Remarks to the Editor

Both reviews were quite extensive. Therefore, we considerably revised the manuscript.

Although not complained by the reviewers, we change the usage of the term "climate sensitivity", as we were not precise with its usage. We meant "strength of the SMB forcing over Greenland" by "climate sensitivity".

Our point-to-point response to the reviewers follow below.

Point to Point Response to all Reviewers

The comments by the reviewers are in indented blocks and italic fonts.

Response to Reviewer 1

Response to specific comments of reviewer 1

> *The manuscript is well organized and clearly written, but I find the discussion of the results and comparison with data or other studies quite qualitative and think it would improve the paper if the authors would make more quantitative analysis of their results, see numerous comments below.*

We quantified our analysis in all cases possibly. Details follow below.

> *The description of the forcing scheme is not clear (section 2.6) as it is not clear how the model applies the temperature defined in equation 8, as it seems that according to equation 10 the mass balance is only the difference between the modelled and measured elevation, divided by relaxation constant. Is this really the case, or is missing a description of a positive degree day method to compute the surface mass balance during the spin up period and current equation 10 would be one term of that forcing? 1*

The surface temperature defined in Eq. 8 applies directly to the surface of the ice sheet. Indeed, the first paragraph of this section is not fully clear.

We made considerable changes to this section. We changed the first paragraph of Section 2.6 (counting of Discussion paper, now Section 2.7), writing explicitly what are our boundary conditions for the palaeo run. To make it further clear we wrote which equations define our surface temperature and surface mass balance used for the palaeo simulations. Further, we now state explicitly that there is no additional forcing needed in Eq. (10), because this equation specifies an iteration for surface mass balance and surface elevation by running the ice sheet model in time.

> *The naming of the implied flux (equation 11) is confusing, suggest to call it something that indicates surface mass balance.*

We renamed "implied flux" into "implied SMB".

The introduction section is comprehensive and gives a good overview of the current state of development of large scale ice sheet models, and there is an impressive reference list for this study. I find, however, that the first part of the introduction should have more references for the general statements (page 1, lines 21 and 22, as well as page 2 lines 1 and 2) or at least indicate that these are not the only papers stating those broad things, with "e.g." before that one reference.

We added five new references to the paragraph. For the theme on accelerating mass loss, we included an "e.g."

Response to minor comments of reviewer 1

Page 1 line 19 suggest to replace "melting" with "retreat"

As we simulate submarine melting in our runs the term "melting" is correct here. No changes made to the manuscript.

Page 2 line 4-5, suggest to add something about the peripheral glaciers, that are not attached to the ice sheets, as those are actually contributing considerably to the sea level rise, see for example: Machguth, H., P. Rastner, T. Bolch, N. Mölg, L. Sandberg Sørensen, G. AÃˇralgeirsdottir, J. H. van Angelen, M. R. van den Broeke and X. Fettweis. 2013. The future sea-level rise contribution of Greenland's glaciers and ice caps. Environ. Res. Lett. 8, 025005 doi:10.1088/1748-9326/8/2/025005

The contribution of these glaciers sea level rise is small compared to the contribution of the ice sheet and their outlet glaciers. We added sentences on this topic to that paragraph and cite Forsberg et al. (2017) therein.

Page 2, line 6, suggest to add "a" before interplay

We added an "an".

Page 2, line 8 suggest to replace "thought" with "intended"

Done.

Page 2, line 10, suggest to replace "treating" with "treat"

We replaced "treating" with "to treat".

Page 2, line 11 suggest to replace "fast" with something like "computationally effiecent"

We rephrased to "computational efficiency".

Page 2 line 15, suggest to add references after "studies exist"

The reference to the model studies follow already after the introductory sentence of the paragraph. There is no change necessary here.

Page 2, line 19, suggest to replace "can serve" with "serves"

Done.

Page 2 line 20 suggest to delete "as" after "and"

The word "as" is fine there, as it introduces a subordinate clause. No changes made.

Page 2 line 22 suggest to replace "elevation" with "ice thickness"

Done.

Page 2 line 23 suggest to replace "elevation" with "ice thickness"

Done.

Page 2 line 24 suggest to replace "due" with "according"

Done.

Page 2 line 25 suggest to add "the" before "projected"

We think it correct, because we meant ice volume in general. No changes made.

Page 2 line 27, suggest to delete "in" after "yielding" and add "of" before "the present day"

We replaced "of yielding in" with "that it provides" and we added "of" before "the present day".

Page 2, line 28 suggest to replace "avoiding" with "avoids"

We replace "avoiding" with "prevents".

Page 2 line 32 and 33, suggest to add further references, suggest Rae, J. G. L. , G. AÃ˘ralgeirsdóttir, T. Edwards, X. Fettweis, J. Gregory, H. Hewitt, J. Lowe, P. Lucas-Picher, R. Mottram, A. Payne, J. Ridley, S. Shannon, W. J. Van de Berg, R. Van de Wal, M. Van den Broeke. 2012. Greenland ice sheet surface mass balance: evaluating simulations and making projections with regional climate models, The Cryosphere, 6, 1275-1294, doi:10.5194/tc-6-1275-2012, 2012 - for line 33, as it compares several RCMs

We added the reference.

Page 3, line 20 suggest to replace "their" with "the"

We rephrased the entire subordinate clause.

Page 3, line 23-24, this sentence is strange, perhaps replace "of being" with "to be"?

Right. We replaced "of being capable" with "to be suitable".

Page 5, line 8, add explanation of what H stands for

Done.

Page 6, line 27, is there missing "-1" on the unit for the decay parameter?

There was indeed a problem. We changed Eq. (7). The decay parameter is in the denominator now.

Page 6, line 28, suggest to add ",B," after Basal melt

Done.

Page 6, line 26, suggest to add ",W," after basal water layer

This makes no sense. "W" is already defined on page 6, line 22. No change made.

Page 7, line 6, "closest to the coordinates by Rignot and Mouginot (2012)" is no clear, suggest to edit to clarify what is meant here

Rignot and Mouginot (2012) give geographical positions of outlet glaciers. We clarified this in the revised manuscript.

Page 7, line 15 that statements "our method is already able to determine the subglacial discharge for each glacier" needs some quantification, how realistically do authors think the model results are?

We understand. That sentence has not too much meaning. We deleted the sentence. The section describes the coupling of SICOPLOIS-HYDRO with the plume model. At this place, there is nothing to quantify. Our results on this are already quantified in the results section of the original paper and are further quantified in the revised version of the paper. See below.

Page 8, line 16, see comment above, is there something missing in the equation that uses the T from equation 9, to compute the surface ablation?

Equation 10 is correct. No further input variables are necessary therein. We addressed the issue in our response to specific comments of reviewer 1 already. We improved Section 2.6 (Discussion paper counting, Section 2.7 revised manuscript) considerably.

Page 8, line 20 and 28, see comment above, suggest to rename "implied flux" to something that has surface mass balance in the name

Done. We use the term implied SMB now.

Page 9, line 10, suggest to rename "implied flux", see comment above

Done. We use the term implied SMB now.

Page 9, line 24, suggest to add "in the 21st century"

Done.

Page 9, line 28, Why do you need forcing until 2300?

In other papers, projections beyond 2100 (until the year 2200 or 2300) are often done in context of ice sheet modelling. This way we can compare with the other publications, what we do in the paper.

We rephrased the sentence and explained why we extended the forcing to year 2300.

Page 10, line 14 how is the ice load changing? If the mass balance follows equation 10 the thickness is always kept close to the observed one?

This is correct. The change in ice load is minor. We opted for free bedrock, because this is a standard setup for SICOPOLIS palaeo runs. We are aware that the change in bedrock is minor if the ice surface relaxes to observed. Nothing is wrong with such setup.

Now we write: "Using a standard setting of SICOPOLIS for palaeo runs, …"

Page 12, line 18, what does "almost no change" mean?

We deleted that sentence and quantified the drift.

Page 12 line 19, what does "a tiny volume change" mean? And "comparably small scale" ?

The point here is that the drift is small compared with the sea level change in the projections. We made this clearer now. Additionally, we improved the quantification of our model drift.

Page 12, line 32 strange beginning of sentence, suggest to edit (Certainly much stronger. ...)

We rephrased the sentence.

Page 13, line 6, what does "is minor" mean can it be quantified

Right. This sentence is about results we have not shown in the paper. Consequently, we erased the entire sentence now.

Page 13, line 14, what does "much higher" mean?

For mass of the ice sheet the effect amounts in at the end of 2100 AD 150 Gt/yr and 290 Gt/yr at the end of 2300 AD. We added a sentence for quantification.

Page 14, line 1, suggest to add "to" after "compared"

Right. Done.

Page 14, line 8, what does "does not show big impacts" mean, can you quantify?

At the end of 2100 AD, the effect increases the annual submarine melt rate of Helheim Glacier and Store Glacier by 76 m/yr and 81 m/yr, respectively. We added some sentences for detailed quantification.

Page 14, line 24-25, strange sentence, it is not clear to what is being referred to, needs more explanation or clarification

Right. We added more explanation and clarification to that paragraph.

Page 14, line 31-32, also here is strange sentence that needs clarification, deepening of basal topography by what?

We deleted that sentence.

Page 15, line 9, what does "an even smaller impact" mean?

Right. We quantified this by comparing our results with the results by Edwards et al. (2014).

Page 15, line 22, also here some quantification would be interesting "reasonably well" and "compared well" does not say enough about the success of the study - and line 33 "relatively large" does not give enough information

First point (Page 15, line 22): We moved the first sentence of that paragraph to the first paragraph of the conclusions Section. We deleted the remainder of that paragraph. Second point: We gave an RMS error for "very large" there and added a reference to Fig. 3a (counting of discussion paper, Fig 5a revised paper).

Page 16, line 5, what does "showed to be important" mean?

We rephrased the sentence.

Page 17, lines 3 and 8, it is not clear what "horizontal time slices" mean, is this spatially distributed?

Specified: Now we use the term MAR fields (the longitudinal-latitudinal distributed fields for annual mean of surface temperature, SMB and monthly surface runoff for a given year simulated by MAR).

Page 17, line 8-12, this whole section is not clear, what is "favorable sampling interval"? and "not enough time slices" is not clear and "interval too long" also needs clarification

We considerably rewrote the entire section. We added a new figure to that section for illustration (Fig A1).

Page 17, line 14, what does this mean? How much overestimation?

Same as before: we rewrote the entire section.

Page 17, lines 20-21 the sentence "Note that the totals… " needs clarification

Same as before: we rewrote the entire section.

Page 25, table 2, Rows and columns have been put in wrong lines (interchange)

The caption of Table 2 was wrong. Therein, we changed "columns" with "rows" in the figure caption.

Figure 3, it is no clear at what time this comparison is made, would be useful to add that information to the figure caption

It is for present day. We added that information to the figure caption (now Fig. 5).

Figure 5, is 4b the difference between 5a and 5b? it would be useful to add third column with the difference and suggest to add boxes on 5C to show the location of smaller figures in Figure 6

First point: Figure 4 (Discussion paper counting, now Fig. 6) is from SIA runs in 10 km resolution, while Fig. 5 (now Fig. 7) is from a hybrid run in 5 km resolution. We explicitly say this in the paper. Fig. 4 (now Fig. 6) was to illustrate the dependence of the model on the relaxation time. Fig. 5 (now Fig. 7) shows the resulting elevation with an optimal relaxation. Differences in simulated elevation to observed one are rather small. Moreover, as we already zoom in to velocity in Fig. 6 (now Fig. 8), we do not think that it makes sense to discus more details in a differences plot. Second point: We added rectangles to Fig. 5c (now Fig. 7c) to indicate the position of the enlarged areas in Fig. 6 (now Fig. 8). We changed the figure captions accordingly.

Figure 6, suggest to add reference for observations, same as in Figure 5?

We added the reference to the caption of Fig. 6 (now Fig. 8).

Figure 8, see comment above about renaming the term "implied flux"

Right. We change the inset of the figure: "IMPL" is changed to "iSMB". We changed the figure caption: "implied flux" is changed to "implied SMB (iSMB)". Now Fig. 10.

Figure 9, is the green line, the total basal melt constant? Then give the exact number value in caption

We intended to show that the total basal melt is small compared to the total surface runoff. Furthermore, the total basal melt is nearly constant and range between 10 and 12 Gt/yr. We found that our 15 Gt/yr were wrong after checking the model output. We added clarification to the main text and the figure caption for Fig. 9 (now Fig. 11).

Response to Reviewer 2

Response to general comments of reviewer 2

It is not clear, at least to me, how you deal with ocean-ice interaction. Does SICOPOLIS simulate any ice shelves in your experiment? If yes, what do you choose as subshelf melting rate? Please make this clearer.

In the model setup for this paper, SICOPOLIS does not simulate ice shelves. We included a new subsection (Section 2.1 “Overview of IGLOO”). This subsection clarifies the point with the ice shelves.

Whilst the different model sub-components are generally well described I had difficulties in understanding what is the 1-D glacier model presented in Figure 1. My guess is that it corresponds to the coupler between the plume model and SICOPOLIS? Similarly as for your other arrows, you could add in Fig. 1 what is exchanged between the 1-D glacier model and the other components. In particular I do not understand what the right-to-left arrows stand for since the models are not interactively coupled yet.

We added a new subsection (Section 2.1 “Overview of IGLOO”) to the beginning Section 2 to specify the used model components of IGLOO. Further, we improved Fig. 1 by adding the variables for the submerge part of the outlet glaciers and for their submarine melt. The exchange arrows between the ice sheet and the outlet glaciers in Fig. 1 are dashed now. Additionally, we refer to Beckmann et al. (2018), who explains more details on the glacier and plume models.

Response to specific comments of reviewer 2

P3L29-30 Please show on a map where these two glaciers are located.

We added a new figure (Fig. 2) with a map, which shows the location of the two glaciers.

P6L7 How is defined the “shape of the glacier front”?

We meant the submerged part of the outlet glacier and changed the term accordingly.

P7L8 "surface melt/runoff" ! It is not clear what this is. I assume it is the runoff provided by the MAR model (_rain minus retention due to refreezing?) and not surface ablation only?

Understood: melt is not synonym of runoff. It is the runoff, in our case from the MAR model. Due to the demands of reviewer 1, we erased the entire sentence. However, in the revised version of the paper, we no more use the term "melt/runoff".

P7L8-9 In your framework, you use the routing scheme of HYDRO (i.e. based on effective pressure) to route the water generated by surface melt. A fair amount of surface melt could be routed using surface gradient instead. Could you comment on how this can affect the pattern of discharge to the ocean?

We commented to this in our online response: We fully agree with the reviewer that the routing scheme in HYDRO can lead to different patterns of subglacial discharge reaching the ocean compared to a scheme, which uses surface gradients only. However, in case of complex bedrock with deeply incised structures, the routing scheme in HYDRO is more accurate. Apart from surface melt, there is also basal melt, which is important for winter and which our model accounts for.

We made no changes in the revised manuscript on that, because we have not made calculations with routing via surface gradient. Therefore, we cannot quantify this effect.

P7L29-30 The gradients are not well defined for the accumulation regime because of precipitation that has a much more complex spatial pattern than temperature (and by extension ablation). I would guess that the vertical gradients for runoff are not well defined neither since a large part of runoff is composed by liquid precipitation. Could you comment on that?

Response: We evaluated the data on surface mass from the MAR regional climate model and made the observation that the SMB gradients in the ablation area are much better defined compared to the gradients in the accumulation area. The issue can be seen in Fig. 2 by Helsen et al. (2012) too. The data in the accumulation area appears noisier compared to the data in the ablation area. Therefore, it makes sense to determine a regression line solely in the ablation area, while in the accumulation area one can just assume zero gradients.

Changes to the manuscript: We added some more explanation on this to that section.

P8L8-9 How is the surface temperature elevation gradient computed?

We compute the surface temperature gradient via representative local gradients inside a search radius, as Helsen et al (2012) did for surface mass balance. This is in further detail explained in Section 2.5 (Section 2.6 in the revised paper).

Changes made in the manuscript: We improved Section 2.5 (now Section 2.6) to make more visible that the surface temperature gradient is determined analogous to the SMB gradient. In Section 2.6 (now Section 2.7) and Section 2.7 (now Section 2.8) we refer to Section 2.5 (now Section 2.6) concerning the gradients now.

P8L25-27 A list of limitations of such an approach is welcome here, thanks. However I think you should expand more on the discussions of these. In particular, I think that taking into account the free-evolving topography during past cycles will have a large impact on simulated temperature profile as the ice thickness has considerably changed during the last termination (Vinther et al., 2009) and the stress regime will be largely different with an ice sheet extended towards the continental shelf in glacial conditions. An other limitation of the SMB anomaly method to drive the spin-up is that an artificial SMB term is used to compensate all the model deficiencies in terms of ice dynamics. A discussion on these points would be much appreciated.

Response: Such a discussion makes sense.

Changes made: We moved the paragraph describing the limitation of the relaxation approach to the discussion section (Section 6), where we extended considerably our discussion on the model limitations.

P9L27-29 Please rephrase.

We considerably improved that paragraph.

P10L9 Can't we expect a regional freshening due to the Greenland ice sheet melt? Why this could not be tested here as well with idealised scenarios (as for temperature)?

As IGLOO does not include interactive ocean, we are not able to test this hypothesis.

P10L22-24 The change in resolution certainly has an impact on the stress regime simulated by SICOPOLIS. Could you compare the state of your Greenland ice sheet (internal temperature) and your inferred surface mass balance (Mimpl) for present-day at 5 and 10 km resolution? Maybe you could add a few words on why doing the spin-up at a coarser resolution is not a problem in your case. It could also be interesting to have the future projections at 10 km resolution.

We commented to this in our online response and included a figure therein. The simulation with the resolution switch already shows a finer structure of the basal temperature field compared to that with lower resolution.

Changes to the manuscript: We added a new paragraph on the first switch to the section on initialization.

P10L25-26 Similarly to the change in resolution: have you tried to switch the hybrid mode before 500 years or after? How big is the impact on the simulated Greenland ice sheet? I assume that the thermal regime might be largely affected by the change in dynamics …

We commented to this in our online response. Mainly, we promised to investigate this in more detail, what did for the revised manuscript. We made a test calculation, which investigates the switch from shallow ice approximation to hybrid.

Changes in the manuscript: We added a new paragraph and a new figure on the second and third switch to Section 3 (Model initialization via palaeo runs).

> *P11L4-5 Would it make sense to discuss the RMSE in SMB instead of the total difference in SMB?*

Response: The choice of the measure is often arbitrarily. However, in our case, we used the total difference in SMB, because we can compare it with the total SMB from simulations with regional models of the Greenland ice sheet. Indeed, we use the difference in SMB for discussion in the discussion section.

No changes made to the manuscript.

> *P11L6-11 In the paper it could be worth discussing the spatial pattern of your difference in surface elevation and surface mass balance. From b-c we can really see that you have an important model drift at the margins: except for the NEGIS region your velocities seem too high (confirmed by Fig. 5) leading to negative surface elevation difference (compensated by artificial positive SMB anomaly). Related to this, why could you not find a Cb value minimising this model drift?*

Indeed this is missing, as we solely focussed the text on the dependence of the field patterns on the relaxation time.

Changes made: We discuss the elevation differences and how they relate to implied SMB and velocity pattern now. Concerning proposal by the referee to use Cb to minimize surface elevation difference, we commented to this in our online response.

> *P11L13-14 Do you need to spin-up HYDRO as well?*

HYDRO does not need any spin-up, as it is a diagnostic model. We added a clarifying sentence on this matter to the first paragraph of Section 2.2 (Section 2.3 revised manuscript).

> *P11L24-25 The simulated surface velocity does not seem smaller to me when you look at the western flank of the ice sheet.*

We wrote about the ridges and not about the flanks. We reformulated the sentence.

> *P11L26-27 Please reformulate.*

We fixed the sentence.

> *P11L29-30 Why?*

In this paper, we operate SICOPOLIS in a setting, which does not treat ice shelves. We clarified this.

P12L18 There is no change in ice volume visible but it does not mean that the ice thickness is not changing as you have compensating errors.

We commented to this in our online response: We are aware that in principle there can always be regional biases, which compensate each other in total. In our approach, the implied flux compensates particularly the regional errors of the ice sheet model. Indeed inspection of evolution of the ice thickness field could be used to refine our approach. We are grateful for this hint. In this paper, we use ice volume as indicator for the quality of our approach. As the change in ice volume in the run forced with implied flux only is very small, we regard ice volume as indicator as far sufficient.

No changes made to the manuscript.

P12L20-21 How this correction has been made? Do you use a point by point correction of ice thickness or do you simply use a correction of volume based on an averaged number? If the latter, how large would have been the difference when using the point by point correction?

We subtract the ice volume gained by the run forced with implied flux only from the ice volume resulting from the projection runs. This is done for every point in time. We clarified this in Section 5.1.

P12L30 The effect is stronger with RCP8.5 when looking at the absolute value but relative changes are in fact smaller. Which is in agreement with Vizcaino et al. (2015).

This is correct. We added some sentences about the relative impact of this feedback to Section 5.1.

P28 Fig. 3 Maybe you could add to these the evolution of total volume and RMSE of SMB for the different relaxation time.

As there is already a panel for the difference between the totals for simulated and observed SMB, this would not give to much new information. We would like to keep the figure as it is.

P29 Fig. 4 Could you add more levels to your colour scale?

We added more levels to both colour bars (now Fig. 6).

P31 Fig. 6 Do you have floating points? If yes, you should highlight them on this plot.

In our model setting, the ice is restricted to land and cannot move into the ocean. Therefore, we do not have floating points. This means that there is not need for changes in the manuscript.

P32 Fig. 7 Could you comment on why you have sub-glacial lakes when the base is frozen?

We commented to this in our online response and added several sentences on this to the end of Section 4.

Response to technical corrections of reviewer 2

> *P3L31 Dot instead of semicolon*

Done.

> *P6L2 No capital S for "submarine"*

Done.

> *P35 Fig. 10 Typo towards the end of the second line.*

Indeed, there was a problem with the sentence. We simplified the sentence.

References

Beckmann, J., Perrette, M., Beyer, S., Calov, R., Willeit, M., and Ganopolski, A.: Modeling the response of Greenland outlet glaciers to global warming using a coupled flowline-plume model, The Cryosphere Discussions, pp. 1–32, doi:10.5194/tc-2018-89, in review, 2018.

Forsberg, R., Sørensen, L., and Simonsen, S.: Greenland and Antarctica ice sheet mass changes and effects on global sea level, Surveys in Geophysics, 38, 89–104, doi:10.1007/s10712-016-9398-7, 2017.

Simulation of the future sea level contribution of Greenland with a new glacial system model

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Abstract. We introduce the coupled model of the Greenland glacial system IGLOO 1.0, including the polythermal ice sheet model SICOPOLIS (version 3.3) with hybrid dynamics, the model of basal hydrology HYDRO and a parameterization of submarine melt for marine-terminated outlet glaciers. Aim of this glacial system model is to gain a better understanding of the processes important for the future contribution of the Greenland ice sheet to sea level rise under future climate change scenarios. The ice sheet is initialized via a relaxation towards observed surface elevation, imposing the palaeo-surface temperature over the last glacial cycle. As a present-day reference, we use the ~~1961-1990~~ 1961–1990 standard climatology derived from simulations of the regional atmosphere model MAR with ERA reanalysis boundary conditions. For the palaeo-part of the spin-up, we add the temperature anomaly derived from the GRIP ice core to the years ~~1961-1990~~ 1961–1990 average surface temperature field. For our projections, we apply surface temperature and surface mass balance anomalies derived from RCP 4.5 and RCP 8.5 scenarios created by MAR with boundary conditions from simulations with three CMIP5 models. The hybrid ice sheet model is fully coupled with the model of basal hydrology. With this model and the MAR scenarios, we perform simulations to estimate the contribution of the Greenland ice sheet to future sea level rise until the end of the 21st and 23rd centuries. Further on, the impact of elevation–surface mass balance feedback, introduced via the MAR data, on future sea level rise is inspected. In our projections, we found the Greenland ice sheet to contribute to global sea level rise between 1.9 and 13.0 cm until the year 2100 and between 3.5 and 76.4 cm until the year 2300, including our simulated additional sea level rise due to elevation–surface mass balance feedback. Translated into additional sea level rise, the strength of this feedback in the year 2100 varies from 0.4 to 1.7 cm, and in the year 2300 it ranges from 1.7 to 21.8 cm. Additionally, taking Helheim and Store Glaciers as examples, we investigate the role of ocean warming and surface runoff change for the melting of outlet glaciers. It shows that ocean temperature and subglacial discharge are about equally important for the melting of the examined outlet glaciers.

1 Introduction

Since the last decade of the 20th century, the Greenland ice sheet (GrIS) loses mass with accelerating speed ~~(Helm et al., 2014)~~ (e. g. Helm e shaping one of the most important contributors to sea level rise ~~(Rietbroek et al., 2016)~~ (Shepherd et al., 2012; Rietbroek et al., 2016; Forsb This mass loss is not only driven by decreasing surface mass balance (SMB), but also by increasing ice discharge via outlet glaciers~~(van den Broeke et al., 2016)~~. The partition between these two contributions to GrIS mass loss is about equal ~~(Enderlin et al., 2014)~~ (Rignot et al., 2011; Box and Colgan, 2013; Enderlin et al., 2014; van den Broeke et al., 2016) . Understanding the processes determining the GrIS ice loss is vital for estimates of its contribution to future sea level rise.

Nowadays, the scientific community recognizes the large Greenland island as a complex system mainly composed of the ice sheet and numerous outlet glaciers (Joughin et al., 2010; Rignot and Mouginot, 2012), in subtle interaction with the surrounding ocean via fjord circulation (Straneo et al., 2012; Murray et al., 2010), and uprising meltwater plumes in an interplay with the calving outlet glaciers (O'Leary and Christoffersen, 2013). In our paper, we introduce the model IGLOO (Ice sheet model for Greenland including Ocean and Outlet glaciers, Fig. 1) ~~thought~~ intended to represent the major processes important for the future mass changes of the GrIS on timescales of some centuries. The ~~idea of this model~~ contribution to future sea level rise of the several glaciers and ice caps detached from the GrIS is small compared to the ice sheet and its attached outlet glaciers (Forsberg et al., 2017) . Anyhow, these glacial bodies are treated by SICOPOLIS in our approach. The idea of IGLOO is to capture the complexity of the system by its involved model components and, at the same time, ~~treating~~ to treat the description of all single components as detailed as necessary (Claussen et al., 2002). We aim to have a tool ~~which is sufficiently fast~~ with sufficient computational efficiency to enable large ensemble simulations on timescales important for future climate change.

Knowledge of the present-day state of the GrIS has been improving considerably. Not only that there are reliable data from numerous observations (e. g. Velicogna and Wahr, 2005; Bales et al., 2009; Morlighem et al., 2014), but also several modelling studies exist. Present-day GrIS velocities are resolved by ice sheet models in horizontal resolutions as high as 600 m, including flow patterns of outlet glaciers (Aschwanden et al., 2016). Robinson et al. (2012) explicitly demonstrated the multistable-hysteresis behaviour of the GrIS with a threshold of 1.6 °C above present-day global temperature for the decay of the GrIS; although such a decay will last at least about 1000 years. The past climate is an important element for GrIS ice sheet modelling as well, as it ~~can serve as~~ serves as a constraint for parameters particularly capturing the present-day GrIS (Robinson et al., 2011; Stone et al., 2013), and as it provides the history of the temperature field inside the present-day GrIS (Goelzer et al., 2013), which is important for the initialization of the GrIS in future warming simulations. However, palaeo-simulations with free surface have the drawback that their resulting present-day ~~elevation~~ ice thickness can differ considerably from observations (e. g. Calov et al., 2015). Such a simulated ~~elevation~~ ice thickness is an unfavourable initial condition for projections because, in this case, the future simulation would start with ice which resides at the wrong locations or is absent at ~~position~~ positions where it should reside ~~due~~ according to observations. This leads to an erroneous drift in projected ice volume evolution. Therefore, we opt for a fixed domain approach (Calov and Hutter, 1996) in our palaeo-spin-ups or, more precisely, for a scheme that relaxes the simulated surface elevation towards the observed one (Aschwanden et al., 2013). This approach has the advantage ~~of yielding in~~ that it provides a good approximation of the present-day temperature-velocity field for initialization and at the same time ~~avoiding~~ prevents a spurious response in ~~ice~~ volume during future simulations of ~~some~~ several hundred years. Different initialization methods are discussed by Saito et al. (2016).

There are several approaches to project future ice mass change of the GrIS, often with a special focus on a certain component of the Greenland glacial system. Classical ~~surface mass balance~~ SMB approaches assume a passive ice sheet, but resolve the atmosphere with general circulation models of the atmosphere (e. g. Gregory and Huybrechts, 2006) or additionally with a regional model ~~(van Angelen et al., 2012; Fettweis et al., 2013)~~ (van Angelen et al., 2012; Rae et al., 2012; Fettweis et al., 2013) . Several pioneering studies used three-dimensional dynamic ice sheet models in the shallow ice approximation (SIA) for projections of GrIS sea level contribution (e. g. Huybrechts and de Wolde, 1999; Greve, 2000). Later, higher-order (Fürst et al., 2013) or even full-Stokes (Gillet-Chaulet et al., 2012; Seddik et al., 2012) ice dynamics was included for GrIS future projections. In a higher-order ice sheet model, Fürst et al. (2015) parameterizes ice sliding via ocean-temperature rise due to future climate change to investigate the impact of ocean warming on future projections of GrIS sea level contribution. Studies with an atmosphere-ocean general circulation model coupled to a SIA ice sheet model via surface-energy fluxes were undertaken by Vizcaino et al. (2015). Inspections of GrIS sea level contribution with a special focus on outlet glaciers were accomplished with a 3-D ice sheet model by Peano et al. (2017) or with a 1-D shallow shelf model (Nick et al., 2013).

Here, we opt for the new version of SICOPOLIS v3.3 (Bernales et al., 2017). This version includes hybrid dynamics, which incorporates via the shelfy stream approximation (SStA; MacAyeal, 1989) longitudinal and lateral stresses, which are important for nearer-margin fast flow areas, along with horizontal plane shear (Hindmarsh, 2004) via the shallow ice approximation (SIA), important for the slow-flow regions in the more central regions of the ice sheet. Hybrid models have been developed before by Pollard and DeConto (2007, 2012); Bueler and Brown (2009); Hubbard et al. (2009); Winkelmann et al. (2011); Fürst et al. (2013); Pattyn (2017). They are a compromise between the shallow ice approximation and the full-Stokes approach. Key of these hybrid models is that SIA and SStA operate on a common domain, although there are other approaches to treat longitudinal and lateral stresses (Ritz et al., 2001). Compared to the SIA, the hybrid dynamics is more promising in reproducing the velocity field of the GrIS in the catchment area of ice streams, where there is already fast flow (Rignot and Mouginot, 2012). We do not employ ice-shelf dynamics in SICOPOLIS because the dynamics of outlet glaciers, which can have a floating ice tongue, is part of the outlet glacier component of IGLOO. We investigate the response of GrIS outlet glaciers to global warming (including ocean warming) with IGLOO in a separate paper (Beckmann et al., 2018a) .

Models assuming a basal water layer for treatment of subglacial hydrology (Shreve, 1972) were often applied to the Antarctic ice sheet (Le Brocq et al., 2009; Kleiner and Humbert, 2014). Here, we apply such a model to the GrIS, because it captures the major pathways of basal water toward the outlet glaciers (Livingstone et al., 2013), i. e. the model resolves in a good approximation the partition of basal water for the main GrIS outlet glaciers. This is important for reproducing the subglacial discharge of outlet glaciers, which is ~~input for~~ fed into our model of ~~their~~ meltwater plumes. Further on, our model for basal hydrology simulates a thickening of the basal water layer toward the major GrIS outlet glaciers, regions over which the ice velocity becomes higher (Rignot and Mouginot, 2012). Therefore, we couple the ice velocities to the basal water layer, while the basal melt rate of the ice sheet model provides the input to the model of basal hydrology. We expect this approach ~~of being capable~~ to be suitable for large-scale ~~ice sheet modelling~~ modelling of ice sheets on decadal timescales.

Simulating submarine melt rates at tidewater glaciers has been accomplished with different models that all share the core of the buoyant-plume theory (Sciascia et al., 2013; Xu et al., 2013; Slater et al., 2015; Carroll et al., 2015; Cowton et al., 2015; Slater et al., 2017). Recent studies ~~(Jackson et al., 2017; ?)~~ (Jackson et al., 2017; Beckmann et al., 2018b) show that the line plume model by Jenkins (2011) is an adequate tool to determine submarine melt rates for tidewater glaciers. In our paper, we apply ~~a~~ the recently developed line plume model ~~(?) after the equations of Jenkins (2011)~~ by Beckmann et al. (2018b) , based on the equations by Jenkins (2011) , to two outlet glaciers, Store and Helheim Glaciers (Fig. 2), of the Greenland ice sheet. We have chosen Helheim and Store Glaciers for investigating the impact of future warming on glacier melt and for testing our methods because they are well examined glaciers~~:~~. Numerous studies on these glaciers and their connecting fjord systems to the open ocean exist (Straneo et al., 2011; Sutherland and Straneo, 2012; Rignot et al., 2015; Jackson et al., 2014; Chauché et al., 2014). Some provide data on temperature- and salinity profiles inside the fjord from conductivity-temperature-depth (CTD) measurements or moorings.

We start with a description of the elements of the glacial system model IGLOO 1.0, including the future and past forcings utilized in our paper (Section 2). In Section 3, we describe our initialization method, while Section 4 compares the simulated present-day surface elevation and velocity with observations. Further on, modelled basal properties are compared with findings of other works. In Section 5, we present projections of the GrIS sea level contribution, ~~of~~ the GrIS total basal and surface runoff and ~~of~~ the submarine melt rates for two GrIS outlet glaciers (Store and Helheim Glaciers). The paper closes with a discussion (Section 6) and the conclusions (Section 7).

2 Ice sheet model for Greenland including ocean and outlet glaciers (IGLOO), version 1.0

2.1 Overview of IGLOO

IGLOO is designed for better understanding the response of the Greenland glacial system to climate change on centennial timescales. It consists of sub-models for the Greenland ice sheet, the basal hydrology, the outlet glaciers and the turbulent meltwater plumes. For initialization via palaeo-runs, the model is forced by the temperature anomaly from GRIP ice core data, while for future projections we make use of data from climate models. The design of IGLOO is shown in Fig. 1.

While the ice sheet model SICOPOLIS is coupled bi-directionally with the model for basal hydrology HYDRO, the coupling between the turbulent meltwater plume and HYDRO is only implemented off-line yet. In the model set-up of this paper, SICOPOLIS does not simulate ice shelves. However, the impact of ocean temperature and subglacial ice discharge on submarine melt in future warming scenarios is investigated by the offline coupling with HYDRO for Store Glacier and Helheim Glacier (Fig. 2). The coupling between outlet glaciers and turbulent meltwater plumes, and future warming scenarios with this model configuration, are described in a accompanying paper by Beckmann et al. (2018a) . The coupling between SICOPOLIS and outlet glaciers is not implemented yet.

In the following subsections, we will present the parts of IGLOO relevant for this paper, including coupling and forcing of these model components.

2.2 Ice sheet model SICOPOLIS version 3.3

SICOPOLIS (SImulation COde for POLythermal Ice Sheets; www.sicopolis.net) is a dynamic/thermodynamic ice sheet model that was originally created by Greve (1995, 1997) in a version for the GrIS. Since then, SICOPOLIS has been developed continuously and applied to problems of past, present and future glaciation of Greenland (e.g., Robinson et al., 2011), Antarctica (e.g., Kusahara et al., 2015), the Eurasian ice sheet including subglacial water (Gudlaugsson et al., 2017), the entire Northern hemisphere (Ganopolski and Calov, 2011), the polar ice caps of the planet Mars and others (see www.sicopolis.net/publ for a comprehensive publication list). The description given here follows Greve et al. (2017) very closely.

The model simulates the large-scale dynamics and thermodynamics (ice extent, thickness, velocity, temperature, water content and age) of ice sheets three-dimensionally and as a function of time. It is based on the shallow ice approximation for grounded ice (Hutter, 1983; Morland, 1984) and the shallow shelf approximation for floating ice (Morland, 1987; MacAyeal, 1989). Recently, hybrid shallow-ice/shelfy-stream dynamics has been added as an option for ice streams (Bernales et al., 2017). The rheology is that of an incompressible, heat-conducting, power-law fluid (Glen's flow law; e.g., Greve and Blatter, 2009).

A particular feature of SICOPOLIS is its very detailed treatment of ice thermodynamics. A variety of different thermodynamics solvers are available, namely the polythermal two-layer method, two versions of the one-layer enthalpy method, the cold-ice method and the isothermal method (Greve and Blatter, 2016). The polythermal and enthalpy methods account in a physically adequate way for the possible co-existence of cold ice (with a temperature below the pressure-melting point) and temperate ice (with a temperature at the pressure-melting point) in the ice body, a condition that is referred to as "polythermal". It is hereby assumed that cold ice makes up the largest part of the ice volume, while temperate ice exists as thin layers overlying a temperate base. In the temperate ice layers, the water content is computed, and its reducing effect on the ice viscosity is taken into account (Lliboutry and Duval, 1985).

SICOPOLIS is coded in Fortran and uses finite difference discretization techniques on a staggered Arakawa C grid, the velocity components being taken between grid points (Arakawa and Lamb, 1977). For solving the thickness evolution equation, we added a further option to the SICOPOLIS code (Appendix A). The simulations of the GrIS discussed here are carried out in a stereographic plane (WGS 84 reference ellipsoid, standard parallel 71°N, central meridian 39°W), spanned by the Cartesian coordinates x and y. The coordinate z points upward.

2.3 Subglacial hydrology model HYDRO

~~Subglacial water flux and storage are governed by the hydraulic~~ HYDRO is a diagnostic model that determines the subglacial water fluxes instantaneously via the hydrological potential Φ, which depends on the elevation potential and the water pressure p_{w} (Shreve, 1972):

$$\Phi = \rho_{\mathrm{w}} g b + p_{\mathrm{w}}, \tag{1}$$

with the ice base b, acceleration due to gravity g and density of water $\rho_\mathrm{w} = 1000\,\mathrm{kg\,m^{-3}}$. The water pressure depends on the ice overburden pressure and the effective pressure N (normal stress at the bed minus water pressure):

$$p_\mathrm{w} = \rho_\mathrm{i} g H - N, \tag{2}$$

wherein $\rho_\mathrm{i} = 910\,\mathrm{kg\,m^{-3}}$ is the density of ice and H is the ice thickness.

Following previous authors such as Le Brocq et al. (2009) and Livingstone et al. (2013), we assume the water moving in a thin (a few mm) and distributed water film. Under this premise, the water pressure and the ice overburden pressure are in equilibrium, and therefore the effective pressure is zero. This enables us to reformulate Eq. (1) as

$$\Phi = \rho_\mathrm{w} g\, b + \rho_\mathrm{i} g H, \tag{3}$$

and then computing the water flux with a simple flux routing scheme as described by Le Brocq et al. (2006). This approach is only valid at large (km) scales and is not able to include local processes such as channels.

The flux routing method requires that every cell has a defined flow direction and that, by successively following these directions, the boundary of the study area is reached. Therefore, local sinks and flat areas must be removed prior to applying the routing scheme. We accomplish this by using a Priority-Flood algorithm as described in Barnes et al. (2014), which fills depressions in a single pass and then add a small gradient to the resulting flats. Adding a gradient towards the outlet of the depression ensures that the hydraulic potential is altered in the smallest possible way. This procedure is a very efficient way to guarantee that all water is drained into the ocean.

The hydraulic potential is computed following Eq. ~~1~~(1), and we use the basal melt rates from SICOPOLIS as the water input for the routing scheme (see Section 2.5.1). The timescales of the water flow are much smaller than for the ice flow, thus, the steady-state water flux ψ_w can be obtained by integrating the basal melt rate along the hydraulic potential.

From the resulting water flux ψ_w, we can compute the water layer thickness W (Weertman, 1972, 1966):

$$W = \left(\frac{12\mu_\mathrm{w}\psi_\mathrm{w}}{\mathrm{grad}\,\Phi}\right)^{1/3}. \tag{4}$$

At locations where sinks in the hydraulic potential have been filled, we set W to a very high value (10 m) to account for the presence of a subglacial lake.

2.4 Meltwater plume model

A further element of IGLOO is the line plume model by ~~? (after Jenkins (2011))~~Beckmann et al. (2018b) , based on Jenkins (2011) . It simulates the width-averaged submarine melt rate of a glacier and accounts for a uniformly distributed subglacial discharge along the grounding line. The plume model describes buoyancy-driven rise of subglacial meltwater until it reaches either neutral buoyancy or the water surface. Two counteracting processes control the maintenance or reduction of the plume buoyancy: ~~Submarine~~ submarine melting at the ice-ocean interface preserves the plume buoyancy, while simultaneously turbulent entrainment and mixing with the surrounding salty fjord water reduces it. The line plume equations are derived under the assumption that the plume is in equilibrium and are thus time-independent. The melt rate is determined by the plume velocity and temperature, which adapts to the boundary conditions along the glacier front or under the floating tongue. As input parameters, it requires the ~~grounding line depth Z ($Z < 0$), the shape~~ submerged part of the glacier front ~~,~~ d and the subglacial discharge Q that leaves the glaciers grounding line over the whole glacier width, and a temperature-salinity depth (TSD) profile close to the glacier. The determination of the input parameters of the plume model is described in section 2.5.2.

2.5 Coupling of model components

2.5.1 Coupling of SICOPOLIS with HYDRO

We use a slightly modified version of the Weertman-type sliding law proposed by Kleiner and Humbert (2014) to couple the basal hydrology model to the ice dynamics:

$$\boldsymbol{v}_{\mathrm{b}} = -f_{\mathrm{b}} C_{\mathrm{b}} |\tau_{\mathrm{b}}|^{p-1} \tau_{\mathrm{n}}^{-q} \tau_{\mathrm{b}}, \tag{5}$$

with the sliding velocity $\boldsymbol{v}_{\mathrm{b}}$, basal sliding parameter C_{b}, basal shear stress τ_{b}, basal normal pressure τ_{n} (assumed as the ice overburden pressure) and the stress and pressure exponents $p = 3$ and $q = 2$. We introduce the dimensionless factor

$$f_{\mathrm{b}} = f_T \left((1 - c_{\mathrm{w}}) + c_{\mathrm{w}} f_{\mathrm{w}}\right), \quad c_{\mathrm{w}} \in [0, 0.9], \tag{6}$$

with

$$f_T = \cancel{\exp\left(\nu(T - T_{\mathrm{pmp}})\right)} \exp\left(\frac{T - T_{\mathrm{pmp}}}{\nu}\right) \qquad \text{and} \qquad f_{\mathrm{w}} = \left(1 - \exp\left(-\frac{W}{W_0}\right)\right), \tag{7}$$

where f_{T} and f_{w} incorporate sub-melt sliding and basal hydrology respectively. Sub-melt sliding allows sliding below the pressure melting point T_{pmp} according to the decay parameter ν (Hindmarsh and Le Meur, 2001), whereas the basal hydrology term depends on the water layer thickness W divided by a typical scale of the layer thickness W_0.

The parameter c_{w} in Eq. ~~6~~ (6) is a weighting factor between "background sliding" – determined by C_{b} – and enhanced sliding due to the basal water layer. Using $c_{\mathrm{w}} = 0$ yields the standard model without any effect of basal hydrology, while $c_{\mathrm{w}} = 0.9$ leads to the same expression as Kleiner and Humbert (2014). In our simulation with basal hydrology, we apply their parameter value, i. e. we set $c_{\mathrm{w}} = 0.9$, while we specify the typical scale of the layer thickness by $W_0 = 0.005\ \mathrm{m}$. Further, our decay parameter is $\nu = 1^{\circ}\mathrm{C}$.

The coupling is bi-directional. Basal melt B (including the water drainage from the temperate basal layer of the ice sheet) computed by SICOPOLIS is used to calculate the thickness of the basal water layer in HYDRO, which in turn affects the basal sliding (Eq. 7). Components and data exchange of the complete coupled model IGLOO are illustrated in Fig. 1.

2.5.2 Off-line coupling of SICOPOLIS and HYDRO with the plume model

We establish a procedure of determining submarine melt rates with our line plume model (Section 2.4) for all Greenland outlet glaciers. This procedure applies only off-line yet, i. e., the input and output of the model components are exchanged manually via data files.

~~For the required subglacial discharge, we use HYDRO to route the basal melt of SICOPOLIS to the grounding line of the glaciers. The grounding line position of each glacier is determined by the ice-land mask of SICOPOLIS of the year 1900 closest to the coordinates by Rignot and Mouginot (2012) .~~

~~Furthermore, we assume that the surface melt/runoff penetrates directly down to the bedrock and route it from there as basal water to the grounding lines.~~ To clarify, as this coupling is off-line, the sliding of ice (Section 2.5.1) is affected solely by basal melt, while the surface melt and basal melt can impact the meltwater plume. ~~Thus, with a method based on distance criteria, the basal water is routed to the determined grounding line position of each glacier and enters the corresponding fjords as subglacial discharge~~

For the subglacial discharge required by the plume model, we use HYDRO to route both the basal melt of SICOPOLIS and the surface runoff by MAR as basal water to the grounding lines of the outlet glaciers. We route on a monthly timescale to resolve seasonality. For the surface runoff, we assume that it penetrates directly down to the bedrock. Among others, Rignot and Mouginot (2012) provide data of the geographical position of many outlet glaciers. We use these data to allocate the water leaving the ice sheet to the individual outlet glaciers.

Although we simulate future scenarios, the grounding line position is considered to be fixed for this procedure. Of course, for glaciers close to another that share the same catchment area, a moving grounding line position might have severe effects. We will account for these dynamic glacier processes in the next version of IGLOO. ~~Despite the assumption of a fixed grounding line, our method is already able to determine the subglacial discharge for each glacier.~~

2.6 Evaluating the data from the regional atmosphere model MAR

The ice sheet model needs the mean annual surface temperature and ~~surface mass balance (SMB)~~ SMB as climate forcings at the surface. In addition, the plume model requires monthly runoff. ~~We construct our future climate forcing from~~ Here, we explain how we derive these forcing fields and their gradients from data of simulations by the MAR regional climate model (Fettweis et al., 2013). These fields and their gradients serve to define our climate forcing of the GrIS for the past (Section 2.7) and for the future (Section 2.8).

Historical MAR simulations using different climate reanalysis products to define the boundary conditions for the regional simulations are available. The boundary conditions for MAR future projections up to 2100 are provided by the output of several CMIP5 general circulation models for different RCP scenarios. Since the MAR simulations are performed for fixed surface elevation of the GrIS, and we expect substantial changes in the ice elevation under future warming scenarios, we correct the regional model output for the change in surface elevation by applying the gradient method of Helsen et al. (2012). In their method, they derived a representative local elevation gradient of the SMB in each grid point from a regression of simulated SMB and surface elevation within a given radius. Helsen et al. (2012) did this separately for accumulation and ablation regimes. Here, we extend their method by applying it also to surface temperature and runoff. ~~The~~

In our scheme, the search radius is set to 100 km, but is extended until it includes at least 100 grid points ~~,~~ if necessary. ~~For the surface mass balance, we apply the gradient method only to the ablation regime, because the~~ Our evaluation of the MAR data for the SMB revealed that the regression is in many cases not well defined for the accumulation regime~~(Helsen et al., 2012) .~~

. This issue can be also seen in Fig. 2 by Helsen et al. (2012) . Therefore, we apply the gradient method only to the ablation regime and set the SMB elevation gradient for the accumulation regime to zero.

2.7 Past climate forcing and implied ~~flux~~ SMB of the GrIS

~~The~~ Our past climate forcing ~~serves two purposes. First, it is used to~~ consists of the surface temperature and the SMB. By running the model over one glacial cycle, we determine an initial temperature-velocity field for our future warming scenarios~~, and second, to yield the implied flux~~. In particular, we determine the implied SMB for the present day, which is used in our future simulations as the climatological present-day ~~surface mass balance~~SMB.

The surface temperature for the past simulation is computed from the sum of the climatological field of the present-day surface temperature simulated by MAR, the temperature anomaly from the GRIP ice core and our temperature elevation correction obtained from the present-day MAR simulations:

$$T_s(x,y,t) = T_{s\,\mathrm{MAR(rean)}}^{\,\mathrm{Clim\ 1961-1990}}(x,y) + \Delta T_{\mathrm{GRIP}}(t) + \left(\frac{\partial T_s}{\partial z}\right)_{\mathrm{MAR(rean)}}^{\mathrm{Clim\ 1961-1990}}(x,y) \quad \Delta z(x,y,t). \tag{8}$$

The elevation correction in the last term of Eq. ~~8~~ (8) is the surface temperature elevation gradient (Section 2.6) from the MAR reanalysis data times a surface elevation difference, which reads

$$\Delta z(x,y,t) = z(x,y,t) - z_0(x,y), \tag{9}$$

with the surface elevation z, simulated with the ice sheet model SICOPOLIS, and the observed surface elevation z_0. For the observed surface elevation, we use the one by Bamber et al. (2013), which is the same as that utilized by Fettweis et al. (2013).

Here, the ~~surface mass balance~~ SMB M is defined by relaxing the ice sheet's surface elevation towards the observed surface elevation as

$$M(x,y,t) = \frac{z_0(x,y) - z(x,y,t)}{\tau_{\mathrm{relax}}}, \tag{10}$$

where τ_{relax} is a relaxation constant. With this relaxation method, we follow Aschwanden et al. (2013, 2016). Outside the ice sheet, we assign the high negative value of $M = -1000$ m ice/yr, which prevents the ice to flow outside its domain. ~~With these forcings, we~~ Running the ice sheet model in time by applying Eq. (10) specifies an iteration for the surface elevation and the SMB. Therefore, we do not need any further input here.

Applying the forcing fields for the surface temperature (Eq. 8) and the SMB (Eq. 10), we run the model over one glacial cycle. As we start the palaeo-simulation with the observed surface elevation, the simulated surface elevation relaxes soon towards the observed one. The relaxation constant τ_{relax} determines how close the simulated surface elevation is to the observed one. When the model reaches its present-day state ($t = 0$), we ~~yield the implied flux~~ obtain the present-day implied SMB M_{impl}, which is used in ~~future simulations~~ the future simulations, as

$$M_{\mathrm{impl}}(x,y) := M(x,y,0).$$

~~Through Eq. 10, the simulated surface elevation tends to approach the observed one, with a strength determined by $1/\tau_{\mathrm{relax}}$. If τ_{relax} equalled the scheme's time step for its ice sheet topography, the simulated surface elevation would fully match the observed one. This would correspond to a fixed domain, or more precisely, to a fixed surface simulation.~~

~~We made here the following simplifications: (1) We ignored changes in elevation andspatial extent of the GrIS during the past glacial cycle, (2) we assumed that the GRIP temperature record can be applied to the entire GrIS and (3) we assumed that the derived present-day elevation correction is valid for the entire glacial climate state~~

$$M_{\mathrm{impl}}(x,y) := M(x,y,0). \tag{11}$$

This SMB field corresponds approximately to the observed SMB, but compensates for errors of the ice sheet model. We will discuss further the relaxation approach and its limitations in Sections 3 and 6.

Outputs of this procedure are the present-day implied ~~flux and a full~~ SMB and a nearly present-day topography set (surface and bedrock elevation) belonging to this implied ~~flux~~SMB. Later on, the present-day implied ~~flux field~~ SMB field (Eq. 11) is added to the anomaly forcing of future climate simulations (see Eq. ~~(13)), as this implied flux field is assumed to be the present-day SMB including the errors of the model~~13).

2.8 Future climate forcing of the GrIS

The surface temperature forcing is computed from the climatological temperature of MAR simulations for 1961–1990 forced by the ERA reanalysis boundary conditions, the anomalies from MAR simulations forced by CMIP5 model output and a temperature elevation correction as:

$$\begin{aligned} T_s(x,y,t) = T_{s\,\mathrm{MAR(rean)}}^{\mathrm{Clim}\ 1961-1990}(x,y) \quad &+ \quad (T_{s\,\mathrm{MAR(CMIP5)}}(x,y,t) - T_{s\,\mathrm{MAR(CMIP5)}}^{\mathrm{Clim}\ 1961-1990}(x,y)) \\ &+ \quad \left(\frac{\partial T_s}{\partial z}\right)_{\mathrm{MAR(CMIP5)}}(x,y,t) \quad \Delta z(x,y,t). \end{aligned} \tag{12}$$

Here, the temperature elevation correction is determined via the product of the surface temperature elevation gradient (Section 2.6) of the MAR model with boundary condition from the CMIP5 models and the elevation anomalies simulated with the ice sheet model SICOPOLIS. As for the palaeoclimate, $\Delta z(x,y,t)$ are the simulated surface elevation anomalies (Eq. 9).

The SMB for future projections is computed as the sum of the implied ~~flux~~SMB, simulated SMB anomalies relative to the reference period 1961–1990 and an elevation SMB correction as follows:

$$\begin{aligned} M(x,y,t) = M_{\mathrm{impl}}(x,y) \quad &+ \quad (M_{\mathrm{MAR(CMIP5)}}(x,y,t) - M_{\mathrm{MAR(CMIP5)}}^{\mathrm{Clim}\ 1961-1990}(x,y)) \\ &+ \quad \left(\frac{\partial M}{\partial z}\right)_{\mathrm{MAR(CMIP5)}}(x,y,t) \quad \Delta z(x,y,t). \end{aligned} \tag{13}$$

Similar to temperature, the elevation SMB correction is calculated from the SMB elevation gradient (Section 2.6) of the MAR model with boundary condition from the CMIP5 models, multiplied by the simulated surface elevation anomalies.

Surface runoff is computed for each month from the climatological runoff of MAR simulations for 1961–1990 forced by the ERA reanalysis boundary conditions, the anomalies from MAR simulations forced by CMIP5 models output and a runoff elevation correction (Section 2.6), which again is computed similarly to the temperature elevation correction:

$$R(x,y,t) = R^{\text{Clim 1961-1990}}_{\text{MAR(rean)}}(x,y) \quad + \quad (R_{\text{MAR(CMIP5)}}(x,y,t) - R^{\text{Clim 1961-1990}}_{\text{MAR(CMIP5)}}(x,y)) \\ + \quad \left(\frac{\partial R}{\partial z}\right)_{\text{MAR(CMIP5)}}(x,y,t) \quad \Delta z(x,y,t). \tag{14}$$

Negative runoff values, which can result from this approach, are set to zero.

Figure 3 shows time series derived from the MAR data. During the 20th century, all curves show rather minor changes in average, besides a visible climate variability. This is in line with general knowledge (e .g. Box et al., 2009; Box and Colgan, 2013). ~~The climate sensitivity~~ In the 21st century, the anomaly of the SMB over Greenland is strongest for CanESM2, weakest for NorESM1, and MIROC5 lies in between. Of course, these ~~21st century~~ 21st-century warming trends correspond to IPCC AR5 (Collins et al., 2013) because the MAR forcing is from the CMIP5 models. The annual average temperature change over Greenland is stronger than the global one.

~~Since for NorESM1 and CanESM2, MAR data are missing for 1900–1949, and we do not have access to any MAR data beyond~~ Over the years 1900–1949, MAR provides data only for MIROC5, and after the year 2100~~(even though we need forcings until 2300), we filled the gaps by an extrapolation method that is explained in more~~ , MAR does not provide data for any of the GCMs used. To obtain complete forcings for the years 1900–2300, we closed the data gaps with an extrapolation procedure (described in detail in Appendix B~~.~~). We extended our forcings to the year 2300 for the sake of comparability with other studies involving ice sheet models, in which also long projections were made (e. g. Edwards et al., 2014) .

2.9 Future climate forcing of the plume model

As future forcing of the plume model, we employ the subglacial discharge from HYDRO and SICOPOLIS (Section 2.5.2) under the RCP 8.5 scenario (Section 2.8) from MAR with MIROC5 only. Additionally, a scenario of the temperature and salinity profiles is needed to project future submarine melting. Even for the present day, measurements inside fjords are rare and do not cover all of Greenland's fjords. We use CTD profiles close to the glaciers obtained for the year 2016 for Store Glacier (data from NASA's OMG mission(https://omg.jpl.nasa.gov/portal/)) and the year 2012 for Helheim Glacier (Carroll et al., 2016). For the ocean warming scenario, we assume a linear temperature trend of 0.03 °C per year over the years 2000–2100 for the entire profiles.

The 3 °C ocean warming in 100 years lies in the upper range found by Yin et al. (2011) for SE and W Greenland. The determined temperature and salinity profiles, in combination with the HYDRO output, serve as the input parameters for the line plume model to determine present and future submarine melting for the Greenland outlet glaciers.

3 Model initialization via palaeo-runs

For the initialization of the ice sheet model, we use the forcings for the surface temperature and the ~~surface mass balance~~ SMB as described in Section 2.7. ~~Here~~Using a standard setting of SICOPOLIS for palaeo runs, isostatic depression and rebound of the lithosphere due to changing ice load is modelled assuming a local lithosphere with relaxing asthenosphere with an isostatic time lag (LLRA approach, Le Meur and Huybrechts, 1996). For the geothermal heat, we use the spatial dependent data by Purucker (2012). In order to cover one full glacial cycle, we run the model over 135 kyrs. Initial conditions of these runs are the present-day ice thickness and elevation by Bamber et al. (2013). The original data with 1 km horizontal resolution are downsampled to 5 km and 10 km grid spacings. To perform a simulation in 5 km horizontal resolution over the entire glacial cycle with the hybrid model is illusive, as it takes 1 day for 100 model years on one HLRS2015 Lenovo NeXtScale nx360M5 processor. Therefore, we opt to perform the first 130 kyrs of the glacial cycle in 10 km horizontal resolution with the classical shallow ice approximation (SIA) employing the diffusivity method with an over-implicit ice-thickness solver. The last 5 kyr of the palaeo-run are performed in 5 km horizontal resolution. As we use different model hierarchies and settings, we devote some more explanation to these last 5 kyr.

During the last 5 kyrs of the run, we have three switches: one for refining the horizontal resolution, one for switching from SIA mode to hybrid mode, and a further one for switching from relaxing ice surface to free ice surface. The first switch at 5 kyr BP refines the horizontal resolution of the model from 10 km to 5 km. The second switch at 500 years BP changes from SIA to hybrid mode and additionally applies the mass conservation scheme for the evolution equation of ice thickness (Eq. A1). The third switch, which releases the relaxing ice surface to free development, is imposed at 100 years BP (year 1900). We introduced ~~this switch~~ the three switches at different times because they represent different regime changes.

We chose the time of the resolution switch furthest back in time (5 kyr BP) to allow the ice temperature to adapt to the refined basal topography. Indeed, by comparing the simulated present-day basal temperature in 10 km resolution with that in 5 km resolution from our simulation with the switches, one can observe that the 5 km basal temperature shows a much finer structure of the temperate basal regions at locations of outlet glaciers and their catchment areas; see also Fig. S1c and d in Calov et al. (2018) . As these temperate regions determine areas with basal sliding, enabling fast flow important for ice dynamics, we regard a time of 5 kyr as sufficient for resolving the major characteristics of the ice temperature field in the 5 km resolution.

Another important aspect for the choice of an adequate timing of the switches is the rate of sea level change, which quantifies the model drift and shows directly how the model recovers from the transition shock. We demonstrate this by inspecting different switching times and comparing our favourite initialization run with a test simulation (Fig. 4). While our initialization run has the switch from SIA to hybrid at the year 1500 and the switch from relaxing surface to free surface at the year 1900, the test simulation has both of these switches at the year 1900. Both curves of the rate of sea level change first rise fast to a maximum and then drop slowly towards smaller values. For the test simulation, the maximum is larger (0.06 cm/yr) compared to that of our initialization run (0.05 cm/yr). More importantly, the test simulation has not enough time to recover. While the rate of sea level change of our initialization run at the year 2000 amounts only to -0.0007 cm/yr, it is much larger (0.03 cm/yr) for the test simulation. In future projections, a drift of 0.03 cm/yr would yield a non-negligible contribution to modelled sea level rise of 3 cm in 100 ~~years earlier than the start of our future sea level scenarios (Section 5.1)in order to avoid spurious trends in ice volume change in our scenarios, which can happen when the ice sheet is released to free surface evolution suddenly~~yrs. Therefore, our initialization with separate switches for regime changes from SIA to hybrid and from relaxed to free surface is much more favourable and thus used for our projections (Section 5.1).

The choice of the relaxation constant rests on numerous simulations in 10 km horizontal resolution in SIA mode, running the model over one glacial cycle until the present day. Figure 5 shows the root mean square error (RMSE) in surface elevation and the total difference in SMB (the total implied ~~flux~~ SMB over the GrIS minus the total SMB simulated by MAR). With increasing relaxation constant, the RMSE in surface elevation increases moderately, while the total difference in SMB decreases strongly, i. e., there is a tradeoff between the RMSE in elevation and the total difference in SMB.

Figure 6 shows the spatial differences between the observed and modelled surface elevation and SMB for different relaxation constants. Again, the tradeoff for representing both surface elevation and SMB is visible. While the simulated elevation is very close to the observation for small relaxation constants, the SMB deviation is very high, even in the interior of the ice sheet, where the deviations reach the amount of magnitude of the accumulation rate. Therefore, too small relaxation constants should be excluded. For larger relaxation constants, both difference fields become smoother, but rather high deviations in surface elevation appear over vast areas of the GrIS. Therefore, too high relaxation constants should be excluded too. The spatial differences in Fig. 6b, e are from the medium relaxation constant of $\tau_{\mathrm{relax}} = 100$ years. One can see that the simulated surface elevation is too low over the north-western and south-eastern marginal regions (Fig. 6b). Over these regions, the simulated velocities are too high, which is compensated by an increased implied SMB (Fig. 6e) compared to the SMB from MAR. For most regions of the Northeast Greenland Ice Stream (NEGIS), the opposite situation occurs. The simulated surface elevation is too high, while the implied SMB is mostly negative over that region, which compensates the too low simulated velocities, see also Fig. 7. Again, the choice of the relaxation constant τ_{relax} is a tradeoff between the errors in the surface elevation and the errors in the implied SMB.

4 Present-day Greenland ice sheet

Here, we present our optimal simulation of the GrIS using the SICOPOLIS model version 3.3 with hybrid dynamics and the model for basal hydrology (HYDRO). Both models are fully coupled (see Section 2.5.1), and the horizontal resolution is always 5 km from now on. In the hybrid mode, a threshold of $r_{\mathrm{thr}} = 0$ applies to the slip ratio (Eq. 8 in Bernales et al. (2017)), i. e., the SStA equations are solved over the entire ice sheet, and the ice velocities are the weighted sum from the SIA and SStA velocities with the slip ratio as weight. The boundary conditions and initialization method to yield the present-day GrIS are described in Sections 2.7 and 3, respectively. As relaxation constant for the surface elevation we use $\tau_{\mathrm{relax}} = 100$ years. Optimal values for the sliding parameters are found by minimizing the error of simulated horizontal surface velocities for values $> 50\ \mathrm{m/yr}$, using observations by Rignot and Mouginot (2012). For such velocities, we expect basal sliding and hybrid ice dynamics to be relevant. We found $C_{\mathrm{b}} = 25\ \mathrm{m/(Pa\ yr)}$ to be optimal for the hybrid model with basal hydrology.

By design of the initialization, the simulated surface elevation compares overall well with the observed one, see Fig. 7a,b. However, as our surface relaxation method leaves the ice sheet's surface a certain degree of freedom (see also Fig. 6), the simulated ice surface over Summit and South Dome as well as on the ridge in between them is slightly lower. The simulated surface ~~velocity along the ridges is~~ velocities ($< 2\ \mathrm{m/yr}$) over the slow flow regions in the vicinity of the ridges are somewhat smaller compared to the observed ~~one~~surface velocities. Such (small) mismatches also appear with other higher-order models, even in higher resolution (Aschwanden et al., 2016). ~~Recall that~~ As we adjusted the sliding parameter C_b to match ~~higher~~ velocities higher than 50 m/yr with observations~~. The~~ , it is obvious that the model cannot resolve every detail over the slow flow regions. Still, the model resolves the major flow patterns of the GrIS, including the flow over the catchment area of the outlet glaciers and the fast flow of the major outlet glaciers and ice streams. Only the smaller-scale outlet glaciers, e. g. in north-west Greenland, are not fully resolved. ~~Further~~Since we have excluded ice shelves in SICOPOLIS, we cannot ~~model~~ reproduce outlet glaciers with floating tongues, such as Petermann, Nioghalvfjerdsbræ and Zachariæ Isstrøm. The ~~Northeast Greenland Ice Stream (NEGIS)~~ NEGIS is the only larger scale feature which we cannot reproduce properly. This feature cannot be simulated without additional assumptions (see the Discussion section).

Figure 8 zooms in Jakobshavn Isbræ and the two major outlet glaciers Helheim and Kangerdlugssuaq. Here, the ability of the model to resolve the catchment areas of these outlet glaciers in a 50 to 500 m/yrs range can be seen in more detail. However, the high-velocity patterns near the glacier termini do not fully match the simulations. In particular, the tributaries of Helheim and Kangerdlugssuaq glaciers and the tip of Jakobshavn Isbræ appear rather smooth compared to the observation.

Fast flow mainly appears over regions with a temperate ice bed. The simulated basal temperature in Fig. 9a shows a pattern which agrees basically with the reconstruction by MacGregor et al. (2016). Regions where there is basal melt, mainly caused by basal friction, exhibit a 1 to 5 mm thick water layer (Fig. 9b). There is a pronounced thickening of the water layer with our Shreve-flow modelling toward major ice streams and outlet glaciers, which is most visible for NEGIS, Jakobshavn Isbræ and Helheim Glacier. Moreover, smaller outlet glaciers like Store Glacier and Daugaard-Jensen Glacier receive intensified basal water supply too. The red dots over central-east Greenland correspond to sinks in the hydrological potential. These sinks are interpreted as subglacial lakes, following Livingstone et al. (2013) who reported similar results. In our simulations, these sinks can appear over a frozen base too, because we operate HYDRO with an option that allows computation of the hydrological potential over the entire ice area. This has the advantage that all basal water can safely reach the ocean. Allowing a water layer over a frozen bed for such sinks in the hydrological potential affects ice dynamics only very little because fully developed sliding appears only over temperate basal areas, while sub-melt sliding decays rapidly with decreasing temperature (Section 2.5.1).

5 Greenland glacial system projections

5.1 Projections of the GrIS's sea level contribution

For our projections of the contribution of the GrIS to global sea level rise, the GrIS is forced by SMB anomalies and surface temperatures derived from the MAR regional climate model (Section 2.8), making use of the initial ice sheet configurations explained in Section 3. As for the last 500 years of initialization, the fully coupled hybrid model including basal hydrology is utilized. Outside of the present-day GrIS area, similarly to the initialization, the prohibiting negative SMB is applied. In Fig. 10, we show the GrIS sea level contribution referenced to the year 2000. The control simulation forced solely with the implied ~~flux~~ SMB illustrates the characteristics of our initialization method. ~~Indeed, with this forcing, there is almost no change in ice volume visible.Only after~~ The sea level change from the control run is always small compared to that of all projections, see Fig. 10. Only compared to the RCP 4.5 projection over 300 years, a ~~tiny ice volume change can be detected in~~ small change in ice volume is visible for the control run (Fig. 10b~~, due to the comparably small scale in the y-axis therein. This model drift amounts about 2 mm sea level contribution per 100 years . In spite of~~). At the year 2300, the control run shows a sea level rise of 3.0 mm. This corresponds to a model drift of 1.0 mm per hundred years between 2000 and 2300. Despite such a small change, we correct our simulated sea level contribution of the GrIS in the simulation with MAR forcing for the ~~implied-flux-only simulations.~~ implied-SMB-only simulations. This correction is based on the ice volumes from the future simulations forced with the MAR data and the ice volume from the run forced with implied SMB only. All simulated ice volumes – those from the runs with MAR and that from the run forced with implied SMB only – are referenced to their respective year 2000 volumes.

Our projections of the GrIS sea level contribution for the year 2100 are close to simulations with a fixed present-day GrIS applying the cumulative SMB method (Church et al., 2013). This is in line with simulations with an active ice sheet model by Goelzer et al. (2013), who found that SMB is the major factor determining the GrIS sea level contribution over the 21st century. Our simulated GrIS sea level contribution for 2100 ranges from 1.9 cm (RCP 4.5, NorESM1) to 13.0 cm (RCP 8.5, CanESM2), see Table 2. Still, the ice dynamics (deformation and sliding velocities) plays a role in our simulations, indirectly via the SMB change. This can be seen when comparing the simulations with and without elevation SMB correction $(\partial M/\partial z)\Delta z$, Eq. ~~13~~(13). Ignoring the elevation SMB correction diminishes simulated 21st-century GrIS sea level contribution between 0.4 and 1.7 cm. Of course, this effect is strongest for the extreme RCP 8.5 scenario together with CanESM2, which is the CMIP5 model ~~exhibiting the most climate sensitivity.~~ used here with the largest SMB anomaly. Interestingly, the relative effect of the elevation SMB feedback at the year 2100 is smaller for RCP 8.5 compared to RCP 4.5. This corresponds to the findings by Vizcaino et al. (2015) , who used the ECHAM5 AOGCM coupled to an ice sheet model for their projections.

~~Certainly much stronger than for the 21st~~ At the end of the 23rd century, the ~~sea level~~ contribution of the GrIS ~~for the year 2300~~ to sea level rise ranges from 3.5 cm to 76.4 cm. The importance of the elevation SMB feedback clearly increases with the elapsed time of the projections, as the respective curves with $\partial M/\partial z$ on/off diverge more and more from each other. For RCP 8.5 with CanESM2, the relative increase of additional loss in ice volume due to elevation SMB correction nearly triples from 2100 to 2300, from 15 % to 40 %. This increase of the relative effect of this feedback with projection time was also observed by Edwards et al. (2014) and Vizcaino et al. (2015) . Detailed numbers for the sea level contributions of the GrIS for the years 2100 and 2300 are listed in Table 2.

Overall, our simulations show a strong dependence of the GrIS sea level contribution both on the RCP scenarios and on the ~~model~~ models used to force MAR. ~~Besides, the impact of the description of ice dynamics on the GrIS sea level rise contribution – i. e., whether SIA or hybrid is used – is minor, although the velocities over the catchment areas of the ice streams are better represented in the hybrid model compared to the SIA model (not shown).~~

5.2 Projections of the GrIS's total basal and surface runoff

For these projections, we use the basal melt from the two simulations by SICOPOLIS (Section 5.1) forced by the MAR data for which MAR used the MIROC5 GCM under the RCP 8.5 scenario. Surface and basal melt are routed over the ice base and distributed to the GrIS outlet glaciers. The details are explained in Section 2.5.2. Figure 11 depicts the total subglacial discharge split into surface runoff and basal melt. ~~The~~ Over the entire simulation time, the total basal melt ~~amounts to about 15 Gtper year~~is small and ranges between 10 and 12 Gt/yr, while the total surface runoff increases up to a peak value of 1700 Gt~~per year~~/yr. Note that, after the year 2100, the ~~surface runoff is decreasing~~ total surface runoff decreases due to the shrinking ice sheet area. Simultaneously, the effect of the elevation SMB feedback becomes more important after the year 2100. At 2100, ~~leading to much higher surface runoff than without the feedback (Eq. 14).~~ the difference between the total surface runoffs with and without elevation SMB feedback amounts to only 150 Gt/yr, while the same difference for 2300 nearly doubles to 290 Gt/yr.

5.3 Projections of submarine melt rate for the GrIS outlet glaciers Helheim and Store

Here, we inspect the impact of global warming under the RCP 8.5 scenario for two outlet glaciers: Helheim Glacier and Store Glacier. In detail, we analyse the impact of both subglacial discharge and ocean warming – as single and combined effects – on the submarine melt rate of these outlet glaciers. While the subglacial discharge comes from simulations with SICOPOLIS and HYDRO under the RCP 8.5 scenario, the ocean warming originates from a scenario similar to RCP 8.5 (Section 2.9). For analysing the impact of the elevation SMB feedback on submarine melt, the plume model is forced by subglacial discharge computed with and without the surface elevation correction of surface runoff (Eq. 14). We calculate all submarine melt rates under the assumptions of both glaciers being tidewater glaciers (no floating tongues) and of their grounding-line depths and widths remaining constant in time. These depths and widths are acquired from present-day observations and amount to 500 m depth and 5 km width for Store Glacier (Chauché et al., 2014) and 650 m depth (Carroll et al., 2016) and 6 km width (Straneo et al., 2016) for Helheim Glacier. We chose the entrainment parameter to be $E_0 = 0.036$ as recommended by ~~?~~ Beckmann et al. (2018b) .

Figures 12 and 13 illustrate the monthly subglacial discharge and the temperature profiles for the years 2000 and 2100 and the resulting submarine melt rates for the RCP 8.5 scenario. For both glaciers, the increasing subglacial discharge and the increasing ocean temperature have an about equal effect on the rising submarine melt, with the ocean temperature becoming more important towards the end of the year 2100. However, the combined effect of increased subglacial discharge and temperature exceeds the single effects alone. As a result, submarine melt exhibits a 2.5-fold increase for Helheim Glacier and a 4-fold increase for Store Glacier in the year 2100 (Figs. 12c and 13c). Although for the year 2000 the amount of basal melt (38 m^3/s for Helheim, 5 m^3/s for Store) is small compared to summer subglacial discharge (818 m^3/s for Helheim, 439 m^3/s for Store), it has a significant effect on the annual submarine melt rate. Due to the basal melt in the winter months (including early spring and late autumn), the submarine melt rate enlarges in those months substantially as illustrated by Fig. 14 for Helheim Glacier. The slight increase in subglacial discharge for all months (Fig. 14a) shows clearly the biggest increase in submarine melt rate for the winter months (Fig. 14b) due to the cubic root dependence of submarine melt rate on subglacial discharge (Jenkins, 2011). On the annual average, this effect leads, for the year 2000, to an increase of submarine melt for Helheim Glacier by 40 % and for Store Glacier by 20 % compared to the case when basal melt was not accounted for (Figs. 12c and 13c). ~~The missing effect of surface elevation correction does not show big impacts on~~ At 2100, the elevation SMB feedback increases the submarine melt rate ~~when turned off~~ of Helheim Glacier and Store Glacier from 787 by 76 to 863 m/yr and from

804 by 81 to 723 m/yr, respectively (Figs. 12c and 13c). With about 10 % for both outlet glaciers, this effect is relatively small. However, as Fig. 11 suggests, ~~this~~ the effect will become more important for the submarine melt rate after the year 2100.

In these experiments, the future submarine melt rate was calculated assuming a constant glacier terminus position and geometry. These calculation have to be seen as a first approximation because we neglect several factors that may influence the submarine melt rate. For instance, if the glacier retreats, the resulting grounding line depth may change depending on the underlying bedrock. Another factor that might change the melt rate estimation considerably is the distribution of subglacial discharge within the year. Here, we assumed no time lag in between runoff and its emergence as subglacial discharge. Due to the cubic root dependence of submarine melting on subglacial discharge, we already see the possible strong effect of basal runoff from the ice sheet on the distribution of the submarine melt rate of an outlet glacier over the year (see Fig. 14). Thus, an inefficient drainage system that is delayed by, e. g., storage of water in subglacial lakes (Nienow et al., 2017) might affect the seasonal distribution of subglacial discharge and thus the annual submarine melt rate substantially.

6 Discussion

In Section 3, we investigated the role of the relaxation constant for initialization. For very small relaxation constants, i. e., an essentially fixed ice surface, the difference between implied and observed SMB at present day becomes very large (more than 2000 Gt/yr, compared to an insignificant amount for τ_{relax} = 100 years). Note that the present-day magnitude of observed total SMB is only about 500 Gt/yr (e .g. Ettema et al., 2009). This means that ~~computation~~ computations with fully fixed surface should be treated with care, as the ~~total artificial mass needed to keep the ice sheet close to observation~~ required total implied SMB is very high. A further ~~factor is a smooth surface elevation~~advantage of the relaxation of the ice surface is that this smoothes the ice surface because it has a certain degree of freedom due to the relaxation constant while solving the ice thickness equation. A similar smoothing effect when running an ice sheet model with free surface evolution over 100 yrs was already observed by Calov and Hutter (1996) . Furthermore, they demonstrated that a smooth ice surface avoids irregular variations in the vertical velocity field.

For our initialization method, we made a number of simplifications. We ignored changes in surface elevation and spatial extent of the GrIS during the past glacial cycle. Anyhow, the elevation changes during the last glacial cycle over central parts of Greenland like Summit (about 100 m, Raynaud et al., 1997) or Dye 3 (some 100 m, Vinther et al., 2009) were rather small compared to the ice thickness. These rather small differences in elevation limit the inaccuracies in the ice temperature field caused by our relaxation approach, also because the maxima of these changes appeared about 10,~~the importance of which was already observed earlier by Calov and Hutter (1996) .~~ 000 years ago during the Holocene. Furthermore, due to slowly flowing ice there, a slightly different temperature-velocity field will not play a too important role for the overall ice dynamics. Also, ignoring the larger areal extension (Funder and Hansen, 1996) at the Last Glacial Maximum (LGM) should not affect the ice temperature field too strongly because over the more peripheral regions of the ice sheet the velocities are rather high and the LGM is 21,000 years ago. Here, the ice temperature will adapt faster and has more time to adapt compared to the central regions of the ice sheet. The assumption that the derived present-day elevation correction is valid over the entire glacial cycle is reasonable. We have examined this by evaluating the temporal dependence of the vertical temperature gradient from the MAR data. Indeed, for the future, this temperature gradient does not change too strongly in time, which does not mean that the elevation correction itself can be neglected here. Anyhow, for the palaeo-run, our ice surface is rather constant in time due to our relaxation approach, which makes the elevation correction small in this case. As a further simplification, we applied the temperature anomaly from the GRIP record to the entire GrIS. This is a standard approach, but still, we are aware that this anomaly is spatially dependent in reality. However, possible errors in this anomaly will become less and less important the nearer the palaeo-simulation approaches the present day. That our present-day temperature-velocity field is reasonable is supported by the fact that it lies in the range found by MacGregor et al. (2016) . In this context, we would like to repeat that our initialization serves the purpose to have the simulated surface elevation at present-day as close as possible to the observed one to minimize the drift in the future projections.

In our simulations, we cannot reproduce the NEGIS ~~ice stream~~ correctly. Certainly, one reason is that we do not optimize the surface velocity by a spatially dependent basal sliding coefficient. With spatially dependent basal sliding coefficients, other studies such as Price et al. (2011) and more recently Peano et al. (2017) simulated the NEGIS in better agreement with observations. Nowadays, there are process-oriented approaches to capture effects important for basal sliding. For example, stronger basal melting at the onset of the NEGIS caused by increased geothermal heat due to a palaeo-hotspot (Rogozhina et al., 2016) could be one factor speeding up the simulated NEGIS velocity. ~~A further factor can be a deepening of the basal topography in this region (Vallelonga et al., 2014) .~~

For our 300-year sea level projections, which reach beyond the 21st century, we prolong the forcing data of the MAR model until the year 2300. Because we merely held the forcing constant between 2101 and 2300, the real RCP 8.5 forcing could be larger, i.e., we expect our simulations with the RCP 8.5 scenario to be a lower estimate of sea level contribution of the GrIS, i.e., the estimate is a rather conservative one. Most certainly, even all our projections including RCP 4.5 are a conservative estimate ~~,~~ because a full coupling with ice–ocean interactions is missing in our model yet, and Fürst et al. (2015) found that ocean warming caused additional mass loss of the GrIS in his projections applying a parameterization of ocean warming.

Our additional sea level rise for the year 2100 due to elevation SMB feedback is somewhat higher than that by Le clec'h et al. (2017), who used the regional model MAR actively coupled to an ice sheet model for their simulations. ~~In contrast, Edwards et al. (2014) found an even smaller impact of this feedback than Le clec'h et al. (2017) , possibly due to an underestimation of its spatial dependence in Edward's parameterization.~~ For the year 2100, Edwards et al. (2014) give an additional contribution to sea level rise of the GrIS due to this feedback ranging between 0.25 and 0.32 cm under the SRES A1B scenario; we excluded their outlier of 0.1 cm. Still, their estimates of this feedback are rather small compared to our ones lying between 0.6 and 1.7 cm, which were produced with the RCP 8.5 scenario though. The SRES A1B scenario is somewhat more moderate that the RCP 8.5 scenario. However, this cannot explain such low numbers for the elevation SMB feedback. As demonstrated to be important by Le clec'h et al. (2017) with fully interactive two-way coupling, this feedback deserves a detailed inspection in the future.

Our presented projections for the GrIS contribution to global sea level rise in the 21st century (1.9-13.0 cm) are consistent with previous publications. However, they do not account for the dynamic response of Greenland outlet glaciers to ocean warming and increase of subglacial discharge. This effect will be account for in a forthcoming paper. We also intend to couple the 3-D ice sheet model SICOPOLIS with the 1-D model for many outlet glaciers.

7 Conclusions

We introduced the coupled Greenland glacial system model IGLOO 1.0 designed to describe the most important parts of the Greenland glacial system: the ice sheet, the subglacial hydrological system, the outlet glaciers and the ice-ocean interaction in the Greenland fjords. The applicability of the hybrid mode of the ice sheet model SICOPOLIS 3.3 to the Greenland ice sheet was demonstrated. Full coupling between the ice sheet model and the model of subglacial water HYDRO has been accomplished, while the coupling between HYDRO and the meltwater plume works only off-line yet.

~~The applicability of the hybrid mode of the ice sheet model SICOPOLIS 3.3 to the Greenland ice sheet was demonstrated. It showed that the model performs reasonably well, as the simulated velocity field compared well with observations, including the two major outlet glaciers Helheim Glacier and Kangerdlugssuaq Glacier and the Jakobshavn Isbræice stream. Further, for simulating optimal velocities, it is reasonable that the sliding coefficient for the model in hybrid mode is larger than that for the SIA model, as lateral strain partly compensates the effect of basal drag.~~

As initialization, we used a relaxation method similar to Aschwanden et al. (2013), but with a somewhat higher relaxation constant of 100 years. For this choice of the relaxation constant, we varied it systematically and investigated the resulting model behaviour by inspecting the RMS error in surface elevation as well as the difference between total simulated SMB and total SMB from the MAR regional climate model. It showed that, for a relaxation constant of 100 years, the deviation of our simulated total SMB from the MAR SMB is about zero, while – at the same time – the RMS of the simulated error in surface elevation stays reasonably small. Additionally, we showed that medium-value relaxation times lead to smooth 2-D fields of the implied SMB, while for too small relaxation times the fields become rather noisy, and for too large relaxation times regional deviations of the simulated elevation from the observed one become relatively large ~~.~~ (RMS error of 95 m for τ_{relax} = 300 years, see Fig. 5a).

Furthermore, we performed projections of the contribution of the GrIS to sea level rise until the year 2300 with hybrid ice dynamics forced by SMB anomalies from the MAR regional model. For the RCP 4.5 and 8.5 scenarios generated by MAR, three CMIP5 GCMs with different climate sensitivity were applied. Altogether, our projected GrIS sea level contribution for the year 2100 obtained with elevation SMB feedback ranges from 1.9 to 13.0 cm, and for the year 2300 from 3.5 to 76.4 cm. The effect of elevation SMB feedback ~~showed to be important~~contributes clearly to our simulated contribution of the GrIS to sea level rise. Generally, its impact increases in the long run with decreasing surface elevation (see Table 2).

Moreover, we demonstrated the importance of the different factors determining the increase of the melt rate of Greenland outlet glaciers under the extreme RCP 8.5 scenario, using Store and Helheim Glaciers as examples. It showed that the knowledge of near-terminus temperature and subglacial discharge in the fjord are both about equally important to determine the future melt of these two outlet glaciers. This underlines the importance of our approach with the Greenland system model IGLOO 1.0.

Code and data availability. SICOPOLIS is available as free and open-source software at www.sicopolis.net. The HYDRO module is not included in the repository yet. The MAR data used as the basis for our forcing are available at ftp://ftp.climato.be/fettweis/MARv3.5/Greenland/.

Appendix A: ~~Mass conservating~~ Mass-conserving scheme for ice thickness evolution

We included a new numerical scheme into SICOPOLIS 3.3, which discretizes the advection term of the ice thickness equation by a strictly mass-conserving scheme in an upwind flux form:

$$A = \frac{(\bar{v}_x(i+1/2,j)H_x^+ - \bar{v}_x(i-1/2,j)H_x^-)\Delta y + (\bar{v}_y(i,j+1/2)H_y^+ - \bar{v}_y(i,j-1/2)H_y^-)\Delta x}{\Delta x \Delta y}, \tag{A1}$$

where A is the advection term and $\bar{v}_x$, $\bar{v}_y$ are the x- and y-components of the depth averaged velocity, respectively. Further, Δx and Δy are the horizontal spacings. The upwind coefficients read:

$$H_x^- = \begin{cases} H(i-1,j), & \bar{v}_x(i-1/2,j) \geq 0, \\ H(i,j), & \bar{v}_x(i-1/2,j) < 0, \end{cases} \qquad H_x^+ = \begin{cases} H(i,j), & \bar{v}_x(i+1/2,j) \geq 0, \\ H(i+1,j), & \bar{v}_x(i+1/2,j) < 0, \end{cases}$$

$$H_y^- = \begin{cases} H(i,j-1), & \bar{v}_y(i,j-1/2) \geq 0, \\ H(i,j), & \bar{v}_y(i,j-1/2) < 0, \end{cases} \qquad H_y^+ = \begin{cases} H(i,j), & \bar{v}_y(i,j+1/2) \geq 0, \\ H(i,j+1), & \bar{v}_y(i,j+1/2) < 0, \end{cases}$$

with the ice thickness H. The pairs (i,j), $(i+1/2,j)$ etc. indicate the indices of the staggered Arakawa C grid.

Appendix B: ~~Adapting MAR~~ Filling the data ~~for~~ gaps of the MAR forcing for initialization and future simulations

~~One element of our initialization method (see Section 3) is the prevention of a model shock (Aschwanden et al., 2013) when we start the projections from the palaeo-spin-up and switch from fixed domain to free surface. Starting the free-surface simulations as early as possible is preferable in order to give the model the chance to recover from possible perturbations at the beginning.~~ While the MIROC5 model provides data starting at the year 1900, the CanESM2 and NorESM1 models start later in time at the year 1950. ~~For the latter two models, we randomly reshuffled the horizontal time slices (annual mean of~~ Therefore, there are no CanESM2 and NorESM1 data for the years 1900–1949. The MAR data consist of the MAR fields, which are longitudinally-latitudinally distributed fields for annual mean surface temperature, SMB and monthly surface runoff~~) from the~~ . Because the climate over Greenland changed relatively little during the 20th century, we took the years 1950–1999 ~~back in time to the years 1900–1949. This yields forcing data for the years 1900–2100 for all three~~ as sampling interval and determined randomly years out of this interval. The MAR fields of a single CMIP5 ~~models.~~

~~As ice sheets react on longer timescales, we needed longer scenarios and opted to prolong the scenario data until the year 2300. However,~~ model for these random years are assigned to MAR fields of the subsequent years inside the target interval, which is defined by the years 1900–1949. By this, we close the data gaps of CanESM2 and NorESM1 for the years 1900–1949.

For the years 2101–2300, there are no direct scenario data available from MAR for any of the three used CMIP5 models. ~~In particular, for RCP 8.5, we have the problem to choose a favourable sampling interval for the horizontal time slices. If we~~

~~choose~~ For the MAR fields from the RCP 4.5 scenario, we apply the same random procedure as described above to fill the data gaps, but with the sampling interval ~~too short, there are not enough time slices to be assigned to the time beyond 2100, and there is almost no variability. If we choose the interval too long, there is on overestimation of variability during the artificially prolonged interval 2101-2300 due the already present~~ 2091–2100 and the target interval 2101–2300. While for RCP 4.5 forcing the climate-warming trend ~~in the MAR~~ over the 21st century is moderate, it is stronger for RCP 8.5~~forcing~~ , see Fig. 3.

We found that the variability for the years ~~towards the year 2100. This problem is particularly prevalent for the anomaly in SMB.A sampling length of 10 years (years 2091–2100) is agood choice. Over this sampling interval, the horizontal time slices are repeatedly and randomly reshuffled forward in time to the years 2101–2300. We found that there still was an overestimation of variability in the prolonged data.~~

~~We circumvented this overestimation of variability for (and only for) the~~ 2100–2300 was relatively high for RCP 8.5 ~~scenarios by computing over the sampling interval the~~ scenarios, see Fig. A1, panel a. Therefore, we modified our random procedure for the MAR fields from RCP 8.5. For RCP 8.5 forcing, we took only those MAR fields from the sampling interval that belong to the lower 70 % of their total SMB anomaly. In other words, we excluded the MAR fields for the upper 30 % of the total SMB anomaly, see Fig. A1, panel b. Mathematically this reads

$$\Delta M_{\text{tot}} > \Delta M_{\text{tot}}^{\text{ave}} + 0.3 \cdot (\Delta M_{\text{tot}}^{\text{max}} - \Delta M_{\text{tot}}^{\text{min}})/2, \tag{B1}$$

with the temporal average, the maximum and the minimum of the anomaly of the total SMB, $\Delta M_{\text{tot}}^{\text{ave}}$, $\Delta M_{\text{tot}}^{\text{min}}$ and $\Delta M_{\text{tot}}^{\text{max}}$, respectively. ~~Then, we apply the condition~~

$$\Delta M_{\text{tot}} > \Delta M_{\text{tot}}^{\text{ave}} + 0.3 \cdot (\Delta M_{\text{tot}}^{\text{max}} - \Delta M_{\text{tot}}^{\text{min}})/2$$

~~in order to exclude time slices a with too positive total SMB anomaly. In fact, we consider 2-D fields where ΔM totals are below its average, while we consider only about the first 1/3 where ΔM totals are above its average. Note that the totals of anomalies of surface mass balance are negative in these scenarios.~~

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Tables

Table 1. Abbreviations in Fig. 1.

Abbreviations	Physical meaning
z_0	Observed present-day elevation of the GrIS
z	Simulated elevation of the GrIS
ΔT_{GRIP}	Reconstruction of temperature anomaly from GRIP ice core
ΔT_{s}	Anomaly of surface temperature simulated by MAR
ΔM	Anomaly of ~~surface mass balance~~ SMB simulated by MAR
ΔR	Anomaly of runoff simulated by MAR
T_{s}	Surface temperature
M	Surface mass balance (SMB)
R	Surface runoff
Q	Subglacial discharge into the given fjord
B	Bottom melt simulated by SICOPOLIS
W	Thickness of basal water layer
T	Ocean temperature (function of depth)
S	Ocean salinity (function of depth)
d	Submerged part of the outlet glaciers
M_{s}	Submarine melt of the outlet glaciers

Table 2. Simulated GrIS contribution to sea level rise for the years 2100 and 2300 in cm. ~~Columns~~ Rows specify the different GCMs used by MAR. ~~Rows~~ Columns list the RCP scenarios used by the MAR GCMs and whether we excluded or included the elevation SMB feedback $\partial M/\partial z$ in our simulation.

MAR GCM	Year 2100 [cm]				Year 2300 [cm]			
	RCP 4.5		RCP 8.5		RCP 4.5		RCP 8.5	
$\partial M/\partial z$	off	on	off	on	off	on	off	on
NorESM1	1.5	1.9	4.0	4.6	1.8	3.5	18.8	25.5
MIROC5	3.7	4.3	7.7	8.8	8.5	10.8	33.7	46.3
CanESM2	4.6	5.6	11.3	13.0	11.2	17.1	54.6	76.4

Figures

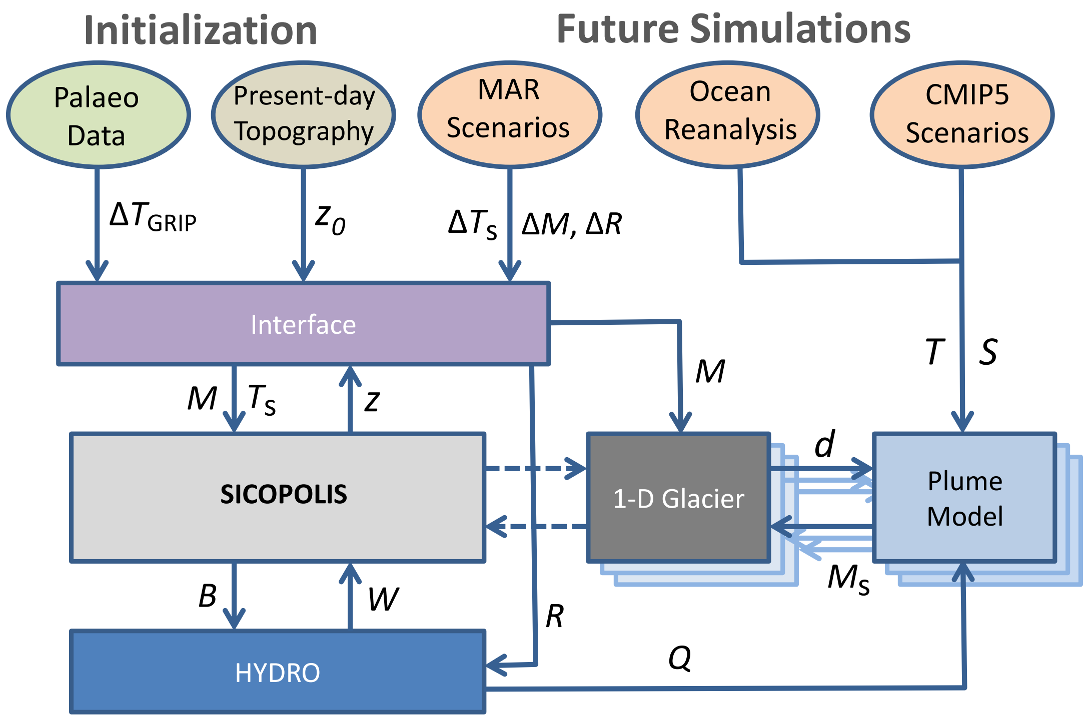

Figure 1. Flow diagram of the model IGLOO and the interaction between its components. The 1-D outlet glacier and plume models are generic models, i. e., they can be applied to each outlet glacier of the Greenland ice sheet. ~~Abbreviations~~ Coupling between the ice sheet model and the generic outlet glacier models is not implemented yet, denoted by dashed arrows. In this paper, coupling between HYDRO and the plume model is off-line. Simulations with the coupled generic outlet glacier models and plume models as well as details on the coupling between them are described in Beckmann et al. (2018a) . The exchange variables are explained in Table 1.

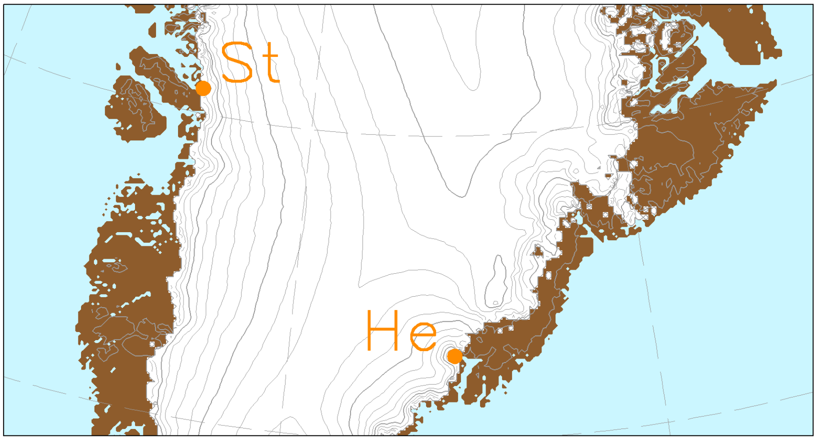

Figure 2. Geographical position of the outlet glaciers mentioned in the main text. “St” indicates the location of Store Glacier, while “He” marks the position of Helheim Glacier.

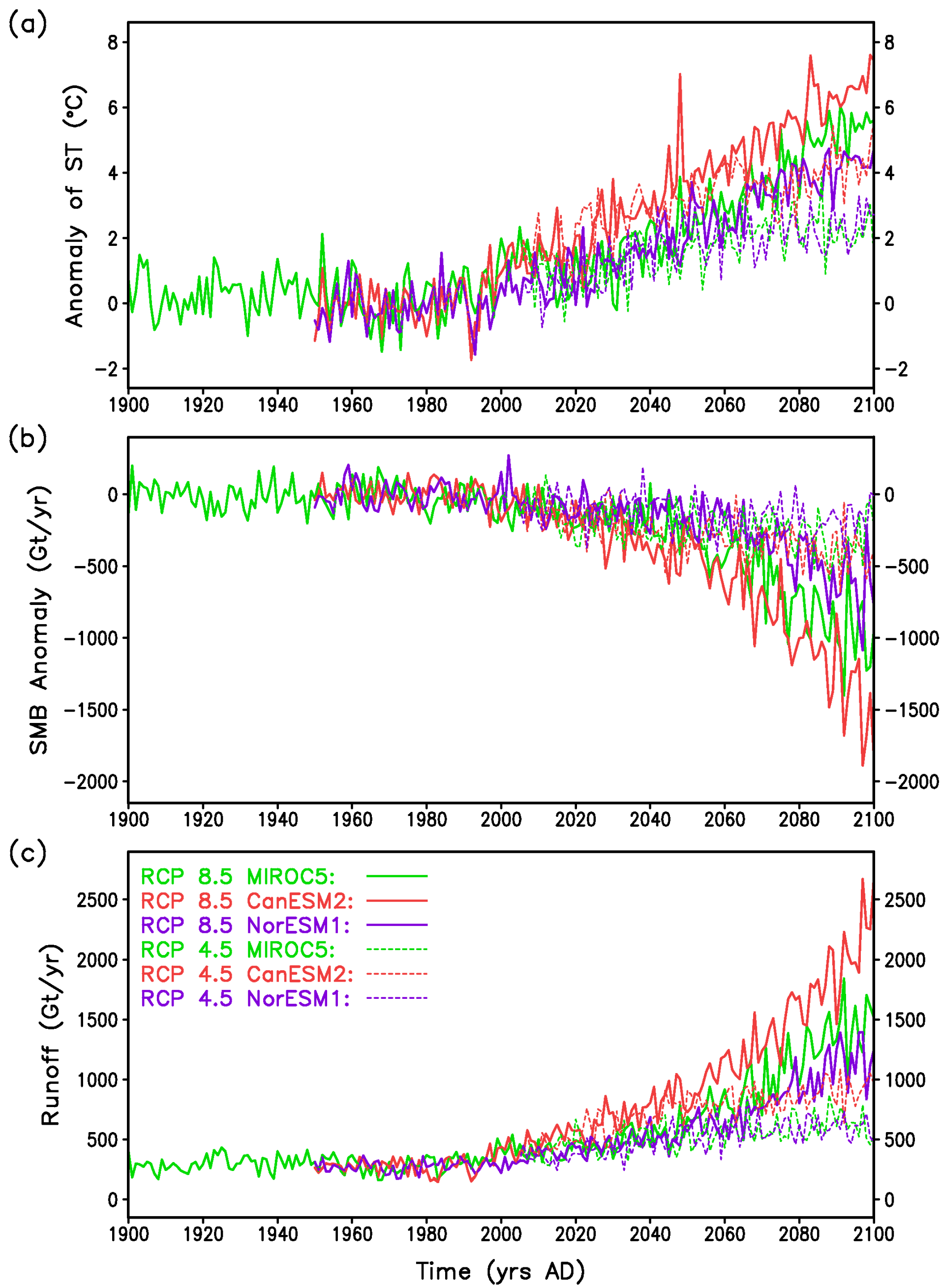

Figure 3. Forcings derived from the MAR regional model. (a) Anomaly of annual average surface temperature, (b) total annual ~~surface mass balance~~ SMB anomaly, and (c) total annual runoff. Anomalies are taken with respect to the period 1961–1990 from the respective CMIP5 models. RCP 8.5 scenarios are indicated by the solid lines, while RCP 4.5 scenarios are shown by the dashed lines.

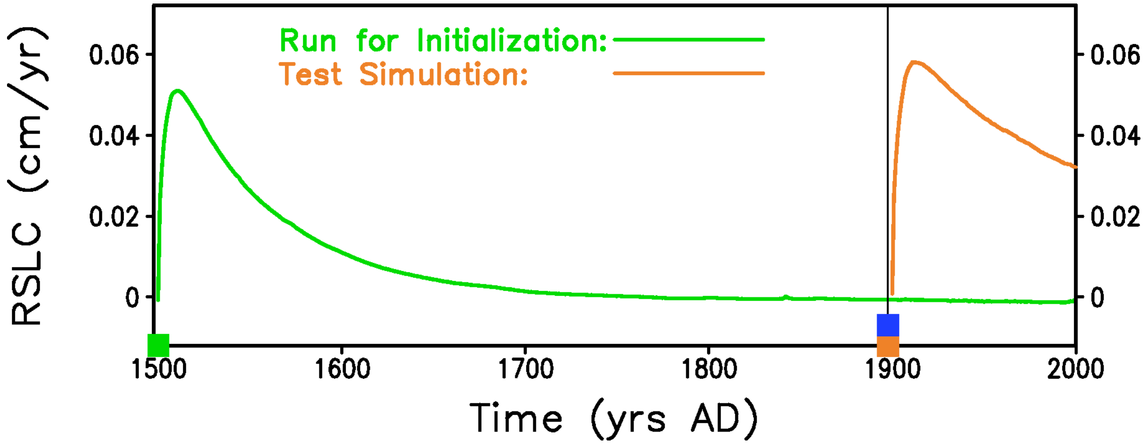

Figure 4. Time series of the rate of sea level change produced by the GrIS, illustrating the impact of the timing of the last two switches during the initialization. The green curve is from the simulation used for initialization of our future projections, which switches from the shallow ice approximation to the hybrid mode at the year 1500 (green square). The orange curve is from a test simulation (not used for initialization of our future projections), where we switch from the shallow ice approximation to the hybrid mode at the year 1900 (orange square). Both simulations switch from relaxing surface to free surface at the year 1900 (blue square), i. e., for the test simulation, the two switches appear at the same time.

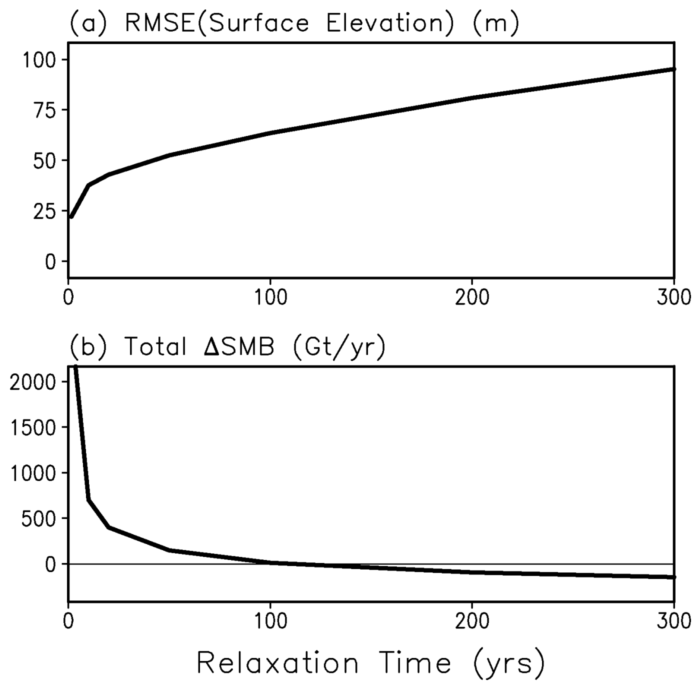

Figure 5. Total ice sheet quantities at present day against relaxation constant. (a) Root mean square error (RMSE) of modelled to observed surface elevation. (b) Total difference between our simulated surface mass balance and the ~~surface mass balance~~ SMB from the regional model MAR using ERA reanalysis 1961-1990 climatology.

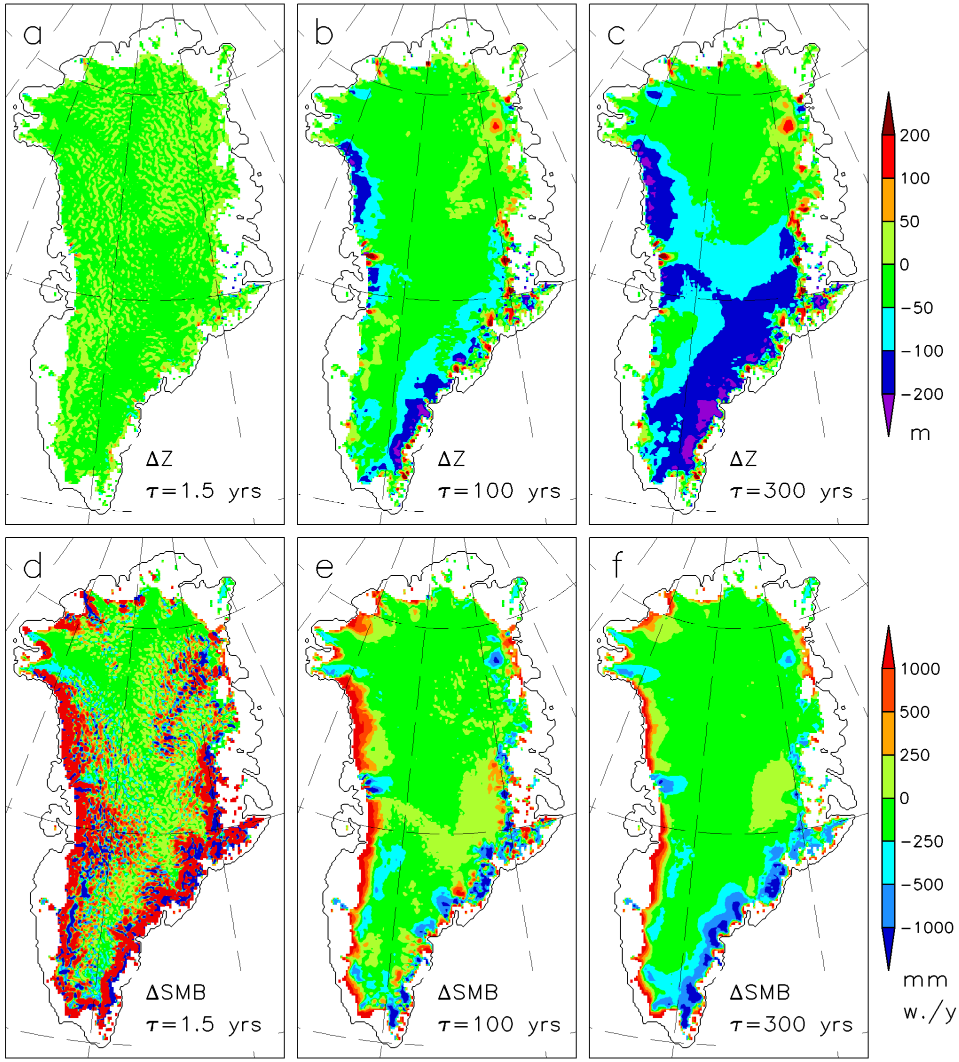

Figure 6. Differences between simulated and observed present-day 2-D fields for various relaxation constants, i. e., 1.5, 100 and 300 years. (a), (b) and (c): deviation of surface elevation from observed. (d), (e) and (f): deviation of our implied ~~surface mass balance~~ SMB from the ~~surface mass balance~~ SMB from the regional model MAR.

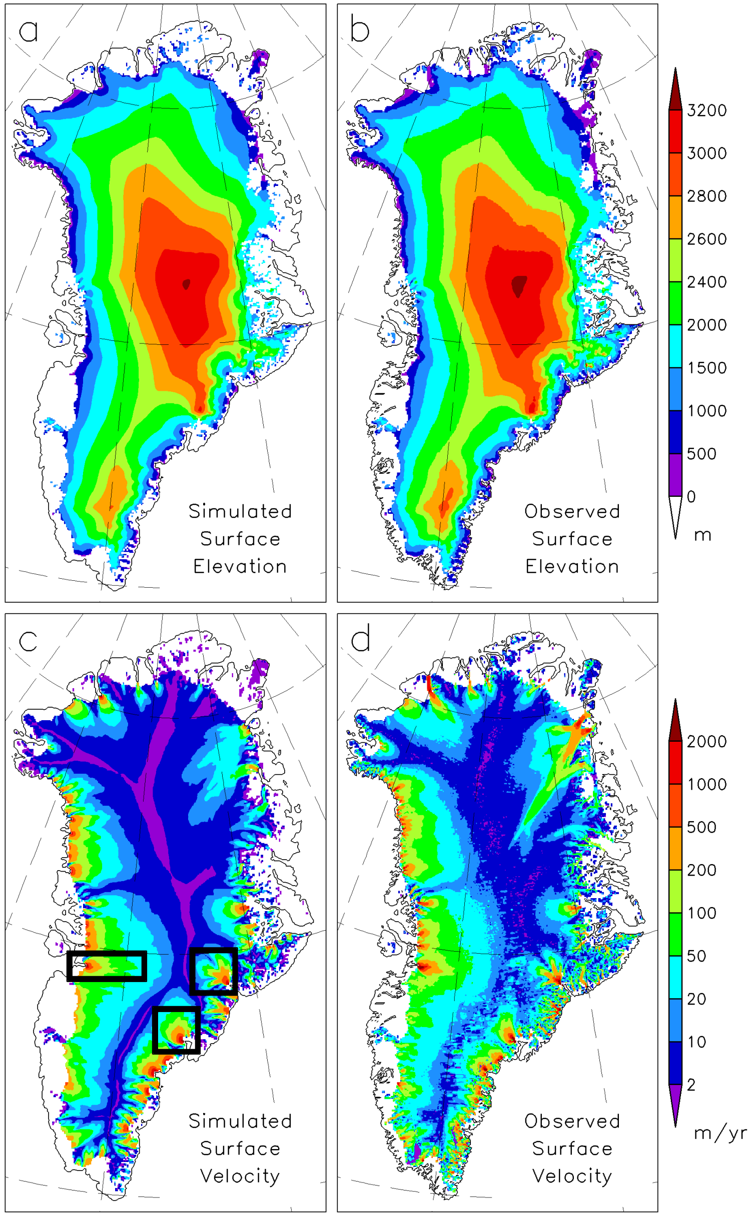

Figure 7. Comparison of ~~our~~ simulated ~~with~~ and observed present-day 2-D fields ~~for present-day~~ with 100 yrs relaxation constant. (a) Simulated surface elevation, (b) surface elevation by Bamber et al. (2013), (c) simulated horizontal surface velocity, and (d) horizontal surface velocity by Rignot and Mouginot (2012). The rectangles in panel c indicate the regions around Jakobshavn Isbræ, Helheim Glacier and Kangerdlugssuaq Glacier, which are enlarged in Fig. 8.

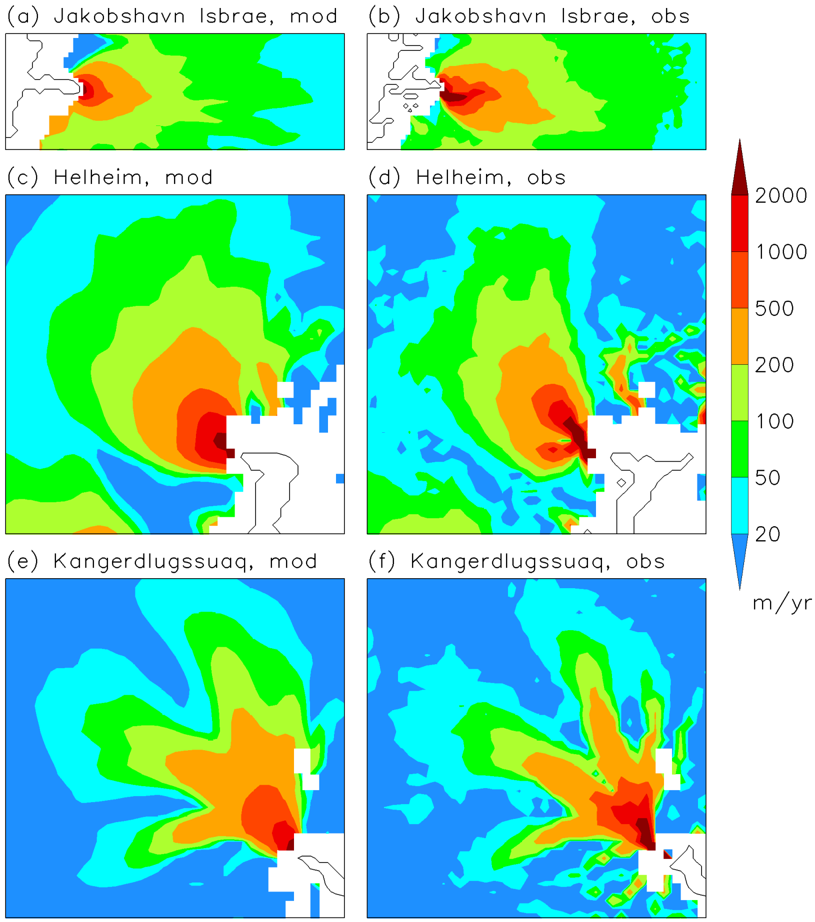

Figure 8. Comparison of ~~observed and~~ simulated and observed (Rignot and Mouginot, 2012) velocity for major ice streams and outlet glaciers. Left side: modelled, right side: observed. (a, b) Jakobshavn Isbræ, (c, d) Helheim Glacier, and (e, f) Kangerdlugssuaq Glacier. The location of the three regions in Greenland is shown in Fig. 7c.

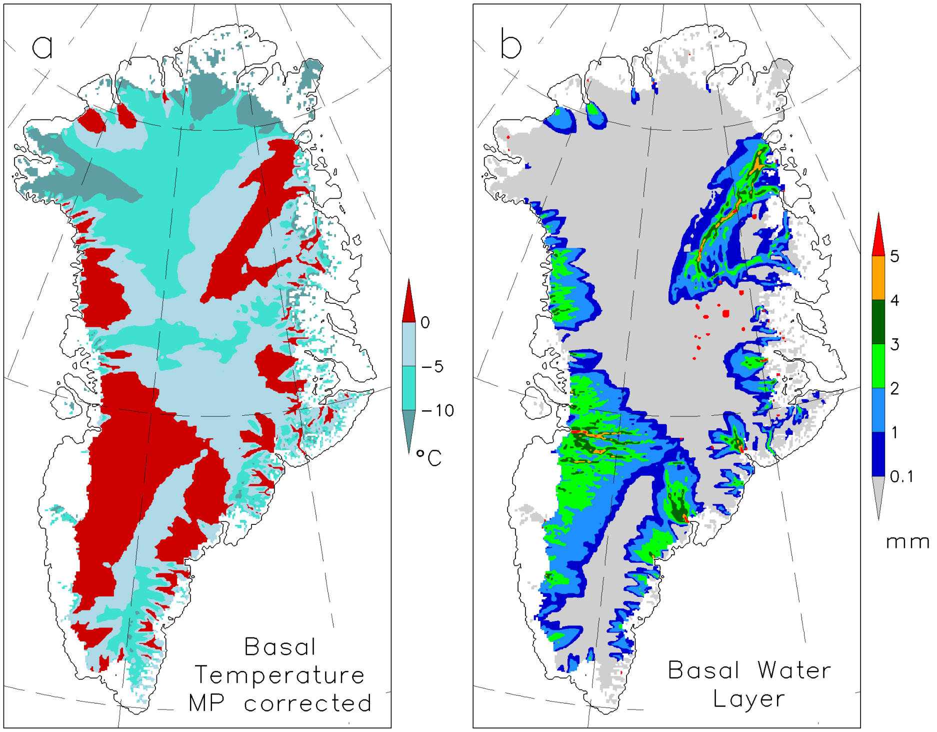

Figure 9. Simulated 2-D basal fields. (a) basal ~~temperatures~~ temperature relative to pressure melting (in $^\circ$C), (b) thickness of basal water layer (in mm).

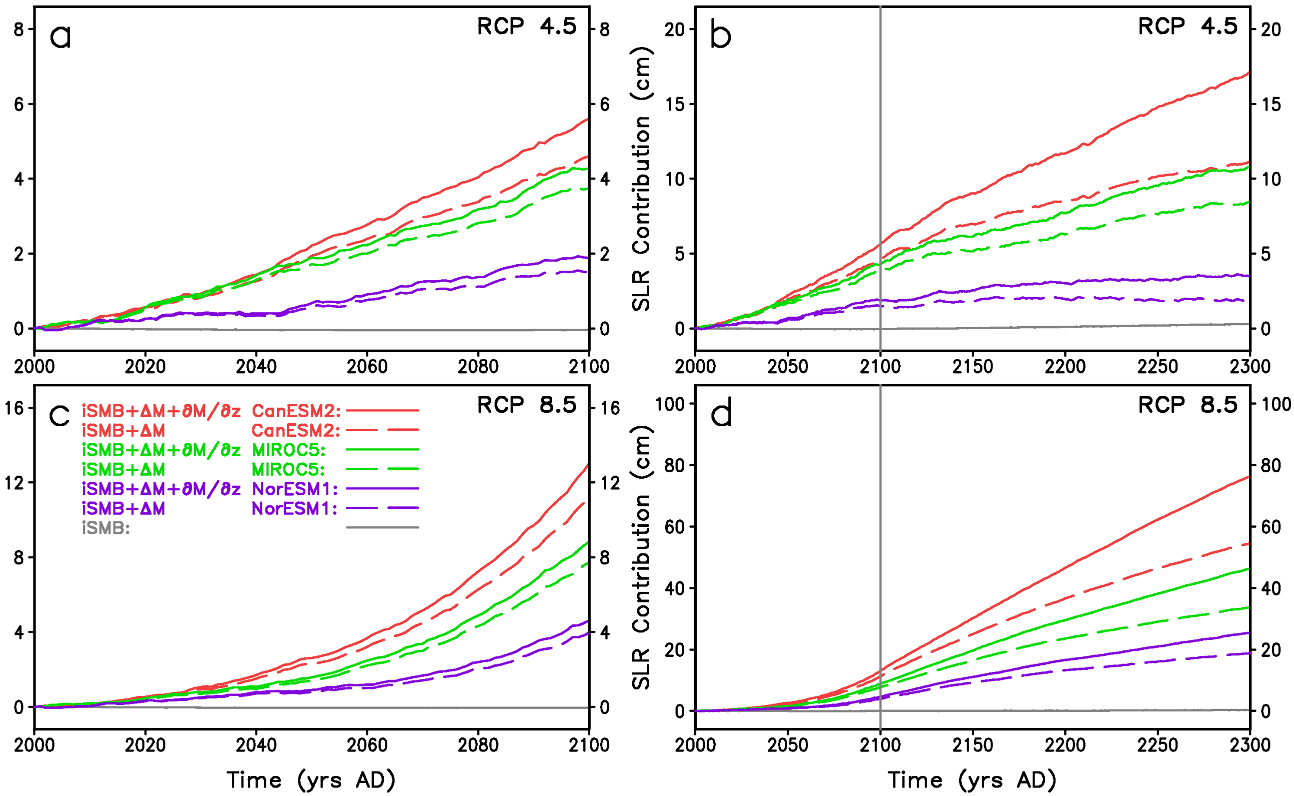

Figure 10. Contribution of the Greenland ice sheet to future sea level rise under MAR forcing for different scenarios. Sea level rise is referenced to the year 2000. Beyond 2100, the forcings of the projections are from prolongations of the original MAR data (see main text for details). This is indicated by the vertical grey line at the year 2100 in panels (b) and (d). RCP 4.5 projections: (a) years 2000–2100 and (b) years 2000–2300. RCP 8.5 projections: (c) years 2000–2100 and (d) years 2000–2300. The different CMIP5 general circulation models utilized by MAR are indicated by colours. Different line characteristics specify optimal simulations with (solid) and without (long dashed) elevation correction for the ~~surface mass balance~~SMB. The grey curves in panels (a) to (d) indicate a control simulation with solely the implied ~~flux~~ SMB (iSMB) as forcing. All simulations are with hybrid ice dynamics and HYDRO basal hydrology.

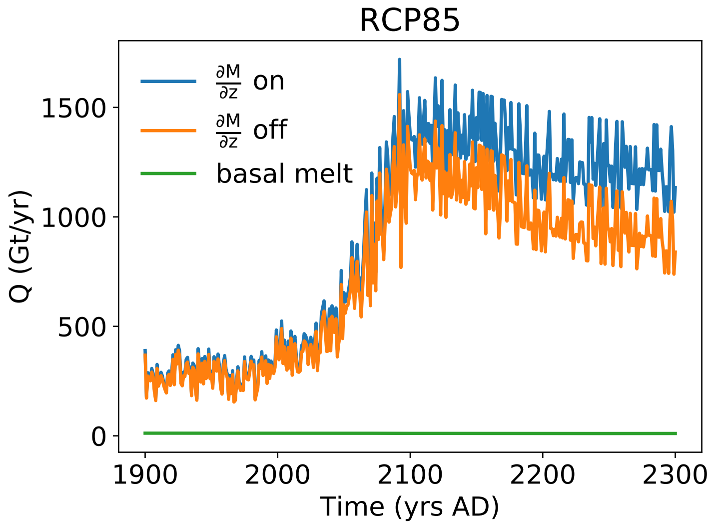

Figure 11. Time series of the components of subglacial discharge. The total basal melt (green) ~~amounts~~ is nearly constant in time and ranges from 10 to ~~approximately 15 Gt yr^{-1}~~12 Gt/yr. Total surface runoff with surface elevation SMB feedback (blue) and without the feedback (orange).

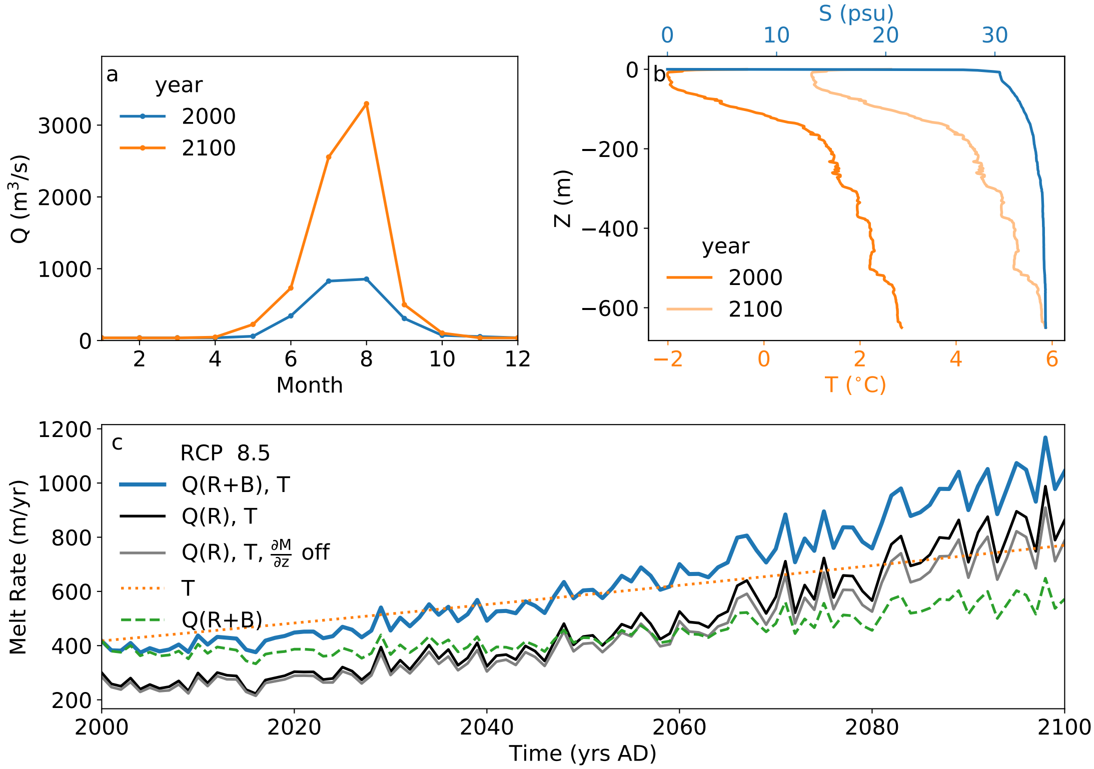

Figure 12. (a) Monthly subglacial discharge derived from runoff and basal melt (R+B) for Helheim Glacier and the scenario RCP 8.5 in the years 2000 and 2100. (b) Temperature-depth and salinity-depth profiles obtained from measurements ~~, with the temperature profile de- and increased by 0.03°C/a^{-1}~~ for the years 2000 and ~~2100 (Section 2.9).~~ 2100. The corresponding submarine melt rates are depicted in (c). The effects of increased temperature and discharge only (orange dotted and green dashed lines respectively), as well as the combined effect (solid lines) are displayed until the year 2100. Melt rates with subglacial discharge ~~or only~~ from solely surface runoff are depicted in black. Melt rates of subglacial discharge containing only surface runoff that was calculated without the surface elevation feedback are depicted in grey.

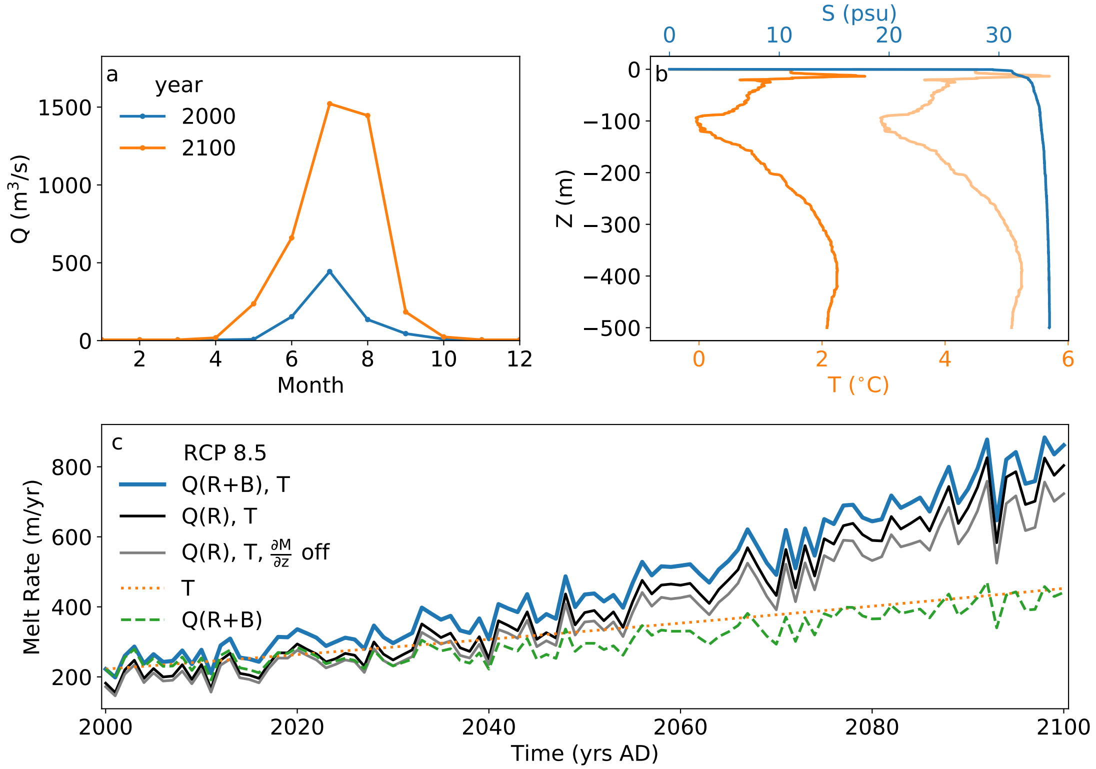

Figure 13. Similar to Figure 12, but for Store Glacier.

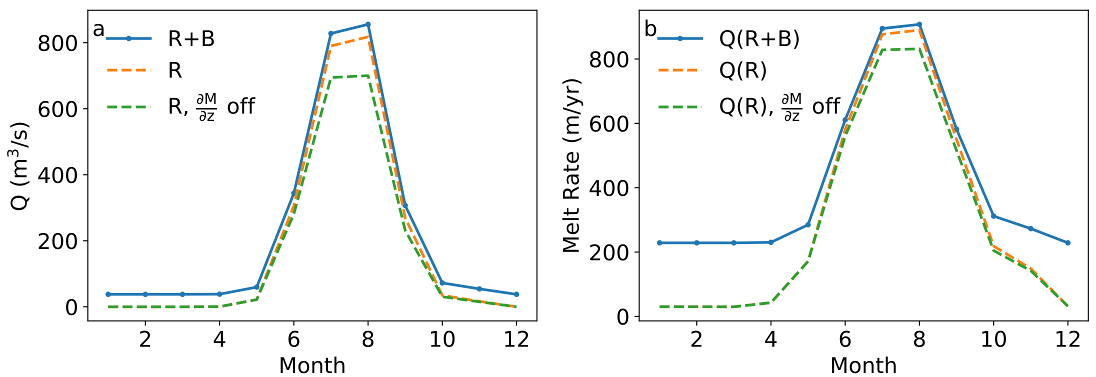

Figure 14. Subglacial discharge of Helheim Glacier (a) for the year 2000 determined by runoff (R) only (dashed lines), with and without surface elevation feedback (orange, green) and runoff together with basal melt (R+B, blue solid line). The corresponding submarine melt rates (b) with the same line colour and line style.

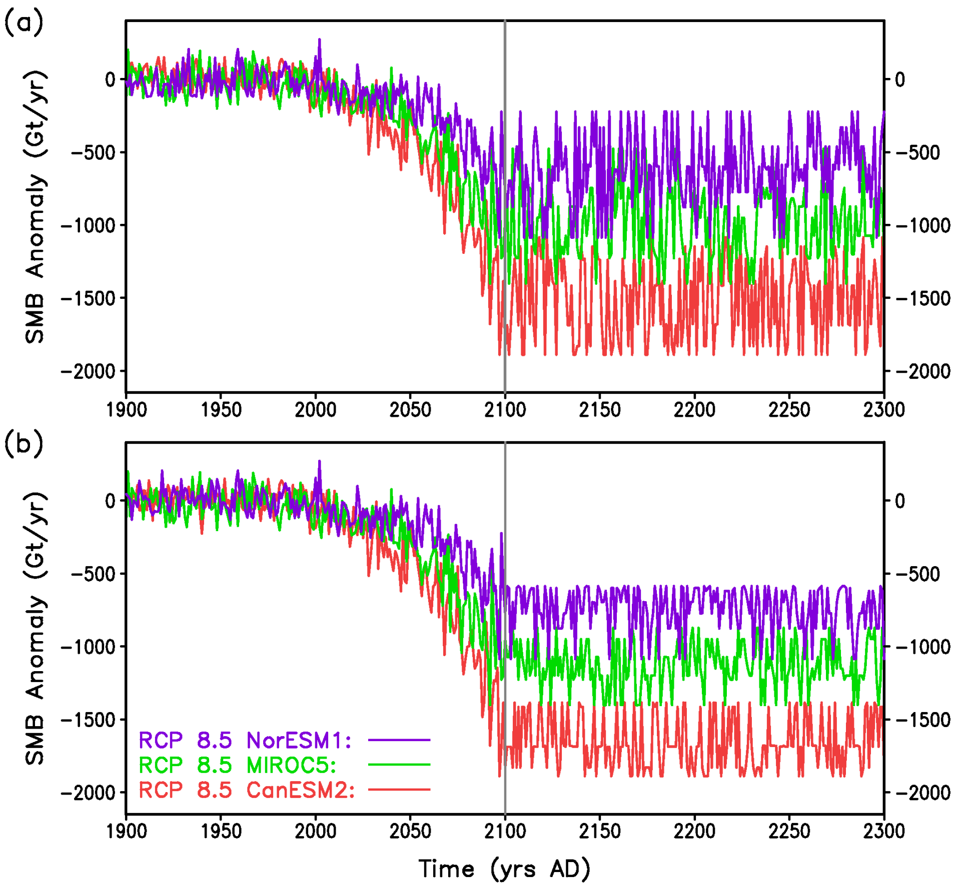

Figure A1. Prolongation of the MAR forcing illustrated for the SMB anomaly. (a) For the years 2101–2300, the SMB anomalies are taken from random years in between 2091–2100. (b) Variability for the years 2101–2300 reduced by 30 %. The vertical grey line at the year 2100 indicates the beginning of the prolongation.