We would like to thank the two anonymous reviewers for their careful reading of our manuscript and their constructive comments. In the following, we reply point by point to each individual comment (referee's comments are in green and italicised). Our new manuscript in which we highlighted changes from the original version can be found at the end of this document.

Anonymous Referee #1

This paper is clearly written and the figures are good. It describes the results of following the ISMIP6 Greenland experimental protocol with a particular dynamical ice-sheet model. Although this is information of use to assessing uncertainties in projections, the scientific gain is not clear. It would be useful if the authors could emphasise scientific lessons we learn from studying this model in particular, beyond its inclusion in the ISMIP6 comparisons, for example? Looking at the conclusions alone, I think a reader who is familiar with the literature of the last several years would find nothing new or surprising, for instance. However, in the paper there are a few new things which ISMIP6 is helping to clarify, and there are moreover useful things which have or could be done with this model, because it is computationally cheap, to test sensitivities.

It seems to us that such papers that show an individual group contribution to a large intercomparison exercise present three main added values:
- It is a way to document a specific model response for a set of forcings. For example, here, GRISLI shows a sensitivity to climate forcing close to the mean ISMIP6 participating models. This is a potential important information to analyse any further GRISLI results in a broader context.
- The uncertainty that arises from climate evolution (atmospheric and oceanic forcing) can be better quantify in such paper. Although it could also be quantify in the community paper, it is nonetheless only partially address in Goelzer et al. (2020) because of too large material to cover.
- Finally, the ISMIP6 participating models use a wide range of initialisation procedure and they show various biases and model drift. Such issues cannot be discussed in the community paper while it is extensively shown here.
We have added a few information in the introduction section:
“The aim of this paper is to discuss the role of the forcing uncertainties for future projections of the Greenland ice sheet contribution to global sea level rise when using our model. This individual model response can be put in perspective with respect to the multi-model spread discussed in Goelzer et al. (2020). This paper discusses additional experiments not included in the community paper (CMIP6 forcing and separate effects of the oceanic with respect to atmospheric forcing). Compared to Goelzer et al. (2020), we provide here a more detailed description of the initial state and its associated biases and model drift. A companion paper (Quiquet and Dumas, submitted) describes the results for the Antarctic ice sheet.”

A few of my comments relate to the importance of the SMB forcing, which the paper demonstrates. It would be useful to quantify (graphically or in numbers) how much of the spread among GCMs and scenario is due simply to the time-integral of the SMB forcing (as applied to the ice-sheet model), and not affected by the ice-sheet model itself. While it is certainly necessary to use a dynamical ice-sheet model to study large changes in ice-sheets, it would be useful if the authors could present evidence for the need to use one for the 21st century (when not coupled to the atmosphere or ocean), especially as doing so introduces complications of drift and spinup, as described by the paper.

The spread among GCMs is now shown with a plot of the time evolution of the yearly SMB spatially integrated over the ice sheet.

If we are correct, with the time-integral of the SMB, the reviewer wants to see the SMB contribution to the Greenland melt with respect to the dynamical contribution. However, the time-integral of the SMB as applied to the ice sheet model already accounts indirectly for the dynamical changes because of: i- the SMB correction for the surface elevation change and; ii- the ice mask change. As a result, the time-integral of the SMB will not reflect the impact of SMB only but also, in part, the dynamics. An alternative would be to compute the time-integral of the SMB over a constant ice sheet topography instead of using the one simulated by GRISLI. Such methods has been widely used in the past (e.g. Fettweis et al., 2013; Meyssignac et al., 2017) as it allows to compute an ice sheet contribution to sea level rise from an atmospheric model only. However this is a crude approximation since the sum will aggregate the strongly negative SMB values at the margin of the ice sheet where the ice will soon disappear and hence overestimate the ice sheet contribution to sea level rise. This overestimation has been quantified with GRISLI to be about 6% (Le clec'h et al., 2019) in 2150 (for 150 simulated years).

We think that the best way to separate the two effect is to compute the dynamical contribution to ice thickness change as explained in Sec. 3.2.3. Note that we also show in this response (Fig. R2) the integrated surface mass balance together with the dynamical contribution to ice thickness change and the ice thickness change, with the same colour scale.

I have some concern about the prescription of the large melting near the edge (o4 line 12) and the retreat masks (p4 line 32). With both of these enforced, is the dynamical behaviour of the model distorted?

The very negative SMB outside the present-day ice mask can be seen as a way to correct two type of biases:
- For some areas, the atmospheric forcing computed by MAR presents a positive annual value (ice accumulation) outside the observed present-day ice sheet mask. Uncorrected, this atmospheric forcing bias will result in an overestimation of the ice sheet extent and thickness.
- We use an inverse procedure to infer the basal drag coefficient that best represent the observed ice sheet thickness. By constraining the extent of the ice sheet with an artificial negative SMB, we infer a basal drag coefficient that best reproduce the dynamical behaviour of the ice sheet since the marginal slopes are closer to the observations.

This artificial negative SMB correction does not directly alter the dynamical behaviour of the model but it prevents any ice advance in the future. However, it is probably very marginal for the Greenland ice sheet in the future.

The glacier retreat parametrisation is slightly different. It could alter the dynamics since it is related to an imposed changed in ice thickness. However this is done on purpose, in order to account for a sub-grid process that is not accounted for otherwise. The effect of the glacier retreat can be quantified thanks to the AO experiments. To answer your concern, we have computed the dynamical contribution to ice thickness change (former Fig. 8) for the AO experiments compared to the full forcing experiment. In fact there is only very limited difference between the two (less than 10 metres difference).

p1 line 10-11. I would not jump to such a strong general conclusion (also on p7).

Reformulated to:
"Amongst the models tested in ISMIP6, the CMIP6 models produce larger ice sheet retreat than their CMIP5 counterparts."

p1 line 17-18. I don't think that this statement (of a most likely contribution of 1 m from ice sheets by 2100) is a correct representation of the current state of scientific knowledge. In the first place, *you can't state a likelihood independent of scenario, since there are no probabilities for scenarios. Bamber et al. write "For a +5degC temperature scenario, more consistent with unchecked emissions growth, the [median and 95-percentile] are 51 and 178 cm, respectively." I'm not sure what "most likely" means, but 1 m is twice their median. Also, Bamber et al. report an expert elicitation, whose reliability is debatable since it's opaque. For comparison, the AR5 assessment of the likely range of ice-sheet contributions by 2100 under RCP8.5 is 0.09 to 0.28 m from Greenland and -0.08 to 0.14 m from Antarctica.*

We agree with the reviewer. We have chosen to cite the Special Report on the Ocean and Cryosphere in a Changing Climate (SROCC, Oppenheimer et al., 2019) instead of Bamber et al. (2019) here. We have reformulated:
"Amongst the different contributions, the Greenland and Antarctic ice sheets have a potential to raise substantially the global mean sea level, with a weakly constrained trajectory (Oppenheimer et al., 2019)."

p2 line 1. Why "asymptotic"?

"approximations" has been replaced by "expansions" since SIA and SSA are the series expansion truncated at the order 0 of the Stokes equation. As such, they are asymptotic expansions.

p3 line 4. I suppose that strictly you could say an ice-sheet model satisfied momentum conservation, but as far as I know this model and others used for such purposes do not contain terms for acceleration or inertia. That is, momentum is always negligible, and they assume a balance of forces at all times.

Rephrased to: "[…] that solve the mass conservation and force balance equations".

p3 Eq 1. I think that BM is a single quantity, isn't it? Typeset like this in a formula it looks exactly like the product of two quantities B and M (like Ubar H is a product). It would be clearer to use a single symbol. Is it just the surface mass balance, or is basal mass balance included too?

BM has been replaced by M. It is the total mass balance (including basal mass balance). It is now specified in the text.

p3 line 11. "the total velocity is simply to superposition of the two main approximation". I would suggest "the total velocity is the sum of the velocities predicted in their respective areas by the two main approximations".

Thanks for your suggestion. We prefer to avoid the use of "respective areas" though, since both velocities are computed for all glaciated grid point. We reformulated as:
"For the whole geographical domain, we assume that the total velocity is the sum of the velocities predicted by the two main approximations: [...]"

p3 line 15. "for which there is infinite, respectively none, friction at the base." I think this should read "for which there is infinite friction at the base or none, respectively." "None" is a pronoun, not an adjective.

Thank you, it has been corrected.

p4 line 5. How accurate are the SMB and the surface topography in the control state?

The surface mass balance used for the control simulation comes from MAR v3.9. This present-day reference climate is also the one used for the initialisation procedure. This has been clarified:
"This present-day reference climate forcing is used for the initialisation procedure and for the control experiment *ctrl*."
MAR v3.9 is one of the few regional climate models that have been extensively validated against observations. On top of the two reference papers cited, there is an extensive literature that shows the model performance. We think that MAR offers an accurate representation of the present-day climate over Greenland even though, as any model, it might present some biases (for example a possible overestimation of the precipitation in South-East Greenland, discussed in the manuscript).

Since there is virtually no floating points in the model, the simulated surface topography in the model is the sum of the bedrock topography with the ice thickness. Isostasy being desactivated (now stated in the manuscript), the bedrock topography remains to the one in the observational dataset (Morlighem et al., 2017). Thus, the simulated topography accuracy in the control experiment can be measured by the error on the ice thickness, discussed in Sec. 3.1.

p4 line 11-14. Does this term strongly interfere with, or even overwhelm, the simulated discharge across the grounding line?

No, it is only a way to prescribe an ice extent that fits the ice sheet mask in the observations. It has consequences on the initial ice mask and topography and as such it defines the ice dynamics in the initial state (through surface slopes and basal drag coefficient). However, it does not interfere with potential changes in the ice dynamics.

p4 line 24. State that these are vertical gradients. I would say that they are vertical gradients of surface quantities in the atmosphere model, rather than in the atmosphere.

Right, we have followed your suggestion:
"[...] yearly values of vertical gradients in the atmospheric model for these two surface variables are also provided."

p5 line 11. branched to -> branched from.

Corrected.

p4 line 21-22. What do you need the surface temperature for, if you're using SMB as forcing?

Surface temperature is a boundary condition for the temperature diffusion equation. Since the model is thermo-mechanically coupled, temperature affects ice velocities (through viscosity). It will also play a role on the thermal conditions at the base of the ice sheet which also affect ice velocities (frozen grid-points have an infinite friction at the base).

p5 line 22, p8 line 3, p11 line 11, Fig 5 caption. Although the reader may sympathise with the authors, it's better to avoid "pessimistic" and "optimistic", which are value-judgements.

Replaced by high emission and low emission scenarios.

p5 last para. I don't understand the reason for these two experiments. Do they start from the same initial state? Since they have the same forcing, they ought to evolve identically.

The experiments *ctrl* and *ctrl_proj* have two different initial states since they start at two different dates: 1995 and 2015, respectively. The *ctrl* experiment can be used to quantify the drift in our

model during the whole time period (including the historical and the projection). In turn, the advantage of the *ctrl_proj* experiment is to be directly comparable to the projection experiments as they cover the same time period and they use the same initial state (which was not the case with the *ctrl* experiment). To clarify this point, we added the following:
“The *ctrl* experiment can be used to quantify the simulated model drift over the whole time period (1995-2100). Instead, the *ctrl_proj* can be directly used to quantify the importance of climate forcing evolution since it uses the same initial state in 2015 as the different projection experiments.”

p5 line 34. alike -> like.

Corrected.

p6 line 22. best -> better.

Corrected.

p6 line 24. "In doing so" means doing what? - absolute or logarithm? I would have assumed logarithm, but the next sentence suggests otherwise. What are the units of 0.55? What are the units of velocity before taking the logarithm? (Strictly you can only take the log of a dimensionless quantity, but the conversion factor between different velocity scales will be a constant offset in the log so doesn’t affect its RMSE, I suppose.)

We meant logarithm of the velocity (expressed in metre per year but as you rightly point out an other choice will not affect the RMSE). We have rephrased to:
“When using the logarithm of the velocity GRISLI slightly improves compared to the other participating models since the RMSE is about 0.55 (eleventh worst value out of 21). This means that the error [...]”

p6 line 30. Why is this "on the contrary"? If I read this correctly, all the errors are in the same direction (too slow in the model). Can you suggest the reason for this systematic bias? What implication does it have for projections?

It should have been “On the contrary, the South East glaciers, Kangerdlugssuaq and Helheim, are too fast in the model.” (and not “too slow”). There is no systematic biases for the velocity: amongst the largest ice streams, the NEGIS, Petermann and Jakobshavn are too slow but the Kangerdlugssuaq and Helheim are too fast.

p7 line 4-5. What implication will this bias in SMB have for projections?

It is difficult to give a definite answer to this question. An overestimation of the precipitation might moderate the effect of the expected decrease in SMB in the future. However a too wet climate could also be the sign of a too intense penetration of warm (thus humid) air over this area which could also facilitate melting at high elevation. Such atmospheric processes are best quantified with dedicated atmospheric model experiments.

p7 line 9 and 15. Are these large drifts in thickness and velocity related? What effect will they have on projections? It’s not obvious that you can simply subtract an unforced drift when it’s large compared with the forced response.

The drift in thickness and velocities are related since the two variables are tightly coupled together in the model. However, we think that the velocity drift mostly derived from the ice thickness drift.

For example, the ice thickness drift in South-East Greenland near the Helheim glacier is negative, which induces a decrease of the ice velocity (less ice to export).

In the paper, the plots of the time evolution of integrated variables show the control experiments (i.e. the drift) as well as the projections without the drift subtraction. We subtract the drift only for 2D maps to better highlight the impact of climate change. However, the drift shown in Fig. 1 (original manuscript) is small when compared to the ice thickness change induced by climate change.

p7 line 18. start by -> start with.

Corrected.

p7 line 21-24. Presumably this spread comes mostly from the spread in SMB forcing from the GCMs. Could you also add the ice-sheet area- and time-integral of the SMB perturbation to the graphs?

In addition to the response we made earlier on your main comment, we can add a few information here. We have preferred to not plot the time integral of the mean SMB over the ice sheet since it may be more difficult to interpret than the yearly evolution. The time integral of this variable is essentially positive with only negative values for some models towards the end of the century. This is because the area-integrated SMB becomes negative only after 2060 for some models (and remains positive for others). Since the simulated ice sheet shows only a small drift in the control experiment, the positive area-integrated at the beginning of the simulation is almost balanced by the melt at the base of the ice sheet and the calving flux. Thus, the time integral of the spatial mean SMB draws an incomplete picture of the evolution of ice volume and does not allow for a separation of the ice dynamics versus SMB contribution.
Nonetheless, to show the spread amongst GCMs, we have added the time evolution of the SMB integrated over the ice sheet mask and added a few sentences:
“The differences in ice volume evolution are tightly linked to the surface mass balance evolution for the different climate forcing. Amongst the CMIP5 climate models, IPSL-CM5-MR and MIROC5 simulate a mean surface mass balance negative as early as 2060 while it remains positive over the next century for CSIRO-Mk3.6 (Fig. 4).”

p7 line 21-24. It seems that these projections imply quite a low sensitivity to climate change compared with the models on which the AR5 was based; their assessment of the Greenland contribution by 2100 under RCP8.5 is 90-280 mm, of which 40-220 mm is from SMB change.

Although slightly smaller perhaps, GRISLI shows a climate sensitivity close to the mean of the ISMIP6 participating model. The community paper (Goelzer et al., 2020) reports a range of 70-135 mm (mean of 100 mm) using MIROC5 RCP8.5 while GRISLI shows a range of 75-95 mm (low to high oceanic sensitivity) under the same forcing. This is now specified in the text:
“The 2100 sea level contribution simulated by GRISLI is close to the mean model response amongst the ISMIP6 participating models.”
The numbers in the AR5 for RCP8.5 were larger (Table 13.5 reports 0.07 to 0.21 m from which 0.03 to 0.16 m due to SMB change). However, these estimates were derived only from a small number of studies/models, compared to the 21 ice sheet models in ISMIP6. They were also obtained sometimes with a simpler methodology : the Special Report on the Ocean and the Cryosphere in a Changing Climate (SROCC) reports a median value for process-based approaches of 11.9 cm under RCP8.5.

p7 line 25. What sort of "tipping point" do you have in mind, that you might see in the volume evolution? Can you give references to relevant suggestions?

We were imprecisely referring to a sharp change of slope. This has been reformulated:
“However, we can not distinguish any sharp inflexion in the volume evolution over the next century.”

p7 line 26-27. I think we should be more cautious in drawing conclusions. There are only four CMIP6 models considered in this study, out of dozens in total, and two of the four are at the edge of the CMIP5 distribution in your projections. Only two show much greater sensitivity, and those results are within the AR5 range.

Thanks for pointing this out. We have reformulated:
“CMIP6 models show generally a much larger Earth climate sensitivity than their equivalent in the former CMIP5 generation (Forster et al., 2020). In particular, the CMIP6 models used in ISMIP6 have an Earth climate sensitivity from 4.8 to 5.3, i.e. larger than the CMIP5 models used here, which show a range from 2.7 to 4.6 (Meehl et al., 2020).”

p8 line 6-7. It’s not the GHG itself which is the driver, but the warming it produces; that is also the reason why the rate of mass loss goes up with time, and the main reason for the spread among models.

Reformulated:
“The future atmospheric and oceanic warming induced by the greenhouse gas mixing ratio is thus a major driver for the Greenland ice mass at the century time scale.”

p8 line 13-14. Since the point you wish to make is the similarity of the patterns, it would be better to show these maps divided by the integrated change in each case i.e. normalised to the same GMSLR contribution. That would reveal the patterns themselves, so they could be compared, which I agree should be the purpose of this figure.

Such figure is shown below in this response (Fig. R1). It is true that the new figure shows nicely the similarity of the patterns for the different GCMs. However, we think that the absolute ice thickness change for a given climate forcing is more informative for the reader as it is a way to show how the volume change (integrated value) translates into ice thickness change. However, if the reviewer believes that we should add this figure in the supplementary material, we would be happy to do so.

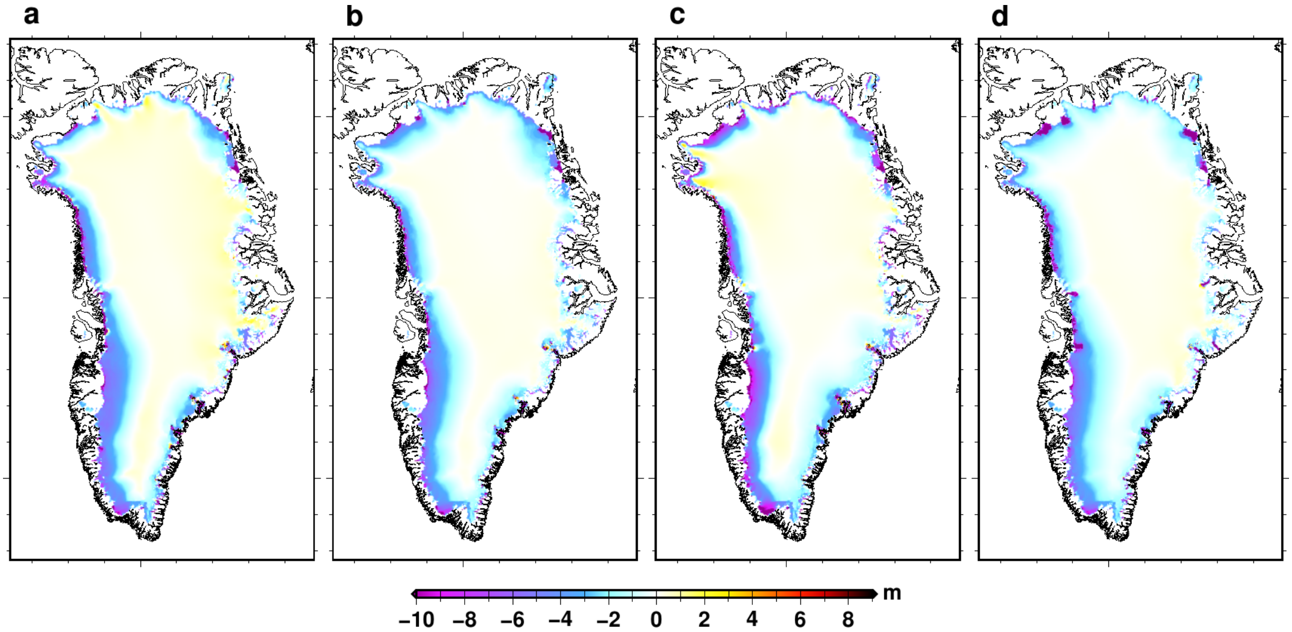

Figure R1. Simulated ice thickness change (2100 – 2015) normalised its spatial average (i.e. volume change) for: (a) CSIRO-Mk3.6 (RCP8.5); (b) MIROC5 (RCP8.5); (c) MIROC5 (RCP2.6) and; (d) UKESM-CM6 (SSP585) climate forcing. The medium oceanic sensitivity has been used for this figure, except for UKESM-CM6 (d) for which we use the high oceanic sensitivity.

p9 line 6. It would be interesting to see the time-integral of the applied SMB perturbation here, to compare with the AO experiments (as I also suggested on p7 for Fig 3). Any difference is due to the dynamical response to the SMB forcing.

The integrated SMB indirectly accounts for dynamical changes. First through the elevation feedback on SMB with the vertical lapserate. Second because the ice mask can change due to ice dynamics. will reflect indirectly the dynamical response, through the elevation change correction and ice mask change. We do not think that such a figure will allow to distinguish the dynamical response from the SMB forcing.

In Fig. R2 of this response, we show the integrated surface mass balance together with the dynamical contribution to ice thickness change and the ice thickness change, with the same colour scale.

p9 lines 16-23. The text says "Fig. 8b shows the difference in ice flux convergence in 2100", and the fig caption says "change in the dynamic contribution to ice thickness change in 2100". I don't think either of those is a correct description, if I have understood correctly. You also say, "This can be considered as the dynamical contribution to ice thickness change," which I think is correct. The quantity shown is the difference (change in topography during the experiment) minus (time-integral during the experiment of the local mass balance change with respect to control) - is that right? It would be useful to compare this difference with the change in topography in the same experiment, using the same color scale, in order to see the relative importance of the dynamical change. If it's a small fraction, you might argue that there's no need to use a dynamical ice-sheet model for projections on this timescale. Where it's not small, you can comment. Part of the dynamical contribution near the coast is a response to the ocean forcing, I presume. Therefore it would also be useful to show the same comparison for the AO experiment. That is, would it be good enough to make the projection without a dynamical model, simply by time-integrating the local SMB perturbation?

Yes you are right with the definition and thank you for pointing this terminology inconsistencies. It is now referred as “dynamical contribution to ice thickness change” throughout the manuscript.

We have added the ice thickness difference in Fig 9, to compare with the dynamical contribution to ice thickness change and added a few information in this manuscript:
"The integration in time of Eq. 1 over 2015-2100 suggests that the integrated ice flux convergence is the difference between the ice thickness change from 2015 to 2100 and the integrated mass balance (surface and basal mass balance and calving) over this period. The integrated ice flux convergence can be considered as the dynamical contribution to ice thickness change. It should be noted that the integrated mass balance here also includes the effect of ice mask change and surface elevation change. As such, it is not comparable to what would have been obtained with an atmospheric model only. Fig. 9b shows the difference of the dynamical contribution in 2100 for a selected climate forcing with respect to the control ctrl_proj experiment. The pattern mostly follows the one of velocity change (Fig. 9a). There is an important positive dynamical contribution to ice thickness change (ice flux convergence) at the margins that tends to partially compensate the decrease in surface mass balance. Conversely, upstream regions show a slightly negative dynamical contribution (ice flux divergence). This pattern is similar amongst the different climate forcings. To compare the relative importance of the dynamical contribution with respect to surface mass balance to explain the ice thickness change we show the ice thickness change in 2100 with the same colour scale in Fig. 9c. The dynamical contribution shows generally much smaller value suggesting that surface mass balance explains the largest changes in ice thickness. However, locally, for example in the South-East and central West regions the dynamical contribution can be the largest driver of ice thickness change."

Fig. R2 is the same as Fig. 9 in the paper, the only difference is that it shows the integrated surface mass balance as well. The dynamical contribution is directly constructed from the difference of the ice thickness change and the integrated total mass balance (from which surface mass balance is the main driver). In the paper, we keep the version of the figure with the dynamical contribution to ice thickness change together with the ice thickness change, but we omit the integrated surface mass balance since we do not think it brings an additional value.

There is virtually no change in the dynamical contribution to ice thickness change when comparing the standard experiment to the AO experiment. The glacier retreat parametrisation can be seen as a calving process. It implies a slightly greater ice thickness change but its effect is affected to the integrated mass balance change (which include surface and basal mass balance in addition to calving). The difference in thickness and surface slope change between the AO and standard experiment does not seem to be sufficiently large to affect the ice dynamics.

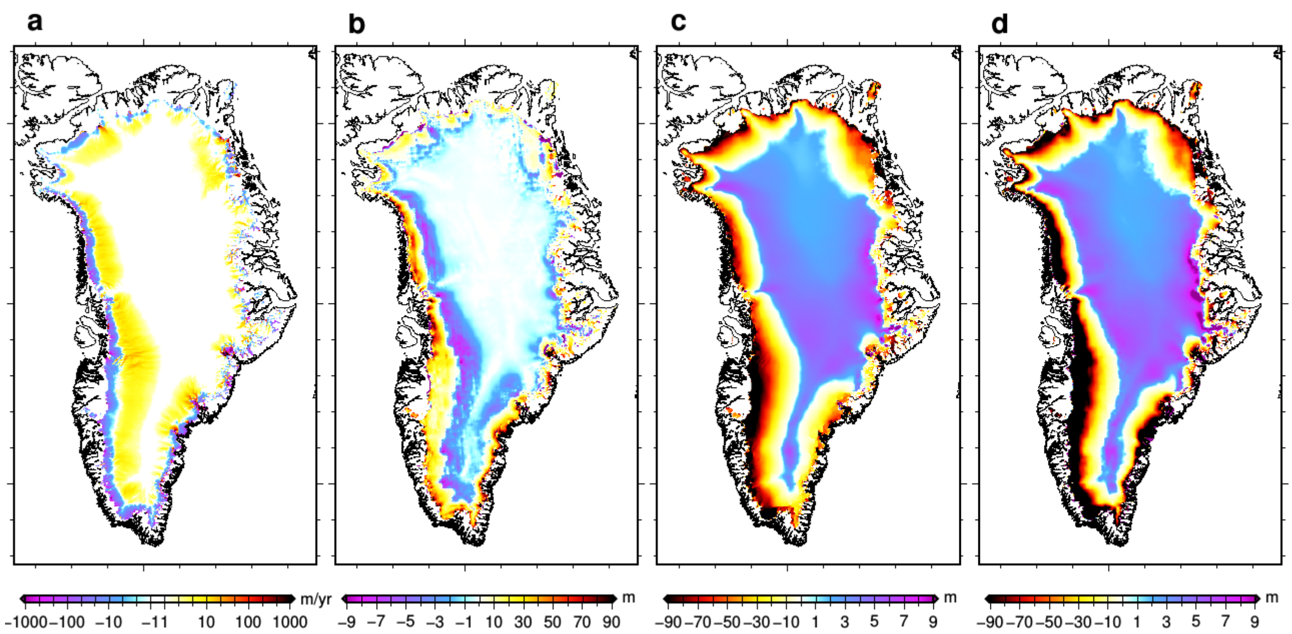

Figure R2. Simulated surface velocity change during the projection run (2096-2100 with respect to 2015-2019) using MIROC5 forcing under RCP8.5 with a medium oceanic sensitivity. **b**: change in the dynamical contribution to ice thickness change in 2100 (see text for definition) for this same experiment. **c**: simulated ice thickness change (2100-2015). **d**: time integral of the surface mass balance (2015-2100). For all panels, we corrected the changes by the ones simulated in the control experiment *ctrl_proj* over the same period. Note that the colour scale is not symmetrical for **(b)**, **(c)** and **(d)**.

p9 line 30. As a guide to the possible magnitude of this underestimate, you could state what the presently observed ice-sheet imbalance would give if it continued as a constant rate to 2100 and compare with your projected changes in response to forcing.

This is a very interesting point indeed, and maybe one of the major point of this paper but also the community paper. Our ice sheet models do not reproduce the recent accelerations and as such probably bias our projections towards low estimates. We have added the following:
"This means that, by constructions, our simulations underestimate the Greenland ice sheet contribution to future sea level rise. A simple linear extrapolation of the 2006-2016 rate (0.77 mm yr^{-1}, Oppenheimer et al., 2019) up to 2100 would result in a 6.5 cmSLE from the Greenland ice sheet. This number is large compared to the GRISLI results discussed in this paper, and more generally it is large compared to the spread amongst ISMIP6 models (3.5 to 14 cmSLE, Goelzer et al., 2020). This suggests that model initialisation is one of the largest source of uncertainty for model projections. Instead of using a methodology that produces ice sheet at equilibrium, some promising alternatives exist, [...]"

p10 line 4-16. This is useful, but it's not really discussion, I'd say. It's another sensitivity test, and it would go well in sect 3.2.3 about change in ice dynamics.

We have moved this part in the Sec. 3.2.3.

p10 line 20. Why is it necessarily an overestimate?

Because the diffusion of the cold temperature within the ice sheet is not accounted for. This is now clarified:
"Our internal temperature field is the result of a long thermo-mechanical equilibrium under perpetual present-day forcing and as such, it is necessarily overestimated since the diffusion in the ice sheet of the cold temperature of the glacial period is not accounted for."

p10 line 27-29. Yes, it would! Since your model is particularly computationally inexpensive, please could you do it and tell us the answer? :-)

Since we think that it makes little sense to perform long multi-millenial integrations with a constant prescribed basal drag coefficient, we are currently working on the calibration of the model parameters for an interactive computation of the basal drag coefficient as in Quiquet et al. 2018. However, although our model is relatively cheap it nonetheless currently requires 11 days on our local computers to perform 10 kyr with the 5-km grid resolution used in the paper. Hopefully in the future we will be able to show the behaviour of our model for two completely independent initialisation procedure.

p10 line 31-32. Could you quantify the elevation-SMB feedback here, or earlier, and compare it with Edwards et al. (Cryosphere, 2014)? You could directly quantify it by running a sensitivity test in which the lapse-rate adjustment is excluded, I suppose.

We have performed a sensitivity experiment using MIROC5 RCP8.5 and the medium oceanic sensitivity in which we did not account for the lapse-rate correction. We found a reduction by 5.1% of the Greenland contribution to sea level rise in this experiment with respect to its counterpart in which the correction is applied. This number is close to the 4.3 reported by Edwards et al. (2014). We have added the following:
“The forcing methodology used for ISMIP6-Greenland does account for the vertical elevation feedback on temperature and surface mass balance. In order to quantify the impact of this correction on the simulated evolution of the ice sheet, we run a sensitivity experiment in which this correction is not accounted for. Using MIROC5 under RCP8.5 scenario with a medium oceanic sensitivity, we simulate a Greenland contribution to future sea level rise 5.1% smaller in this sensitivity experiment compared to the same experiment in which the vertical correction is applied. This number is slightly higher than the effect reported by Edwards et al. (2014) and Le clec’h et al. (2019a) (4.3 and 4.2% respectively) but smaller to Vizcaino et al. (2015) (8-11%) and Calov et al. (2018) (about 13%). Differences in resolution and/or physical processes implemented in the atmospheric model could explain this diversity.”

p11 line 8. is systematically loosing -> systematically loses

Corrected.

Fig 1 caption. Does "respective to" mean "with respect to"? For clarify please state the years of the end of the historical and end of ctrl_proj.

Done.

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Anonymous Referee #2

In this manuscript, the authors report on their ISMIP6 Greenland projections with the model GRISLI. The paper is easy and straightforward to follow. Its scientific value beyond the community publication (Goelzer et al., 2020, in press) lies in a more detailed description of the set-up of GRISLI, a more detailed analysis of the results and the fact that the entire suite of ISMIP6 experiments (Tier 1-3) are dealt with.

Overall, I found the results interesting and the presentation adequate. I'd only like to raise some issues that should be dealt with as follows:

Thank you for your positive evaluation, we reply to your individual comments below.

The English writing clearly has some room for improvements. I am not going to point out all the issues, but just some examples from the first page: P. 1, l. 3/4: "an increase_d_ mass loss". P. 1, l. 5: "the largest single source contribution _after_ the thermosteric contribution". P. 1, l. 19/20: Assessment of projections? Either "need for assessment of future SLR by projections" or "need for projections of future SLR". P. 1, l. 22: Strange formulation: "from changing boundary conditions such as climate change". Before resubmission, the entire manuscript should be very carefully proof-read by a (near-) native speaker or a professional language editing service.

Thank you for your corrections, we have followed all your suggestions. We are indeed non-native English speakers but we put a lot of effort to write in English since it is the international language for Science. Even after careful reading, we are aware that our manuscript will contain grammatical errors or poorly formulated sentences but we think that it is generally understandable. If not, we will be more than happy to follow your corrections. Also, it might be relevant to note that The Cryosphere journal includes a language editing service for all accepted manuscripts.

Throughout MS (e.g., p. 1, l. 10, l. 14): "mmSLE" -> "mm SLE"

Corrected.

Throughout MS (e.g., p. 2, l. 4): "in term of" -> "in terms of"

Corrected.

P. 1, l. 17/18, "most-likely amplitude exceeding 1 metre in 2100": This is not what has been found in the ISMIP6 ensemble projections (Goelzer et al., 2020, TC, in press; Seroussi et al., 2020, TC, in press). Even with the most sensitive model results, it is less than half a metre combined. At some point in the paper, this should be mentioned.

We apologise, this was an exaggerated statement since the range of Bamber et al. (2019) is 51 to 178 cm for unmitigated emissions. We have chosen to cite the Special Report on the Ocean and Cryosphere in a Changing Climate (SROCC, Oppenheimer et al., 2019) instead of the expert judgement of Bamber et al. (2019) here:
“Amongst the different contributions, the Greenland and Antarctic ice sheets have a potential to raise substantially the global mean sea level, with a weakly constrained trajectory (Oppenheimer et al., 2019).”

The numbers for the contribution of the Greenland ice sheet to 2100 sea level rise in Goelzer et al. (2020) and in our manuscript are within the range of the SROCC.

P. 3, l. 7, 17: Add commas after the displayed equations.

Done.

P. 3, l. 11-13: The description of the SIA and SSA is over-simplified. Starting from full Stokes, in both cases, some horizontal and some vertical derivatives of the components of the stress tensor are neglected. In very compact form, this is shown in the tutorial at http://doi.org/10.5281/zenodo.3739009, p. 22 (for SIA) and p. 24 (for SSA).

Reformulated:
"For the whole geographical domain, we assume that the total velocity is the sum of the velocities predicted by the two main approximations: the shallow ice approximation (SIA) in which the deformation is entirely driven by the vertical shear and the shallow shelf approximation (SSA) in which the vertical shear is neglected and the horizontal stresses are predominant."

P. 3, l. 14/15: Is floating ice included in the simulations? If so, what is assumed for the sub-ice-shelf melt rate?

Only few glaciers in Greenland present a floating tongue and when they do it is located in very narrow valleys. The 5-km grid used in our model is not precise enough to represent these floating tongues and the physical processes related. This is why we have imposed a very large basal melting rate in our simulations (200 m yr^{-1}) to avoid floating points. This is now stated in the manuscript:
"Since 5 km is too coarse to represent Greenland floating ice tongues, sub-shelf melting rate has been set to a large value (200 m yr^{-1}) to discard simulated floating points."
We agree that is a simplification that can bias the future projections. In fact, this is not a problem specific to GRISLI since most ISMIP6 participating models do not have the resolution needed to represent such floating ice tongues. The ISMIP6 glacier retreat parametrisation has been developed (Slater et al., 2019) to account for such a process in models that would not otherwise.

P. 3, l. 16: "till _layer_"?

Corrected.

P. 3, l. 24ff: 30 kyr is likely not long enough to reach thermal equilibrium for an ice body as large as the Greenland ice sheet. This should be commented on. Further, does the inferred sliding depend on the basal thermal state, or is basal sliding applied everywhere?

We agree. This is why we performed more than ten cycles (thermal equilibrium + multiple 200-yr simulations). The basal drag coefficient and the basal temperature are coupled and in doing more cycles, we expect the two variables to be consistent with each other. This is now more clearly stated in the manuscript:
"After a few 200-yr experiments, we repeat the thermal equilibrium computation restarting from the previous equilibrium state with the newly inferred basal drag coefficient. In doing so, the basal drag coefficient and the temperature at the base are consistent with each other."

P. 6, l. 18ff: I cannot see it so well in Fig. 2, but it seems to me that the simulations do not include/reproduce the floating ice tongues (at least off the NEGIS). If so, this may also be partly responsible for the velocity misfits because buttressing effects are missing.

You are correct, we do not simulate the floating ice tongues which can exert buttressing. However, the simulated velocity in the NEGIS is underestimated so it cannot be linked to the missing buttressing (which would reduce the velocity even more). The velocity misfit is most probably linked to the basal drag coefficient.

P. 9, l. 7: It would be interesting to quantify this. How large (e.g., in per cent) is the difference between full forcing and (AO+OO)?

The AO+OO explains 93.6%, 91.6% and 92% of the full forcing for MIROC5, NorESM1 and CSIRO-Mk3.6 respectively. This is now stated in the manuscript:
“Also, the sum of the ice loss of AO and OO experiments approximate closely the ice loss simulated when using the full forcing (92 to 94% of the full forcing). ”

P. 10, l. 15: This is the first time in the paper that the enhancement factor is mentioned. It should be defined and specified earlier in the paper (section 2.1).

Added in the description of the model:
“Like most ice sheet model, GRISLI uses a flow enhancement factor that favours longitudinal deformation in the SIA (Quiquet et al., 2018). However, here we use a flow enhancement factor set to 1 (no enhancement). Similarly, the flow enhancement factor for the SSA is also set to 1.”

P. 11, l. 18/19: This should be made a proper reference and cited here as Quiquet and Dumas (2020). And, BTW: _Z_enodo.

Done.

The GRISLI-LSCE contribution to ISMIP6, Part 1: projections of the Greenland ice sheet evolution by the end of the 21st century

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Abstract.

Polar amplification will result in amplified temperature changes in the Arctic with respect to the rest of the globe making the Greenland ice sheet particularly vulnerable to global warming. While the ice sheet has been showing an ~~increase~~ increased mass loss in the past decades, its contribution to global sea level rise in the future is of primary importance since it is at present the largest single source contribution ~~behind~~ after the thermosteric contribution. The question of the fate of the Greenland and Antarctic ice sheets for the next century has recently gathered various ice sheet models in a common framework within the Ice Sheet Model Intercomparison Project for CMIP6. While in a companion paper we present the GRISLI-LSCE contribution to ISMIP6-Antarctica, we present here the GRISLI-LSCE contribution to ISMIP6-Greenland. We show an important spread in the simulated Greenland ice loss in the future depending on the climate forcing used. The contribution of the ice sheet to global sea level rise in 2100 can be thus as low as 20 ~~mmSLE~~ mm of sea level equivalent (SLE) to as high as 160 ~~mmSLE. The~~ mm SLE. Amongst the models tested in ISMIP6, the CMIP6 models produce ~~much~~ larger ice sheet retreat than their CMIP5 counterparts. Low emission scenarios in the future drastically reduce the ice mass loss. The mass loss is mostly driven by atmospheric warming and associated ablation at the ice sheet margin while oceanic forcing contributes to about 10 ~~mmSLE~~ mm SLE in 2100 in our simulations.

1 Introduction

The relative contribution of land ice to global mean sea level rise has considerably increased in the recent decades, now larger than the thermosteric effect (Nerem et al., 2018). Amongst the different contributions, the Greenland and Antarctic ice sheets have ~~the largest~~ a potential to raise substantially the global mean sea level, with a ~~most-likely amplitude exceeding 1 metre in 2100 (Bamber et al., 2019)~~weakly constrained trajectory (Oppenheimer et al., 2019). While observational datasets show a dramatic increase in mass loss over the last decades for both ice sheets (Mouginot et al., 2019; Rignot et al., 2019), there is an urgent need for robust assessment of future sea level rise by projections obtained with numerical models.

Most of the time, these projections involve comprehensive ice sheet models that compute the ice thickness change that results from ~~changing boundary conditions~~ evolving forcings, such as climate change. On top of uncertainties related to future climate evolution, there are important differences amongst existing ice sheet models, and these differences represent a major source of uncertainty for the fate of the ice sheets in the future. First, in order to save computing time, most of the ice sheet models use various asymptotic ~~approximations~~ expansions (e.g. the shallow ice and shallow shelf approximations or higher-order models) even if~~more recently~~, more recently, models that account explicitly for all the stress components of the Stokes equation at the ice sheet scale have emerged (e.g. Seddik et al., 2012). This difference in ~~term~~ terms of ice sheet model complexity is a source of uncertainty for future projections. Second, ice sheets respond to a wide spectrum of timescales, from sub-annual to multi-millenial. As a result, diverse methodologies to initialise the models for projection purposes have been developed. For Greenland ice sheet models, these differences in methodologies lead to an even larger uncertainty for future projections than model complexity and explain most of the multi-model spread (Goelzer et al., 2018). A last source of uncertainty lies in poorly known processes, such as sub-glacial processes, or processes that are not included in models due to their complexity or too fine spatial scale, such as outlet glacier dynamics or fracturing. Large international intercomparison exercises are a useful way to quantify these different uncertainties and to infer robust sea level projections into the future.

The Ice Sheet Model Intercomparison Project for CMIP6 (ISMIP6, Nowicki et al., 2016), endorsed by the Coupled Model Intercomparison Project – phase 6 (CMIP6), aims at investigating the role of dynamic Greenland and Antarctic ice sheets in the climate system and to reduce the uncertainty in ice sheet contribution to global sea level rise in the future. Within this framework, stand-alone ice sheet model experiments have recently been carried out by world-wide research groups. Many model experiments using both CMIP5 and CMIP6 climate forcing scenarios until 2100 were conducted with ice sheet models spanning a range of model complexities and using different initialisation techniques. To date, this is the most ambitious intercomparison exercise dedicated to the fate of the Greenland and Antarctic ice sheets in the future. At the Laboratoire des Sciences du Climat et de l'Environnement (LSCE), we participated to this stand-alone intercomparison with the GRISLI model (Quiquet et al., 2018). This model uses the shallow ice and shallow shelf approximations and is relatively inexpensive in ~~term~~ terms of computational cost. We were thus able to perform all the different experiments of ISMIP6.

The aim of this paper is to discuss the role of the forcing uncertainties for future projections of the Greenland ice sheet contribution to global sea level rise when using our model. This individual model response can be put in perspective with respect to the multi-model spread discussed in Goelzer et al. (2020). This paper discusses additional experiments not included in the community paper (CMIP6 forcing and separate effects of the oceanic with respect to atmospheric forcing). Compared to Goelzer et al. (2020), we provide here a more detailed description of the initial state and its associated biases and model drift. A companion paper (Quiquet and Dumas, submitted) describes the results for the Antarctic ice sheet.

In Sec. 2 we describe briefly the GRISLI ice sheet model as well as the procedure used for its initialisation. We also provide information on the ISMIP6-Greenland forcing methodology and we provide an overview of the different experiments performed. In Sec. 3 we discuss the results for the different experiments in ~~term~~ terms of geometry and dynamical changes. We discuss these results in a broader context in Sec. 4 and we conclude in Sec. 5.

2 Methods

2.1 Model and initialisation

For this work, we use the GRISLI ice sheet model. The model is a 3D thermo-mechanically coupled ice sheet model that solve the mass ~~and momentum conservation~~ conservation and force balance equations. The model is fully described in Quiquet et al. (2018) and we only provide here a brief overview of its characteristics.

Assuming incompressibility, the mass conservation equation for a grid element is:

$$\frac{\partial H}{\partial t} = BM - \nabla \left(\bar{\mathbf{U}} H\right)$$

$$\frac{\partial H}{\partial t} = M - \nabla \left(\bar{\mathbf{U}} H\right), \tag{1}$$

with H the local ice thickness, ~~BM the~~ M the total mass balance and $\bar{\mathbf{U}}$ the vertically averaged horizontal velocity vector. $\nabla \left(\bar{\mathbf{U}} H\right)$ is thus the ice flux divergence.

The Stokes momentum equation is solved using asymptotic zero-order ~~approximations~~expansions. For the whole geographical domain, we assume that the total velocity is ~~simply to superposition of the~~ the sum of the velocities predicted by the two main approximations: the shallow ice approximation (SIA) in which the ~~horizontal derivatives of the stress tensor are neglected (vertical shear driven deformation)~~ deformation is entirely driven by the vertical shear and the shallow shelf approximation (SSA) in which the vertical ~~derivatives of the stress tensor are neglected (longitudinal stresses predominant)~~shear is neglected and the horizontal stresses are predominant. Practically, this means that we use the SSA equation as a sliding law ~~(Bueler and Brown, 2009)~~(Bueler and Brown, 2009; Winkelmann et al., 2011). Grounded cold base and floating shelves are special cases for which there is infinite ~~, respectively none,~~ friction at the base or none, respectively. Elsewhere, friction is assumed to follow a Weertman (1957) power law with a till layer that allows viscous deformation:

$$\tau_{\mathbf{b}} = -\beta\, \mathbf{U_b}$$

$$\tau_{\mathbf{b}} = -\beta\, \mathbf{U_b}, \tag{2}$$

where τ_b is the basal drag, β is the basal drag coefficient and $\mathbf{U_b}$ is the basal velocity. The basal drag coefficient is spatially variable but constant in time (except in specific cases such as during the inversion procedure).

Like most ice sheet model, GRISLI uses a flow enhancement factor that favours longitudinal deformation in the SIA (Quiquet et al., 2018) . However, here we use a flow enhancement factor set to 1 (no enhancement). Similarly, the flow enhancement factor for the SSA is also set to 1.

Similarly to what has been done with GRISLI for the initMIP-Greenland experiments (Goelzer et al., 2018), we used here an inverse procedure to initialise the model at the start of the historical experiment. We mostly followed the iterative method of

Le clec'h et al. (2019b) which consists at yielding the map of the basal drag coefficient β that minimises the ice thickness error with respect to observations. To this aim, we first run a 30 kyr experiment with fixed topography and perpetual present-day climate forcing in order to compute the thermal state of the ice sheet in agreement with the boundary conditions. From this, we do multiple 200-yr long experiments under constant present-day climate forcing but with an evolving topography. During the first 20 years of these experiments, we adjust the basal drag coefficient to minimise the ice thickness mismatch with respect to the observations. Each iteration starts from the exact same initial condition, except that the basal drag coefficient is different. Also, the ice thickness error at the end of the 200-yr long experiment is used to facilitate convergence towards the observed ice thickness through a local basal drag modification. This ~~modification on the basal drag consist~~ basal drag coefficient modification consists in finding an ice flux on the simulated topography as close as possible to the balance ice flux on the observed topography. After a few 200-yr experiments, we repeat the thermal equilibrium computation ~~so that it accounts for the~~ restarting from the previous equilibrium state with the newly inferred basal drag coefficient. In doing so, the basal drag coefficient and the temperature at the base are consistent with each other. For this work we performed more than ten thermal equilibrium experiments, each one followed by five iterations of 200 years.

At the end of the iterative process, we use the last inferred basal drag coefficient together with the corresponding thermal state to run a short relaxation experiment of 20 years. The end of this relaxation experiment defines our initial state which is used to begin the historical experiment *hist* and the control experiment *ctrl* (see Sec. 2.3).

Our ice thickness and bedrock topography of reference is the BedMachine v.3 (Morlighem et al., 2017). This dataset is used as a target for our iterative procedure to infer the basal drag coefficient. It is also used as the starting topography for the short relaxation that defines our initial state. Our present-day reference climate forcing, namely annual near-surface air temperature and annual surface mass balance, comes from the MAR v3.9 (Fettweis et al., 2013, 2017) forced at its boundary by MIROC5, averaged over the 1994-2015 period. On top of this climate forcing, we also add a strongly negative surface mass balance term of -15 m yr^{-1} outside the present-day ice mask in the observational dataset in order to avoid inconsistencies between the climate forcing and the initial ice sheet geometry. This present-day reference climate forcing is used for the initialisation procedure and for the control experiment *ctrl*. The model is run on a Cartesian grid at 5 km resolution covering the Greenland ice sheet using a stereographic projection. Since 5 km is too coarse to represent Greenland floating ice tongues, sub-shelf melting rate has been set to a large value (200 m yr^{-1}) to discard simulated floating points. Glacial isostatic adjustment has been deactivated for all the experiments shown in this manuscript.

2.2 ISMIP6-Greenland forcing methodology

The ISMIP6-Greenland working group distributed atmospheric and oceanic forcings to drive individual ice sheet models. They also suggest a forcing methodology so that participating models are run using a common framework. Full description of the methodology is available in Nowicki et al. (2020) and only a summary is presented here.

For the atmospheric forcing, MAR v3.9 has been run from 1950 until 2100 forced at its boundaries by a selection of CMIP5 and CMIP6 general circulation model (GCM) outputs. To force the ice sheet models, yearly anomalies of near-surface air temperature and surface mass balance are provided. These anomalies were constructed as the difference of a given yearly value with the climatology over the reference period 1960-1989. In addition, to account for the surface elevation feedback on temperature and surface mass balance, yearly values of ~~atmospheric gradients~~ vertical gradients in the atmospheric model for these two surface variables are also provided. These spatially variable gradients were evaluated with the MAR model with the method of Franco et al. (2012).

Ice–ocean interactions for the Greenland ice sheet is most of the time poorly represented amongst ISMIP6-Greenland participating models. This is mostly due to the fact that the spatial scale needed to represent such interactions is out of reach for most models. This is also the case for GRISLI, where the 5 km resolution grid is too coarse to capture marine-terminating outlet glaciers. To cope with this problem, retreat masks for outlet glaciers have been available in ISMIP6-Greenland. They were obtained with simple parametrisations calibrated and tested against observational datasets (Slater et al., 2019). These masks provide, for a given resolution, the fraction of the grid that becomes ice free and they are used to impose a specific retreat rate of the marine front. For each climate forcing, three retreat masks are available for different oceanic sensitivities (*low*, *medium* and *high*). Since our model does not account for partially glaciated grid cell, the fractional information given by the retreat masks is used to reduce the local ice thickness with respect to a reference ice thickness (i.e. the ice thickness evolution for the outlet glaciers is imposed). The reference ice thickness could have been chosen as the ice thickness at a specific time (e.g. the ice thickness at the end of the historical experiment). However, in doing so, we may create strong discontinuities in ice thickness when the retreat mask is used for the first time. For this reason, we choose instead the value of the local ice thickness at the time when the imposed retreat starts to play as a reference ice thickness.

2.3 List of experiments

The ice sheet state inferred at the end of the initialisation procedure (Sec. 2.1) is used as initial condition for the historical experiment *hist*. In our case, the historical experiment starts in January 1995 and ends in December 2014. For this historical experiment, we use the climate forcing of MAR forced at its boundary by the MIROC5 climate model. The projection experiments described in the following are all branched ~~to~~ from the end of the year 2014 of this historical experiment and span 2015-2100 (86 simulated years).

ISMIP6-Greenland listed a large ensemble of experiments to be performed with individual ice sheet models (Tab. 1). The ensemble of experiments is large enough to assess: ice sheet sensitivity to the chosen climate forcing, CMIP5 with respect to CMIP6, sensitivity to the greenhouse gas emissions scenarios, the respective role of oceanic forcing with respect to atmospheric forcing, and to quantify the uncertainty regarding the oceanic forcing. The core experiments consist (Tier 1) in a selection of three CMIP5 climate models (MIROC5, NorESM and HadGEM2-ES) run under the RCP8.5 scenario for greenhouse gases.

In addition, MIROC5 was chosen to be run with a different RCP scenario (RCP2.6) and using different oceanic sensitivities (*high* and *low* in addition to *medium*). Tier 2 has two subsets: an extended ensemble with three additional CMIP5 models using RCP8.5 and an other with four CMIP6 models. Amongst CMIP6 models, CNRM-CM6 has been run under two scenarios: a ~~pessimistic~~ high (SSP585) and ~~an optimistic~~ a low (SSP126) emission scenario. Tier 3 has also two subsets. The first one aims at quantifying the respective role of the oceanic forcing with respect to atmospheric forcing, running the ice sheet models only with one of this forcing at a time. Three climate models were selected (MIROC5, CSIRO-Mk3.6 and NorESM) and as in Tier 1, MIROC5 was run for two greenhouse gases scenarios and different oceanic sensitivities. Finally, the second subset of Tier 3 contains the ten climate models (CMIP5 and CMIP6) each time run with the two additional oceanic sensitivities (*high* and *low*). CNRM-CM6 is the only one in this subset that has run under two emission scenario (SSP585 and SSP126). We performed all these experiments with the GRISLI ice sheet model.

In addition to the these projection experiments, we also perform two control experiments in which the climate forcing remains unchanged, being our reference climate forcing used during the initialisation procedure (zero anomaly). The control experiment *ctrl* starts from the initial state resulting from our initialisation procedure and covers the 1995-2100 period (106 years). The *ctrl_proj* experiment starts in January 2015, ~~alike~~ like the projection experiments, and runs for 86 years under a constant climate forcing. The *ctrl* experiment can be used to quantify the simulated model drift over the whole time period (1995-2100). Instead, the *ctrl_proj* can be directly used to quantify the importance of climate forcing evolution since it uses the same initial state in 2015 as the different projection experiments.

3 Results

We aim here at providing a detailed description of the historical experiment *hist* and the model response under the various forcings of the projection experiments. While some information is given in this section, the reader is invited to refer to Goelzer et al. (2020) to compare in details the response of GRISLI to other participating models. A map of Greenland with the names of the major ice streams discussed in the following is shown in Fig. 1.

3.1 Present-day simulated ice sheet

At the end of the historical experiments *hist*, with a value smaller than 30 m, GRISLI shows the lowest ice thickness root mean squared error (RMSE) with respect to the observations of Morlighem et al. (2017) amongst the ISMIP6-Greenland participating models (Goelzer et al., 2020). This is a result of the initialisation procedure we use (Sec. 2.1) that includes only a short relaxation of 20 years. With an historical experiment of 20 years only, the model has no time to depart strongly from the observations. The map of the ice thickness difference with respect to observations is shown in Fig. 2a. The model shows a very good agreement with the observations for most of the ice sheet, except at specific locations at the margin. In particular, South-East Greenland is the least well reproduced with local errors greater than 200 metres. In the region of Kangerdlugssuaq and Helheim glaciers, there is an ice thickness overestimation near the glacier termini and an underestimation upstream. These differences with the observations can be due to the fact that this area is particularly difficult to model since it has a complex surface mass balance pattern with very strong horizontal gradients and also a rough topography that is not necessary well captured at 5 km resolution.

Some of the ice thickness mismatch with respect to the observations can be partly explained by error related to ice dynamics. GRISLI has indeed an ice velocity RMSE with respect to the observations (Joughin et al., 2016) of about 35 m yr^{-1}, making the model the sixth worst model out of 21 (Goelzer et al., 2020). Our initialisation procedure favours a good match of the simulated ice thickness with respect to observations but it does not include any constraints on the ice velocity. It is thus not particularly surprising that GRISLI performs ~~best in term~~ better in terms of ice thickness than in ~~term~~ terms of ice velocity. Since ice velocity is a very heterogeneous variable, it is sometimes convenient to use the logarithm of the velocity instead of the absolute velocity. ~~In doing so, the RMSE is about 0.55 (eleventh worst value out of 21). In~~ When using the logarithm of the velocity ~~,~~ GRISLI slightly improves compared to the other participating models ~~, meaning~~ since the RMSE is about 0.55 (eleventh worst value out of 21). This means that the errors are mostly localised in areas of high velocities. Fig. 3a shows the absolute simulated velocity, to be compared to the observations in Fig. 3b. The pattern is generally well reproduced and the model is able to reproduce the localisation of the major existing ice streams. However, the velocity of the ice streams is not always in agreement with the observational data. The ~~Northern and Western~~ northern and western ice streams are generally too slow with an underestimation reaching more than 500 m yr^{-1} for the Jakobshavn, Petermann and North East Greenland ice stream (NEGIS) glacier termini (Fig. 3c). On the contrary, the South East glaciers, Kangerdlugssuaq and Helheim, are too ~~slow~~ fast in the model. For the northern and western regions, the errors in ice thickness are small meaning that the ice velocity mismatch there cannot be reduced within our initialisation procedure which only minimises the ice thickness error. This is somewhat different for the ~~South Eastern~~ south-eastern region, where there are important errors in ice thickness. However, there is an important positive bias in ice thickness at the ice sheet margin that tends to produce very high ice flow (very low basal drag coefficient to reduce this bias). Since the SSA equation is elliptic, the low basal drag at the margin has a regional impact on ice flow, which tends to produce an underestimation of the ice thickness further inland. While we strongly overestimate the velocity in this area, the ice thickness at the margin is still overestimated. This suggests that the surface mass balance used in our reference climate is probably overestimated in this region.

The ice sheet model drift can be assessed by examining Fig. 2c. The ice thickness drift in the control experiment *ctrl_proj* is generally very small (lower than 10 metres) with only a few regions with higher values. Here again, the Kangerdlugssuaq and Helheim glacier regions show the largest model drift with a local increase in ice thickness of more than 100 metres near the glacier termini. Overall the ice ~~volume~~ mass drift is negligible over the duration of the control experiment (86 years), also because of some compensating biases (see also Fig. 4). In addition to ~~a~~ the ice thickness drift, the model simulates a drift in velocities during the duration of the control experiment (Fig. 3d). For most of the ice sheet the velocity change is small and only reaches more than 1 m yr^{-1} at the ice sheet margins. The largest changes concern the glaciers in South-East Greenland such as the Helheim and the Kangerdlugssuaq glaciers where locally, at the termini, there can be an ~~acceleration~~ increase in velocity by more than 1000 m yr^{-1}.

3.2 Ice sheet evolution projections

3.2.1 Sensitivity to climate forcing

Amongst the different experiments, we start ~~by~~ with the description of the simulated ice sheet evolution under the RCP8.5 scenarios for the 6 available CMIP5 models (Tier 1 and Tier 2). The simulated ~~volume~~ total ice mass evolution over the 1995-2100 period is shown in Fig. 4 (expressed in total ~~ice volume~~ mass and in contribution to global sea level rise). In 2100, the total ice loss ranges from about ~~-15000 km^3 to -35000 km^3~~ -15 to -35×10^3Gt. This translates to a Greenland ice sheet melt contribution to global sea level rise of 35 mm of sea level equivalent (~~mmSLE~~mm SLE) to 80 ~~mmSLE. The~~ mm SLE. The 2100 sea level contribution simulated by GRISLI is close to the mean model response amongst the ISMIP6 participating models (Goelzer et al., 2020). The spread amongst the different climate forcings of about ~~20000 km^3~~ 20×10^3Gt (or 45 ~~mmSLE~~mm SLE) is thus larger than the ~~ice volume~~ mass change yielded with the GCM providing the smallest ice sheet response (CSIRO-Mk3.6). The evolution of ice loss over the 86 simulated years is not linear, with an acceleration for all climatic scenarios. However, we can not distinguish any ~~tipping point~~ sharp inflexion in the total mass evolution over the next century. The differences in mass evolution are tightly linked to the surface mass balance evolution for the different climate forcing. Amongst the CMIP5 climate models, IPSL-CM5-MR and MIROC5 simulate a mean surface mass balance negative as early as 2060 while it remains positive over the next century for CSIRO-Mk3.6 (Fig. 5).

CMIP6 models show generally a much larger Earth climate sensitivity than their equivalent in the former CMIP5 generation (Forster et al., 2020). In particular, the CMIP6 models used in ISMIP6 have an Earth climate sensitivity from 4.8 to 5.3, i.e. larger than the CMIP5 models used here, which show a range from 2.7 to 4.6 (Meehl et al., 2020). This has important consequences on the projected Greenland ice sheet. The ~~ice volume~~ total ice mass evolution for the four CMIP6 models under SSP585 scenario is shown in Fig. 6. The CMIP6 models produce systematically higher ice loss than the CMIP5 models. The two most sensitive CMIP6 models (UKESM1-CM6 and CESM2) almost double the ice loss with respect to the most sensitive CMIP5 models (IPSL-CM5-MR and MIROC5). The ice loss thus reaches ~~-60000 km^3~~ -60×10^3Gt (140 ~~mmSLE~~mm SLE) by the end of the century.

Two climate models have been run under two scenarios for the evolution of future atmospheric greenhouse gases. The ice loss for the two scenarios of the climate models is shown in Fig. 7. The CMIP5 (MIROC5) and CMIP6 (CNRM-CM6) responses to the change in greenhouse gas scenario (RCP8.5 to RCP2.6 and SSP585 to SSP126 respectively) is very similar. There are very small differences for the first half of the century but after 2060 the ~~pessimistic~~ high emission scenario produces substantial additional mass loss with respect to the ~~optimistic scenarios~~low emission scenario. By the end of the century, the ~~pessimistic greenhouse gas~~ high emission scenario produces roughly ~~-25000 km^3~~ -25×10^3Gt (55 ~~mmSLE~~mm SLE) of additional ice loss with respect to the ~~optimistic~~ low emission scenario. The future ~~evolution of atmospheric~~ atmospheric and oceanic warming induced by the greenhouse gas mixing ratio is thus a major driver for the Greenland ice mass loss at the century time scale.

The spatial pattern of ice loss by the end of this century is shown in Fig. 8. For this figure we have chosen four projection experiments that show contrasted integrated ice mass loss by 2100: the CISRO-Mk3.6 under RCP8.5 which produces a small integrated ice loss (Fig. 8a), the MIROC5 under RCP8.5 which produces an important mass loss (Fig. 8b), the MIROC5 under RCP2.6 to show the impact of low emission scenario (Fig. 8c) and UKESM-CM6 under SSP585 with a high oceanic sensitivity which produces the highest mass loss amongst all the different experiments (Fig. 8d). While the amplitude of ice thickness change is drastically different amongst these experiments, the spatial pattern is similar. The major signal is a substantial widespread ice thickness decrease at the margin of the ice sheet. If the ice thickness decrease is about 50 m for the least sensitive model (CSIR-Mk3.6), it can reaches more than 200 m for the most sensitive model (MIROC5 or IPSL-CM5-MR). The south-western region shows the largest ice sheet thinning. On the contrary, the central region shows a slight increase in ice thickness which can reach about 50 m at places for the most sensitive climate scenario. This increase in ice thickness is related to the slight increase in precipitation simulated by some GCMs in the course of the century. The central eastern region shows only limited ice thickness changes regardless of the climate forcing used. The use of the RCP2.6 emission scenario reduces drastically the ice thickness changes.

3.2.2 Importance of the oceanic forcing

The uncertainty that arises from the oceanic forcing can be evaluated thanks to the different glacier retreat scenarios (*low*, *medium* and *high* sensitivity to oceanic forcing). In Fig. 4 is represented on the right-hand side the uncertainty that arises from the oceanic forcing for the individual CMIP5 models. In 2100 the ice ~~volume~~ mass loss difference between the *low* and *high* oceanic sensitivities is generally of about ~~-5000 km^3~~ -5$\times$10^{3}Gt (less than 10 ~~mmSLE~~mm SLE). Without being negligible, the oceanic sensitivity for a given climate scenario is nonetheless relatively small compared to the spread amongst the different CMIP5 climate models used. For the CMIP6 experiments, the uncertainty that comes from the oceanic forcing is almost doubled with respect to the CMIP5 experiments, with about ~~10000 km^3~~10$\times$10^{3}Gt (~20 ~~mmSLE~~mm SLE) of ice loss difference from the *low* to *high* oceanic sensitivity (Fig. 6) but these CMIP6 models also produce much greater ice loss.

We also performed experiments in which we isolate the response of the model that arises from the atmospheric forcing only (first subset of Tier 3). For the *atmosphere only* (AO) experiments, we do not impose a retreat rate for the outlet glaciers and only the atmospheric perturbation is taken into account. Conversely, for the *ocean only* (OO) experiments, there is no atmospheric perturbation (as in the control *ctrl_hist* experiment) but we do impose a retreat rate for the outlet glaciers. The ice ~~volume~~ mass evolution for these experiments is shown in Fig. 9. The OO experiments produce almost identical ~~ice volume~~ mass evolutions amongst the different GCMs. This means that even if the glacier retreat is subject to uncertainties, with the methodology of Slater et al. (2019) it is nonetheless only weakly sensitive to the differences in the climate forcing used to elaborate it. Fig. 9 also shows that the atmospheric forcing is the main driver for ice loss for the GCMs that produce important ice loss. Also, the sum of the ice loss of AO and OO experiments approximate closely the ice loss simulated when using the full forcing (92 to 94% of the full forcing).

3.2.3 Change in ice dynamics

Climate forcing, and its associated ice sheet geometry change, leads to a change in the dynamics of the Greenland ice sheet. Fig. 10a shows the change in the simulated surface velocities at the end of the century with respect to the year 2015 for a given climate forcing. On the one hand, consistently with what has been found in previous studies (e.g. Peano et al., 2017; Le clec'h et al., 2019a), there is a decrease in simulated velocities related to ice thinning at the margins. On the other hand, the increase in surface slopes due to ice thinning at the margin leads to increased velocities further upstream.

The change in ice dynamics can also be assessed by investigating the different terms of the mass conservation equation. The integration in time of Eq. 1 over 2015-2100 suggests that the integrated ice flux convergence is the difference between the ice thickness change from 2015 to 2100 and the integrated mass balance (surface and basal mass balance and calving) over this period. The integrated ice flux convergence can be considered as the dynamical contribution to ice thickness change. It should be noted that the integrated mass balance here also includes the effect of ice mask change and surface elevation change. As such, it is not comparable to what would have been obtained with an atmospheric model only. Fig. 10b shows the difference ~~in ice flux convergence in~~ of the dynamical contribution in 2100 for a selected climate forcing with respect to the control *ctrl_proj* experiment. ~~This can be considered as the dynamical contribution to ice thickness change.~~ The pattern mostly follows the one of velocity change (Fig. 10a). There is an important ~~ice flux convergence~~ positive dynamical contribution to ice thickness change (ice flux convergence) at the margins that tends to partially compensate the decrease in surface mass balance. Conversely, upstream regions show a slightly negative ~~convergence~~dynamical contribution (ice flux divergence). This pattern is similar amongst the different climate forcings.

4 ~~Discussion~~

~~In order to minimise the initial error in ice thickness~~ To compare the relative importance of the dynamical contribution with respect to surface mass balance to explain the ice thickness change we show the ice thickness change in 2100 with the same colour scale in Fig. 10c. The dynamical contribution shows generally much smaller values suggesting that surface mass balance explains the largest changes in ice thickness. However, locally, for example in the ~~observations, we have used an inverse procedure that optimally tune the basal drag coefficient. In doing so, we produce a simulated ice sheet that is in quasi-equilibrium with the climate forcing (minimal ice thickness drift).In reality, the Greenland ice sheet is far from being at equilibrium with present-day climate since it has been loosing ice at an accelerated rate over the last four decades (Mouginot et al., 2019). This means that , by constructions, our simulations underestimate the Greenland ice sheet contribution~~

~~to future sea level rise. Some promising alternatives exist, for example using data assimilation of observed velocities in a transient ice sheet simulation (Gillet-Chaulet, 2020). These methods require however a complex data assimilation framework, currently not implemented in our ice sheet model. Instead, we plan to modify the inverse procedure of Le clec'h et al. (2019b) by incorporating the ice thickness change inferred by gravimetry as an additional constraint in order to improve on the initial state of the Greenland ice sheet~~South-East and central West regions the dynamical contribution can be the largest driver of ice thickness change.

~~In addition, the~~ The inferred basal drag coefficient during the initialisation procedure is left unchanged for the duration of the historical and projection experiments. This is probably an important and unjustified approximation since the basal conditions are susceptible to respond to changes in ice geometry and, eventually, basal hydrology. To assess the importance of basal drag coefficient changes for our projections, we perform a new set of experiments using the MIROC5 climate forcing under RCP8.5 with a medium oceanic sensitivity. For these simulations, we apply a spatially uniform modification factor to reduce or increase the value of the basal drag coefficient after the year 2045. The ~~ice volume~~ modification ranges from -90% (reduction to 10% of the initial value) to +100% (doubling of the initial value). The total ice mass difference in 2100 with respect to the experiment with no modification of the basal drag coefficient is shown in Fig. 11a,b. The ~~ice volume~~ mass change in response to small perturbations of the basal drag coefficient is relatively linear and limited. Thus, a perturbation of 20% results in less than ~~5000 km^{3}volume~~ 5×10^3Gt mass change, which translates to less than 10 ~~mmSLE~~mm SLE. This means that it is unlikely that basal condition changes in the future could produce a drastically different ~~ice volume~~ total ice mass change in 2100. This also suggests that a slightly different basal drag coefficient inferred during our initialisation procedure will produce a similar ~~ice volume~~ mass evolution in the projection experiments. ~~We repeat this kind of sensitivity experiment for the~~ In order to further assess the sensitivity of our projections to the choice of mechanical parameters, we repeated these perturbation experiments for the SIA flow enhancement factor (Fig. 11c,d). ~~Here again~~We varied the enhancement factor from 0.4 to 6 with respect to the standard value of 1. As for the basal drag coefficient perturbation, the response in ~~term of ice volume~~ terms of ice mass loss is small and relatively linear.

To assess the range of acceptable values for the basal drag coefficient perturbation and the enhancement factor, we also performed similar sensitivity experiments to the control experiment *ctrl_proj*. The range of acceptable perturbations is thus defined as the perturbed control experiments that produce less than 0.1% total mass change with respect to the standard control experiment. 0.1% total mass change corresponds to one tenth of the total mass change in 2100 with respect to 2015 using the MIROC5 climate forcing under RCP8.5 with a medium oceanic sensitivity. The acceptable perturbations of the basal drag coefficient range from -15% to 20% and the acceptable enhancement factors range from 0.8 to 1.2. Interestingly, the effect of the perturbations (basal drag coefficient or enhancement factor) on the mass change is almost identical for the projection experiments (blue dots in Fig. 11) and for the control experiments (light blue dots in Fig. 11). This means that, in our model, different mechanical parameters do not enhance nor mitigate the mass loss due to climate change.

4 Discussion

In order to minimise the initial error in ice thickness with respect to the observations, we have used an inverse procedure that optimally tune the basal drag coefficient. In doing so, we produce a simulated ice sheet that is in quasi-equilibrium with the climate forcing (minimal ice thickness drift). In reality, the Greenland ice sheet is far from being at equilibrium with present-day climate since it has been loosing ice at an accelerated rate over the last four decades (Mouginot et al., 2019). This means that, by constructions, our simulations underestimate the Greenland ice sheet contribution to future sea level rise. A simple linear extrapolation of the 2006-2016 rate (0.77 mm yr^{-1}, Oppenheimer et al., 2019) up to 2100 would result in a 6.5 cm SLE from the Greenland ice sheet. This number is large compared to the GRISLI results discussed in this paper, and more generally it is large compared to the spread amongst ISMIP6 models (3.5 to 14 cm SLE, Goelzer et al., 2020). This suggests that model initialisation is one of the largest source of uncertainty for model projections. Instead of using a methodology that produces ice sheet at equilibrium, some promising alternatives exist, for example using data assimilation of observed velocities in a transient ice sheet simulation (Gillet-Chaulet, 2020). These methods require however a complex data assimilation framework, currently not implemented in our ice sheet model. Instead, we plan to modify the inverse procedure of Le clec'h et al. (2019b) by incorporating the ice thickness change inferred by gravimetry/altimetry as an additional constraint in order to improve on the initial state of the Greenland ice sheet.

One additional limitation of the inverse procedure is that it does not take into account the impact of the last glacial cycle on ice temperatures. Our internal temperature field is the result of a long thermo-mechanical equilibrium under perpetual present-day forcing and as such, it is necessarily overestimated since the diffusion in the ice sheet of the cold temperature of the glacial period is not accounted for. In addition to an underestimated ice viscosity, this has also consequences on the simulated basal temperature and, as a result, on the regions where sliding occurs. This might affect the dynamical response of the model to future climate change. If earlier studies have already identified these limitations (e.g. Rogozhina et al., 2011; Yan et al., 2013; Seroussi et al., 2013; Le clec'h et al., 2019b), our inverse procedure does not allow for a quantification of these limitations. Given the relatively low computational cost of GRISLI, one alternative would be to perform muti-millenial palaeo integrations to infer the initial state used for the projections. This strategy generally leads to a larger ice thickness error with respect to present-day observations but has the advantage to have a thermal state consistent with the model physics and with the palaeo temperatures. While the ISMIP6-Greenland participating models either choose one or the other initialisation technique (Goelzer et al., 2020), it would be very informative to have two drastically different initialisation methods for a given ice sheet model.

~~Finally, the~~ The forcing methodology used for ISMIP6-Greenland does account for the vertical elevation feedback on temperature and surface mass balance. ~~However, other feedbacks are~~ In order to quantify the impact of this correction on the simulated evolution of the ice sheet, we run a sensitivity experiment in which this correction is not accounted for. Using MIROC5 under RCP8.5 scenario with a medium oceanic sensitivity, we simulate a Greenland contribution to future sea level rise 5.1%

smaller in this sensitivity experiment compared to the same experiment in which the vertical correction is applied. This number is slightly higher than the effect reported by Edwards et al. (2014) and Le clec'h et al. (2019a) (4.3 and 4.2 % respectively) but smaller to Vizcaino et al. (2015) (8-11%) and Calov et al. (2018) (about 13%). Differences in resolution and/or physical processes implemented in the atmospheric model could explain this diversity.

In addition to the vertical elevation feedback on surface mass balance, other feedbacks at play for the future evolution of the Greenland ice sheet are not accounted for in the ISMIP6-Greenland methodology. Notably, the MAR model used to compute the forcing fields does not account for topography and ice mask changes. The effect of these changes is probably limited for moderate ice sheet retreat (Le clec'h et al., 2019a). However, since the CMIP6 models used here produce a much greater retreat than the CMIP5 models, they could also induce more important feedbacks if MAR was bi-directionally coupled to an ice sheet model. In addition, the effect of Greenland ice loss on the ocean is also not taken into account with the forcing methodology followed. While, the oceanic forcing seems not to be the major driver for future Greenland ice loss, glacier retreat in the future should ideally take into account the oceanic circulation changes in the fjords related to freshwater discharge from ice sheet melting.

5 Conclusions

In this paper we have presented the GRISLI-LSCE contribution to ISMIP6-Greenland. Independently from the climate forcing used to drive the ice sheet model, we have shown that the Greenland ice sheet ~~is systematically loosing~~ systematically looses ice in the future. However, the magnitude of the mass loss by 2100 is very sensitive to the climate forcing. Under a *business as usual* scenario for the greenhouse gas emission (RCP8.5 or SSP585), the mass loss translates into a Greenland ice sheet contribution to global sea level rise that ranges from 35 to 160 ~~mmSLE~~mm SLE. However, with ~~an optimistic greenhouse gas emission scenario~~ a low emission scenario for greenhouse gases (RCP2.6 or SSP126) the mass loss can be significantly reduced. The CMIP6 models selected for ISMIP6 tend to produce ~~much~~ a larger ice loss due to their higher climate sensitivity with respect to the one of the CMIP5 models. While the oceanic forcing contributes to ice loss by about 10 ~~mmSLE~~ mm SLE in 2100, the Greenland ice mass loss in the future is mostly driven by a larger ablation at the ice sheet margin~~in the future~~. This suggests that this process should be carefully implemented in ice sheet models aiming at simulating the Greenland ice sheet evolution at the century scale.

6 Data availability

The GRISLI outputs from the experiments described in this paper are available on the ~~zenodo~~ Zenodo repository with digital object identifier 10.5281/zenodo.3784665 (Quiquet and Dumas, 2020). The outputs in the ~~zenodo~~ Zenodo repository are the standard GRISLI outputs on the native 5 km grid and, as a result, they may slightly differ from the post-processed outputs available on the official CMIP6 archive on the Earth System Grid Federation (ESGF). In order to document CMIP6's scientific impact and enable ongoing support of CMIP, users are obligated to acknowledge CMIP6, the participating modelling groups, and the ESGF centres (see details on the CMIP Panel website at http://www.~~wcrp- climate~~wcrp-climate.org/index.php/wgcm-cmip/about-cmip). The forcing datasets are available through the ISMIP6 wiki and are also made publicly available via https://doi.org/xxx.

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Table 1. List of ISMIP6-Greenland experiments performed in this work.

exp_id	scenario	GCM	Ocean	
exp05	RCP8.5	MIROC5	Medium	Core experiments – Tier 1
exp06	RCP8.5	NorESM	Medium	
exp07	RCP2.6	MIROC5	Medium	
exp08	RCP8.5	HadGEM2-ES	Medium	
exp09	RCP8.5	MIROC5	High	
exp10	RCP8.5	MIROC5	Low	
expa01	RCP8.5	IPSL-CM5-MR	Medium	Extended ensemble – Tier 2
expa02	RCP8.5	CSIRO-Mk3.6	Medium	
expa03	RCP8.5	ACCESS1.3	Medium	
expb01	SSP585	CNRM-CM6	Medium	CMIP6 extension – Tier 2
expb02	SSP126	CNRM-CM6	Medium	
expb03	SSP585	UKESM1-CM6	Medium	
expb04	SSP585	CESM2	Medium	
expb05	SSP585	CNRM-ESM2	Medium	
expc01	RCP8.5	MIROC5 AO	Medium	Ocean only (OO) and Atmos. only (AO) – Tier 3
expc02	RCP8.5	MIROC5 OO	Medium	
expc03	RCP8.5	CSIRO-Mk3.6 AO	Medium	
expc04	RCP8.5	CSIRO-Mk3.6 OO	Medium	
expc05	RCP2.6	MIROC5 AO	Medium	
expc06	RCP2.6	MIROC5 OO	Medium	
expc07	RCP8.5	NorESM AO	Medium	
expc08	RCP8.5	NorESM OO	Medium	
expc09	RCP8.5	MIROC5 AO	Low	
expc10	RCP8.5	MIROC5 OO	High	

exp_id	scenario	GCM	Ocean	
expd01	RCP8.5	NorESM	High	Ocean sensitivity – Tier 3
expd02	RCP8.5	NorESM	Low	
expd03	RCP8.5	HadGEM2-ES	High	
expd04	RCP8.5	HadGEM2-ES	Low	
expd05	RCP8.5	MIROC5	High	
expd06	RCP8.5	MIROC5	Low	
expd07	RCP8.5	IPSL-CM5-MR	High	
expd08	RCP8.5	IPSL-CM5-MR	Low	
expd09	RCP8.5	CSIRO-Mk3.6	High	
expd10	RCP8.5	CSIRO-Mk3.6	Low	
expd11	RCP8.5	ACCESS1.3	High	
expd12	RCP8.5	ACCESS1.3	Low	
expd13	SSP585	CNRM-CM6	High	
expd14	SSP585	CNRM-CM6	Low	
expd15	SSP126	CNRM-CM6	High	
expd16	SSP126	CNRM-CM6	Low	
expd17	SSP585	UKESM-CM6	High	
expd18	SSP585	UKESM-CM6	Low	
expd19	SSP585	CESM2	High	
expd20	SSP585	CESM2	Low	
expd21	SSP585	CNRM-ESM2	High	
expd22	SSP585	CNRM-ESM2	Low	

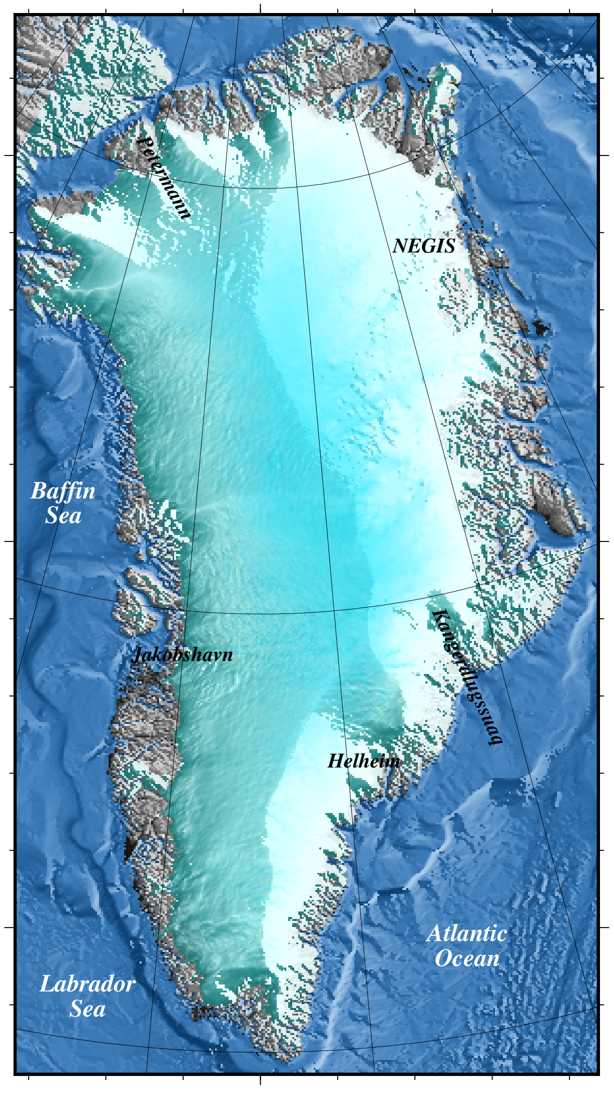

Figure 1. The Greenland ice sheet with the major ice streams discussed in the text.

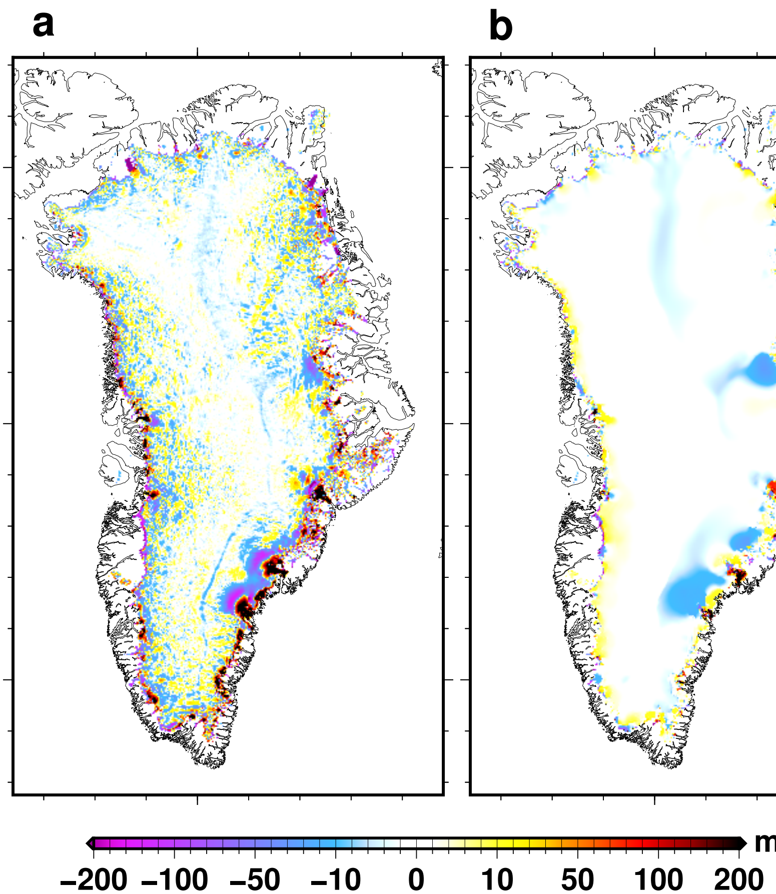

Figure 2. Ice thickness difference: **(a)** end of the historical experiment *hist* ~~respective~~ (2015) with respect to the observations (Morlighem et al., 2017); **(b)** end of the control experiment *ctrl_proj* (2100) respective to the end of historical experiment *hist* (2015).

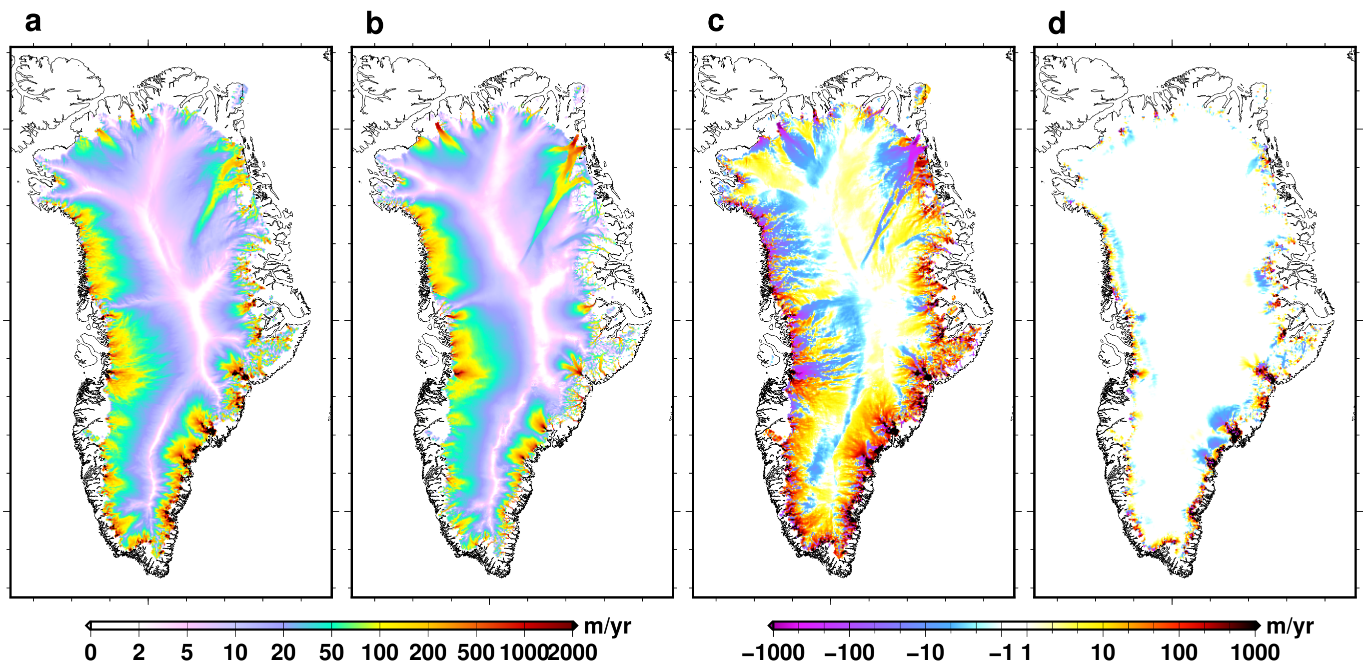

Figure 3. Surface velocity magnitude: **(a)** simulated at the end (2011-2015) of the historical experiment *hist*; **(b)** in the observational datasets of Joughin et al. (2016); **(c)** difference between (a) and (b). The surface velocity magnitude change from 2011-2015 to 2096-2100 in the control experiment *ctrl_proj* is shown in **d**. We use a 5 year mean for the simulated velocity to reduce the impact of interannual variability. The range -1 to 1 m yr^{-1} is set to white for velocity difference (**c** and **d**).

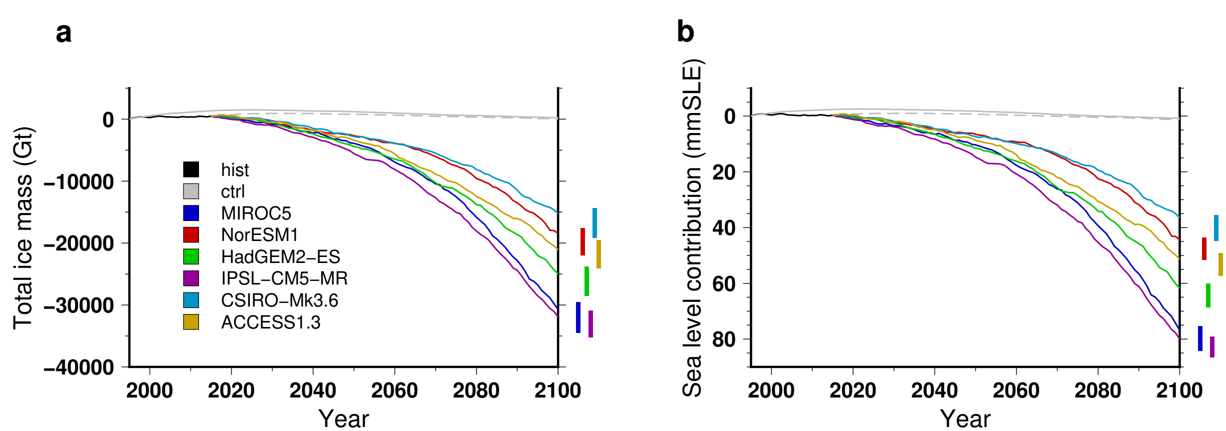

Figure 4. Simulated total ice ~~volume~~ mass change for the historical simulation *hist* (1995-2015), the control experiments *ctrl* (solid grey lines) and *ctrl_proj* (dashed grey lines) and the projections under the different CMIP5 forcings using the RCP8.5 scenario and the medium oceanic sensitivity: **(a)** total ~~ice volume~~ mass change and **(b)** ice volume contributing to sea level rise. For each projection experiment the right-hand side vertical bar shows the minimal and maximal ~~volume change~~ changes associated with the oceanic forcing uncertainty (*low* and *high* scenarios).

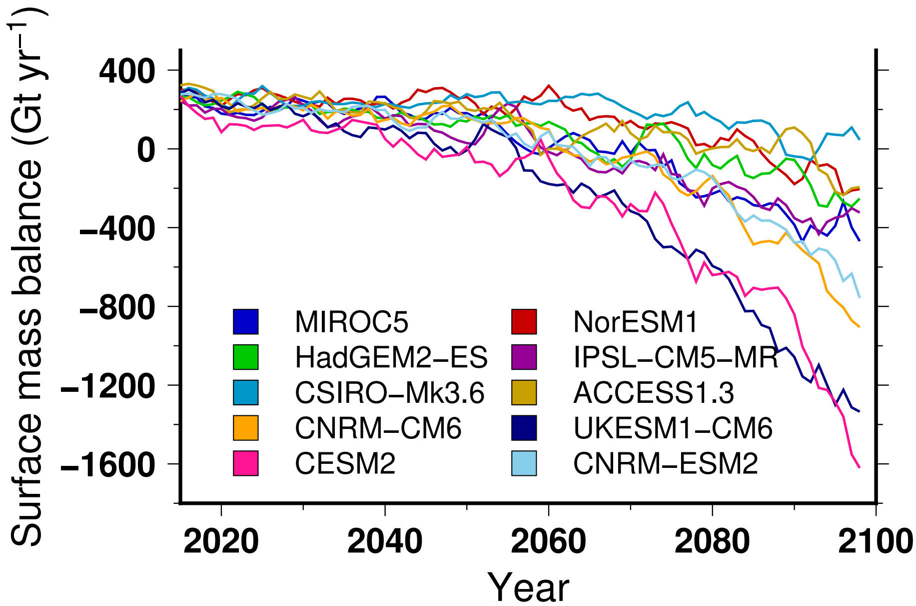

Figure 5. Simulated surface mass balance, integrated over the ice sheet, for different CMIP5 and CMIP6 climate forcings using the RCP8.5 scenario and SSP585 scenario, respectively. The projection experiments shown in this figure use the medium oceanic sensitivity. For this figure we use a 5-year running mean in order to smooth the interannual variability.

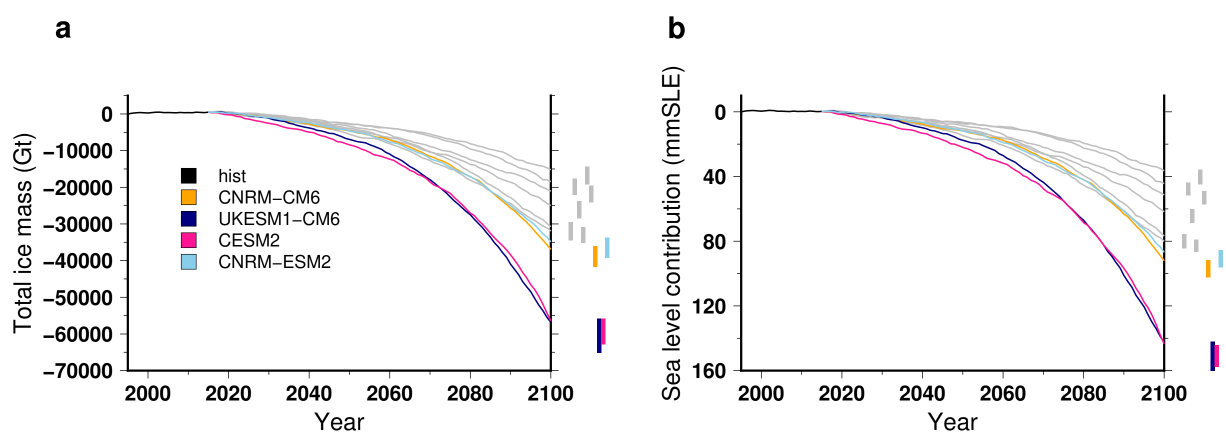

Figure 6. Simulated total ice ~~volume~~ mass change for the historical experiment *hist* (1995-2015) and the projections under the different CMIP6 forcings using the SSP585 scenario and the medium oceanic sensitivity: **(a)** total ~~ice volume~~ mass change and **(b)** ice volume contributing to sea level rise. For each projection experiment the right-hand side vertical bar shows the minimal and maximal ~~volume change~~ changes associated with the oceanic forcing uncertainty (*low* and *high* scenarios). The grey lines are the ~~volume change~~ changes under the CMIP5 forcings shown in Fig. 4.

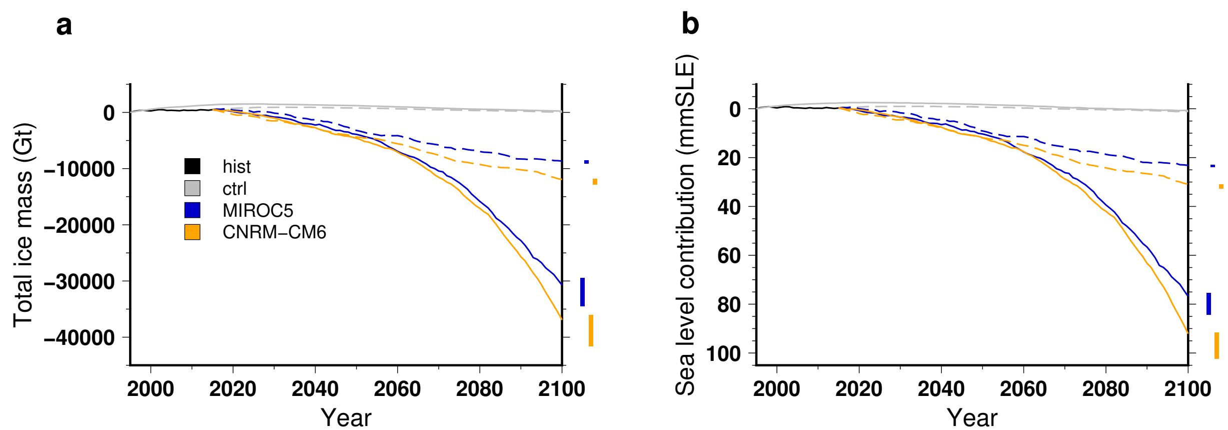

Figure 7. Simulated total ice ~~volume~~ mass change for the historical experiment *hist* (1995-2015), the control experiments *ctrl* (solid grey lines) and *ctrl_proj* (dashed grey lines) and for the projections using two climate models run under a ~~pessimistic greenhouse~~ high emission scenario for greenhouse gases (solid lines, RCP8.5 for MIROC5 and SSP585 for CNRM-CM6) and ~~an optimistic greenhouse~~ a low emission scenario (dashed lines, RCP2.6 for MIROC5 and SSP126 for CNRM-CM6) with a medium oceanic sensitivity, expressed as: **(a)** total ~~ice volume~~ mass change and **(b)** ice volume contributing to sea level rise. For each projection experiment, the right-hand side vertical bar shows the minimal and maximal ~~volume change~~ changes associated with the oceanic forcing uncertainty (*low* and *high* scenarios)

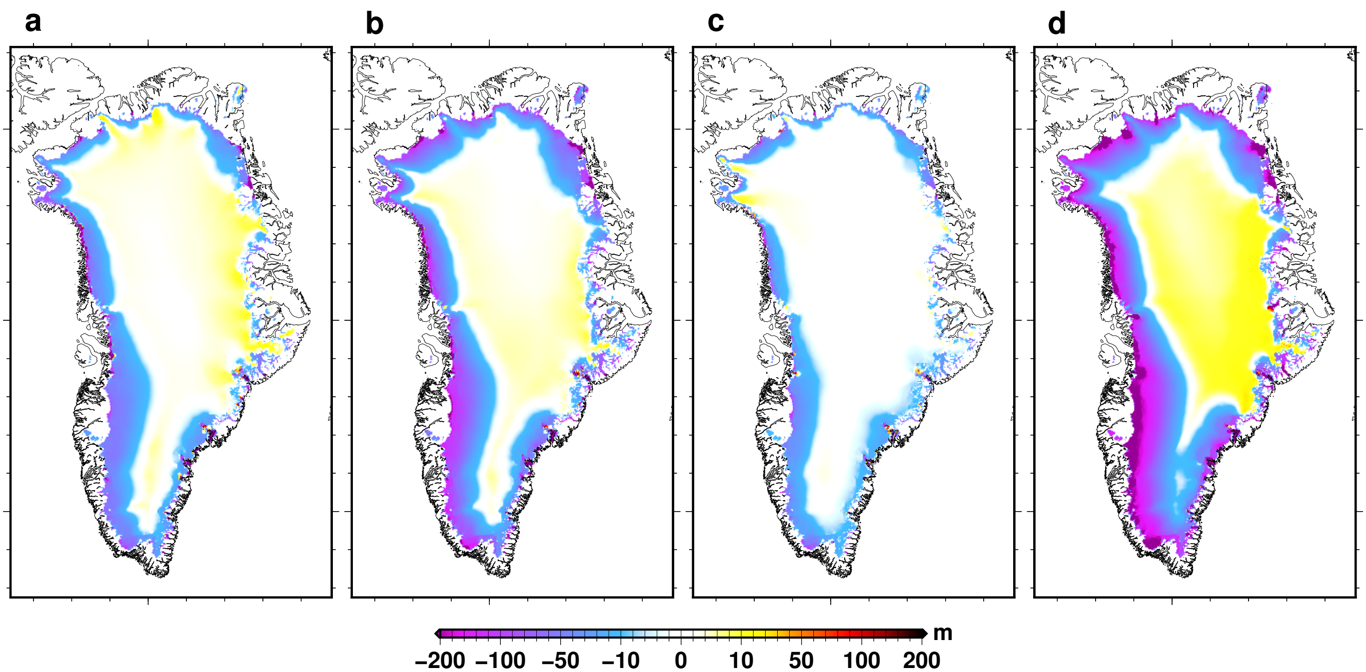

Figure 8. Simulated ice thickness change (2100 - 2015) for: **(a)** CSIRO-Mk3.6 (RCP8.5); **(b)** MIROC5 (RCP8.5); **(c)** MIROC5 (RCP2.6) and; **(d)** UKESM-CM6 (SSP585) climate forcing. The medium oceanic sensitivity has been used for this figure, except for UKESM-CM6 (d) for which we use the high oceanic sensitivity. The ice thickness change shown here is corrected for the ice thickness change (2100-2015) in the control experiment *ctrl_proj*.

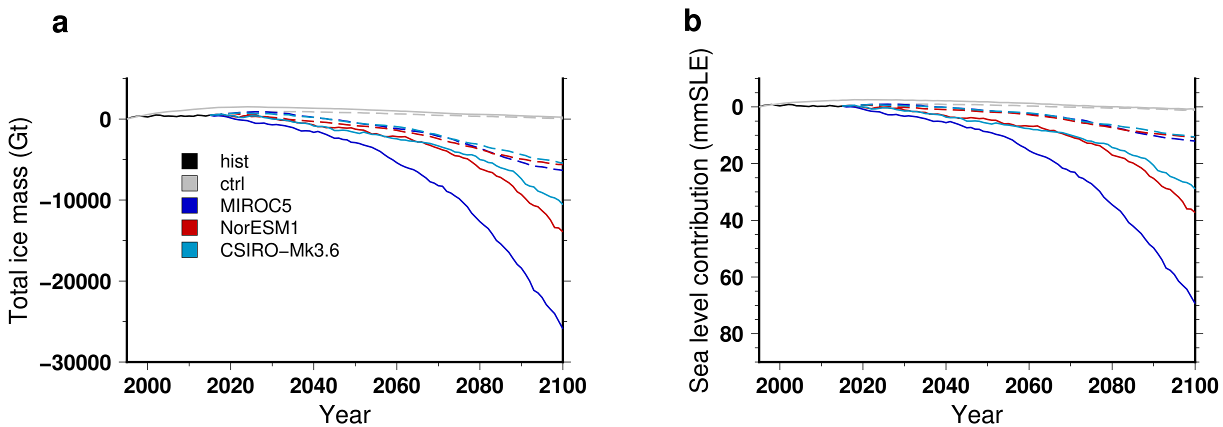

Figure 9. Simulated total ice ~~volume~~ mass change for the historical experiment *hist* (1995-2015), the control experiments *ctrl* (solid grey lines) and *ctrl_proj* (dashed grey lines) and the projections under different CMIP5 forcings using the RCP8.5 scenario. For the projections, the solid lines stand for experiments under atmospheric forcing change only (no imposed outlet glacier retreat, AO) while the dashed lines stand for experiments under oceanic forcing change only (no change in surface mass balance, OO). The ~~volume change is~~ changes are expressed as: **(a)** total ~~ice volume~~ mass change and **(b)** ice volume contributing to sea level rise. The medium oceanic sensitivity has been used for the oceanic only experiments (OO).

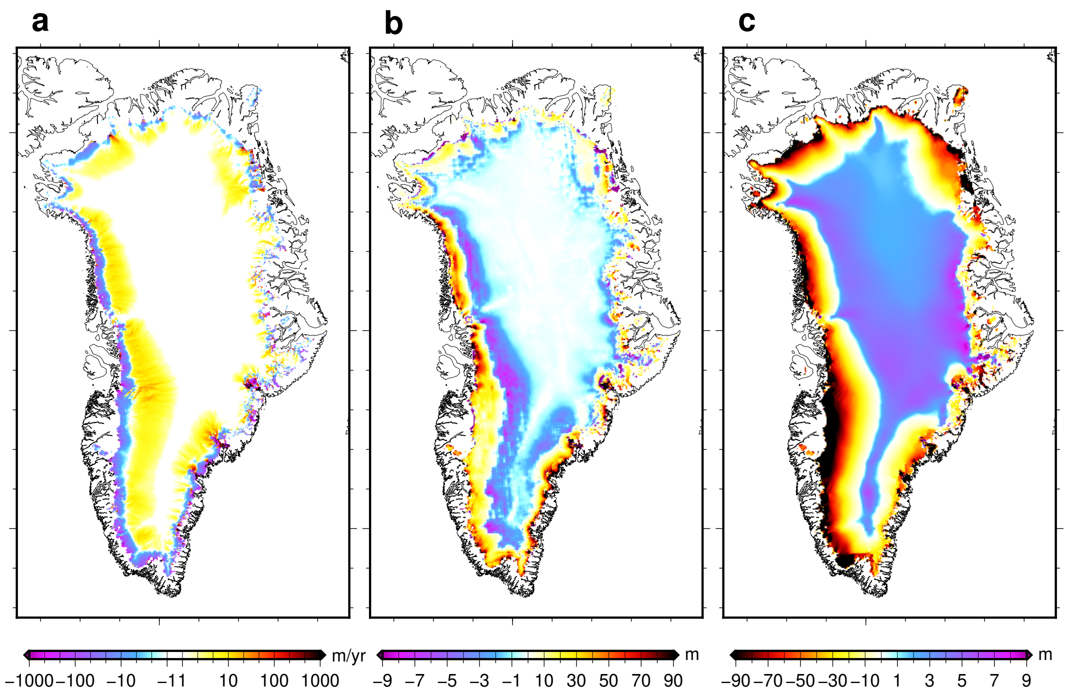

Figure 10. **(a)**: Simulated surface velocity change during the projection run (2096-2100 with respect to 2015-2019) using MIROC5 forcing under RCP8.5 with a medium oceanic sensitivity. **(b)**: change in the ~~dynamic~~ dynamical contribution to ice thickness change in 2100 (see text for definition) for this same experiment. **(c)**: simulated ice thickness change (2100-2015). For ~~both~~ all panels, we corrected the changes by the ones simulated in the control experiment *ctrl_proj* over the same period. The range -1 to 1 m yr^{-1} is set to white for velocity difference (**a**). The colour scale is not symmetrical for **(b)** and **(c)**.

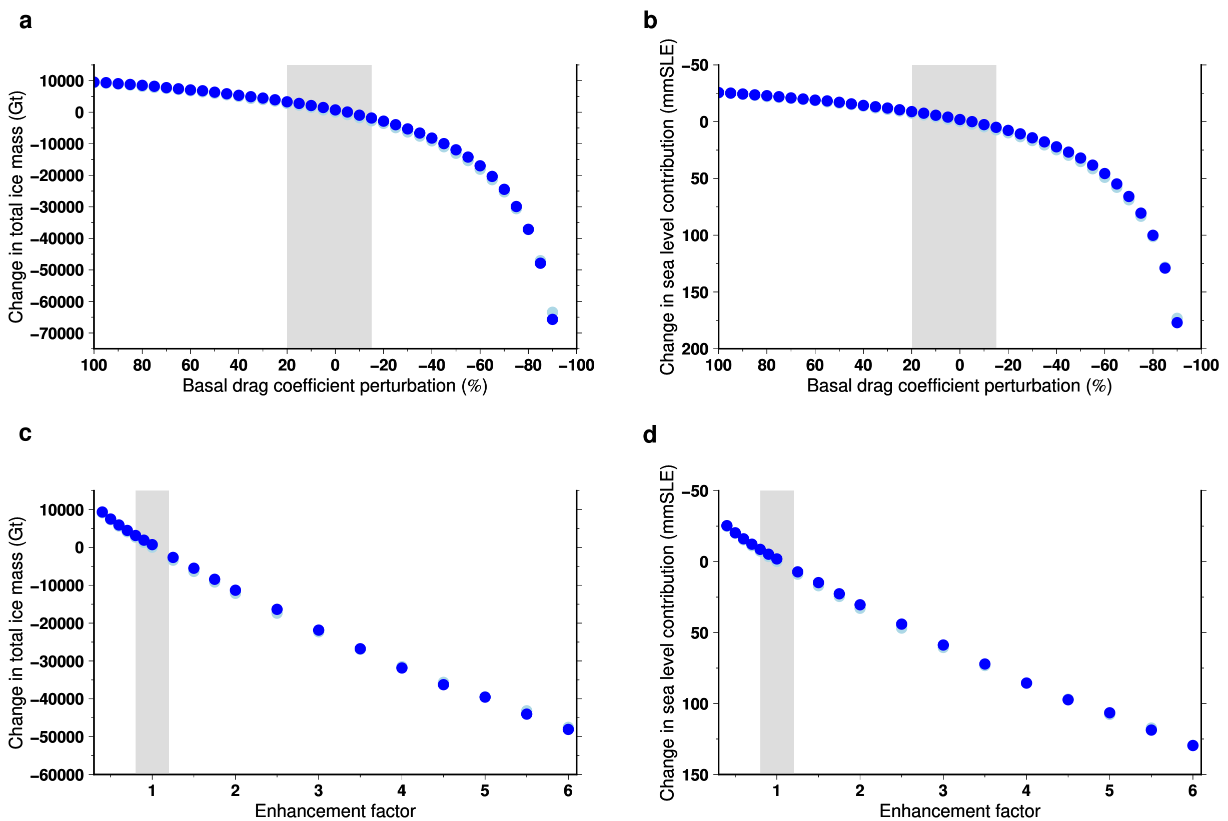

Figure 11. Change in ice volume for a modification of the basal drag coefficient (~~(a)~~ **a** and ~~(b)~~**b**) and different values of the enhancement factor (~~(c)~~ **c** and ~~(d)~~**d**). In this figure, each dot represents the ~~ice volume~~ difference in 2100 with respect to the standard ~~projection~~ experiment (no basal drag coefficient perturbation and enhancement factor at 1). The ~~climate forcing used for this figure is~~ dark blue dots are projection experiments that use MIROC5 under RCP8.5 with a medium oceanic sensitivity. The ~~perturbation is~~ light blue dots are control experiments *ctrl_proj*. Some control experiments can be hidden by the projection experiments if they imply a similar volume change. The perturbations are applied starting at year 2045. The vertical grey band stands for the range of perturbations that produce a 0.1% of total mass change in the perturbed control experiment with respect to the standard control experiment. The difference is expressed in total ice ~~volume~~ mass (~~(a)~~ **a** and ~~(c)~~**c**) and ice volume contributing to sea level rise (~~(b)~~ **b** and ~~(d)~~**d**).