Ice dynamic projections of the Greenland ice sheet in response to atmospheric and oceanic warming

Reply to List of Comments

by J.J. Fürst, H. Goelzer and P. Huybrechts

First of all we want to thank the two reviewers for the critical and useful comments they gave on the manuscript. All comments are considered and helped to significantly improve the quality of our work. In the following the responses to the reviewers comments are denoted in italic and indented.

Anonymous Referee #1

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General comments

The paper uses numerical modelling to assess the relative contributions of surface mass balance change and accelerated discharge from marine terminating outlet glaciers to Greenland ice loss during the next three centuries. Overall the paper is interesting and makes a good contribution. However, a couple of elements of the methodology need explaining in more detail, particularly the section linking ice velocities and oceanic forcing. There are a number of small errors in grammar / syntax that should be addressed. I give my detailed comments below.

> *We want to thank the reviewer for the overall constructive comments. These comments helped to improve the legibility and clarity of the manuscript. We have addressed almost all the comments in the revised manuscript.*

Specific comments

Abstract Line2: specify dates

> ***Corrected.*** *The last decade refers to the period 2000-2010, as is now added.*

Line 9: What about other potential controls on ice discharge, e.g. hydrofracture causing an increase in calving or loss of sea ice?

> *It is clear that ice discharge is influenced and controlled by many small-scale processes which effects are not well quantified to this day. On the floating tongue,*

these concern crevasse/damage formation and advection, buttressing from a mélange cover, weakening of shear margins, basal melting that depends on the fjord circulation and the basal melt-water discharge, etc. On the grounded part, ice discharge is potentially affected by a loss of lateral flow resistance or changes in basal friction due to variable water discharge through the basal drainage system. On top of this, the superposition of these processes are likely different for each individual glacier setup. These processes are not directly taken into account in this model application as its large-scale character inhibits to resolve them. The introduced parameterisation rather assumes a first order control from oceanic and atmospheric temperatures. We have clarified in the manuscript that the parameterisation does not aim at reproducing this complexity.

Adjusted Sect.2.2 (now 2.3) accordingly.

Line12: Why use only the low emissions scenarios for the longer runs? The introduction seems unnecessarily long and gives a lot of detail for each example, although the examples themselves are relevant. I suggest restructuring around the idea of oceanic forcing being important (which is the main point), so that the argument is more tightly focused and uses the examples more effectively.

The introduction has been shortened to be more concise and to the point.

For the reason why only the low emission scenarios were prolonged up to the year 2300, see the response to the comment by this reviewer made at P 3870 Line 12 further below.

P 3853 Line 15: Contradicts the previous sentence.

***Sentences removed** during the shortening of the introduction.*

P 3853 Line 19: Add Carr et al 2013.

Included citation.

P 3854 Line 13-17 & 25: Needs referencing.

Re-referenced Murray et al. (2010).

P 3854 Line 26: There is only evidence that the water can overtop sills and a few Greenland glaciers (the Straneo paper looks at 6). This implies it over tops all around Greenland, so please re-phrase.

Adjusted accordingly.

P 3854 Line 28: Indicate how the melange cover can impact on glacier retreat/calving.

The ice mélange is believed to play a role on the seasonal cycles at the glacier front by the mechanical back-stress it applies on the calving face, but the underlying processes are poorly understood.

A sentence was added to that effect.

P 3858 Line 13: How were they adjusted?

Added the reference to Goelzer et al. (2013). *There the following is specified:*

'We use a Cartesian grid on a polar stereographic projection with standard parallel at 71°N and standard meridian at 44°W, which differs from the standard meridian of 39°W used by Bamber and others (2013). Their dataset is re-projected and interpolated from the original 1km grid to the ice-sheet model grid of 5 km and a geoid correction is applied to reference the dataset to mean sea level. Since the model does not treat floating ice shelves, all floating ice is removed using a flotation criterion for an effective ice density of 910 kg m^3 and a sea-water density of 1028 kg m^3.'

P 3858: Line 21: Peak melt water doesn't necessarily mean peak ice velocities (Schoof, 2010; Sundal et al., 2011; Vieli et al., 2004). Also, the Rignot et al., 2010 paper focuses on submarine melting, rather than seasonal velocities, so I suggest using another reference.

Corrected accordingly *as follows: 'Observational studies often report on successive distinct speed-up events during the melt season (Zwally et al., 2002; Bartholomew et al., 2011; Andrews et al., 2014; Tedstone et al., 2014).'*

P 3859 Line23 & P 3860 line 24: How representative are these values for the rest of the ice sheet and how much does the choice of these parameters effect your results?

As mentioned in the manuscript, there are only a few direct observations on the link between speed-up, runoff and basal drainage. We also mention that the observations might be biased to the western flank of the Greenland ice sheet. Therefore the values are not representative for the entire ice sheet. Due to that reason, Shannon et al. (2013) conducted a comprehensive sensitivity study showing the effect of choosing these values on the centennial ice sheet evolution. They conclude that in any case the impact of basal lubrication on the centennial ice loss is no more than 5% of the integrated SMB. We already referred to this study but we clarify that the sensitivity of the projections to these values is considered small. The manuscript now reads:

'Within a comprehensive uncertainty study on the lubrication effect, Shannon et al. (2013) find that its effect is of secondary importance in terms of the centennial ice volume evolution. Therefore, only one set of parameters is used for the projections here.'

Adjusted Sect. 2.1 (now 2.2) accordingly.

P 3860: Line 15: Why not include these in the calculation of the parameter values?

Observations at these locations either are affected by the vicinity of the marine front or can only serve to give limits on the functional dependence.

Not adjusted.

P 3861 Line 1-11: This approach seems a substantial over simplification of the relationship between ocean melt and glacier velocities. First, there is very large variability in the response of individual glaciers to oceanic forcing within each region, which is likely due to localised topography, so we cannot assume that glaciers will respond to future ocean forcing in the same way as they have done in the past (e.g. if they are now on the far side of an overdeepening). It assumes that all changes in velocity are ocean driven and proportional to the forcing applied, which is not necessarily the case. Looking at the south-east in particular, the speed up between 2000-10 consists of acceleration, followed by deceleration, so looking at a decadal scale response could disguise important detail. I also do not understand how / why these are scaled to the entire ice sheet: is the assumption that all areas of the ice sheet respond to ocean temperatures in the same way? To take an extreme example, we cannot say that northern Greenland glaciers (with extensive floating tongues) will respond in the same way as south-western Greenland glaciers. I think this section needs better explanation and justification.

The reviewer raises important issues that we are not able to answer within the scope of this paper. Apart from the many involved processes that are not well quantified, the small-scale geometric details of these glaciers, that set the frame for their individual responses, are not resolved in this large-scale model application. Despite the expected non-uniform response of glaciers, the pattern of recent glacier accelerations is, to a certain degree, consistent with the variability in offshore ocean temperatures around Greenland (Straneo et al., 2012). Here, we make a first attempt at linking changes in ocean temperature to changes in ice discharge from a limited amount of direct observations and poorly understood processes. The parameterisation is therefore kept simple but is still believed to capture first-order processes informed by more detailed modelling of the most important outflow glaciers (work from Faezeh Nick).

In response to the reviewers concerns, we clarify the deliberate choice for a simple relationship between changes in ocean temperatures and in ice discharge.

Added a passage in Sect. 2.2 (now 2.3) and adjusted the conclusion accordingly.

P 3862 Line 1: Has this distance been used in other modelling studies, e.g. further south on GrIS?

This treatment was inspired by previous work on the inland transmission of marginal pertubations at marine margins with the same ice-dynamic model. This distance was found as the typical distance over which changes at the calving front were instantaneously transmitted by longitudinal stress coupling (Fürst el al., 2013)

Added a reference to Fürst et al. (2013).

P 3863 Line 5: Why were these parameter ranges selected?

Parameter ranges are centred around values obtained from a previous tuning of this flow model within ranges informed from past experience with the model. The range for the positive degree-day parameters is chosen rather narrow as the *model spin-up is very sensitive to them. The ranges for the sliding coefficient and the rate factor are larger, as they are not as well constrained. The range selection obtains additional justification because the best-fit parameter combination does not hit the limits.*

Added a sentence for clarification.

P 3865 Line 7: Do you see spatial patterns in precipitation, as well as temperature?

As we could not find a general pattern in the AOGCM precipitation changes, we focussed on the average changes over Greenland, showing an increase.

Added the following sentence. *'The patterns of future precipitation changes are also AOGCM dependent and cannot be generalised'*

P 3865 Line 15 onwards: needs references and could be explained more easily by adding these currents and water body names to Fig 4.

Added reference and locations of ocean currents and water bodies to Fig.4.

P 3866 Line 3-6: How representative are these offshore ocean temperatures of what is happening at the glacier front? Is there sufficiently detailed bathymetric data available to identify sills that might block warmer water from entering glacier fjords? Also, how valid is it to use lines of latitude as boundaries for your oceanic units? E.g. using a divider at 70 N means that glaciers located up to 3 degrees north of the Denmark Strait are included in the south-east region. However, it is unlikely that warm water from the Irminger Sea will penetrate this far north and will be much less prevalent than on the south-east coast of Greenland.

We are aware of all the complications of waters penetrating to calving fronts, details of which are badly understood. That's why we introduced a parameterisation linking modelled ocean temperatures to observed ice discharge fluctuations calibrated to observations and more detailed modelling results. In this parameterisation, offshore ocean temperatures are an indicator for the potential temperature changes near the ice front. This finds justification from the fact that warm water at depth near six glaciers all around Greenland has an Atlantic source (Straneo et al., 2012; Jackson et al. 2014). Intrinsically, the parameterisation assumes a direct scaling relation between the offshore and fjord temperature changes. This assumption very likely breaks for short-term warming events and ignores any delays in the ocean system. Yet, adding more details is not warranted by the current understanding of the issues at stake. We assume that the parameterisation captures first-order effects in a more or less realistic way.

Concerning the delineation, we agree with the reviewer that this is just a first attempt to distinguish regional differences. Concerning the dynamic response, results are presented and discussed with respect to an average ocean warming in all basins. Consequently, a sensitivity study on the basin delineation would not add much to the general findings of this manuscript.

Added paragraph on limitations in Sect. 2.2 (now 2.3).

P 3866 Line 10: How much does your choice of depth averaging influence your results?

We present a comprehensive discussion on the sensitivity to the single value (now α) in the parameterisation for ocean-induced discharge changes (Eq. 3). The manuscript was completed by a sensitivity analysis on α. We expect that different choices in depth-averaging of the temperature will show a similar sensitivity, as both together determine the effect on the discharge. Yet we admit, that a decrease in the layer thickness, used for averaging will increase the amplitude of any variability and favour short-term changes. As the parameterisation is meant for long-term trends in temperature changes, the smaller the layer thicknesses the less appropriate the approximation. Averaging over a larger depth is expected to reduce the signal which could be accommodated by increasing α. We therefore think that the added sensitivity study on α largely covers the sensitivity to changes in the temperature averaging.

Addressed by added sensitivity analysis (Sect. 5 + additional table) on other parameter in response to comment from reviewer #2

P 3867 Line 9: Can you quantify 'fairly well'? Some areas look quite different, e.g. north east Greenland.

As the reviewer is certainly aware of, a direct differencing will only highlight inherent differences that arise from the spin-up method. We therefore followed a qualitative assessment to the best of our abilities, as the spin-up is not tuned to exactly reproduce the observed velocity field. Even for inversion methods that aim at reducing the mismatch between modelled and observed velocities under a given geometry, a qualitative comparison is preferable (Morlighem et al., 2013). As we argue in the text, it seems more relevant to capture the regional distribution of ice discharge.

No extra quantification necessary *as presented differences in ice discharge seem more relevant.*

P 3870 Line 12: Why use the two lowest scenarios? Why not do the highest ones as well?

For RCP6.0, many AOGCMs simply did not continue the projections beyond 2300. For the high impact scenario, available AOGCM data was limited to a few models that showed a highly divergent response, making our ensemble approach questionable. A prolongation of the RCP8.5 ice volume projections would therefore only reflect the dominant uncertainties in the modelled climate response, and undermine the relevance of climate modelling. Therefore, we decided to focus on the first century response.

Reformulated sentence as follows: *'As AOGCM input was not available for RCP6.0 beyond 2100 and as the divergent response of the few AOGCMs under RCP8.5 is not considered compatible with our ensemble approach, projections were continued until 2300 AD only for the two lowest scenarios.'*

Technical corrections

There are a number of minor errors in grammar / syntax and placing of brackets around references throughout the paper, but particularly in the earlier sections. I have highlighted some here, but it would benefit from a detailed proof read.

> ***Corrected.*** *In the processes of re-editing the manuscript, all authors paid attention to detect minor issues on grammar and syntax. We therefore hope to have been able to remove most of them during this revision. We verified the placement of brackets around citations/abbreviations and could remove some inconsistencies.*

Throughout the paper, 'but' is used at the start of certain sentences. Although this is not technically incorrect, it looks colloquial, so please change.

> ***Corrected by adjusting the respective sentences.***

Abstract Line 5: 'with a relative contribution of 40 and 60% respectively'.

> ***Corrected*** *placement of 'respectively'.*

Line 13: This sentence has grammatical errors and is hard to follow.

> ***Corrected by splitting the sentence to improve legibility.***

P 3853 Line 9: In>During

> ***Corrected as suggested.***

P 3854 Line 18: Petermann Glacier (delete 'the')

> ***Corrected as suggested.***

P 3854 Line 24: with>of

> ***Corrected as suggested.***

P 3856: Line 17: 'Here we include more: : :'

> ***Corrected as suggested.***

P 3856: Line 19: 'with the aim of better assessing: : :'

> ***Corrected as suggested.***

P 3856: Line 20: (RCP) (Moss et al., 2010).

> ***Corrected*** *by rearranging the sentence structure*

P3857 Line 1-3: The model evaluation in the recent past: : :.. and the sea level projections for the Greenland ice sheet are presented in Section 5'.

> ***Corrected*** *by reformulating the passage following to the comment.*

P3857Line 19: delete 'representative'.

Corrected as suggested.

P 3858: Line 21: on>of

Corrected as suggested.

P 3859: Line 2: 'the annual increase in sliding, relative to the winter reference'.

Corrected *in the course of reformulating this passage.*

P 3859: Line 13: is preferred> develops.

Corrected as suggested.

P 3860: Line 2: beyond>above

Corrected as suggested.

P 3860: Line 7: 'the mass balance model used here'

Corrected.

P 3862 Line 11: In order to initialise to the: : :

Corrected as suggested.

P 3862 Line 13: : : :.regional surface temperatures, precipitation and sea level.

Corrected as suggested.

P 3862 Line 14: Although the general approach is unchanged [unchanged from what?]: : :. (Appendix A).

Corrected *by adding reference to Huybrechts (2002).*

P 3863 Line 9: separate these criteria using semicolons for clarity.

Corrected as suggested.

P 3864 Line 4: as is often done> as in previous studies.

Corrected as suggested.

P 3864 Line 16: : : :to avoid any potential bias associated with the mean states: : :.

Corrected *according to suggestion from the reviewer.*

P 3864 Line 21: precipitations>precipitation.

Corrected as suggested.

P 3865 Line 5: For a given latitude, the difference in warming between the east and west of the ice sheet depends strongly on the individual AOGCM.

Corrected as suggested.

P 3865 Line 14: inspired> based on / determined from

Corrected as suggested.

P 3869 Line 21: Combine this and the previous sentence and alter to 'but the AR5 is the first to attempt to quantify: : :.'

Corrected as suggested.

P 3869 Line 24: ': : :suffers from including multiple studies that do not have forcing factors or setups that are directly comparable when: : :..

Corrected *by reformulating this passages to make it more concisive.*

P 3870 Line 8-9: twice as high as other RCPs.

Corrected as suggested.

P 3870 Line 24: For runs extending to both 210 and 2300, the sensitivity: : :.

Corrected as suggested.

P 3871 Line 7: This comprises both directly induced: : :.

Corrected as suggested.

P 3872 Line 26: this is because many of the smaller glaciers: : :

Corrected as suggested.

Anonymous Referee #2

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This manuscript assesses the response of Greenland Ice Sheet over the next 1-3 centuries to climate change as simulated by climate model output. This is a useful and pertinent exercise (but which has been carried out in various forms by other groups, and also previously by these authors). Thus, the paper makes a contribution to our understanding of the cryospheric response to climate change, though I think the authors need to contrast against previous studies to highlight why exactly this study is unique enough to justify a new publication. More importantly, I have identified a few major issues in their methodology and presentation which I have tried to highlight below, that I feel should be addressed before this manuscript can be published.

Indeed, other studies have focused on single aspects such as the lubrication effect (Shannon et al. 2013) or the generalisation of the dynamic response of individual outlet glaciers under future warming (Goelzer et al., 2013). The former study presents a comprehensive uncertainty analysis while the latter aims at a best projection relying on additional input for speedup from a flow-line model and SMB changes from a regional climate model. Consequently, both studies concentrated on a single climate scenario. Similarly, SMB projections with regional climate models often focus on individual climate trajectories. Here, we follow an ensemble approach for a suite of AOGCMs to assess the range of the future ice sheet response. In addition, this study comprises all 4 climate scenarios / emission pathways, suggested for the last IPCC AR5. Informed by the earlier studies, our approach accounts for basal lubrication and ice discharge increase due to ocean warming, and uses a higher-order ice-flow model incorporating longitudinal stresses. On the basis of a poor knowledge on how efficient offshore waters enter the local fjord systems on Greenland and assuming a first-order control from atmospheric and oceanic temperatures on the variety of small-scale processes controlling ice discharge, we decided for a simple parameterisation in our large-scale model application. The parameterisation finds justification from the fact that the regional acceleration pattern is consistent with an assessment of offshore temperature variability around Greenland (Straneo et al., 2012; Jackson et al., 2014). The manuscript is therefore a first attempt to explain and project the ice discharge response with temperature changes in the ocean.

The above discussion is recurrently addressed under different angles throughout the manuscript (and prominently in the abstract and the conclusion). The authors are therefore convinced that this study is sufficiently contrasted against previous work.

General comments

General: enhancement of mass loss due to warmer ocean temperatures seems an important result, however, this is based on a very simple (one-equation) parameterization of the influence of ocean forcing on the ice sheet. Thus, the quality of this important parameterization becomes quite important to one of the ‘takeaways’ of this manuscript. Yet no example of the sensitivity of this parameterization to, e.g.: the value 5.2 in Equation 3, the breakdown of ocean basins, the distance up-glacier to which the parameterization is applied, the assumption of uniform outlet glacier response, etc. etc., is documented. It seems incumbent on the authors to assess the sensitivity of their results on the nature and constant values of this equation, so that they can assert that the apparently significant effect of marine-based acceleration they find is potentially realistic, given the very simple parameterization used.

This comment is largely congruent with the major concern of reviewer #1. Main aspects have been addressed in detail in the answers to this reviewer. The new aspect here is the sensitivity of the projections to the single parameter (now called α) in Equation (3). We agree that without knowing the sensitivity, the projections are difficult to assess.

Corrected by including a paragraph and a table on the sensitivity. *When the model was initially set up for the presented projections, the determination of this parameter was an essential step. For 3 values, projections for the entire ensemble and all scenarios were computed. These results are now briefly presented in Sect. 5. The main messages are:*

1. The choice of this parameter mainly influences the present day contribution from ice discharge increase to total ice loss. If the parameter is chosen such that this partitioning reproduces the observational record, we expect a sensitivity of the projected sea-level contribution of about 15% by 2100 and less than 10% by 2300. This is significantly below the spread in projected ice loss introduced by the climate models.

2. The sensitivity of the sea-level projections decreases with increasing amplitude of the future warming and with the length of the projections.

3. The RMS deviation of the ensemble ice loss for single RCPs is very robust under the different choices of this parameter.

General: there is significant emphasis placed on matching recent ice discharge trends, yet the model is forced with ocean forcing from ocean models, which cannot be expected to match the phase of observed climate variability.

We admit that the model is not forced with ocean temperature reanalysis data for the reference period 1958-2005. As we do not expect that the AOGCMs reproduce recent trends in atmospheric and oceanic warming around Greenland, it indeed has no sense to tune the discharge increase to observations. As mentioned in Sect. 2.2 (now 2.3), the tuning aims at the relative contribution of ice discharge and surface mass balance to the total ice loss of the model ensemble. The only assumption is therefore that AOGCMs are able to reproduce the relation between oceanic and atmospheric warming realistically, as this directly controls the relative ice loss contributions under a warming.

Rephrased passages *in Sect. 4. (in line with later comments) to clarify the difficulty of a direct comparison of observed and modelled changes in ice discharge.*

General: some main conclusions of the paper are not new, in that they have already been documented in other papers (including papers by the present authors).

This comment is similar to aspects raised in the reviewer's initial remark and is briefly discussed here. ***For details refer to answer to the first remark.***

The novelty of this study lies in the ensemble approach covering multiple AOGCMs and the four RCP emission scenarios, suggested for the AR5. Informed by earlier studies, two parameterisations are suggested that address the dynamic response of the ice sheet to ocean and atmospheric forcing. Altogether, this allows an assessment of the full range of the future evolution of the Greenland ice sheet under many possible climate trajectories. In this way, the presented study clearly stands out against earlier work.

General: ensure all figures are referenced in the text, and in order.

We found that all figures were referenced in the right order. We assured this also for the revised manuscript.

No correction necessary.

Specific comments:

Abstract:

Line 6: 'initialized' often seems to refer to inverse or adjoint-based procedures these days. I recommend 'spun up' or similar term, instead

Corrected *in the entire manuscript as suggested.*

Line 11: SLR referenced to which base sea level? Present day?

Corrected. *Base line is the year 2000.*

Line 12: why only low emission scenarios considered to 2300?

Refer to response to reviewer #1.

Added explanation in Sect. 5.

Line 19: refer to other papers which also find a minimal contribution from basal sliding to long-term response

Corrected by adjusting the abstract. *The references itself are given in the main text of the manuscript.*

Line 23: also refer to other papers which find large climate-side uncertainty (mentioned more below)

Corrected *according to reply to later comment in this review.*

P3853L7: suggest, refer to Enderlin et al for updated breakdown of surface melt vs. marine ice loss (system is trending towards relatively more surface melt)

Corrected. *Two references were added and main findings are explicitly specified.*

P3853-3854: this detailed overview of recent yearly-scale ice sheet 'weather' is interesting, but perhaps out of place or at least overly detailed, in a paper that looks at the long-term ice sheet 'climate' response. Suggest a summary that is relevant to the simulations detailed later in the paper.

The introduction has been shortened to be more concise and to the point.

P3855L2: if there's a physical explanation it isn't a coincidence

Adjusted the formulation *as follows: '[…] simultaneous occurrence […]'*

P3855L8: suggest referencing recent meltwater–velocity studies that suggest that annually-averaged effect might be small.

Corrected. *A recent reference is added (Tedstone et al., 2014) suggesting that the effect from lubrication on summer to winter differences might be small.*

P3855: negative trend since 1990s may be largely due to NAO, and not a long-term secular trend - see Fettweis work.

Reformulated. *We want to avoid invoking any notion on whether the recent SMB decrease is linked to long or short-term climate variations/trends.*

P3856,L5: "The physical complexity..." this sentence is unclear to me.

Reformulated *as follows: 'Computational constraints typically limit RCM applications […]'*

P3856,L16: "This feedback..." I do not understand this sentence. For example, it is not only ice discharge that affects geometry, and therefore SMB. In fact SMB affects geometry as well.

Deleted sentence *as it did not add anything.*

Section 2: suggest making separate "Ice sheet dynamics" and "SMB" sections

Adjusted *by adding the relevant sections.*

P3856L17: Perhaps it is worthwhile to note other efforts to model more ice dynamical processes in ice sheet models... for example, others have certainly pubilshed ocean forcing effects, and/or runoff-based basal lubrication experiments. To what extent are your experiments a unique contribution?

The comment addresses a passage where a short overview of the manuscript is presented, not suited for full referencing. Relevant literature is given in Sect. 2 where the parameterisations of the dynamic processes are specified and in Sect.5 during the discussion of the results.

No action taken.

P3857L11: I think you need to expand on the description of the nature of the 'higher order' ice physics in this paper, even if it is described more fully in other publications. Especially given that a main point of your paper is "Here we use a higher-order ice flow model..." (third sentence of abstract). At this point the nature of the higher order physics is mysterious to readers of this paper.

Added specific terminology. *Reference to Hindmarsh (2004).*

P3857L17: the term "parametric SMB model" is not clear - to me, this refers to an SMB model that, for example, parameterizes things as a function of latitude... which I don't think is the case here.

Though we think that the term is self-explicatory, the authors want to avoid any misunderstanding and the concerned passages were rephrased.

Rephrased passage for clarity.

P3857L26: perhaps more justification would be good here, as to why you think PDD method is robust, especially in the far future, and with constant assumed variability in daily temperatures.

The PDD model lends its robustness from a basically correct simulation of present-day annual mass balance characteristics and an acceptable simulation of the past ice-sheet history. No attempt was made to change these parameters in the future owing to a lack of solid information on how these might change.

No action undertaken.

P3858L1: does the snow model have multiple layers? Is there a multi-year memory of capillary water, and how does the capillary space compact with time? More explanation here as to the complexity/simplifications of the snow model would be good.

The snow model consists of only one layer. All capillary water is assumed to refreeze over the next winter turning the saturated snow into superimposed ice.

Additional details mentioned in text.

P3858L8: suggest a quantitative comparison to back up this currently unsupported statement of a good comparison to RACMO variability.

R^2 coefficients of determination between annual mass-balance components from our model and RACMO2.1/GR for the period 1958-2010 were 0.79 for SMB, 0.84 for precipitation, and 0.75 for runoff (Hanna et al., 2011). Such values for R^2 are considered to indicate a good agreement. A quantitative comparison between the SMB model, used here, and other RCMs is presented in Sect. 4 as evaluation during the reference period.

R^2 correlation coefficients added in text. *Quantitative comparison in Sect. 4.*

P3858,L10: suggest moving this description of ISM discretization up to an "Ice dynamics" section

Corrected as suggested.

P3858L12: "geometric input": do you mean bed topography?

Indeed, the model primarily requires the bed topography as input. However, information on ice thickness and surface elevation is also required for the initial model setup and when it comes to the evaluation of the present-day state after the free evolving spin-up phase (see Sect.2.3 (now 2.4)). Therefore we keep 'geometric input'.

No adjustment was considered necessary.

P3858L13: "slight adjustments": describe briefly for completeness

Added reference to Goelzer et al. (2013). *For details see response to review #1.*

P3858L17: can you describe these Gaussian functions more, or provide a reference that explains them?

The Gaussian functions correct the background geothermal heat flux in the vicinity of ice-core sites with (i) the value derived from temperature calculations at ice-core sites where basal temperature has been measured (ii) a gradual decrease of the difference with respect to the background field with distance according to a Gauss function with a standard deviation of 100 km radial distance. The method is described more fully in Pattyn (2010): Antarctic subglacial conditions inferred from a hybrid ice sheet/ice stream model, Earth and Planetary Science Letters 295 (2010) 451-461, doi: 10.1016/j.epsl.2010.04.025

Additional explanation and the reference to Pattyn (2010) is added.

P3859: Schoof (2010) and others don't so much relate sliding to annual average (or cumulative?) runoff, as much as to large, individual events. Also, see, e.g. new paper by (Andrews et al., Nature). So, it isn't so much the values integrated over a year, but more the amount of discrete events: : : Based on this, I think your justification for your particular sliding law needs more justification, even if it turns out it isn't important.

We agree with the reviewer that the research interest is often in singular or successive but distinct speed-up events rather than the cumulative effect. The melt component of the SMB model however works with monthly mean temperature fields using a statistic to account for daily variations. Thus the model is not meant to produce individual melt peaks and therefore our interest is in the integrated effect. Sundal et al. (2011) argue that mean summer speed-up is positively correlated with daily runoff as long as runoff rates do not exceed a certain threshold. As this timescale is not explicitly resolved in the SMB model, we decided to work with annual values. We therefore accept that a certain value for local annual runoff can be obtained in various ways over the melt season. In general however, the annual value should scale with the periods during which a critical daily runoff is exceeded.

Reformulated paragraph with main reference to Sundal et al. (2010).

P3859: is S_BL spatially varying? Or applied ice-sheet-wide?

Yes, S_BL varies spatially. An ice sheet-wide average would not have much meaning.

No correction necessary.

P3859,L14: so if no surface runoff, then S_BL=1?

The reviewer is correct. ***No action undertaken.***

Equation 2: Can you possibly refer to the plot of this relationship here, instead of a few sentences lower?

Corrected as suggested.

Equation 2: can you provide more physical and/or theoretical basis for this equation? At present is seems quite arbitrary to the naive reader why this form was chosen.

We justify our choice by referring to the best-fit parameterisation of the lubrication effect in Shannon et al. (2013).

Added reference.

Equation 2: are you solving for the basal drag as part of your force balance, or does it come from some prescribed basal drag field?

In this model variant, the basal sliding coefficient (A_S) is modified spatially by the two parameterisations. If runoff is locally non-zero, sliding is increased following the lubrication argumentation. Close to the marine margins, A_S is increased/decreased according to ocean temperature changes. So yes we solve for the basal drag as part of the solution.

No corrections necessary.

P3859L27:"...with annual accelerations of up to 20% above the winter background: : :": it is not clear what this sentence means. For example, acceleration is not the same units as winter background (velocity?). Do you mean the annually-averaged velocity is 20% greater than the winter velocity? Or peak summer velocity is 20% greater?

Corrected as suggested. *We mean annual mean velocities.*

P3860L5: again, do you mean acceleration peak, or velocity peak?

Corrected for velocity peak.

P3860L15: did you intend to 'hold back' Swiss Camp velocity data to use as validation (as opposed to tuning)? If so, perhaps state this clearly. Else, bundle the Swiss Camp comparison into earlier discussion of the K-transect data-based estimates of a,b and c.

Observations at these locations either are affected by the vicinity of the marine front or can only serve to give limits on the functional dependence.

Not adjusted.

P3860L15: Now, for Swiss Camp, you are discussing 'annual motion increases' and not 'accelerations'. Suggest using the same metric for all discussion.

Corrected as suggested.

P3860,L20: the term 'annual speedup' is unclear. Do you mean, increases to annually averaged velocities, relative to other years? Also, the 'of not more' is confusing. What happens for runoffs of greater than, e.g. 1m/yr (for the Swiss Camp discussion)?

Adjusted *according to previous comments on terminology.*

P3860L25: "Yet the approach: : :" this sentence is perhaps in the wrong spot? It appears to be an assessment of the runoff component of the SMB model. There is no mention of ice velocity changes here. Also, which is the unnamed model that is mentioned here?

Removed the sentence. *Repetition. The effect from differences between modelled and observed runoff on the lubrication parameterisation is already discussed earlier.*

P3861L5:"...to temperature variabliity diagnosed from five ocean basins in available AOGCMs for the decade 2000-2010." AOGCMs do not capture the 'absolute' timing of climate variability. So there is no reason to expect that AOGCM ocean variability is at all syncronized with observed ice sheet variability.

We fully agree with the reviewer on this point and the issue is amply discussed further below in the discussion Section of the manuscript.

No correction necessary.

P3861L13: how can regional climate models infer ice discharge variability?

Reformulated to clarify. *We meant that observations on volume change and SMB estimates from regional climate models can serve to discern the variability in ice discharge.*

P3861L20: "... support the choice: : :": in what way?

The exponential relation was adopted because this gave the closest agreement of increased discharge with the mentioned studies for a similar warming scenario.

No correction necessary.

P3861L21:"The selected relationship is calibrated such that the ice sheet model reproduces the relative contribution of the discharge increase to the total ice loss over the last decade in response to the considered climate models": As commented before, I am not convinced that this is a robust approach, given that AOGCMs aren't expected to actually simulate the phase of decadal-scale variability in an absolute sense (if they do, it is simply a coincidence). So, calibrating the ocean-discharge to the 2000-2010 period almost certainly introduces aliasing due to inaccurate sampling of the simulated climate record.

Why not calibrate rather to something much closer to observations, such as e.g. the World Ocean Atlas?

Here we want to refer to our detailed reply to the second comment by the same reviewer. We agree that AOGCMs are not expected to reproduce decadal scale temperature variability in an absolute sense (as mentioned in the manuscript). Therefore we decided to calibrate the relative contribution of ice discharge increase to total ice loss. In this way, we only assume that oceanic and atmospheric warming are related such that the ice-sheet model reproduces the observed partitioning of the ice loss. Generally, the climate models show a warming during the decade 2000-2010, but the simulated ensemble-mean ice loss is lower than was observed. This just reflects the incapability of the climate models to reproduce the magnitude and timing of the observed warming over Greenland. Yet our tuning allows the reproduction of a realistic partitioning of the ice loss, assuming that the relative warming in the ocean as compared to the atmosphere warming is reliable in the AOGCMs. Therefore we believe that our tuning holds.

Not adjusted *as a similar discussion is included in Sect. 4.*

Ocean reanalysis data would have certainly been preferable but multi-decadal time series were not available to us at the time. Though surface water temperatures are well known, we have doubts that reanalysis temperatures are well constrained at intermediate depth around Greenland. That's where we require reliable information on temperature variations to tune the parameterisation.

Equation 3: what determines the constant values in this equation?

Sensitivity study and table added. *For details see above reply to general comments.*

Equation 3: as previously mentioned in General Comments, given the importance of this equation to the final results, I think the authors need to do more work to assess uncertainty their results coming from uncertainty in this parameterization.

Sensitivity study and table added. *For details see above reply to general comments.*

P3862L5: "more regular than the amplification of the sliding coefficient: : :" this point statement is unclear to me.

Reformulated text. *We removed the unclear formulation and rephrased this passage as follows:*

'For a one-degree centennial warming, the sliding coefficient is increased by a factor 5.2 after 100 years. Yet, ice discharge does not even double. One reason is a geometric adjustment and thinning at the marine margins that limits the attainable ice export (Fürst et al., 2013). Another reason is that basal velocities do not necessarily scale linearly with changes of A_S *in a higher-order flow model.'*

P3862L11:"As initialisation"->"For initialization"

Corrected as suggested.

P3862L21: "Experiments have shown"... for this statement and others like it, without a reference I think the authors need to provide some form of (even just basic) quantitative description.

We refer here to our own experiments.

Reformulated the sentence to that effect.

P3863L2: LHS technique has also been used by several other ice-sheet-specific studies (e.g. Applegate et al 2010, Fyke et al 2014).

Added references.

P3863L3: DDF factors were previously stated to be definite values: : :"Melt rates are then determined: : : with degree-day factors : : : of : : : 0.0030 and 0.0079: : :" but here they are allowed to vary as LHS parameters. Perhaps the text could be made more consistent.

***Corrected** by adjusting the description of the SMB model such that it is clear that these values are determined during the model tuning in Sect. 2.3 (now 2.4).*

P3863L6: these +/- ranges seem arbitrary (e.g. 36-450% for m).

***No correction necessary.** As reviewer #1 makes a similar comment, we refer to our elaborate reply there and keep the discussion brief here. The chosen parameter ranges allow significant influence on the spin-up. Together with the fact that the best-fit combination does not hit any range limits, our selection receives justification.*

P3863L19: It would seem important to state the size of the LHS ensemble, to ensure that the parameter volume is sampled statistically sufficiently (rule of thumb 10 ensemble members/free parameter).

***Corrected by adding the number.** We used 100 samples, thus twice the amount according to the reviewer's rule of thumb.*

P3863L18: By what method are the criteria actually combined to determined the 'best' ensemble member?

The authors combined all criteria into a single value by a linear combination. Yet this value could not be used for the final decision. We want to illustrate the issues that arose on the basis of the ice volume criterion. As mentioned in the text, the simulated ice-sheet margin is typically too thick. In order to match the total ice-sheet volume, parameter combinations with thinner ice in the interior would be preferred. It is difficult to assess what is an acceptable offset/uncertainty in this criterion for such a spin-up without loosing in efficiency to discriminate sample members. This argumentation is readily transferable to some other criteria and inhibited to find the best fit from a single value assessment. We therefore opted for a more qualitative assessment of all sample members to select the best-fit.

Corrected by adding: *'One best-fit, reference parameter set and 7 additional combinations were selected on the basis of a qualitative assessment of respectively all or individual criteria (Table 1).'*

P3863L19: Table 1 shows 7 parameter sets, not one.

Corrected *by reformulating passage (see previous reply).*

P3863L25: To what extent does switching to ECMWF anomalies, then to AOGCMbased anomalies, introduce step functions in the SMB forcing?

The background fields of precipitation and temperature remain unaffected by switching to ECMWF anomalies in 1958 and later to AOGCM anomalies after 2005. As all anomalies are formed with respect to the period 1960-1990, the moments where anomalies are spliced fall into a 15-yr range. In atmospheric sciences, periods of more than 30 years are considered climatologically significant and thus changes on shorter time scales are considered inter-annual variability. Consequently, the discontinuity at 2005 is generally below the magnitude of the inter-annual variability in temperature and precipitation. Prior to 1958, the anomalies from ice core records are used which are also referenced to the period 1960-1990. By definition, this transition is also dominated by inter-annual variability.

Adjusted by adding sentence in Sect. 3.1: *'Discontinuities in these anomalies, when switching the forcing in 1958 and 2005, generally fall below the inter-annual variability.'*

P3864L5: what are the baseline ocean temperatures onto which anomalies are applied? What oceanic anomalies are used for the historical period?

As mentioned in the manuscript, the baseline ocean temperatures are the 1960-1990 average fields for the AOGCMs. For the historic, spin-up period, no ocean anomalies are applied. Ocean forcing gradually starts in the reference period.

No correction necessary.

P3864L12: "... and their capability: : :" what exactly is being assessed here in terms of AOGCM performance? And what measure is taken to assess whether the particular metric for model performance isn't being compensated for other climate model behaviour that could affect future simulations?

Our assessment is based on the temperature product of all AOGCMs. This involved a comparison of 1960-1990 averages to the ECMWF reference. In addition, models where rejected if the mean surface air-temperature increase over the projection period was a clear outlier as compared to the entire AOGCM ensemble.

Reformulated by specifying: *'The selection of climate models was based on the scenario coverage, the covered projection period and whether surface air temperatures, averaged for 1960-1990, generally agreed with the ECMWF product. Outliers in terms of average warming by 2100 and 2300 where identified from the AOGCM ensemble and hence rejected.'*

P3864L16: "...is used to avoid a bias by the mean states of the AOGCMs". But future projections could still be potentially affected by AOGCM biases. For example, if the climate model is too cold around the GrIS margins, then the warming to the point where a 0C surface (the maximum surface temperature of snow/ice surfaces) is obtained will be greater. When applied as an anomaly, this will artificially appear as greater warming (due to the initially cold state). Can the authors potentially assess in any way that AOGCM biases are not suspiciously correlated to the temperature changes they simulate in future projections?

We are aware of this possibility. However, we believe that our prior AOGCM selection could already reject badly performing models that were clearly deficient in reproducing Greenland surface temperature during our initial model selection. Since we don't know the details of the surface schemes employed by all AOGCMs we have to accept any remaining shortcoming in the selected CMIP5 suite and feel unable to make the assessment the reviewer point to.

No action taken.

P3864L21: Are the monthly SAT and P anomalies area-mean anomalies over the entire ice sheet, or spatial fields? It seems spatial fields are used, but it is not quite clear that this is the case, from the text.

These are spatial fields. Spatial patterns appearing in the anomaly fields are described in more detail in the next paragraph. Therefore we think, it is very clear that 2D anomalies are applied.

No correction necessary.

P3865L3: "... north south gradient: : :" perhaps note which direction this gradient goes (presumably, more warming farther polewards?)

Corrected *by mentioning direction of gradient in this paragraph.*

P3867L3: I think a plot of the difference between observed and simulated ice thicknesses would be very important for the reader to see.

The overestimation of margin thicknesses after spinning up an ice flow model for large-scale geometries is a well-known and well-quantified effect (refer to the 4 references given in the text). The next comment of the reviewer even refers to this as common ice-sheet model deficiencies. In conclusion, the authors consider it a redundant exercise to present and quantify these differences here.

No additional figure introduced.

P3867L6: It would seem to me that thicker margins would actually cause a faster velocities right near the margins (due to steeper surface slope to the margin). Perhaps instead, the lower margin ice velocities can be attributed to the relative lack of ice streams or other (common ice sheet model) deficiencies?

Without regard to the reasons of these higher ice margins, ice thicknesses are increased over some tens of kilometres upstream. The resultant relative changes in surface slopes are typically much higher than in the thickness field. From a *simple SIA point of view, this will lead to reduced velocities. In addition, consider a steady-state ice sheet that has to export a certain accumulation. If a certain flux has to pass an area of thicker ice, lower flow velocities are required (irrespective of whether this is controlled by sliding or deformation).*

No corrections required.

P3867L18: "On 5 km resolution, ice flow toward the margin is more channelised: : :": relative to what?

The comparison is with respect to earlier model versions on coarser grids.

***Corrected** by referring to model versions with coarser grid sizes.*

P3868L20: As noted previously: climate model simulations cannot be expected to capture the absolute phasing of climate variability. Thus, while SMB from ECMWF-based atmospheric forcing can be assessed compared to, e.g. 2005-2010 period, the '2005-2010' ocean forcing from climate models cannot be assumed to be on the same climate variability pathway as the real world. So, it is likely that HadGEM2-ES fortuitously simulated the ocean T change over this period correctly (was this why it was somewhat arbitrarily highlighted?). I note that the authors do seem aware of this general point, from the statement "not all AOGCMs are expected to correctly reproduce the real trend over such a short time period". I would strengthen this statement to something like: "no AOGCMs are expected to correctly reproduce the real trend over such a short time period, except by pure good luck." and ensure that this fact is represented throughout the manuscript and methodology.

We fully agree with the reviewer's comment. This issue has already been discussed in response to various comments from both reviewers (see above). Therefore we only specify our undertaken action for this passage.

***Corrected the statement as suggested to:** 'This reflects that AOGCMs are not expected to correctly reproduce the real trend over such a short time period.'*

P3869L3: An average increase, relative to what?

Relative to the 1990-2000 average.

***Added reference period as follows:** '[…]with respect to the average value in the 1990s.'*

P3869L3: Also, again, it is not clear that assessing ensemble performance over a 5 year period is useful, given that the ensemble average of a set of climate models cannot be expected to reflect the real phasing of climate variability.

Apart from the phasing of the climate variability between model fields and observations, which was discussed in our replies on several comments throughout the rebuttal, the reviewer is certainly aware of the fact, that our assessment, here, is limited by the short observational record on ice discharge. At least, our discussion indicates that the sign of the observed warming by 2010 is generally reproduced by the AOGCMs.

No correction possible *as limited by the length of the observational record.*

Figure 3: Mean annual surface air temperature anomaly is much less important than summer margin air temperature anomaly (which is the temperature subset that actually determines melting in a PDD model). Suggest plotting this instead, or in addition to, mean annual SAT anomaly.

We agree with the assessment of the reviewer on the subject of the importance of the summer temperature anomalies over the ice sheet margin for the simulated melting. Yet it is not evident to define the margin under such variable forcing and such long projections periods. In the most extreme scenarios, the melt area extents over most of the ice sheet. The two options are thus to prescribe a fixed mask or to trace the melt area through time to determine an average temperature increase. In both cases we introduce a bias in the average field. Therefore, we keep the original and objective method to average over the entire ice sheet domain. In addition, the annual average is kept, as we rather want to inform on the general warming over the ice sheet.

Not corrected.

Figure 3: does not show oceanic warming trends, but the text refers to ocean warming trends in Figure 3.

Corrected. *Indeed the wrong figure was referenced. We now refer to Fig.* 5.

Table 4: "mean atmospheric and oceanic warming": mean around GrIS? Global?

The caption for Table 4 indeed did not specify the character of the presented averages. We added this now.

Corrected. *Caption reads now as follows:*

'Ice sheet-wide mean atmospheric warming, basin-mean oceanic warming, and ensemble-average contribution of the Greenland ice sheet to global sea-level change by 2100 AD and 2300 AD. Sea-level changes are calculated with respect to the year 2000. Ensemble averages for each scenario use equal weights for individual AOGCMs. The root mean square (RMS) deviation from the mean ensemble realisation is added to estimate the variability.'

Table 4: is it correct to call the +/- values 'error estimates'? Or are they more accurately 'uncertainty ranges'?

The ranges given here are calculated as an RMS deviation from the ensemble mean, as was specified initially. We reformulated the caption to avoid any misunderstanding.

Corrected. *See previous action.*

P3869L14: It seems strange that if the IPCC AR4 SLR range is smaller, yet you say it additionally considers 'additional uncertainty arising from the SMB model'.

The AR4 sea-level projections comprise a very conservative estimate for future ice discharge evolution. Yet even the AR4 states, that when scaling dynamic ice loss *trends with global temperature change, the 2100 sea-level contribution from both ice sheets could be increase by up to 20 cm. This value is equal to the full range from the SMB contribution (both ice sheets) even under the highest warming. As future discharge changes are directly estimated in this study and as they explain about 40% of the entire sea-level signal from the Greenland ice sheet, it seems not surprising that AR4 found lower values relying on the conservative approach for ice dynamics. The variability in the SMB models therefore seems to be lower than the dynamic signal or this variability was simply underestimated by the time of the AR4.*

No correction required.

P3869L19: '...for the future discharge increase: : :" yet you say in the abstract that "enhanced discharge decreases over time: : :". Are these statements compatible?

This sentence could indeed be misunderstood and imply an incompatibility with the rest of the manuscript.

***Corrected by adapting the sentence** as follows: 'Yet the larger range is attributed to directly accounting for future changes in ice discharge.'*

P3869L23: "The new AR5 suffers from: : :": this statement is a little unclear – suggest rewriting.

***Reformulated** as follows: 'The new AR5 is however not able to quantify the importance of the interaction between ice dynamics and surface mass balance, as it suffers from the fact that the considered studies are not directly comparable either in terms of forcing or setup.'*

P3870L12: Can you explain why the high-emissions scenarios weren't continued to 2300?

The authors want to only briefly answer this comment as a more comprehensive reply is given to reviewer #1. For RCP6.0, the reason for this decision is that many AOGCMs were not continued beyond 2100. For RCP8.5, available AOGCM data was limited to a few models, showing a very large warming spread, making our ensemble approach questionable.

***Reformulated sentence as follows:** 'As AOGCM input was not available for RCP6.0 beyond 2100 and as the divergent response of the few AOGCMs under RCP8.5 is not considered compatible with our ensemble approach, projections were continued until 2300 AD only for the two lowest scenarios.'*

P3870L19: For the ISM runs forced with RCP26 and RCP45, is SMB typically negative at the end of 2300? This would give some indication as to whether a true 'stable' ice sheet configuration has actually been reached.

We are very grateful for this comment from the reviewer, as it adds another nuance to the brief stability discussion in Sect. 5.

For RCP2.6, all models show still a positive total SMB when averaged over the last 10 years of the experiment. This certainly is a first indicator for future stability. The picture changes for RCP4.5, where the SMB by 2300 falls below zero for most *of the AOGCMs. For the three AOGCMs with lowest warming, the ice sheet shows an SMB close to zero or slightly positive.*

Added information on the sign of the SMB by 2300.

P3871L2: "With forcing from MIROC-ESM-CHEM": perhaps note for completeness why this particular model was used.

Replaced by general sensitivity analysis covering all AOGCMs. *Particular AOGCM reference no longer mentioned as a general sensitivity study covering all climate models is introduced.*

Figure 9: This figure is nice, but really dense (and too small). Perhaps a clearer form of conveying this information? Also, it is not clear what the % values in each panel actually represent. Also, the 'overcompensation' due to negative discharge cumulative effects is not clear to me, at least via the graphics representation.

This is the central figure of our study and the authors worked hard on a representation that summarises the effect of ice discharge on the future ice volume evolution for each AOGCM and each scenario. Deliberately this figure is very dense. In this way, it might not be evident at first what the stated relative contributions mean in detail. Consequently, the authors reformulated the caption to better clarify and specify these subtleties.

Reformulated parts of the caption *to clarify the denoted relative values.*

P3873L11: doesn't Figure 10b show the relative thickening effect?

The references to Eq. (2) and Fig.1 were meant to guide the reader back to the respective parameterisation. The reviewer is right that the relative thickening is shown in Fig. 10b. This figure is already referenced a few sentences before.

Corrected. *Added reference to this figure.*

P3873L26: Again, I'm not confident that the model architecture is suitable for making statements of rates of ice mass change over the very short 2005-2010 period. To that extent, I would suggest that the good comparison to, e.g. Shepherd et al., 2012, is only fortuitous.

Reformulated passage according to the above discussion on the raised issue here.

Corrected as follows: *'When considering climate forcing from ECMWF reanalysis data and ocean temperatures from an AOGCM that shows an expressed warming over the period 2005-2010, we find an ice loss rate of 0.62 mm/yr over the same period that is explained by ~40% from increased ice-discharge, in agreement with the observational range.'*

P3874L25: This conclusion, that the largest source of uncertainty in ice sheet mass changes comes from SMB (i.e., climate), has been demonstrated by others previously: see, for example, Pollard et al 2000 (10.1016/S0921-8181(99)00071-5), Quiquiet et al 2012 (10.5194/tc-6-999-2012), Yoshimori et al 2012 (10.1175/2011JCLI4011.1), Fyke et al 2014 (10.1007/s00382-014-2050-7), and probably others as well. Suggest the authors reference and discuss at least some of these existing studies, with respect to this finding.

This conclusion certainly supports many studies on ice sheet volume projections on century and millennium time scales, including some of the early work of one of the authors (Huybrechts). Yet this conclusion is continuously challenged, especially when looking at (very) short observational records of changes in calving rates, and therefore merits to be stressed once more. Our study moreover adds a side-aspect in that uncertainties on volume evolution arising from future ice discharge changes are comparably smaller than the climate uncertainty. This conclusion finds even more back-up from the sensitivity study added during the revision of this manuscript. Nevertheless, we agree to acknowledge the more recent studies on this subject (Yoshimori & Abe-Ouchi, 2011, Quiquet et al., 2012, Fyke et al., 2014), not yet referenced in this manuscript.

Corrected *by including and referencing these studies in the discussion of our results in Sect. 5.*

P3875L6: Similarly, the finding that ice discharge at calving fronts is self-limited by ice dynamics (and the competition from SMB increases) has been shown previously, for example, even in some nice papers by the authors of the present manuscript (Goelzer et al., 2013), but also, Gillet-Chaulet (2012/2013, :10.5194/tc-6-1561-2012), Lipscomb et al., 2013 (10.1175/JCLI-D-12-00557.1), and perhaps others.

We agree that the concept of self-limitation by geometric changes is by itself not a new finding and has been suspected by many earlier on. The novelty is that even when scaling ice discharge with climatic variables, as in our study, ice discharge is still destined to decreases on the long term.

Added and discuss references *in Sect. 5 where this effect is assessed in detail.*

List of major changes.

P3-P4 Reformulation and shortening of the introduction.

P6-P8 Restructured model description, giving more details on SMB model, ice-dynamic component and data input.

P8-P9 Clarified idea of parameterisation and its applicability for lubrication effect.

P11-P12 Specified single parameter tuning, assumptions and limitation for ocean warming-induced discharge increase parameterisation.

P14 Specified LHS sampling and evaluation of criteria used for parameter tuning during the model spin-up.

P23-P24 Added detailed sensitivity discussion in Sect. 5. of the effect of changing the single value in the parameterisation of ocean warming-induced discharge increase.

P42 Added Table 5 for sensitivity to change in parameterisation of ocean warming-induced discharge increase.

P56 Expanded on details in caption of Fig.9. as this figure holds dense information.

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Ice-dynamic projections of the Greenland ice sheet in response to atmospheric and oceanic warming

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Abstract

Continuing global warming will have a strong impact on the Greenland ice sheet in the coming centuries. During the last decade (2000-2010), both increased surface melting and enhanced ice discharge from calving glaciers have contributed $0.6 \pm 0.1\,\mathrm{mm\,yr^{-1}}$ to global sea-level rise, with a relative contribution of 60 and 40% respectively. Here we use a higher-order ice flow model, spun up to present day, to simulate future ice volume changes driven by both atmospheric and oceanic temperature changes. For these projections, the flow model accounts for runoff-induced basal lubrication and ocean warming-induced discharge increase at the marine margins. For a suite of ten Atmosphere and Ocean General Circulation Models and four Representative Concentration Pathway scenarios, the projected sea-level rise between 2000 and 2100 lies in the range of +1.4 to +16.6 cm. For two low emission scenarios, the projections are conducted up to 2300. Ice loss rates are found to abate for the most favorable scenario where the warming already peaks in this century, which allows to preserve the ice sheet in a geometry close to the present-day state. For the other moderate scenario, loss rates remain at a constant level over three hundred years. The volume loss is predominantly caused by increased surface melting as the contribution from enhanced ice discharge decreases over time and is self-limited by thinning and retreat of the marine margin, reducing the ice–ocean contact area. As confirmed by other studies, the effect of enhanced basal lubrication on the volume evolution is found to be negligible on centennial time scales. The presented projections show that the observed rates of volume change over the last decades cannot simply be extrapolated over the 21st century on account of a different balance of processes causing ice loss over time. The results also indicate that the largest source of uncertainty arises from the surface mass balance and the underlying climate change projections, and not from ice dynamics.

1 Introduction

Volume changes of the Greenland ice sheet result from a balance between ice accumulation on its surface and ice loss around its margin by both melt-water runoff and ice discharge into the surrounding ocean. In the thirty year period prior to 1990, the ice-sheet geometry has been in a virtual balance with the prevailing climate but has since been losing mass at an increasing rate (Rignot et al., 2011; Zwally et al., 2011; Shepherd et al., 2012; Sasgen et al., 2012). Almost half of this recent mass loss is attributed to increased ice discharge at the marine margins (van den Broeke et al., 2009; Shepherd et al., 2012; Sasgen et al., 2012; Vaughan et al., 2013), with a tendency towards relatively more surface melting since 2005 (Csatho et al., 2014; Enderlin et al., 2014). During the period 1972 to 1995, glacier terminus positions and ice flow were rather stable around Greenland (Moon and Joughin, 2008; Howat and Eddy, 2011; Bevan et al., 2012). Over the last decade, however, ice sheet-wide surface velocity observations reveal complex spatial and temporal changes with accelerated glacier flow in the northwest, more variability in the southeast and relatively steady flow elsewhere (Moon et al., 2012; Carr et al., 2013; Moon et al., 2014).

A prominent example for recent dynamic changes of outlet glaciers in west Greenland is Jakobshavn Isbræ. Starting in 1998, its frontal zone sped up from about 6 to 12 $\mathrm{km\,yr^{-1}}$ within five years (Joughin et al., 2004, 2008c). One hypothesis links the acceleration to a successive loss of buttressing on the grounded ice as the floating ice tongue destabilised and collapsed while another hypothesis points to a speedup initiated by a weakening of the ice at the lateral glacier margins (van der Veen et al., 2011). In any case, the initiation of the glacier acceleration and retreat coincides with an intrusion of warm Atlantic Water into Disco Bay that likely entered the local fjord systems (Holland et al., 2008).

In southeast Greenland, speedup and retreat peaked in 2005 for Helheim and Kangerlussuaq Glaciers, which are both located at the end of ~ 80 km long fjords. Though the acceleration peak occurred simultaneously for both glaciers, the speed and retreat behaviour leading to this event differs (Stearns and Hamilton, 2007; Joughin et al., 2008b). While Helheim shows a continuous acceleration starting in 2002 with a cumulative retreat of the ice front of 8 km, Kangerlussuaq exhibited an abrupt retreat and acceleration between 2004 and 2005. Yet for both glaciers, the acceleration events were temporary and glacier speeds dropped again to the pre-speedup level (Bevan et al., 2012). In recent years, there is growing evidence that changes in fjord circulation and fjord stratification were the first-order control on this regional retreat and acceleration (Murray et al., 2010; Straneo et al., 2011; Straneo and Heimbach, 2013; Inall et al., 2014; Jackson et al., 2014; Sutherland et al., 2014). The southeast glacier acceleration was preceded by a period of low runoff from the ice sheet, which weakened the East Greenland Coastal Current and allowed warm waters from the Irminger Sea to reach the coast (Murray et al., 2010). Subsequently, the coastal current regained strength and provided again cold Arctic waters, which presumably has led to the regional re-stabilisation.

At the northern margin of the Greenland ice sheet, Petermann Glacier recently lost a major part of its 80 km long floating tongue. On 4 August 2010, about one fifth of the ice tongue broke off and drifted out of the fjord into Nares Strait (Falkner et al., 2011). In line with the above speedup examples, this breakup event was also preceded by ocean warming in the hundred meters above the 300 m deep sill at the southern end of Nares Strait (Münchow et al., 2011).

Warm and saline waters of tropical origin are in fact found at intermediate depth beyond the continental shelf break all around Greenland. There is evidence that these waters overcome the sills of individual fjord systems around Greenland (Straneo et al., 2012). Warming of deep fjord water can intensify submarine melt below an existing ice shelf or mélange cover (Motyka et al., 2011), or directly at the calving front (Rignot et al., 2010). The ice mélange is thought to play a role in the mechanical backstress it applies on the calving face. Thinning in the frontal zone, in turn, reduces the buttressing on the upstream glacier trunk and alters the local stress regime in favour of glacier acceleration (Nick et al., 2009). This provides a physical explanation of the simultaneous occurrence of recent glacier accelerations with warm waters reaching the respective shorelines.

Apart from the oceanic influence, the ice flow towards the margin is also affected by seasonal meltwater production at the surface that finds its way to the ice-sheet base (Schoof, 2010). Observations on both ice velocity and local runoff at various positions along the western flank of the Greenland ice sheet show distinct speed-up events during the melt season (Zwally et al., 2002; van de Wal et al., 2008; Bartholomew et al., 2011; Sundal et al., 2011). Though observations and simulations indicate that the effect might be small on annual time scales (Shannon et al., 2013; Tedstone et al., 2014), basal lubrication is hypothesised to enhance ice flow towards the marine margin and thereby influence ice discharge.

While ice discharge changes explain about 40 % of the recent ice loss on Greenland, the remainder is attributed to a decreasing surface mass balance (Sasgen et al., 2012). Most direct observations of the surface mass balance (SMB) components have local and at most regional character and are limited to the last decade (van den Broeke et al., 2011). Therefore, they are too short and not representative to directly infer ice-sheet wide trends. Yet SMB modelling has improved with the availability of validation data. Regional climate models are now capable of producing a physically-based, ice sheet-wide SMB estimate (Ettema et al., 2009; Fettweis et al., 2011; Vernon et al., 2013). SMB model results show that the five years with highest annual meltwater runoff since 1870 fall into the period after 1998 (Hanna et al., 2011). This concentrated occurrence of years with peak runoff exemplifies the general increase in runoff or decrease in SMB since the late 1990s (Ettema et al., 2009). In addition, the cumulated melt area has continuously increased, and melt extents since 2000 are on average twice as large as in the early 1980s (Fettweis et al., 2011).

For ice loss on Greenland over the next few centuries, a major contribution is expected from a decreasing SMB, or more precisely an increase in surface meltwater runoff (Cazenave et al., 2013). By now, the modelling community has managed to improve regional climate models (RCM) to the point that they reproduce past and present changes in various components of the SMB rather well (Vernon et al., 2013). Owing to a shortage in the observational coverage, the largest source of model uncertainty remains in the treatment and quantification of meltwater percolation and refreezing within the snowpack. Computational constraints typically limit RCM applications on ice sheet-wide scales to coarse grid resolution (often beyond 10 km). Yet it is within a narrow band of several tens of kilometres around the ice-sheet margin that the largest SMB changes are expected under atmospheric warming. Assuming small perturbations, RCM simulations often use a fixed ice-sheet geometry. Under strong future warming, margin thinning can attain a level, where SMB models need to account for these geometric changes. Instead of using a downsampling procedure for RCM SMB fields (Franco et al., 2012), SMB models often rely on temperature-index approaches for surface melting in high-resolution ice-flow models (Huybrechts, 2002; Robinson et al., 2011; Greve et al., 2011). Though such approaches rely on parameterisations of individual SMB components, ice volume projections can account for the feedback between geometric adjustments and SMB changes.

Here we include more ice-dynamical processes in a thermo-mechanically-coupled, three-dimensional ice flow model (Huybrechts and de Wolde, 1999) with the aim of better assessing the impact of ice dynamics on ice volume projections. These projections are driven by the 4 Representative Concentration Pathways (RCPs), specified by Moss et al. (2010) and commonly used for the IPCC's Fifth Assessment Report (AR5; IPCC, 2013). The ice dynamic model component includes parameterisations for ocean warming-induced discharge increase and runoff-induced basal lubrication (Sect. 2). To sample the range of climate sensitivities, a selection of ten Atmosphere and Ocean General Circulation Models (AOGCM) from the CMIP5 dataset (Taylor et al., 2012) is used. From this climatic input, both atmospheric and oceanic forcing is prepared as anomalies to drive the ice-sheet model (Sect. 3). The model evolution is evaluated for the recent past (Sect. 4) and the influence of ice discharge changes is assessed on the contribution of the Greenland ice sheet to future sea-level change (Sect. 5).

2 Model description and spin-up

2.1 The ice-sheet model

The three-dimensional thermo-mechanically coupled ice-sheet model comprises three main components that respectively describe the mass balance at the upper and lower ice-sheet boundaries, the ice dynamic behaviour and the isostatic adjustment of the Earth lithosphere (Huybrechts and de Wolde, 1999; Huybrechts, 2002; Fürst et al., 2011)

2.1.1 Ice-sheet dynamics

The simulated ice flow arises as a viscous response of the material to gravitational forcing. Using a higher-order approximation to the force balance, the model accounts for effects from horizontal gradients in membrane stresses (Fürst et al., 2011). More specifically, the model adopts a multilayer longitudinal stresses approximation of the force balance, abbreviated as LMLa in Hindmarsh (2004). This ice-dynamic core allows for a more realistic inland transmission of perturbations at the ice-sheet margin (Fürst et al., 2013). The model is run on a 5 km equidistant grid in the horizontal plane and uses 30 non-equidistant layers in the vertical. The vertical grid spacing is refined towards the bottom where vertical shearing is concentrated. The flow component of the ice-sheet model also accounts for the direct effect of ocean warming on ice discharge and for runoff-induced lubrication. Both effects are parameterised and presented in following sections.

2.1.2 Surface mass balance

The SMB model comprises snow accumulation, meltwater runoff and meltwater retention in the snowpack. The background field for surface accumulation is based on the Bales et al. (2009) accumulation map for the period 1950–2000. For the ablation component, the melt and runoff model relies on the widely used positive degree-day runoff/retention approach (Janssens and Huybrechts, 2000; Gregory and Huybrechts, 2006). This approach first determines the positive degree-day sum from monthly air temperature input assuming a statistical variability of daily near-surface temperatures around the monthly mean (with a standard deviation of 4.2 °C). Melt rates are then determined with different degree-day factors for snow and ice. Their values are determined by tuning the model spin-up (Sect. 2.4 and Table 1). Initial surface melt is first stored as capillary water until the snowpack becomes saturated and runoff occurs. In the snowpack model, formation of superimposed ice occurs when water-saturated snow survives above the impermeable ice layer until the end of the season, and subsequently refreezes. The SMB model relies on a parameterisation of the surface temperature calibrated for the period 1960–1990. The model is forced by monthly surface air temperature and annual precipitation anomalies relative to the 1960–1990 mean. For the period 1958–2010, the positive degree-day runoff/retention approach has been compared to RACMO2.1/GR, a physical snow model coupled to a high-resolution model for atmosphere dynamics (Hanna et al., 2011). Both approaches for SMB agree well in terms of interannual variability (R^2 coefficients of determination of 0.79 for SMB, 0.84 for precipitation, and 0.75 for runoff).

2.1.3 Input data

Geometric input has been updated from the Bamber et al. (2013) data set with slight adjustments for our specific model requirements (Goelzer et al., 2013). A geoid correction is applied to reference the data set to mean sea level, which is subsequently re-projected and interpolated from the original 1 km grid to the ice-sheet model grid. The geothermal heat flux is inferred from seismic data (Shapiro and Ritzwoller, 2004). The values were adjusted with Gaussian functions at the deep ice core sites (NEEM, GRIP, NGRIP, Dye3 and Camp Century) to reproduce observed basal temperatures, assuming a radius of 100 km to gradually blend in the difference with the background field (Pattyn, 2010).

2.2 Effect of surface runoff on basal lubrication

Observations of ice velocities show seasonal speed-up in the summer melt-period (Zwally et al., 2002; Bartholomew et al., 2011; Sundal et al., 2011). Surface runoff generally finds a way into the ice body through moulins and the water is assumed to reach the bed near the ice-sheet margin. The rate of basal melt-water discharge determines the two-fold character of the subglacial drainage system, which in turn controls lubrication and its effect on the sliding velocity (Schoof, 2010). Observational studies often report on successive distinct speed-up events during the melt season (Zwally et al., 2002; Bartholomew et al., 2011; Andrews et al., 2014; Tedstone et al., 2014). For our model application, however, the interest is on their integrated effect over one year. Sundal et al. (2011) find that mean summer speed-up is positively correlated with daily runoff, as long as runoff rates do not exceed a certain threshold. Above this threshold, average speedup is somewhat reduced as exemplified in the two-fold character of basal drainage. The annual runoff will strongly depend on the amount of days for which this threshold is exceeded. Therefore, we assume a relation between the annual surface runoff and the annual increase in sliding relative to the winter reference. In this way, the speedup parameterisation will not distinguish between years of comparable annual runoff, caused primarily by moderate but constant melting during the entire summer season or by individual high melt peaks. In the ice flow model, the Weertman sliding relation is therefore extended with a multiplier S_{BL} that depends on the annual rate of basal meltwater discharge.

$$v_{\mathrm{b}} = S_{\mathrm{BL}} \frac{A_{\mathrm{S}}}{H} \tau_{\mathrm{b}}^{3} \tag{1}$$

Here sliding velocities are denoted with v_{b}, basal drag with τ_{b}, the sliding factor with A_{S} and the ice thickness with H. In this parameterisation, the basal meltwater discharge rate is assumed equal to the local surface runoff R, whilst neglecting contributions from basal melting or meltwater routing beneath the ice sheet. Theoretical work on subglacial drainage systems indicates a speedup peak for a specific rate of basal water discharge

(Schoof, 2010). Above this discharge rate, a channelised basal drainage system develops, which is associated with lower relative speed-up values. In the absence of local runoff, no lubrication effect is simulated (S_{BL}=1). Informed by the best-fit parameteriation in Shannon et al. (2013), a Poisson-like functional dependence (Fig. 1) between relative speedup and runoff is chosen.

$$S_{\mathrm{BL}} = 1 + cR^a \cdot \exp\left(-bR\right) \tag{2}$$

In this notation, the unknown parameters a, b and c are assumed positive. Within a comprehensive uncertainty study on the lubrication effect, Shannon et al. (2013) find that its effect is of secondary importance in terms of the centennial ice volume evolution. Therefore, only one set of parameters is used for the projections here.

The three unknown parameters are determined using observational data on annual velocity increase and runoff at two locations along the western flank of the Greenland ice sheet (Fig. 1). The first location is east of Kangerlussuaq and upstream of Russell glacier, often referred to as the K-transect (van de Wal et al., 2008; Shepherd et al., 2009; Bartholomew et al., 2010, 2011). Here a consistent picture emerges with annual mean velocities of up to 20 % above the winter background for runoff rates below 3.5 m ice equivalent yr^{-1}. For the Russell glacier transect, Bartholomew et al. (2011) find the highest velocities for observed runoff rates above 3 $\mathrm{m\,yr}^{-1}$. In the larger vicinity of the K-transect, Sundal et al. (2011) link the speedup of several glaciers to runoff extracted from a monthly degree-day surface meltwater runoff/retention model. Their findings indicate a velocity peak for an annual runoff below 1 m. This difference between observed and modelled critical runoff rates is considered in our functional dependence. For our simulated ice-sheet geometry, our mass balance model gives annual runoff rates of up to 4 $\mathrm{m\,yr}^{-1}$ near the K-transect. Due to a faster inland decrease in modelled runoff, as compared to observations, upstream speedup would be underestimated. Taking this into account, the following parameter values are chosen: $a = 1.8$, $b = 0.9\,\mathrm{yr\,m}^{-1}$ and $c = 0.43$. For these parameters, the maximal annual velocity lies 25 % above the winter reference for an annual runoff rate of 2 $\mathrm{m\,yr}^{-1}$ (Fig. 1). The magnitude of the runoff rate causing maximal speedup agrees with theoretical estimates using a plastic ice-sheet geometry (Schoof, 2010).

Observations near Swiss Camp upstream of Jakobshavn Isbræ serve as extra validation for the chosen functional dependence (Zwally et al., 2002; Joughin et al., 2008a; Colgan et al., 2011). Near Swiss Camp, observed annual flow increases by 2 % for an annual runoff of not more than 1 $m\,yr^{-1}$. Further down the glacier and considering other outlet glaciers in the vicinity of Jakobshavn Isbræ (Joughin et al., 2008a; Colgan et al., 2011), a different picture emerges with 10 % annual velocity increase for runoff rates of about 1 $m\,yr^{-1}$. At these locations however, the velocity variations are also influenced by seasonal changes at the marine termini. These observational estimates are covered by the suggested functional dependence. The presented parameterisation might be affected by the observational bias towards the western flank of the Greenland ice sheet. Yet the approach accounts for differences between observed and simulated runoff values.

2.3 Effect of ocean warming on ice discharge

With the aim to parameterise ocean-induced changes in ice discharge, outlet glacier accelerations are linked to oceanic warming assuming a uniform functional dependence. This choice ignores the details of the many processes that may affect the dynamics of calving glaciers and thus the ice discharge. Their representation is limited by the large-scale character of the envisaged simulation, not resolving geometric details. We therefore assume that ocean temperature changes have a first-order control on the discharge response, being aware that the individual response highly depends on the local fjord and glacier geometries (e.g. Moon et al., 2014). Despite this non-uniform behaviour from glacier to glacier, the pattern of recent glacier accelerations is, to a certain degree, consistent with the variability in offshore ocean temperatures around Greenland (Straneo et al., 2012; Jackson et al., 2014). The functional dependence is derived by relating velocity observations (Rignot and Kanagaratnam, 2006; Moon et al., 2012) to temperature variability diagnosed from five ocean basins in available AOGCMs for the decade 2000–2010. Observations during this decade show an average speedup of outlet glaciers in the southeast of 34 % and in the northwest of 28 %, while other regions show no significant trend (Moon et al., 2012). Scaling these accelerations to the entire ice sheet and weighting them with the regional discharge distribution (Rignot and Kanagaratnam, 2006) results in an average ice discharge increase of about 10 to 15 %. This increase shows an almost linear trend over the last decade (Rignot et al., 2011). Using the residual between observed volume changes and SMB estimates from RCMs as indicator for ice discharge changes (Sasgen et al., 2012), the decadal discharge increase explains between 25 and 40 % of the total mass loss (Shepherd et al., 2012). Considering the oceanic temperature forcing at hand together with the fast marginal adjustment properties of the ice-sheet model (Fürst et al., 2013), a linear increase in discharge is best simulated by a non-linear relation between ocean temperatures and sliding velocities. In addition, results from a generalisation of the flow-line response of individual outlet glaciers to a large-scale Greenland ice-sheet application (Nick et al., 2013; Goelzer et al., 2013) support the choice for a exponential dependence. The selected relationship is calibrated such that the ice-sheet model reproduces the relative contribution of the discharge increase to the total ice loss over the last decade in response to the considered climate models.

$$A_{\mathsf{S}}^{\mathsf{outlet}} = A_{\mathsf{S}} \cdot \alpha^{(\Delta T_{\mathsf{ocean}}/1\,^{\circ}\mathsf{C})} \quad (3)$$

For the tuning goal described above, we find $\alpha = 5.2$. The sensitivity of the projections to changes in parameter α is described in Sect. 5. The amplification of the sliding factor $A_{\mathsf{S}}^{\mathsf{outlet}}$ applies exclusively to marine-terminated glaciers using the temperature anomaly $\Delta T_{\mathsf{ocean}}$ in the adjacent ocean basins. In this way, we circumvent to directly quantify how efficiently offshore waters enter the fjords to facilitate melt at the glacier fronts. Consequently, the parameterisation is assumed to be valid for long-term gradual ocean warming and is not applicable for short-term warming events. In addition, any delays in the ocean system are intrinsically neglected. The forcing is applied up to 20 km inland from the calving front for ice grounded below sea-level to account for a far-reaching loss in back-stress on a length scale appropriate to longitudinal stress coupling (Nick et al., 2012; Fürst et al., 2013).

Prescribing an experiment with a linear increase in ocean temperatures under constant atmospheric forcing, the ice-sheet model shows an increase in ice discharge (Fig. 2). For a one-degree centennial warming, the sliding coefficient is increased by a factor 5.2 after one hundred years. Yet, ice discharge does not even double. One reason is a geometric adjustment and thinning at the marine margins that limits the attainable ice export (Fürst et al., 2013). Another reason is that basal velocities do not necessarily scale linearly with changes of A_S in a higher-order flow model. After one hundred years, ocean temperatures are kept constant and ice discharge remains at an increased level. Yet the ongoing geometric adjustment causes a general decrease of the ice discharge in this latter period.

2.4 Glacial cycle spin-up

In order to initialise to the present day, the model is spun up over a full glacial cycle as described in Huybrechts (2002). The ice sheet geometry evolves freely in response to past changes in regional surface temperatures, precipitation and sea level.Although the general approach is unchanged from earlier applications of this model (Huybrechts, 2002), the underlying reconstruction for past temperature changes is updated with recent proxy information from several ice cores (for details see Appendix A). A new compilation of accumulation observations over the Greenland ice sheet (Bales et al., 2009) is used as basis for scaling past precipitation changes with the mean annual temperature change (by 5 % $^{\circ}C^{-1}$). Finally, a new parameterisation to improve the retreat history from the Last Glacial Maximum is applied (Simpson et al., 2009), which is constrained by proxies for relative sea level. Switching at 3 kyr BP from a shallow ice approximation to the higher-order formulation appeared to be sufficiently early to resolve the main effects of including horizontal stress gradients by the present day.

Using an unconstrained model evolution during the spin-up phase guarantees a self-consistent model state in the present day but the geometry will deviate from the observed state. Therefore, key model parameters are tuned to minimise geometric and dynamic differences after the spin-up. For a statistically sufficient and efficient coverage of the parameter space, a Latin hypercube sampling (LHS) was chosen (McKay et al., 1979), relying on hundred combinations. This sampling technique was previously used for assessing the parameter sensitivity when spinning up ice-sheet models (Stone et al., 2010; Applegate et al., 2012; Fyke et al., 2014). We vary the positive degree-day factors for both ice (DDF_{ice}) and snow (DDF_{snow}) together with an enhancement factor (m) to the rate factor and the sliding coefficient (A_S). These four parameters control both the SMB and the dynamic state of the modelled ice sheet. Parameters are selected in ranges of 75–125 % for the degree-day factors, 36–450 % for the enhancement factor m and 50–200 % for A_S with respect to a previous calibration. Parameter ranges were estimated from the respective sensitivity of the model, known from previous tuning.

Eight criteria were chosen to quantify differences between the modelled ice sheet and the observed present-day state. The minimisation reduces the mismatch of the following quantities: total ice volume; ice-covered area; ice area above 3000 m and below 1500 m surface elevation; southwest position of the land-terminated ice margin; global ice thickness and surface elevation. Instead of exclusively focussing on geometric tuning diagnostics, as in Stone et al. (2010), a final criterion evaluates the dynamic state of the ice sheet. Ice discharge in the decades prior to 1990 is assumed to have compensated for ~ 60 % of the average accumulation (Ettema et al., 2009). This additional criterion considerably reduces the parameter space. One best-fit, reference parameter set and 7 additional combinations were selected on the basis of a qualitative assessment of respectively all or individual criteria (Table 1). Very similar positive degree-day factors were found as compared to a previous tuning while parameters controlling the ice flow magnitude are slightly reduced. This reduction is necessary because of higher velocities in the ablation zone when using the parameterisation for runoff-induced speedup.

3 Climatic forcing

3.1 Reference period

For the period 1958 to 2005, the SMB model is forced with monthly temperature anomalies and annual precipitation ratios from a combination of ECMWF ERA-meteorological reanalysis and ECMWF operational analysis data as described in Hanna et al. (2011). Anomalies and ratios are calculated with respect to the period 1960–1990. This assumes that the ice sheet was in quasi-equilibrium with the prevailing climate of that time, as in previous studies (e.g. Hanna et al., 2005). The reference precipitation is from Bales et al. (2009). In the same way, the oceanic temperature anomalies are calculated from the Atmosphere and Ocean General Circulation Models (AOGCMs). Discontinuities in these anomalies, when switching the forcing in 1958 and 2005, generally fall below the inter-annual variability.

3.2 Future scenarios

For future ice-sheet simulations, climate projection data from ten AOGCMs were selected from the WCRP's CMIP5 multi-model dataset prepared for the IPCC AR5 (Taylor et al., 2012). The selection of climate models was based on the scenario coverage, the covered projection period and whether surface air temperatures, averaged for 1960-1990, generally agreed with the ECMWF product. Outliers in terms of average warming by 2100 and 2300 were identified from the AOGCM ensemble and hence rejected. Table B1 gives a complete overview of the considered AOGCMs). For these projections, the AOGCMs were forced with four CMIP5 Representative Concentration Pathway (RCP) scenarios (Moss et al., 2010). The same anomaly approach as for the reference period is used to avoid any potential bias associated with the mean states. Monthly surface air temperature anomalies, annual precipitation ratios and annual ocean temperature anomalies are therefore considered with respect to the same 1960–1990 reference period.

3.2.1 Atmospheric forcing

Monthly surface air temperature anomalies and annual precipitation ratios are derived for each individual AOGCM over the ice-sheet model domain. These future atmospheric anomalies drive the SMB model starting from the year 2005. In most cases, the data covers the period up to 2100 or 2300 AD. Missing data in the last year of two AOGCMs were filled by repeating the previous year.

The annual air temperature anomaly averaged over the present ice-sheet extent (Fig. 3) is instructive as a general trend but conceals the 2-D pattern of the warming (not shown). In general, the spatial pattern of the temperature forcing shows an expressed north-south gradient of up to 10 °C by 2100 AD, with stronger warming in the north. This latitudinal gradient depends on the climate sensitivity and the polar amplification of each AOGCM. For a given latitude, the difference in warming between the east and west of the ice sheet depends strongly on the individual AOGCM. The patterns of future precipitation changes are also AOGCM dependent and cannot be generalised. Yet the average precipitation increases and scales with the scenario intensity. By 2100, the ensemble averages per RCP show 13, 19, 23 and 37 % additional precipitation for RCP2.6, RCP4.5, RCP6.0 and RCP8.5, respectively. For RCP2.6 and RCP4.5, these values increase to respectively 19 and 31 % by 2300.

3.2.2 Ocean forcing

Oceanic forcing is decomposed into time series for five different oceanic basins. Their delineation is based on the circulation pattern of Atlantic Water (AW) around Greenland (Straneo et al., 2012, and references therein), cf. Fig. 4. The North Atlantic Current brings warm and saline water from the Atlantic Ocean and splits into the Irminger current and the Norwegian Atlantic Current. The latter enters the Nordic Seas where sinking occurs but AW partly submerges under fresh Polar Waters and continues northwards to Fram Strait. There, one portion enters the Arctic Ocean ultimately reaching the north Greenland continental shelf break (northern region). The other portion turns back at Fram Strait along the eastern flank of Greenland at intermediate depth (northeastern region). South of Denmark Strait, it joins warmer AW provided by the Irminger Current and continues southwards along the shelf break (southeastern region). At the southern tip of Greenland, it feeds into the Labrador Sea where further sinking occurs (southwestern region). A fraction of these waters remain at intermediate depth flowing northward and potentially overcome the sill into Baffin Bay (northwestern region). Warm AW with subtropical origin is therefore found at intermediate depth all around Greenland. For our projections, ocean temperature changes in these basins are related to ice discharge changes at the marine-terminated margin of the Greenland ice sheet.

Ocean circulation in the deeper ocean around Greenland, off the continental shelf, is resolved in most AOGCMs. Ocean basins are latitudinally delineated by the 60° N, 70° N, 80° N parallels and the North Pole at 90° N, and confined by the Greenland coastline (Fig. 4). In each individual basin, AOGCM grid box centres that lie within a 300 km radius from the Greenland coastline are considered. This belt covers the continental shelf and a part of the deep ocean beyond the shelf break. The resulting basin temperature anomalies are not very sensitive to a radius increase to 500 km. In the vertical, temperatures are averaged over a depth of 200 to 600 m. The upper limit is inspired by the average freshwater layer thickness in Greenlandic fjords (Straneo et al., 2010, 2011, 2012) together with intermediate depth locations of offshore AW (Holland et al., 2008). The latter argument combined with the fact that Greenlandic fjords have typical sill depths of several hundred metres gives rise to the lower bound. Area- and depth-averaging of all AOGCM grid points in each basin provides five temperature time series for each AOGCM and each RCP.

Ocean temperature anomalies for each basin are considered with respect to the 1960–1990 average (Fig. 5). For each basin, the annual temperature anomaly records are filtered with a 5-yr moving average. This is necessary to prevent high frequency oscillations when forcing the ice-dynamic model. Though there is a tendency for stronger warming in the northern ocean basins in many of the AOGCMs, differing trends within the five basins are highly dependent on the individual climate model.

4 Ice sheet evolution in the recent past

After the glacial-cycle spin-up, the present-day ice-sheet geometry is in a self-consistent state concerning ice geometry, dynamics, temperature and SMB. The geometry and temperature naturally carry the long-term memory of the ice-sheet evolution. The main shortcoming from such a spin-up is that for the present day the modelled geometry does not exactly match observations. Like in other studies with a similar spin-up technique, ice thicknesses near the margin tend to be overestimated and therefore the ice extent is somewhat larger (e.g. Huybrechts, 2002; Robinson et al., 2011; Greve et al., 2011; Graversen et al., 2011). Though the geometric mismatch biases the SMB near the margin, the ice sheet-wide SMB compares well with other approaches (see below). Thicker margins also affect the modelled ice flow, with tendency to underestimated margin ice velocities (Fig. 6). A side-by-side comparison shows that the locations and the magnitudes of channelised ice flow towards the marine margin are well reproduced on the 5 km grid. In this spin-up technique, regions of fast flow naturally arise from the interplay between deformation, sliding and thermo-dynamics.

Since velocities generally drop below 100 $\mathrm{m\,yr^{-1}}$ within some tens of kilometres from the ice-sheet margin, regions further upstream are not expected to directly contribute to ice discharge within one century. More meaningful than matching velocities at the margin is therefore that the model is capable of reproducing ice discharge rates and their regional distribution around Greenland (Table 2). The simulated present-day state shows a total ice discharge that slightly exceeds otherwise inferred values (Rignot and Kanagaratnam, 2006). The 5% overestimation mostly arises from simulated ice-ocean contact in regions where no ice-sheet cover is observed, i.e. in the north and the east. A 20 km model spin-up is only capable of reproducing the large-scale regional distribution and the total ice discharge. Compared to this coarser model version, ice flow towards the margin is more channelised for the presented 5-km grid and the regional agreement between modelled and inferred discharge improves, on a regional level and down to the level of major outlet glaciers. The match on a drainage basin level arises naturally without specific model tuning. In this regard, the glacial-cycle spin-up method is preferable to another initialisation technique that aims at inverting for observed ice velocities using the observed geometry (Gillet-Chaulet et al., 2012). Though it reproduces observed velocities, this latter initialisation technique is confronted with a strong initial model drift. Therefore, we believe that the free-geometry spin-up, using a model with increased dynamic complexity on high resolution, provides a useful initial state for projecting the future dynamic response of the Greenland ice sheet on centennial time scales.

Averaged over the 1960–1990 period, the positive-degree-day runoff/retention approach gives a total SMB of 373 $\mathrm{Gt\,yr^{-1}}$, when forced with ECMWF ERA-reanalyses data. Other physically-based models show a spread between 341 to 479 $\mathrm{Gt\,yr^{-1}}$ in the same period (Vernon et al., 2013). Somewhat at the lower end, the difference in our model might arise from the underlying reference precipitation map (Bales et al., 2009; Hanna et al., 2011). Moreover, recent changes in the total SMB agree fairly well between inferred values and the used positive-degree-day approach (Table 3). SMB changes estimated from observations and given by various other model approaches (Sasgen et al., 2012; Vernon et al., 2013) can be compared on the basis of six main drainage basins (Hardy et al., 2000). On this drainage basin level, differences between various methods become more expressed. For one drainage basin (in southeast Greenland; C in Table 3), discharge-corrected observations from GRACE cannot be reconciled with any model estimate. This indicates some large remaining uncertainties in both modelled SMB changes and otherwise inferred estimates. However, in most cases our SMB model reproduces the trends of other models within stated uncertainty bounds.

When forced with ECMWF atmospheric reanalysis data and using ocean temperatures from one climate model with expressed warming over that period (i.e. HadGEM2-ES in Table B1), the simulated ice sheet loses mass at a rate of 0.62 $\mathrm{mm\,yr^{-1}}$ for the period 2005–2010. This is in good agreement with the inferred average trend of 0.7 ± 0.1 $\mathrm{mm\,yr^{-1}}$ (Shepherd et al., 2012). A 41 % share (or 0.25 $\mathrm{mm\,yr^{-1}}$) of the mass loss arises from increased ice discharge. For the full ensemble of climate models, the average mass loss rate for the period 2005–2010 is lower at 0.32 $\mathrm{mm\,yr^{-1}}$. This reflects that AOGCMs are not expected to correctly reproduce the real trend over such a short time period. For the ensemble member with the highest initial oceanic and atmospheric warming, the sea-level contribution reaches a maximum rate of 0.71 $\mathrm{mm\,yr^{-1}}$ for the period 2005–2010. This suggests that the Greenland ice sheet is for now responding to the upper end of temperature changes provided by the CMIP5 climate model ensemble.

Over all climate models and scenarios, this approach gives an average increase in ice discharge of about 0.14 $\mathrm{mm\,yr^{-1}}$ with a maximum of 0.23 $\mathrm{mm\,yr^{-1}}$ for the period 2005 to 2010 with respect to the average value in the 1990s. The average increase in discharge caused by the climate model ensemble produces the inferred $\sim 40\,\%$ share of the total mass loss. However, the mean is at the lower end of observations during this period and results from a weak oceanic warming around Greenland over the last decade in the used climate models (Fig. 5).

5 Future projections

Figure 7 and Table 4 summarise the volume projections of the Greenland ice sheet for all models and all scenarios under investigation. A breakdown by individual climate models is presented in Appendix B. By 2100 AD, the full model and scenario range of Greenland sea-level contributions is between 1.4 and 16.6 cm (Fig. 7 and Table B1). This range is slightly higher than the 1–12 cm found for the IPCC AR4 (Meehl et al., 2007), which included the additional uncertainty arising from the SMB model. The higher maximum in sea-level projections is somewhat unexpected because the RCP scenarios have a reduced upper bound for radiative forcing by 2100, when compared to the previously used scenarios. Yet the larger range is attributed to directly accounting for future changes in ice discharge. In terms of the SMB contribution to future ice loss, the IPCC AR5 (Cazenave et al., 2013) gives a range of 1–11 cm, confirming the results of the previous AR4. Yet the AR5 is the first to attempt to quantify the contribution from future changes in ice discharge. It states an additional contribution from dynamic changes of 1–9 cm for all RCP scenarios. The new AR5 is however not able to quantify the importance of the interaction between ice dynamics and surface mass balance, as it suffers from the fact that the considered studies are not directly comparable either in terms of forcing or setup.

Until 2050 AD, there is hardly any difference in the mean sea-level contribution between the four scenarios. This is in agreement with a similar behaviour for the underlying atmospheric and oceanic forcing (Sect. 3.2.1). The ensemble spread in sea-level evolution for each scenario arises from the different climate trajectories followed by the individual AOGCMs. This spread is largely overlapping during the first century for three scenarios. The exception is RCP8.5, a high-impact scenario assuming a high-emission fossil-fuel orientated world. This scenario causes a mean centennial sea-level contribution of 10.2 cm, which is about twice as large as for other RCPs. The reason is an average warming of $\sim 7\,^{\circ}$C over Greenland that is also more than twice as high as for other RCPs. In addition, RCP8.5 is the only scenario for which mass loss rates significantly increase throughout the next century.

As AOGCM input was not available for RCP6.0 beyond 2100 and as the divergent response of the few AOGCMs under RCP8.5 is not considered compatible with our ensemble approach, projections were continued until 2300 AD only for the two lowest scenarios. Both assume a stringent climate policy with focus either on terrestrial carbon for mitigation (RCP4.5) or on negative emissions (RCP2.6). Both scenarios aim for a climate stabilisation but only RCP2.6 has a peak greenhouse gas concentration before 2100 AD and declines afterwards (Moss et al., 2010). For both scenarios, the Greenland contribution to global sea-level rise increases continuously but for RCP2.6 the rate of increase gradually levels off. In this case, the SMB remains positive in the last decade of the projection. Therefore, it appears that a new ice-sheet equilibrium with limited ice loss (<20 cm of sea-level rise) is attainable. For RCP4.5, the rate of mass loss is almost constant over three hundred years with a total volume loss equivalent to 20.1 cm sea-level increase. Average SMB values during the last decade are negative for most ensemble members. A typical thinning pattern for RCP4.5 shows extensive marginal thinning and inland retreat of calving fronts after three hundred years (Fig. 8). Mass loss near the margin is partially balanced by increased snow accumulation and thickening in the interior.

In all climate scenarios, oceanic warming causes additional mass loss from the ice sheet by 2100 AD (upper dark blue columns in Fig. 9). This comprises both the directly induced changes in ice discharge but also their effect on the SMB via geometric adjustments. For individual AOGCM projections, the inclusion of oceanic forcing can explain more than 50 % of the total contribution to sea-level rise by a given time period with an average increase of the total mass loss by $\sim$ 40 %. In absolute terms, the ocean-induced contribution to sea-level change ranges from 1.8 to 2.6 cm (scenario averages) and 1.1 to 3.2 (full spread) after one century, and from 3.8 to 5.4 cm after three centuries (full spread is 2.3 to 7.4 cm). The oceanic influence on the total ice loss becomes relatively less important for more intense atmospheric warming. It explains about half of the mass loss for RCP2.6 while the share is reduced to 27 % for RCP8.5. This indicates that decreasing SMB and increasing discharge are mutually competitive processes for ice removal at the marine margin. In addition, ice further upstream is efficiently removed by ablation before it actually reaches the marine margin for calving. The oceanic forcing typically induces a diffusive thinning wave at the marine margin which is gradually transmitted inland (Fig. 10a). In areas with a marine margin, this additional thinning wave explains a large share of the total thinning including surface melting under atmospheric warming (Figs. 8 and 10a).

In Fig. 9, we also attribute simulated mass changes to either changes in ice discharge, arising from oceanic forcing and inland ice dynamics, or from changes of the mass balance at the ice sheet surface or base (although in all cases, basal melting contributes less than 3 % of the total land ice loss). While increased discharge explains about 40 % of the average mass loss between 2000 and 2010 (light blue columns), its relative contribution generally decreases afterwards and changes in SMB become the dominant factor for mass loss. This is because total ice export across calving fronts eventually falls below year 2000 levels, despite warmer ocean temperatures. Limitations on the ice discharge increase are a direct result of gradual thinning at the marine margins with a fast adjustment of the ice inflow from upstream (Fürst et al., 2013) but are also a consequence of a retreat of the ice-sheet margin back on land. For the CanESM2 model under RCP4.5, the ice sheet loses more than half of its contact area with the ocean by 2300 (Fig. 8). In general, ice discharge increase is more relevant for the total mass loss in scenarios with higher mitigation efforts (RCP2.6, RCP4.5). The reason is that an ice discharge increase also causes dynamic thinning inland and thereby intensifies surface melting. Surface melting in turn competes with the discharge increase by removing ice before it reaches the marine margin. Margin thinning and retreat limit the ice discharge and increase the relative importance of surface melting in the future volume evolution. The total 2100 ice loss, from SMB changes only, increases by more than 70 % when including ice–ocean interaction. This share is about 42 % of the combined total ice loss in 2100 (Fig. 9b) but only 10 % of it are directly caused by ice discharge increase at the marine margin. By 2300 AD, the cumulative effect from ice discharge changes are even negative as ice discharge rates have on average fallen below the pre-2000 level between 2000 and 2300.

Detailed flow-line projections of the ice discharge evolution of four major outlet glaciers on Greenland show a general increase by 2100 and 2200 AD (Nick et al., 2013). Such a widespread increase of ice discharge is not confirmed by our projections. The glaciers in the Nick et al. (2013) study are however driven with only one specific climate model and only represent the response of four individual, well-studied outlet glaciers. In our large-scale model approach, ice discharge of main outlet glaciers can also show a significant increase while the ice sheet-wide discharge increase is more moderate. This is because many of the smaller glaciers become land-based soon. Therefore, scaling up the discharge response of only those glaciers with the most prolific ice export is not necessarily representative for the future ice-dynamic evolution of an entire ice sheet. A generalisation of the discharge evolution of the four outlet glaciers modelled in Nick et al. (2013) to the entire ice sheet is in line with our finding that the relative importance of ice discharge changes to the future ice loss is self-limited by thinning and retreat of ice in contact with the ocean (Goelzer et al., 2013). Though not linking ice discharge changes directly to climatic variables, other projections of the Greenland ice sheet under future warming also found evidence for this self-limiting effect (Gillet-Chaulet et al., 2012; Lipscomb et al., 2013)

In all experiments, the additional effect of basal lubrication on total mass loss is very small, corresponding to an additional sea-level contribution of less than 1 % (Fig. 10b). This is in agreement with recent observational evidence (Tedstone et al., 2014) and results from a parametric approach to link runoff to basal lubrication (Shannon et al., 2013). Lubrication-induced speedup displaces inland ice mass but does in general not remove it. In the upper ablation area, the ice thins as it accelerates, while for melt rates exceeding 2 m yr^{-1} near the margin, the relative speedup decreases under warming, causing a relative thickening (Fig. 10b, also see Eq. 2 and Fig. 1). The reason is that when meltwater export rates exceed a threshold, a channellisation of the basal drainage system is assumed with concurrent reduction of basal lubrication. Ice flow is mainly enhanced close to the equilibrium line where runoff rates cause maximal speedup. This even leads to a negative feedback as the relative thickening of the ablation zone reduces runoff rates through the height-mass balance feedback (Huybrechts et al., 2002).

For both projections periods to 2100 and 2300, the mass loss projections do not depend much on the parameters tuned during the model spin-up (Sect. 2.4). For seven additional and acceptable parameter sets (Table 1), the future sea-level contribution lies within 4 % of the reference model (i.e. ± 2 or ± 12 mm by 2100 or 2300, respectively). The sensitivity of the projections to the parameterisation for warming-induced discharge increase (Eq. 3) is assessed from additional results for the full ensemble obtained with $\alpha = 1.8$ and 2.6. For the period 2000-2010, we find that the relative contribution from ice discharge to total mass loss is ∼20 %, ∼27 % or ∼40 % for α equal to 1.8, 2.6 or 5.2, respectively. The effect on the projections is however somewhat reduced, as ice discharge increase is even more limited. For the sea-level projections (Table 5), variations reach ∼25 % compared to the reference run ($\alpha = 5.2$). Relative to increasing values for α, a saturation of the ice loss increase can be stated. The RMS deviation around these ensemble values is not much affected by the choice of α, and differences mostly fall below 10 %. If one excludes the value 1.8, as the 2000-2010 contribution from ice discharge in this case is rather low, differences between ensemble-mean mass loss lie within 15 % of the standard results. In this case, the sensitivity to changes in α of the mass loss in 2300 is about 10 %, even lower than in 2100. For α-values of 2.6 and 5.2, ocean forcing explains about 30 % or 40 % of the total mass loss in 2100, respectively. By increasing α beyond 5.2, the present-day ice discharge can certainly be increased further. If the value is chosen such that the present discharge contribution stays in an realistic range, we do however not expected that the projection results will qualitatively change. In summary, the projections are sensitive to the choice of α but the sensitivity decreases with the length of the projection period and the warming magnitude. Despite this sensitivity, the spread in future ice loss, introduced by the climate model ensemble, is several times larger (Table 4). This is in line with other studies recognising the importance of the climate trajectory as the main source for the large spread in sea-level projections of the Greenland ice sheet (e.g. Yoshimori et al., 2011; Quiquet et al., 2011; Fyke et al., 2014).

6 Summary and conclusion

In this study, we included more dynamic processes in a thermo-mechanically-coupled, three-dimensional ice flow model with the aim to better assess the impact of ice dynamics on the future evolution of the Greenland ice sheet. We suggested parameterisations that link ice discharge increase to ocean warming and allow for runoff-induced lubrication. To assess the likely range of the future contribution from the Greenland ice sheet to sea-level change, climate anomalies were taken from a suite of ten Atmosphere–Ocean General Circulation Models (Table B1). They were selected from the WCRP's CMIP5 multi-model dataset prepared for the IPCC AR5 (Taylor et al., 2012) and forced by four Representative Concentration Pathway climate scenarios. When considering climate forcing from ECMWF reanalysis data and ocean temperatures from an AOGCM that shows an expressed warming over the period 2005–2010, we find an ice loss rate of 0.62 $\mathrm{mm\,yr^{-1}}$ over the same period that is explained by $\sim 40\,\%$ from increased ice-discharge, in agreement with the observational range. Changes in ice discharge are attributed to oceanic warming in the surrounding ocean basins. The mean ice volume loss for the CMIP5 ensemble is however biased low with 0.32 $\mathrm{mm\,yr^{-1}}$. This bias arises from the spread in climate models that are not expected to correctly simulate the observed trend over such a short period of time. The ensemble maximum of the ice loss during this recent period is 0.71 $\mathrm{mm\,yr^{-1}}$ and equally covers values inferred from observations. For the climate model ensemble, increased ice discharge also explains $\sim 40\,\%$ of the total mass loss during the last decade.

Accounting for the four RCP scenarios, we find a Greenland ice-sheet contribution to global sea-level rise of between 1.4 and 16.6 cm by 2100 AD. For the two low-impact scenarios, ice loss attains respectively 11.1 and 32.0 cm by 2300 AD. Despite an average increase in mass loss of $\sim 40\,\%$ in 2100, when accounting for ice–ocean interaction, mass loss is predominantly caused by changes in SMB. The reason is that ice discharge is limited by margin thinning and retreat but also by a competition with surface melting that removes ice before it reaches the calving fronts. These geometric limits on ice discharge explain that most of the mass loss by 2100 is caused by changes in SMB. Beyond 2100, modelled ice discharge rates fall below the pre-2000 level and this decrease is compensated by the dominant changes in SMB. The results therefore suggest that the largest source of uncertainty in future mass loss arises from the SMB and the underlying climate change projections, and not from ice dynamics.

Our results have implications for attempts to estimate the role of ice discharge on the future mass loss of the Greenland ice sheet. Observed rates of change over the last decade cannot simply be extrapolated over the 21st century on account of a different balance of processes causing mass loss over time. Extrapolating recently observed mass trend changes to a century time scale (Rignot et al., 2011) or linking observed Greenland sea-level trends to temperature change (Rahmstorf, 2007) implies continued glacier acceleration and a multifold increase of the ice discharge (Pfeffer et al., 2008) that is not found attainable in numerical ice-sheet models. Ice discharge at calving fronts is self-limited by ice dynamics, supporting the view that centennial mass changes are dominantly driven by SMB changes, and thus by changes in surface climate conditions.

Appendix A: Climate conditions over the last glacial cycle

Temperature history

The model spin-up over several glacial cycles requires information on the past climate, which is reconstructed from ice core data. The glacial temperature forcing is obtained from synthesised isotope records representative for central Greenland conditions. For the period prior to 122.6 kyr BP, the forcing reconstruction is based on a synthesised Greenland δ^{18}O record derived from Antarctica Dome C using a bipolar seesaw model (Barker et al., 2011). Subsequently, the NGRIP δ^{18}O record (Andersen et al., 2004) is used before switching to GRIP information at 103.8 kyr BP (Dansgaard et al., 1993). For the last 4 kyr, a direct reconstruction of snow temperatures is available based on a a $\delta^{15}\mathrm{N}/\delta^{40}\mathrm{Ar}$ record from GISP2 (Kobashi et al., 2011).

The synthesised δ^{18}O record from Barker et al. (2011) matches well with the GRIP record. Therefore, the fabricated isotope values are transformed into temperature changes according to one single transfer function as given by Huybrechts (2002). For the NGRIP record the same transfer function gives lower temperatures during the LGM compared to the GRIP reconstruction. For the purpose of splicing NGRIP to GRIP, an overlap period for rescaling the transfer function is defined between 102.4 and 90.9 kyr BP. Since present day δ^{18}O values match between GRIP and NGRIP, only the scaling factor is adjusted from 2.40 to 2.13 K ‰$^{-1}$. By replacing information from GRIP with NGRIP during the period 122.6–103.8 kyr BP, the spliced record does not contain the disturbed lower part of the GRIP ice core. The Kobashi et al. (2011) snow temperature reconstruction for the last 4 kyr is offset by its average of $-19.6\,^{\circ}$C during the reference period 1960–1990. Thereafter, the temperature reconstruction shows a mismatch of $0.4\,^{\circ}$C with the GRIP reconstruction at 4 kyr BP. Before splicing these two records, the Kobashi et al. (2011) temperatures are lowered over time with a linear function that removes the past mismatch but keeps the present-day values (Fig. A1). In a final step, the temperature reconstruction is linearly interpolated on time intervals of ten years.

Assembling the forcing record in this way prolongs any records exclusively based on Greenland ice cores by several hundred millennia. In addition, the intermediate switch to the NGRIP record gives more reliable information during the late Eemian period than GRIP. This is because of known disturbances in the lower parts of the GRIP ice core prior to 105 kyr BP. The last splice with surface snow temperature reconstructions at GISP2 seems favourable because this reconstruction method was validated against observations and model reconstructions starting in 1850 AD. One remarkable feature of our assembled temperature forcing record is the Little Ice Age cooling on the Greenland ice sheet (Fig. A1). This cold period 200–500 years ago influences our spin-up into the present-day, and causes ice-sheet growth up to the beginning of the 20th century.

Appendix B: Breakdown of projections by climate model

For most of the climate model ensemble members (Table B1), air temperature anomalies correlate better with the centennial contribution of the Greenland ice sheet to sea-level change than ocean temperature anomalies. Linear correlation coefficients for air temperature in general exceed 0.7 while this threshold is not surpassed for ocean temperatures except in RCP8.5. By 2300, the correlation with ocean forcing dominates for RCP2.6.

The spread in centennial sea-level contributions and atmospheric warming (Fig. B1) reflects both uncertainties in the realised future scenario and differences in the respective AOGCM. Up to 2100 AD, this spread is explained by differences in individual AOGCM projections rather than scenario differences. In particular the three low impact scenarios show a large overlap in AOGCM realisations. By 2300, the spread introduced by the different scenarios is largest. For the two lowest scenarios, the 2300 temperature spread remains similar to the centennial spread while deviations in sea-level contribution become more than twice as large.

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Table 1. Sensitivity of future sea-level change to main model parameters. Values from a previous tuning are indicated together with the reference values for this study. Mean and RMS values are given for the ensemble projections forced with CanESM2/RCP4.5. Positive degree day factors are given in ice equivalent (i.e.).

	Degree-day factor for snow [m i.e. d^{-1} $°C^{-1}$]	Degree-day factor for ice [m i.e. d^{-1} $°C^{-1}$]	Enhancement factor [–]	Sliding coefficient [10^{-10}] [m^2 yr^{-1} Pa^{-3}]	2100 AD Sea level contribution [cm s.l.e.]	2300 AD Sea level contribution [cm s.l.e.]
Previous tuning	0.00300	0.00800	3.50	1.000		
Reference values	0.00297	0.00791	3.28	0.83	9.3	32.7
parameter set 1	0.00303	0.00800	3.22	0.936	9.3	32.0
parameter set 2	0.00294	0.00800	3.28	0.828	9.0	30.5
parameter set 3	0.00267	0.00776	3.47	0.972	8.7	28.6
parameter set 4	0.00276	0.00749	3.40	0.936	8.4	28.2
parameter set 5	0.00285	0.00749	3.40	1.080	8.9	29.8
parameter set 6	0.00303	0.00749	3.28	1.080	9.0	30.4
parameter set 7	0.00322	0.00749	3.40	0.792	9.1	30.6
Mean	0.00293	0.00770	3.34	0.932	9.0	30.1
RMS deviation	±0.00016	±0.00022	±0.08	±0.10	±0.2	±1.1

Table 2. Ice discharge prior to 2000 as inferred by Rignot and Kanagaratnam (2006) and as simulated with the ice sheet model using two resolutions. Observationally inferred values are representative for 1996 (or 2000) while simulated values are averaged over the period 1960–1990. These values therefore represent ice discharge prior to any major acceleration in the outlet glaciers. All values are given in $\mathrm{km^3\,yr^{-1}}$.

	Observations	20 km model	5 km model
North	**50.0**	**76.7**	**76.4**
Humboldt	3.7	14.2	6.1
Petermann	11.8	5.1	12.2
Storstrømmen	0.1	5.0	0.8
Nioghalvfjerdsbræ and Zachariae Isbræ	23.4	28.0	20.2
West	**165.8**	**132.9**	**129.0**
Jakobshavn	23.6	15.8	21.9
Rink Glacier	11.8	2.2	4.1
East	**141.0**	**141.1**	**165.9**
Helheim	26.3	9.9	26.2
Kangerlussuaq	27.8	16.9	22.0
Total	**356.8**	**350.7**	**371.3**

Table 3. Recent SMB changes in six main drainage basins. Values for four SMB model estimates are averaged from Vernon et al. (2013). The GRACE observational mass change record is corrected for ice discharge D based on Fig. 2 in Sasgen et al. (2012). SMB changes are given in $Gt\,yr^{-1}$.

Drainage basin	SMB models mean ± RMS (1996–2008)	GRACE + D (2002–2010)	Ice sheet model SMB component 1996–2008
A	-19 ± 6.9	-17	-14
B	-15 ± 6.8	-15	-12
C	-4 ± 5.4	-16	-35
D + E	-33 ± 15.2	-21	-46
F	-54 ± 19.4	-30	-56
G	-40 ± 7.0	-46	-29
Total change	-165 ± 55.6	**−145**	**−203**

Table 4. Ice sheet-wide mean atmospheric warming, basin-mean oceanic warming, and ensemble-average contribution of the Greenland ice sheet to global sea-level change by 2100 AD and 2300 AD. Sea-level changes are calculated with respect to the year 2000. Ensemble averages for each scenario use equal weights for individual AOGCMs. The root mean square (RMS) deviation from the mean ensemble realisation is added to estimate the variability.

Climate scenario	2100 AD Atmospheric warming [°C]	2100 AD Oceanic warming [°C]	2100 AD Sea-level contribution [cm s.l.e.]	2300 AD Atmospheric warming [°C]	2300 AD Oceanic warming [°C]	2300 AD Sea-level contribution [cm s.l.e.]
RCP2.6	2.10 ± 1.53	1.12 ± 0.57	4.23 ± 1.80	2.59 ± 1.62	1.32 ± 0.73	8.82 ± 4.48
RCP4.5	3.56 ± 1.86	1.62 ± 0.67	5.50 ± 1.86	5.27 ± 1.62	2.77 ± 1.18	20.11 ± 8.03
RCP6.0	4.00 ± 1.59	1.43 ± 0.22	5.40 ± 1.49	–	–	–
RCP8.5	7.15 ± 1.98	2.68 ± 0.94	10.15 ± 3.24	–	–	–

Table 5. Sensitivity of future sea-level contribution from the Greenland ice sheet to the parameterisation of ocean warming-induced discharge increase. Values are ensemble averages with respect to the year 2000, given in cm s.l.e.

Climate scenario	$\alpha = 1.8$ sea-level contribution by 2100 / 2300	$\alpha = 2.6$ sea-level contribution by 2100 / 2300	$\alpha = 5.2$ sea-level contribution by 2100 / 2300
RCP2.6	3.18 / 6.86	3.58 / 7.77	4.23 / 8.82
RCP4.5	4.36 / 17.46	4.77 / 18.63	5.50 / 20.11
RCP6.0	4.38 / -	4.77 / -	5.40 / -
RCP8.5	8.65 / -	9.29 / -	10.15 / -

Table B1. Atmospheric and oceanic temperature forcing as provided by the AOGCMs given together with the resulting Greenland ice sheet contribution to sea-level change by 2100 and 2300. Sea-level contribution is determined with respect to 2000 AD. Ocean temperatures are basin-averages. Also provided are model means and root mean square (RMS) deviations from the mean for each RCP scenario. Hyphens indicate no data for the selected model and period. Ensemble averages are given in bold.

	2100 AD			2300 AD		
Climate scenario and model	Air temp. change [°C]	Ocean temp. change [°C]	Sea level contr. [cm s.l.e.]	Air temp. change [°C]	Ocean temp. change [°C]	Sea level contr. [cm s.l.e.]
RCP2.6						
CanESM2	4.0	2.6	7.8	3.5	2.7	16.3
CCSM4	2.6	1.3	4.1	–	–	–
CSIRO Mk3 6	1.2	1.2	1.4	–	–	–
GFDL ESM2G	0.3	0.6	2.8	–	–	–
GISS E2 R	0.1	0.8	1.9	1.7	1.2	3.4
HadGEM2 ES	4.7	1.0	4.4	3.6	1.3	11.1
IPSL CM5A LR	2.9	0.8	4.7	4.3	1.0	7.2
MIROC5	1.0	0.5	5.0	–	–	–
MPI ESM LR	0.9	1.0	4.4	−0.2	0.5	6.0
NorESM1 M	3.3	1.4	5.0	–	–	–
Model mean	**2.10**	**1.12**	**4.23**	**2.59**	**1.32**	**8.82**
RMS deviation	±1.53	±0.57	±1.80	±1.62	±0.73	±4.48

Table B1. Continued.

Climate scenario and model	2100 AD Air temp. change [°C]	2100 AD Ocean temp. change [°C]	2100 AD Sea level contr. [cm s.l.e.]	2300 AD Air temp. change [°C]	2300 AD Ocean temp. change [°C]	2300 AD Sea level contr. [cm s.l.e.]
RCP4.5						
CanESM2	6.1	3.3	9.3	6.8	5.3	32.0
CCSM4	3.5	1.5	4.5	–	–	–
CSIRO Mk3 6	0.6	1.7	2.8	4.8	3.1	14.4
GFDL ESM2G	1.8	0.9	4.2	–	–	–
GISS E2 R	2.3	1.0	3.3	2.6	1.3	6.7
HadGEM2 ES	6.2	1.5	7.0	7.8	2.5	26.9
IPSL CM5A LR	5.1	1.2	5.7	5.7	1.8	19.4
MIROC5	4.4	1.4	6.5	–	–	–
MPI ESM LR	1.4	1.6	5.0	3.8	2.5	15.6
NorESM1 M	4.1	1.8	6.7	5.4	2.7	25.7
Model mean	**3.56**	**1.62**	**5.50**	**5.27**	**2.77**	**20.11**
RMS deviation	±1.86	±0.67	±1.86	±1.62	±1.18	±8.03

Table B1. Continued.

Climate scenario and model	2100 AD Air temp. change [°C]	2100 AD Ocean temp. change [°C]	2100 AD Sea level contr. [cm s.l.e.]	2300 AD Air temp. change [°C]	2300 AD Ocean temp. change [°C]	2300 AD Sea level contr. [cm s.l.e.]
RCP6.0						
CanESM2	–	–	–	–	–	–
CCSM4	5.2	1.7	5.8	–	–	–
CSIRO Mk3 6	1.2	1.5	2.7	–	–	–
GFDL ESM2G	2.7	1.2	4.3	–	–	–
GISS E2 R	2.5	1.1	3.8	–	–	–
HadGEM2 ES	6.3	1.7	6.9	–	–	–
IPSL CM5A LR	5.0	1.2	6.4	–	–	–
MIROC5	4.4	1.3	6.3	–	–	–
MPI ESM LR	–	–	–	–	–	–
NorESM1 M	4.7	1.7	7.1	–	–	–
Model mean	**4.00**	**1.43**	**5.40**	–	–	–
RMS deviation	±1.59	±0.22	±1.49	–	–	–

Table B1. Continued.

Climate scenario and model	2100 AD Air temp. change [°C]	2100 AD Ocean temp. change [°C]	2100 AD Sea level contr. [cm s.l.e.]	2300 AD Air temp. change [°C]	2300 AD Ocean temp. change [°C]	2300 AD Sea level contr. [cm s.l.e.]
RCP8.5						
CanESM2	8.6	5.0	16.6	–	–	–
CCSM4	6.7	2.0	8.7	–	–	–
CSIRO Mk3 6	5.9	2.9	6.8	–	–	–
GFDL ESM2G	6.1	2.1	7.1	–	–	–
GISS E2 R	4.1	1.1	5.1	–	–	–
HadGEM2 ES	11.1	2.9	11.7	–	–	–
IPSL CM5A LR	7.8	2.7	11.2	–	–	–
MIROC5	9.4	2.8	13.0	–	–	–
MPI ESM LR	5.3	2.7	9.1	–	–	–
NorESM1 M	6.5	2.3	11.9	–	–	–
Model mean	**7.15**	**2.68**	**10.15**	–	–	–
RMS deviation	±1.98	±0.94	±3.25	–	–	–

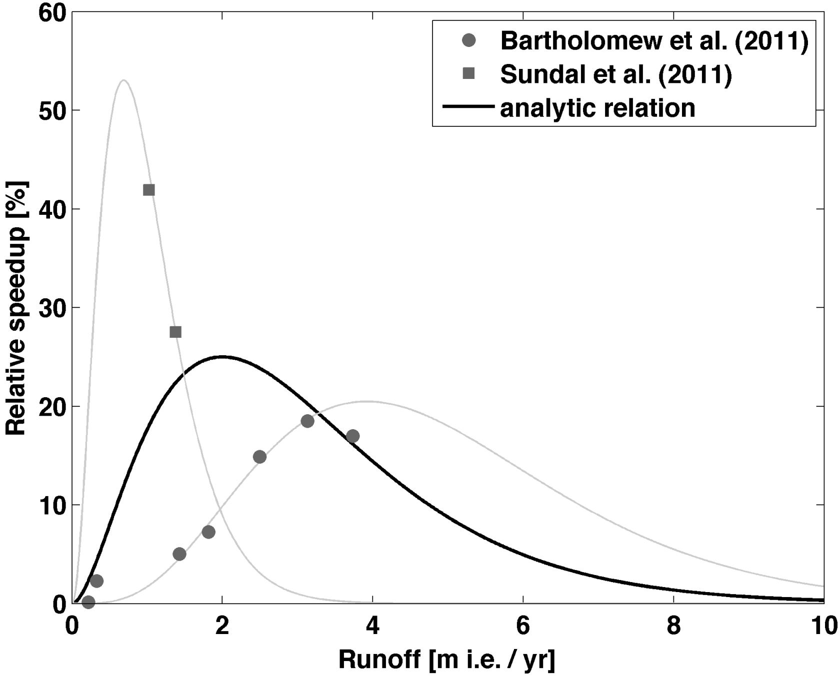

Figure 1. Functional dependence of relative annual speedup to local runoff. Grey symbols indicate either direct field observations (Bartholomew et al., 2011) or observed speedup combined with output from a SMB model (Sundal et al., 2011). Observational data originate from Russell Glacier, east of Kangerlussuaq. The parameterisation considers a functional dependence (black line) that is a compromise between all observations. Grey thin lines indicate a best fit to the respective data sets.

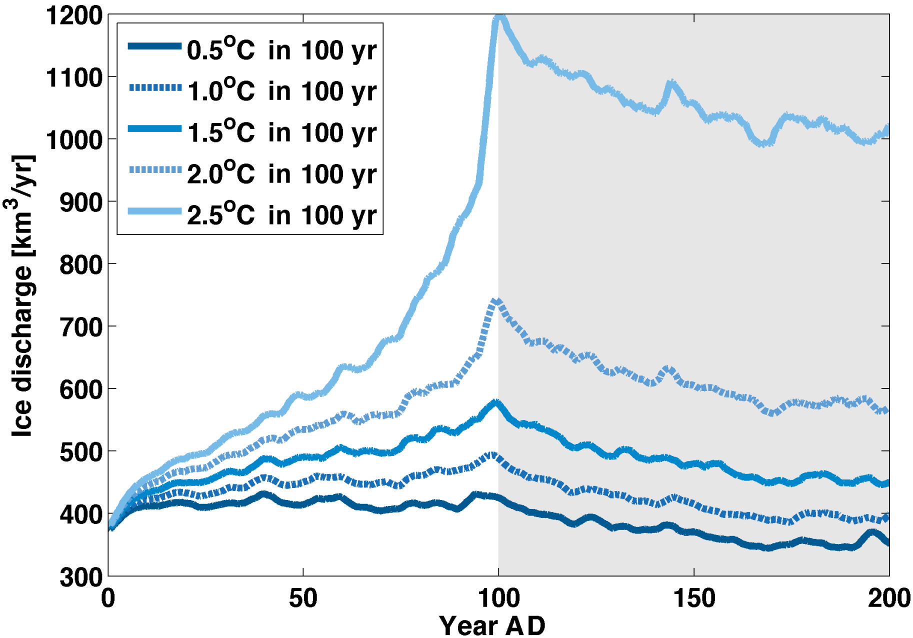

Figure 2. Ice discharge response to a linear increase in ocean temperature. The atmospheric forcing is unchanged and based on the SMB of one climate model (i.e. 2005 MPI-ESM-LR). Ocean temperature increase is linear for hundred years and is then kept at the same level.

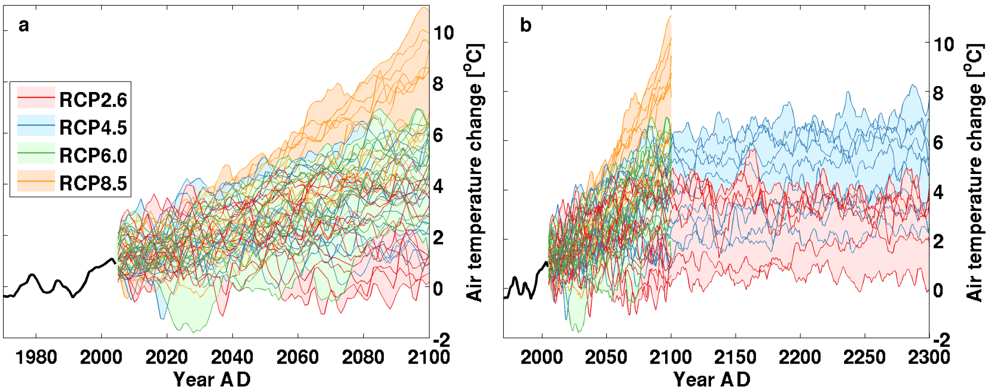

Figure 3. Mean annual surface air temperature anomaly over the present ice sheet extent with respect to the reference period 1960–1990. For illustration, the monthly temperature forcing is smoothed with a 5 year running mean. Panels cover different time periods up to 2100 **(a)** and 2300 AD **(b)**. Thin lines represent individual projections and the lighter background shading covers the area between the minimum and maximum realisation for each RCP except when they overlap with other scenarios. Prior to the year 2005, the temperature forcing comes from the ECMWF ERA-40 and ERA-Interim meteorological reanalyses (black line).

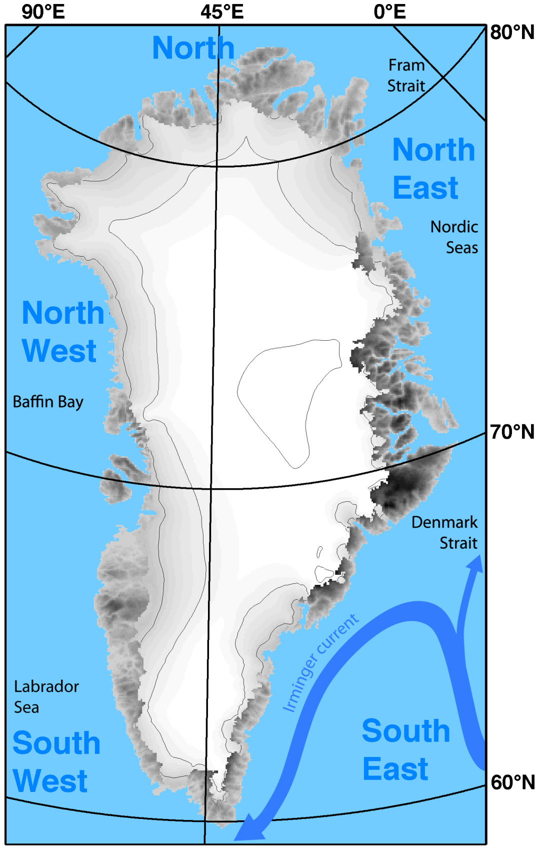

Figure 4. Observed ice sheet geometry. Surface elevation for ice sheet and bed topography are given in different grey shading. Over the ice sheet, contour lines for surface elevation are indicated with 1000 m spacing. The five ocean basins are labelled (bold, dark blue). They are separated by the three shown latitudes and Greenland. Oceanographic names are given in black and the Irminger current is delineated.

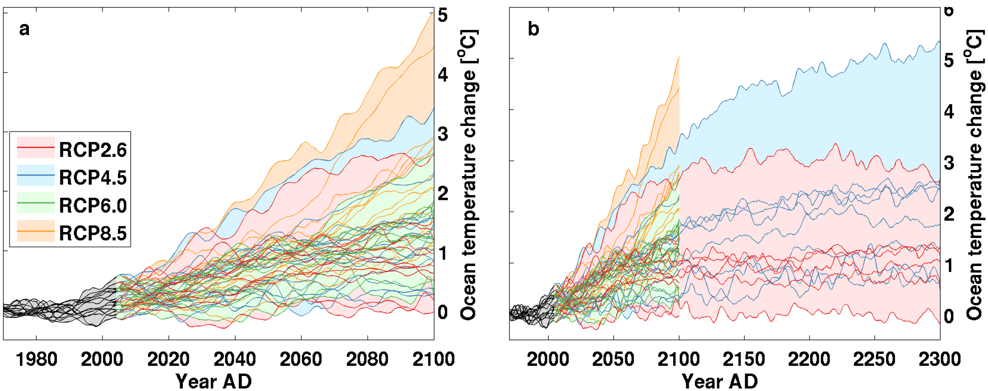

Figure 5. Mean annual ocean temperature anomaly around Greenland with respect to the reference period 1960–1990. Panels cover different time periods up to 2100 **(a)** and 2300 AD **(b)**. Thin lines represent individual projections and the lighter background shading covers the area between the minimum and maximum realisation for each RCP except when they overlap with other scenarios. Temperature anomalies are averaged over the five ocean basins. Prior to 2005, ocean forcing is taken from each individual climate model (grey shading and black lines).

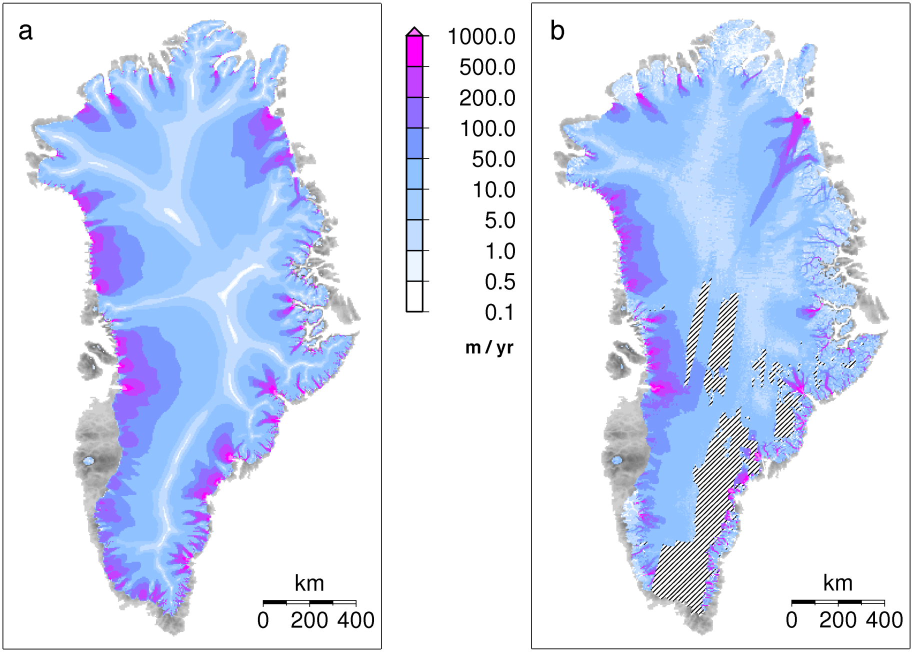

Figure 6. Comparison of present-day modelled **(a)** and observed **(b)** surface velocities. Observations are averaged over the years 2000 and 2005–2008 AD (Joughin et al., 2010).

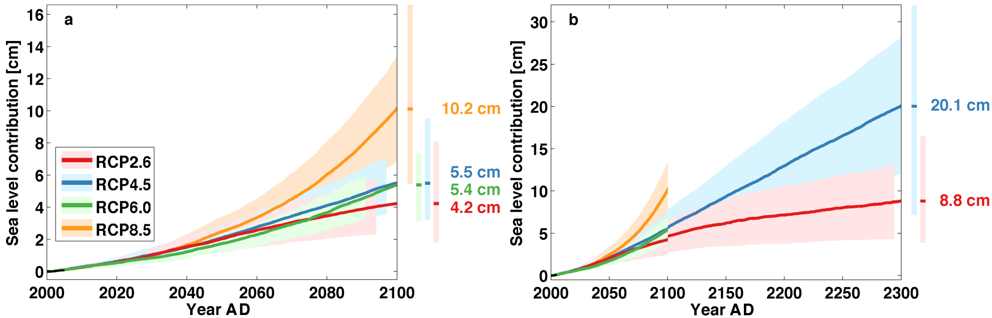

Figure 7. Greenland ice sheet contribution to future global sea-level change. Given are ensemble averages for each scenario during the 21st century **(a)** and the next three centuries **(b)**. The modelled rate of mass loss during the observational period (2000–2010) is on average 0.32 $mm\,yr^{-1}$. Colours indicate the respective RCP scenario and the lighter background colour is for the standard deviation from each mean trajectory. Vertical bars indicate the spread of sea-level contributions arising from individual AOGCMs at the end of each scenario. The jump across the year 2100 in the right panel arises from a different number of climate models.

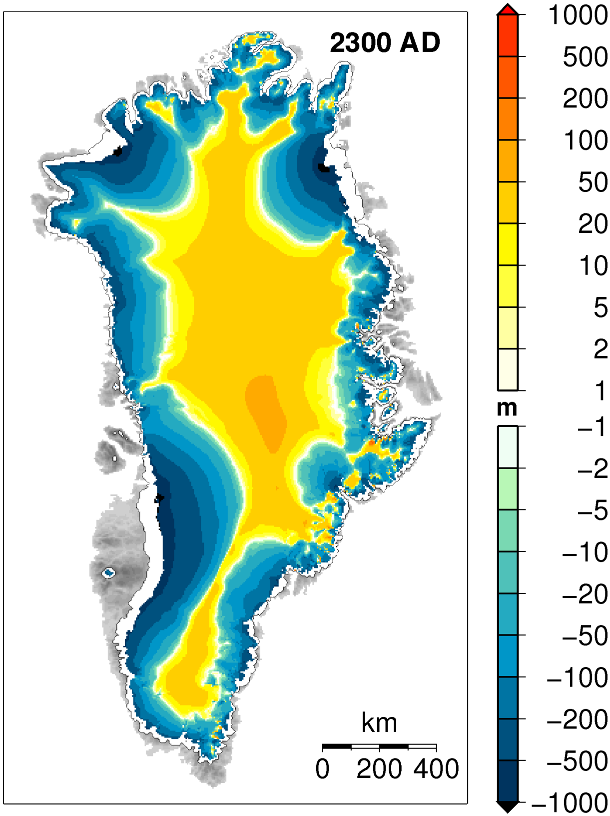

Figure 8. Total ice thickness change by 2300 AD. The initial ice extent is indicated with a black contour line while thickness changes are exclusively shown within the ice extent at the end of the experiment. This particular result was obtained with CanESM2 for RCP4.5, one of the most sensitive climate models in the ensemble. The thinning patterns for other ensemble members are qualitatively similar.

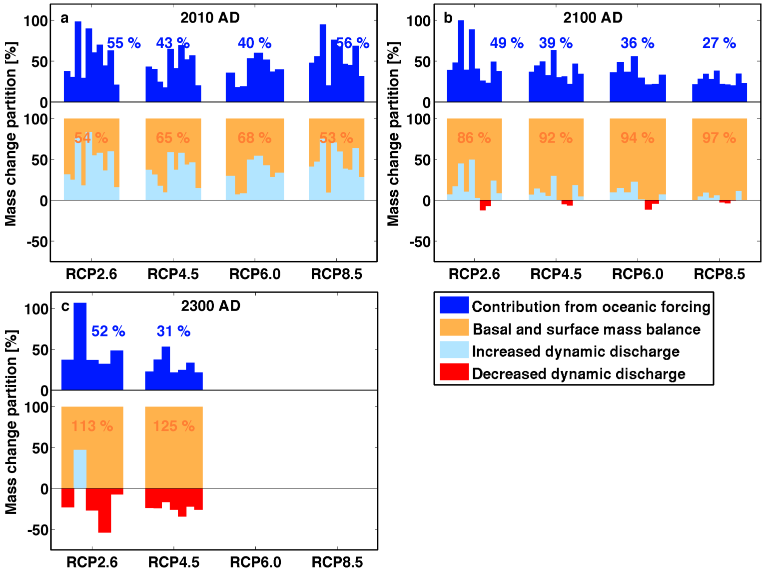

Figure 9. Partitioning of mass changes by 2010 **(a)**, 2100 **(b)** and 2300 AD **(c)**. Values are given relative to the total ice loss of the individual AOGCM projection and grouped by climate scenario. Each vertical column represents one AOGCM projection. The dark blue columns denote the contribution to the total mass change arising from oceanic forcing, diagnosed from a control run with SMB forcing only. The diagnostics comprise the directly induced ice discharge changes but also the indirect feedback with the SMB via the ice geometry. The mass change of the projections is subsequently partitioned into contributions from changes in both basal melt and SMB (orange columns) or in ice discharge (light blue and red columns). A decreasing ice discharge (red) is overcompensated by the respective contribution from SMB to close the total sea-level contribution. The presented partitioning of the mass change is cumulative. Changes with respect to the average 1990-2000 values of all contributors in the mass budget are integrated over time. The scenario averages are then given in per cent.

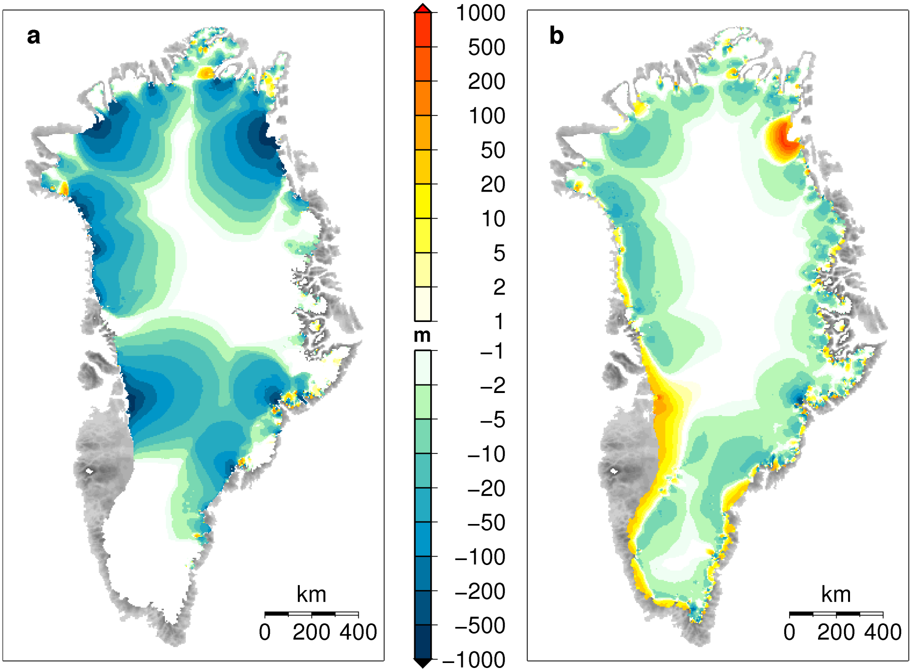

Figure 10. Additional ice thickness changes from ocean warming-induced discharge increase **(a)** and runoff-induced lubrication **(b).** In this particular experiment, obtained with CanESM2 for RCP4.5, additional oceanic forcing accounts for 7.4 cm of the total sea-level contribution of 32.0 cm. The effect of basal lubrication increases mass loss by 0.1 cm. This small extra contribution results from a general ice displacement expressed by relative thinning of the upper ablation area and resulting thickening of the marine margin.

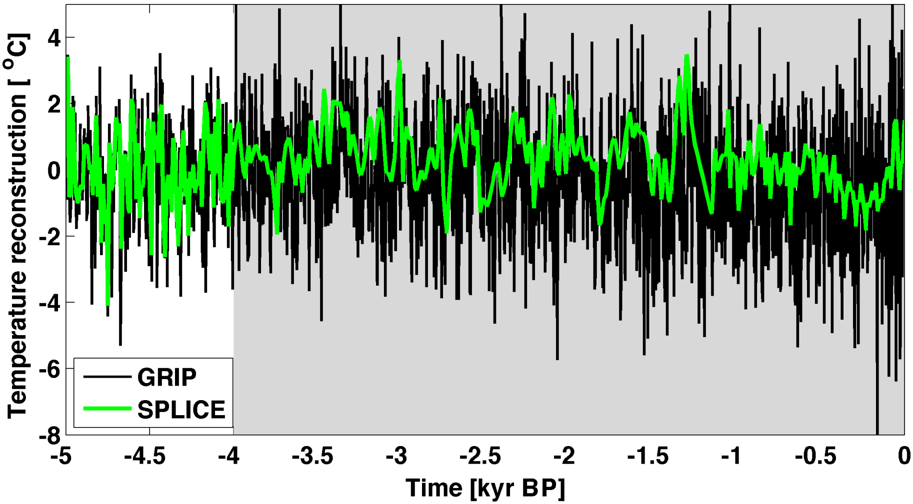

Figure A1. Assembled temperature forcing during the last 5 kyr based on the δ^{18}O GRIP ice core record and a direct temperature reconstruction. The splicing point of these two records is indicated by a change in the background shading at 4 kyr BP. Note that the original GRIP record shows sub-decadal resolution during the Holocene period while the used temperature forcing is linearly interpolated for a decadal sampling rate.

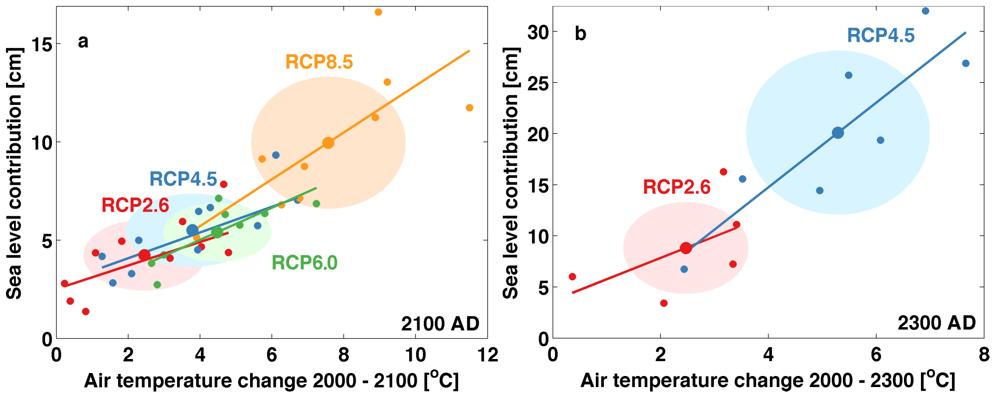

Figure B1. Greenland ice sheet contribution to global sea level change as a function of regional atmospheric warming by 2100 **(a)** and 2300 AD **(b)**. Temperature changes are taken as differences between 10-yr averages at either end of the projection period. Small dots represent each individual realisation with colours indicating the RCP scenario. The respectively coloured lines are a linear fit to each RCP response. Larger dots indicate the model averages for each RCP. Ellipses indicate RMS deviations in both temperature change and sea-level change.