========= Reviewer comment 1 (anonymous) =============

I think the paper provides an important contribution to our understanding of the possible past behaviors of the Antarctic Ice Sheet and will be of interest to the community. I recommend publication after minor (if any) corrections.

We appreciate the positive review from Referee#1 and thank them for their valuable suggestions. Below you can find our response to each comment. We would like to point out that simulations have been repeated with an improved version of the Yelmo model (Robinson et al., 2020). Although specific values for each simulation have slightly changed, our main conclusions remain robust.

1.) The authors assume a relaxation time of 3,000 years for the GIA component (Page5, line 8). The community is undergoing a shift in ideas on the rheology of the Earth beneath the Antarctic Ice Sheet (e.g. Whitehouse et al., 2019; Barletta et al., 2018). How sensitive is your model to this relaxation time? What happens if you use a weaker rheology?

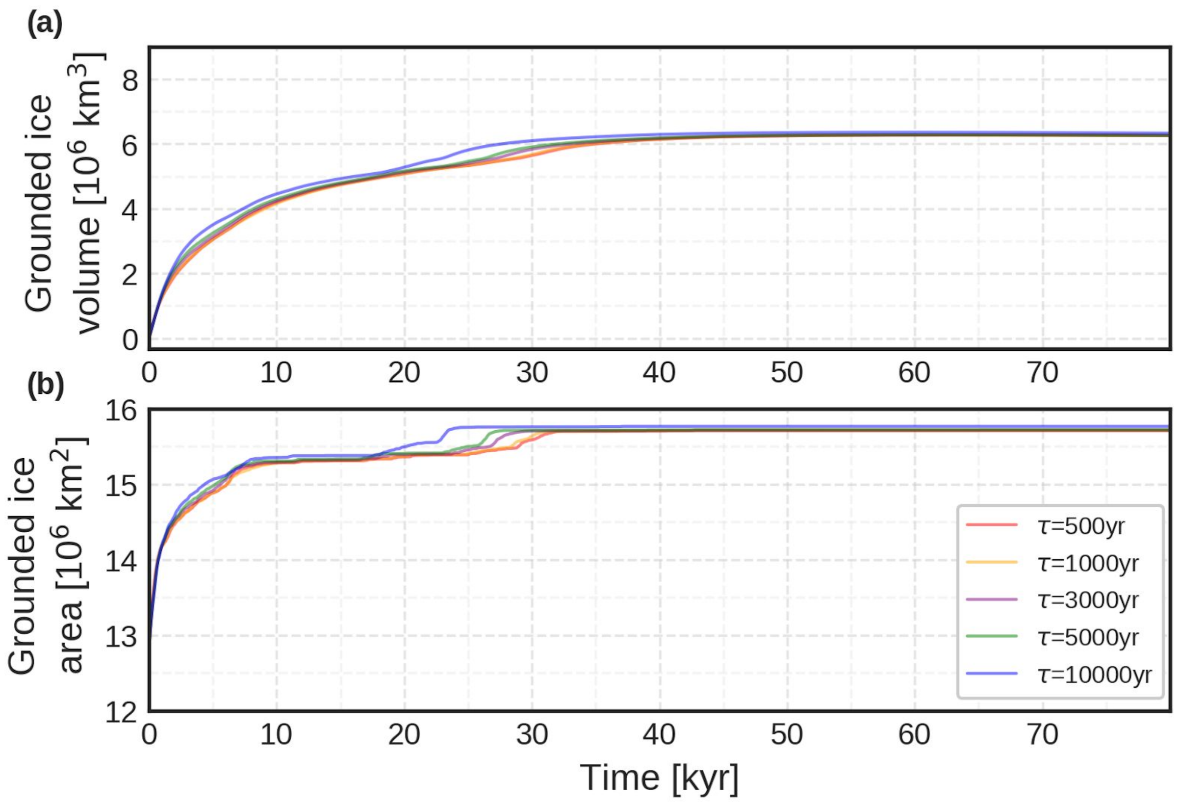

*Fig. 1 Evolution of the **(a)** grounded ice volume; **(b)** grounded ice area for different relaxation times (from 500 yr - 10000 yr) for the friction values z0 = -125 m and c_min = $5*10^{-5}$ yr/m.*

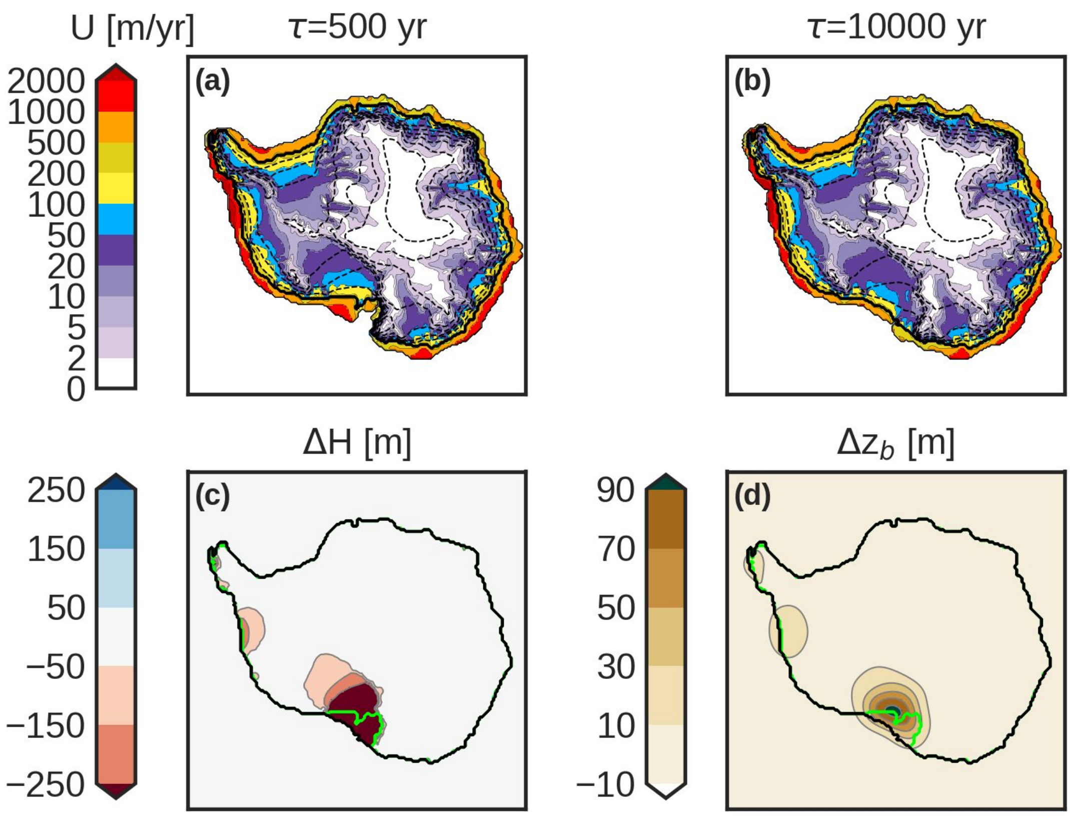

Fig.2: Simulated LGM AIS ice velocity for a relaxation time ***(a)*** *of τ=500 years;* ***(b)*** *τ=10000 years after 25000 years of cold climate evolution. Dashed lines represent surface elevation contours every 500 meters up to 3500 meters. The thick black line represents the grounding-line position.* ***(c)*** *ice thickness* ***(d)*** *bedrock elevation anomaly (****(a)*** *minus* ***(b)****). The thick green/black line represents the grounding-line of* ***(a)/(b)****.*

Figure 1 shows the time evolution of the grounded ice volume and ice area for the reference friction parameters and the Average PMIP3 fields. All simulations yield a similar equilibrated end-state. However, not all of them reach the continental-shelf break at the same time. Figure 2 shows the simulated LGM ice sheet for a weaker rheology (a; τ=500 years) and a stronger rheology (b; τ=10000 years) after 25000 years. These results show that a weaker rheology simulates a lower ice sheet in the WAIS and especially at the Ross shelf, where it does not fully advance (Fig. 2c). Comparing the bedrock elevation differences (Fig. 2d), a weaker rheology has a more elevated bedrock at the Ross shelf, the Bellingshausen Sea and the Amundsen sea, which impedes there a complete advance.

2.) This might just be a reflection of my ignorance with models but your model is allowed to run for 80 ka (Page 6, line 12), I assume to reach some sort of equilibrium but how do we know that the ice sheet was in equilibrium. How important are the dynamics of the ice sheet leading up into the LGM for its LGM behavior?

Indeed, we ran the model for 80 ka to reach an equilibrated state. In this way we can analyze the effect of dynamics and LGM climatologies without accounting for the transient character of the ice sheet. Nonetheless, fully-LGM conditions occurred only for a couple of millennia. After the Last Interglacial (LIG; around 120 ka), global temperatures decreased slowly until they reached the LGM, at around 21 ka. Ice core records of the AIS show that temperatures were around 10 degrees colder than the PD (Jouzel et al., 2007). As temperatures became colder, the AIS advanced up to the continental-shelf break. Due to the steep slope of the continental-shelf break, the AIS is not capable of advancing further.

In order to reach an equilibrium, the total mass balance of the AIS has to be zero. Because ablation in the LGM state is most likely negligible (we argue this in the next question), only calving at the ice front and basal melt at the continental-shelf break lead to mass loss. Hence, given that during the LGM the AIS advanced to the continental-shelf break, it is very likely that accumulation rates were compensated with calving events (and potentially melting for ice shelves below the continental-shelf break, Kusahara et al., 2015), leading to an equilibrated state.

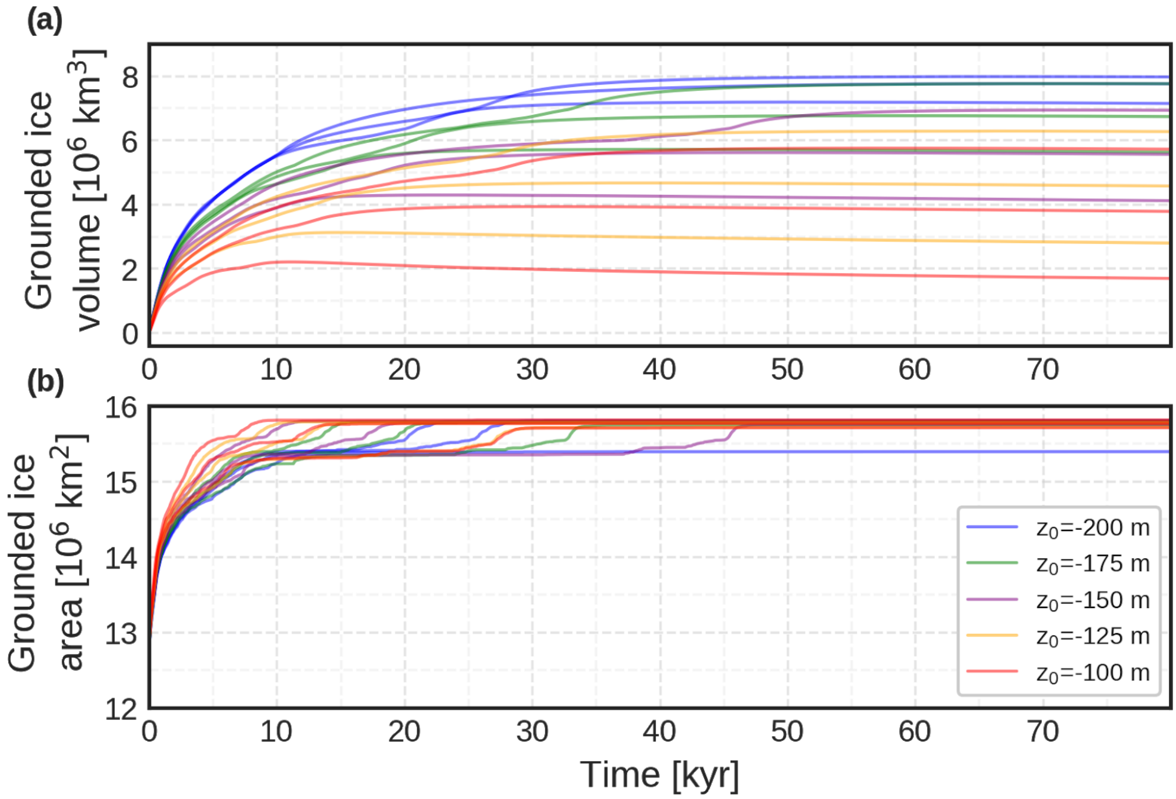

*Fig.3: Simulated **(a)** ice volume and **(b)** ice extent time evolution for the friction parameters ensemble. Coloured lines represent each a z0 value.*

From geomorphological records it is possible to estimate the grounding-line retreat since the LGM to the PD, but there is no ice-extent record from the LIG to the LGM. Dynamics play a crucial role in the evolution of the AIS towards its LGM state. Whereas faster dynamics facilitate a more rapid advance towards the continental-shelf break (red lines in Fig. 3b), slower dynamics need more time to reach the borders (blue lines). Nonetheless, because the simulations that reach the continental-shelf break earlier have faster ice streams, this translates into lower ice volumes (Fig. 3a).

3.)Page 3, line 3 – please give a reference for “ablation and basal melting were probably negligible at the LGM.” Probably, but you could use some justification of this assumption.

Even at PD, ablation rates are almost negligible in the AIS domain except for localized regions, such as the Antarctic Peninsula (vanWessem et al., 2016, 2018). Ice core records show that the AIS was on average 10 degrees colder than the PD at the LGM (Jouzel et al., 2007), thus, by applying a spatially homogeneous cooling, ablation rates turn to be almost negligible. Nonetheless, in this study ablation rates are computed using the output fields from the PMIP3 models. As shown in the Supplementary Material (SM), some models do show ablation in the AIS domain but we consider this very unrealistic.

Basal-melting rates on the other hand are more difficult to infer. In order to do so, it would be necessary to have a spatial map of subsurface oceanic temperatures and salinity. However, to our knowledge, there are no such paleoceanographic records for the Southern Ocean. From a modelling perspective, PMIP3 fields from the LGM also give the simulated outputs of salinity and oceanic surface temperature. However, these models use an AIS LGM state up to the continental-shelf break (Abe-Ouchi et al., 2015) and hence it is not valid for computing at the interior of the continental-shelf. Therefore, because including basal-melting rates would add a degree of difficulty and the aim was to simulate a fully advanced AIS, basal-melting rates were set to zero for the sake of simplicity in this work.

We rephrased the above sentence in the manuscript for:
"Ablation rates at the PD are almost zero except for localized areas (van Wessem et al., 2016, 2018). Because the LGM is a colder period, around 10 degrees as shown by ice core records (Jouzel et al., 2007), ablation rates in the LGM would have been probably negligible. On the other hand, basal melting rates from the LGM are difficult to estimate due to the scarcity of oceanic temperature reconstructions. Nonetheless, geomorphological records point to a fully advanced AIS during the LGM (The Raised Consortium, 2014). This could hint to low basal-melting rates inside the continental-shelf break."

Other minor editorial suggestions:
1.) Page 2, line 14: remove "up"
Done.

2.) Page 9,line 12-13: "...grounding-line from thickening, as a..."
Done.

3.) Page 9, line 14: "...viscosity such as GISS-E2-R-150..."
Done.

4.) Page 9, line 15: "...Amery Trough."
Done.

5.) Page 9, line 24: "...temperatures, which result in low viscosities. Therefore..."
Done.

6.) Page 10, line 10: "...pronounced; however, inland..."
Done.

7.) Page 10, line 30 – Please explain what you mean by "specially determinant"
Because this statement can cause confusion we changed the sentence

"However, the importance of saturated tills is specially determinant for transient simulations with a retreating grounding line."

to:

"However, the aim of this work was to study the uncertainty associated with the basal drag parameters, rather than assessing the uncertainty for different friction laws."

========= Reviewer comment 2 (Johannes Sutter) =============

The manuscript is generally clearly structured and easy to follow. However, there are some issues which should be addressed to make this a valuable contribution to The Cryosphere. Below I list my major concerns followed by some minor stilistic/editorial aspects.

We are grateful to Johannes Sutter for bringing up several key points that will serve to improve the manuscript. We have addressed these concerns below. We would like to point out that simulations have been repeated with an improved version of the Yelmo model (Robinson et al., 2020). Although specific values for each simulation have slightly changed, our main conclusions are robust.

1. The authors omit a discussion as to how their initial ice sheet configuration affects their results and conclusions:
1.1 How does the model spin up affect the final LGM extent of the AIS. Is there a thermal spin up, paleo-spin up or a "cold start". A more detailed discussion of the initial state of the ice sheet would be useful. I suggest one additional figure (this could be figure 1 or 2) which gives an overview over the initial state of the ice sheet and the present day (PD) tuning simulations (best fit, ice thickness change vs observations, ice volume and sea level equivalent, grounding line configuration, surface velocity). There are some figures in the supplement but I think an overview figure in the main manuscript is needed.

The LGM and the PD simulations start from the same initial state, mainly the PD topographic variables (bedrock, ice thickness, masks, etc.). The remaining variables, namely dynamics and thermodynamics, are derived from boundary conditions. Then LGM and PD conditions are run for 80 kyr under the respective constant climatic conditions, hence a “cold” start for the LGM.

We added a figure of the best simulated PD (which from now on is the new reference state in the manuscript, Fig. 1). A discussion can be found in the next point.

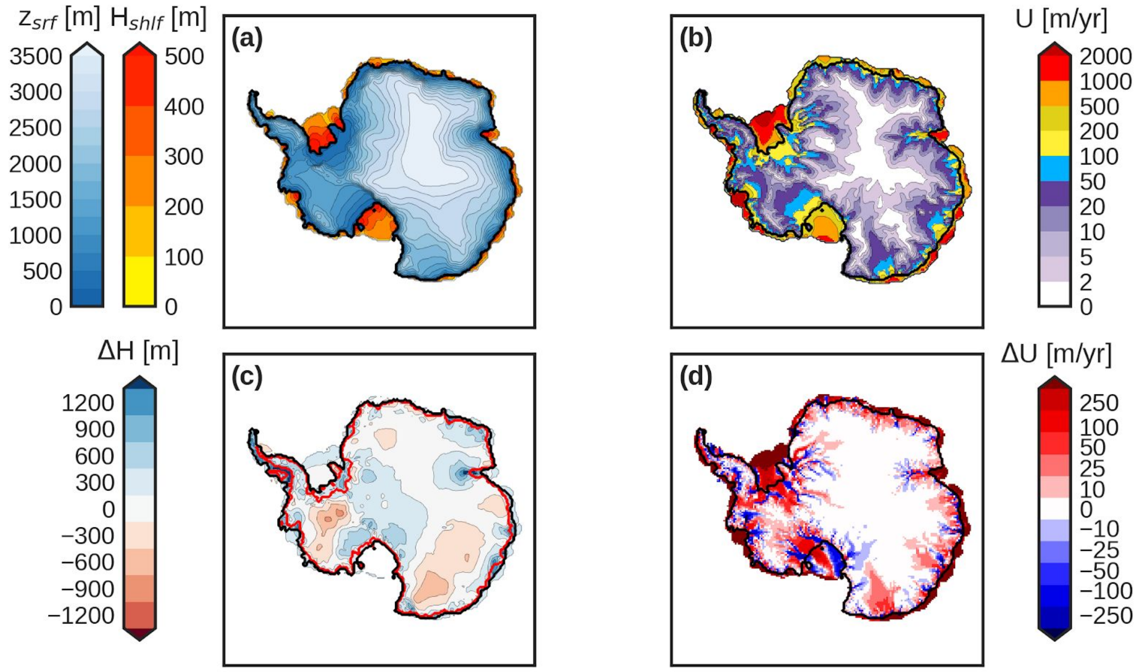

*Figure 1: Simulated PD AIS **(a)** surface elevation (blue) and ice-shelf thickness (orange); **(b)** ice velocity; **(c)** ice thickness anomaly (simulated minus observations); **(d)** surface velocity anomaly, for the best match PD of all the ensemble mean. The thick black line corresponds to the simulated grounding-line position. The thick red line in **(c)** represents the actual grounding-line position.*

1.2 Arguably, the authors chose a relatively loose definition of a "good" present day fit with respect to sea level equivalent ice volume change (-3m to +3m). The ice volume spread at PD due to the parameterisation regime is about the same as the total LGM volume spread in their ensemble. How would the LGM spread change if more rigid conditions are applied for PD (e.g. pm 1m?). Also it seems that ice shelves are extensive in the PD tuning runs (supplementary figure 2). How does this affect grounding line advance (buttressing) as well as SMB (a very large area is gaining mass right away,whereas in reality there might have been no ice shelves).

The extension of the PD ice shelves was improved with the new version of Yelmo and is in better agreement with observations (Fig. 1 and new SM figures). If we apply more rigid conditions, such as ±1 m, then the spread reduces to 3.8 m (from 10.3 to 14.1 msle). We now only focus on that range rather than ±3 m in the new manuscript version.

We added in the Discussion section:

“The simulated PD configurations show a slightly more advanced grounding line in the WAIS compared to the observations, especially at the Ronne shelf. Also the ice thickness in the interior of the WAIS is systematically lower than observations. Both features can be partially explained by the basal-drag parameterisation used. Our parameterisation enhances sliding for deeper bedrock. The WAIS is in its vast majority a marine ice sheet, where bedrock depths can reach up to 2000 m in the interior regions. Thus we systematically simulate a lower WAIS, as we overestimate the ice flow at the interior. This, in addition, promotes the grounding line to advance. Nonetheless, this parameterisation allows for a precise tracing of ice streams. Except in the Larsen embayment, ice shelves generally show a slightly larger extension than observations. Because larger ice shelves allow for more ice accumulation and exert a backward force, it also helps the grounding-line to advance. Thus, the more advanced grounding line in the Ronne, Amundsen sea and Amery shelves could be additionally explained by the backward force exerted by ice shelves. Nonetheless, the overall picture of the simulated AIS fits well with observations in terms of grounding-line position as well as simulated ice volumes.”

2. The experimental setup assumes a steady state LGM-forcing for 80 ka. I understand that Blasco et al. chose an idealised setup in order to fully focus on the effects of different climatological boundary conditions and ice flow parameterisation. This is fine,however the fact should be discussed, so the reader can appreciate the potential impact on the results. In reality, full LGM-forcing was only sustained for maybe a couple of millennia, preceded by a long cooling period starting at the end of the last interglacial.The authors should include a discussion of the transient evolution of the AIS from the initial present day (PD) state to the final LGM state. How fast is equilibrium reached?Does it take several tens of thousands of years or only a couple of millennia? Is the relative homogeneity of the grounding line extent due to the long integration time under LGM conditions or the forcing? What role does the sea level boundary condition play?Actually, reading the text I was missing information whether sea level was set to LGM conditions (ca. -120 m) or PD or something in between? This is important information,as sea level alone

exerts a big influence on the grounding line position via the flotation criterion. Here an additional figure would be nice which shows the transient growth of the AIS under constant LGM forcing for each ensemble member. This would elucidate the inter-ensemble differences in the pace of AIS grounding line advance and ice volume change.

Indeed, we left out information about the boundary sea-level stand. In the simulations, it is set to -120 m. This has been added to the Methods section.

Yes, the large integration time contributes to a similar extension for all the simulations. Nonetheless, this extension is reached in all simulations at most after 45kyr. Assuming that the LGP occurred for almost 100 kyr we think that it is realistic that all the ice sheets fully expanded at the LGM. We added in the Discussion section:

"In this study we assumed steady-state LGM and PD conditions to investigate the effect of climatological boundary conditions and basal drag parameterisation. Of course, this represents a simplification of reality, as full LGM conditions only occurred for a couple of millennia. In a transient simulation, the results would additionally include a potential internal drift, which we tried to avoid. Although simulations were forced during 80 kyr under steady LGM conditions, equilibrated states were reached after only 30 to 40 kyr (see SM). Given that the LGP was a cold and sufficiently long period in the Antarctic domain, constant LGM conditions should be enough to stabilize the AIS near its real LGM state."

We also added to the SM the transient evolution of the whole ensemble (Figures 2, 3).

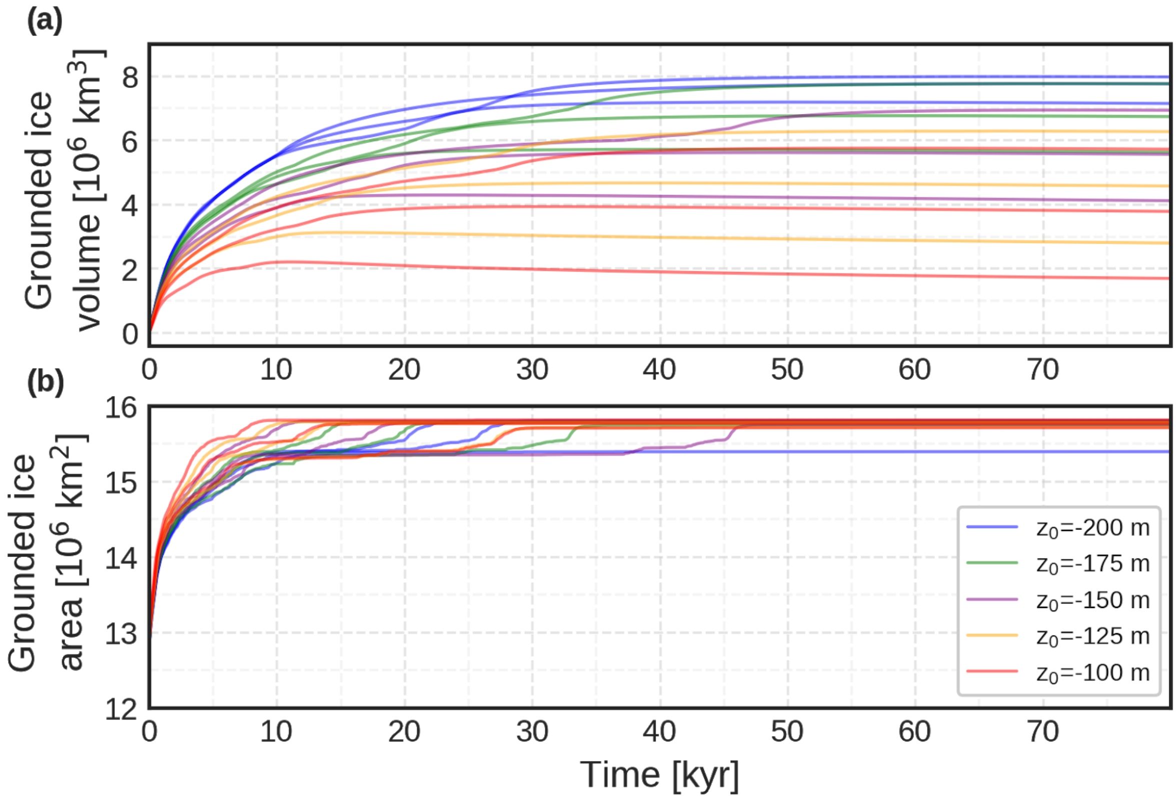

Figure 2: Simulated ***(a)*** *ice volume and* ***(b)*** *ice extent time evolution for the friction parameters ensemble. Coloured lines represent each a z_0 value.*

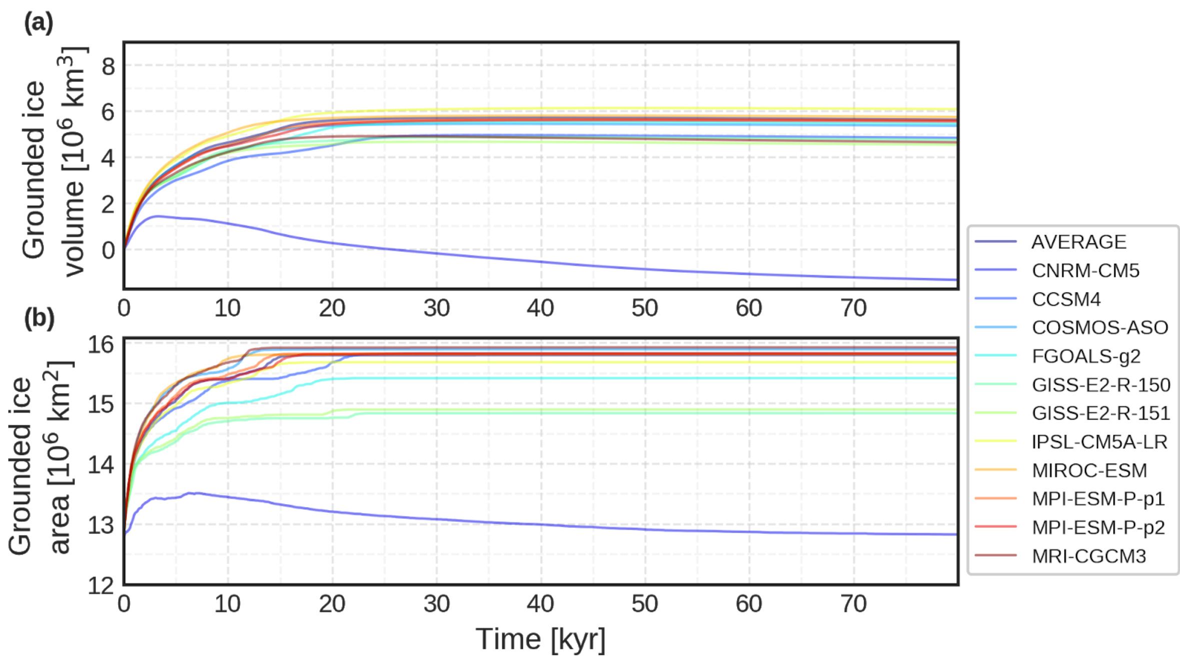

Figure 3: Simulated ***(a)*** *ice volume and* ***(b)*** *ice extent time evolution for the whole PMIP3 ensemble and the reference friction parameters z_0 = -125m and c_{min} = $5*10^{-5}$ yr/m.*

3. Model resolution. This is a somewhat nasty argument as in theory very high spatial resolutions are required to adequately resolve grounding line migration. However coarse resolutions are a tried and tested instrument to allow for larger paleo ice sheet ensembles and the authors do use a sub-grid grounding line procedure to accommodate for the coarse resolution. Still, 32 km are on the rough end of currently used grid-spacing and it would be interesting to see the effect of say doubling resolution(16-km) on final LGM ice volume and extent. This does not have to be done for each and every run, but picking one single member and maybe the GCM-mean forcing would show the impact of resolution on LGM ice sheet configuration. This would mean only two additional simulations and should not take too much time.

We have followed the reviewer's suggestion. Here we show the results in terms of sea-level equivalent (SLE) for 32km and 16km for two PMIP3 members: the COSMOS-ASO as well as the mean forcing of the whole ensemble (AVERAGE). In addition, simulations for PD forcing based on observations were carried out for both spatial resolutions.

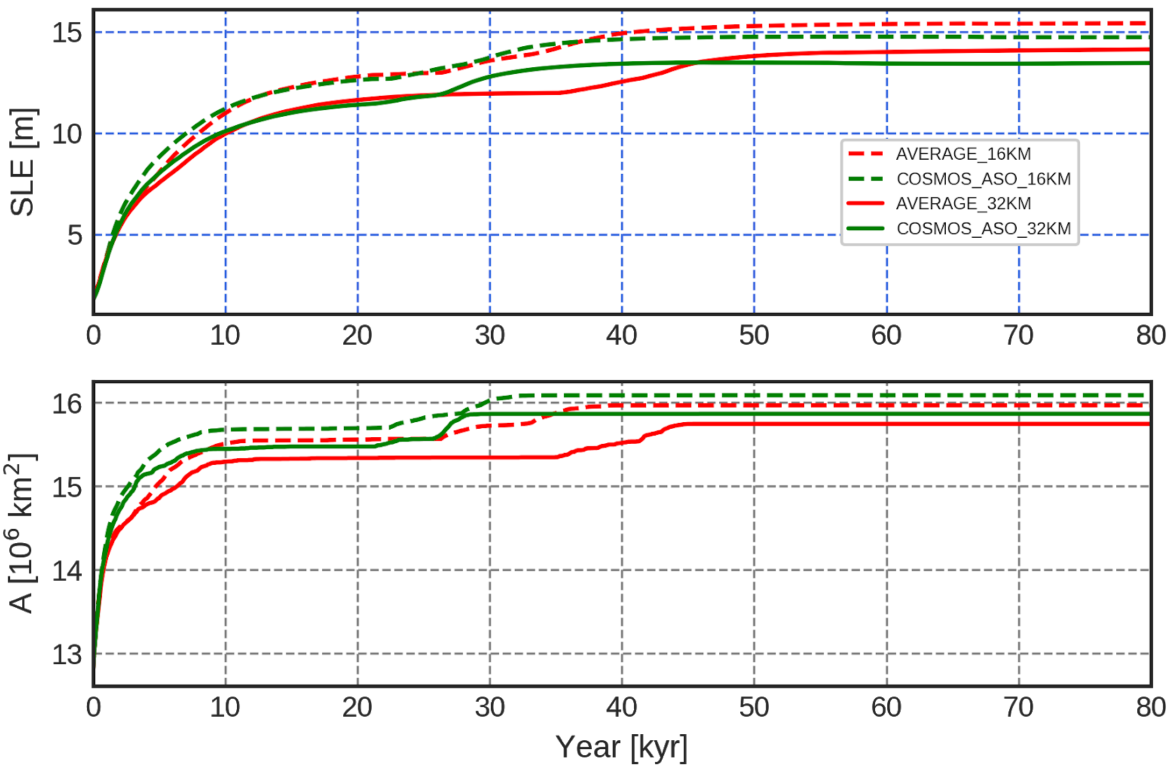

Figure 4: Ice volume evolution in terms of SLE and grounded ice area evolution for two PMIP3 members and the average with 32km and 16km resolution.

	V 32km [msle]	V 16km [msle]	A 32km [10^6 km^2]	A 16km [10^6 km^2]
AVERAGE	72.8	73.0	15.7	16.0
COSMOS-ASO	72.2	72.3	15.9	16.1
PD	58.7	57.6	12.9	13.0

Table summarizing the simulated ice volume (in msle) and grounded ice extension (in 10^6 km^2) for different resolutions.

The simulated LGM state for AVERAGE 16km and COSMOS-ASO 16km has a similar ice volume than for 32km resolution. However, the simulated PD state is smaller for 16km resolution than for 32km (around 1 msle), which creates a larger LGM ice volume anomaly as it is measured with respect to the simulated PD state. The simulated LGM state for 16km is more extended (0.3 and 0.2 million km^2 respectively) which allows for more ice accumulation, and results in a slightly larger LGM ice volume per se.

Overall, the simulated LGM snapshots are similar for both resolutions, with a similar ice thickness anomaly pattern (Figure 5, 6).

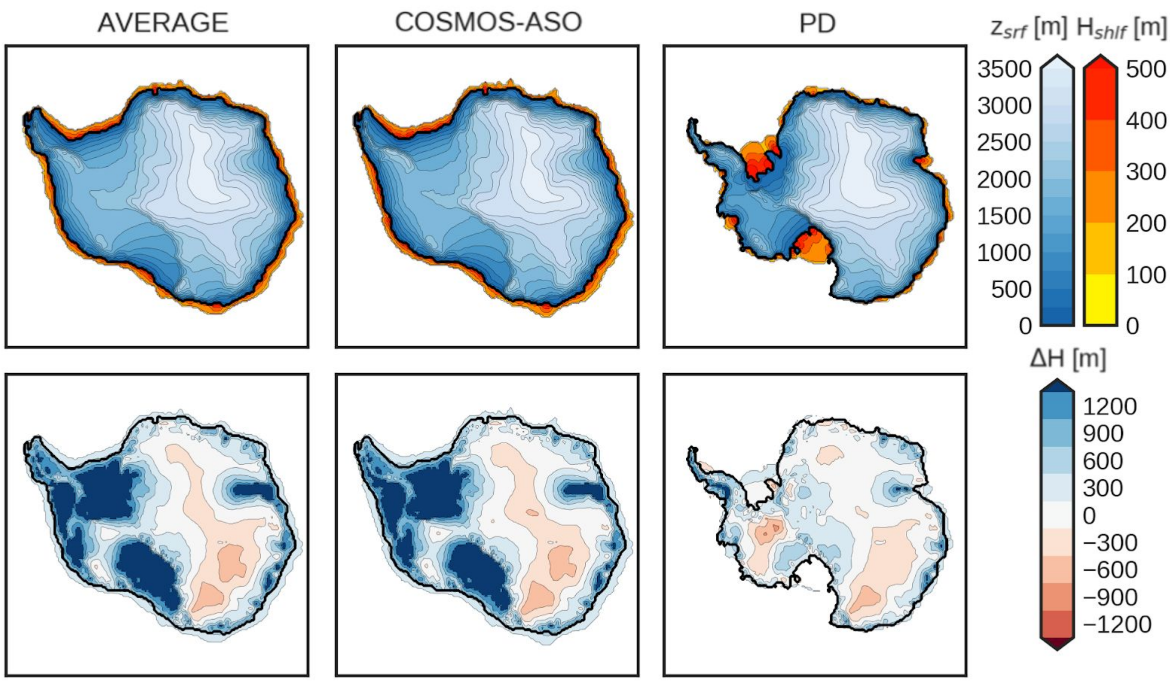

Figure 5: Upper row: Simulated surface elevation and ice shelf thickness for AVERAGE, COSMOS-ASO and PD at 32km resolution. Lower row: ice thickness anomaly with respect to the simulated PD (LGM-PD). In the case of the PD the ice thickness anomaly is drawn with respect to observations (simulated PD - observed PD).

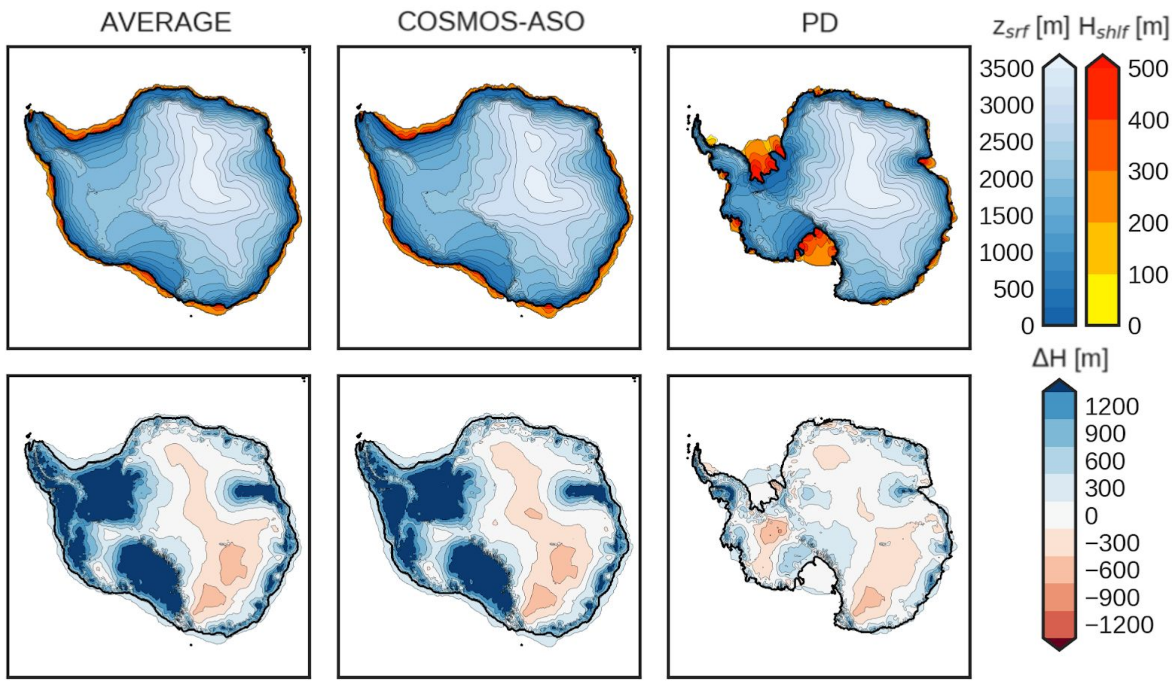

Figure 6: As Figure 5 but for a horizontal resolution of 16 km.

We added these Figures to the SM and added this paragraph in the discussion section:

“**Model limitations**

In this study we employed a coarse resolution of 32km. The simulation of large continental marine ice sheets has been found to be very sensitive to spatial resolution, especially at the grounding line (Pattyn et al., 2012). Grounding-line migration is a subgrid-scale process at such coarse resolutions. Ice-sheet models often use subgridding parameterisations to mimic higher resolutions at the grounding line. Nonetheless, even these parameterisations are often unable to trace the grounding-line migration correctly (Seroussi et al, 2014; Gladstone et al., 2017). Yelmo computes the fraction of grounded ice at the grounding line via subgrid and scales the basal friction at the grounding line with the grounded ice fraction (Robinson et al., 2019). To analyze the potential implications of a higher spatial resolution, we additionally performed two LGM experiments (namely AVERAGE and COSMOS-ASO) together with the simulated PD state at 16km.

We find that the simulated LGM state for a fully advanced AIS simulates a similar volume (a difference of 0.2-0.3 msle) and has a slightly larger extension (0.2 to 0.3 million km^2) for both resolutions (SM). Nonetheless, the simulated PD state is smaller for 16km resolution than for 32km (around 1 msle), which creates a larger LGM ice volume anomaly for 16km. Overall, the simulated pattern and grounding-line position is similar for both resolutions (SM). However, it is important to mention that the equilibrated state is reached at different times for different resolution (SM), pointing to the importance of resolution for assessing grounding-line migrations."

General comments text: The manuscript is generally well written but contains a couple of stilistic issues, redundancies, unclear sentences etc. of which I try to note a couple in the following:

Title: I think the title is a little misleading, as you do not explicitly simulate the Last Glacial Maximum Antarctic Ice Sheet configuration per-se but rather potential equilibrium states of the AIS under LGM conditions. Below is an attempt at a slightly modified title. Exploring the impact of atmospheric forcing and basal drag on Antarctic Ice Sheet equilibrium extent and volume under Last Glacial Maximum conditions.

We appreciate the suggestion, and changed the title from
"Exploring the impact of atmospheric forcing and basal boundary conditions on the simulation of the Antarctic ice sheet at the Last Glacial Maximum"
to
"Exploring the impact of atmospheric forcing and basal drag on the Antarctic ice sheet under Last Glacial Maximum conditions"

Abstract: I think one interesting outcome of this study is that the ensemble spread with regard to sea level equivalent ice volume change is about the same for the tested parameterisations of basal drag as for the different GCM forcings used (both ca. 6 m).This should be mentioned in the abstract and discussion.
We added in the Abstract:
"Overall, we find that the spread in the simulated ice volume for the tested basal drag parameterisations is about the same range as for the different GCM forcings (4 to 5 m)."

p3 l35: you assume a priori zero basal melt underneath ice shelves. For me this is fine,but how do you legitimize this choice? Relatively little is known about the state of CDW during the LGM, but I guess it is not to be excluded that regionally if the grounding line is located at sufficient depth, some basal melt is possible even during the LGM. Maybea reference would be helpful here.

We added the sentence:
"Ablation rates at the PD are almost zero except for localized areas (van Wessem et al., 2016, 2018). Because the LGM is a colder period, around 10 degrees as shown by ice core records (Jouzel et al., 2007), ablation rates in the LGM would have been probably negligible. On the other hand, basal melting rates from the LGM are difficult to estimate due to the scarcity of oceanic temperature reconstructions. Nonetheless, geomorphological records point to a fully advanced AIS during the LGM (The Raised Consortium, 2014). This could hint to low basal-melting rates inside the continental-shelf break."

We also added two paragraphs in the Discussion section saying:

"In this study, no basal melting was considered during the LGM. Of course, this is a vast simplification of reality. Unfortunately, reconstructions of ocean subsurface temperatures at the LGM are not available, so that the geological evidence for basal melt is lacking. As shown in Golledge et al., (2012), oceanic forcing leads to a dynamic response of LGM ice streams in the WAIS. If basal melt would have been considered, this would have most likely reduced the total LGM ice volume and affected its extension. Thus, our results represent an upper limit which would reduce when oceanic forcing is considered.

From the point of view of modelling, there have been some attempts to infer basal-melting rates. Kusahara et al., (2015) used a coupled ice-shelf-sea-ice-ocean model with a fixed LGM AIS extension, up to the continental-shelf break. In their model results, they obtained a larger basal melt value of ice shelves than PD. These large basal-melting rates occurred because the ice shelves were located at the edge of the continental-shelf break, where ice shelves are in contact with the warm CDW. However, these basal-melting values cannot be applied to the interior of the continental shelf as these waters do not penetrate so easily there. On the other hand, Obase et al., (2017) simulated basal-melting rates on an idealized PD AIS to investigate the response of basal melt rate to a changing climate. However, these basal-melting rates are not realistic and cannot be applied directly to the AIS as the grounding-line advances during the LGM affect the climatic conditions and subshelf melting. In order to investigate the impact of realistic basal-melting rates it would be necessary to account for comprehensive parameterisations or coupled ice-sheet-ocean models (Lazeroms et al., 2018; Reese et al., 2018; Favier et al., 2019; Pelle et al., 2020), which is outside of the scope of this study. Furthermore, since our aim was to simulate a fully advanced AIS, as suggested by geomorphological records (The Raised Consortium, 2014), basal-melting rates were set to zero for the sake of simplicity in this work."

p4 l 31: The SMB is obtained from the difference between ice accumulation …

Done

p5 l10: how does this relaxation time relate to other figures used in the field? How does it affect the results?

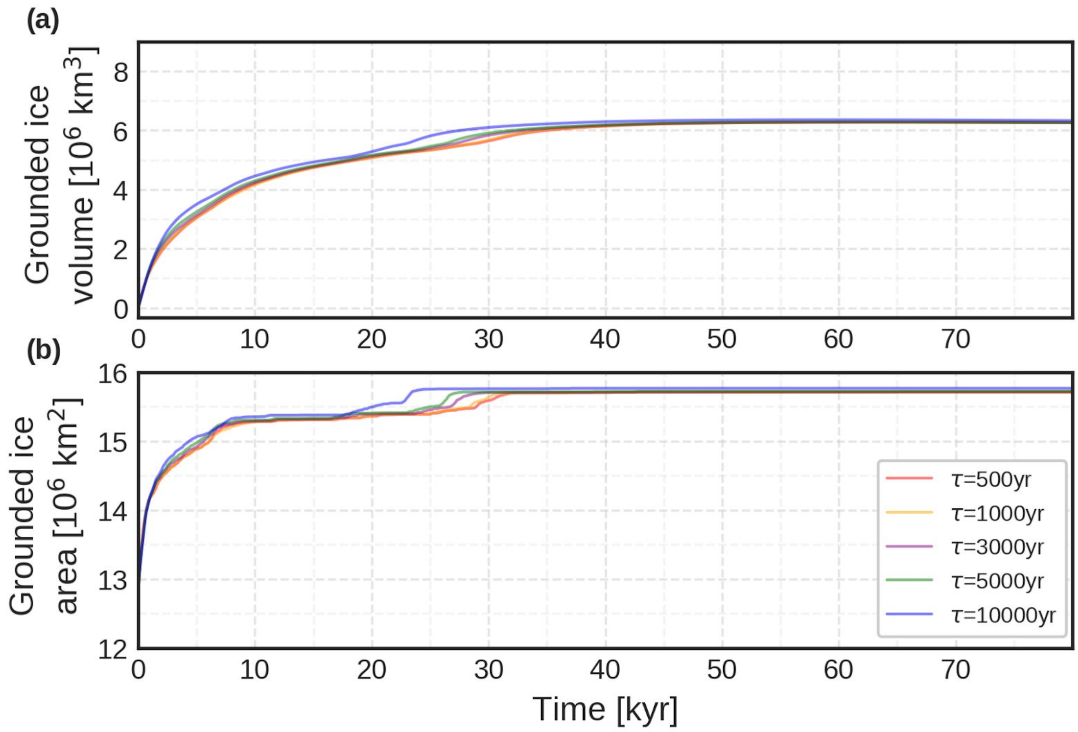

*Fig. 7 Evolution of the **(a)** grounded ice volume; **(b)** grounded ice area for different relaxation times (from 500 yr - 10000 yr) for the friction values z_0 = -125 m and c_{min} = 5*10^{-5} yr/m.*

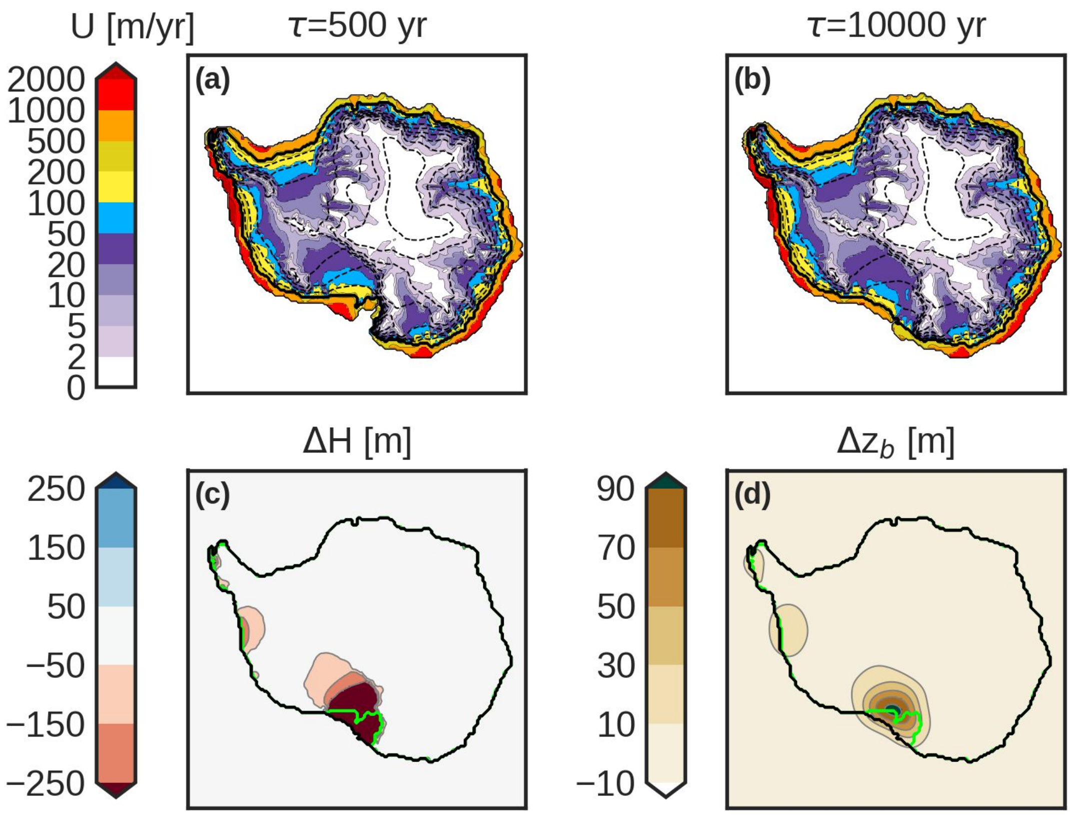

Fig.8: Simulated LGM AIS ice velocity for a relaxation time ***(a)*** *of τ=500 years;* ***(b)*** *τ=10000 years after 25000 years of cold climate evolution. Dashed lines represent surface elevation contours every 500 meters up to 3500 meters. The thick black line represents the grounding-line position.* ***(c)*** *ice thickness* ***(d)*** *bedrock elevation anomaly (****(a)*** *minus* ***(b)****). The thick green/black line represents the grounding-line of* ***(a)/(b)****.*

Figure 7 shows the time evolution of the grounded ice volume and ice area for the reference friction parameters and the Average PMIP3 fields. All simulations yield a similar equilibrated end-state. However, not all of them reach the continental-shelf break at the same time. Figure 2 shows the simulated LGM ice sheet for a weaker rheology (a; τ=500 years) and a stronger rheology (b; τ=10000 years) after 25000 years. These results show that a weaker rheology simulates a lower ice sheet in the WAIS and especially at the Ross shelf, where it does not fully advance (Fig. 2c). Comparing the bedrock elevation differences (Fig. 2d), a weaker rheology has a more elevated bedrock at the Ross shelf, the Bellingshausen Sea and the Amundsen sea, which impedes there a complete advance.

p5 l 11 ... so an "enhancement factor" is used …
Done

p5 l27: This sentence is a little confusing, maybe change to : "For lower values of z_0,c_b falls more rapidly ..."
Done

p6 l27: how realistic is this assumption? I guess in some regions basal freeze on could be extensive and for other regions basal melt is theoretically possible. A short discussion would be helpful.
Discussion added (see above).

p7 l 10-12: as mentioned in my comment 1.2, how does the large SL spread in the PD simulations affect the spread at LGM. What happens if you only account for those simulations with a spread of e.g. pm 1 m.
The new manuscript version only accounts for simulations with a spread of ±1 m.

p7 l22: suggest to change to: Here we present the simulated AIS equilibrium configuration under LGM conditions for different basal friction parameters.
Done

p7 l22: change to: Ice volume change is converted into …
Done

p7 l28 : change to : ...basal friction reduces basal sliding…
Done

p7 l29 : change to : ...also reduces ice volume…
Done

p7 l30 : suggest to change to: We do not identify a strong impact of marine basal friction on equilibrium grounded ice area, as the final grounding line configuration is similar in all ensemble members (Fig. 2b). Comment: is this mainly due to the long integration time? How quickly is the final ice extent reached?

Done. As mentioned above, this is partly due to the long integration time, however, assuming that the LGP occurred for almost 100kyr we think that it is realistic that all the simulated ice sheets fully expand at the LGM.

p8 l11 ...a slowly decreasing basal friction
Done

p8 l26: ...a spread of 6.2 msle.
Done

p8 l32: use other word than "appreciable", maybe "strong" ?
Done

p9 l6 ...identify the surface temperature …
Done

p9 l7 maybe change to: Whereas low surface temperatures lead to similar ice extend,relatively warm surface temperature forcing results in smaller equilibrium grounding line advance.
Done

p9 l8 change to: given the overall low surface temperature at LGM, ablation can generally be discarded as the …
Done

p9 l16 it is unclear here what these hypotheses are. I assume you mean:1. The more slippery the bed the farther the ice extend and the lower the volume. 2. the colder the surface temperature the larger the ice volume 3. the higher the precip the larger the ice volume. For the reader it would be nice if the authors main hypotheses are spelled out in the beginning.

We changed the paragraph to:

“The CNRM-CM5 model simulates the smallest AIS LGM for all the PMIP3 models. This model expands partly at the Ross shelf and Antarctic Peninsula zone, but collapses completely in the Ronne and Amery shelf, leading to ice free zones in the EAIS and a lower ice volume than the PD (Fig. 5). This occurs due to the presence of ablation in these regions (see SI, Fig. S8). Such a configuration is highly unlikely compared with sea-level and ice extension reconstructions from the LGM. We will discuss later possible explanations for this behaviour.

In summary, we find that the choice of the boundary climate conditions is crucial for the simulated LGM ice sheet. On one hand, the atmospheric temperatures near the coastal regions control the ice extension through viscosity. If the viscosity is low, then the ice flows too fast, preventing the necessary thickening for advancing towards the continental-shelf break. Particularly, if the bedrock is too deep, the ice sheet’s expansion will be hampered. Secondly, if the ice sheet extends close to the continental-shelf break, then the accumulation pattern will determine the total amount of ice volume. We find that for fully extended ice sheets (IPSL-CM5A-LR and MRI-CGCM3), the sea-level difference due to accumulation differences is about 4.2 msle.”

p9 l20 what is abnormal? change phrasing.
Done

p9 l20 suggest to omit "unexpected"
Done

p9 l22 ... the regions with grounding line advance to the continental shelf break (e.g.the Ross Basin)... Comment: how do they contribute to low ice temperature? Due to lapse rate effects? Clarify.
This part was changed in the new manuscript, however, the lower ice temperatures occur partly due to lapse rate effects but also to the employed PMIP3 field, which can have warmer temperatures at the coastal regions.

p9 l27 too low for what? Suggest to change to: If the viscosity is low …
Done

p9 l28 necessary for what?? grounding line advance? I suggest to rephrase the whole last paragraph, beginning at "In summary". You mix climate effects on the ice sheets rheology with topographic effects due to bedrock configurations and the location of the continental shelf break under the header of "Impact of climate forcing". The last sentence provides an important finding as it shows the impact of different SMB regimes under similar ice sheet configurations.
The new paragraph reads now:
“In summary, we find that the choice of the boundary climate conditions is crucial for the simulated LGM ice sheet. Atmospheric temperatures have a direct impact on the ice flow of ice sheets. Warmer temperatures lead to lower ice viscosities, enhancing ice flow. A faster flow leads to thinner ice.

On one hand, the atmospheric temperatures near the coastal regions control the ice extension through viscosity. If the viscosity is too low, then the ice flows too fast, preventing the necessary thickening. Particularly, if the bedrock is too deep, the ice sheet’s expansion will be hampered. Secondly, if the ice sheet extends close to the continental-shelf break, then the accumulation pattern will determine the total amount of ice volume. We find that for similarly extended ice sheets (IPSL-CM5A-LR and MRI-CGCM3), the sea-level difference due to accumulation differences is about 3.5 msle”

p9 l33. I think at the current state of art in the field it is unclear what approach is "valid"given the large persisting uncertainties in paleo ice sheet modelling (as well as ice sheet projections). I therefore suggest to rephrase to: ... is a common approach…
Done

p10 l3. As of yet it is unclear what a "realistic" SLE is, this is something you rightly state at the beginning of the manuscript. Therefore I suggest to rephrase to : All simulations produce SLE ice volume in the range of previously suggested figures and ice extend similar to reconstructions (e.g. Bentley et al. 2014) if using the same coefficients for basal friction and different climate forcings. Overall, consistently …
Done

p10 l4 change to: This is solely due to the difference in forcing, as the parameterisation of ice flow is identical.
Done

p10 l5 change to: Since surface temperatures are not sufficient to cause surface melt,differences in ice volume and extent are exclusively due to differences in accumulation anomalies.
Done

p10 l7. It is evident that the main source of ice volume differences is due to changes in the WAIS configuration.
Done

p10 suggest to change header of 4.1 to: "Role of basal friction" or similar
Done

p10 l21 : between the end members
Done

p10 l25 I know what you mean with "still agree with PD observations" but I suggest to rephrase the sentence or split it in two.
We changed the paragraph to “The dynamical state of the LGM remains a source of uncertainty as there are no observations from that time period of the AIS configuration. To study potentially possible AIS LGM dynamical states, we covered a range of friction values which lead to realistic LGM and PD configurations. [...] For example, an AIS that extends up to the continental-shelf break, but with a relatively low volume increase, can be achieved through a very dynamically active ice sheet. In that case, marine-based regions, and more specifically the WAIS, have the potential to maintain fast ice streams at the LGM.”

p10 l26 change to: The choice of the friction law …
Done

p10 l30 suggest to change wording to: ...is especially relevant…
To avoid the confusion pointed out by one of the reviewers, we changed the sentence to:
“However, the aim of this work was to study the uncertainty associated with the bedrock friction parameter, rather than assessing the uncertainty for different friction laws.”

p11 l7 The simulated grounding line advance is strongly influenced by air temperature.
Done

p11 l12 if temperatures are sufficiently cold ($< 20^\circ$C) ice full advances …
Done

p11 l13 The RAISED consortium shows a similar grounding line extend, albeit with two large ice shelves ... Comment: to my knowledge Bentley et al. show grounding line extend but not the presence of ice shelves but I might be mistaken? Please clarify.
Indeed, they show grounding-line extent but no ice shelves. We changed the sentence to:
“The RAISED Consortium has a similar extension, but presents retreated areas at the margins of the Ronne shelf, which we are not able to simulate.”

p11 l29 Overall, homogenous climate anomaly-forcing relative to present day leads to a …
Done

p11 l32 Thus, recent paleo ice sheet model exercises utilise climate forcing derived from GCMs
Done

p11 l1 Nevertheless, ... resulting from different assumptions of basal drag.
Done

p12 l4 change to: By design the modelled ice sheet could be expected to be driven towards the configuration used as a boundary condition in PMIP3. However, …
p12 l6 ...the comparison with proxy-observations.
p12 l8 ... more accurate paleo-climate forcing will hopefully be available.
This last part has changed as suggested by another reviewer.

p12 l21 Imposing the PMIP3 fields, which explicitly assume an LGM ice sheet configuration, leads to higher preci…
We deleted "whose climate simulations include dynamic adjustment to the LGM boundary conditions," as this is not that simple. As pointed out by another reviewer "On one hand, the very thick and extensive PMIP3 LGM ice sheet can induce a drastic expansion of LGM AIS due to the large decrease in surface air temperature. However, on the other hand, the thick ice sheet will reduce the amount of precipitation, which will cause a thinning of LGM AIS, opposite to PMIP3 LGM ice sheet"

p12 l24 I guess you mean WAIS not AIS here? You show in your results that the uncertainties regarding basal conditions are as high as the uncertainties regarding climate forcing, this should be restated in the conclusions as I think this is an important finding of this work.
Yes, we meant WAIS there. We added the sentence "Our results show that the uncertainty in sea-level LGM estimates due to basal drag is similar to the uncertainty resulting from the background climatic conditions derived from PMIP3.".

General comments figures:
Figure 3. Cmap different to read. Suggest to use simpler colormap (e.g. Red-Blue)and plot ice thickness changes relative to PD. For the surface contours I suggest using one color (e.g. gray).
Done. Surface contours were changed to discontinuous black lines.

Figure 4. The figure size seems overly large given that it shows less information than the following figures.
This figure was moved to SM.

Figure 5. Tough too discern features with this color scale, I suggest something simpler (e.g. Red-Blue or similar) and plotting delta thickness with respect to present day observations instead of LGM surface elevation. This way it is easier to identify regional changes caused by the different GCM-forcings.
Done

Figure 7. With the "jet/rainbow" color scale it is tough to discern between different ensemble members, I suggest different marker styles ("x o , ." etc) for each GCM in addition to the colors.
Done (also for Figure 6)

Figure 8. same as Figure 5. Suggest different color scale and delta thickness instead of surface elevation. You can keep the surface contours for reference. Why is ice thickness lower in the coastal regions of the Bellinghausen Seas for the PMIP3av even though accumulation is higher? It can’t be basal shelf melt as this is set to zero?
Done. These results have slightly changed with the new Yelmo version.

========= Reviewer comment 3 (anonymous) =============

I think the content of this study matches the interest of the reader of The Cryosphere. Furthermore, as a climate modeler, I find this result quite interesting, and think it offers valuable information to both ice sheet and climate communities. Below, I address several concerns mostly focusing on the discussion of the results.

We thank the thoughtful and constructive review from Referee#3. Below you can find our response to each comment. We would like to point out that simulations have been repeated with an improved version of the Yelmo model (Robinson et al., 2020). Although specific values for each simulation have slightly changed, our main conclusions remain robust.

General comments:

1. I think the authors should discuss the uncertainty in the glacial atmospheric forcing arising from the ice sheet configuration used in PMIP3 LGM simulation. In the PMIP3 LGM simulations, the climate models are forced with PMIP3 LGM AIS, which has a volume of 22.3 meter SLE compared with PI (Abe-Ouchi et al. 2015). This value largely overestimates the reconstructed value of the LGM AIS (Less than + 15 m SLE), and causes an inconsistency between the LGM AIS used for climate model simulations and the simulated LGM AIS with the ice sheet model. Therefore, the author should address this problem, and suggest the climate modeling community to perform LGM simulations with a more realistic AIS, which matches the reconstruction. In addition,I have a comment on a sentence starting from P12L5 “ A way to potentially test the plausibility of the employed climatic fields is to compare with ice proxies.” I agree to this sentence, but again, the inconsistency in the LGM AIS used in climate models and the reconstructed LGM AIS bothers me. For example, even if some PMIP3 glacial atmospheric forcing show consistent results with available ice core data, and regarded as reasonable glacial atmospheric forcing, I don’t think it is physically correct. Please add a discussion on this point in section 4.3.

Indeed, the employed LGM AIS is clearly larger, not only than the simulated in this work, but also in comparison with other recent studies (Simms et al., 2019).

We added a paragraph in the Discussion section:

“The cryosphere is a component of the Earth System that also interacts with other components, such as the atmosphere or the ocean. Therefore the configuration of the AIS (as well as other ice sheets) for the PMIP3 LGM simulations plays a crucial role in LGM climatologies. We note that the PMIP3 LGM simulations were forced with an AIS with an ice volume of 22.3 msle compared to PI (Abe-Ouchi et al., 2015). This ice volume largely overestimates the AIS volume change inferred from the latest studies (Simms et al., 2019). It is clear that a significant larger AIS will create a colder and drier environment than a smaller ice sheet. In order to compare with PMIP3 results, the first preliminary results of PMIP4 are forced with the same AIS LGM configuration (Kageyama et al., 2020). Nonetheless, given the fact that the latest studies point to a lower ice volume, new PMIP experiments should consider the effect of a fully advanced, but smaller AIS. The alternative would be to employ fully coupled ice-sheet--climate models to simulate both the LGM climatologies and the LGM ice sheets.”

With deleted the sentence:

“A way to potentially test the plausibility of the employed climatic fields is to compare with ice proxies.”

We agree with the reviewer’s opinion that in order to compare the model output with ice proxies it is first necessary to have consistency between the employed LGM AIS for climate models and reconstructions.

2. The basal melting of the ice shelf is fixed to zero in the simulations. I think this is a reasonable simplification to focus on the main topic of this study, however you should at least discuss the potential effect of the simplification you made. For example, while Obase et al. (2017, JCLIM) show that the basal melting at the LGM largely reduced compared with PI in their simulations with regional ocean model, the basal melt of LGM was still more than 50% of the PI experiment. Based on their estimates, the simulated area and volume in your experiment can be considered as the maximum estimate, and that the

uncertainties in the ice shelf basal melting can have an impact on the LGM AIS. Please add a discussion on this topic.

Indeed, adding basal melt would add a degree of difficulty. As shown in Golledge et al., (2012), oceanic forcing leads to a dynamic response of rapid ice streams, especially in the WAIS. Thus, including basal melt would most likely reduce the total ice volume and potentially affect the ice extension.

We found two studies that particularly address the problem of basal melting rates during the LGM, namely Obase et al., (2017) and Kusahara et al., (2015). Kusahara et al. (2015) used a fully advanced AIS to estimate the basal melting rates of ice shelves located at the border of the continental-shelf break. The basal-melting rates obtained were higher than PD values due to a greater exposure of the ice shelves to warm CDW. Nonetheless, these ice shelves are located at the continental-shelf break and these high melting rates do not necessarily apply in the interior of the continental shelf because of a more limited penetration of CDW. In our experimental setup the applied basal-melting rate refers to the interior of the continental shelf.

On the other hand, Obase et al. (2017) applied LGM conditions to an idealized PD configuration to investigate the response of basal-melt rates to a changing climate. However, as they pointed out, the changing basal mass balance actually modifies the thickness of the ice shelf and the positions of the grounding lines. This in turn affects the sea ice and the ocean around the ice shelves, which affects basal melting. Therefore, these melting values inside the continental-shelf are not valid for our experimental setup. We choose to set the basal-melting rates to zero to allow for a maximum ice extent.

We added two paragraphs in the Discussion section saying:
"In this study, no basal melting was considered during the LGM. Of course, this is a vast simplification of reality. Unfortunately, reconstructions of ocean subsurface temperatures at the LGM are not available, so that the geological evidence for basal melt is lacking. As shown in Golledge et al., (2012), oceanic forcing leads to a dynamic response of LGM ice streams in the WAIS. If basal melt would have been considered, this would have most likely reduced the total LGM ice volume and affected its extension. Thus, our

results represent an upper limit which would reduce when oceanic forcing is considered.

From the point of view of modelling, there have been some attempts to infer basal-melting rates. Kusahara et al., (2015) used a coupled ice-shelf-sea-ice-ocean model with a fixed LGM AIS extension, up to the continental-shelf break. In their model results, they obtained a larger basal melt value of ice shelves than PD. These large basal-melting rates occurred because the ice shelves were located at the edge of the continental-shelf break, where ice shelves are in contact with the warm CDW. However, these basal-melting values cannot be applied to the interior of the continental shelf as these waters do not penetrate so easily there. On the other hand, Obase et al., (2017) simulated basal-melting rates on an idealized PD AIS to investigate the response of basal melt rate to a changing climate. However, these basal-melting rates are not realistic and cannot be applied directly to the AIS as the grounding-line advances during the LGM affect the climatic conditions and sub-shelf melting. In order to investigate the impact of realistic basal-melting rates it would be necessary to account for comprehensive parameterisations, such as PICO or PICOS (Reese et al., 2018; Pelle et al., 2019), or coupled ice-sheet-ocean models. This is out of the scope of this study. Furthermore, since our aim was to simulate a fully advanced AIS, as suggested by geomorphological records (The Raised Consortium, 2014), basal-melting rates were set to zero for the sake of simplicity in this work."

3. It is interesting to see that the differences in glacial atmospheric forcing caused large discrepancies in the simulated LGM AIS, especially for that of CNRM5. While it is not the main topic of this study to understand the cause of the difference in atmospheric forcing, I think it is valuable to discuss some possible reasons. For example,the result of CNRM5 reminds me of a study by Marzocchi and Jansen (2017, GRL)who compared the sea ice among PMIP3 LGM simulation. In their Fig. 3, you can find that CNRM5 simulates the smallest austral summer sea ice extent in LGM among PMIP3 models. This will cause warmer summer temperature over the marginal region of AIS and contribute to the negative mass balance. Perhaps, you may add one or two sentences on this point.

The CNRM5 model simulates the warmest LGM temperature, not only for the Antarctic domain, but this was also found in the NH (Niu et al., 2019). Our new results show even a smaller AIS due to the abnormal presence of ablation (Supplementary Material). In fact, in Kageyama et al., (2020) this model is represented as an outlier. We did not investigate the possible reasons for this, but truly Fig. 3 from Marzocchi and Jansen (2017) could hint to a potential explanation for these warm temperatures.

We added in the Discussion section:
"Nonetheless, it seems unrealistic that air temperatures were high enough to produce ablation during the LGM as seen in CNRM-CM5. The model CNRM-CM5 simulates the warmest LGM temperatures not only in the SH, but it has been also shown to simulate the lowest LGM volumes for the NH (Niu et al., 2019). A potential explanation for this behaviour can be due to sea-ice formation. As shown in Marzocchi and Jansen (2017), the CNRM-CM5 model simulates the lowest austral sea-ice extent. Such a low extent would increase surface temperatures through sea-ice albedo feedback. Hence, this could point to sea-ice formation as a crucial element in driving fully LGM conditions."

Specific comments:
P2L30-31: This sentence describes several processes, which affect the estimate of the volume of LGM AIS. However it is unclear how the modifications affect the estimate.Please add some explanations on this point. You may focus on one or two processes,which are relevant to this study.

The new sentence reads now:
"Whereas older studies estimated large sea-level contributions generally above 15 m (e.g. Nakada et al. (2000);Huybrechts (2002); Peltier and Fairbanks (2006); Philippon et al. (2006); Bassett et al. (2007)), more recent modelling studies and reconstructions have lowered these estimates to 7.5-13.5 m (Mackintosh et al., 2011; Whitehouse et al., 2012a; Golledge et al., 2012, 2014; Gomez et al., 2013; Argus et al., 2014b; Briggs et al., 2014; Maris et al., 2014; Sutter et al., 2019). This lowering in ice volume can be explained by the fact that the first ice-sheet models were based purely on the Shallow Ice Approximation (SIA) for inland ice. This solution solves for slow moving ice, based on shear deformation. However, later models include more sophisticated approximations (e.g. Shallow Shelf

Approximation, Full Stokes) with a better representation of fast flowing ice streams. These fast flowing regions contribute to a decrease in ice volume. Nevertheless, the latest LGM AIS volume estimates still differ by more than 5 msle. Part of this difference can be explained by spatial resolution and sub-grid scale grounding-line treatment (e.g. Goelzer et al. 2017; Pattyn 2018). Other possible explanations include the implementation of external processes, like the GIA (e.g., Whitehouse et al., 2019), or, as this work, the effect of uncertain climatologies and ice-sheet dynamics."

P3L3: I mostly agree with this sentence, but is it really true that the basal melting is negligible during LGM? For example, Obase et al. (2017, JCLIM) showed with regional ocean model that there is still some basal melting occurring at LGM. Please modify this sentence in a more modest way.

This issue was discussed above. The new sentence reads now:
"Ablation rates at the PD are almost zero except for localized areas (van Wessem et al., 2016, 2018). Because the LGM is a colder period, around 10 degrees as shown by ice core records (Jouzel et al., 2007), ablation rates in the LGM would have been probably negligible. On the other hand, basal melting rates from the LGM are difficult to estimate due to the scarcity of oceanic temperature reconstructions. Nonetheless, geomorphological records point to a fully advanced AIS during the LGM (The Raised Consortium, 2014). This could hint to low basal-melting rates inside the continental-shelf break."

P4L10: This sentence is difficult to read. Do you mean that in some models, the simulated results largely differ from ice core reconstructions? Please modify this sentence.

Yes, that is what we meant to say. The new sentence reads

"However, in some models, the simulated results differ from ice core reconstructions (Cauquoin et al., 2015). This may lead to an unrealistic configuration and thus it is necessary to evaluate the accuracy of model outputs."

P6L15-20: I had difficulty understanding this sentence, since I'm not familiar with an ice sheet model. Please describe this sentence in more detail. Why

do you use PD temperature field at sea level rather than surface? How does RACMO calculate the sea level temperature field? Do they assume a constant lapse rate in converting the temperature? If so, is the value of the lapse rate identical to what you chose in your ice sheet model?
P6L19: How did you decide this value of the lapse rate? Do you have any reference for this?

RACMO does not compute temperatures at sea-level, but at the surface. Yelmo, as well as other ice-sheet models, uses a lapse rate to scale the temperatures down to sea level and then scale them back up to the simulated surface elevation, to take into account changes in temperature and precipitation due to the elevation. Hence, Yelmo needs the surface temperatures simulated by RACMO as well as the PD surface elevation, to convert these temperatures to sea level. The same occurs for the LGM climatologies: the LGM surface elevations provided by Abe-Ouchi et al., (2015) are needed to correct the climatologies with the elevation

The lapse rate value is an imposed value in ice-sheet models. It is not a uniform value over the whole Antarctic domain, but ranges from 0.015 K/m in most interior regions to 0.005K/m in the coastal zones (Fortuin and Oerlemans, 1990). Ice-sheet models commonly set this value to 0.008 K/m over the whole continent for simplicity (DeConto and Pollard, 2016; Quiquet et al.,2018; Albrecht et al., 2020). It accounts for the fact that changes in surface elevation imply also a change in temperature (colder temperatures at higher elevations). In order to improve the employed methodology we take into account changes in humidity by imposing two values, one for summer and another for annual temperatures.

The new paragraph reads:

“We apply a lapse rate correction that accounts for LGM minus PD changes in elevation (0.008K $m^{-1}$for annual temperatures and 0.0065K $m^{-1}$for summer temperatures) in concordance with other ice-sheet models (Ritz et al., 1997, DeConto and Pollard, 2016; Quiquet et al.,2018; Albrecht et al., 2020).”

P8L14: Are these results consistent with previous studies?

Yes, these results are similar to the basal sliding map of Golledge et al. (2012). We added the sentence
“These zones of fast flowing areas are similar to the predicted occurence of basal sliding from Golledge et al. (2012).”

P9L6: How did you define ground line temperature? Does the location of grounding line depend on simulations?

The grounding-line temperature is defined as the mean temperature of the ice column of all grounding-line points. The location of the grounding line is defined in Yelmo through the flotation criterium, hence it is different for each simulation.

The new sentence reads:
“Further inspection allows us to identify the atmospheric temperature close to the grounding line (Fig 7d) as a critical factor in determining how far the AIS advances. The grounding-line temperature is defined as the mean temperature of the ice column for all the grounding-line grid points.”

P11L20-25: Please add a discussion on the role of basal melting in this subsection.
Done, we added there the paragraph of General Comment #2.

P12L1-2: I like this finding.
Thank you. As suggested by another reviewer we have highlighted this finding in the abstract.

P12L4-5: I don’t think it’s that simple. On one hand, the very thick and extensive PMIP3 LGM ice sheet can induce a drastic expansion of LGM AIS due to the large decrease in surface air temperature. However, on the other hand, the thick ice sheet will reduce the amount of precipitation, which will cause a thinning of LGM AIS, opposite to PMIP3 LGM ice sheet.
Indeed, a thicker ice sheet will tend to produce a colder and drier climate which can hamper the formation of a large ice sheet. We removed this last part of the article and finished with the discussion about the role of the employed LGM AIS in the CMIP3 experiments.

P12L8: You may cite a recent article by Kageyama et al. (2020, Climate Past Discussion), which discusses preliminary results of PMIP4 LGM experiments.
Done, thanks for pointing this work out.

Fig. 5: It's hard to see the contour of the surface topography. Please modify this figure.
Done

FIg.S1: I think this figure contains some important information on the reproducibly of modern Antarctic ice sheet. Please move it to the main manuscript
Done

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Exploring the impact of atmospheric forcing and basal ~~boundary conditions~~ drag on the ~~simulation of the~~ Antarctic ice sheet ~~at the~~ under Last Glacial Maximum conditions

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Abstract. Little is known about the distribution of ice in the Antarctic ice sheet (AIS) during the Last Glacial Maximum (LGM). Whereas marine and terrestrial geological data indicate that the grounded ice advanced to a position close to the continental-shelf break, the total ice volume is unclear. Glacial boundary conditions are potentially important sources of uncertainty, in particular basal friction and climatic boundary conditions. Basal friction exerts a strong control on the large-scale dynamics of the ice sheet and thus affects its size, and is not well constrained. Glacial climatic boundary conditions determine the net accumulation and ice temperature, and are also poorly known. Here we explore the effect of the uncertainty in both features on the total simulated ice storage of the AIS at the LGM. For this purpose we use a hybrid ice-sheet-shelf model that is forced with different basal-drag choices and glacial background climatic conditions obtained from the LGM ensemble climate simulations of the third phase of the Paleoclimate Modelling Intercomparison Project (PMIP3). Overall, we find that the spread in the simulated ice volume for the tested basal drag parameterisations is about the same range as for the differente GCM forcings (4 to 6 m sea level equivalent). For a wide range of plausible basal friction configurations, the simulated ice dynamics vary widely but all simulations produce fully extended ice sheets towards the continental-shelf break. More dynamically active ice sheets correspond to lower ice volumes, while they remain consistent with the available constraints on ice extent. Thus, this work points to the possibility of an AIS with very active ice streams during the LGM. In addition, we find that the surface boundary temperature field plays a crucial role in determining the ice extent through its effect on viscosity. For ice sheets of a similar extent and comparable dynamics, we find that the precipitation field determines the total AIS volume. However, precipitation is highly uncertain. Climatic fields simulated by climate models show more precipitation in coastal regions than a spatially uniform anomaly, which can lead to larger ice volumes. ~~We~~ Our results strongly support using these paleoclimatic fields to simulate and study the LGM and potentially other time periods like the Last Interglacial. However, their accuracy must be assessed as well, as differences between climate model forcing ~~lead to a range~~ leads to a large spread in the simulated ice volume and extension~~of about 6 sea-level equivalent and one million~~.

1 Introduction

Sea-level variations on long timescales are driven by the waxing and waning of large continental ice sheets. The characterisation of the sensitivity of ice sheets to past climate changes is fundamental to gaining insight into their underlying dynamics as well as their response to future climate change. In addition, understanding past sea-level changes is important for quantifying sea-level rise (Nicholls and Cazenave, 2010; Defrance et al., 2017; King and Harrington, 2018; Golledge et al., 2019; Robel et al., 2019) and for assessing the risk of crossing tipping points within the Earth System, such as the collapse of the West Antarctic Ice Sheet (Kopp et al., 2009; Sutter et al., 2016; Pattyn et al., 2018).

The Antarctic Ice Sheet (AIS), in particular, plays a fundamental role as it is the largest ice sheet on Earth and stores ca. 58 meters of sea-level equivalent (msle; Fretwell et al. (2013)). Due to its size it is potentially the largest contributor to future sea-level projections, but it is also the most uncertain (Collins et al., 2013). Assessing the AIS contribution to the total sea-level budget at different time periods has proven to be challenging. The Last Glacial Maximum (LGM, 21 ka BP) represents an ideal benchmark period since there is a large availability and variety of proxy data that, furthermore, indicate important AIS changes relative to present day (PD). Both, marine and terrestrial geological data, indicate that at the LGM, the AIS extended ~~up~~ to the continental-shelf break (Anderson et al., 2002, 2014; Hillenbrand et al., 2012, 2014; The RAISED Consortium, 2014; Mackintosh et al., 2014). However, its exact extent is not well constrained everywhere. Whereas its advance in the Amundsen region, the Bellingshausen Sea and the Antarctic Peninsula is well established, in the Ross Sea and the East Antarctic region it remains controversial (Stolldorf et al., 2012; The RAISED Consortium, 2014). Furthermore, the total AIS ice volume is even less well constrained (Simms et al. (2019) and references therein). Geological data furthermore do not provide direct information on past thickness and volume of ice sheets, which must hence be inferred. There have been several approaches to infer past ice-volume change of an individual ice sheet as the AIS. One approach is to use direct ice-sheet modelling to simulate the volume of the AIS at the LGM (e.g Huybrechts (2002); Whitehouse et al. (2012a); Golledge et al. (2012); Gomez et al. (2013); Maris et al. (2014); Briggs et al. (2014); Quiquet et al. (2018)). An alternative is to use Glacial Isostatic Adjustment (GIA) modelling, which describes the viscous response of the solid Earth to past changes in surface loading by ice and water (e.g. Ivins and James (2005); Bassett et al. (2007)). This approach has also been used in combination with direct ice-sheet modelling (e.g. Whitehouse et al. (2012b)) and/or by making use of constraints on ice-thickness from reconstructions based on exposure age dating, as well as satellite observations of current uplift (Whitehouse et al., 2012b; Ivins et al., 2013; Argus et al., 2014b). Whereas older studies estimated large sea-level contributions generally above 15 m (e.g. Nakada et al. (2000); Huybrechts (2002); Peltier and Fairbanks (2006); Philippon et al. (2006); Bassett et al. (2007)), more recent modelling studies and reconstructions have lowered these estimates to 7.5-13.5 m (Mackintosh et al., 2011; Whitehouse et al., 2012a; Golledge et al., 2012, 2014; Gomez et al., 2013; Argus et al., 2014b; Briggs et al., 2014; Maris et al., 2014; Sutter et al., 2019). ~~Several factors have contributed to a decrease in the estimate of the LGM AIS volume. On one hand, the state of the art of~~ This lowering in ice volume can be explained by the fact that the first ice-sheet ~~modelling has considerably advanced in the last years, for example through the inclusion of more complex physics, increased~~ models were based purely on the Shallow Ice Approximation for inland ice. This solution solves for slow moving ice, based on shear deformation.

However, later models include more sophisticated approximations (e.g. Shallow Shelf Approximation, Full Stokes) with a better representation of fast flowing ice streams. These fast flowing regions contribute to a decrease in ice volume. Nevertheless, the latest LGM AIS volume estimates still differ by more than 5 m. Part of this difference can be explained by spatial resolution and sub-grid scale grounding-line treatment (e.g. Goelzer et al. (2017); Pattyn (2018)). ~~On the other hand,~~ Other possible explanations include the implementation of external processes, like the ~~ice-ocean interaction or the GIA , are now treated with more accurate parameterisations and models~~ GIA (e.g., ~~Reese et al. (2018); Whitehouse et al. (2019)). Nevertheless, the latest LGM AIS volume estimates still differ by more than 5 msle~~Whitehouse et al. (2019)), or, as this work, the effect of uncertain climatologies and ice-sheet dynamics.

~~Given that ablation and basal melting were probably negligibleat the LGM in the AIS,~~ Ablation rates at the PD are almost zero except for localized areas (van Wessem et al., 2016, 2018). Because the LGM is a colder period, around 10 degrees as shown by ice core records (Jouzel et al., 2007), ablation rates in the LGM would have been probably negligible. On the other hand, basal melting rates from the LGM are difficult to estimate due to the scarcity of oceanic temperature reconstructions. Nonetheless, geomorphological records point to a fully advanced AIS during the LGM (The RAISED Consortium, 2014). This could hint to low basal-melting rates inside the continental-shelf break. Therefore ice-sheet dynamics and accumulation must have been the two main factors controlling ice-mass gain during this period. The representation of ice dynamics in ice-sheet models is a key feature that can potentially lead to important discrepancies. Most ice-sheet models simulating the past long-term evolution of large-scale ice sheets are hybrid models that rely on the Shallow Ice Approximation (SIA) and the Shallow Shelf Approximation (SSA). Moreover, there is no universally accepted friction law, and basal friction is treated in different manners in ice-sheet models. Ritz et al. (2015) emphasize the importance of the basal friction, as it can favour the occurrence of the marine instability in future AIS projections. Generally, basal stress follows either a power-law formulation on the basal ice velocity (a special case being the Weertman (1957) friction law) or a Coulomb friction law (Schoof, 2005) with different power-law coefficients, a friction coefficient and potentially a regularization term. Ice-sheet models thus use friction formulations that can range from linear viscous and regularized Coulomb friction laws, typical of hard bedrock sliding (Larour et al., 2012; Pattyn et al., 2013; Joughin et al., 2019) to Coulomb-plastic deformation, characteristic of ice flow over a soft bedrock with filled cavities (Schoof, 2005, 2006; Nowicki et al., 2013). In the simplest cases a constant friction coefficient is prescribed over the whole domain (Golledge et al., 2012), but generally this parameter incorporates the dependency of basal friction on the effective pressure exerted by the ice, as well as on bedrock characteristics by making use of assumed till properties (Winkelmann et al., 2011; Albrecht et al., 2019; Sutter et al., 2019) or basal temperature conditions (Pattyn, 2017; Quiquet et al., 2018). The sensitivity of the simulated ice volume to these features is substantial. For instance, Briggs et al. (2013) obtained differences of more than 5 msle for an Antarctic LGM state depending only on the friction coefficients used for hard and soft beds. Some studies have attempted to overcome the uncertainty in basal friction by optimising the friction coefficient through inversion methods in order to obtain an accurate PD ice-sheet state (Morlighem et al., 2013; Le clec'h et al., 2019). However, these optimizations are based on a particular configuration of the PD state, and it is unclear whether they remain valid for glacial conditions. All in all, basal friction is poorly characterised, and the potential consequences of the associated uncertainty should be considered in ice-sheet modeling.

Glacial atmospheric boundary conditions over Antarctica are also far from being well constrained. It is clear from ice-core records and marine deep-sea sediment data that, at the continental scale, temperatures were lower than today and that the climate was drier (Frieler et al., 2015; Fudge et al., 2016). Typically, ice-sheet models use two approaches for simulating the atmospheric conditions at the LGM. On one hand, some studies prescribe a spatially-uniform temperature anomaly (generally between 8 K and 10 K below PD) and a uniform reduction in precipitation (generally by 40-50% compared to PD), as inferred from individual ice-core records (Huybrechts, 2002; Golledge et al., 2012; Whitehouse et al., 2012a; Gomez et al., 2013; Quiquet et al., 2018). However, this approach provides only a crude representation of glacial climate anomalies. In reality, even if ice cores show a similar temperature decrease, estimated precipitation changes are less homogeneous. Thus imposing a constant change over the whole domain will potentially misrepresent climatologies in localized areas (Frieler et al., 2015; Fudge et al., 2016). In addition, ice cores are extracted from domes and the recorded changes are not necessarily representative of coastal regions. Because the LGM is a cold state, with presumably no (or negligible) ablation and oceanic basal melt, the reduction of precipitation with respect to the PD should have an important impact on the size of the simulated ice sheet. In addition, because the temperature and/or precipitation anomalies are uniform, the PD pattern is imprinted on the LGM atmospheric forcing fields, and changes in atmospheric patterns are thus neglected.

Another commonly used method is to prescribe the LGM temperature and precipitation fields for the whole Antarctic domain from climate simulations (Briggs et al., 2013; Maris et al., 2014; Sutter et al., 2019). Output from simulations using a hierarchy of climate models has been used in the literature, from global general circulation models (GCMs) (Sutter et al., 2019), sometimes downscaled with regional models (Maris et al., 2014), to Earth System Models of Intermediate Complexity (EMICs) (Blasco et al., 2019). Briggs et al. (2013) went a step forward to investigate the effect of uncertainty in the climate forcing fields by assessing the effect of the inter-model variance through an empirical orthogonal function (EOF) analysis. However, some model outputs do not simulate the temperature anomalies correctly at specific sites where proxies are available, such as Vostok or Dome C. This may lead to an unrealistic configuration and thus it is necessary to evaluate the accuracy of model outputs (Cauquoin et al., 2015).

In this work we aim to assess the effects of the uncertainty in basal friction and climatic (in particular atmospheric) boundary conditions on the simulated LGM AIS. We focus on basal-drag choices which can lead to realistic LGM states. For these we then investigate the effect of different temperature and precipitation fields. To this end, we use a thermomechanical ice-sheet-shelf model forced with LGM background conditions. The atmospheric temperature and precipitation fields are obtained from the eleven GCMs participating in the Paleoclimate Modelling Intercomparison Project Phase III (PMIP3) as part of the Coupled Model Intercomparison Project Phase 5 (CMIP5, Taylor et al. (2012)). The article is structured as follows. First, we describe the ice-sheet-shelf model used and the experimental setup (Section 2). Then, we show the results obtained for different basal friction coefficients and atmospheric conditions (Section 3). Finally, the results are discussed (Section 4) and summarized in the conclusions (Section 5).

2 Methods and experimental setup

For this study we use the three-dimensional, hybrid, thermomechanical ice-sheet-shelf model Yelmo (Robinson et al., 2020). The model covers the whole Antarctic domain with 191x191 grid cells of 32km x 32km resolution and 21 layers in sigma-coordinates. The flow of the grounded ice is computed as the sum of the solutions of the Shallow Ice Approximation (SIA, Hutter (1983)) and the Shallow Shelf Approximation (SSA, MacAyeal (1989)). Sliding occurs only within the SSA solution, where the computed basal velocity is ~~corrected~~ modulated with the corresponding basal friction. Ice shelves are solved within the SSA solution without basal drag. The initial topographic conditions (ice thickness, surface and bedrock elevation) are obtained from the RTopo-2 dataset (Schaffer et al., 2016). The internal ice temperature is calculated via the advection-diffusion equation.

Yelmo computes the total mass balance (MB) as a sum of the surface mass balance (SMB), the basal mass balance at the ice base and calving at the ice front. The SMB is obtained ~~as a~~ from the difference between the ice accumulation through precipitation and surface melting using the positive degree-day method (PDD; Reeh (1989)). Although there are more comprehensive methods that account for short-wave radiation for instance (Robinson et al., 2011), the PDD scheme is commonly used in ice models in the Antarctic domain, because ablation at these latitudes is limited (Winkelmann et al., 2011; Pollard and DeConto, 2012; Pattyn, 2017). Furthermore, in this particular study, the transient character of the AIS evolution is not simulated, as we focus on the LGM period. Thus, there is no need ~~for explicitly accounting~~ to explicitly account for the effects of changes in insolation on melting. Calving occurs when the ice-front thickness decreases below an imposed threshold (200 m in this study) and the upstream ice flux is not large enough to provide the necessary ice for maintaining the previous thickness (Peyaud et al., 2007). Present-day basal melting rates at the ice-shelf base and at the grounding line are obtained from Rignot et al. (2013) and extrapolated over all 27 basins identified by Zwally et al. (2012). Below grounded ice, the basal mass balance is determined through the heat equation as in Greve and Blatter (2009), where the geothermal heat flux field is obtained from Shapiro and Ritzwoller (2004). The glacial isostatic adjustment (GIA) is computed with the elastic lithosphere-relaxed asthenosphere (ELRA) method (Le Meur and Huybrechts, 1996), where the relaxation time of the asthenosphere is set to 3000 years.

Yelmo does not explicitly model the impact of ice anisotropy on the ice flow, so ~~the classical~~ an "enhancement factor" ~~are~~ is used as a tuning parameter (Ma et al., 2010; Pollard and DeConto, 2012; Maris et al., 2014; Albrecht et al., 2019). For this study we found realistic PD states for E_{grounded}=1.0 and for ice shelves E_{floating}=0.7.

2.1 Basal-drag law

As mentioned above basal sliding is calculated within the SSA solution, which is a function of the basal stress. Yelmo computes the basal stress at the ice base ($\boldsymbol{\tau}_\mathbf{b}$) through a linear viscous friction law. It depends on the basal ice velocity ($\mathbf{u_b}$), the effective ice pressure (N_{eff}) and a tunable friction coefficient (c_b):

$$\boldsymbol{\tau}_\mathbf{b} = \beta \mathbf{u_b}, \tag{1}$$

and

$$\beta = c_b N_{\text{eff}} \tag{2}$$

is the basal-drag coefficient, in $[\mathrm{kPa\ yr\ m^{-1}}]$. c_b, given in $[\mathrm{yr\ m^{-1}}]$, is a coefficient that reflects the bedrock characteristics, and N_{eff} is the effective ice pressure, given in $[\mathrm{kPa}]$. Here we have parameterized c_b as a function of the bedrock elevation, z_b (positive above sea level), analogous to previous work (e.g., Martin et al. (2011)):

$$c_b = \begin{cases} c_{\text{max}} & \text{if } z_b \geq 0 \\ \max\left[c_{\text{max}} \exp\left(-\frac{z_b}{z_0}\right), c_{\text{min}}\right] & \text{if } z_b < 0 \end{cases} \tag{3}$$

Here, z_0 is an internal parameter that determines the bedrock e-folding depth over which the friction coefficient c_b decreases from a maximum value of c_{max} reached for bedrock elevations above sea level ($z_b \geq 0$) and a minimum threshold value c_{min}. For ~~higher values of z_0 (i.e., lower absolute~~ lower values of z_0~~)~~, c_b falls more rapidly with depth. This parameterisation captures the phenomenon by which the occurrence of sliding (and its intensity) is favoured at low bedrock elevations and specifically within the marine sectors of ice sheets. It follows a similar approach as in Albrecht et al. (2019) and Martin et al. (2011), where the bedrock friction (in their case the “till friction angle”) depends on the bedrock elevation.

The effective pressure is represented by the Leguy et al. (2014) formulation, under the assumption that the subglacial drainage system is hydrologically well connected to the ocean so that there is full support from the ocean wherever the ice-sheet base is below sea level. We thus assume that the exerted basal pressure at the land-ice interface depends on the difference between the overburden pressure and the basal water pressure (i.e. the distance from flotation as measured in ice thickness), hence:

$$N_{\text{eff}} = \rho_i g \left(H - H_f\right) \tag{4}$$

where ρ_i is the density of ice, g is gravity, H is the ice thickness and H_f is the flotation thickness, given by $H_f = \max\left[0, -\frac{\rho_w}{\rho_i} z_b\right]$, where ρ_w is the seawater density, respectively, and z_b is the bedrock elevation (positive above sea-level). In this way, far from the grounding line, $H_f = 0$ and $N_{\text{eff}} = \rho_i g H$, while at the grounding line, where $H = H_f$, $N_{\text{eff}} = 0$. This ensures continuity of $\boldsymbol{\tau}_{\mathbf{b}}$ at the grounding line.

2.2 Climate forcing

To simulate the AIS at the LGM, Yelmo is run over 80 kyr with constant LGM conditions from PD observations. Sea level was set at -120 m during the LGM. The atmospheric forcing field is given by the following equation:

$$T^{atm}_{\text{LGM}} = T^{atm}_0 + \Delta T^{atm}_{\text{LGM-PD}} \tag{5}$$

where T^{atm}_0 is the PD temperature field at sea level obtained from RACMO2.3 forced by the ERA-Interim reanalysis data (Van Wessem et al., 2014) and $\Delta T^{atm}_{\text{LGM-PD}}$ is the LGM surface temperature anomaly relative to the PD. The monthly-mean temperature fields are obtained from each of the the eleven PMIP3 models, as well as by the ensemble mean (Fig. 1a). We apply

a lapse rate correction that accounts for changes in elevation (0.008 $\mathrm{K\ m^{-1}}$ for annual temperatures and 0.0065 $\mathrm{K\ m^{-1}}$ for summer temperatures) in concordance with other ice-sheet models (Ritz et al., 1997; DeConto and Pollard, 2016; Quiquet et al., 2018; Albrech
.

The LGM precipitation is calculated as

$$P_{\text{LGM}} = P_0 \delta P_{\text{LGM/PD}} \tag{6}$$

where P_0 is the PD monthly-mean precipitation obtained in the same way as the PD temperature and $\delta P_{\text{LGM/PD}}$ is the relative anomaly between the LGM and PD obtained from the PMIP3 ensemble. Figure 1b shows the resulting precipitation field, P_{LGM}, for the PMIP3 ensemble mean. Precipitation is corrected with local temperature anomalies through Clausius-Clapeyron scaling ~~,~~ which assumes more accumulation for warmer temperatures and therefore lower elevations (5 $\%\mathrm{K^{-1}}$; Frieler et al. 2015). Note that precipitation is given in water equivalent and transformed into accumulation via changes in density (i.e. 1 $\mathrm{m\ yr^{-1}}$ water equivalent ca. 1.09 m ice). Basal-melting rates for floating ice shelves are set to zero in the LGM state for simplicity.

2.3 Experimental set-up of the sensitivity studies

Basal friction

To investigate the impact of changes in basal friction on the LGM AIS we assess the sensitivity to the friction in marine zones via the minimum friction allowed (c_{min}) and the elevation parameter (z_0) in Eq. 3 that controls how quickly friction decreases with depth. For this purpose we force Yelmo with a single reference climatic state obtained from the average anomaly of the PMIP3 ensemble for the LGM climate (Fig. 1) and a range of friction parameters. This range was determined in two steps. First, PD AIS simulations were carried out. Values of c_{max}= 200·10^{-5} $\mathrm{yr\ m^{-1}}$ were found to simulate the PD AIS in good agreement with observations in terms of grounded ice volume and grounding-line advance for the selected range of values of c_{min} = 1·10^{-5}, 3·10^{-5} and 5·10^{-5} $\mathrm{yr\ m^{-1}}$ and of z_0 = -100, -125, -150, -175 and -200 m (Fig. 2; see Supplementary Information, Fig. S1, S2 for 2D-snapshots). The parameter range for the LGM AIS simulations was then selected under the criterion that the simulated volume of ice above flotation in the corresponding PD AIS simulation is within ±~~3.5~~ 1 msle of that calculated from PD observations as in Schaffer et al. (2016) (~~see Supplementary Information, Figure S1~~grey band in Fig.2).

Climatic fields

To understand the impact of changes in climatic forcing on the ice sheet, we fix the friction parameter values to a single, reference set of values ~~(~~which simulate the best PD state (Fig. 3, z_0 = ~~-175~~ -150 m and c_{min} = ~~1~~5·10^{-5} $\mathrm{yr\ m^{-1}}$) and analyze the AIS simulated at the LGM for the climatic forcing derived from each of the 11 models in the PMIP3 ensemble, using the aforementioned forcings for temperature (Eq. 5) and precipitation (Eq. 6). We focus on how the temperature and precipitation fields control the size and extent of the ice sheet. In all experiments the sea-level change estimates are computed with respect to the simulated PD state for the reference friction parameter values.

3 Results

3.1 Impact of basal friction

Here we present ~~our LGM simulated AIS~~ the simulated AIS equilibrium configuration under LGM conditions for different basal friction parameters. Ice volume change is converted into a sea-level contribution by subtracting the floating portion and taking isostatic depression of the bedrock into account (Goelzer et al., 2019). Figure 4a shows how the simulated ice volume (in ~~msle~~msle) varies with the mean basal-drag coefficient (β) of the marine zones for c_{min} = $1{\cdot}10^{-5}$ yr m^{-1} (circles), $3{\cdot}10^{-5}$ yr m^{-1} (crosses) and $5{\cdot}10^{-5}$ yr m^{-1} (diamonds) (SM, Fig. S3 for individual snapshots and Fig. S4 for time evolution). A higher mean marine friction (associated with lower z_0 values) is found to result in a larger ice volume. Sea-level differences between a case with rapidly decreasing marine friction (e.g. z_0=-100 m; in red) and a case with more gradually decreasing friction (e.g. z_0=-200 m, in blue) are about 7 ~~msle~~msle. This can be explained by the fact that ~~a higher basal friction slows the basal velocity~~ basal friction reduces basal sliding and hence the ice flow, translating into thicker ice. Faster sliding in the deepest areas (lowest c_{min} values) also ~~contributes to reduce the~~ reduces ice volume, ~~but only by about 2 msle~~ by about 5 msle for the range of parameters explored. We do not identify a ~~clear dependency of the simulated grounded area on the~~ strong impact of marine basal friction ~~exerted~~on equilibrium grounded ice area, as the ~~grounding-line position~~ final grounding line configuration is similar in all ~~cases~~ ensemble members (Fig. 4b). However, as discussed later, this can be due to the long integration time (SM, Fig. S4). Our results fit well within the range of previous studies both in terms of simulated msle (Simms et al. (2019) and references therein) and reconstructions of ice extension from ICE-6G (Argus et al., 2014a; Peltier et al., 2015, 2018), The RAISED Consortium (2014) and the ANU reconstruction (Lambeck and Johnston, 1998; Lambeck and Chappell, 2001; Lambeck et al., 2002, 2003). Note that in order to avoid biases due to Yelmo's coarse spatial resolution, these extensions were computed using the ice-sheet margins of each of the reconstructions at Yelmo's spatial resolution ~~. The three lowest bound simulations correspond to cases for which the corresponding PD AIS ice volume deviates from PD observations by more than 3.5 msle (see SI~~(SM, Fig. ~~S1, S2).~~ S5). For the simulations that matched PD AIS volumes within ±1 msle to observations, LGM ice volumes differences between 12.3 to 15.1 msle and ice extension about 16 million km^2 were computed.

~~The simulated surface velocity pattern shows a distribution with low values near the summit and increasing values towards the margins (Fig. 5). Our friction parameterisation reproduces the fact that ice streams become faster on topographic lows with the Amery, Wilkes and Victorias Land showing active ice streams of more than 50 (Fig. 5a,b). The WAIS, due to its marine character, is also a very active sector. Ice volume differences between~~ Looking at the simulated ice thickness between the LGM and the PD state we find a similar pattern for a slowly decreasing basal friction (z_0=-200 m; Fig. 5b) and a more rapidly decreasing friction (z_0=-150 m~~)~~ ; Fig. 5a). The main source of the LGM volume difference comes primary from the WAIS, especially from the Ross and Ronne shelf, as they advanced up to the continental-shelf break. Also a slight ice thickness decrease is found in the center of the EAIS. Performing an anomaly study between these two states allows to analyze the effect of the employed basal friction parameterisation (Fig. 5c). Ice volume differences primarily originate in the WAIS and the coastal marine regions of the EAIS and its surroundings~~(Fig. 5c), and are the result of higher basal velocities with lower friction values~~ . This occurs as a consequence of ice streams which become faster on topographic lows, such as the Amery, Wilkes and Victorias Land (Fig. 5d) leading to thinner ice. These zones of fast flowing areas are similar to the predicted occurrence of basal sliding from (Golledge et al., 2012).

Subtle differences are found when comparing the extension of grounded ice in our simulated AIS with previous reconstructions~~(Fig. ??); given the lack of sensitivity of the AIS extension to the friction coefficients explored here, the results are shown only for one set of parameters, c_{min}= $1{\cdot}10^{-5}$ and z_0=-175 , an intermediate case between the high and low friction values previously discussed (hereafter our reference run)~~. Our simulated grounded area ~~(thick black line)~~ covers almost 16 million km^2 of the 17 million km^2 of the continental-shelf break (i.e. defined by the contour z_b=-2000 m~~; grey shaded area~~). Our simulated extension stands between the ICE-6G model ~~(green line in Fig. ??)~~ and the RAISED Consortium ~~(red line)~~ and the ANU ~~(blue line)~~ model. The largest discrepancies between models occur on the Ross shelf (SM, Fig. S5). Whereas ANU and RAISED estimate an advance close to the continental-shelf break, ICE-6G is more retreated, while our results support a nearly complete advance except for z_0=-200 m and c_{min} =$5{\cdot}10^{-5}$ yr m^{-1}.

3.2 Impact of climatic forcing

Here we present the simulated LGM AIS of each individual PMIP3 model for the reference friction parameters (Fig. 6) (SM, Fig. S6 for time evolution and Fig. S7 for velocity distribution). The simulated ice-volume anomaly ranges from ~~7.8 msle to 14.0~~ 9.6 msle to 15.4 msle (Fig. 7), a ~~difference of 6.2 msle.~~ spread of 5.8 msle. We excluded in this range the model CNRM-CM5, which we will discuss later. The total ice extension ranges from ~~14.6~~ 15.9 million km^2 to ~~15.8~~ 14.6 million km^2 , a difference of ~~1.2~~ 1.3 million km^2. Thus, while the spread in ice volume is somewhat smaller than found when investigating the sensitivity to friction, the spread in extension is significantly larger.

Because the underlying dynamics in Yelmo are the same in all cases, the differences in size and extension can only be explained by differences in the climatic fields. To determine the causes underlying these differences, we investigate the sensitivity of the ice thickness and extension to the climatic fields used to force the ice-sheet model (Fig. 8). We find that higher accumulation results in a thicker ice sheet (Fig. 8a), but has no ~~appreciable~~ strong effect on the ice extension (Fig. 8b). For model climatologies for which the LGM ice sheet extends close to the continental-shelf break (an extension of around 15.5 million km^2, see Fig 8d), the AIS ice volume increases with increasing accumulation (Fig. 8c). However, there are four climate models (CNRM-CM5, GISS-E2-R-150, GISS-E2-R-151, FGOALS-g2) that despite having higher accumulation on average than the ensemble mean, do not allow the ice sheet to advance as much as the other models, leading in all cases to extensions below 15 million km^2 (Fig. 8b). Therefore, the simulated AIS volume is smaller for these less advanced ice sheets, despite the relatively high accumulation rates imposed. For all the others, for which extension is around 15.5 million km^2, the AIS ice volume clearly increases with increasing accumulation (Fig~~7c~~. 8c).

Further inspection allows us to identify the ~~atmospheric~~ surface temperature close to the grounding line (Fig 8d) as a critical factor in determining how far the AIS advances. ~~Whereas low temperatures present similar ice extension,as it becomes warmer the ice sheet is more retreated~~The grounding-line temperature is defined as the mean temperature of the ice column for all the grounding-line grid points. Whereas low surface temperatures lead to similar ice extend,relatively warm surface temperature forcing results in smaller equilibrium grounding line advance. Given the ~~low temperature values~~overall low surface temperature at LGM, ablation can ~~be generally~~ generally be discarded as the source of this behaviour (SI Fig. ~~S3~~S8; there is, however, one exception, as discussed below and a small area of ablation rates in the Antarctic Peninsula for GISS models), so we turn our attention to ice viscosity. A necessary condition for marine-based ice sheets to advance is that the ice thickness at the grounding line overcomes the flotation criterion as sustained through accumulation and/or by inland ice flow. This condition is fulfilled when the ocean depth (z_b) is shallower than $\sim$90% of the ice thickness. Warmer ice temperatures lower the ice viscosity (Fig. 8e) and prevent the grounding-line ~~to thicken~~from thickening, as a consequence of enhanced ice flow, and advance towards more depressed bedrock zones. Therefore, simulations with lower ice viscosity such as GISS-E2-R-150, GISS-E2-R-151 and FGOALS-g2 do not fully advance in the Ross shelf, Pine Island or the Amery Through (Fig. 6,7).

~~Finally, CNRM-CM5 is a particular case which does not fulfil any of our proposed hypotheses. Viscosity describes the fluidity of a material, therefore warmer temperatures enhance ice flow. Thus, following the same reasoning as before, one would expect a low viscosity as a consequence of a warmer ice column for~~ The CNRM-CM5 ~~, which is not the case (Fig. 8e)~~model simulates the smallest AIS LGM for all the PMIP3 models. This model expands ~~fully~~ partly at the Ross shelf and Antarctic Peninsula zone, but collapses completely in the Ronne and Amery shelf, leading to ice free zones in the EAIS and a lower ice volume than the ~~Ronne shelf is far from the grounding-line and the Amery shelfis even more retreated than~~ PD (Fig. 6). ~~The ice sheet does not advance in these regions~~ This occurs due to the presence of ~~abnormal ablation , which impedes the ice expansion~~ ablation in these regions (see SI, Fig. ~~S3). We argue that the unexpected large viscosity is a consequence of two competing effects. The fully advanced regions, as the Ross basin, contribute to a rather low ice temperature and hence a high viscosity. On the other hand, the ablation zones such as the Ronne and Amery basin, have warmer ice temperatures which conclude into low viscosity. Therefore Fig. 8e shows that CNRM-CM5 has on average a warm ice column and a high viscosity. A similar reasoning can be applied to Figure 8a where the mean ice thickness is low despite its high accumulation~~S8). Such a configuration is highly unlikely compared with sea-level and ice extension reconstructions from the LGM. We will discuss later possible explanations for this behaviour.

In summary, we find that the choice of the boundary climate conditions is crucial for the simulated LGM ice sheet. On one hand, the atmospheric temperatures near the coastal regions control the ice extension through viscosity. If the viscosity is ~~too~~ low, then the ice flows too fast, preventing the necessary thickening for advancing towards the continental-shelf break. Particularly, if the bedrock is too deep, the ice sheet's expansion will be hampered. Secondly, if the ice sheet extends close to the continental-shelf break, then the accumulation pattern will determine the total amount of ice volume. We find that for ~~similarly~~ fully extended ice sheets (IPSL-CM5A-LR and MRI-CGCM3), the sea-level difference due to accumulation differences is about ~~3.5~~ 4.2 msle.

Spatially homogeneous approach

Applying a simple scheme that lowers the ice accumulation and surface temperature homogeneously over the whole domain is a common ~~and valid~~ approach at first order, because during the LGM, at continental scale, a colder and drier climate is expected (Huybrechts, 2002; Golledge et al., 2012; Whitehouse et al., 2012a; Gomez et al., 2013; Quiquet et al., 2018). We thus tested a spatially homogeneous scaling (hereafter, the homogeneous method) for comparison. All simulations ~~simulated~~

~~realistic SLE~~ produce SLE ice volume in the range of previous studies and ice extensions ~~during the LGM for the same friction coefficients . In overall~~similar to reconstructions (e.g. The RAISED Consortium 2014) if using the same coefficients for basal friction and different climate forcings. Overall, consistently lower ice volumes ~~as well as reduced ice extensions~~ are simulated with the homogeneous method~~(Fig. S4). Again, because the ice dynamics are the same, this difference can only be explained by the climatic forcing. Moreover, because temperatures are not sufficiently high to produce ablation it points to ice accumulation differences~~, up to 1.5 msle (except for one case SM, Fig. S8). This is solely due to the difference in forcing, as the parameterisation of ice flow is identical. Fig. 9c illustrates the ice thickness difference between the two methods for a similar ice extension (Fig. 9a,b). ~~The anomaly shows~~ It is evident that the main source of ~~this difference in ice volume and extension comes from the WAIS~~ ice volume differences is due to changes in the WAIS configuration. The Antarctic Peninsula in particular shows a high positive thickness anomaly for the average PMIP3 climatic fields relative to the homogeneous case~~because the grounding-line does not advance there in the latter case~~. In the EAIS, the anomalies are not so pronounced~~, however~~; however, inland ice is slightly thinner, whereas closer to the coast it is thicker. This anomaly pattern can be explained by the difference between the accumulation fields (Fig. 9d). The spatially homogeneous method accumulates more ice inland and and leads to a reduced accumulation towards the continental-shelf break, especially at the Ross shelf, Pine Island and the Antarctic Peninsula. Because ice cores are generally extracted from dome regions with colder conditions, it is expected that precipitation and air temperatures near the coast are underestimated by the homogeneous approach. Nonetheless, the grounding-line is slightly more advanced in the western region of the Antarctic Peninsula. Similar as with the different PMIP3 fields, we argue that this difference is due to changes in viscosity due to atmospheric temperatures (SM, Fig. S8).

4 Discussion

4.1 ~~Basal dragging law~~Steady-state simulations

In this study we assumed steady-state LGM and PD conditions to investigate the effect of climatological boundary conditions and basal drag parameterisation. Of course, this represents a simplification of reality, as full LGM conditions only occurred for a couple of millennia. In a transient simulation, the results would additionally include a potential internal drift, which we tried to avoid. Although simulations were forced during 80 kyr under steady LGM conditions, equilibrated states were reached after only 30 to 40 kyr (see SM, Fig. S4, S6). Given that the LGP was a cold and sufficiently long period in the Antarctic domain, constant LGM conditions should be enough to stabilize the AIS near its real LGM state.

The simulated PD configurations show a slightly more advanced grounding line in the WAIS compared to the observations, especially at the Ronne shelf (Fig. 3, SM Fig. S1, S2). Also the ice thickness in the interior of the WAIS is systematically lower than observations. Both features can be partially explained by the basal-drag parameterisation used. Our parameterisation enhances sliding for deeper bedrock. The WAIS is in its vast majority a marine ice sheet, where bedrock depths can reach up to 2000 m in the interior regions. Thus we systematically simulate a lower WAIS, as we overestimate the ice flow at the interior. This, in addition, promotes the grounding line to advance. Nonetheless, this parameterisation allows for a precise tracing of ice streams. Except in the Larsen embayment, ice shelves generally show a slightly larger extension than observations. Because larger ice shelves allow for more ice accumulation and exert a backward force, it also helps the grounding-line to advance. Thus, the more advanced grounding line in the Ronne, Amundsen sea and Amery shelves could be additionally explained by the backward force exerted by ice shelves. Nonetheless, the overall picture of the simulated AIS fits well with observations in terms of grounding-line position as well as simulated ice volumes.

4.2 Role of basal friction

Even at present-day it is difficult to estimate bed properties like basal temperature or ice velocities, which could improve our understanding of basal friction. Therefore, estimating bed properties at the LGM, where the total ice volume and extension is not fully constrained, adds a degree of difficulty. ~~We~~ The dynamical state of the LGM remains a source of uncertainty as there are no observations from that time period of the AIS configuration. To study potentially possible AIS LGM dynamical states, we covered a range of friction values which lead to realistic LGM and PD configurations. The simulated sea-level differences were about ~~7~~ 4 msle between the ~~extreme cases~~ end members (Fig. 4). We found that the choice of different bedrock frictions has an impact on ice-stream activity in marine-based regions. For example, an AIS that extends up to the continental-shelf break, but with a relatively low volume increase, can be achieved through a very dynamically active ice sheet. In that case, ~~marine~~ marine-based regions, and more specifically the WAIS, have the potential to maintain fast ice streams at the LGM~~and still agree with PD observations~~.

The choice of ~~a given and unique~~ the friction law for the whole AIS is still somewhat arbitrary and unconstrained. We focused on a linear viscous friction law commonly used in other studies (Morlighem et al., 2013; Quiquet et al., 2018; Alvarez-Solas et al., 2019). We are aware that other types of friction laws could have been tested, such as a regularized Coulomb law (Joughin et al., 2019) or a Coulomb-plastic behaviour (Nowicki et al., 2013), typically for ice flowing over a bedrock filled with cavities. However, the ~~importance of saturated tills is specially determinant for transient simulations with a retreating grounding line~~aim of this work was to study the uncertainty associated with the basal drag parameters, rather than assessing the uncertainty for different friction laws. Given the large uncertainty we quantified for only one friction formulation, we expect that this range would increase further considering additional formulations.

4.3 Sea-level and ice extent uncertainty

For our reference friction parameters we used the individual climate simulations of the participating PMIP3 groups as surface boundary forcing. The sea-level difference between the models was about ~~6.2~~ 5.8 msle. The lowest sea-level contribution was ~~7.8 msle (~~9.6 msle (CCSM4, with exception of CNRM-CM5) and the largest ~~14.0 msle~~ 15.2 msle (IPSL-CM5A-LR). These sea-level estimates were inside the range of other studies and reconstructions. From this point of view, we were not able to discard any specific model field. ~~Nonetheless, it seems unrealistic that air temperatures were high enough to produce ablation during the LGM as seen in~~

The CNRM-CM5 model is a particular model which simulates lower sea-level contributions than PD and more retreated grounding-lines in the Ronne sector and zones of the EAIS. The model CNRM-CM5 simulates the warmest LGM temperatures not only in the SH, but it has been also shown to simulate the lowest LGM volumes for the NH (Niu et al., 2019). A potential explanation for this behaviour can be due to sea-ice formation. As shown in Marzocchi and Jansen (2017), the CNRM-CM5 model simulates the lowest austral sea-ice extent. Such a low extent would increase surface temperatures through sea-ice albedo feedback. Hence, this could point to sea-ice formation as a crucial element in driving fully LGM conditions.

The simulated ~~ice extension is determined through air temperatures~~grounding line advance is strongly influenced by air temperature. Warmer temperatures lower the ice viscosity. Due to the marine character of the AIS, a lower viscosity enhances ice flow leading to thin ice in regions where the bedrock is too deep, which prevents a complete advance towards the continental-shelf break. Forcing from the models CCSM4, FGOALS-g2, GISS-E2-R-150 and GISS-E2-R-151 for instance do not allow a full advance in the Ross shelf, resembling the ICE-6G reconstruction (Fig. ~~??). Similarly, with FGOALS-g2 the advance into the Pine Island region or the Amery shelf advance is impeded (Fig.~~ 6). On the other hand, if temperatures are sufficiently cold ~~, less than~~ (<-20ºC~~or so, then the ice fully~~) ice full advances as in the ANU reconstruction (~~Fig. ??~~SM, Fig. S4). The RAISED Consortium has a similar extension, but presents ~~two large ice shelves~~ retreated areas at the margins of the Ronne shelf, which we are not able to simulate. Again, the simulated ice extensions were inside the range of the reconstructions, and we could not exclude any case. But we found that in addition to the precipitation field, temperature fields play a crucial role as they have the potential to accelerate the ice by lowering the viscosity and determine the total grounded ice area, which in turn affects the grounded ice volume.

~~Of course there are several sources which could impact AIS volume estimates, aside from the climatology and basal friction~~In this study, no basal melting was considered during the LGM. Of course, this is a vast simplification of reality. Unfortunately, reconstructions of ocean subsurface temperatures at the LGM are not available, so that the geological evidence for basal melt is lacking. As shown in Golledge et al. (2012), oceanic forcing leads to a dynamic response of LGM ice streams in the WAIS. If basal melt would have been considered, this would have most likely reduced the total LGM ice volume and affected its extension. Thus, our results represent an upper limit which would reduce when oceanic forcing is considered.

From the point of view of modelling, there have been some attempts to infer basal-melting rates. Kusahara et al. (2015) used a coupled ice-shelf-sea-ice-ocean model with a fixed LGM AIS extension, up to the continental-shelf break. In their model results, they obtained a larger basal melt value of ice shelves than PD. These large basal-melting rates occurred because the ice shelves were located at the edge of the continental-shelf break, where ice shelves are in contact with the warm CDW. However, these basal-melting values cannot be applied to the interior of the continental shelf as these waters do not penetrate so easily there. On the other hand, Obase et al. (2017) simulated basal-melting rates on an idealized PD AIS to investigate the response of basal melt rate to a changing climate. However, these basal-melting rates are not realistic and cannot be applied directly to the AIS as the grounding-line advances during the LGM affect the climatic conditions and subshelf melting. In order to investigate the impact of realistic basal-melting rates it would be necessary to account for comprehensive parameterisations or coupled ice-sheet-ocean models (Lazeroms et al., 2018; Reese et al., 2018; Favier et al., 2019; Pelle et al., 2019), which is outside of the scope of this study. Furthermore, since our and the aim was to simulate a fully advanced AIS, as suggested by geomorphological records (The RAISED Consortium, 2014), basal-melting rates were set to zero for the sake of simplicity in this work.

Another potential source of uncertainty is the employed bedrock relaxation time. A change in bedrock depth, for instance, has profound implications on the simulated AIS, as it does not only change the local sea level, but it can also facilitate (or impede) the ice advance and retreat (Philippon et al., 2006). Here we used a simple ~~parameterization~~ parameterisation that accounts for the elasticity of the lithosphere and a non-local response caused by lateral shift (Le Meur and Huybrechts, 1996). This formulation does not capture differences in the mantle viscosity as it applies the same spatially homogeneous time response. Nonetheless, the Antarctic bedrock is a complex component with different rheological properties. The WAIS for instance is a low-viscosity region where the bedrock deformation happens on a shorter timescale (Whitehouse, 2018; Whitehouse et al., 2019). The next generation of ice-sheet models coupled to GIA models may produce more realistic bedrock responses and hence help to improve the sea-level budget at the LGM. This can be helpful for instance to constrain the phase space of friction parameters.

4.4 Forcing methods

Overall, ~~a homogeneous anomaly~~ homogeneous climate anomaly-forcing relative to present day ~~simulates~~ leads to a lower ice volume as a consequence of low accumulation near the ice-sheet margins (Fig. 9b). This indicates that the AIS could have stored more ice at the LGM than estimated by studies applying such a scheme. As opposed to a spatially homogeneous method, GCM outputs are capable of representing local atmospheric effects, such as atmospheric circulation changes or localized precipitation structures. Thus, ~~the latest ice-sheet models have begun to be forced by more detailed and arguably more realistic climatic fields~~ recent paleo ice sheet model exercises utilise climate forcing derived from GCMs (Briggs et al., 2013; Maris et al., 2014; Sutter et al., 2019). Nevertheless, we have shown here that the spread of the simulated ice volume and ice extension for different climatic outputs can be equal to or larger than that resulting from different ~~basal-dragging choices. The~~ assumptions of basal drag. The cryosphere is a component of the Earth System that also interacts with other components, such as the atmosphere or the ocean. Therefore the configuration of the AIS (as well as other ice sheets) for the PMIP3 LGM ~~climatologies are built with a prescribed ice extension and surface elevation (Abe-Ouchi et al., 2015).~~ simulations is crucial in assessing the LGM climatologies. The PMIP3 LGM simulations were forced with an AIS volume of 22.3 msle compared to PI (Abe-Ouchi et al., 2015). This ice volume largely overestimates the obtained values in this work, as well as from latest studies (Simms et al., 2019). It is clear ~~then, that by construction, ice models should be driven towards these particular configurations. NonethelessGCM models may exhibit biases in the temperatures and precipitation in localized regions . A way to potentially test the plausibility of the employed climatic fields is to compare with ice proxies. We strongly recommend that paleo~~ that a significant larger AIS will create a colder and drier environment than a smaller ice sheet. Part of this effect can be partially compensated in ice-sheet ~~simulations should be performed with GCM outputs, as they capture more complex processes than a spatially homogeneous method, but the choice of the climatic fields has to be consistent with reconstructions. In the future with~~ models with the elevation lapse rate. Nonetheless, wind currents for instance which could affect the cloud formation and accumulation at localized regions could not be taken into account. In order to compare with PMIP3 results, the first preliminary results of PMIP4 ~~results, more accurate climatic fields are expected~~are forced with the same AIS LGM configuration (Kageyama et al., 2020). Nonetheless, given the fact that the latest studies point to a lower ice volume, new PMIP experiments could consider the effect of a fully advanced, but smaller AIS. Another possibility is to employ fully coupled models to evaluate the LGM climatologies and the simulated LGM ice sheets.

4.5 Model limitations

In this study we employed a coarse resolution of 32 km. The simulation of large continental marine ice sheets has been found to be very sensitive to spatial resolution, especially at the grounding line (Pattyn et al., 2013). Grounding-line migration is a subgrid-scale process at such coarse resolutions. Ice-sheet models often use subgridding parameterisations to mimic higher resolutions at the grounding line. Nonetheless, even these parameterisations are often unable to trace the grounding-line migration correctly (Seroussi et al., 2014; Gladstone et al., 2017). Yelmo computes the fraction of grounded ice at the grounding line via subgrid and scales the basal friction at the grounding line with the grounded ice fraction (Robinson et al., 2020). To analyze the potential implications of a higher spatial resolution, we additionally performed two LGM experiments (namely AVERAGE and COSMOS-ASO) together with the simulated PD state at 16km. We find that the simulated LGM state for a fully advanced AIS simulates a similar volume (a difference of 0.2-0.3 msle) and has a slightly larger extension (0.2 to 0.3 million km^2) for both resolutions (SM, Table and Fig. S10, S11). Nonetheless, the simulated PD state is smaller for 16 km resolution than for 32 km (around 1 msle), which creates a larger LGM ice volume anomaly for 16 km. Overall, the simulated pattern and grounding-line position is similar for both resolutions (SM, Fig. S10, S11). However, it is important to mention that the equilibrated state is reached at different times for different resolution (SM, Fig. S12), pointing to the importance of resolution for assessing grounding-line migrations.

5 Conclusions

The ice dynamics and the boundary climatology are two essential building blocks for the simulation of an Antarctic LGM state. Here we studied the uncertainty in LGM ice volume associated with these two factors, by investigating the effect of the representation of basal friction and of the atmospheric forcing, respectively, in simulations. First, we tested a range of potential basal friction values of marine zones which simulated plausible LGM states. We found that for a simple linear friction law lower (larger) friction values enhance (diminish) the ice dynamics of marine zones and result in ice sheet configurations with less (more) ice volume, but still similar grounded ice extension. This led to several potential configurations of the AIS with a sea-level difference with respect to today in the range of ~~11.2 msle and 17.5 msle~~ 12.3 msle to 15.1 msle and with a total ice extension in the range of ~~15 to 16~~ 15.7 to 15.8 million km^2. Then, for a particular friction configuration within the estimates of ice volume and extension, we studied the individual sea-level contribution from simulations driven by LGM climates provided by the eleven PMIP3 participating groups. We found ice volume anomalies ranging from ~~7.8-14.0~~ 9.6 to 15.4 msle and extensions of 14.6 to ~~15.8~~ 15.9 million km^2. Our results show that the uncertainty in sea-level LGM estimates due to basal drag is similar to the uncertainty resulting from the background climatic conditions derived from PMIP3. Imposing the PMIP3 fields ~~, whose climate simulations include dynamic adjustment to the LGM boundary conditions, translate into~~ leads to higher precipitation rates along the Antarctic coast ~~, hence leading~~ and hence to a larger simulated ice volume compared to using a homogeneous anomaly method. The grounding-line advance is strongly determined by the atmospheric temperatures as well. Higher temperatures enhance ice flow reducing the ice viscosity. Because of the marine character of the ~~AIS~~WAIS, relatively high temperatures near the coast can prevent ice expansion. Thus, along with improved knowledge of basal conditions, constraining broader possible climatic changes during the LGM is imperative to be able to reduce uncertainty in the AIS volume estimates for this time period.

Code and data availability

Yelmo is maintained as a git repository hosted at https://github.com/palma-ice/yelmo under the licence GPL-3.0. Model documentation can be found at https://palma-ice.github.io/yelmo-docs/. The results used in this paper are archived on Zenodo (http://doi.org/10.5281/zenodo~~.3701258~~.4139169).

Author contributions. JB carried out the simulations, analyzed the results and wrote the paper. All other authors contributed to designing the simulations, analyzing the results and writing the paper.

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

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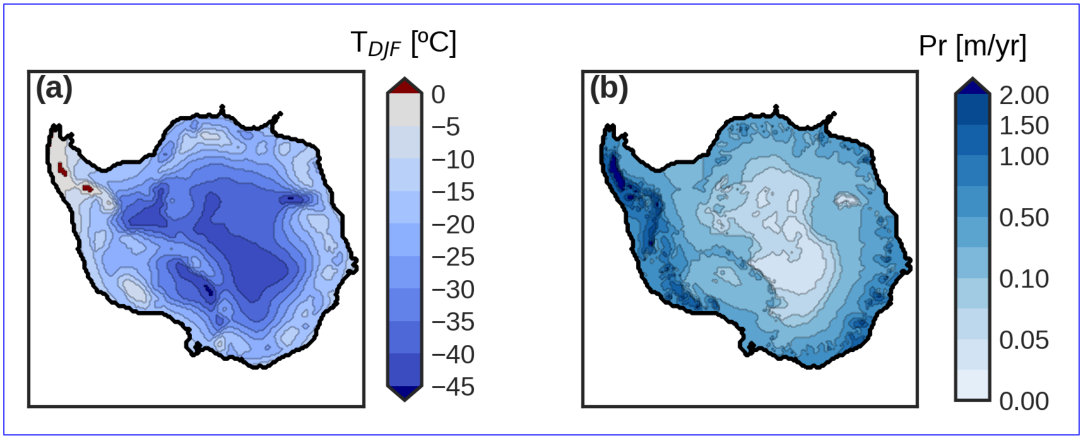

Figure 1. PMIP3 ensemble mean **(a)** surface summer temperature (in ºCelsius) and **(b)** annual precipitation (in $\mathrm{m\ yr^{-1}}$ water equivalent) at sea level. The thick black line shows the 2000 m-depth contour.

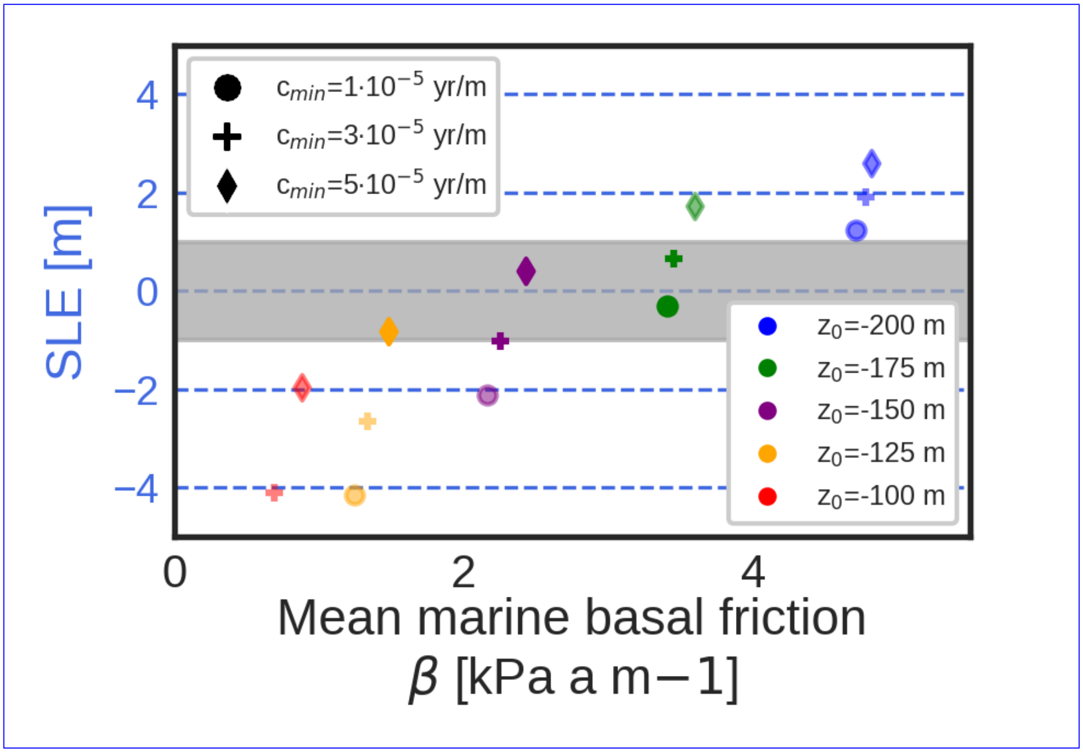

Figure 2. Present-day (PD) Antarctic Ice Sheet (AIS) ice volume above flotation and sea level equivalent (SLE) simulated for the explored values of friction parameters for c_{max} =200·10^{-5} yr m^{-1}. The grey band represents a desviations of ± 1 m from PD observations (Schaffer et al., 2016). Full colors represent simulations that fall inside the grey band.

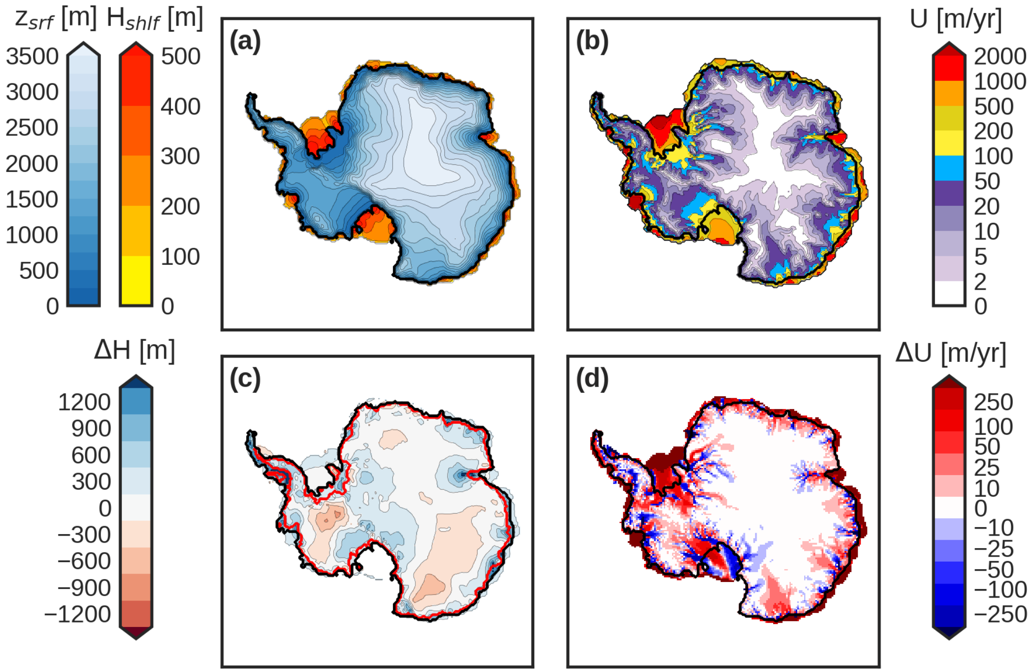

Figure 3. Simulated PD AIS **(a)** surface elevation (blue) and ice-shelf thickness (orange); **(b)** ice velocity; **(c)** ice thickness anomaly (simulated minus observations); **(d)** surface velocity anomaly, for the best match PD of all the ensemble mean. The thick black line corresponds to the simulated grounding-line position. The thick red line in **(c)** represents the actual grounding-line position.

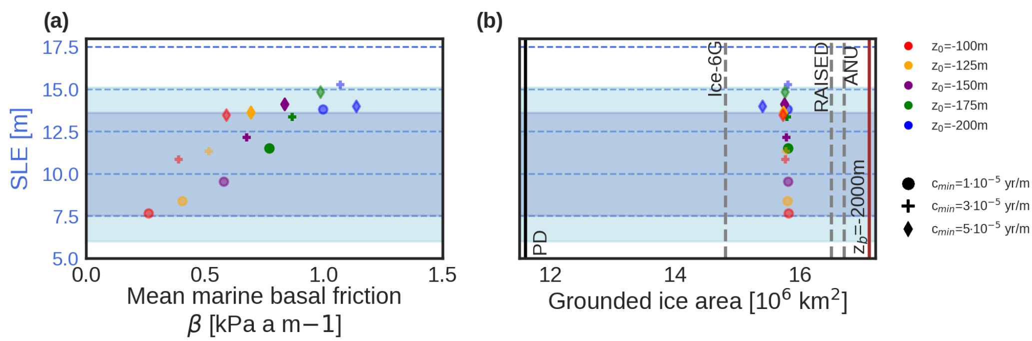

Figure 4. Scatter plot of the simulated LGM ice-volume anomaly (in msle, positive means ice-volume increase at the LGM) with respect to **(a)** the mean basal-drag coefficient and **(b)** the simulated grounded ice area, for the LGM simulations corresponding to different friction parameters. The dark blue horizontal area represents the SLE LGM estimates summarized by Simms et al. (2019) since 2010. The light blue area includes the uncertainties of the two extreme cases. The grey shaded vertical lines in **(b)** show the ice extension estimates from ICE-6G, The RAISED Consortium and the ANU reconstruction at the spatial resolution of our simulations (see main text). The black vertical line is the PD extension and the brown vertical line represents the computed ice area within the continental-shelf break defined as z_b>-2000 m. ~~Grey-colour symbols~~ Full colors represent simulations that ~~did not produce~~ simulate a ~~realistic~~ PD state ~~(see Supplementary Information)~~ ± 1 m from PD observations.

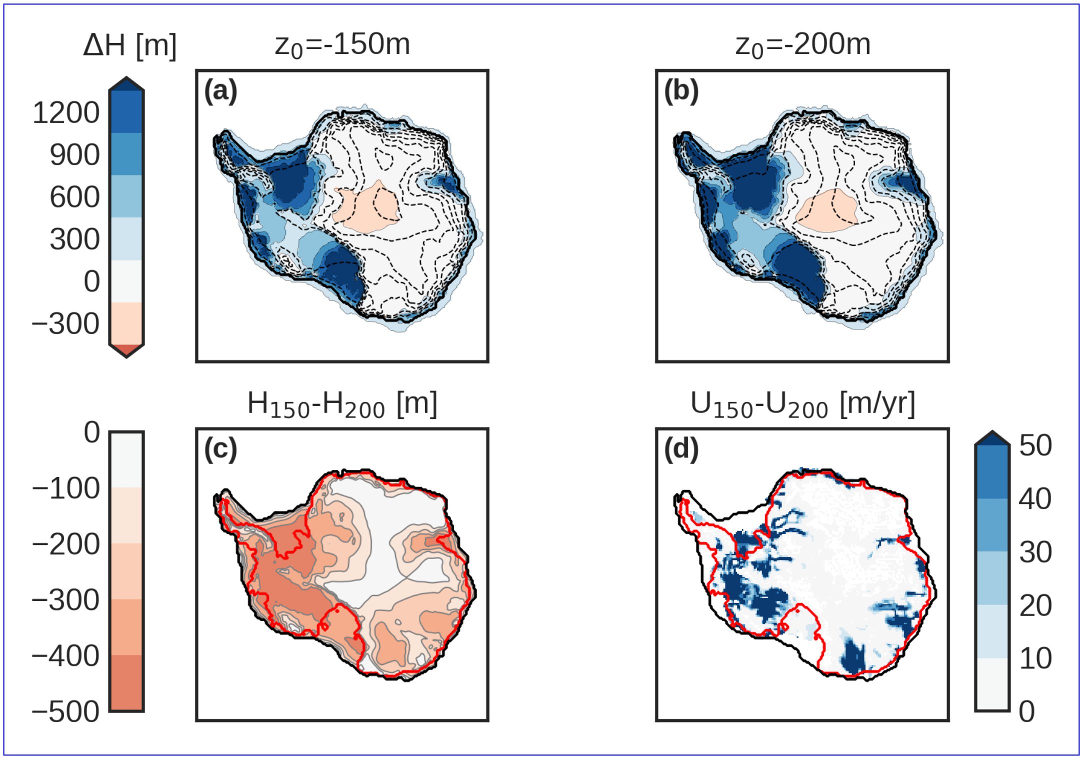

Figure 5. Simulated ice thickness anomaly between the simulated LGM ~~surface elevation~~ and ~~velocity~~ PD state (LGM minus PD) for c_{min}=$1\cdot10^{-5}$ yr m^{-1} for **(a)** z_0=-150 m and **(b)** z_0=-200 m; ~~brown~~ black discontinuous contours show surface elevation in 500 m intervals up to 3500 m above sea level. Difference in **(c)** ice thickness and **(d)** basal velocity between ~~(a) and (b)~~ the two simulated LGM states (a minus b); the thick black line shows the simulated grounding-line position of z_0=-200 m and the ~~green~~ thick red line the simulated PD grounding-line position.

~~Grounded ice extensions reconstructions from RAISED Consortium in red; ANU in blue and ICE-6G in green. In black, the simulated ice extension in this study for z_0=-175 and $c_{\min}$= $1\cdot10^{-5}$. The grey dark area shows the PD grounded ice. The area between the PD grounded area and the continental-shelf break (z_b<-2000) is shown in light grey.~~

~~LGM AIS ice elevation (brown contours) and velocity (colors) simulated using the LGM minus PD anomalies of each of the PMIP3 ensemble-members as forcing (see main text). The thick black line shows the grounding-line position. The brown contours show surface elevation in 500 intervals up to 3500 above sea level.~~

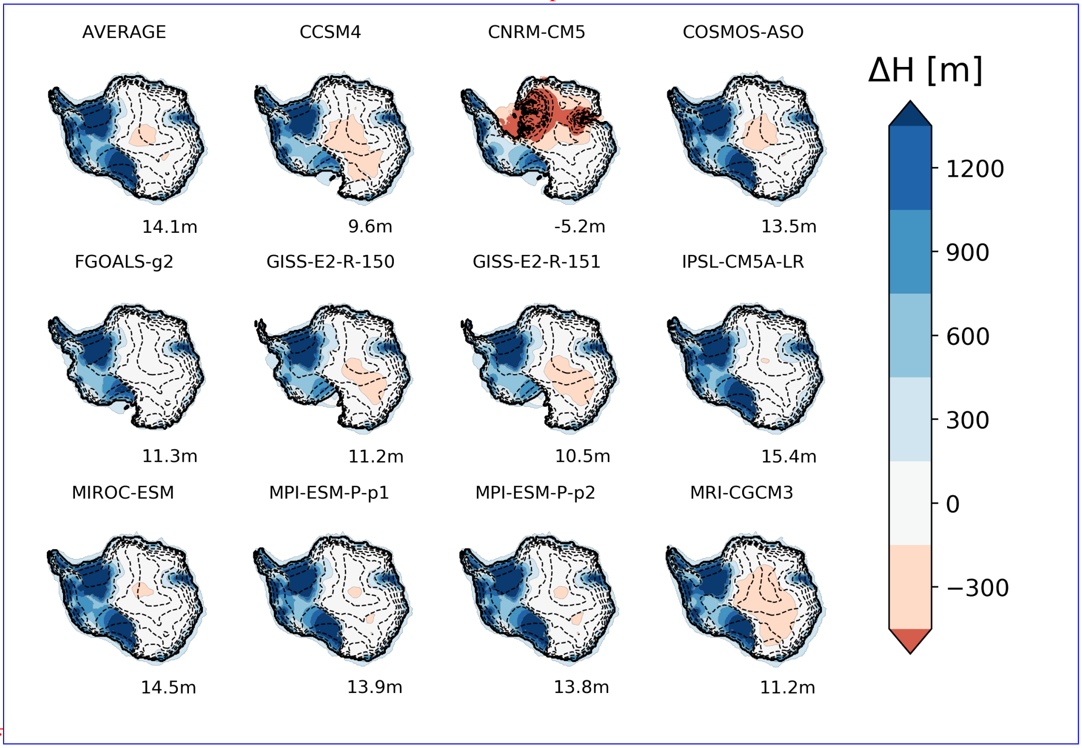

Figure 6. Ice thickness anomaly between the simulated LGM and PD for the PMIP3 ensemble. Black line represents the simulated LGM grounding-line position. Black discontinuous contours show surface elevation in 500 m intervals up to 3500 m. The number in each panel shows the ice volume difference between the simulated LGM and PD (LGM minus PD) in terms of msle.

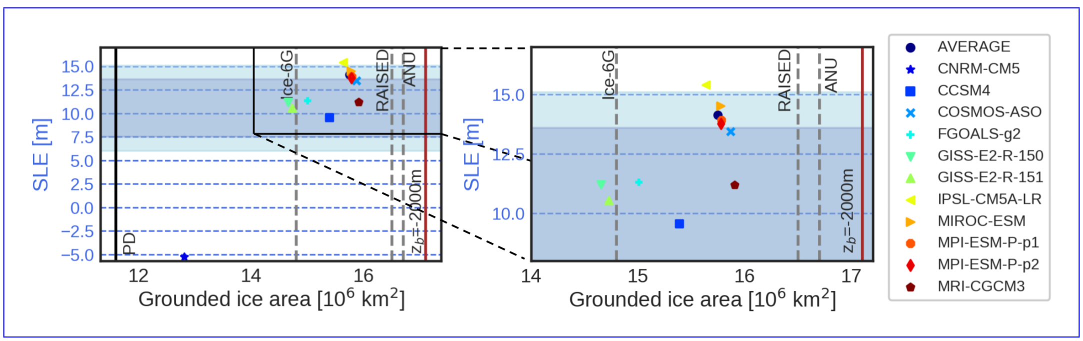

Figure 7. Scatter plot, as in Fig. ~~2~~4, of the simulated LGM ice volume anomaly (SLE) against the grounded ice area for the PMIP3 ensemble and reference values of z_0= ~~-175~~ -150 m and $c_{\min}$= ~~1~~5$\cdot 10^{-5}$ yr m^{-1}.

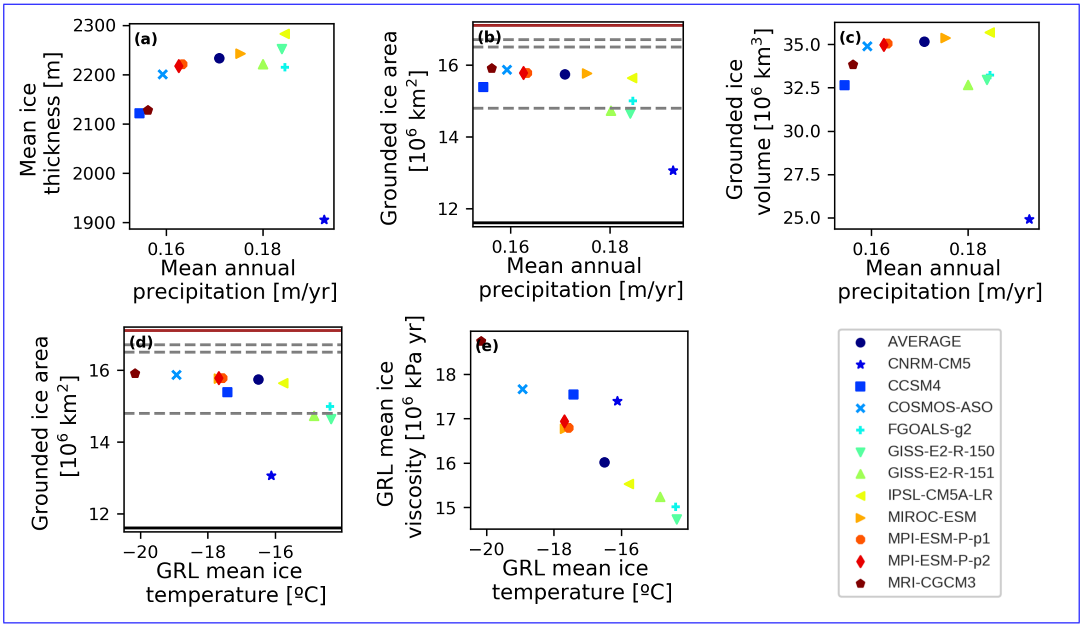

Figure 8. Scatter plots of **(a)** the mean ice thickness vs. the mean annual precipitation of the grounded grid points; **(b)** the grounded ice area vs. the mean annual precipitation of the grounded grid points; **(c)** the grounded ice volume vs. the mean annual precipitation of the grounded grid points; **(d)** the grounded ice area vs. the mean ice temperature at the grounding line; **(e)** the mean ice viscosity at the grounding line vs. the mean ice temperature at the grounding line. The horizontal lines in **(b)** and **(d)** represent the ice extensions described in Fig. 4.

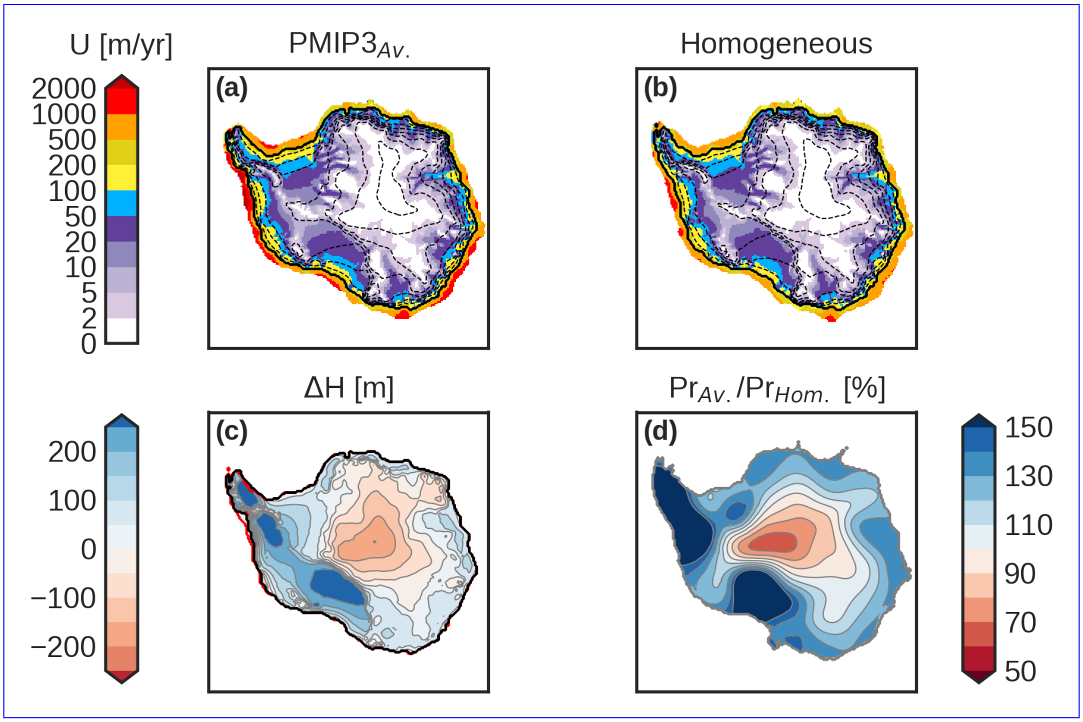

Figure 9. Simulated ~~AIS LGM surface elevation and velocity when forcing with~~ ice thickness anomaly (LGM-PD) for **(a)** the ~~spatially homogeneous method and **(b)** the~~ PMIP3 average snapshot and **(b)** the spatially homogeneous method with z_0=~~-175~~ -150 m and $c_{\min}$=~~1~~5$\cdot 10^{-5}$ yrm^{-1}. The ~~thick~~ black ~~line shows the grounding-line position. The brown~~ discontinuous contours show surface elevation ~~in~~ every 500 m intervals up to 3500 m above sea level. Panel **(c)** shows the ice thickness difference **(a)** minus **(b)**, where the thick ~~green~~ red and black lines show the grounding-line position from the simulation with homogeneous and PMIP3 climatic forcing, respectively. Panel **(d)** shows the ratio of precipitation in the PMIP3 forced simulation to that of the homogeneous simulation ~~and the grey line~~ up to the continental-shelf break (z_b=-2000 m).