Response to Reviewer Comments

Journal: The Cryosphere
Title: **Impact of the melt-albedo feedback on the future evolution of the Greenland Ice Sheet with PISM-dEBM-simple**
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First of all, we would like to thank the editor Kerim Nisancioglu and the two reviewers, Signe Hillerup Larsen and Mario Krapp for their helpful and excellent comments and their efforts to create the detailed reviews! In our revision of the manuscript we addressed the main issues:

1. We have now included a comparison with the positive degree day melt model, which is widely used in the ice-sheet modeling community.
2. We have rewritten the methods section in order to increase clarity.
3. We have clarified the framing of the results and included some of the discussion points closer to the results.

We provide detailed answers to all comments below. The reviewers' comments are given in black and the authors' in blue. The changes made to the main document can be found at the end of this document (created with latexdiff).

Signe Hillerup Larsen

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Review of Zeitz et al in the cryosphere discussion
Summary
The study presents a new module for the ice flow model PISM, making it possible to include effects of changes in global radiation and albedo, in the model. In order to make a computationally efficient experiments, MAR albedo and global radiation is not used directly but implemented as parametrisations. This model setup is then used to explore the effect of the melt-albedo feedback in PISM. The manuscript covers a description of the method and performed experiments testing the effect of the implemented module and parametrisation on model prescribed ice mass loss.
The originality is that the study presents a new module to calculate melt in PISM. This module is as simple to implement as the pdd melt model, given that albedo and transmissivity is parametrised, but then offers the benefit of being able to add both effects of changing surface reflection, and incoming shortwave radiation. In this way it is possible to model the effect of changing incoming radiation on melt in for example the Eemian interglacial as well as investigating the effect of the melt-albedo feedback in predictive ice-sheet models.

The significance of the study is the implementation of a new module in PISM, making it possible to experiment with the melt-albedo feedback as well as orbital parameters in PISM, and that the modelled ice loss is significantly altered when using the melt-albedo feedback. A revision of the presentation of the study is my opinion needed in order for the paper to be published, but I think it is a matter making sure that the conclusions are discussed in the right context. See my comments below:

General comments

My impression of the manuscript is that conclusions are drawn on the mass loss due to the melt albedo feedback in general. This is, in my opinion, a job for a focussed study using an advanced (regional) climate model. I suggest to alter the focus of the manuscript to investigate the effect of adding the melt-albedo feedback in PISM projections. As an example changing the following sentence from the conclusions:

"Using PISM-dEBM-simple we find that the melt–albedo feedback can lead to additional 12 cm sea-level equivalent of mass loss in RCP2.6 and additional 70 cm in RCP8.5 until the year 2300"

to something like this:

"Using PISM-dEBM-simple we find that the melt–albedo feedback can lead to additional 12 cm sea-level equivalent of mass loss in RCP2.6 and additional 70 cm in RCP8.5 in the projected mass loss from PISM until the year 2300"

This is a good suggestestion to be clear about the distinction between the simple approach used in this manuscript and a full climate model. We incorporated the suggested framing in the manuscript.

The model performance is investigated from many different angles, but I think a comparison between using dEBM-simple with using the simple pdd module, on the MAR historical time series, could add some insights into that dEMB is actually a more physically based model – that as is also shown also makes it possible to make more realistic experiments further back in time where pdd factors are certain not to be the same, as is shown in the Eem experiment.

We now show the melt rates computed with pdd from the historic validation experiment with spatially resolved temperature fields (similar to Figure 1 and 2) and discuss them in Appendix D. We find that the pdd module, with standard parameters, gives a similar root mean square error over the historic period, both over the cumulated melt per year (39.85 Gt with pdd vs. 32.92 Gt with dEBM-simple) and over the spatial June, July and August melt (0.47 mWE/year with pdd vs. 0.36 mWE/year with dEBM simple). However, the pdd melt introduces a North-South gradient to the melt anomaly (compared to MAR data), overestimating the melt in the North and underestimating the melt in the South which underlines that the additional physics modelled with dEBM-simple improves the melt calculation in comparison to the pdd.

Abstract:

Page 1

Line 1- 3: The melt-albedo feedback that is investigated here is only on the snow part of the ice sheet.

While it is true that the albedo is only varied between a snow-albedo and an ice-albedo value, the feedback itself does not depend on the presence of snow. The parametrization can be easily adjusted to go below the ice-albedo value, possibly with a different slope (as we tested in Appendix B).

We have slightly adjusted the wording in the abstract, to reflect that the albedo can change over the whole ice sheet ("... may lower the reflectivity over the ice-sheet surface...")

Line 3: add a sentence like: In order to test the effect of melt-albedo feedback in a prognostic ice sheet model, we implement...

done

See also general comments about focus of the conclusions.

We have adapted the conclusions accordingly.

Introduction:

From reading the introduction, I don't think it is entirely clear what the "-simple" refer to? Does it refer to the simple version of dEBM presented in the 2018 paper or does it refer to simplifications made in this study?

Thank you for the good question. dEBM-simple refers to the simplifications made in this manuscript. It is based on the version presented in (Krebs-Kanzow et al., 2018) paper. The main changes are adapted empirical parameters and the parameterization of albedo and atmospheric transmissivity. It is implemented in the Parallel Ice Sheet Model (PISM) and can replace the Positive Degree Day method for the calculation of climatic mass balance.

We have clarified the abstract and the introduction in that regard.

Page 2

Line 5-6: These are in fact areas that are not considered or discussed in the present study, later on this should be discussed in more detail.

Those areas serve as a motivation to perform the experiment with lowered ice albedo and overall the darkening experiment. We now mention this at the end of the introduction.

Line 7: Replace "As darker snow ..." with something like: As darker surface absorb more radiation than lighter surfaces, the effect of darkening due to increased melt could trigger...

done

Line 14: Replace "covered by meltwater" with: at melting point

done

Line 24-25: perhaps move the references up into the text: The insolation-temperature-melt equations defined/used by van den Berg and Robinson ...

done

Line 32: This is the first time you mention PROMICE – you need to add info on what that is. Or rewrite to say that the model showed good correlation with observations.

done

Page 3

Line 3: Delete "in addition". It is not in addition, but in order to do what you do as you state later in the sentence.

"In addition" should have referred to the fact that the parametrizations of albedo and transmissivity are in addition to the work of Krebs-Kanzow et al. (2018). This is clarified now in the manuscript.

Line 7: This is the first time you mention MAR – spell it out and add references

done

Line 18 – 19: "PISM was shown ..." I guess this is actually mostly true when run on a spatial grid below 1km. Here you run on 4.5 km (and that is completely fine for this purpose), but perhaps state the resolution issues here somehow.

done. We now say "PISM was shown to be capable of reproducing the complex flow patterns evident in Greenland's outlet glaciers at high resolution of less than 1km" and mention explicitly the lower resolution used in this manuscript at the end of the subsection.

Methods

The parametrisation of the melt-albedo feedback is the weakest point of the study, and care should be taken that aspects of the consequences of the simple melt-albedo feedback parametrisation are discussed thoroughly perhaps already in the methods section. For example – has this been done in other studies before? Discussion of in particular the melt albedo feedback parametrization is needed in order to be able to draw any conclusions of the contribution of this to the ice sheet mass balance. After reading the methods section. I have questions such as: What is the consequences of only looking at the snow zone – and thereby neglecting albedo increase in the bare ice zone? Where has the ice sheet been observed to have the alpha_min value that the study is using? How large is the part of the ablation zone that is snow covered compared to the bare ice zone?

As far as we know, this is the first study to parameterize the albedo with the melt rate directly. Many other studies, which use simple albedo parameterizations, include the snow (and sometimes firn) thickness explicitly, linearly or exponentially scaled (see Robinson et al., 2010 or Krapp et al., 2017). Some consider wet snow explicitly (e.g. Robinson et al., 2010). These methods do not in fact consider an albedo decrease in the bare ice zone, in contrast to e.g. the regional climate model MAR, which explicitly considers excess surface melt water, but not impurities, algae or bacteria or changes in the ice structure.

In our setup, scaling albedo with snow thickness led consistently to an underestimation of albedo and an overestimation of melt in the dry and cold northern parts of Greenland, which motivated us to find an alternative parameterization.

Lowering $\alpha_{\min}$ expands the range of the feedback and serves as a first approximation to changes in the bare ice albedo. In order to distinguish albedo changes in snow and ice, one could use a different slope for albedo changes in ice. However, for simplicity the experiments as in Figure 5, where we explore the sensitivity of the melt to the minimal ice albedo, can be used as a first approximation. We find that, with the parametrization presented here, only high melt scenarios are sensitive to changes in minimal albedo.

We also include an analysis of daily observed MODIS albedos over Greenland in the Appendix A in order to show the areas where the albedo is equal to or lower than the minimal albedo αmin. We also discuss the parameterization in the methods in a more detailed way.

Page 3

Line 32: "We neglect ..." Do you allow for shelfs when you have fixed calving front position?

Yes, we do allow for shelves. As mentioned in the manuscript the outline is given by the BedMachine Data, (Morlighem et al., 2017). We do not allow the shelves to grow and do not consider changes in sub-shelf melt due to increasing temperatures.

Page 4

Line 9-11: I think this becomes a bit confusing. I suggest that you consider the following two paragraphs as the place where you introduce all the different parts of equation (1). This means that you do not need to mention albedo and transmissivity in line 9, as you will go through them later and the following sentences describing c1 and c2, should go down after you introduce Teff.

Thank you for the suggestion. We cleaned up the subsection and hope that it is clearer now.

Page 5

Line 17: Does this basically mean that the transmissivity is an average over 2019? A sentence or two of what this actually means in relation to the real transmissivity would be informative in order to understand the prognostic potential of the parametrisation.

The average is over the years 1958 to 2019, considering only the summer months (June, July and August). The parametrization relies on the assumption that the mean transmissivity does not change in a changing climate. In particular the impact of extremes, e.g. Greenland blocking, which might become more frequent in future, is not captured with this approach.
We have added this explanation to the manuscript.

Line 24: Perhaps mention here what you then neglect by introducing the linear relation, like effect of clouds.
Added in the paragraph above

Line 28: delete the sentence starting with "Regional climate models..." And start next sentence with something like "Snow albedo in MAR is calculated using a snowpack model, explicitly ..."
done

Page 6

Line 1: The sentence starting with "Ice albedo ..." could be rephrased to: "In MAR, ice albedo is explicitly ..."
done

Line 6: add information about MAR version and reference to data
done

Line 7: "Allow us to capture melt processes". I am not sure exactly what is meant here
We clarified the sentence.

Line 9: Introducing alpha_min: This means that you do not consider the darkening of ice at all. This should be pointed out somehow, maybe here or somewhere else, but it is an important point, and also, have alpha min been observed at anytime across the ice sheet during melt events?
We have pointed out now that the ice is not darkened in this framework, but it could be easily expanded into this direction. We also include an analysis of observed MODIS albedos in Appendix A and refer the reader to it. This analysis shows that albedo values between 0.45 and 0.5 are reached regionally for up to 50 days a year (average over 2000-2019). However, the majority of albedo values is close to the snow albedo value (with the 0.82 being the most frequent value).

Line 18: Is the geometry kept fixed?
During the spin-up the geometry is not kept fixed. We clarify that paragraph because it is misleading.

Line 20: perhaps just write that precipitation is kept constant – in stead of writing that it is not scaled...
done

Line 25-31: I think this needs to be elaborated. You do this calibration experiment where you do not use PISM-dEBM-simple but force with the data that you have parametrized.
We do use PISM-dEBM-simple, but it still needs precipitation and near-surface air-temperatures. Those are taken from the MAR data. We have clarified the paragraph.
I think this is more or less what it already says, but I think it needs to be reformulated. You need to explain why you do the Eem test? I suppose it is to test the sensitivity to insolation values - but it needs to be clearer.
Exactly, the Eemian experiment is designed to show how the melt rates depend on a change in insolation values, in particular because the temperature field is not changed in this case. We clarified this in the text.

Page 7

Line 4-7: Rephrase: Does these monthly temperature fields come from MAR? And where does these scalar temperature anomalies come from?
done

From line 8: I would like to have the reasons for each of the experiment series before the method is describe. Basically moving the information from page 8 line 8-20 up. Would it somehow be possible to group the experiments into four groups, so that it is easier to follow which group of experiments that are being discussed in section 4?

We have reorganized the section and introduced a table with seven groups of experiments. Now the section starts with the motivation for the experiments.

Page 8

Line 21: Should this title refer better to sec 2.5? By using the word calibration perhaps?

Now "validation" is used for both sections, since it describes more accurately what is done in this section.

Line 22-23: I am missing a sentence like: As described in the Calibration experiments in the methods section?

done

Section 3.1:

This section is not completely clear to me. Perhaps spell out a bit more what is calibration and what is validation.

Line 27: Is this an experiment? I thought perhaps it could be called a calibration run?

Line 31: The root mean square error of what field? Melt?

Indeed, we show the root mean square error of the melt field. It is now clarified in the text.

Page 9

Line 4: Maybe add a sentence here to sort of conclude that using the RMSE method described, you find the parametrisation constants?

done

Line 5: "Yearly total melt computed with PISM" While using the dEBM-simple method?

This is now clarified in the text.

Line 12: Could this also relate to the fact that you do not consider the darkening of ice?

This is indeed possible. This thought is taken up in the manuscript.

Page 10

Section 3.2:

I think that it needs to clarified throughout the text that this experiment is done to test the sensitivity to the orbital parameters (or something like that).

done

Page 11

Line 4: clarify which historic variability?

done

Line 5-6: "This is in line with ..." So I guess this is the point of the experiment - basically to test if you get similar results to others.

Exactly. We have added a sentence in the beginning of the subsection, so that this information does not come as a surprise to the reader.

Section 4

It would be nice if it was clear here which of the experiments are being discussed. Perhaps if they are put into four groups as suggested above, this would be easier. Keep in mind that the resolution of 4.5km actually prevents the model from resolving the ice streams properly - this could have an influence on the surface-elevation feedback. Then on the other side, the fixed calving front must add some effect of inducing ice streaming.

Page 12

Line 1-2: Needs to be rephrased. Here the lower bounds of the experiments in this study. It is a lower bound for the model ice losses. I do no believe that this model set-up is able or discussed to high enough detail to give the lower bound for actual ice losses.

done

Page 13

Line 1-2: "But also ..." rephrase sentence

done

Line 7: By this point I have forgotten the timescales that the experiments are being conducted at. Perhaps remind the reader

Line 9: melt-albedo

done

Line 11-12: Explain why the melt-albedo feedback becomes less important with time? I suspect that this is due to the fact that the entire ice sheet gets the minimum albedo.

Exactly. We have added this interpretation to the manuscript.

Page 15

Line 23: Reducing the frequency how? I think actually, the frequency is really interesting. What effect do we get if we get more extreme years like 2012, and with this module this could actually be tested.

The darkening experiment as described in the main text is an extreme scenario, where the reflectivity drops over the whole ice sheet for the entire summer for each year.

Reducing the frequency of those events, i.e. the reflectivity drops over the whole ice sheet for the entire summer, but not each year but every two years, leads to less ice losses compared to the more extreme case. If darkening in every year increases ice losses by 70%, the increase would drop to approx. 35% if only every second year experiences a darkening event.

Knowing the projected frequency of extreme melt years could thus give an estimate of the additional mass losses, which we think also would be a nice future application of the model.

We have clarified this section.

And why do you think June is most sensitive to darkening?

Two effects might play a role: 1) The days are longest and the insolation highest. 2) In the beginning of the melt season the albedo is still high. The artificial reduction of albedo has the strongest effect then.

We briefly discuss this in the manuscript now.

Discussion

I like the discussion, and it shows that considerations

Page 16

Line 15: Sentence starting with "This is because..." Perhaps the sentence should be slightly rephrased, however, this is the kind of reasoning I think is missing in the two sections above.

done

Line 30 - page 17 line 4: Paragraph starting with "Therefore the only ..." This is a great paragraph and it really frames the whole study.

Thank you!

Page 17

Line 6-7: Perhaps mention why the model overestimates early melt and underestimates late melt.

done

Line 17-18: "It is a coarse representation of what is important of the albedo of snow and ice". Actually it is only a representation of the albedo of snow – as the minimum albedo is clean ice?

We reformulated the sentence in the manuscript.

Conclusion

Page 19

Line 11-12: I think this sentence could be expanded to something like: Using dEBM-simple we find that the melt-albedo feed back can lead to additional 12 cm SLE ... in the projected mass loss in PISM.

done

Mario Krapp

Received and published: 21 June 2021

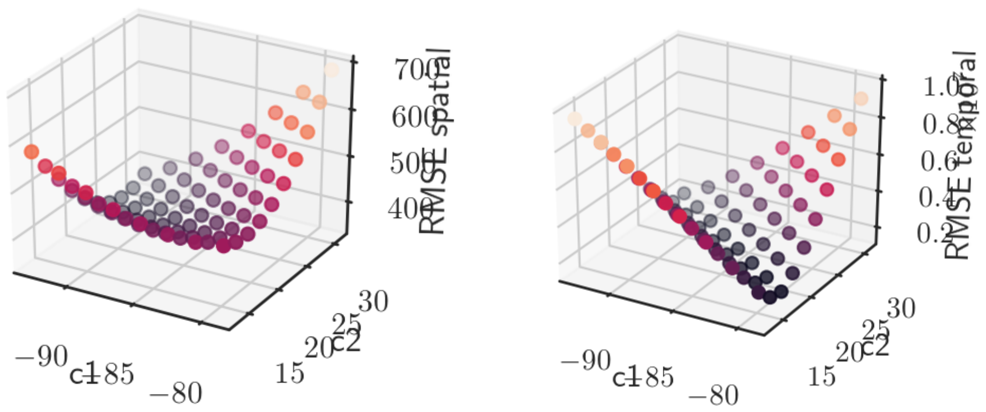
The paper describes how a melt parametrisation affects surface melting on the Greenland Ice Sheet for different forcing scenarios. It is a simplified version of a previously publishes version of a melt scheme and uses temperature and insolation as inputs to calculate surface melt rates. This scheme is fast and simple (in terms of its input) and can therefore replace the positive-degree-day melt scheme which is the current melt scheme used in the numerical ice sheet model PISM. This paper is well written and it can be a valuable contribution for the ice sheet modelling community and should therefore, certainly find a home in TC. However, I cannot recommend this paper being published in its current form as it leaves some open questions that I feel need to be addressed first. The authors have put a lot of effort in the experimental setup but, I think, they almost tried to do too many things at once. I would recommend to them sorting out priorities of what experiments add to the story they want to tell and why, and how they then tell it. Therefore, it requires major revisions for which I have some, hopefully helpful, comments and suggestions. Find below a list of major and minor comments that should be addressed by the authors.

Major Comments

- **Abstract:** the relative changes in surface melt don't mean much without a reference value to relate them to. What do you compare your ice loss with the new melt scheme with? I assume the default melt scheme for PISM is PDD, so why don't you compare your results against that?

  The focus of the manuscript is not the comparison between PDD and dEBM-simple. Therefore we do not compare those two schemes in the abstract. However, as pointed out correctly, a relative change is meaningless without a reference value. We now make explicit that the reference value in this case is the simulation with fixed albedo values.

- It is not easy to follow the results **(Sect 3 and 4)**. The flow of the paper is interrupted by the need too flip pages back and forth too many time to find the respective experiment for the respective results (flip between Sect 2.5 and Sect 3/4). My suggestions: Present the experiments in order with their results in the appropriate (sub)sections. Provide a matrix of the experiments and what you are testing with those (in a table) and briefly describe that matrix in Sect 2.6.
  We have rewritten and clarified the methods- and results sections, so that they are easier to follow. We have also introduced a table summarizing the experiments.
- **Sect 2.5:** It's not clear to me what the calibration experiments are (2.5) What model parameters have been calibrated and what parametrisations have been tested?
  Indeed, the title of the section was misleading. We have changed the title to "validation experiments", since this title reflects more precisely what has been done. We do not present a thorough calibration in this manuscript. See Krebs-Kanzow et al. (2018) for a calibration.
  These experiments are performed with the coupled PISM-dEBM-simple, however, the topography is held fixed in order to ensure comparability to the MAR simulations, also performed with a fixed ice-sheet geometry.



  However, see the spatial and temporal RMSE for the parameters we have tested before running the simulations of the study.
- **Sect 2.6**: Reading on from the previous section, are these experiments now coupled and what is turned on and off?
  These experiments are also performed with PISM-dEBM-simple, however, compared to the previous section, the ice-sheet geometry is now allowed to evolve due to changes in climatic mass balance and ice flow. Changes in the topography induce the melt-elevation feedback via an atmospheric temperature lapse rate.
  We have clarified these points in the manuscript.
- What the major uncertainties/limitations of your melt scheme? For example, cloud cover is not explicitly considered here. Is that a problem and how much can the transmissivity parametrisation account for that? (Extreme) surface melt "events" as mentioned in the introduction are another example. Do they even exists in your forcing? For how much surface melt do they account for?
  Figure 1 shows that the melt rate in most years is well represented without considering changing atmospheric transmissivity. Indeed, changes in cloud cover are not considered. To the contrary, the

parameterization of the atmospheric transmissivity with surface altitude entails the assumption that average cloud cover does not change in the future.

Forcing with historic MAR temperature data shows that indeed, the extreme melt in the years 2012 and 2019 is underestimated in the PISM-dEBM-simple scheme (Figure 1). While the melt of the year 2019 is underestimated due to the parametrization of albedo and transmissivity and can be improved by taking the shortwave downward radiation and the albedo fields as input rather than parametrizing (see Figure A3), the melt in the year 2012 can not be reproduced. However, in the full dEBM model, the cloud cover strengthens the 2012 melt (Krebs-Kanzow, personal communication). Otherwise cloud cover changes do not seem to drive the variability (Krebs-Kanzow, personal communication)

Moreover, the forcing for the future scenarios, as RCP2.6 or RCP8.5, does not take extremes specifically into account but uses the temperature data from CMIP5, which is likely to underestimate temperature extremes, since the associated strong negative NAO index that led to persistent anticyclonic pressure heights over Greenland (Tedesco et al., 2020, Hofer et al., 2017, Bevas et al., 2019) is absent in any future CMIP5 projection(Hanna et al., 2018).

- I don't understand how PISM is used here, offline or interactive simulations, and which experiments are what? In the abstract you refer to "dynamic simulations" of the GIS, but I can't seem to find which of your experiments are coupled and which are uncoupled

  PISM-dEBM-simple is always used in coupled mode, since dEBM-simple is implemented as a surface module in the Parallel Ice Sheet Model PISM. However, the validation simulations use a fixed topography, mainly in order to ensure the comparability to MAR simulations.

  The forward simulations with the RCP forcings use the full ice dynamics and allow for changes in topography, which also allows for the melt-elevation feedback. We have added a clarification in Section 2.3.

- One aim of this paper is to provide a fast scheme for centennial- to millennial scale time scales. I was just wondering, in case it has been coupled, it would be interesting for the readers to see what the long-term effect of using the new melt scheme over PDD might be.

  We now show the melt rates computed with pdd from the historic validation experiment with spatially resolved temperature fields (similar to Figure 1 and 2). We find that the pdd module, with standard parameters, gives a similar root mean square error over the historic period, both over the cumulated melt per year (39.85 Gt with pdd vs. 32.92 Gt with dEBM-simple) and over the spatial June, July and August melt (0.47 mWE/year with pdd vs. 0.36 mWE/year with dEBM simple). The pdd melt introduces a North-South gradient to the melt anomaly (compared to MAR data), overestimating the melt in the North and underestimating the melt in the South.

  We now also present the ice losses of the RCP2.6 and the RCP8.5 scenarios as computed with the standard pdd, finding that the mass loss computed with pdd is greater than with dEBM simple, in particular with the RCP8.5 scenario. Here we find a relative increase of 12% in the year 2100 and of 47% in the year 2300. This could be due to the fact that the melt factors in pdd are optimized for a present day melt rate and might be not valid in a future warming scenario. In pdd, the sensitivity of ice melt to T_eff is given by the degree day factor, usually assumed to be 8 mm liquid-water-equivalent / pos degree day. The temperature dependent melt of dEBM on the other hand scales with $\Delta t_{\Phi} / \Delta t * 1/(\rho * L_m) * c_1 \cong 4.37$ mm liquid-water-equivalent / pos degree day (if expressed in the same units. Thus, once the snow cover is gone pdd will react more sensitively to temperature changes.

In addition, increased ice losses are amplified via increasing temperatures via the atmospheric temperature lapse rate, as the melt-elevation feedback sets in.

- It would be interesting to see what is the relative importance of the parameters ta, c1 and c2 in Eq 1, (and s In Eq 3, is that a parameter or calculated from the daily temperatures over a month?) in their contribution the melt rate M would be. Could you show how that partitioning between the temperature-driven and the insolation driven melt looks like (a spatial map of sorts, or a stacked time series line plot)?

  We now include a map with average relative importance of temperature driven melt over the historic time period.

- For the RCPs the surface mass loss is expressed as SLE but the comparison with MAR is done as melt rates. I suggest you show the same quantity in Fig 1.

  Both plots show different quantities. In Fig. 1 we explicitly compare the melt rates of PISM-dEBM-simple to the MAR melt rates. Therefore, we think it does make sense to show the melt rates in Gt/year. This quantity can not be expressed in SLE in a meaningful way, since it does not include the full mass balance. In contrast, we show the total ice losses in Figure 4, which include the full climatic mass balance, basal mass balance, and discharge into the ocean. Even though we do not aim to provide projections, we still think that SLE (which is the accumulated mass change rate) is a useful unit to express these results by making them more comparable to previous results. We have added a conversion from mSLE to Gt in the figure captions.

- I'm not fully convinced that the RCP scenarios are helpful for the conclusions. Wouldn't you want to show that the new melt scheme is better than the old scheme in a controllable way? I understand that you get a bigger signal with RCP8.5 but that doesn't seem to be the point here. For instance, let's assume that PDD is the better approach, than whatever you show with an alternative model doesn't really matter. Of course, I'm pretty convinced that your melt scheme is superior. That's why I want you to make sure that this is the point of the paper and that you have demonstrated it.

  In fact proving that dEBM-simple is superior to pdd is not the point of the paper. The melt scheme is very comparable to the Krebs-Kanzow et al. (2018), only simplified by the parametrizations for albedo and atmospheric transmissivity. The above mentioned paper rigorously compares the dEBM to pdd and benchmarks both against MAR data.

  In this manuscript, after showing that the parametrizations for albedo and transmissivity and the implementation in PISM work as expected, we aim to apply the melt scheme in RCP scenarios, with the influence of the melt-albedo feedback being one of the main scientific questions.

  We make sure to refer to Krebs-Kanzow et al. (2018) more prominently, so that the interested reader can convince themselves about the differences in dEBM and PDD melt scheme. Further statistics about the performance of the full dEBM model are found in Fettweis et al. (2020). We make sure to cite this more prominently. We have included a comparison with PDD in Appendix D.

- In your current experimental setup, you can't be certain that the optimal parameters for the different linear fits, i.e., the individual terms as diagnosed from MAR (i.e., one for melt, one for transmissivity, etc, Fig. A1 and 2), would minimise your error with respect to the melt rates. I would therefore suggest that you do one single optimisation sweep. The other parameterisations are just means to combine different terms into a single equation (Eq. 1). For example, there is no point in minimising ta but would be still a useful diagnostic to check after your optimisation.

In this manuscript, we made a distinction between the dEBM-simple melt equation (Equation 1) and the parametrizations for the albedo and transmissivity fields. optimizing both individually.

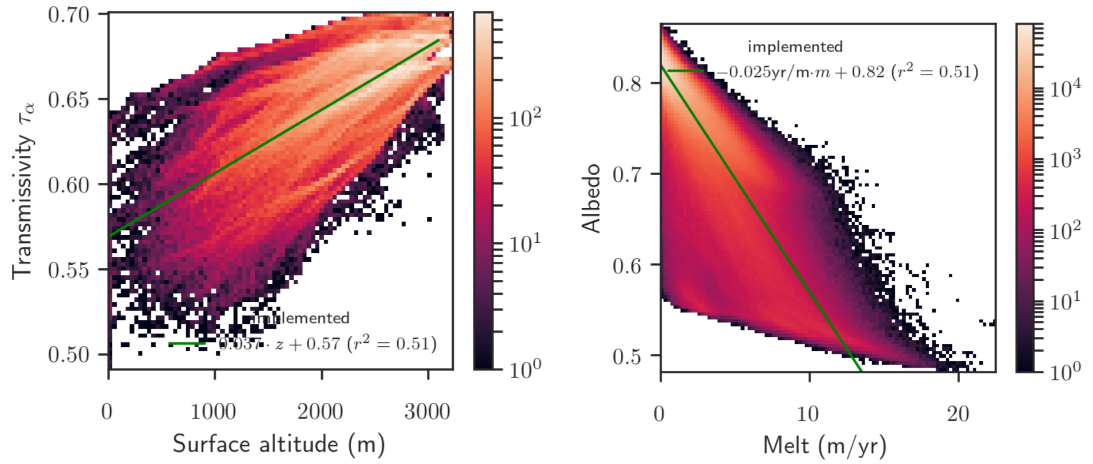
It is true that the total fit could be optimized by including the values chosen for the parametrization of albedo and transmissivity.

We selected the optimal values based on MAR data, but another set of parameters might require a different transmissivity or a different albedo parameterization, while keeping the other dEBM parameters constant.

We have included this point in the discussion of the results.

- The albedo-melt relationship (Fig. A1) and the transmissivity-altitude (Fig. A2) don't seem to be centred very much around the line that has been fitted. Furthermore, these relationships are not time-invariant, which means that the surface response differently to local climate, depending when and where it is. Maybe summer snowfall events play a role (as you said, they interrupt the whole melt process) and you can show that in the MAR data.

  Right, the temporal variation is not transformed into the parametrization. We here show all points from all grid cells on the ice sheet over 62 years for the months June, July and August individually. This visualization showcases the differences between those months. Moreover, it has been difficult to choose a transparency value, which both does not saturate where the density of markers is high and is still visible enough in a low density. However, if all data is taken together and visualized in a 2d-histograms with a logarithmic color scale, the visualization shows that the fit is more centered:



  Summer snowfall events might alter the albedo on a submonthly time scale, just as well as aging of snow. Other processes, which might increase the spread in albedo are shading, wind exposure or rain spells. We now discuss those in the manuscript.

- A thorough uncertainty assessment of your free model parameters is necessary, and, in my view, this would add credibility to your melt scheme and your paper. This can be done by randomly sampling from a multi-variate distribution whose mean is given by the optimised parameter set and whose scaling is given by the standard error thereof. I'm not asking for a full ensemble of coupled PISM simulations (that would be perfect, of course) but to analyse the variable space more systematically, e.g., in the (Te f f , Sf , M) hyper volume

Thank you very much for the suggestion. We have simulated an ensemble of 100 members by drawing the parameters for c_1, c_2, the slope for the albedo parameterization α and both, slope and intercept for the transmissivity parameterization τ. The volume changes under RCP 8.5 forcing are in between the upper and lower bound given by the experiments with albedo forcing (darkening and interannually constant yearly cycle respectively). These simulations are fully coupled to PISM.

The analysis of the ensemble allows us to identify the parameters which increase the temperature sensitivity of the ice sheet. While an ensemble of 100 members is too small to draw statistically sound conclusions, we can deduce that the dEBM parameter c_2 and the slope of the transmissivity parametrization do not seem to have a large effect on the variability of ice losses (with a Spearsman correlation coefficient of r = -0.06 and r = 0.25 respectively), in contrast the slope of the albedo and the intercept of the transmissivity τ seem to have a larger effect.

We have included this analysis in Appendix E and in the discussion.

In addition see Krebs-Kanzow et al., 2018 and Krebs-Kanzow et al. 2021 for further sensitivity analysis, e.g. for the sensitivity to the solar angle $\boldsymbol{\Phi}$.

- Summary statistics for fits (Fig. A1,A2) are missing (R2, standard errors or confidence intervals of slope and intercepts, etc.)

  Done, we added the coefficient of determination to the plots. The standard errors of the slope and the interval are not meaningful (i.e. zero) due to the large number of data points over which the fit is performed.
- Is it possible to compare the melt rate with observational (AWS) data and use AWS data as input, if they are available at all? e.g., PROMICE, GEUS

  The historic data in MAR is already tuned to AWS. As we want to compare the data to a full field, we think that MAR is more appropriate than extrapolating the individual points from AWS data.

Minor Comments

- P2/L13: ",where when large parts"

  done
- P2/L32: reference to PROMICE is missing

  We removed the direct reference to PROMICE, as suggested by the first reviewer Signe Hillup Larssen. The reference to the SMB-MIP remains, so the interested reader can read further.
- P3/L26: What is the present-day reference period?

  it is from 1971-1990, we added that to the manuscript
- P3/L29: "which is are modelled"

  done
- P3/L32: How is the snowfall determined from rainfalls and near-surface temperature?

  temperatures below 0°C lead to snow only, temperatures above 2°C lead to rain only, with linear interpolation in between
- P4/L1: Where does the "0.047" come from?

There was a typo, it's supposed to be 0.05193 m/yr and it's the default value in PISM. Compared to observed values, the default value used here is likely to underestimate the mass losses. However, at the resolution of 4.5km the ice losses through sub-shelf melting are not the main driver. The area of ice shelves in the simulation constitutes approximately 2000 km^2, compared to approximately 2*10^6 km^2 ice sheet area (0.1% of the ice area is floating ice in this simulation). Even in the RCP8.5 simulations, as the ice area decreases and the area of floating ice increases, it does not exceed 0.5% of the total ice area.

- Eq 1: the a in ta suggests link with albedo, maybe change the subscript to "a" or "A" for atmosphere
  done
- P5/L8: add "(TOA)" after "top of the atmosphere"
  done
- L15: It is unclear whether the cosine approximation is used here or the version by Berger (1978), which you refer to in A2. The Berger (1978) values are likely to differ from that present-day approximation.
  For the present day, the Liou (2002) expansion is used, for paleo simulations the Berger approximation is used.
- P5/LL1: I'm confused as to why the melt module is evaluated weekly if Te f f is monthly
  The melt is calculated as a function of insolation, atmospheric transmissivity, 2m air-temperature, and albedo. While it is true that the 2m air-temperature, an input field, has only monthly values, the insolation, the transmissivity and most importantly the albedo can change on a time scale shorter than one month. This is taken into account with the weekly evaluation of the melt scheme. In particular, as the albedo depends on the melt, frequent evaluation of the melt and the albedo reduces the error in albedo calculation while being more computationally fast compared to an iterative scheme in each, monthly, time-step.
  This line of thought is now reflected in the manuscript.
- how is refreezing calculated?
  It is a fixed value of 0.6, which means that 60% of the melt refreezes independently of space or time. Only snow melt can refreeze, it is assumed that ice melt runs off into the ocean before it can refreeze.
- P5/L6: Can you explain why ice/snow can melt below freezing point?
  If the monthly mean temperature is below freezing it does not mean that each day of this month would have a temperature below freezing. So there can be melt days in a month, even if the average temperature of this month is below freezing point.
  This is also the reasoning behind the positive degree day approach. Now Equation (1) uses for T_eff the same effective positive temperature as pdd would (see Eq. 3), which is always T_eff >0. But while the melt in pdd would simply approach 0 for small temperature, as the number of positive degree days goes to zero, the dEBM melt has an additional positive term coming from the insolation, so in contrast to a pdd scheme a lower threshold for melt is needed to avoid artificial insolation driven melt under very cold conditions. The cutoff at monthly mean temperatures below -6.5°C is in line with observations (see Krebs-Kanzow et al., 2018 and the references cited therein).
- P5/L29 and P6/LL1: There is something wrong with this sentence
  The sentence is changed based on comments of Signe.

- P6/L3-5: I understand that ”several iterations” mean that melt rate converges to some equilibrium value. Is that it?
  Yes, if the melt is evaluated under otherwise same conditions. See also the reasoning above, to why the melt rate should be evaluated weekly rather than monthly.
- ”Beckmann and Winckelmann” is a pre-print available for others to read? Otherwise, you need to explain the details.
  We have now included the statistics over the initial state in the supplementary information.
- P8/L22 ”as and example”
  done
- L25/26 there is a duplication of ”as described in Section 2.5”
  Removed the duplicated sentence.
- P11/L9 move the comma to after ”Here” in ”Here we analyse, how”
  done
- Fig. 4: The caption doesn’t say what the shading means (check the other figure captions, too)
  corrected in the manuscript
- P13/L1,2: Too many ”also”
  done
- P16/L27: Add comma after ”In dEBM”
  done
- P17/l1: remove ”classical”
  replaced with “widely-used”
- P18/L15: ”parameterized”
  done

Impact of the melt–albedo feedback on the future evolution of the Greenland Ice Sheet with PISM-dEBM-simple

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Abstract. Surface melting of the Greenland Ice Sheet contributes a large amount to current and future sea-level rise. Increased surface melt ~~, algae growth, debris, and dust deposition~~ may lower the reflectivity of the ~~ice~~ ice-sheet surface and thereby increase melt rates: the so-called melt–albedo feedback describes this ~~potentially~~ self-sustaining increase in surface melting. ~~Here we present~~ In order to test the effect of the melt-albedo feedback in a prognostic ice sheet model, we implement dEBM-simple, a simplified version of the diurnal Energy Balance Model ~~(dEBM-simple) which is implemented as a surface melt module~~ dEBM, in the Parallel Ice Sheet Model (PISM). ~~dEBM-simple is a modification of diurnal Energy Balance Model (dEBM), a surface melt scheme of intermediate complexity useful for simulations over centennial to multi-millennial timescales. dEBM-simple is computationally efficient, suitable for standalone ice-sheet modeling and~~ The implementation includes a simple representation of the melt–albedo feedback ~~. Using dEBM-simple and PISM~~and can thereby replace the positive degree day melt scheme. Using PISM-dEBM-simple, we find that this feedback increases ice loss until 2300 through surface warming by 60 % for the high-emission scenario RCP8.5 compared to a a scenario in which albedo remains constant at it's present day values. With an increase of 90 % compared to a fixed-albedo scenario, the effect is more pronounced for lower surface warming under RCP2.6. Furthermore, assuming an immediate darkening of the ice surface over all summer months, we estimate an upper bound for this effect to be +70% in the RCP8.5 scenario and a more than fourfold increase under RCP2.6. With dEBM-simple implemented in PISM, we find that the melt–albedo feedback is an essential contributor to mass loss in dynamic simulations of the Greenland Ice Sheet under future warming.

1 Introduction

The Greenland Ice Sheet is currently one of the main contributors to sea-level rise (Frederikse et al., 2020). Roughly 35% of the observed mass loss during the last 40 years is attributed to changes in surface mass balance and 65% of the mass loss is due to an increase in discharge fluxes (Mouginot et al., 2019). Overall, the contribution of changes in surface mass balance is expected to increase with ongoing warming (Shepherd et al., 2020).

Observations show that the surface of the Greenland Ice Sheet has been darkening over the last decades (He et al., 2013; Tedesco et al., 2016), and projections show that it is likely to darken further with increasing warming (Tedesco et al., 2016). Changes in albedo are driven by melt, the retreat of the snow line, black carbon, dust, and algae growth (Cook et al., 2020; Williamson et al., 2020; Box et al., 2012; Box, 2013; Box et al., 2017; Tedstone et al., 2017, 2020; Ryan et al., 2019). In particular the dark zone in the south-west of the Greenland Ice Sheet is subject to increased darkening, where ice-albedo values reach values as low as 0.27 due to surface water and impurities at the surface (for comparison, clean ice has typically an albedo between 0.45 and 0.55) (Ryan et al., 2019). As darker ~~snow and bare ice~~ surfaces absorb more radiation ~~, thereby increasing melt rates, these processes~~ than lighter surface, the effect of darkening due to increased melt could trigger a positive feedback mechanism: surface darkening increases melting, which in turn can lead to further darkening (Stroeve, 2001). ~~Studies~~ In addition to the darkening through melt, studies suggest a positive feedback mechanism between microbes, minerals, and melting, where algae-induced melting releases ice-bound dust, which in turn increases glacier algal blooms, leading to more melt (Di Mauro et al., 2020; McCutcheon et al., 2021). ~~This~~ The melt–albedo feedback is usually interrupted by winter snow accumulation and snow events in summer (Gardner and Sharp, 2010; Noël et al., 2015). In the light of recent extreme melting events as in 2010 (Tedesco et al., 2011), 2012 (Nghiem et al., 2012) or 2019 (Tedesco and Fettweis, 2020), ~~where~~ when large parts of the surface area were ~~covered by melt water~~ at melting point and therefore darker than usual, it is important to model the response of the ice sheet to such large-scale changes in albedo.

To assess the influence of the atmosphere on the surface mass balance of ice sheets a range of models is available and typically used: from process-based snow pack models coupled to regional climate models, which explicitly compute the regional climate and energy fluxes in the snow and at the ice surface (Fettweis et al., 2013, 2017; Noël et al., 2015; Langen et al., 2015; Niwano et al., 2018; Krapp et al., 2017), to simpler parameterizations like the positive degree day (PDD) models (e.g. Reeh, 1991). Regional climate models, for example, can be coupled with an ice-sheet model to compute interactions of the ice and the atmosphere while explicitly resolving all relevant feedbacks (Le clec'h et al., 2019). Since this is computationally expensive, often a simpler approach is required in order to run simulations over centuries to millennia or large ensembles of simulations. In such cases, the surface mass balance is typically calculated from near-surface air temperatures with a positive degree day approach (Wilton et al., 2017; Aschwanden et al., 2019; Rückamp et al., 2019), which is computationally much less expensive but lacks the direct contribution of shortwave radiation and albedo to melting. The insolation–temperature–melt equation ~~is one of the few examples, which~~ used by van den Berg et al. (2008) and Robinson et al. (2010) uses explicit albedo and insolation on long time scales~~(van den Berg et al., 2008; Robinson et al., 2010).~~ . The Surface Energy and Mass balance model of Intermediate Complexity (SEMIC) uses the explicit energy balance and albedo parameterization and an implicit diurnal cycle for the temperature (Krapp et al., 2017) and is therefore capable to predict present and future melt.

The recent development of the diurnal energy balance model dEBM presented by ~~Krebs-Kanzow et al. (2020)~~ Krebs-Kanzow et al. (2021 (with a simpler version introduced in Krebs-Kanzow et al. (2018)) is computationally efficient, works well for the Greenland Ice Sheet, and can represent melt contributions from changes in albedo as well as seasonal and latitudinal variations of the diurnal cycle. In the Greenland Surface Mass Balance Model Intercomparison Project (Goelzer et al., 2020), dEBM shows a good correlation with ~~ice core data, air-borne radar and PROMICE data and~~ observations and is among the models which compare the closest with observed integrated mass losses from 2003-2012 (Fettweis et al., 2020a). Thus it fills the gap between a process-based snow pack model coupled to a regional climate model and a simple and efficient temperature-index approach such as PDD.

~~To this end, we~~ We here expand the Parallel Ice Sheet Model (PISM) (the PISM authors, 2018; Winkelmann et al., 2011; Bueler and Brown, 2009) by the surface module dEBM-simple, which includes melt driven by changes in albedo based on Krebs-Kanzow et al. (2018), in order to explore their effects on the future ice evolution. ~~In addition, we~~ Beyond the work of Krebs-Kanzow et al. (2018), we additionally introduce parameterizations of albedo and atmospheric transmissivity to make it possible to run the model in standalone, prognostic mode (see Section 2). In particular the nonlinear albedo–melt relation (see Section 2.3.2) serves as an approximation to the melt–albedo feedback and allows to estimate its importance. ~~We~~ First, we compare the model against regional climate model simulations from ~~MAR~~ the Regional Atmosphere Model (Modèle Atmosphérique Régional, MAR, Fettweis et al. (2013, 2017) and find a good fit (Sect. 3). ~~We~~ In order to explore the minimal and maximal contribution of the melt–albedo feedback to future mass losses ~~and~~ we test the effect of albedo changes on future mass loss under RCP2.6 and RCP8.5 warming (Sect. 4). ~~The results~~ Here we distinguish between simulations, which do not allow for changes in albedo, simulations with adaptive albedo and darkening simulations, where the surface of the whole ice sheet is at the bare ice value. While the latter experiments are inspired by the large-scale melt events (see Sect. 4.3), the dark zone in Greenland serves as motivation to explore the influence of the ice-albedo value (see Sect. 4.4 and Appendix B). We compare dEBM simple with PDD and find a better performance for the historic period (Appendix D). A detailed comparison of dEBM and PDD can be found in (Krebs-Kanzow et al., 2018, 2021). The results considering future warming are discussed in Sect. 5.

2 Methods

We first present the Parallel Ice Sheet Model PISM, then describe the diurnal Energy Balance Model as introduced in Krebs-Kanzow et al. (2018) and its implementation in PISM. To be able to run the model in standalone, prognostic mode, we introduce parameterizations of the surface albedo and transmissivity in Sections 2.3.2 and 2.3.1 (see Appendix A for more detail). In the last two subsections, we describe the spin-up of PISM for the Greenland Ice Sheet and the experiments conducted in the next sections.

2.1 Ice-sheet model PISM

PISM is a thermomechanically coupled ice-sheet model which uses a superposition of the shallow-ice approximation (SIA) for slow-flowing ice, and the shallow shelf approximation (SSA) for fast-flowing ice streams and ice shelves (Bueler and Brown, 2009; Winkelmann et al., 2011; the PISM authors, 2018). PISM was shown to be capable of reproducing the complex flow patterns evident in Greenland's outlet glaciers at high resolution of less than 1km (Aschwanden et al., 2016).

The SSA basal sliding velocities are related to basal shear stress via a power law with a Mohr–Coulomb criterion that relates the yield stress to parameterized till material properties and the effective pressure of the overlaying ice on the saturated till (Bueler and Pelt, 2015). We use a non-conserving simple hydrology model that connects the till water content to the basal melt rate (Bueler and van Pelt, 2015). The internal deformation of the ice is described by *Glen's flow law* with the flow exponent $n = 3$ for both, SIA and SSA flow and with the enhancement factors $E_{\text{SSA}} = 1$ and $E_{\text{SIA}} = 1.5$ for SSA and SIA flow respectively.

Using PISM, we first create an initial configuration of the Greenland Ice Sheet under present-day climate conditions, using a climatology averaged over 1971-1990. Then we run a suite of experiments to validate dEBM simple for present-day as well as to test the role of insolation and temperature melting in future warming scenarios.

In this manuscript we concentrate on the changes in the surface mass balance, which ~~is~~ are modelled using the newly implemented dEBM-simple. The atmospheric conditions, namely the monthly 2D temperature and precipitation fields, are read in as input fields. The precipitation fields remain fixed and the share of snowfall and rain is determined from the local near-surface air temperature, with rain at temperatures above 2°C and snow at temperatures below 0°C. We neglect effects from changing ocean temperatures, thus the sub-shelf melting is constant in space and time at ~~0.047~~0.05193 m/yr. Also, calving is not modeled explicitly, but induced by a fixed calving front at its present-day location based on Morlighem et al. (2017), thus changes in mass losses from ice-ocean interaction are not considered here. Isostatic adjustment of the bedrock is not considered here. All experiments were run on a 4.5 km horizontal grid with a constant vertical resolution of 16 m. While this resolution is too low to reproduce the details of the outlet glacier flow, it still preserves the general flow pattern. Moreover the focus of the paper on climatic mass balance justifies this choice.

2.2 Adapted diurnal energy balance model dEBM-simple

For the simple version of the diurnal energy balance model (dEBM-simple) we follow the parametrization as laid out by Krebs-Kanzow et al. (2018). The melt equation reads

$$M = \frac{\Delta t_\Phi}{\Delta t \rho_w L_m}\left(\tau_{\cancel{\alpha}\mathrm{A}}(1-\alpha_s)\bar{S_\Phi} + c_1 T_{\text{eff}} + c_2\right), \tag{1}$$

with fresh water density ρ_w, latent heat of fusion L_m ~~,~~ and the surface albedo α_s~~, atmospheric transmissivity τ_α. The parameters describing the effective temperature influence on melting c_1 and the melt intercept c_2 are estimated from MAR v3.11 simulations (Fettweis et al., 2017), see Section 3. The values used here are given in Table A2.~~ . ~~dEBM~~ dEBM-simple is based on the assumption that melting occurs only during daytime, when the sun is above an elevation angle Φ, which is estimated to be constant in space and time $\Phi = 17.5°$, see (Krebs-Kanzow et al., 2018). The time period of a day when the sun is above the elevation angle Φ is denoted by Δt_Φ. The length of a day is Δt and the fractional time that the sun is above the elevation angle Φ is given by:

$$\frac{\Delta t_\Phi}{\Delta t} = \frac{h_\Phi}{\pi} = \frac{1}{\pi}\frac{\sin\Phi - \sin\varphi\sin\delta}{\cos\varphi\cos\delta} \tag{2}$$

with h_Φ being the hour angle when the sun has an elevation angle of at least Φ, δ being the solar declination angle and φ the latitude. The incoming radiation over the time Δt_Φ~~(when the sun is above the declination angle Φ) drives the melting. It is~~ , obtained from the insolation at the top of the atmosphere (TOA), $\bar{S}_\Phi$, and the parameterized transmissivity ~~τ_α~~ τ_A (see equation (6)), drives the insolation melt described in the first term of the Equation (1).

~~In addition, the~~ The temperature-dependent melting ~~is not~~ described in the second term of Equation (1) is not simply a function of the air temperature directly as in, e.g., Pellicciotti et al. (2005); van den Berg et al. (2008), but a function of the cumulative temperature exceeding the melting point in a given month T_{eff} as in ~~Krebs-Kanzow et al. (2018, 2020)~~ Krebs-Kanzow et al. (2018, 2021).

$$T_{\text{eff}}(\bar{T},\sigma) = \frac{1}{\sigma\sqrt{2\pi}} \int_0^\infty \mathrm{d}T\, T \exp\left(-\frac{(T-\bar{T})^2}{2\sigma^2}\right) \tag{3}$$

Here, T is the fluctuating daily temperature, $\bar{T}$ is the monthly average temperature, which is used as an input, and σ is the standard deviation of the temperature. The melting point is at $0\,^\circ\mathrm{C}$.

The parameters describing the effective temperature influence on melting c_1 and the melt intercept c_2 are estimated from MAR v3.11 simulations (Fettweis et al., 2017), see Section 3. The values used here are given in Table A2.

Refreezing is assumed to be constant, with 60% of snow melt refreezing independent of temperature or melt. Ice melt does not refreeze but immediately runs off.

2.3 Implementation of dEBM-simple in PISM

The diurnal energy balance model is implemented in PISM as a climatic mass balance module. It takes the local near-surface air temperature and the precipitation as an input and computes the local climatic mass balance. The shortwave downward radiation and the broadband albedo are not needed as inputs, as they are parameterized internally, with the possibilty to use other orbital parameters than the present day values. The melt module is evaluated at least weekly, independent of the adaptive time step used for the ice dynamics in PISM. The amount of melt is balanced with refreezing and snowfall before the surface mass balance is aggregated over the adaptive time steps in PISM. The aggregated values feed into the update of the ice geometry. On the other hand, the ice geometry is used as an input for the dEBM melt, as it feeds into the parameterization of the atmospheric transmissivity (see Equation (6)) and updates the local temperature when the atmospheric temperature lapse rate is considered. If a run is started without knowledge about melting in the previous time step, the albedo is assumed to be at the fresh snow value everywhere on the ice sheet or can be read in from an input file. In line with Krebs-Kanzow et al. (2018), no melting is allowed below $-6.5\,^\circ\mathrm{C}$ even if the insolation alone would be sufficient to cause melting.

2.3.1 Parametrization of shortwave downward radiation

The shortwave downward radiation is computed from the top of the atmosphere (TOA) insolation and a linear model of the transmissivity of the atmosphere. Daily average TOA insolation $\bar{Q}^{\text{day}}$ is computed from

$$\bar{Q}^{\text{day}} = \frac{S_0}{\pi}\left(\frac{\bar{d}}{d}\right)^2 (h_0 \sin\varphi \sin\delta + \cos\varphi\cos\delta\sin h_0), \tag{4}$$

where $S_0 = 1367\,\mathrm{W/m^2}$ is the annual mean of the total solar irradiance (solar constant), $\bar{d}$ is the annual mean distance of the earth to the sun, d is the current distance, φ is the latitude, δ is the declination angle, and h_0 is the hour angle of sunrise and sunset. The average TOA insolation during the daily melt period $\bar{S}_\Phi$ is given by

$$\bar{S_\Phi} = \frac{S_0}{\Delta t_\Phi \pi}\left(\frac{\bar{d}}{d}\right)^2 (h_\Phi \sin\varphi \sin\delta + \cos\varphi\cos\delta\sin h_\Phi) \tag{5}$$

Under present day conditions the declination angle δ and the sun-earth distance d are approximated with trigonometric expansions depending on the day of the year, see Liou (2002, Chapter 2.2.). This approximation is used, as long as the used does not specifically demand paleo simulations.

To scale the insolation to the ice surface, we assume that the transmissivity of the atmosphere depends only on the local surface altitude in a linear way (similar to Robinson et al. (2010)). The linear fit for the shortwave downward radiation from TOA insolation is obtained from a linear regression of MAR v3.11 data averaged over the years 1958 to 2019, considering only the summer months (June, July and August~~from 1958 to 2019~~) (see Appendix A2 and Figure A3). The ~~transmissivity is~~ parametrization relies on the assumption that the mean transmissivity does not change in a changing climate. In particular the impact of cloud conditions and events like e.g. Greenland blocking, which might become more frequent in future, is not captured with this approach. The transmissivity is given by:

$$\tau_{\cancel{\alpha}\mathrm{A}} = a + b\cdot z, \tag{6}$$

where a and b are parameters (here $a = 0.57$ and $b = 0.037\mathrm{km}^{-1}$) and z is the surface altitude in km. The approach to calculate shortwave downward radiation is further described in Appendix A2, in particular it is described how TOA conditions different than present-day, e.g. during the Eemian, can be modeled.

2.3.2 Albedo parametrization

PISM-dEBM-simple allows to read in the albedo field as an input. However, in order to keep the input requirements for a standalone version of the model minimal and to allow for a melt-dependent albedo, a simple albedo parametrization is implemented. ~~Regional climate models such as MAR or RACMO use~~ Snow albedo in MAR is calculated using a snowpack model~~to determine the albedo (van Dalum et al., 2020). Snow albedo there is explicitly calculated~~ , explicitly based on snow grain size, cloud optical thickness, solar zenith angle and impurity concentration in snow ~~or parameterized. Ice albedo there can be~~ (van Dalum et al., 2020). In MAR, ice albedo is explicitly calculated as a function of ice density, time of the day, solar angle, spectrum of the solar radiation etc. Here, in contrast, albedo is paramaterized in an ad-hoc way as a function of the melt in the last (weekly) timestep. As the time step in the climatic mass balance module is typically smaller than the adaptive ice-dynamics time-step, and the temporal resolution of the 2d air-temperature input, this approach allows for several iterations of the melt-dependent albedo under otherwise same conditions.

The corresponding parameters are fitted using ~~MAR data~~ MARv3.11 data (Fettweis). The advantage of this approach is that it requires no further information in PISM (e.g., a fully resolved firn-layer) but still ~~allows us to capture melt processes~~ captures melt processes driven by changes in albedo or insolation. In this approach, the albedo decreases linearly with increasing melt from the maximal value $\alpha_{\max} = 0.81$ (close to the fresh-snow albedo) for regions with no melting to $\alpha_{\min} = 0.47$ (close to the bare-ice albedo). The albedo cannot drop below the value of $\alpha_{\min}$.

$$\alpha_s = \max[\alpha_{\max} - c \cdot M, \alpha_{\min}] \tag{7}$$

The slope $c = -0.025$ yr/m is estimated from MAR data (see appendix Figure A1 and Section A1). We will later on test the sensitivity of the melting to the ~~parameters.~~ slope and the value of $\alpha_{\min}$. Lowering the value of $\alpha_{\min}$ may indicate the sensitivity to darker ice. While an explicit darkening of the ice, possibly with a different albedo to melt relation, is not captured in this framework, it can be easily expanded to incorporate darkening ice.

For comparison, the observed albedos are shown in Figure A2.

2.4 Initial state

All simulations are started from a spun-up state and run with full ice dynamics (SIA and SSA as well as temperature evolution and a thermomechanical coupling). ~~For the spin-up the~~ The procedure detailed in Aschwanden et al. (2019) is used for the spin-up: a temperature anomaly is applied over the last 125 ka to the climatological mean (1971 to 1990 monthly averages) of the 2D temperature field of MAR v3.9 in order to obtain a realistic temperature distribution within the ice ~~.~~ while the topography is allowed to evolve. ~~Ice geometry and bedrock topography are from BedMachine V3 (Morlighem et al., 2017). Basal heat flux is obtained from Maule (2005)~~During the spin-up the more conventional positive-degree day model is used to compute changes in climatic mass balance. In the simulations, surface temperatures are scaled with changes surface elevation (atmospheric temperature lapse rate of -6 K/km) to include the melt–elevation feedback in the simulations. ~~Precipitation is not scaled with changing temperature. The spin-up is performed using the more conventional positive-degree day model~~

Initial ice geometry and bedrock topography are taken from BedMachine V3 (Morlighem et al., 2017). Basal heat flux is obtained from Maule (2005). The yearly cycle of precipitation is kept fixed but during the spin-up the precipitation fields are scaled: for each degree of warming we apply 7.3% precipitation increase for each degree of surface warming (Huybrechts, 2002)
.

~~The thickness and~~ The root mean square error in thickness amounts to 237 m, overestimating the thickness values in the West and North West (Morlighem et al., 2017). The velocity anomalies of the initial state ~~at the end of the spin-up are detailed in Beckmann and Winkelmann (submitted).~~ show a root mean square error of 145 m/yr compared to observed data (Rignot and Mouginot, 2012). The North East Greenland Ice Stream and several other fast flowing outlet glaciers are underestimated in the surface velocities. See Supplementary Figure S1 for anomaly maps.

2.5 ~~Calibration~~ Validation experiments

To calibrate the model parameters and test the parameterizations, we perform diagnostic experiments with PISM-dEBM-simple over the period between 1958 and 2019. In order to disentangle the surface module from indirect effects of ice dynamics, e.g. dynamic thinning and thus a temperature increase through the temperature lapse rate, changes in the ice topography are suppressed in these diagnostic experiments. Monthly MAR v3.11 near-surface air temperature and precipitation fields from 1958 to 2019 are used as atmosphere input while the albedo, the transmissivity and the melt rate are computed as shown above.

~~The sensitivity to~~ In order to explore the sensitivity to insolation, Eemian insolation values ~~is tested~~ are used in an analogous experiment where only the orbital parameters, which determine the top of the atmosphere insolation, are changed. The temperature and precipitation input remains the same as described above. Precipitaiton and albedo are calculated using the respecitve parametrizations.

2.6 Warming and darkening experiments over the next centuries

Here, we describe the series of experiments which are performed to assess the impact of the melt–albedo feedback onto the mass losses of the Greenland Ice Sheet. All experiments start from the same initial state, described in Sect. 2.4 and run over the period from 2000 to 2300. In contrast to the previously described experiments in Section 2.5, the ice topography is now allowed to change and the results are expressed in terms of cumulated mass losses since 2000 in meters sea level equivalent (m SLE). Melt rates are calculated by PISM-dEBM-simple using periodic monthly temperature fields given by the climatological mean ~~from~~ over the period 1971 to ~~1990. The~~ 1990 from the regional climate model MAR v3.11. MAR was forced with ERA reanalysis data (ERA-40 from 1958-1978 and ERA-5 after) (Kittel et al., 2021; Fettweis et al., 2020b). In the warming experiments, scalar temperature anomalies are applied uniformly over the entire ice sheet. The temperature anomalies are obtained from averaging the output of the IPSL-CM5A-LR model (Dufresne et al., 2013), which is one of four CMIP5 models extended until 2300, over the simulation domain containing the Greenland Ice Sheet and computing the anomaly relative to the 1971-1990 period over the same domain.

In addition to forced temperature changes, the local near-surface air temperature adapts to topography changes of the ice sheet with a lapse rate of -6 K / km, thus taking the melt–elevation feedback into account. Note that in all experiments changes in albedo do not feed back onto the atmosphere, in particular albedo changes do not affect near-surface air temperatures.

~~We perform the following experiments:~~

- ~~**Ctrl:** Control run without darkening or temperature anomalies, using the parameters from Table A2 consistent with the present day experiments presented in Section 3.~~

- ~~**RCP2.6** and **RCP8.5:** Similar to Ctrl, with a scalar temperature anomaly from the RCP2.6 and the RCP8.5 scenarios of the IPSL-CM5A-LR model (Dufresne et al., 2013), respectively, which is one of four CMIP5 models extended until 2300. The scalar temperature anomaly was obtained by averaging the monthly near-surface temperature over the Greenland Ice Sheet and computing the anomalies to the average yearly cycle of 1971-1990 over the same spatial domain.~~

- ~~**Ctrl** α_{1990}, **RCP2.6** α_{1990}, and **RCP8.5** α_{1990}: Control and warming experiments with a fixed yearly cycle of albedo values instead of the melt-dependent parameterization. The 2 dimensional albedo fields are obtained from monthly averages from the MAR v3.11 data, averaged over the period before 1990. Scalar temperature anomalies as detailed above in RCP2.6, RCP8.5.~~

- ~~**Ctrl** $\alpha_{\mathbf{dark}}$, **RCP2.6** $\alpha_{\mathbf{dark}}$, and **RCP8.5** $\alpha_{\mathbf{dark}}$: In these experiments the albedo over the entire ice sheet is lowered to the bare ice value of 0.47 during the summer months June, July and August. During all other months the albedo parametrization (with the optimized values as described in Appendix A1) is used to compute the albedo. Scalar temperature anomalies as detailed above in RCP2.6, RCP8.5.~~

- ~~**Ctrl** $\alpha_{\mathbf{ls}}$, **RCP2.6** $\alpha_{\mathbf{ls}}$, and **RCP8.5** $\alpha_{\mathbf{ls}}$: The slope of the albedo parametrization is decreased to half the optimal value (which correspond to -0.013 yr/m). Thus the bare ice value of the albedo is reached at melt rates of 28 m/yr, (compared to 14 m/yr with the optimal value for the slope). Scalar temperature forcing for the RCP experiments as described above.~~

- ~~**Ctrl** $\alpha_{\mathbf{hs}}$, **RCP2.6** $\alpha_{\mathbf{hs}}$, and **RCP8.5** $\alpha_{\mathbf{hs}}$: The slope of the albedo parametrization is increased to twice the optimal value (which correspond to -0.05 yr/m). Thus the bare ice value of the albedo is reached at melt rates of 7 m/yr, (compared to 14 m/yr with the optimal value for the slope). Scalar temperature forcing for the RCP experiments as described above.~~

- ~~**Ctrl** $\alpha_{\mathbf{min}}$**x**, **RCP2.6** $\alpha_{\mathbf{min}}$**x**, and **RCP8.5** $\alpha_{\mathbf{min}}$**x**: The lower limit for albedo values is reduced from the standard value of 0.47 to x=0.4 and x=0.3. The slope of the albedo parametrization remains unchanged, but albedo values decrease further for melt rates above 14 m/yr until the new minimal albedo value is reached. Scalar temperature forcing for the RCP experiments as detailed above.~~

- ~~**Ctrl** $T_{\mathbf{noLR}}$, **RCP2.6** $T_{\mathbf{noLR}}$, and **RCP8.5** $T_{\mathbf{noLR}}$: Experiments analogous to Ctrl, RCP2.6 and RCP8.5, but neglecting the atmospheric temperature lapse rate. Thus the local temperature is independent of the ice-sheet topography and the representation of the melt–elevation feedback is interrupted.~~

The ~~α_{1990} experiments use a constant~~ experiments can be summarized into seven groups: The $\alpha_{\mathbf{1990}}$ experiments use an interannually constant yearly albedo cycle, and therefore suppress the adaptation of albedo to increased melt rates under warming. They are used to estimate future ice loss without the melt–albedo feedback ~~. The~~ and serve as a reference. The **std** experiments include the melt-albedo feedback through the standard parameterization for albedo. The α_{dark} experiments represent an extreme scenario, assuming that the whole ice surface will be snow free or covered with melt water during the months June, July and August in each year. This is not a realistic scenario but rather an upper limit to the possible impact of albedo changes on ice losses. The ~~, and~~ $\alpha_{\mathbf{ls}}$, $\alpha_{\mathbf{hs}}$ and $\alpha_{\mathbf{min}}$ experiments explore the uncertainty from the albedo parameterization. Doubling the slope to -0.05 yr/m leads to a steeper decrease in albedo with increasing melt rates, which is closer to the conditions in August (see Appendix Figure A1). Assuming that in the future the melting period over the Greenland Ice Sheet is longer and therefore the conditions we observe in August might be more characteristic over the melting period, justifies to explore the impact of an increased sensitivity in the $\alpha_{\mathbf{hs}}$ experiments. However, in this approach the minimal albedo for bare ice remains at 0.47, and is therefore reached with melt rates of 7 m/yr (instead of 14 m/yr in the standard parameterization). Halving the slope to -0.013 yr/m explores the lower boundary of albedo-melt sensitivity (see Figure A1) in the $\alpha_{\mathbf{ls}}$ experiments. In the $\alpha_{\mathbf{min}}$ experiments, we test the influence of a reduced minimum albedo, as observed today in the dark zone of the Greenland

Table 1. Overview over the experiments performed in this study

Experiment group	Name	Temperature forcing	Albedo	Lapse rate
$\boldsymbol{\alpha_{1990}}$	Ctrlα_{1990}	none	fixed yearly cycle	-6K/km
	RCP2.6α_{1990}	RCP2.6	fixed yearly cycle	-6K/km
	RCP8.5α_{1990}	RCP8.5	fixed yearly cycle	-6K/km
std	Ctrl	none	std parameterized	-6K/km
	RCP2.6	RCP2.6	std parameterized	-6K/km
	RCP8.5	RCP8.5	std parameterized	-6K/km
$\boldsymbol{\alpha_{dark}}$	Ctrlα_{dark}	none	bare ice value	-6K/km
	RCP2.6α_{dark}	RCP2.6	bare ice value	-6K/km
	RCP8.5α_{dark}	RCP8.5	bare ice value	-6K/km
$\boldsymbol{\alpha_{ls}}$	Ctrlα_{ls}	none	parameterized, low slope	-6K/km
	RCP2.6α_{ls}	RCP2.6	parameterized, low slope	-6K/km
	RCP8.5α_{ls}	RCP8.5	parameterized, low slope	-6K/km
$\boldsymbol{\alpha_{hs}}$	Ctrlα_{hs}	none	parameterized, high slope	-6K/km
	RCP2.6α_{hs}	RCP2.6	parameterized, high slope	-6K/km
	RCP8.5α_{hs}	RCP8.5	parameterized, high slope	-6K/km
$\boldsymbol{\alpha_{min}X}$	Ctrlα_{min}X	none	parameterized, changed ice albedo	-6K/km
	RCP2.6α_{min}X	RCP2.6	parameterized, changed ice albedo	-6K/km
	RCP8.5α_{min}X	RCP8.5	parameterized, changed ice albedo	-6K/km
$\boldsymbol{T_{noLR}}$	CtrlT_{noLR}	none	std parameterized	none
	RCP2.6T_{noLR}	RCP2.6	std parameterized	none
	RCP8.5T_{noLR}	RCP8.5	std parameterized	none

Ice Sheet. The $\boldsymbol{T_{noLR}}$ experiments neglect the atmospheric temperature lapse rate. Thus the local temperature is independent of the ice-sheet topography and the representation of the melt–elevation feedback is interrupted.

An overview over all experiments is given in Table 1.

3 Validation for the Greenland Ice Sheet

In this section, we validate the dEBM-simple melt for present-day conditions and show as ~~and~~ an example melt rates with Eemian insolation. The experiments are performed as described in the Methods Section 2.5.

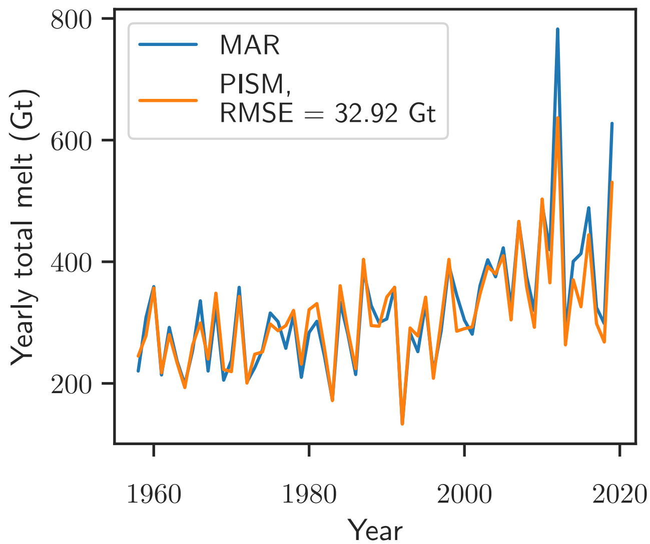

Figure 1. Comparison of annual total melt of the Greenland Ice Sheet as calculated with MAR v3.11 and PISM-dEBM-simple. The diagnostic simulation with PISM-dEBM-simple (orange line) is performed using monthly MAR 2D temperature fields as forcing. The albedo α_s is parameterized with the local melt rate m as $\alpha_s = 0.82 - 0.025\,\mathrm{yr/m} \cdot m$ and the shortwave downward radiation is approximated by the top of the atmosphere radiations and the transmissivity $\cancel{\tau_\alpha}$ τ_A parameterized with surface altitude z as $\cancel{\tau_\alpha = 0.037\,\mathrm{km}^{-1} \cdot z + 0.57}$ $\tau_\mathrm{A} = 0.037\,\mathrm{km}^{-1} \cdot z + 0.57$. The root mean square difference between the PISM-dEBM-simple simulation and total melt as given by MAR (blue line) is 32.92 Gt.

3.1 Present-day melt rates

~~As described in Section 2.5, we use historic and present day experiments to optimize for the melt parameters in equation . We compare the resulting modeled melt~~ Here, we compare the melt modeled with PISM-dEBM-simple over the historic period with ~~melting~~ melt modeled by MARv3.11 ~~in Figure 1. The experimental~~ (see Figure 1). The setup is described in Section 2.5. Note that here the evolution of the ice-sheet topography is suppressed, that is the melt rates are calculated over a fixed geometry corresponding to present day.

Due to the parameterization of the albedo and the transmissivity (and thus the shortwave downward radiation) detailed in Section 2.3 and in Table A2, the parameters of the dEBM-simple model are adjusted from Krebs-Kanzow et al. (2018). The chosen dEBM-simple parameters c_1 and c_2 (see Table A2) minimize the product of spatial and temporal root mean square error in the melt rate over the whole period from 1958 to 2019 while using the parametrizations for albedo and shortwave downward radiation. The temporal root mean square error is computed from a comparison of total yearly melt (see Figure 1) and the spatial root mean square error is computed from a comparison of the 2D fields of summer (JJA) melt rates averaged over the whole period (see Figure 2)~~.~~, both with respect to MAR data. Both of the root mean square errors are minimized by the dEBM-simple parameters c_1 and c_2 given in table A2.

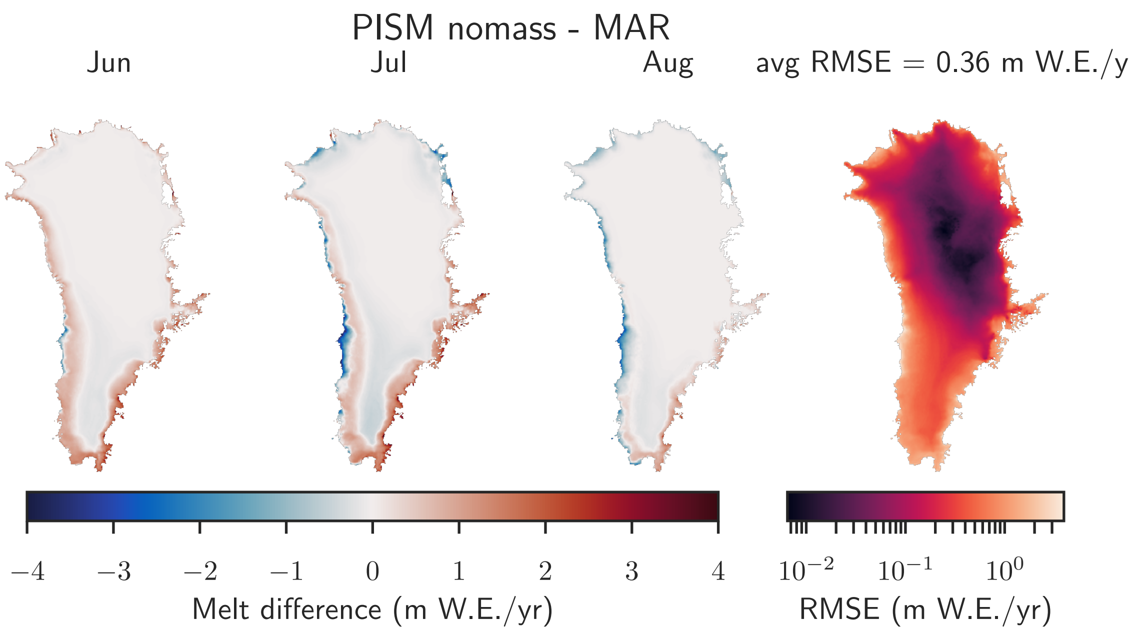

Figure 2. Local differences between the monthly averaged June, July, and August melt rates as diagnosed with PISM-dEBM-simple compared to MAR. The PISM simulation uses monthly 2D temperature fields from MAR as forcing and parametrizes albedo and shortwave downward radiation as detailed in 2.3.1 and 2.3.2. Positive numbers mean that PISM overestimates the melt and negative numbers mean that PISM underestimates the melt. The local root mean square error averaged over June, July and August from 1958-2019 is shown in the right plot. The spatial average of the RMSE is 0.36 m/yr.

Yearly total melt computed with ~~PISM~~ PISM-dEBM-simple follows the MAR data closely (see Figure 1). That extreme melt years such as 2012 and 2019 are underestimated can be explained by the parametrization of shortwave downward radiation, which neglects temporal variability in the cloud cover, one of the drivers of recent mass loss in Greenland (Hanna et al., 2014; Hofer et al., 2017). We also test dEBM-simple with shortwave downward radiation and albedo from MAR directly. Figure A4 shows, that in this case the extreme melt in 2019 is better captured, while melting in 2012 is still underestimated.

As Figure 2 shows, melt is generally overestimated in June, at the beginning of the melt season and underestimated as the melt season progresses. In July dEBM-simple underestimates melt mostly at the western margin, where ablation is highest. In August, toward the end of the melt season, melt is systematically underestimated by the dEBM-simple module, in particular in the regions where the darkest albedo values are observed. The systematic underestimation could be caused by taking a constant minimal value for the ice albedo and not allowing for processes which would lead to a natural darkening of the surface, i.e. algae growths, supra-glacial lakes or darkenig related to the ageing of snow or exposed ice. On the other hand, many of those processes, in particular bio-albedo feedbacks or dust deposition, are not yet represented in the MAR model neither, and thus should not induce a systematic bias when comparing to MAR data.

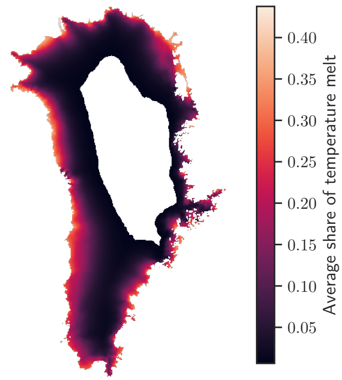

Figure 3. Share of temperature-induced melt. The fraction of tempeature-induced melt, defined as $M_t/(M_t + M_i)$, is diagnosed with PISM-dEBM-simple and averaged over the months June, July and August over the whole simulation period from 1958-2019. The white part in the center illustrates areas of the Greenland Ice Sheet where the average melt is zero at present.

The melt equation (1) can be used to attribute the melt rates of the present day to temperature- or insolation driven melt in a first order approximation. To this end we compare both contributions to the total melt rate, with the temperature driven melt $M_t = \frac{\Delta t_\Phi}{\Delta t \rho_w L_m} c_1 T_{\text{eff}}$ and the insolation driven melt $M_i = \frac{\Delta t_\Phi}{\Delta t \rho_w L_m} \tau_\alpha (1 - \alpha_s) \bar{S_\Phi}$. The share is then defined by $M_t/(M_t + M_i) = (c_1 T_{\text{eff}})/(c_1 T_{\text{eff}} + \tau_\alpha (1 - \alpha_s) \bar{S_\Phi})$ over the regions which experience melt. Under present day conditions, this approach indicates that the melt over the whole ice-sheet is mainly driven by the insolation (see Figure 3). Even at the margins, where monthly mean temperatures and the fraction of temperature driven melt are highest, the fraction does not exceed one half. In particular over the high and cold regions of the ice sheet, the melt seems to be entirely driven by the insolation, with an indirect effect of the temperature only allowing melt if monthly mean air temperatures are above -6.5°C.

The model is able to capture melt patterns of the Greenland ice sheet over the historic period between 1958 - 2019 with a root mean square error of 32.92 Gt for the yearly total melt and an average root mean square error of 0.36 m W.E./yr for the local summer melt rates. A more thorough discussion on the performance and the sensitivity of the melt equation (1) (without the parametrization of albedo and transmissivity) and a comparison to the positive degree day model can also be found in (Krebs-Kanzow et al., 2018). An overview over the performance of the full dEBM model compared with other state of the art models can be found in (Fettweis et al., 2020a).

Overall, the skill of the PISM-dEBM-simple model under present-day conditions and using high resolution forcing from MAR is similar to the skill of the ~~dEBM model (Krebs-Kanzow et al., 2020)~~full dEBM model (Krebs-Kanzow et al., 2021). Compared to MAR, dEBM revealed a RMSE of 27Gt for the annual mean 1979-2016 SMB in an experiment which was forced with reanalysis data (Krebs-Kanzow, pers. com.). ~~However dEBM better reproduces extreme years such as 2012 as this version also~~ The dEBM full model accounts for changes in the atmospheric emissivity and transmissivity, both caused by changes in

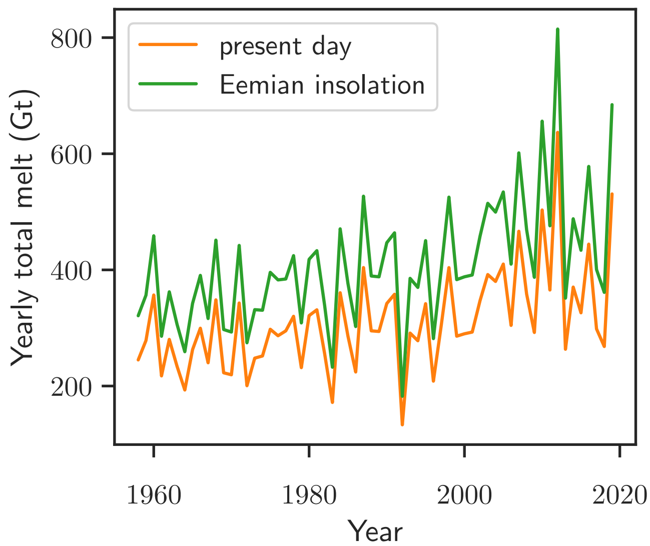

Figure 4. Comparison of Eemian vs. present-day insolation in dEBM simple Yearly total melt of the Greenland Ice Sheet as diagnosed with PISM-dEBM-simple under present day insolation (orange) and Eemian insolation (green). The diagnostic simulation with PISM were performed using monthly MAR 2D temperature fields as forcing and the parameterizations for shortwave downward radiation and albedo mentioned in the text.

cloud cover. As the cloud cover ~~, as drivers of melt , and is therefore~~ was the main driver in the 2012 melt event, the full dEBM model therefore is better suited to ~~represent melt-events such as in 2012.~~ reproduce this and similar melt events. Furthermore dEBM computes the refreezing on the basis of the surface energy balance.

3.2 Sensitivity to Eemian solar radiation

The dEBM approach together with the approximations for albedo and transmissivity allows to include changing orbital parameters for simulations on paleo timescales. Here we ~~use as an example the orbital parameters corresponding to the Eemian ,~~ explore the melt response to Eemian (125 ka before present~~, with the~~) orbital parameters in order to test for the sensitivity to insolation values and compare with other results from the literature. Therefore we use the eccentricity $e = 0.0400$, the obliquity $\varepsilon = 23.79°$ and the longitude of the perihelion $\omega = 307.13°$. The insolation at the top of the atmosphere is then calculated as detailed in Section A1. We use the present day topography for the diagnosis of melt rates and keep the surface air temperature fields unchanged from the previous experiment (MAR v3.11 in the period of 1958-2019).

The increase in solar radiation leads to increased melt, as seen in Figure 4. The inter-annual variability in yearly total melt is very close to the ~~historic variability~~ present day variability computed with MAR, mainly driven by inter-annual temperature changes. Averaged over the whole time period (1958-2019) the yearly total melt increases by 98 Gt/yr, which corresponds to a relative increase of 31%. This is in line with findings of Van De Berg et al. (2011), who find that Eemian insolation alone

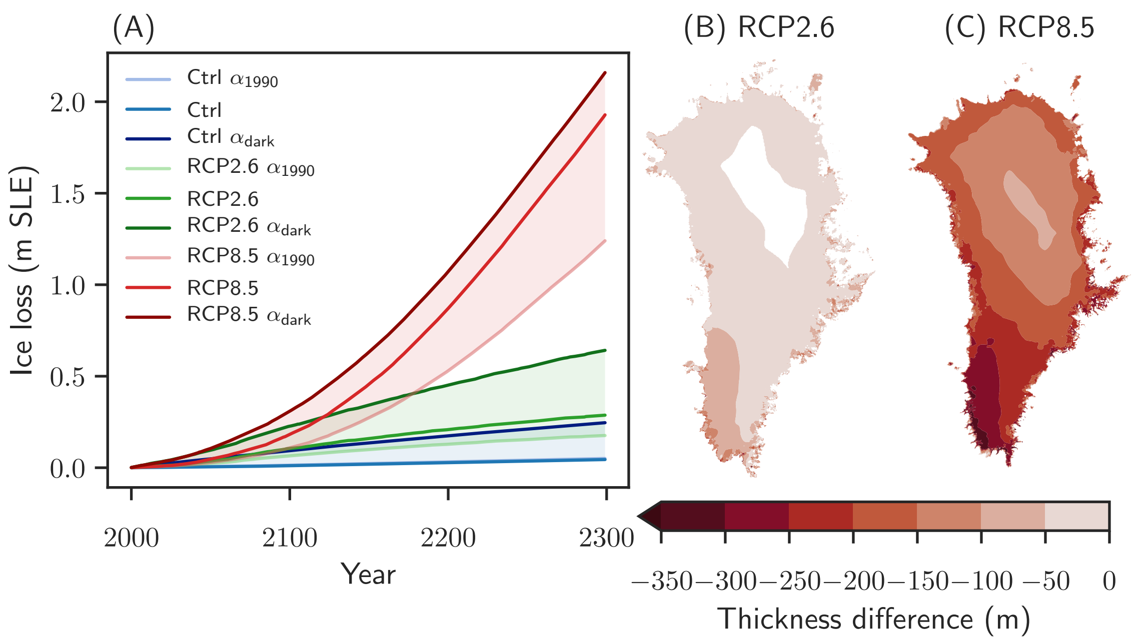

Figure 5. Influence of the melt–albedo feedback on Greenland's ice loss under future warming. The scenarios consist of a control experiment (blue), and temperature forcing with RCP2.6 (green) and RCP8.5 (red). (A) Ctrl, RCP2.6, RCP8.5: ice loss between 2000 and 2300 modelled with PISM-dEBM-simple using the standard parameters and the respective temperature forcing. α_{1990}: ice loss for the respective temperature forcing with the respective temperature forcing with monthly albedo fixed to the average pre-1990 values, thereby interrupting the melt–albedo feedback. α_{dark}: ice loss with the respective temperature forcing with summer albedos (June, July, August) set to the bare-ice value over the whole ice sheet. This yields an upper limit of ice losses driven by albedo changes. The shading is to illustrate the corridor between the lower and upper limit. Ice loss is given in meter sea-level equivalent. 1 mSLE corresponds to approx. 361800 Gt of ice. Panel (B) and (C) show the ice thickness difference between the lower bound experiments α_{1990} and the standard experiments for RCP2.6 (B) and RCP8.5 (C) in the year 2300.

leads to 40 Gt/yr increase in runoff compared to present day and to 113 Gt/yr increase in runoff when compared to preindustrial values.

4 Influence of the melt–albedo feedback on Greenland's ice loss under future warming

Here~~we analyse,~~, we analyse how changes in albedo may impact the melt rates and the ice losses of the Greenland Ice Sheet under the greenhouse gas emission scenarios RCP2.6 and RCP8.5. In particular we focus on the melt–albedo feedback, and on the additional ice losses driven by changes in albedo. The experiments are motivated and described in detail in Section 2.6. The volume of the ice sheet and the mass losses until the years 2100 and 2300 due to the respective warming scenarios are summarized in Table 2.

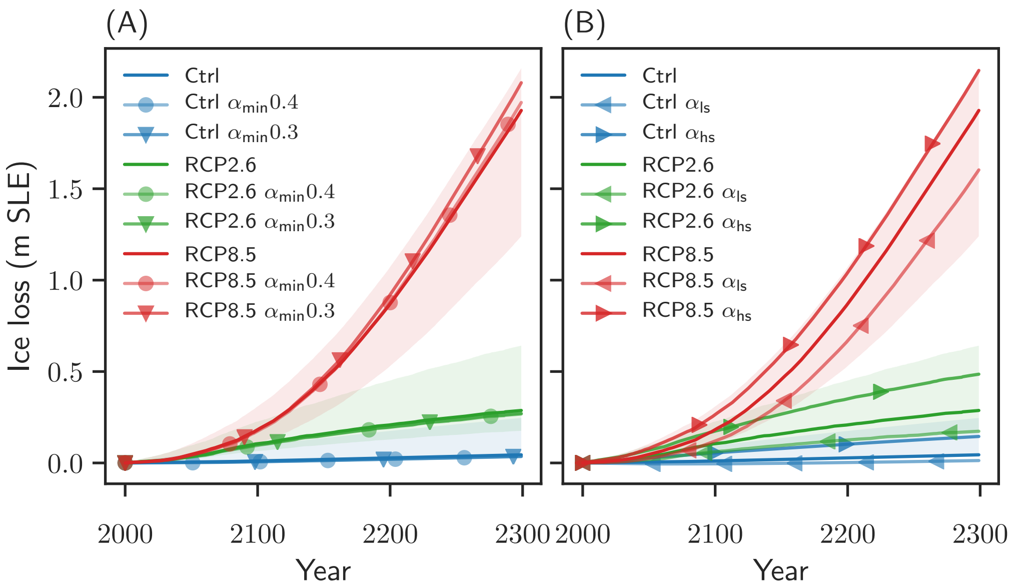

Figure 6. Uncertainty of albedo-change driven ice loss. Ice losses of the Greenland ice sheet under the Ctrl, RCP2.6, and RCP8.5 sceanrios, exploring the effect of different albedo sensitivities, as described in detail in Sect. 2.3.2 and Figure A1. Shaded regions correspond to the corridor between lower and upper bound for ice losses, as shown in Figure 5. (A) Ice losses with variations in the minimal value for albedo. Lower α_{min} corresponds to darker bare-ice. (B) Ice losses with variation in the slope of the albedo parametrization. α_{ls} experiments use a lower slope (half of the standard value), thus the sensitivity of albedo to melt is reduced. α_{hs} experiments use a higher slope (double of the standard value), thus increasing the sensitivity of albedo to melt. Ice loss is given in meter sea-level equivalent. 1 mSLE corresponds to approx. 361800 Gt of ice.

4.1 Ice losses under warming without the melt–albedo feedback

~~Experiments with a~~ The experiments $\boldsymbol{\alpha_{1990}}$ use fixed monthly albedo fields and thereby interrupt the melt–albedo feedback~~and allow to explore the~~. Those experiments illustrate a lower bound of ice-losses due to warming ~~.~~ in this model setup. Note that the lower bound of projected future ice losses under global warming likely differs due to the coarse resolution and the lack of ice-ocean interaction in this study. As described in Section 2.6, the monthly albedo in the ~~experiments the monthly albedo~~ $\boldsymbol{\alpha_{1990}}$ experiments is fixed to an average yearly cycle given by the pre-1990 values in MARv.3.11. In consequence the insolation related melt, as given by the first term of Equation (1), remains constant or even decreases due to decreasing transmissivity of the atmosphere, and only the temperature dependent term increases due to the warming.

In this scenario, the Ctrl α_{1990} experiment remains constant in volume, while the RCP2.6 α_{1990} shows 5.2 cm ice loss until 2100 and 12.6 cm until 2300. In the RCP8.5 α_{1990} experiment the ice losses amount to 9.8 cm until 2100 and to 119 cm until 2300 (see Figure 5). The mass loss until 2100 is in line with the estimate of 9 ± 5 cm in the community-wide ISMIP6 projections (Goelzer et al., 2020). Note that in contrast to ISMIP6, the ocean driven melting remains constant, even under

Table 2. Sea-level relevant volume (in m sea-level rise equivalent) and mass loss (in cm sea-level rise equivalent). All values are relative to the respective control simulation. Only α_{dark} simulations are given in absolute values, since the Ctrl α_{dark} is an extreme scenario which does not qualify as a control experiment.

Experiments	Volume (m SLE)		ΔVolume (cm SLE)	
	2100	2300	2100	2300
Ctrl α_{1990}	7.59	7.55	-	-
Ctrl	7.59	7.55	-	-
Ctrl α_{dark}	7.51	7.35	8.3	20.1
Ctrl α_{ls}	7.60	7.59	-	-
Ctrl α_{hs}	7.54	7.45	-	-
Ctrl $\alpha_{\mathrm{min}}0.4$	7.59	7.56	-	-
Ctrl $\alpha_{\mathrm{min}}0.3$	7.59	7.56	-	-
RCP2.6 α_{1990}	7.53	7.42	5.2	12.6
RCP2.6	7.49	7.31	9.4	24.3
RCP2.6 α_{dark}	7.37	6.96	21.4	59.7
RCP2.6 α_{ls}	7.54	7.43	6.7	16.0
RCP2.6 α_{hs}	7.42	7.11	11.9	34.1
RCP2.6 $\alpha_{\mathrm{min}}0.4$	7.50	7.33	9.0	23.3
RCP2.6 $\alpha_{\mathrm{min}}0.3$	7.50	7.33	9.2	23.5
RCP8.5 α_{1990}	7.49	6.36	9.3	119.0
RCP8.5	7.42	5.67	16.8	188.4
RCP8.5 α_{dark}	7.29	5.44	29.3	211.5
RCP8.5 α_{ls}	7.48	6.00	12.2	159.0
RCP8.5 α_{hs}	7.34	5.45	20.5	200.2
RCP8.5 $\alpha_{\mathrm{min}}0.4$	7.43	5.63	16.4	193.7
RCP8.5 $\alpha_{\mathrm{min}}0.3$	7.42	5.52	17.0	204.4

increased temperatures, and there is no glacier retreat due to ice-ocean interactions. ~~But, also~~ However, the mitigating effect of an precipitation increase ~~under~~ in a warmer climate is also missing.

4.2 Increased ice loss through the melt–albedo feedback

In the following we present the results for ~~ice loss~~ the **std** experiments with PISM-dEBM-simple, taking into account the melt–albedo feedback through the melt-dependent albedo parameterizations as described in Section 2. In the control simulation Ctrl, without temperature forcing or artificial darkening, the ice sheet is stable in volume on the timescale of 300 years.

In the RCP2.6 simulations, the moderate increase in temperatures leads to a approximately linear decline in ice-volume with an ice loss of 9.4 cm until 2100 and 24.3 cm until ~~2300.~~ 2300 in comparison to the year 2000. This is an increase in ice loss of +82 % in 2100 and +93 % in 2300 in comparison to the simulation without ~~melt-alebdo~~ melt-albedo feedback α_{1990} , see Fig. 5 and Table 2. The RCP8.5 simulations show a strong and non-linear decline in ice-volume, with ice losses of 16.8 cm in 2100 and 1.88 m in 2300. This corresponds to a relative increase of +80 % and +58 % respectively due to the melt–albedo feedback. The relative contribution of the melt–albedo feedback to ice loss keeps increasing with time for the RCP2.6 experiment, while it becomes less important with time for the RCP8.5 experiment, as the whole ice sheet approaches the minimal albedo value α_{min}. However, in absolute terms the melt–albedo feedback still contributes almost 70 cm SLE mass loss in the RCP8.5 experiment until the year 2300.

We compare these values with the influence of the melt–elevation feedback, (T_{noLR}, see Fig. C1). This feedback is weaker, it increases the ice loss by 18 % and 13 % in the RCP2.6 and RCP8.5 simulations respectively.

The melt–albedo feedback is particularly important in the south of Greenland, where the insolation averaged over the daily melt period $\bar{S}_{\Phi}$ (see Eq. (1)) is highest. Until the year 2300 it initiates up to 100 m additional thinning in the South-West for the RCP2.6 scenario compared to RCP2.6 α_{1990} (Figure 5 (B)). In the RCP8.5 experiment, the melt–albedo feedback is impacting the thinning over the whole ice sheet (Figure 5 (C)). However, the most important contribution remains in the south-west of Greenland, with additional 300 m thinning compared to RCP8.5 α_{1990} .

4.3 An upper limit for ice loss through extreme surface darkening

As a next step, the upper limit of the melt–albedo feedback is explored via prescribing summer albedos equal to the bare ice albedo over the whole ice sheet in each year (see details in Section 2.6). First, the effect of such a surface darkening is explored without any temperature forcing in the Ctrl α_{dark} scenario. In this experiment approximately linear mass loss is observed, with a rate of 8 mm SLE per decade (see Fig. 5 (A)) and induces ice losses of 8.3 cm until 2100 and 20.1 cm until 2300. The condition that the local monthly mean air temperature needs to be higher than -6.5 °C to allow melt, prevents further melting in the ice sheet's interior. Topographic changes together with the temperature-lapse rate feedback increase the melt area slowly, but do not have a major impact over the 300 years. Note that this extreme darkening Ctrl α_{dark} scenario alone induces more ice loss than the RPC2.6 α_{1990} scenario (see Fig. 5 and table 2).

The RCP2.6 α_{dark} experiment combines the extreme summer darkening with the RCP2.6 temperature anomaly, thereby increasing ice losses from the RCP2.6 α_{1990} experiment by more than a factor of four in comparison to the α_{1990} experiments (Figure 5 and Table 2). This corresponds to a more than fourfold increase in ice losses. The darkening together with the moderate temperature increase induces an expansion of the melt zone and thus strong melt in areas that are not affected in the Ctrl or RCP2.6 experiments. The ice-volume evolution in the RCP2.6 experiment is closer to the lower than to the upper bound of the melt–albedo feedback.

In the RCP8.5 α_{dark} experiment the summer darkening leads to an increase in ice loss of 214 % in 2100 and 77 % in 2300 in comparison to the no-feedback RCP8.5 α_{1990} experiment (see Fig. 5 and table 2). The strong shock of albedo darkening is particularly relevant when overall temperature increases are still low. In contrast to the RCP2.6 α_{dark} experiment, where the additional mass losses increase with time, here the relative impact of extreme summer darkening decreases on long time scales. As the warming progresses the temperature becomes a more important driver to melt.

Reducing the frequency of ~~the~~ darkening in the RCP2.6 α_{dark} and RCP8.5 α_{dark} ~~experimens linearly~~ experiments to darkening events every two or every five years instead of every year reduces the difference in mass loss between the ~~dark RCP~~ RCP α_{dark} and the RCP experiments (see Appendix Figure B1). When the darkening happens every two years the additional ice losses decrease to approximately half of the ice losses with darkening every year. Similarly, a darkening event every five years leads to only 20% of the additional ice losses coming of the darkening in each year. The effect is approximately linear in the frequency of darkening years. This might help to estimate additional albedo-driven ice losses in extreme years as 2012, if projections for the frequency of such extreme event are available.

Reducing the length of the dark period from the whole summer (i.e. June, July and August) to only one month reveals that the month of June is most sensitive to additional darkening, inducing more than half of the additional ice losses between the RCP8.5 and the RCP8.5 α_{dark} experiments (see Appendix Figure B1). The increased sensitivity to darkening in June could be due to the fact that the northern hemisphere receives the most insolation during the month of June. Moreover, in the beginning of the melt season the albedo has not yet decreased due to the melt processes, so an artificial darkening has the strongest effect, compared to the following summer months.

4.4 Exploring uncertainty in albedo-change driven ice losses

The standard parameters for the albedo parameterization used in the RCP2.5 and the RCP8.5 experiments provide the best fit to the MARv3.11 data over the historic period. However, the corridor for possible contributions of the melt–albedo feedback is large, therefore we test how changes in the albedo parameterization affect the ice losses driven by albedo changes. The albedo parametrization can affect the strength of the melt–albedo in two ways: first by changing α_{min} , the lowest albedo possible, and second by changing the sensitivity of albedo to melt via the slope in Equation (7). To ensure that the subsequent mass changes are not primarily due to model drift, they are corrected by a Ctrl experiment without warming but with otherwise same parameters.

A decreased value of α_{min} does not affect regions where melt rates are below 14 m/yr. Consequently, strong melt rates are necessary to observe its impacts: in the RCP8.5 α_{min} simulations with α_{min} = 0.4 the lowered α_{min} value causes additional 5.3 cm of ice loss in 2300 (compared to RCP8.5) and the RCP8.5 α_{min} simulations with α_{min} = 0.3, it causes additional 16 cm until 2300 (Figure 6 (A)).

In contrast to the α_{min} experiments, changing the slope in the albedo parameterization in Equation (7) affects the sensitivity of the albedo to melt already at low melt rates. In the Ctrl α_{hs} experiment, the ice sheet loses 10 cm SLE until 2300 from the increase in the albedo sensitivity alone, ~~indicating that this is not consistent with the standard parameters~~twice as much as with standard parameters. The increased melt sensitivity, although at the upper end of the uncertainty of the parameters presented in Sect. 2.5, might not be optimal in representing historical melt when the other parameters remain unchanged. We test thus the mass losses of the warming scenarios with respect to the Ctrl α_{hs} experiment, in order to explore the interplay of an increased melt–albedo feedback and warming.

The additional effect of the increased sensitivity on ice-volume evolution depends on the warming scenario. The RCP2.6 α_{hs} experiment with moderate warming is affected by a more sensitive albedo parameterization, with up to 40 % increases in ice loss until 2300 with respect to the RCP2.6 experiment (see Figure 6). In contrast, the additional mass loss in the RCP8.5 α_{hs} is lower with +6 % in 2300 compared to the RCP8.5 scenario. This ~~is because already in the~~ can be explained by the fact that in the high melt rates in the RCP8.5 scenario ~~melt is quickly high enough to reach~~ quickly induce the minimal albedo over the whole ice sheet and thereby interrupt the feedback. Once the minimal albedo is reached, further increase in melt rates does not affect the albedo any more, thus the melt is not affected by the stronger feedback any more.

If the sensitivity of albedo to melt is reduced, the ice sheet in the the Ctrl α_{ls} experiment shows slight mass gains (4 cm over 300 years). The lower sensitivity mitigates mass losses from both, the RCP2.6 α_{ls} and the RCP8.5 α_{ls} experiments, with 8.3 cm and 29.4 cm less mass loss until 2300 respectively. However, even with the reduced melt–albedo feedback the ice losses increase by approximately one third when compared to the α_{1990} experiment without albedo-melt feedback.

5 Discussion

We have presented an implementation of a simple version of the diurnal energy balance model (dEBM-simple) as a module in the Parallel Ice Sheet Model PISM. Using this model we evaluate how changes in albedo impact future mass loss of the Greenland Ice Sheet under the RCP2.6 and the RCP8.5 warming scenarios.

5.1 Implementation and validation

In dEBM, the surface melt is calculated as a function of near-surface air temperature and shortwave downward radiation. A first version of the dEBM was tested and validated in Krebs-Kanzow et al. (2018) and a full version was presented in ~~Krebs-Kanzow et al. (2020)~~Krebs-Kanzow et al. (2021). dEBM-simple adapts the approach taken in Krebs-Kanzow et al. (2018) and adds additional modules to calculate the albedo as a function of melt, as well as the shortwave downward radiation. Therefore the only inputs needed to compute the melt rate are two-dimensional near-surface temperature fields including the yearly cycle, and further a precipitation field in order to complete the climatic mass balance. This approach makes the model as input-friendly as a temperature-index model such as the ~~classical~~ widely-used positive degree day model, but with the advantage to capture the melt–albedo feedback. The dEBM-simple surface mass balance module can therefore be used with PISM in a standalone setting to simulate past and future ice-sheet evolution, requiring only a temperature field, a precipitation field, and the time-series of the temperature anomaly as inputs. As PISM is an open-source project, the module can easily be expanded or implemented in other standalone ice-sheet models.

Being a simple model, dEBM-simple does not fully resolve the spatial pattern and temporal evolution of melt over the Greenland Ice Sheet, the melt rates are slightly overestimated towards the beginning of the melt season (June) and underestimated towards the end of the melt season (August) and at the margins of the ice sheet. This is possibly related to the albedo parameterization, which in turn underestimates the albedos in June and overesimates the albedo in August, not capturing important processes like exposure of firn or ice, or darkening of the ice via algae or meltwater. However, the total yearly melt rates match well with those of MAR over the period 1958-2019 and on this timescale the skill of the model is comparable to the dEBM ~~(Krebs-Kanzow et al., 2020)~~(Krebs-Kanzow et al., 2021). The exception of the extreme melt in the years 2012 and 2019, where dEBM-simple clearly underestimates melt rates, are related to changes in cloud cover or blocking events (Delhasse et al., 2021; Hanna et al., 2014; Hofer et al., 2017), which are not captured by the parameterization of the transmissivity of the atmosphere.

Increased insolation values like during the Eemian increase the melt on average by 97 Gt/yr under otherwise same conditions. This is in line with the findings of Van De Berg et al. (2011). While this is only an approximation with several strong assumptions, e.g. the present-day topography of the ice sheet is preserved and we did not apply changes in the temperature, it illustrates the possibility of extending this model to paleo timescales with relatively low efforts.

The implemented parametrization for albedo is based on a phenomenological relation of albedo to the melt rate. It is a coarse representation of the effects, which are important for the ~~albedo of snow and ice~~snow albedo, such as the grain size~~of snow~~, surface water and melt ponds, impurities ~~on snow or ice~~ (e.g. black carbon or algae) or any dependence on the spectral angle or the cloudiness condition of the sky. ~~Moreover, it~~ The possible darkening of ice is considered only indirectly in this approach. In particular lowering the minimal allowed albedo to values which are typical for either dirty ice or supraglacial melt ponds could allow to estimate albedo changes of the ice. Moreover, the parameterization neglects the impact of the snow-cover thickness, which might mitigate melt-driven reduction in albedo after a winter with heavy precipitation (Box et al., 2012). As the parameters of the albedo scheme are fitted against monthly averages of the MAR albedo, processes which happen on a sub-monthly time scale are not well captured. The aging or renewal of snow, associated with the frequency of snowfall events, is not directly represented in the monthly averaged MAR data used to fit the paremeterization. Neither are the influence of shading, wind exposure or rain spells. These could induce additional variability associated to the albedo-melt relations.

Similarly, the parametrization introduced for the shortwave downward radiation does not take into account temporal or spatial patterns. The inter-annual variability of the cloudiness over Greenland and blocking events can therefore not be represented with this approach.

However, the introduced parametrizations do not introduce a systematic bias or a large additional error in comparison to a purely diagnostic mode of dEBM-simple, where instead of parametrized albedo and shortwave downward radiation the 2D fields of MAR output are used to calculate the melt rates (see Figure A4), while all other parameters are kept constant.

In this manuscript we optimize the dEBM parameters c_1 and c_2 independently from the parameters for the albedo and the transmissivity. All parameters are based on MAR data. While this procedure gives an overall good fit, as seen in Section 2.5, it is not necessarily the optimal solution in combination. However, this procedure does keep the parameters independent from the forcing. A first statistical analysis reveals, that the ice loss under a strong warming scenario (RCP8.5) is sensitive towards variations in the slope of the transmissivity parameterization, in the dEBM parameter c_1 and in the slope of the albedo parameterization. The dEBM parameter c_2 and the slope of the transmissivity parameterization seem to have less influence on ice losses (see Appendix E). One could, based on the application, choose to change the parameterizations independently from the dEBM parameters and thereby study the influence on the ice losses, as we have shown in Section 4.4.

5.2 Sensitivity of the Greenland Ice Sheet to warming and surface darkening

In this manuscript we use the PISM-dEBM-simple model in order to assess the influence of albedo changes and surface warming on the Greenland Ice Sheet. The simple surface mass balance model allows a first estimate of the influence of the melt–albedo feedback on the future evolution of the Greenland Ice Sheet in two temperature scenarios: moderate warming RCP2.6, a scenario compatible with the Paris Agreement, and high warming RCP8.5, a worst-case scenario. Experiments with a fixed yearly cycle of the albedo suppress the melt–albedo feedback and thus serve as a lower bound to future ice losses. In contrast, the extreme scenario in the α_{dark} experiments with the surface albedo lowered to the bare ice value over the whole ice sheet for the months June, July and August serves as an upper bound for future melt through the melt–albedo feedback. The experiments with adaptive albedo serve as a more realistic estimate of future mass losses.

This experimental design allows to attribute ice losses to the melt–albedo feedback. Overall we find that the melt–albedo feedback has a strong influence on melt under future warming. For example in the RCP2.6 scenario, the ice loss almost doubles through the albedo feedback (compare RCP2.6 with RCP2.6 α_{1990}). Moreover, the relative amount of ice loss driven by changes in albedo keeps increasing over time. In contrast, the share of melt driven by albedo changes is lower in the high temperature RCP8.5 scenario and decreases as the temperature increases, indicating that temperature is a more important driver under these conditions. Note however, that the absolute increase in mass loss through the feedback is higher for RCP8.5 than for RCP2.6. We also find that extreme darkening alone, without any temperature anomaly, can initiate mass losses comparable to the RCP2.6 scenario.

Moreover, the interaction between the extreme darkening and warming initiates additional ice losses. In particular the RCP2.6 α_{dark} scenario loses 23 % more mass until 2300 than the sum of RCP2.6 and the Ctrl α_{dark} simulations, suggesting that other feedbacks, such as the melt–elevation feedback, enhance the mass loss of the RCP2.6 α_{dark} scenario.

In this setup the melt–elevation feedback has a smaller impact on ice losses than the melt–albedo feedback: experiments which neglect the melt–elevation feedback (here ~~paramterized~~ parameterized through the atmospheric temperature lapse rate of -6K/km) lose 18 % less mass in the RCP2.6 T_{noLR} scenario and 13% less in the RCP8.5 T_{noLR} scenario until 2300, respectively (see Figure C1). This is in line with previous studies (Le clec'h et al., 2019) and suggests that the melt–elevation feedback, although weaker than the melt–albedo feedback, should not be neglected on the time scale of several centuries.

In this study, we assume simplified representations of both the melt–elevation and the melt–albedo feedbacks. However, certain effects such as the feedbacks between the topography of the ice sheet and the atmospheric conditions which affect the surface mass balance cannot be expressed in the atmospheric temperature lapse rate alone. Similarly, the albedo is affected not only by melt, but also the sky conditions, snow events, and impurities. While PISM-dEBM-simple is computationally efficient and represents the ice dynamics well, it cannot compete with an explicit process-based snow-pack model as used by the regional climate models MAR or RACMO (Le clec'h et al., 2019; Kuipers Munneke et al., 2011) or represent the effect of summer snowfall on albedo (Noël et al., 2015). Moreover, here the melt–albedo feedback is represented by a relation linear at low melt rates and obtained from a MAR simulation over the historic period 1958-2019. This relation might not apply under future warming. Therefore we test uncertainties related to the albedo parameterization with the resulting mass losses lying in the corridor between the lower bound, i.e., the no-feedback scenario, and the upper bound, i.e., the extreme darkening scenario. Further analysis of the influence of the melt–albedo feedback with models that fully resolve the firn layer would be helpful to analyse processes that are neglected or simplified in this manuscript.

In observations, long-lasting albedo changes are already found as a consequence of heat waves which initiate strong surface melt (Nghiem et al., 2012; Tedesco and Fettweis, 2020). While the regions with the most rapid darkening in Greenland are located in the ablation zones, ice-sheet wide melt events trigger albedo changes over the whole ice sheet (Tedesco et al., 2016). Studies suggest that heat waves in the Arctic may become more frequent with future warming (Dobricic et al., 2020), with still unknown consequences to ice-sheet melt and albedo. Currently, there are no explicit albedo projections, which take all processes and feedbacks like the distribution of surface meltwater, algae growths, dust deposition, and dust meltout into account. While PISM-dEBM-simple does not explicitly model all these processes, it adds a tool to explore albedo-change scenarios and their influence on the future evolution of the ice sheet in a numerically efficient way, which takes the ice dynamics into account.

6 Conclusion

The module dEBM-simple is implemented in the open-source Parallel Ice Sheet Model PISM and allows to capture albedo and insolation as well as temperature-driven melt in standalone ice-sheet simulations. Due to its simplicity it can be used to perform large-scale ensemble studies or long-term simulations over centuries to millennia. The source code is fully accessible and documented, as we want to encourage improvements and implementation in other ice-sheet models. This includes the adaption to other ice sheets than the Greenland Ice Sheet.

Using PISM-dEBM-simple we find that the melt–albedo feedback can lead to additional 12 cm sea-level equivalent of mass loss in RCP2.6 and additional 70 cm in RCP8.5 in the projected mass loss until the year ~~2300.~~ 2300 with PISM. While our experiments rely on a simple parameterization of albedo with surface melt, they show that future albedo changes can make an important contribution to Greenland's future mass loss.

Code and data availability. The PISM source code including the dEBM-simple module is freely available through https://github.com/mariazeitz/pism/tree/pik/dEBM_dev. The code of the regional climate model MAR is available through https://mar.cnrs.fr/. The MAR data is available at ftp://ftp.climato.be/fettweis/MAR v3.11/Greenland/ERA_1958-2019-10km/monthly_1km/. The CMIP5 datasets for the RCP2.6 and the RCP8.5 warming scenarios are available through https://esgf-node.llnl.gov/search/cmip5/.

Table A1. Variables used in dEBM-simple

Name	Variable	Unit
z	ice surface elevation	km
α_s	albedo	
$\bar{T}$	monthly average near surface temperature	$^\circ$C
T_{eff}	cumulative temperature exceeding the melting point	$^\circ$C
τ_α	transmissivity of the atmosphere	
$\bar{S}_\Phi$	TOA insolation, averaged over Δt_Φ	W/m^2
SW	shortwave downward radiation at the surface	W/m^2
M	melt rate	kg m^{-2}s^{-1}
Δt_Φ	time period with sun above elevation angle Φ	s
$\bar{Q}^{\text{day}}$	daily average TOA insolation	W/m^2
ϵ,ω,e	orbital parameters	$^\circ$,$^\circ$,

Table A2. Parameters used in dEBM-simple

Name	Parameter	Value	I
ρ_w	fresh water density	$1000\,\text{kg/m}^3$	
L_m	latent heat of fusion	$3.34\cdot10^5\,\text{J/kg}$	
c_1,c_2	dEBM parameters	~~29, −93~~ $29\,\text{W/m}^2\text{K}, -93\,\text{W/m}^2$	
S_0	solar constant	$1367\,\text{W/m}^2$	Li
Φ	minimal elevation angle for melt	17.7°	Krebs-Kanzow et
σ	standard deviation of daily temperature	5K	
a, b	parameters for transmissivity	0.57, 0.0037km^{-1}	op
$\alpha_{\max}$, $\alpha_{\min}$	maximal and minimal albedo values	0.82, 0.47	op
$\cancel{c}\ \alpha_{\text{sl}}$	slope in albedo parametrization	-0.025 yr/m	op
γ	atmospheric temperature lapse rate	$-6\,\text{K/km}$	typ
$T_{\min}$	temperature threshold for melt	-6.5°C	~~Krebs-Kanzow et al. (2018, 2020)~~Krebs-Kanzow et al. (20

Appendix A: Parametrizations for standalone ice-sheet models

A1 Parametrization of albedo as a function of melt

Albedo is complicated to parametrize correctly, because of its dependence on a number of factors: the snow or firn albedo depends on grain size, impurities, surface water, refrozen ice, compaction, sky conditions and spectral angle, while the ice albedo depends on impurities, surface water, sky conditions and spectral angle. Here we aim for a very simple phenomenological

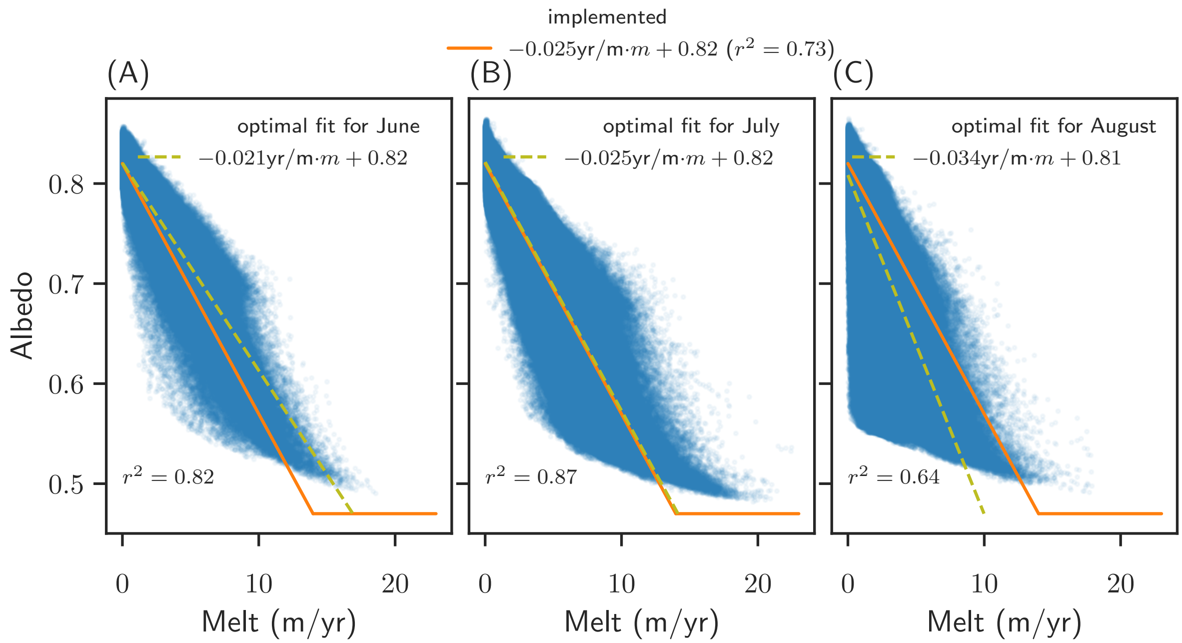

Figure A1. Fit for albedo parametrization Albedo vs. melt of June (A), July (B) and August (C) over the period 1958 - 2019 in the MAR v3.11 dataset. Each dot represents values in one cell of the ice sheet, averaged over a month. Orange lines show the parametrization with parameters as used in PISM. Light green lines show the best linear fit for each month, with the parameters given in the legend.

parametrization of albedo, which is good enough to be valid on large spatial scales and on long time scales. Only the broadband albedo is parametrized here, assuming that the average cloudiness of the sky does not change over long time scales. Further, it is assumed that grain size and surface water can be summarized in a single dependence of the albedo on the melt rate. In the MAR v3.11 dataset, a negative correlation of albedo with melt is found (see Figure A1). The average relation over the months June, July and August in the period of 1958-2019 can be best described by the linear relation $\alpha = -0.025\,\text{yr/m} \cdot m + 0.82$, indicated by the dashed orange lines in Figure A1. The intersection with the y-axis is interpreted as average snow albedo. At very high melt rates the albedo is less sensitive to additional increases with melt, which might be caused when e.g. the snow cover disappeared and bare ice is exposed. In this parametrization we introduce a lower limit to the albedo such that it can not be lower than 0.47 (approximately the value for bare ice (Gardner and Sharp, 2010; Bøggild et al., 2010)). This value is lower than the MAR value for bare ice, but in line with MODIS and RACMO at the ice margin, where impurities can accumulate (Noël et al., 2018; van Dalum et al., 2020; Stroeve et al., 2013). There is a large variance in how sensitively albedo is related to melt, which is due to both, spatial and temporal (intra-annual as well as inter-annual) variability. However, a clear long-term trend of how the albedo depends on surface melt could not be established. In July, the albedo is on average less sensitive to melt, with an average slope of -0.021 yr/m, in June the monthly fit is identical to the the whole summer and in August the albedo decreases on average more strongly with melt, corresponding to a slope of -0.034 yr/m. In addition, in August there is a broad

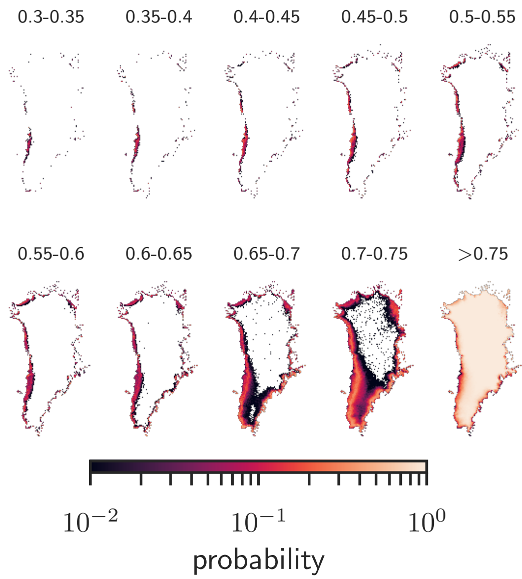

Figure A2. **Observed Albedo** Local probability to find an albedo value within the given bracket on any day in June, July or August from 2000-2019. Data: MODIS Greenland Albedo, see Box et al. (2017).

distribution of albedo values at zero melt, ranging from approximately 0.57 to the fresh snow value, which underlines that the correct albedo depends not only on the current condition but also on the melt during the past months. In order to estimate the sensitivity of future ice evolution to the exact parameters of the albedo parameterization, we vary the slope of the albedo over a broad range, by taking the double or half slope found with the linear regression, here indicated with the grey lines.

We tested ~~discrete albedo classes, that~~ other albedo parameterizations, which are successfully used in other models (e.g. ~~Krebs-Kanzow et al. (2020); Krapp et al. (2017); Robinson et al. (2010)). In our implementation, we found that the continuous relation of albedo to melt was better to predict melt.~~ Krebs-Kanzow et al. (2021); Krapp et al. (2017); Robinson et al. (2010)).

We found that it is also better suited than a parametrization with the snow thickness, successfully used by many models as well (Krapp et al., 2017). The parameterization with snow thickness did lead to too low albedo values and thus too high melting in North West Greenland, where precipitation is generally low. In our implementation, we found that the continuous relation of albedo to melt performed better to predict melt. Our approach comes with the caveat, that the snow thickness is not considered for the calculation of the albedo, although observations suggest that increased winter snow can mitigate summer melt due to the higher albedo of the snow (Box et al., 2012; Riihelä et al., 2019).

The spatial distribution of summer albedos is shown in Figure A2.

A2 Parametrization of shortwave downward radiation

Shortwave downward radiation that reaches the ice sheet's surface depends on the incoming radiation at the top of the atmosphere, the solar zenith angle, the surface altitude, and the cloud cover. In order to get the most correct estimate of shortwave downward radiation at the ice sheet's surface, it would be ideal to know the monthly average cloud cover. Since here, we aim for a parametrization, which makes the model as simple as a temperature index model concerning the inputs needed, we instead parametrize the transmissivity of the atmosphere with the assumption that the average cloud cover does not change, neither during the summer months nor on longer time scales. Following Robinson et al. (2012) we assume that the transmissivity is solely a function of the surface altitude. In order to get a best estimate to this relation, the top of the atmosphere (TOA) radiation, which depends only on season and latitude, is compared to the MAR output for shortwave downward radiation. The daily average TOA radiation $\bar{Q}^{\text{day}}$ is described by Equation (4). The local shortwave downward radiation SW would then be

$$SW = \bar{Q}^{\text{day}} \cdot \tau_\alpha.$$

A linear relation of the transmissivity to the surface altitude is given by

$$\tau_\alpha = a + b \cdot z$$

with the surface elevation in meters z and the fit parameters a and b. The linear fit for the shortwave downward radiation from TOA insolation was obtained from a linear regression of MAR v3.11 data averaged over June, July and August from 1958 to 2019 (see Figure A3).

Because melting occurs predominantly over the summer months June, July and August, we derive the average transmissivity of the atmosphere based on the transmissivity calculated in MAR in June, July and August. The best fit over these three months simultaneously is obtained with $a = 0.57$ and $b = 0.037$ 1/km, as indicated by the orange dashed line in Figure A3. A seasonality can be observed, the transmissivity is on average higher in June ($a = 0.61$ and $b = 0.026$ 1/km) than in July ($a = 0.57$ and $b = 0.040$ 1/km) and August ($a = 0.053$ and $b = 0.046$ 1/km).

For simulations under paleo-conditions, changes in orbital parameters affect the insolation at the top of the atmosphere and the trigonometric expansion used under present day conditions (see Section 2.3.1) does not hold. The declination angle is then described by $\sin\delta = \sin(\epsilon)\sin(\lambda)$ and the sun-earth distance

$$\left(\frac{\bar{d}}{d}\right)^2 = \frac{(1 + e\cos(\lambda - \omega))^2}{(1 - e^2)^2} \tag{A1}$$

with the oblique angle ϵ, the eccentricity e, the precession angle ω and the true longitude of the earth λ. The orbital parameters e, ϵ and ω are given in the input, while λ varies over the time of the year and is computed internally using an approximation of Berger (1978) :

$$\lambda = \lambda_m + \left(2e - \frac{e^3}{4}\right)\sin(\lambda_m - \omega) + \frac{5}{4}e^2\sin(2(\lambda_m - \omega)) + \frac{13}{12}e^3\sin(3(\lambda_m - \omega)), \tag{A2}$$

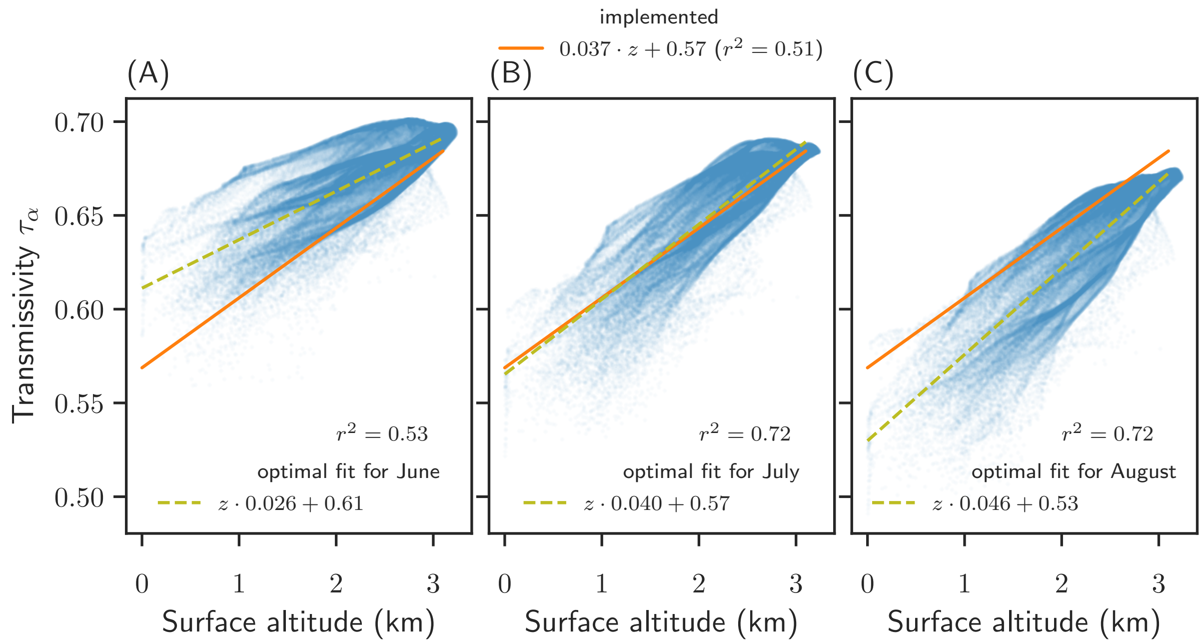

Figure A3. Fit for the parametrization of transmissivity Shortwave downward radiation vs. surface altitude in June (A), July (B) and August (C) over the period 1958 - 2019 in the MAR v3.11 dataset. Each dot represents values in one cell of the ice sheet, averaged over a month. Orange lines show the parametrization with parameters as implemented, which is the best fit over the three months, June, July, and August together. Light green lines show the best linear fit for each month, with the parameters given in the legend.

with

$$\lambda_m = -2\left(\left(\frac{e}{2}+\frac{e^3}{8}\right)\left(1+\sqrt{1-e^2}\right)\sin(-\omega) - \frac{e^2}{4}\left(\frac{1}{2}+\sqrt{1-e^2}\right)\sin(-2\omega) + \frac{e^3}{8}\left(\frac{1}{3}+\sqrt{1-e^2}\right)\sin(-3\omega)\right) + \Delta\lambda$$

$\Delta\lambda = 2\pi(\text{day} - 80)/\text{days per year}.$

Here $\lambda = 0$ at the spring equinox. This approximation is used only for explicit paleo simulations.

A3 Validation of parameterization

In order to asses the validity of the parameterizations in PISM-dEBM-simple, the yearly melt with the fully parameterized model, as shown in Figure 1, is compared to the yearly melt of a diagnostic analysis of the dEBM-simple with otherwise fixed parameters. Instead of computing the albedo and the shortwave downward radiation internally, monthly fields of those variables from the MAR v3.11 date are given as input to compute the melt rates via Equation (1) with the same parameters c_1 and c_2. While the diagnostic experiment performs better in the extreme melt years 2012 and particularly 2019, we find an increased mismatch, in particular in the 1970s and a resulting larger root mean square error. This can be attributed to the fact, that the parameters c_1 and c_2 were optimized for a low temporal and spatial RMSE with the parametrizations for albedo and

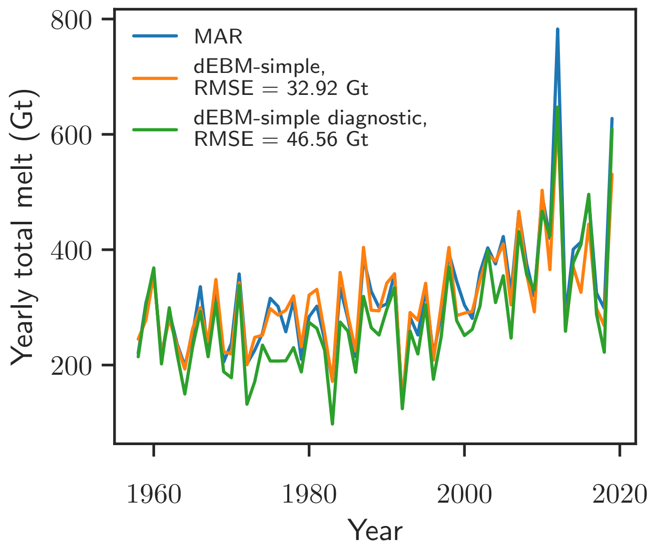

Figure A4. Yearly total melt of the Greenland Ice Sheet as calculated with MAR (blue), diagnosed with the fully parameterized PISM-dEBM-simple simulation (orange), which uses only the monthly 2D temperature fields as input, and diagnosed with a non-paramterized diagnostic dEBM-simple version, which takes the 2D-temperature field, the shortwave downward radiation and the albedo as inputs. The root mean square error for the individual time-series are given in the legend.

transmissivity as desribed above. c_1 and c_2 differ from Krebs-Kanzow et al. (2018) and from an optimal value for the diagnostic melt rate.

Appendix B: Sensitivity to the darkening scenario

In order to test how the results are impacted by a shorter darkening period or even stronger albedo forcing, we study the upper limit RCP8.5 α_{dark} scenario in greater detail.

Shortening the darkening period to only one month reduces, as expected, the impact of darkening. Moreover, it reveals which months are the most vulnerable to darkening. In particular, we observe that darkening in June leads to the highest mass losses (see dash-dotted line in Figure B1 (A)). Darkening in June alone leads to 9.6 cm additional mass loss in 2100 and to 14.8 cm additional mass loss in 2300 compared to the Warming RCP8.5 scenario without darkening. In contrast, darkening in only July or August has a less significant effect, with 4.3 cm and 1.4 cm additional mass loss in 2100 and 7.5 cm and 5,4 cm in 2300. On the one hand this might be caused by the larger insolation and longer days during the month of June. In June average daily insolation at latitudes above 60°N is approximately 7 % larger than in Juny and 50 % larger than in August. Moreover, due to the high melt in the Warming RCP8.5 scenario, albedo values are already low in July and August, even without darkening.

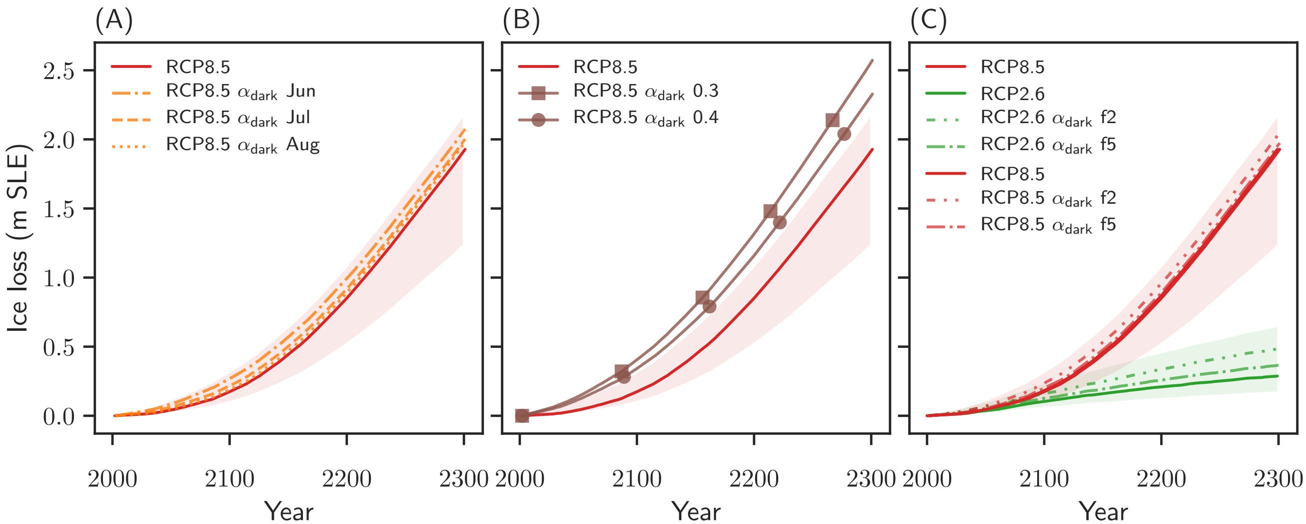

Figure B1. Sensitivity to the darkening scenario Ice volume evolution for different implementations of the darkening scenario. The envelopes of minimal and maximal mass loss, given by α_{1990} and α_{dark} experiments, and the RCP simulations with standard parameters are shown for reference. (A) Period of extreme darkening in the α_{dark} scenario are shortened to one month (orange broken lines). (B) The albedo value for extreme summer darkening is lowered to 0.3 (brown line with square markers) or 0.4 (brown line with circle markers). (C) Reducing the frequency of extreme darkening summers to every 2 (f2) and every 5 years (f5). Ice loss is given in meter sea-level equivalent. 1 mSLE corresponds to approx. 361800 Gt of ice.

Using an albedo value which is lower than the value for bare ice leads to increased ice losses. An albedo value of 0.4 instead of 0.47 over the whole ice sheet increases ice loss by additional 16 % or 4.6 cm by the year 2100 and by additional 8 % or 17 cm by the year 2300 compared to the RCP8.5 α_{dark} scenario. An even lower albedo value of 0.3 increases ice losses by additional 37 % or 11 cm by 2100 and by additional 19 % or 41 cm by 2300 compared to the RCP8.5 α_{dark} scenario.

Reducing the frequency of dark summers to every 2 years leads to additional mass losses which are approximately half of the additional mass losses caused by the darkening in every year for both warming scenarios. A darkening frequency of every 5 years leads to additional mass losses of about 20 % of the additional mass loss with darkening in every year. This suggests that, at least on time scales of 300 years, the effects of more or less frequent darkening remain linear.

Appendix C: Effect of the melt–elevation feedback

The melt–elevation feedback is generally represented in all experiments by adjusting surface temperatures with height changes by 6 K/km. The influence of the feedback on the simulations is tested by switching off this lapse-rate correction, with the resulting mass loss shown in Fig. C1.

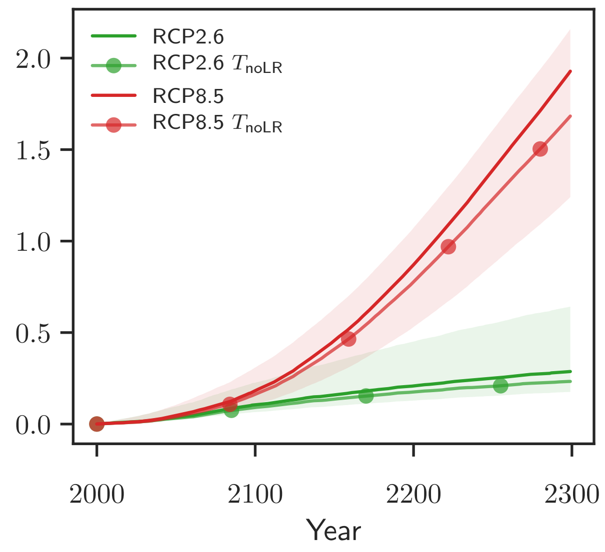

Figure C1. Impact of the melt–elevation feedback. PISM-dEBM-simple simulations of the Greenland Ice Sheet with RCP2.6 (green lines) and RCP8.5 (red lines) warming. Dark solid lines take the melt–elevation-feedback through the atmospheric temperature lapse rate into account. Shaded lines with markers neglect the melt–elevation feedback and assume a zero atmospheric temperature lapse rate. Ice loss is given in meter sea-level equivalent. 1 mSLE corresponds to approx. 361800 Gt of ice.

Appendix D: Surface melt computed with the positive degree day method (PDD)

During the historic validation period, the simulation with the positive degree day method (PDD) for melt has a similar performance to PISM-dEBM-simple (see Figure D1). The standard parameters were used for this simulation: The standard deviation of the temperature $\sigma = 5\,\mathrm{K}$, the melt factor for ice $f_i = 8\,$(mm liquid-water-equivalent)/(pos degree day) and the melt factor for snow $f_s = 3\,$(mm liquid-water-equivalent)/(pos degree day). However, the spatial distribution of melt anomalies shows a distinct north-south-gradient, with an overestimate of melt in the north and an underestimate of melt in the south, see Figure D2.

In the warming simulations, the simulations with pdd melting show increased melt compared to the dEBM simple in the high temperature scenario (see Figure D3). In RCP2.6 the north south bias in the melt rates, compared with PISM-dEBM-simple persists. However, the positive and negative biases balance each other out and lead to mass losses very similar to those computed with PISM-dEBM-simple. In contrast, the melt rates in the RCP8.5 scenario are almost consistently higher with the PDD melt module, only in the south-west PISM-dEBM-simple produces higher melt rates than PDD. We find an increase of ice loss of 12% in the year 2100 and of 47% in the 2300, compared to the standard dEBM run. The difference between ice losses computed with dEBM or with PDD is not only due to different sensitivities to temperature increase

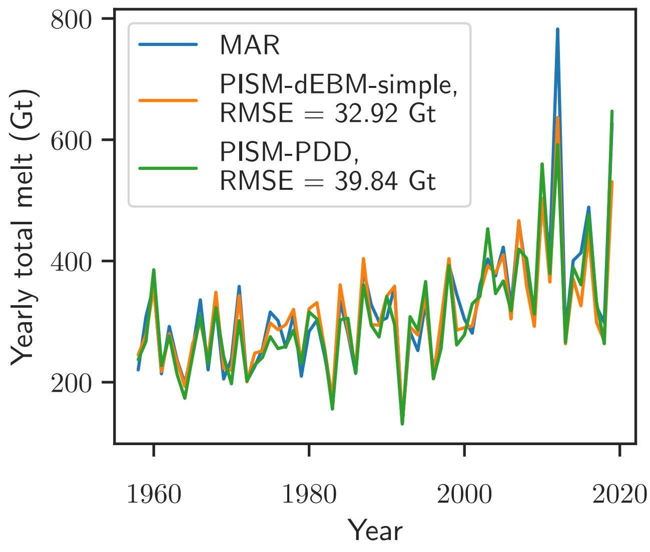

Figure D1. Comparison of annual total melt of the Greenland Ice Sheet as calculated with MAR v3.11 and PISM-PDD. The diagnostic simulation with PISM-PDD (green line) is performed using monthly MAR 2D temperature fields as forcing. The root mean square difference between the PISM-PDD simulation and total melt as given by MAR (blue line) is 39.84 Gt. Details on the PISM-dEBM-simple simulation are found in Figure 1 and in Section 3.

Table E1. Overview over the experiments performed in this study

Name	Variable	Range
c_1	dEBM parameter	$[27, 31]\ \mathrm{W/m^2K}$
c_2	dEBM parameter	$[-95, -93]\ \mathrm{W/m^2}$
α_{sl}	slope in albedo parametrization	$[-0.034, 0.021]\ \mathrm{yr/m}$
$\tau_{\mathrm{A,sl}}$	slope in transmissivity parametrization	$[0.026, 0.046]\ \mathrm{km^{-1}}$
$\tau_{\mathrm{A,in}}$	intercept in transmissivity parametrization	$[0.53, 0.65]$

Appendix E: Variablity of RCP8.5 simulations

In addition to the RCP8.5 simulation with standard parameters, we tested how the variability of the parameters impacts the volume changes under an RCP8.5 forcing. Here, the experimental protocol is analogous to the protocol for standard parameters, described in the main paper in Section 2. However, instead of using only the standard set of parameters, the values for five parameters have been drawn randomly, creating an ensemble of 100 members. The varied parameters are summarized in table E1.

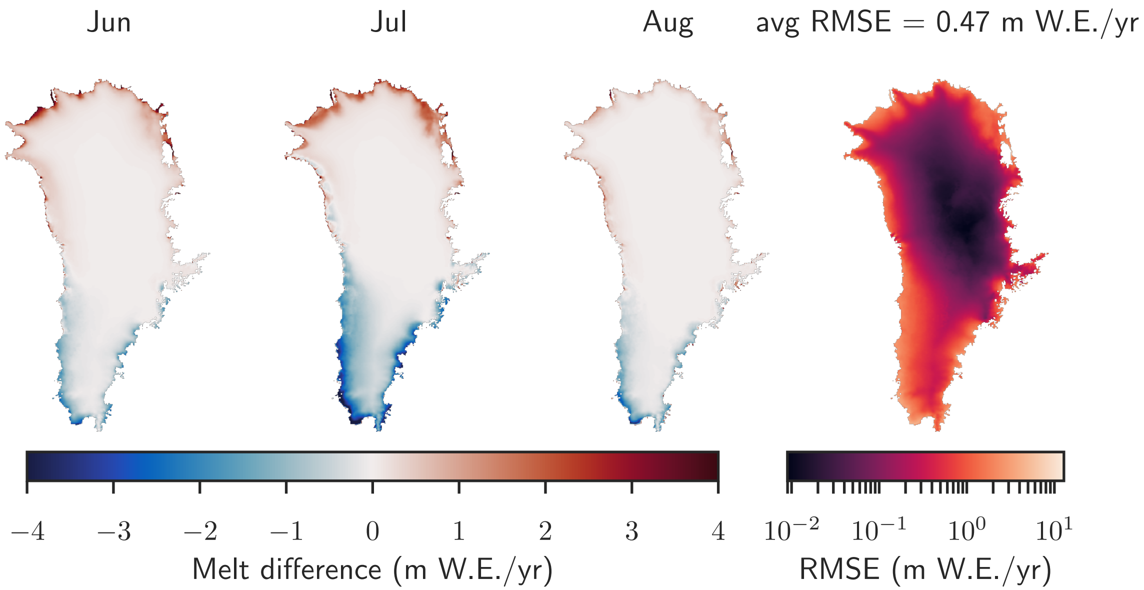

Figure D2. Local differences between the monthly averaged June, July, and August melt rates as diagnosed with PISM-PDD compared to MAR. The PISM simulation uses monthly 2D temperature fields from MAR as forcing. Positive numbers mean that PISM overestimates the melt and negative numbers mean that PISM underestimates the melt. The local root mean square error averaged over June, July and August from 1958-2019 is shown in the right plot. The spatial average of the RMSE is 0.47 m/yr.

We use uniform random distributions instead of Gaussian for all parameters. The dEBM parameters c_1 and c_2 were derived by optimization of historic melt rates (see Section 3, therefore we do not have an estimate of a mean or a standard deviation. The range of parameters, which was used for this ensemble, was chosen such that all parameters c_1 and c_2 for which the root mean squared error in the historic melt rates does not increase by more than 10% compared to the standard values. The parameters which describe the albedo and the transmissivity parameterizations were chosen such that the intra-annual variability is represented (see Appendix A).

The volume change of each ensemble member remains in the envelope given by the RCP8.5 α_{1990} simulations as a lower bound and the RCP8.5 α_{dark} as an upper bound (see Figure E1 (A)). The variability of the intercept of the transmissivity parameterization has the largest influence on the variablity in ice loss after 300 years due to warming. The ice loss until 2300 also seems to be correlated (or anti-correlated) to the dEBM parameter c_1 and the slope of the albedo parameterization, while the dEBM parameter c_2 and the slope of the transmissivity parameterization seem to have only negligible influence on the ice loss due to warming (see Figure E1, (B)-(F)). However, the ensemble size of 100 is not large enough for a thorough statistical analysis.

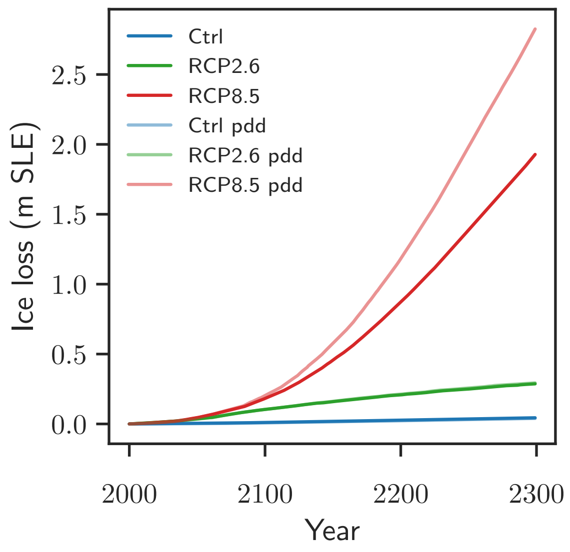

Figure D3. **Comparison with the positive degree day model** PISM-dEBM-simple and PISM-PDD simulations of the Greenland Ice Sheet with RCP2.6 (green lines) and RCP8.5 (red lines) warming. Ice loss is given in meter sea-level equivalent. 1 mSLE corresponds to approx. 361800 Gt of ice.

Author contributions. MZ implemented the dEBM-simple module with the parameterizations for albedo and shortwave downward radiation and performed the analysis. RW conceived the study. JB contributed the initial state for PISM simulations. UKK developed the dEBM model, the basis for dEBM-simple, and provided support with the setup. MZ and RR designed and wrote the manuscript with input and feedback from all coauthors.

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

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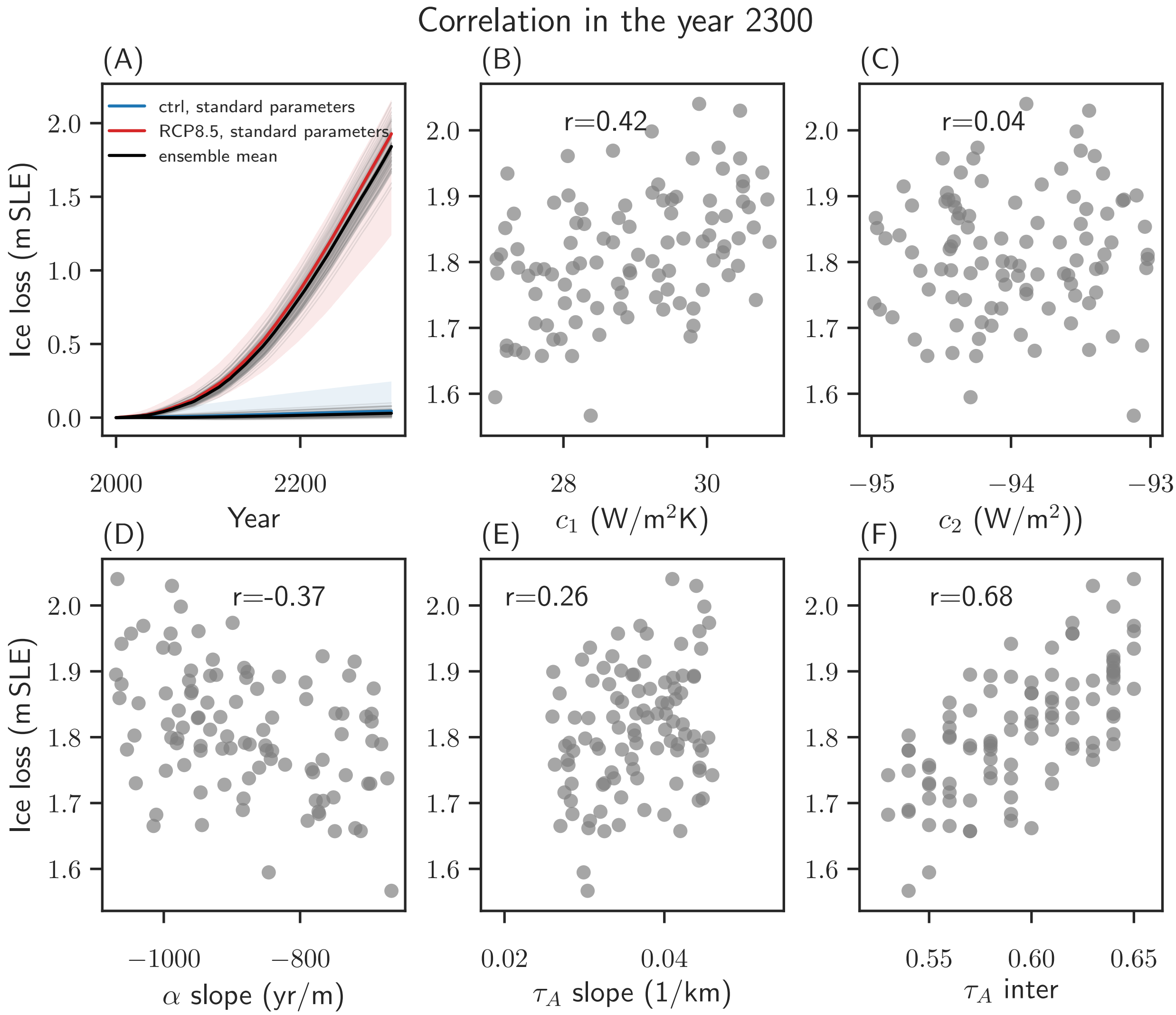

Figure E1. **Impact of the parameter variability.** (A) Timeseries of PISM-dEBM-simple simulations of the Greenland Ice Sheet with RCP8.5 warming and control simulations. The thick red and blue line are simulations with standard parameters, the shading shows the upper and the lower bound of the melt-albedo feedback, as shown in Figure 5 and discussed in Section 4. The thin black lines are the ensemble simulations, with parameters drawn in random an shown in Table E1. Thick black lines show the ensemble average. Ice loss is given in meter sea-level equivalent. 1 mSLE corresponds to approx. 361800 Gt of ice. (B)-(F) Ice loss until year 2300 in mSLE vs. each of the varied parameters. Note that here the ice loss is corrected by the respective control simulation, which uses the same set of parameters but has no temperature forcing. The Spearman correlation coefficient is given in each of the panels.

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