric-topography feedbacks impacting the SMB. Indeed, higher spatial resolution could resolve finer scale ice sheet dynamics to better represent the ice flow in outlet glacier or better represent fine scale atmospheric-topography feedbacks impacting the SMB in these regions. However, a compromise must be reached between the additional computing resources and the required degree of accuracy of sea-level projections.

Regarding the ~~ice-sheet~~ ice sheet model, a 5 km horizontal resolution does not permit to capture the complex ice flow patterns of smallest outlet glaciers, whose characteristic length scale can be less than 1 km (Aschwanden et al., 2016) and to quantify accurately the ice discharge at the marine front. This may have large implications in the sea-level rise estimates. Using a 3D ~~ice-sheet~~ ice sheet model with prescribed outlet glacier retreat, Goelzer et al. (2013) found an additional SLR contribution from outlet glaciers of 0.8 to 1.8 cm in 2100 and 1.3 to 3.8 cm in 2200, with the influence of their dynamics on SLR projections decreasing with time and with the increasing importance of the atmospheric forcing. This is in line with the fact that ice ~~dynamics~~ flow act to counteract ice loss from surface melting (see Sect. 4.2), as previously outlined by several authors ~~(Edwards et al., 2014a; Goelzer et al., 2013; Huybrechts and de Wolde, 1999)~~(Huybrechts and de Wolde, 1999; Goelzer et al., 2013; Edwa
. However, despite the possible decreasing influence of marine terminating glaciers, at the centennial time scale, it ~~seems to be preferable to~~ is essential to evaluate more accurately the impact of ice dynamics and to better capture the complex geometry of fjords surrounding the marine-terminating glaciers.

There is a growing number of evidence for attributing the acceleration of outlet glaciers to the intrusion of warm waters from adjacent oceans in the fjord systems or in the cavity of floating ice tongues ~~(e.g., Straneo et al., 2012; Johnson et al., 2011; Rignot et al., 201~~
(e.g., Johnson et al., 2011; Straneo et al., 2012; Rignot et al., 2015) that can destabilise the glacier front and/or favour the ice-shelf breakup (Gagliardini et al., 2010), decreasing thereby the buttressing effect and increasing the ice calving. In turn, the released freshwater flux in ocean may impact sea-surface temperatures, oceanic circulation and sea-ice cover. Moreover, atmosphere-ocean feedbacks also have an impact on the GrIS. As an example, Fettweis et al. (2013) showed that the disappearance of Arctic sea ice in summer induced by ocean warming enhances surface melting in northern Greenland through a decrease of surface albedo and the subsequent atmospheric warming. Thus, the absence of the oceanic component in our modelling setup appears as a limiting factor, although, the direct impact of ocean via sub-shelf melt at the ice sheet margin will likely be limited in the future as a result of inland retreat of GrIS.

Our ~~spin-up~~ initialisation method adjusts the basal drag coefficient in such a way that the departure between the observed and the initial GRISLI topographies is reduced. The resulting β coefficient is spatially varying but is taken constant in time. This assumption may likely be valid for short-term forward simulations but is probably overly simplistic. On the one hand, the basal drag tends to be smaller towards the margins with respect to the interior. As the ice sheet retreats inland, it can be expected a reduction in basal drag for a specific location, due for example to a decreasing effective pressure. On the other hand, changing basal hydrological conditions can also alter the basal drag. This can occur as a result of rainfall or surface meltwater infiltration that can refreeze at depth or propagate all the way to the bottom of the ice sheet and increase basal lubrication (Kulessa et al., 2017). Therefore, a time constant basal drag coefficient inferred under present-day conditions may underestimate the ice flow acceleration. A few models describing the vertical inflow exist

~~(e.g., Banwell et al., 2016; Clason et al., 2015; Koziol et al., 2017)~~ (e.g., Clason et al., 2015; Banwell et al., 2016; Koziol et al., 2017) but are generally run at the regional scale and at very high spatial resolution (a few tens to a few hundreds of meters at most). Implementing such models in large-scale ~~ice-sheet~~ ice sheet models is currently outside the realm of possibilities. However, as there is a growing interest in performing ~~ice-sheet~~ ice sheet projections over multi-centennial time scale, ~~the GRISLI-like~~ large-scale ice sheet models would undoubtedly benefit from the implementation of simplified infiltration schemes (e.g., Goelzer et al., 2013) so as to account for the impact of ongoing changes in surface meltwater on ice ~~dynamics~~flow.

An additional limitation related to the choice of our spin-up procedure is that the glacial-interglacial signature of past climatic changes is ignored. Neglecting the climate history of the Greenland ice sheet implies too warm ice temperatures. This may have an impact on the future GrIS evolution and on its contribution to sea-level rise. Indeed, the basal drag coefficient inferred from the inverse method may be too high so as to compensate the errors induced by the artificial warming. However, using a higher-order ice flow model, Seroussi et al. (2013) showed that at the centennial time scale the basal conditions and the GrIS projections are only poorly sensitive to the initial vertical temperature profile but are critically dependent on atmospheric conditions.

Despite these limitations, the sea-level projections performed with GRISLI compare well with those conducted with more sophisticated ~~ice-sheet models Edwards et al. (2014a)~~ ice sheet models (Edwards et al., 2014b), and the simulated surface ice velocities present a good agreement with the observed ones (Fig. S11). It appears thus as a good numerical tool to be coupled with a regional climate model with a reasonably good representation of the ice dynamics and ~~a~~ limited computational resources.

6 Conclusions

This study is based on the first regional atmospheric – ~~ice-sheet~~ ice sheet coupled model allowing the GrIS-atmosphere feedbacks to be accounted for. Using this new model, we investigated the GrIS evolution and its contribution to sea-level rise from 2000 to ~~AD~~ 2150 under a prolonged RCP8.5 scenario (2W experiment). The importance of the GrIS-atmosphere feedbacks has been assessed through the comparison of the two-way coupled experiment with two other simulations based on simpler coupling strategies: the NF experiment in which the MAR outputs are directly used as GRISLI forcing and the PF experiment in which the elevation feedbacks are parameterized. In both NF and PF experiments, changes in topography simulated by GRISLI are not updated in the atmospheric model. The main conclusions drawn from this study are the following:

- Accounting for the GrIS-atmosphere feedbacks amplifies the ice mass loss and changes in ~~ice-sheet~~ ice sheet geometry with increased surface slopes from the central regions to margin areas ~~. In turn, changes in the shape of Greenland modify the~~ and consequences on the Greenland ice velocities.

- The effect of accounting for the feedbacks between GrIS and the atmosphere increases ~~over~~ with time and becomes significant at the end of the 21^{st} century, as illustrated by ~~a~~ the 2W-NF difference in GrIS contribution to sea-level rise ~~of 20.4 cm~~ in 2150~~against 7.9 cm only~~ , i.e. 1.9 cm, against 0.3 cm in 2100.

– Accounting for the parameterized elevation feedbacks in the PF experiment leads to an additional SLR contribution of ∼7.6 % in 2150 compared to NF. On the other hand, the parametrization used in PF allows to reduce the mismatch (in terms of SLR projections) between the one-way and the two-way coupled approaches by 73.7 % in 2150, showing that at this time scale, changes in ~~ice-sheet~~ ice sheet geometry appear to be dominated by the SMB-elevation feedback.

– Finally, with our modelling setup, we showed that estimating the GrIS contribution over a fixed ice sheet mask (as in PF and NF experiments) overestimates the SLR contribution by ∼6 %, suggesting that most of RCM-based studies have probably overestimated the ice loss computed from changes in SMB.

7 Data availability

The model output from the simulations described in this paper are freely available from the authors without conditions. The source code of MAR version 3.7 is available on the MAR website: http://mar.cnrs.fr. The GRISLI source code are hosted at https://forge.ipsl.jussieu.fr/grisli, but are not publicly available due to copyright restrictions. Access can be granted on demand by request to Christophe Dumas (christophe.dumas@lsce.ipsl.fr).

Author contributions. The implementation of the three coupling methods as well as the simulations were done by X. Fettweis and C. Wyard. S Le clec'h, S. Charbit and A. Quiquet analysed the results and wrote the manuscript with contributions from M. Kageyama, C. Dumas and X. Fettweis. The GRISLI model was developed by C. Ritz.

Competing interests. The authors declare that they have no conflict of interest.

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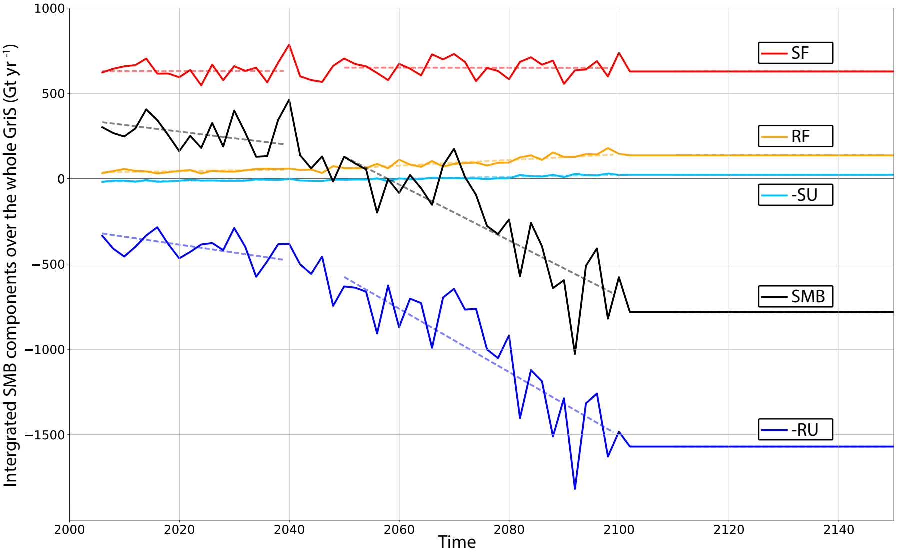

Figure 1. Evolution of the SMB (black line) and its components from 2005 to 2150 (in Gt yr^{-1}) simulated by MAR in the 2W experiment and integrated over the ~~entire GrIS~~ice sheet mask taken from Bamber et al. (2013). The SMB components are: snowfall (SF, red line), rainfall (RF, orange line), sublimation (-SU, light blue line), runoff (-RU, dark blue line). Dashed lines correspond to regression lines ranging from 2000 to 2039 and from 2041 to 2099. The light dark solid line corresponds to the zero line.

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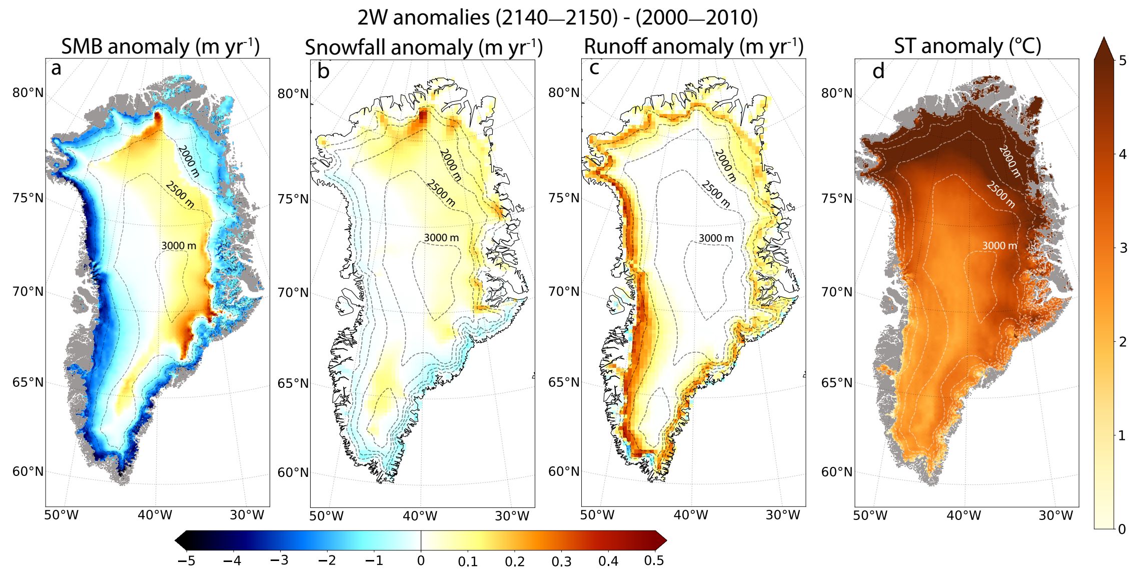

Figure 2. Anomalies of (a) mean annual surface mass balance in m yr^{-1} (b) annual snowfall in m yr^{-1} (c) annual runoff in m yr^{-1} and (d) mean annual surface temperature (in °C). These anomalies are given between the last ~~(2140-2150)~~ 2140—2150 and the first ~~(2000-2010)~~ 2000—2010 ten years of the 2W experiment. (a) and (d) are computed on the GRISLI grid and (b) and (c) are given on the MAR grid. The dashed lines correspond to the 500 m surface elevation iso-contours for the present-day observed topography. The grey shade represents the non ice-covered areas. Note for (a), (b) and (c) that the colour scale is not symmetric for positive and negative values.

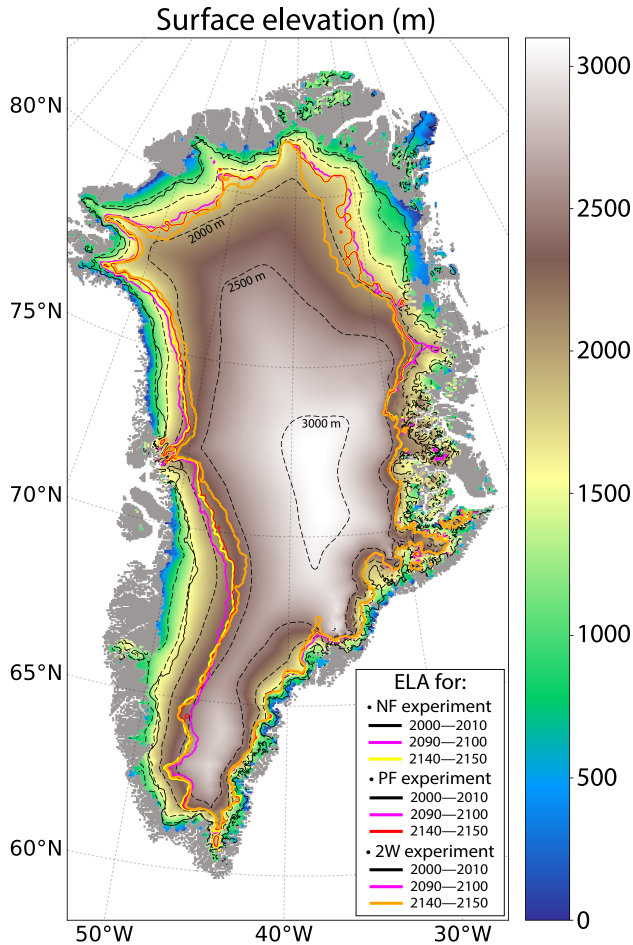

Figure 3. ~~Mean~~ GrIS surface elevation ~~for~~ in 2150 simulated in the ~~last ten years (2140-2150) of the~~ 2W experiment (in m). The solid black ~~line represents~~ and purple lines represent the equilibrium line ~~altitude~~ altitudes (ELA, limit between the accumulation and the ablation zones) in ~~2000-2010 for~~ the ~~NF, PF~~ 2000—2010 and ~~2W experiments. The solid green line represents the ELA of the 2090-2100~~ 2090—2100 mean ~~period~~ periods respectively for the NF, PF and 2W experiments. The ~~blue~~yellow, red and ~~yellow~~ orange solid lines indicate the ELA position ~~of the~~ over 2140—2150 for the NF, PF and 2W experiments respectively~~over 2140-2150~~. The dashed lines correspond to the 500 m surface elevation iso-contours for the present-day observed topography. The grey shade represents the non ice-covered areas.

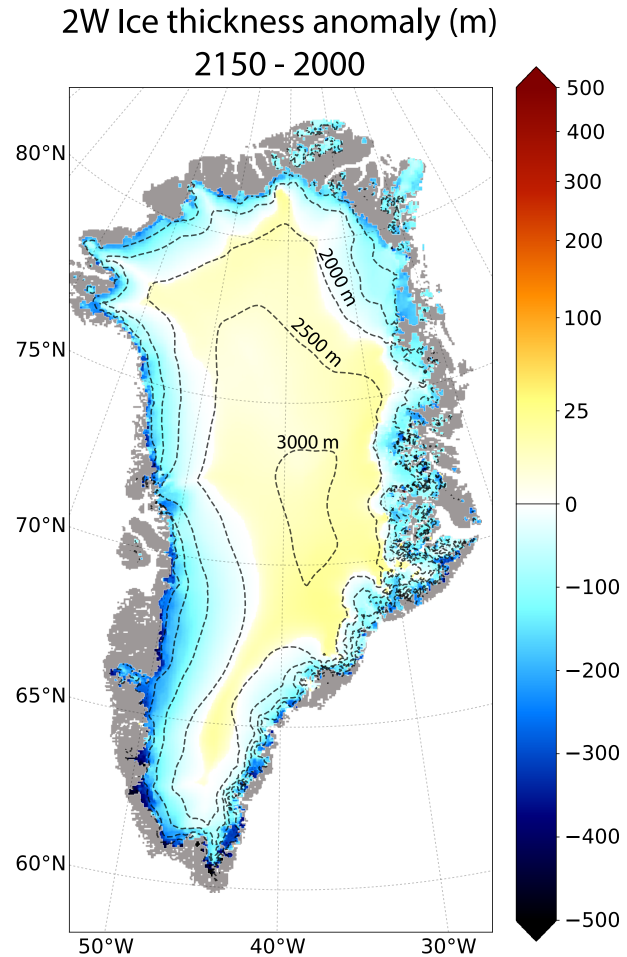

Figure 4. Ice thickness anomaly (~~in m~~2150-2000) ~~between~~ simulated in the ~~last (2140-2150) and the first (2000-2010) ten years of the~~ 2W experiment (in m). The dashed lines correspond to the 500 m surface elevation iso-contours for the present-day observed topography. The grey shade represents the non ice-covered areas. A non-linear color scale is used for positive values.

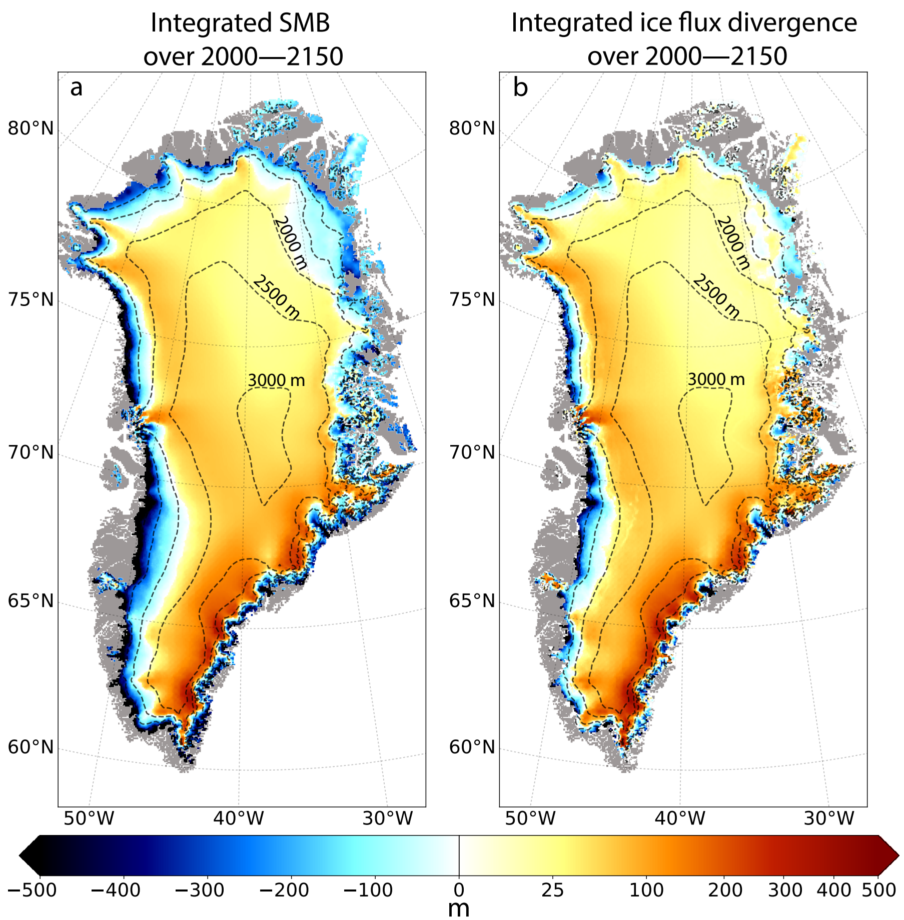

Figure 5. Cumulated SMB (a) and ice flux divergence (b) throughout the entire 2W experiment (~~2000-2150~~from 2000 to 2150) given in meters and computed on the GRISLI grid. The dashed lines correspond to the 500 m surface elevation iso-contours for the present-day observed topography. The grey shade represents the non ice-covered areas. A non-linear color scale is used for positive values.

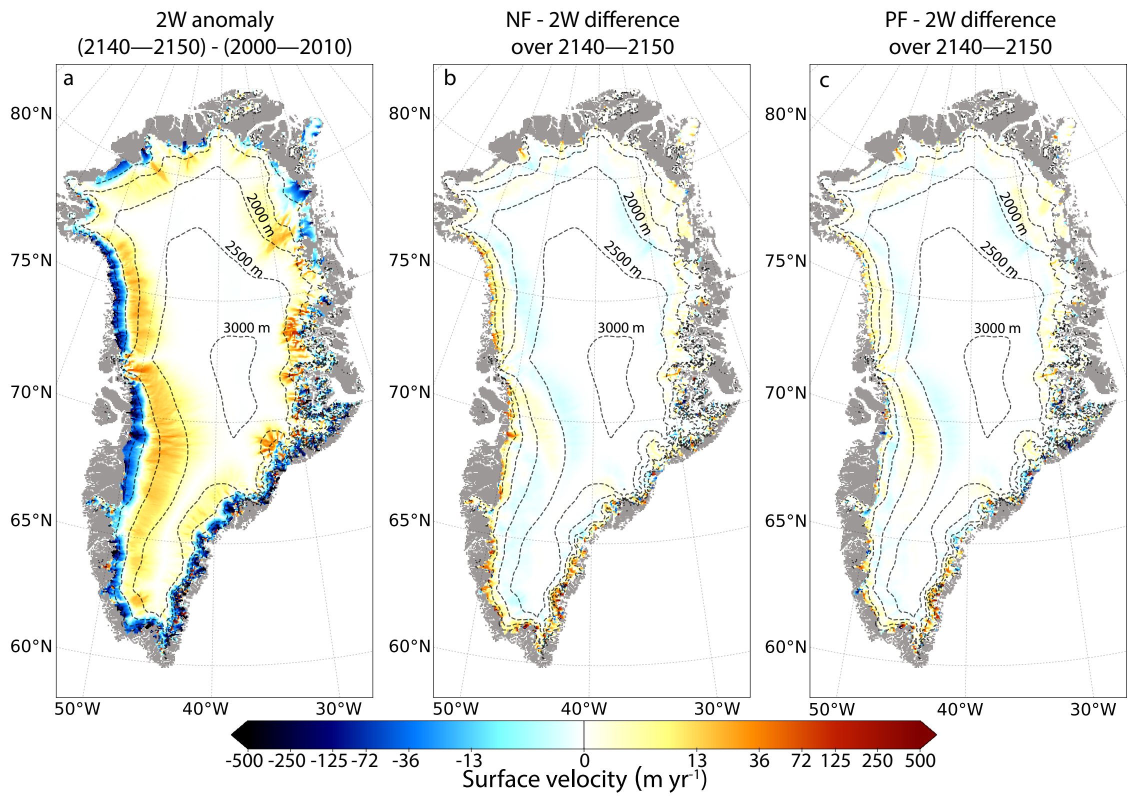

Figure 6. (a) Mean surface velocity ~~anomalies~~ anomaly (in m yr^{-1}) ~~(a)~~ between the ~~2140-2150~~ 2140—2150 and the ~~2000-2010~~ 2000—2010 mean periods for the 2W experiment. (b) Mean surface velocity difference between the ~~2W~~ NF and the ~~NF~~ 2W experiments for the ~~2140-2150~~ 2140—2150 mean period~~and~~. (c) ~~between~~ Same as (b) for the ~~2W and the~~ PF and 2W experiments~~for the 2140-2150 mean period~~. These ~~anomalies~~ differences are computed on the GRISLI grid. The dashed lines correspond to the 500 m surface elevation iso-contours for the present-day observed topography. The grey shade represents the non ice-covered areas. A non-linear color scale is used for both positive and negative values.

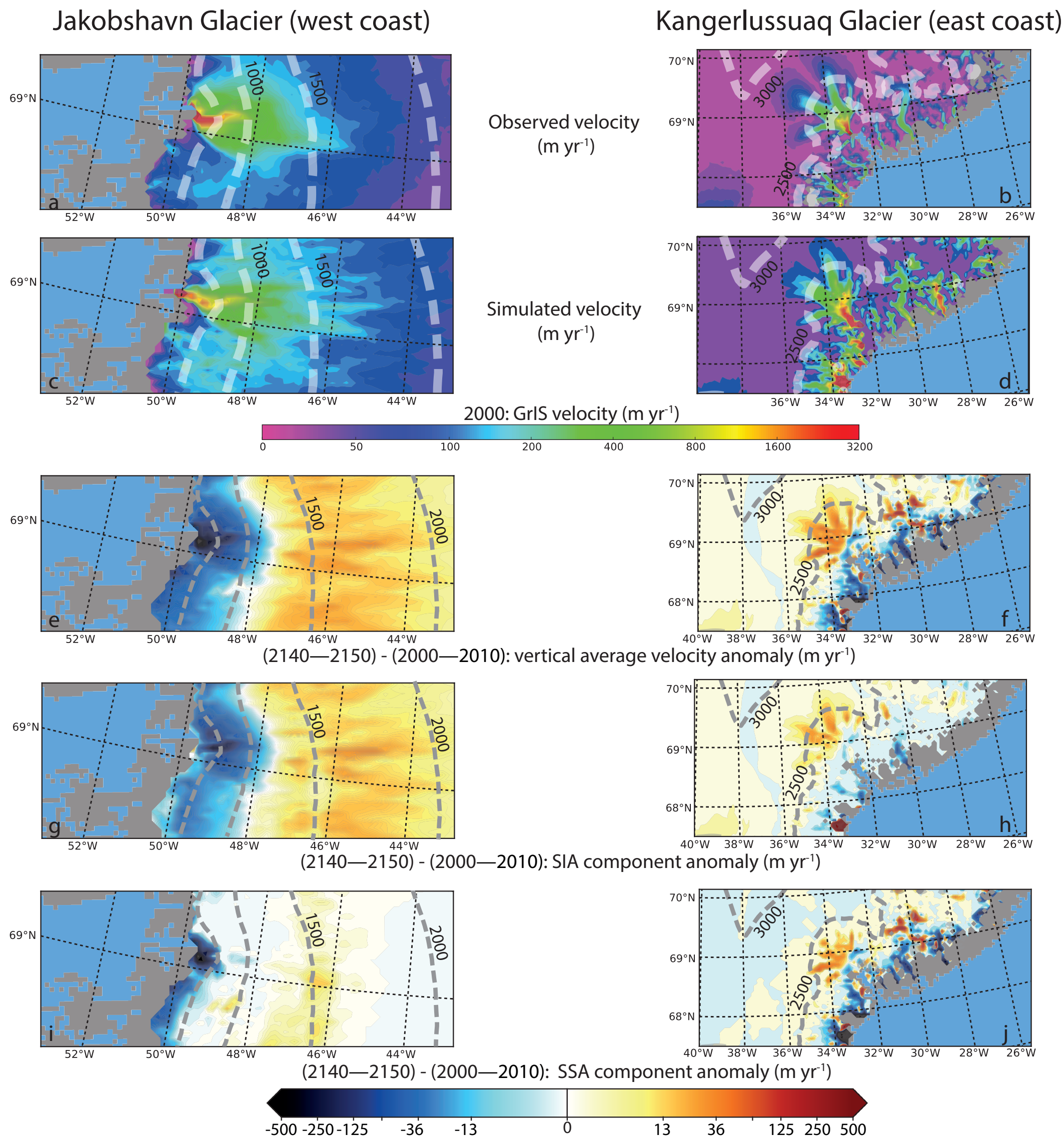

Figure 7. Regional zoom over the Jakobshavn (left ~~panesl~~panels) and the Kangerlussuaq (right panels) glaciers for the 2W experiment. ~~The top panels~~ (~~1st line~~a) ~~represent~~ and (b) are the observed velocities (in m yr^{-1}) from Joughin et al. (2018). (c) and (d) are the simulated ~~surface~~ velocity (in m yr^{-1}) after the initialization procedure. Panels from (~~big boxes~~e) ~~and the observed surface velocities~~ to (~~small boxes~~j) ~~from Joughin et al. (2010). All other panels~~ represent the velocity anomalies ~~between the~~ (~~2140-2150~~in m yr^{-1}) between the 2140—2150 and the ~~(2000-2010)~~ 2000—2010 mean periods of the 2W experiment for the vertically averaged surface velocity (~~2nd line, in m yr^{-1}~~e and f), the SIA velocity component (~~3rd line in m yr^{-1}~~g and h) ~~,~~ and the SSA velocity component (~~4th line, sliding velocities, in m yr^{-1}~~i and j). Note that a logarithmic scale is used for the ~~vertically averaged~~ velocity ~~and for the SIA and SSA components~~anomalies. The dashed lines correspond to the 500 m surface elevation iso-contours for the present-day observed topography. The grey shade represents the non ice-covered areas and the blue shade is the ocean mask.

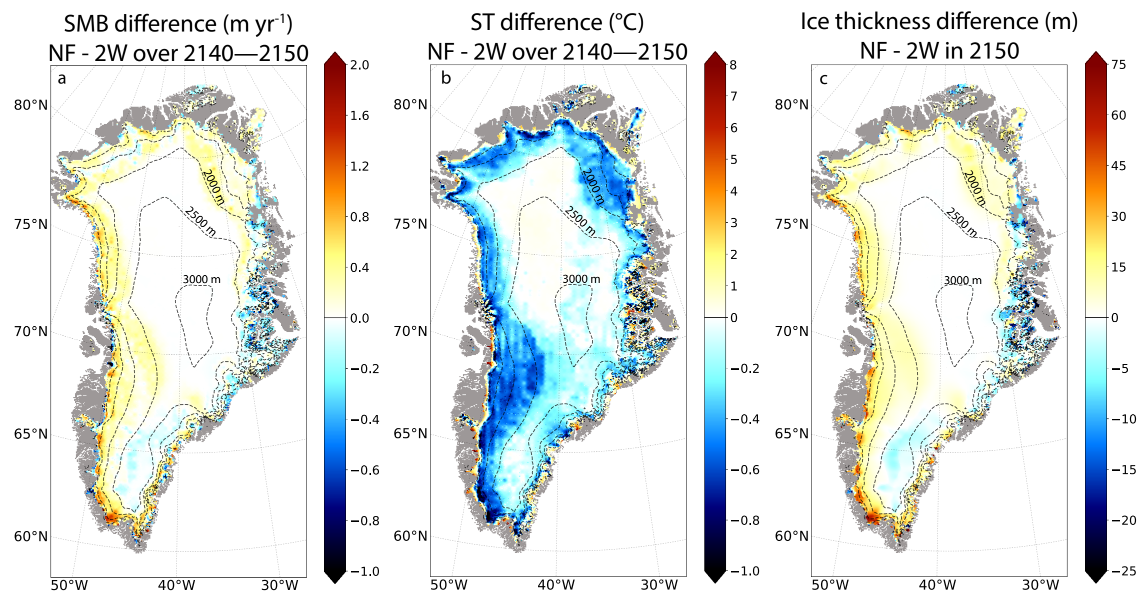

Figure 8. Mean ~~anomalies~~ differences (~~2W – NF~~NF-2W) for the ~~(2140-2150)~~ 2140—2150 mean period (a) Annual surface mass balance (in m yr^{-1}), (b) Annual surface temperature (in °C), (c) Ice thickness (in m) in 2150 between NF and 2W. Note that these differences are computed on the GRISLI grid. The dashed lines correspond to the 500 m surface elevation iso-contours of the present-day observed topography. The grey shade represents the non ice-covered areas.

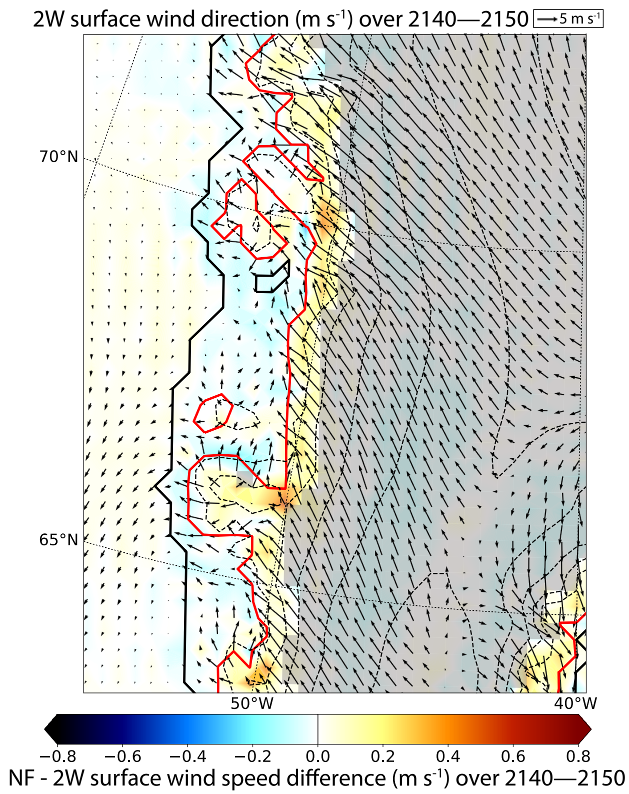

Figure 9. ~~Anomaly of surface~~ Surface wind speed difference (shaded) between the ~~2W~~ NF and the ~~NF~~ 2W experiments for the ~~2140-2150~~ 2140—2150 mean period. Black arrows represent the wind direction in the ~~2140-2150~~ 2140—2150 mean period of the 2W experiment. The length of the arrows indicate the magnitude of wind speed. The grey shaded area stands for the extent of the region for which the permanent ice fraction is 100 % (no tundra). Red solid line indicate the ~~ice-sheet~~ ice sheet extent The dashed black lines correspond to the 500 m surface elevation iso-contours~~(500, 1000, 1500, 2000, 2500 m) for the present-day observed topography~~.

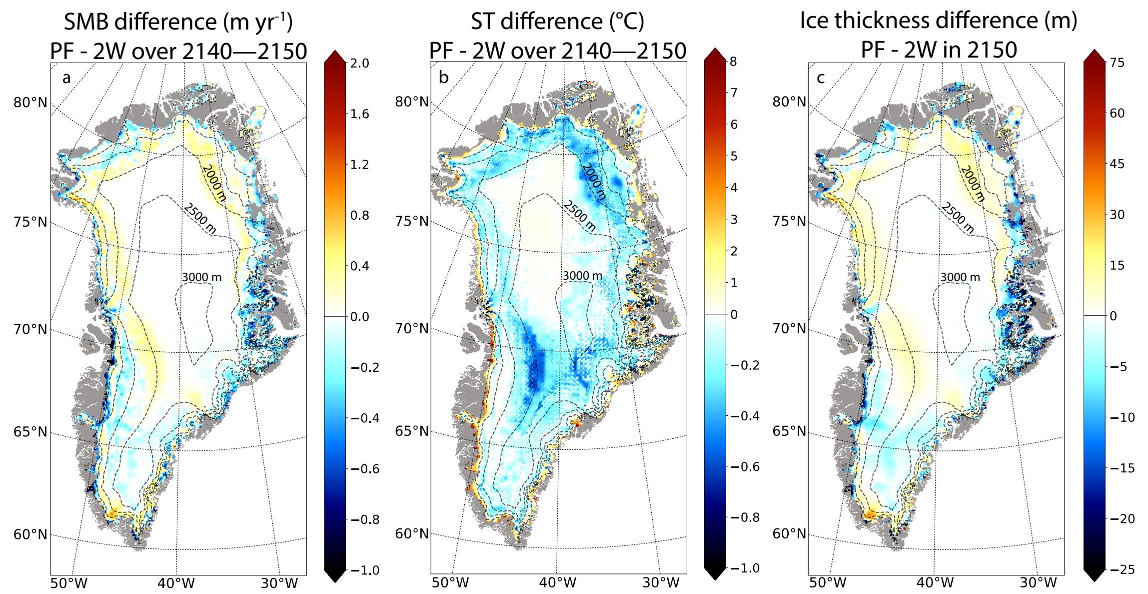

Figure 10. ~~Same as Figure 8~~ Mean differences (PF-2W) for the ~~anomalies~~ 2140—2150 mean period (a) Annual surface mass balance (in m yr^{-1}) and (b) Annual surface temperature (in °C) (c) ice thickness (in m) in 2150 between ~~the 2W~~ PF and 2W. Note that these differences are computed on the ~~PF experiment~~GRISLI grid. The dashed lines correspond to the 500 m surface elevation iso-contours of the present-day observed topography. The grey shade represents the non ice-covered areas.

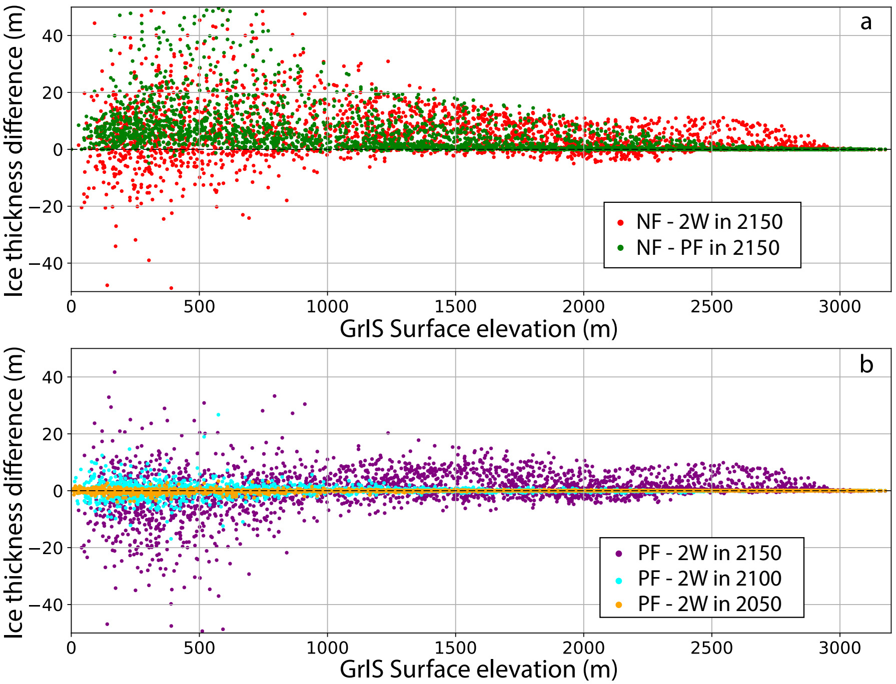

Figure 11. Ice thickness ~~anomalies~~ difference (m) as a function of the GrIS surface ~~altitude~~ elevation (m) for the three coupling experiments. In (a) the red ~~dots represent 2W – NF differences averaged between 2140~~ and ~~2150;~~ green dots represent the ~~PF – NF differences over~~ NF-2W and the ~~same period. Figure~~ NF-2W differences in 2150; (b) ~~represents the ice~~ Ice thickness ~~anomalies~~ difference (~~2W – PF~~PF-2W) ~~averaged over 2040-2050~~ in 2050 (~~red~~ purple dots), ~~2090-2100~~ 2100 (~~green~~ light blue dots) and ~~2140-2150~~ 2150 (~~blue~~ orange dots) mean periods.

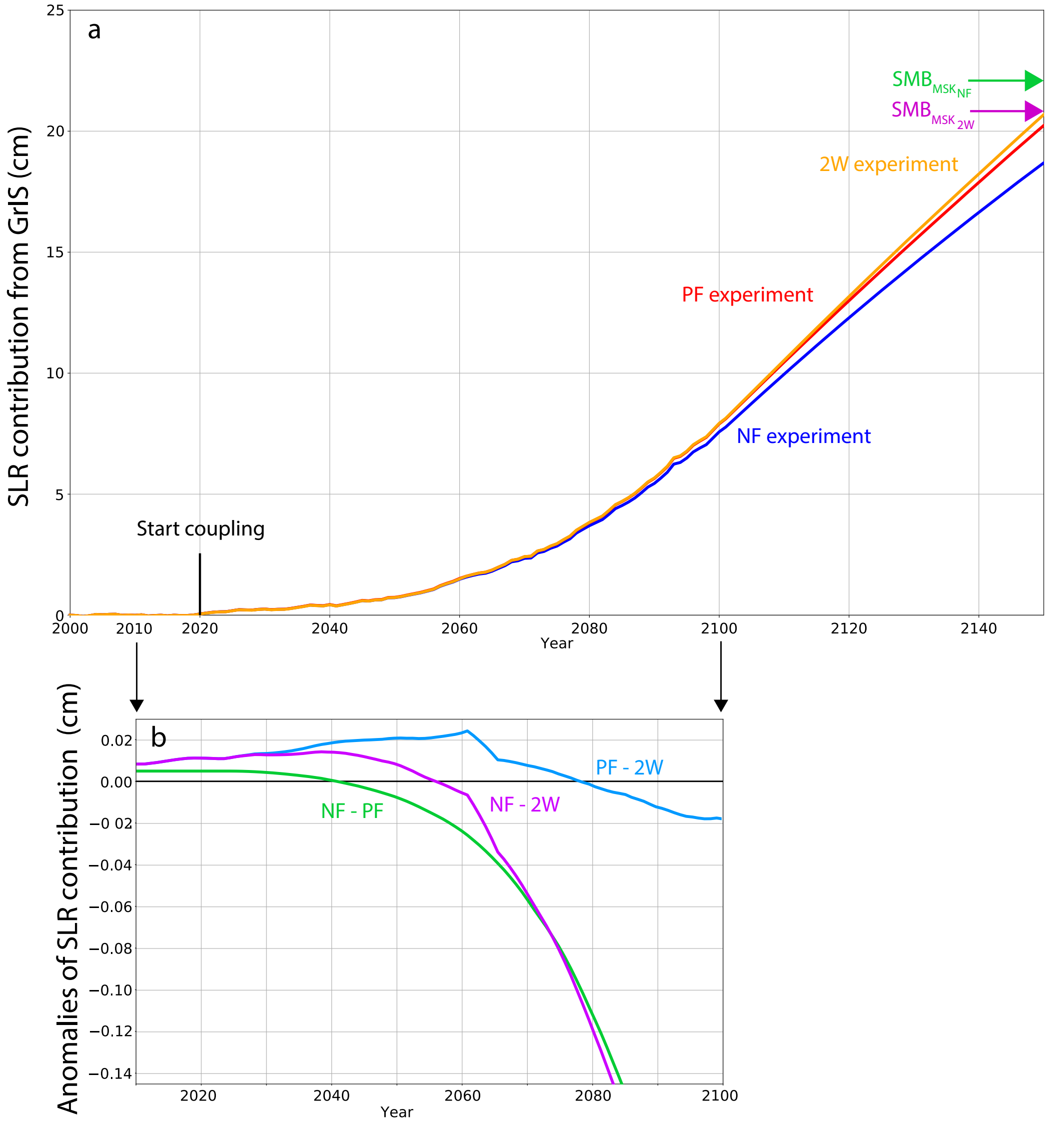

Figure 12. (a) Contribution of the GrIS to sea-level rise (in cm) as simulated in the NF (blue line), PF (red line) and 2W (yellow line) experiments and inferred from the ice thickness changes between the ~~2140-2150~~ 2000 and ~~2000-2010 mean periods.~~ 2150. Green and purple arrows indicate the projected contribution from GrIS inferred from SMB changes integrated ~~between~~ over the same ~~mean periods~~ period over a fixed ~~ice-sheet~~ ice sheet mask (SMB$_{\cancel{MSK-NF}\,MSK_{NF}}$) and a time variable ~~ice-sheet~~ ice sheet mask (SMB$_{\cancel{MSK-2W}\,MSK_{2W}}$). (b) Zoom of the ~~anomalies~~ differences of GrIS contributions to sea-level rise between the PF and the NF experiments (~~red~~ green line), between the 2W and the NF experiments (~~blue~~ purple line) and between the 2W and the PF experiments (~~yellow~~ light blue line).