

**Answer to Interactive comment on “Submarine melt as a potential trigger of the NEGIS margin retreat during MIS-3” by Ilaria Tabone et al. Anonymous Referee #3**

The study of Tabone et al. focuses on the Northeast Greenland Ice Stream (NEGIS) and its response to changes in climate, and in particular submarine melt, during the last glacial period. By applying climate forcing mimicking conditions during the last glacial period, an ice sheet/ice shelf model is used to study the transient evolution of the Greenland ice sheet over the past 120ka years. The evolution of the NEGIS are discussed in light of existing reconstructions of its history.

The study is original in assessing the long term response of the NEGIS to changes in climate, and goes beyond state of the art by comparing the dynamical evolution of the ice stream to proxy records. The paper is well written and the figures are clear. However, there are a several concerns which should be considered before publication in the cryosphere.

We are glad the reviewer valued our work and we thank them for their constructive comments and suggestions that certainly helped to make the manuscript (MS) clearer and more exhaustive. Answers to general and specific comments are reported below.

**GENERAL COMMENTS:**

The results of the study are clearly novel and of great potential in our understanding of the long term evolution of the NEGIS. However, there is a lack of detail in the description of the model results and the full potential of the study is untapped.

Given that the model simulates the entire Greenland Ice sheet these results should be included and discussed. In particular, how well does the model reproduce the present data ice sheet configuration as well as the ice stream. Similarly, how do the model results compare to published simulations and reconstructions of the LGM configuration of Greenland. This should also include an assessment of the transient evolution of the equivalent sea level contribution from Greenland.

For the NEGIS it is not clear how well the ice stream itself is reproduced by the model. To what extent does the model capture the observed geometry and velocities of the ice stream? And in particular, an assessment of the time evolution of the ice stream should be included. In what periods was the ice stream active, and did it change its position through time? If possible the model simulations should be compared with reconstructions from marine sediment archives. To make these comparisons relevant, as more data on the evolution of NEGIS become available, a time series showing the simulated ice flux at the margin of NEGIS should be included.

We agree with the reviewer that we might have been too concise in describing the overall behavior of our ice-sheet-shelf model. The MS was missing a detailed description of how the model is able to reproduce the past and present evolution of the entire Greenland Ice Sheet (GrIS). In response to this, we implemented the MS with a new series of figures that should satisfactorily answer the points raised by the reviewer. Some of these figures have been included as Supplementary Material to limit the MS length.

Specifically, we introduced:

- a comparison between simulated-observed present-day (PD) surface elevation (data from Schaffer et al., 2016) and ice velocities (data from Joughin et al., 2018) for the entire Greenland domain (new Fig. S1 in the MS);
- a map showing the GrIS last glacial maximum (LGM) extent simulated by each experiment compared to available LGM reconstructions inferred from proxy records (Evans et al., 2009; Arndt et al., 2015, 2017; Winkelmann et al., 2010) and one modelling study (Lecavalier et al., 2014) (new Fig. 7 in the MS);
- a figure showing the GrIS sea-level contribution for the last 60 ka (new Fig. S4 in the MS).

Regarding the ability in reproducing the NEGIS stream we do not think this is a relevant point for our study, since here we focus on its margin fluctuation, not much on the evolution of the stream itself. Still, this is an interesting detail to show and discuss. To do so, we included:

- a figure showing the simulated and observed PD ice velocities zoomed on the NEGIS sector (new Fig. S2);
- the spatial distribution of the simulated-observed velocities (new Fig. S3, left panel) and a scatterplot between observed and simulated PD velocities (new Fig. S3, right panel) for the NEGIS sector.

Although the magnitude of the ice flow is reasonably well simulated (very fast ice at the margin, that slows down inland), Fig. S2 clearly shows that our model fails in reproducing the detailed present geometry of the ice stream. The ice flows that feed the 79N and ZI are not well reproduced and also the long penetrating tongue of fast ice typical of the NEGIS is simulated as a wider area of fast flow, but not as fast as the observations.

The past evolution of the stream itself simulated by our model is shown in new Fig. 4 (old Fig. 3). However, to our knowledge there are no available proxy data for the past to compare with. New data from ice-core EGRIP project will certainly provide new insights into the paleo evolution of the NEGIS, but the only available information so far is given for the last 400 yr (Vallelonga et al., 2014). Thus, discussing in detail the past evolution of the stream seems to be pointless.

The lack of marine sediment archives in the northeast region hampers the possibility to compare our results with proxy data for the oceanic conditions offshore the NEGIS. This is also the reason why we chose to not include the millennial-scale variability in the oceanic forcing since, when focusing on a restricted area of the ice sheet, very precise information on the oceanic state is required to add a good degree of realism (see a detailed answer to this point below in this document). Estimates of present submarine melting rate at 79N suggest that values of  $50 \text{ m a}^{-1}$  at the grounding line and  $15 \text{ m a}^{-1}$ , or lower, along the shelf (Wilson et al., 2017). This drop in melt with increasing distance from the grounding line is taken into account by our submarine melt parameterisation. Hopefully new submarine melt estimations in the future will help to better constrain our results.

Nevertheless, we compare our results to other types of reconstruction. Our simulated maximum glacial extent is compared to previous reconstructions based on geological records (Evans et al., 2009; Arndt et al., 2015, 2017; Winkelmann et al., 2010; Funder et al., 2011) showing a fair agreement (new Fig. 7). These and other paleo data investigating the NEGIS margin throughout the last glacial period (LGP) (e.g. Bennike and Weidick (2001), Weidick (1996)) were already considered by Larsen et al. (2018) to construct their grounding-line migration transient profile (Fig 3a of Larsen et al., (2018)). Therefore, by comparing our results to those from Larsen et al. (2018) we are implicitly evaluating our simulated NEGIS front evolution with respect to all these geological records (new Fig. 3, old Fig. 1).

Finally, the transient evolution of the ice flux has been included in the results to help analysing the dynamical effects of the oceanic forcing on the retreat that old Fig. 2 (new Fig. 3 in the MS) may not exhaustively explain. Note that the ice flux has been averaged on the NEGIS sector to smooth the signal. This figure confirms the dynamical reorganization of the NEGIS sector induced by the oceanic forcing.

All the figures cited here have been discussed in the new version of the MS.

Accordingly, we changed the paragraph of Pag. 6 lines 14-19 (old MS) to:

*“Second, our simulated grounding-line advance during the LGM is smaller than the maximum extension suggested by reconstructions from geological records (Fig. 7). This bias furthermore increases with increasing oceanic forcing. However, even in the unperturbed experiment, which allows the largest ice-sheet expansion due to the absence of melting at the marine margins, the grounding-line does not reach the continental shelf break either. This discrepancy in the LGM extents is reflected in the transient GrlS sea level contribution from the LGM to the present (Fig. S4), that is underestimated as compared to other recent modelling work (Lecavalier et al., 2014, Tabone et al., 2018). Nevertheless, our estimation is not far from others (Simpson et al., 2009, Huybrechts 2002) and well within the range proposed by Buizert et al. (2018). Note that although the LGM extent simulated by Lecavalier et al (2014) is smaller than ours in the northeast, their ice volume contribution at the glacial maximum is about 1 m SLE higher. This could be partly due to their larger grounding-line advance in the northwest, but it might be also related to a more active dynamics in our simulations.*”

The volume discrepancy between our two studies performed using the same ice-sheet-shelf model are likely due to differences in the setup dynamics. The main reason seems to be related to the fact that SIA and SSA velocities are here simply summed up instead of mixed through a weighting function as in Tabone et al., 2018. This increases the velocities in the transition zones promoting discharge of ice from the interior and consequently limiting the ice volume accretion. Second, Tabone et al. (2018) accounted for refreezing processes at the base of the ice shelves, which allowed the grounding line to advance easily, leading to a glacial state in which almost all the GrIS margins were able to reach the shelf break. It is clear that this larger extent could account for a substantial part of the ice volume discrepancy. Another possible reason could be that here we increased the basal drag at the base of grounded temperate ice (by increasing its coefficient  $c_t$  of Eq. 2). More friction at the base may foster the production of water at the ice-bed interface through heat release, making the bed more slippery and the ice flow to accelerate. However, we expect that this process is responsible for only a small fraction of the ice volume discrepancy, since it is counteracted by the increase in basal friction itself. Increasing the total ice volume during the glacial would probably require a substantial tuning effort, that is beyond the scope of this study. Our goal is not to provide a perfect match with the LGM but to illustrate a plausible mechanism behind the retreated ice margin at MIS-3 and its subsequent advance.”

Also, this paragraph has been added in the Discussion:

“The comparison of our results with observations is a good strategy to assess the model performance and to comprehensively evaluate the robustness of our results. At large spatial scales our simulations fairly represent the present state of the GrIS (Fig. S1). The maximum differences in surface elevations are found in the southwest and in the east due to a mismatch in ice cover. There, the ice sheet ends in many steep and narrow fiords which are not properly represented by the 10km-resolution model. Also, the NEGIS front is located farther inland than as observed. The velocity field shows a pretty good agreement in the interior of the ice sheet, where ice speeds are expected to be lower than  $50 \text{ m a}^{-1}$  (Joughin et al., 2018). However, the simulated ice flow of outlet glaciers and ice streams shows more discrepancies. The speed of the inland flow is generally overestimated, whilst the velocities of streams as they extend far inland is underestimated. By zooming into our domain of interest we see that this pattern is also shared by the NEGIS (Fig. S2 and left panel of Fig. S3). The stream geometry is not properly recognized, although the spatial distribution of the velocities is somewhat consistent with observations (faster flow at the margins and reduced speed in the interior, as seen in Fig. S2). However, the tributary fast flows that feed the 79N are not reproduced; the SG is faster than expected and, instead of the long penetrating tongue of ice that characterises the NEGIS, the model simulates a stream catching a wider area.

Properly modelling the NEGIS is a well-known problem of ice-sheet models that investigate the evolution of the GrIS at large spatial scales. Most of these models underestimate the stream velocity and do not properly capture its outline (Seddik et al., 2012; Greve and Herzfeld, 2013; Aschwanden et al., 2016; Calov et al., 2018; Golledge et al., 2019). Greve and Otzu (2007) succeed in reproducing a correct magnitude of its speed by increasing the basal sliding under the NEGIS by three orders of magnitude relative to the rest of the ice sheet, but they fail in reproducing its geometry. A good agreement between model and data is found in Price et al. (2011) and Peano et al. (2017), who use a spatially variable basal friction coefficient derived from an iterative inverse method to match the observed velocities.

Our imperfect reproduction of the NEGIS stream is probably related to a combination of still low spatial resolution (10 km) and problems in capturing the dynamics at the base of the ice sheet. Our basal friction coefficient  $\beta$  is a function of the effective water pressure at the base of the ice sheet, which is a significant degree of freedom in ice-sheet models. A better representation of basal hydrology and sliding could help to improve the simulation of the ice stream. In parallel, new studies on the origin of the stream (following Roghozina et al., (2016)), its basal characteristics (following e.g. Keisling et al., (2014), Christianson et al. (2014) and Rivermann et al., (2019)) and new data from the EGRIP ice core (following e.g. Vallelonga et al., 2014) will bring new insights in this direction.“

Another concern is the choice of oceanic forcing applied to the model ice sheet. For simplicity the submarine melt rate is assumed to be spatially uniform around Greenland. Given the lack of data this

can be argued to be a fair assumption. However, the impact of this choice should be documented and discussed in light of existing data from sites along the margins of Greenland. A bigger concern is the inference that past oceanic temperatures below the ice evolve in phase with the atmospheric temperature (eq. 4). Several studies have shown that during glacial periods the subsurface temperatures off Greenland were relatively warm due to the stratification of the water column under an extensive sea ice cover and associated fresh surface layer (see e.g. Alvarez-Solas et al. 2011).

The reviewer is certainly right arguing that subsurface waters of the North Atlantic and Nordic Seas were likely warmer during the stadial phases of the LGP, thus evolving in antiphase with respect to the surface atmospheric temperatures inferred from the ice-core records of Greenland. In another work that investigated millennial-scale variability of the GrIS to abrupt oceanic variations associated with Dansgaard-Oeschger (D-O) cycles (Tabone et al., 2019), we actually forced the ice-sheet model through subsurface temperature anomalies that accounted for this antiphase relation. In that case, the usage of an antiphase signal was highly recommended to correctly interpret the results, since several high-resolved proxy records from marine sediment cores and modelling studies exist corroborating the hypothesis that subsurface and surface water temperatures were decoupled during D-O events. Thus, that choice was sustained by the short (millennial) timescales involved.

On the contrary, here, we focus on the response of the NEGIS margin to variations in the oceanic temperatures at much longer (orbital) timescales. To make the experiment as simple as possible and within the scope of a sensitivity study, we impose a spatially homogeneous temperature anomaly at the base of the ice shelf that follows climatic variations due to orbital changes. This is based on the assumption that at such long timescales the ocean responds as the atmosphere to changes in insolation, although this is likely a simplification of what probably was the real paleo oceanic circulation around Greenland. Also, due to the lack of proxy data that provide information on oceanic temperature variations that cover large portions of the quaternary (at least for the last 120 kyr), an in-phase ocean-atmosphere relation at long timescales cannot be ruled out (this assumption is actually corroborated by a proxy record for the temperatures of the Arctic Ocean waters at intermediate-depth during the last 50 kyr, as discussed further in this document).

A similar approach has already been successfully used in a previous work (Tabone et al., 2018) that investigated the past role of the ocean on the GrIS evolution during the last two glacial cycles. Of course, a more realistic approach would probably consider submarine melt rates that are depth-dependent (possibly following the currently simulated depth of the grounding line) and that spatially vary across the Greenland coasts, but these are improvements that go beyond our simple sensitivity study. However, these features will be taken into account in a future work for a comprehensive investigation of the problem.

#### SPECIFIC COMMENTS:

Line 7, page 1: LGP is not a common acronym. Better to spell out last glacial period and if necessary use common acronyms such as the LGM to specify a specific period within the glacial period where appropriate.

Changed accordingly.

Line 14, page 1: NG - a more common acronym for the 79N glacier in the literature is 79N.

Changed accordingly.

Line 16, page 1: it is stated that 79N is more stable than ZI due to its bed configuration - please elaborate on this.

That sentence has been changed to: *“Since 79N is retreating over an upward-sloping bed (Mouginot et al., 2015), it may be less prone than ZI to an unstable retreat. This has been recently examined through an ice-flow model pointing out that its floating tongue has to lose several tens of km of ice before the glacier becomes unstable (Rathmann et al., 2017). A larger stability of 79N has been*

*recently tested under various future warming scenarios by another modelling study (Choi et al., 2017), suggesting that it may be related to the presence of pinning points (such as ice rises) near the calving front. Despite so, the 79N ice shelf is losing mass since 2001 (Mayer et al., 2018)."*

Line 5, page 2: the slow retreat of 79N suggested by Choi et al. is described as conservative. Why? Please elaborate.

*This sentence has been changed to : "A recent study investigating the response of 79N and ZI to oceanic forcing with the aim of constraining their future stability suggests a further slow retreat of 79N and a complete loss of the ZI ice tongue due to increasing melt rates in the next decades (Choi et al., 2017). However, new evidence of retreat of both glaciers beyond their PD margins during the Holocene suggests that the resistance of 79N to increasing basal and frontal melt modelled by Choi et al., 2017 could be too conservative (Larsen et al., 2018)."*

Line 11, page 2: the ice is thought to have retreated 20-40km being its PD position during MIS3. How is this now? Please elaborate and include an assessment of the uncertainties.

*This sentence has been changed to: "The paleo records emerging from this study, combined with a collection of geological data assembled in the last 20 years (Weidick et al., 1996; Bennike and Weidick, 2001; Evans et al, 2009; Winkelmann et al., 2010; Arndt et al., 2015, 2017), suggest that the ice margin considerably fluctuated in magnitude throughout this period. Around 41-26 ka BP during Marine Isotope Stage 3 (MIS-3, c. 60-25 ka) the NEGIS front was ca. 20-40 km farther inland than today, then advanced by more than 250 km toward the shelf break at the Last Glacial Maximum (LGM) and retreated again during the last deglaciation, at ca. 70 km behind its present-day position, where stopped most of the mid-and late Holocene (7.8-1.2 ka BP)."*

*Also, added in the Results: "This reconstruction is a result of averaging the evolution of three NEGIS outlet glaciers fronts (79N, ZI and SG) inferred from the various geological records with respect to their position at 2014 (Howat et al., 2014). Although it is a valuable tool providing a rough idea of the margin fluctuation during the last 45 ka, caution should be taken before performing one-to-one comparisons with model data. Specifically, while the strong retreat during the Holocene is documented for all those glaciers, records showing their margin position during Marine Isotope Stage 3 (MIS-3, ca. 60-25 ka BP) are available only for ZI and SG, which were behind their present location by ca. 20 and 40 km, respectively. However, since they all shared the same behavior during the Holocene, it is likely that 79N front was as inland as the others during MIS-3 (Larsen et al., 2018)."*

Line 24, page 2: resolve "?"

We do not see any "?" at Pag. 2 Line 24 in the published version of the MS under discussion.

Line 8, page 3: the climate forcing is composed of 3 different ice core reconstructions (Vinther, NGRIP, and NEEM). Substantiate why this is done, instead of using only one ice core record such as NEEM.

*The choice of a composite climatic index is based on the usage of valuable reconstructions specific for certain intervals of time to provide a final accurate temperature signal valid for the whole domain. Also, temperature reconstructions from NEEM ice core have been found to underestimate the temperature anomaly during some D-O events with respect to those inferred at the dome due to spatial temperature gradients across the GrIS (Guillevic et al., 2013). Although we focus on the NEGIS sector in the analysis, our ice-sheet model is applied to the whole GrIS. Thus, using temperature reconstructions derived from the NEEM ice core only would probably not be the best strategy.*

*The bulk of the data comes from the well-known work from Kindler et al. (2014) which reconstructed temperatures in Greenland between 120 and 10 kyr from high-resolution  $\delta^{15}\text{N}$  records from the NGRIP ice core.  $\delta^{15}\text{N}$  is sensitive to processes of the firn layer and is considered as a good alternative to  $\delta^{18}\text{O}$  for inferring surface temperatures. This time series is the longest high-resolution surface temperature reconstruction derived from the NGRIP core so far; calibrated using both proxy*

and diffusion models data, it contains all the D-O events of the last glacial, thus it is also appropriate for investigations at millennial timescales. Then, this reconstruction has been expanded to reach the PD and the LIG through other available records. Vinther et al., 2009 provides a temperature reconstruction for the Holocene based on ice cores from Greenland and Canada. This signal accounts for ice-sheet elevation changes occurred during the deglaciation. This is one of the most valuable surface temperature reconstructions available so far for the Holocene (if we do not consider a more recent work from Lecavalier et al., 2017). Finally, temperature reconstructions from 115 to 130 kyr BP are taken from the NEEM ice core (NEEM community members, 2013), since it is the only available ice core as yet that allows to trace Greenland surface temperatures back to the LIG. The final composite serie leads to a temperature reconstruction from which extracting a climatic index to be applied to the whole GrIS.

Line 9, page: why is the variability below orbital removed? What is the purpose of this? What is the model result given the full variability represented by the climate reconstructions? Is there a reason to believe the millennial scale variability should be neglected in forcing the ice sheet?

The millennial-scale variability is here removed to make the message of the paper more direct and straightforward. The experimental design was chosen to be as simple as possible. Focusing on a specific region, such as that of NEGIS, requires a right dose of caution when considering oceanic changes at rapid timescales to represent the behavior of that area from a more realistic point of view. The best way of doing it would be building a signal that comes from proxy records close to the NEGIS area. However, oceanic temperature reconstructions for the northeastern part of Greenland that, moreover, cover a large portion of the LGP do not exist to our knowledge. Considering orbital timescales only in a sensitivity test allows to avoid this issue, leading to a simpler experimental design that directly drives the reader to the message of the work. Nevertheless, records from the Arctic Ocean and Fram Strait suggest that the oceanic conditions considerably varied during the last 50 kyrs. However, the long-scale variation of these temperature records seems to fairly agree with the submarine melting evolution adopted in our work.

To justify this, we added this paragraph in the Discussion: *“Paleoceanographic records inferred from marine sediments in the Arctic Sea and Fram Strait that provide information on the oceanic state during MIS-3 at high temporal resolution are scarce. However, they all suggest rapid temperature fluctuations as a result of large changes in water masses at different depths. Warmer SST may last for 3-4 kyr before cooling (Muller et al., 2014). Generally, strong variations in the oceanic conditions are found between glacial-interglacial, but also between larger stadial-interstadial transitions (Poirer et al., 2012). A sediment record between the Nordic seas and the Arctic Ocean suggest that high SST and low intermediate water temperatures are typical of interstadials, while the opposite is found during stadials due to intrusion of warmer Atlantic subsurface water (Rasmussen et al., 2014). This strong oceanic temperature variability during the last 50 kyr is also documented by another record based on a stack of sediment records of the Arctic Ocean and the Fram Strait, suggesting the occurrence of several peaks of warmings during MIS-3 reaching temperatures 1-3 °C higher than those recorded for the Holocene (Cronin et al., 2012). However, a qualitative analysis of this temperature record at long (orbital) timescales indicate that its evolution agrees well with that of the melting rate signal used in this work: high melting during MIS-3, prolonged cooling during the LGM and high melting again during the Holocene. Thus, even though we remove some degree of realism by not considering the millennial-scale variability in the ocean, our experimental design could fairly represent the evolution of northern Greenland oceanic conditions at long timescales.”*

Line 11, page 3 and eq. 1: PD is referred to as interglacial. Please be more precise on definition of PD: interglacial, Holocene or present day?

PD refers to present day (sensu lato, that is, preindustrial). To make it clearer, this sentence has been changed to: *“ $T_{LGM, atm} - T_{PD, atm}$  is the glacial minus present-day (meaning preindustrial) atmospheric temperature anomaly simulated by the climate model of intermediate complexity CLIMBER-3 $\alpha$ .”*

Line 13, page 3: why use CLIMBER-3a and not PMIP for the LGM - interglacial climate? What is the impact of the choice of model?

We did not perform simulations with different climatologies since this is beyond the scope of this work. However, we agree with the reviewer that outputs from more recent, comprehensive, and higher resolution climate models might be more appropriate. This will be surely done in our future work. Nevertheless, CLIMBER-3 $\alpha$  has been already used in several modelling studies investigating the paleo evolution of past ice sheets proving reliable results (e.g. Alvarez et al. (2011, 2013), Banderas et al. (2012, 2018), Blasco et al. (2019), Tabone et al. (2018, 2019)). Also, since snapshots from CLIMBER-3 $\alpha$  are used for defining the transient paleo climatology for the sole atmosphere and since this work focuses on the sensitivity of the NEGIS margin to the ocean, we expect that the choice of the climatology has only second-order effects on the results.

Line 15, eq. 2, page 3: What is the rationale behind choosing the same approach for calculating precipitation as for temperature? Is this appropriate? What is the impact of this choice, please document. Note that P\_LGM and P\_PD are not described. How are these calculated?

$P_{LGM, ann}$  and  $P_{PD, ann}$  are the annual precipitations provided as outputs from simulations performed with the CLIMBER-3 $\alpha$  model under weak-AMOC and present-AMOC conditions, respectively (Montoya and Levermann, 2008), as  $T_{LGM, atm}$  and  $T_{PD, atm}$ . Defining the paleo precipitations through an anomaly method with respect to the present, as done with the temperatures, is an approach followed by many models to represent transient past precipitations (e.g., Marshall et al. (2000, 2002), Marshall and Koutnik (2006), Charbit et al. (2002, 2007), Zweck and Huybrechts (2005), Philippon et al. (2006), Colleoni et al. (2014), Banderas et al. (2018)). Note that, here, the ratio of the precipitation anomalies originally comes from the assumption that precipitation depends exponentially on temperature, thus a difference in temperature would result in a ratio in precipitation. Also, it allows to avoid the creation of negative values in the precipitation field, which would not be physical. This approach is especially useful when the ice-sheet model is not coupled to a climate model. Despite its crudity, it allows to better represent past precipitations than directly using paleo records coming from ice cores, since model snapshots describe the spatial precipitation pattern away from the core locations with more accuracy. This method can be then improved for example by incrementing the number of snapshots between the LGM and the PD (Charbit et al., 2002), or correcting the fields with an amplification factor (Banderas et al., 2018). However, for the purpose of investigating the ice-sheet evolution due to oceanic forcing at long timescales the approach used in this work should be sufficiently accurate.

To clarify this point, the sentence at Pag. 3 Line 16 has been changed to: *“The precipitation field is obtained following a similar approach based on the ratio of LGM and present-day precipitations, scaled by  $\alpha(t)$ , as:*

$$P_{ann}(t) = P_{clim, ann} * (\alpha(t) + (1 - \alpha(t)) P_{LGM, ann} / P_{PD, ann})$$

where  $P_{LGM, ann}$  and  $P_{PD, ann}$  are the LGM and PD annual precipitations provided by the same climate simulations of  $T_{LGM, atm}$  and  $T_{PD, atm}$ . This approach has been adopted by many ice-sheet models to represent transient past precipitations when they are not coupled to a climate model (e.g., Marshall et al. (2000, 2002), Marshall and Koutnik (2006), Charbit et al. (2002, 2007), Zweck and Huybrechts (2005), Philippon et al. (2006), Colleoni et al. (2014), Banderas et al. (2018)).”

Line 16, page 3: why use PDD and not scale SMB from MAR which is used for the temperature?

Despite the known shortcomings of the PDD parameterisation, such as its tendency to underestimate the surface melt, especially in warm climates since it does not take into account insolation changes (e.g. Robinson and Goelzer, 2014; Van den Berg et al., 2011), it is still one of the most used approaches in the literature for representing the ablation in paleo modelling (e.g. Peyaud et al., 2007, Born and Nisancioglu 2012, Stone et al., 2013, Quiquet et al., 2013, Banderas et al., 2018). This is because it is a simple method compared to others, such as ITMs (Insolation-temperature melt methods, Robinson et al., 2010) or EBSMs (energy balance and snowpack models, Bougamont et al., 2007), and its results are still rather satisfactory when compared to regional climate models (e.g. Peano et al. (2017), Vernon et al. (2013)). Scaling the SMB directly from MAR would overlook important processes that our ice-sheet model accounts for, such as refreezing of surface waters or the elevation-melt feedback, that can be considered by explicitly calculating the ablation and have

important consequences on the evolution of the ice sheet. The suggested method would probably oversimplify the experiment for our purposes.

Line 17, page 3: it is claimed that using PDD does not “jeopardise” results as focus is no oceanic forcing. Note that this invalidates any comparison of the relative importance of atmospheric and oceanic forcing. Please elaborate on this point + check manuscript for consistency with the discussion of the importance of the oceanic forcing given that its relative role cannot be assessed.

The reviewer is right arguing that the relative role between atmospheric and oceanic forcings in the NEGIS retreat during MIS-3 cannot be determined from our results due to the lack of a sensitivity test on the atmosphere. However, assessing this is beyond the scope of our work, which simply aims to show the potential of the oceanic forcing in driving the margin of the NEGIS throughout the last glacial period.

To make it clear, this paragraph has been changed to: “*Surface ablation is calculated by the simple positive degree (PDD) scheme (Reeh, 1989). Although this method does not account for past insolation changes, and since here we primarily investigate the sensitivity of the NEGIS to the oceanic forcing during glacial times, the choice of this melt scheme instead of another should bring only second-order effects to the overall results of this work.*”

Also this paragraph has been added in the Discussion: “*Further experiments accounting for changes in the atmospheric temperatures and precipitation or variations in the external forcing (i.e. insolation) should be carried out for a full understanding of the mechanisms involved in this retreat, here explained by considering the sole impact of the ocean. Particularly, a sensitivity study on climatic variations performed with a prescribed ocean could help constraining the effect of the atmosphere in this phenomenon to eventually evaluate the relative role between the forcings in driving the NEGIS margin.*”

Line 8, page 4: resolve “?”

We do not see any “?” at Pag. 4 Line 8 in the published version of the MS under discussion.

Figure 3: what is shown here. Please specify time periods for each subplot.

New Fig. 4 (old Fig. 3) now shows the time period for each subplot.

Figure 4: Show A and B in relation to ice margin (e.g. in figure similar to 3). Specify where smb and Bmelt are taken from.

Now the inset map of Fig. 6 (old Fig. 4) shows the maximum glacial extent for the experiment  $\kappa=8 \text{ m a}^{-1} \text{ K}^{-1}$ . Surface mass balance and basal melt are simulated by the model and averaged over the regions A and B.

#### REFERENCES:

Alvarez-Solas et al., 2011. Heinrich event 1: an example of dynamical ice-sheet reaction to oceanic changes." *Climate of the Past* 7.4 (2011): 1297-1306.

Alvarez-Solas et al., 2013. Iceberg discharges of the last glacial period driven by oceanic circulation changes. *Proceedings of the National Academy of Sciences* 110.41 (2013): 16350-16354.

Alvarez-Solas et al., 2018: Oceanic forcing of the Eurasian Ice Sheet on millennial time scales during the Last Glacial Period, *Clim. Past Discuss.*, <https://doi.org/10.5194/cp-2018-89>, in review, 2018.



Arndt et al., 2015. A new bathymetry of the Northeast Greenland continental shelf: Constraints on glacial and other processes. *Geochemistry, Geophysics, Geosystems*, 16, 10 (2015): 3733-3753.

Arndt et al., 2017. The last glaciation and deglaciation of the Northeast Greenland continental shelf revealed by hydro-acoustic data. *Quaternary Science Reviews*, 160 (2017): 45-56.

Aschwanden et al., 2016. Complex Greenland outlet glacier flow captured. *Nature communications* 7 (2016): 10524.

Banderas et al., 2012. Role of CO<sub>2</sub> and Southern Ocean winds in glacial abrupt climate change. *Climate of the Past* 8.3 (2012): 1011-1021.

Banderas et al., 2018. A new approach for simulating the paleo-evolution of the Northern Hemisphere ice sheets. *Geoscientific Model Development* 11.6 (2018).

Bennike and Weidick, 2001. Late Quaternary history around Nioghalvfjærdsfjorden and Jøkelbugten, North-East Greenland. *Boreas* 30.3 (2001): 205-227.

Blasco et al., 2019. The Antarctic Ice Sheet response to glacial millennial-scale variability. *Climate of the Past* 15.1 (2019): 121-133.

Bougamont et al., 2007. Impact of model physics on estimating the surface mass balance of the Greenland ice sheet. *Geophysical Research Letters* 34.17 (2007).

Born, Andreas, and Kerim Hestnes Nisancioglu. "Melting of Northern Greenland during the last interglaciation." (2012).

Bradley, Sarah L., et al. "Simulation of the Greenland Ice Sheet over two glacial-interglacial cycles: investigating a sub-ice-shelf melt parameterization and relative sea level forcing in an ice-sheet-ice-shelf model." *Climate of the Past* 14.5 (2018): 619-635.

Buizert, C., et al. "Greenland-Wide Seasonal Temperatures During the Last Deglaciation." *Geophysical Research Letters* 45.4 (2018): 1905-1914.

Calov et al., 2018. Simulation of the future sea level contribution of Greenland with a new glacial system model. *The Cryosphere* 12.10 (2018): 3097-3121.

Charbit et al., 2002. Simulations of Northern Hemisphere ice-sheet retreat: sensitivity to physical mechanisms involved during the Last Deglaciation. *Quaternary Science Reviews* 21.1-3 (2002): 243-265.

Charbit et al., 2007. Numerical reconstructions of the Northern Hemisphere ice sheets through the last glacial-interglacial cycle. *Climate of the Past* 3.1 (2007): 15-37.

Choi et al., 2017. Modeling the response of Nioghalvfjærdsfjorden and Zachariae Isstrøm Glaciers, Greenland, to ocean forcing over the next century. *Geophysical Research Letters* 44.21 (2017): 11-071.

Christianson et al. (2014). Dilatant till facilitates ice-stream flow in northeast Greenland. *Earth and Planetary Science Letters* 401 (2014): 57-69.

Colleoni et al., 2014. Modeling Northern Hemisphere ice-sheet distribution during MIS 5 and MIS 7 glacial inception. *Climate of the Past* 10.1 (2014): 269-291.

Cronin 2012. Deep Arctic Ocean warming during the last glacial cycle. estimate intermediate water temperatures over the past 50,000 years from the Mg/Ca from Arctic sediment cores.

Evans et al., 2009. Marine geophysical evidence for former expansion and flow of the Greenland Ice Sheet across the north-east Greenland continental shelf. *Journal of Quaternary Science: Published for the Quaternary Research Association*, 24, 3 (2009): 279-293.

Funder, Svend, et al. "The Greenland Ice Sheet during the past 300,000 years: A review." *Developments in Quaternary Sciences*. Vol. 15. Elsevier, 2011. 699-713.

Goelzer et al., 2016. Last Interglacial climate and sea-level evolution from a coupled ice sheet–climate model

Golledge et al., 2019. Global environmental consequences of twenty-first-century ice-sheet melt. *Nature* 566.7742 (2019): 65.

Greve and Herzfeld, 2013. Resolution of ice streams and outlet glaciers in large-scale simulations of the Greenland ice sheet. *Annals of Glaciology* 54.63 (2013): 209-220.

Greve and Otzu, 2007. The effect of the north-east ice stream on the Greenland ice sheet in changing climates. *The Cryosphere Discussions* 1.1 (2007): 41-76.

Guillevic et al., 2013. Spatial gradients of temperature, accumulation and  $\delta^{18}\text{O}$ -ice in Greenland over a series of Dansgaard–Oeschger events. *Clim. Past*, 9 (2013): 1029–1051.

Howat et al., 2014. The Greenland Ice Mapping Project (GIMP) land classification and surface elevation data sets. *The Cryosphere*, 8, 4 (2014): 1509-1518.

Huybrechts, Philippe. "Sea-level changes at the LGM from ice-dynamic reconstructions of the Greenland and Antarctic ice sheets during the glacial cycles." *Quaternary Science Reviews* 21.1-3 (2002): 203-231.

Joughin et al., 2018. Greenland Ice Mapping Project: ice flow velocity variation at sub-monthly to decadal timescales.

Keisling et al, 2014. "Basal conditions and ice dynamics inferred from radar-derived internal stratigraphy of the northeast Greenland ice stream." *Annals of Glaciology* 55.67 (2014): 127-137.

Kindler et al., 2014. Temperature reconstruction from 10 to 120 kyr b2k from the NGRIP ice core. *Climate of the Past* 10.2 (2014): 887-902.

Larsen et al., 2018. Instability of the Northeast Greenland Ice Stream over the last 45,000 years. *Nature communications* 9.1 (2018): 1872.

Lecavalier et al., 2014. A model of Greenland ice sheet deglaciation constrained by observations of relative sea level and ice extent. *Quaternary Science Reviews* 102 (2014): 54-84.

Lecavalier, Benoit S., et al. "High Arctic Holocene temperature record from the Agassiz ice cap and Greenland ice sheet evolution." *Proceedings of the National Academy of Sciences* 114.23 (2017): 5952-5957.

Marshall et al., 2000. Glaciological reconstruction of the Laurentide Ice Sheet: physical processes and modelling challenges. *Canadian Journal of Earth Sciences* 37.5 (2000): 769-793.

Marshall et al., 2002. North American ice sheet reconstructions at the Last Glacial Maximum. *Quaternary Science Reviews* 21.1-3 (2002): 175-192.

Marshall and Koutnik, 2006. Ice sheet action versus reaction: Distinguishing between Heinrich events and Dansgaard-Oeschger cycles in the North Atlantic. *Paleoceanography* 21.2 (2006).

Mayer et al., 2018. Large ice loss variability at Nioghalvfjærdsfjorden Glacier, Northeast-Greenland. *Nature communications* 9.1 (2018): 2768.

Montoya and Levermann, 2008. Surface wind-stress threshold for glacial Atlantic overturning. *Geophysical Research Letters* 35.3 (2008).

Mouginot et al., 2015. Fast retreat of Zachariæ Isstrøm, northeast Greenland. *Science* 350.6266 (2015): 1357-1361.

Müller et al., 2014. High-resolution record of late glacial and deglacial sea ice changes in Fram Strait corroborates ice–ocean interactions during abrupt climate shifts.

NEEM community members, 2013. Eemian interglacial reconstructed from a Greenland folded ice core. *Nature* 493.7433 (2013): 489.

Peano et al., 2017: "Ice flux evolution in fast flowing areas of the Greenland ice sheet over the 20th and 21st centuries." *Journal of Glaciology* 63.239 (2017): 499-513.

Peyaud, V., C. Ritz, and G. Krinner. "Modelling the Early Weichselian Eurasian Ice Sheets: role of ice shelves and influence of ice-dammed lakes." *Climate of the Past Discussions* 3.1 (2007): 221-247.

Philippon et al., 2006. Evolution of the Antarctic ice sheet throughout the last deglaciation: A study with a new coupled climate-north and south hemisphere ice sheet model. *Earth and Planetary Science Letters* 248.3-4 (2006): 750-758.

Poirier 2012. Central Arctic paleoceanography for the last 50 kyr based on ostracode faunal assemblages.

Price, Stephen F., et al. "Committed sea-level rise for the next century from Greenland ice sheet dynamics during the past decade." *Proceedings of the National Academy of Sciences* 108.22 (2011): 8978-8983.

Quiquet et al., 2013: Greenland ice sheet contribution to sea level rise during the last interglacial period: a modelling study driven and instrumentation constrained by ice core data. *Climate of the Past* 9.1 (2013): 353-366.

Rasmussen et al., 2014. , Water mass exchange between the Nordic seas and the Arctic Ocean on millennial timescales during MIS 4–MIS 2.

Rathmann et al., 2017. Highly temporally resolved response to seasonal surface melt of the Zachariae and 79N outlet glaciers in northeast Greenland. *Geophysical Research Letters* 44.19 (2017): 9805-9814.

Reeh 1989. Parameterization of melt rate and surface temperature in the Greenland ice sheet. *Polarforschung* 59.3 (1991): 113-128.

Riverman et al., 2019. Riverman, K. L., et al. "Enhanced Firn Densification in High-Accumulation Shear Margins of the NE Greenland Ice Stream." *Journal of Geophysical Research: Earth Surface* 124.2 (2019): 365-382.

Robinson et al., 2010. An efficient regional energy-moisture balance model for simulation of the Greenland Ice Sheet response to climate change. *The Cryosphere* 4.2 (2010): 129-144.

Robinson, Alexander, and Heiko Goelzer. "The importance of insolation changes for paleo ice sheet modeling." (2014).

Roghozina et al., 2016. Melting at the base of the Greenland ice sheet explained by Iceland hotspot history. *Nature Geoscience* 9.5 (2016): 366.

Schaffer et al., 2016. A global, high-resolution data set of ice sheet topography, cavity geometry, and ocean bathymetry. (2016): 543-557.

Seddik et al., 2012. Simulations of the Greenland ice sheet 100 years into the future with the full Stokes model Elmer/Ice. *Journal of Glaciology* 58.209 (2012): 427-440.

Simpson, Matthew JR, et al. "Calibrating a glaciological model of the Greenland ice sheet from the Last Glacial Maximum to present-day using field observations of relative sea level and ice extent." *Quaternary Science Reviews* 28.17-18 (2009): 1631-1657.

Stone et al., 2013 Quantification of the Greenland ice sheet contribution to Last Interglacial sea level rise. *Climate of the Past*, 9, 621-639. 1

Tabone et al., 2018. The sensitivity of the Greenland Ice Sheet to glacial-interglacial oceanic forcing, *Clim. Past*, 14 (2018): 455-472.

Tabone et al., 2019. Impact of millennial-scale oceanic variability on the Greenland ice-sheet evolution throughout the last glacial period, *Clim. Past*, 15 (2019): 593-609.

Vallelonga et al., 2014. Initial results from geophysical surveys and shallow coring of the Northeast Greenland Ice Stream (NEGIS). *The Cryosphere* 8.4 (2014): 1275-1287.

Van de Berg, Willem Jan, et al. "Significant contribution of insolation to Eemian melting of the Greenland ice sheet." *Nature Geoscience* 4.10 (2011): 679.

Vernon et al., 2013. "Surface mass balance model intercomparison for the Greenland ice sheet." *The Cryosphere* 7 (2013): 599-614.

Vinther, et al., 2009. "Holocene thinning of the Greenland ice sheet." *Nature* 461.7262 (2009): 385.

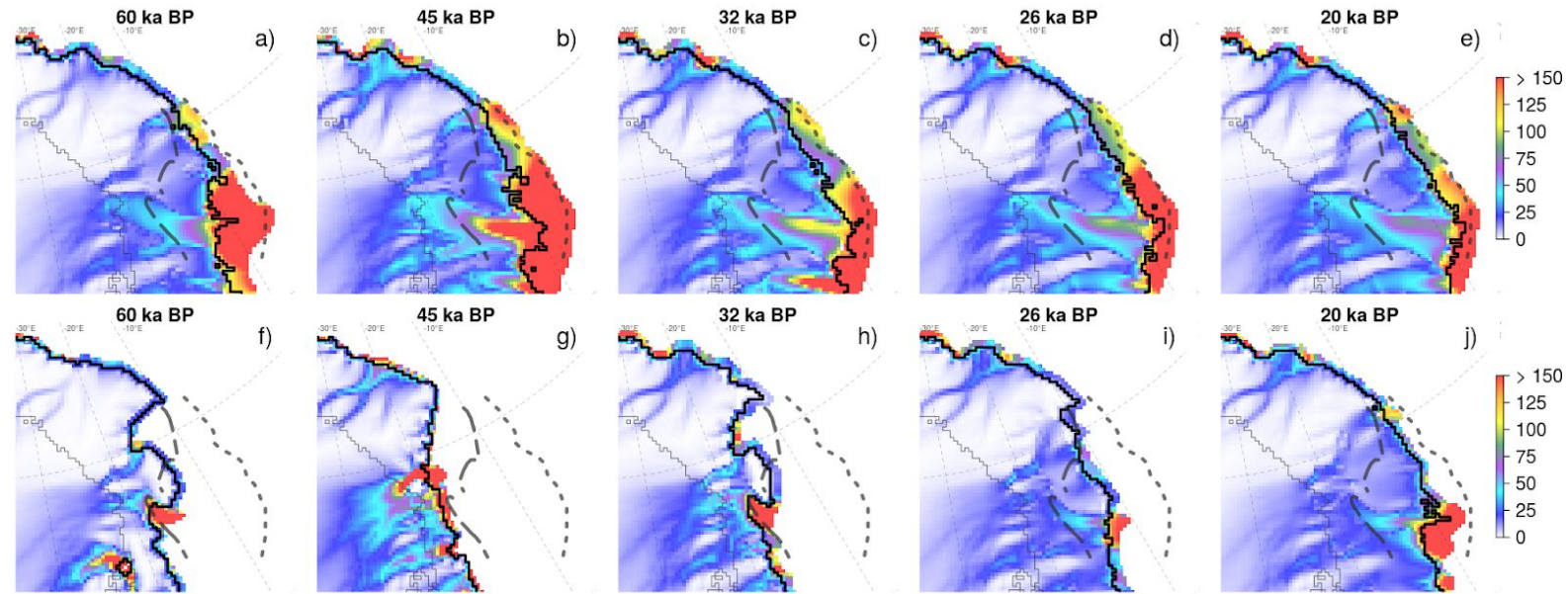
Weidick et al., 1996. Neoglacial glacier changes around Storstrommen, North-East Greenland. *Polarforschung* 64.3 (1996): 95-108.

Wilson, et al., 2017: "Satellite-derived submarine melt rates and mass balance (2011–2015) for Greenland's largest remaining ice tongues." (2017).

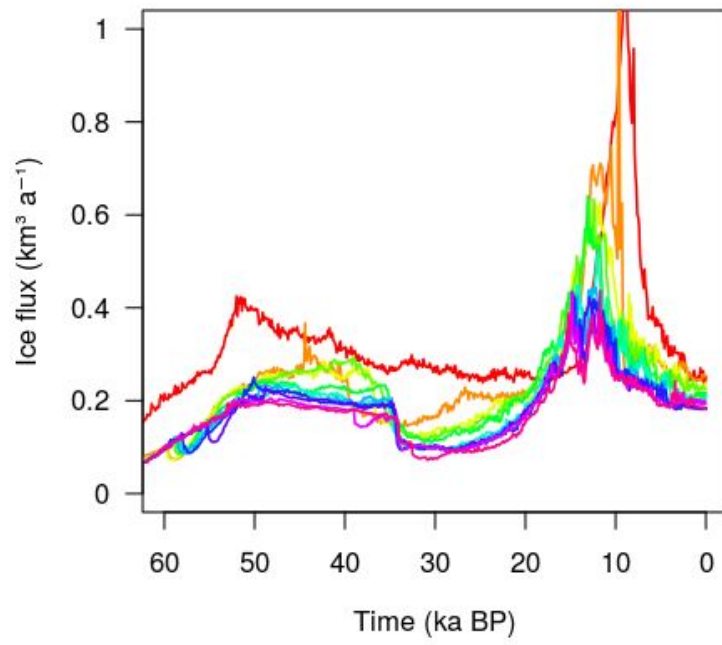
Winkelmann et al., 2010. Submarine end moraines on the continental shelf off NE Greenland—Implications for Late glacial dynamics. *Quaternary Science Reviews*, 29, 9-10 (2010): 1069-1077.

Zweck and Huybrechts, 2005. Modeling of the northern hemisphere ice sheets during the last glacial cycle and glaciological sensitivity. *Journal of Geophysical Research: Atmospheres* 110.D7 (2005).

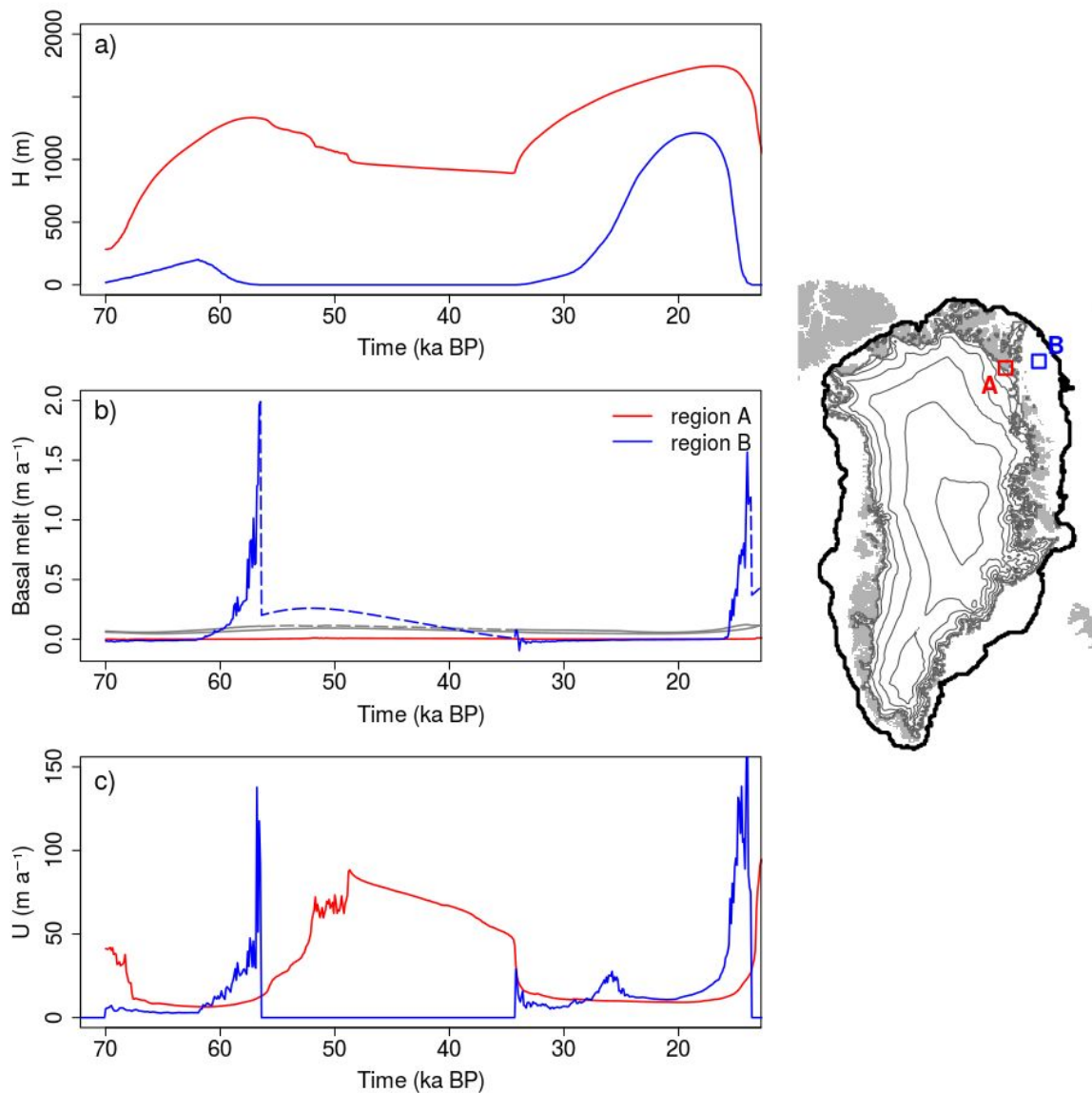
## FIGURES



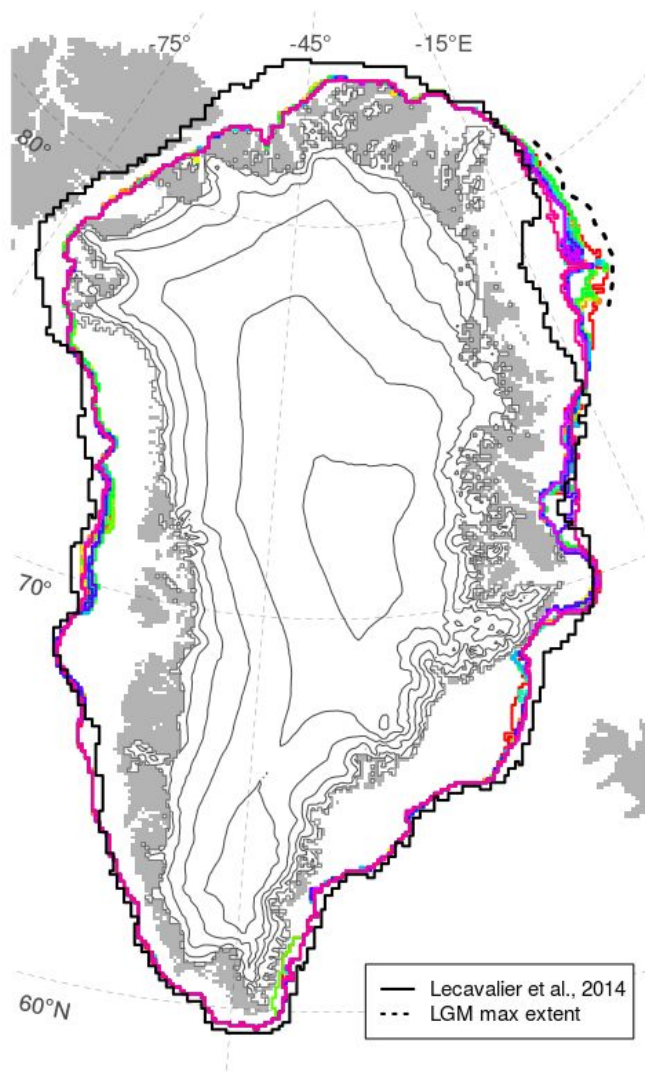
New Fig. 4 (old Fig. 3). Snapshots of  $U$  ( $\text{m a}^{-1}$ ) in total absence of submarine melting (a-e) and in presence of active orbital-driven oceanic forcing ( $\kappa = 8 \text{ m a}^{-1} \text{ K}^{-1}$ ,  $B_{\text{ref}} = 8 \text{ m a}^{-1}$ ) (f-j) at different times along MIS-3 and the LGM. The black line represents the position of the simulated grounding line. Grey thin solid line represents the observed PD grounding-line position (Schaffer et al., 2016). Maximum (dotted black line) and minimum (dashed black line) grounding-line positions reconstructed for the LGM (Funder et al., 2011) are also shown.



New Fig. 5 in the MS. Ice flux at the NEGIS sector for different oceanic forcings. Colors refer to the color scale of new Fig. 3 (old Fig. 1).

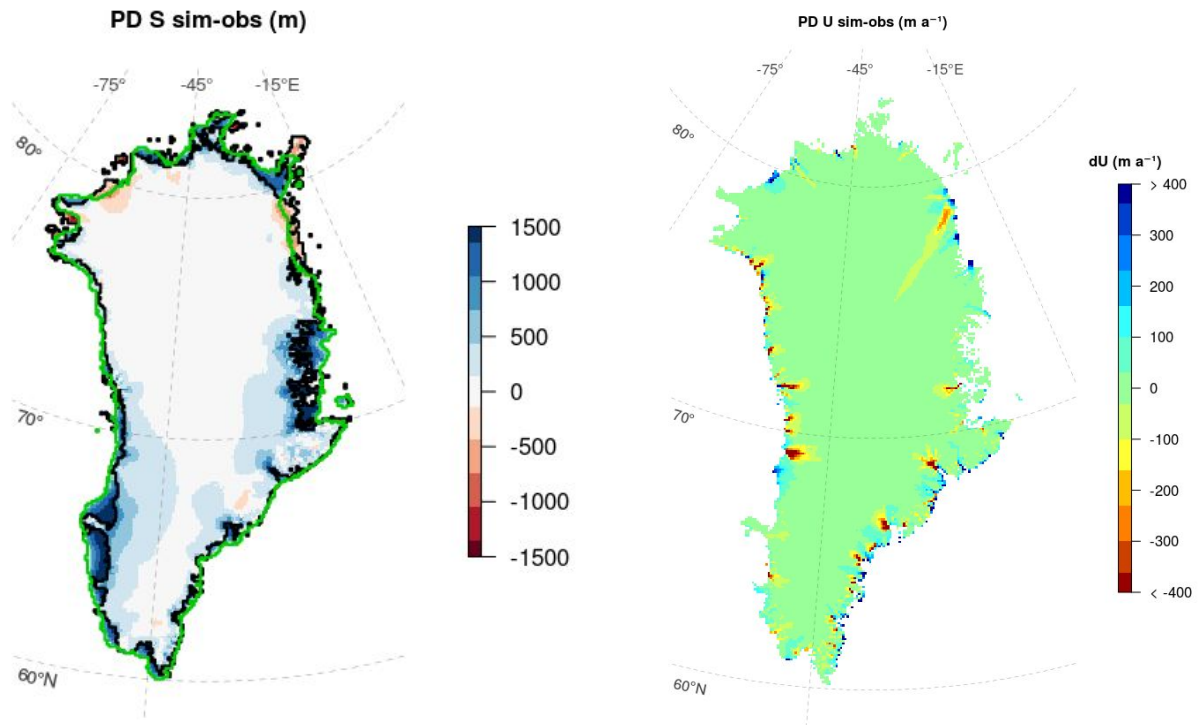


New Fig. 6 (old Fig. 4). Evolution of the ice thickness (a), basal melt (b) and ice velocity (c) averaged within the regions A (red lines) and B (blue lines), simulated in presence of submarine melt during MIS-3 ( $\kappa=8 \text{ m a}^{-1} \text{ K}^{-1}$  experiment). Grey lines in panel b) show the contribution to surface mass balance (accumulation minus ablation) simulated by the model and averaged over the regions A and B. Dashed lines in the same panel show the potential contributions that would be observed if regions A and B were ice covered. The black contour on the side map represents the LGM maximum extent for the  $\kappa=8 \text{ m a}^{-1} \text{ K}^{-1}$  experiment.

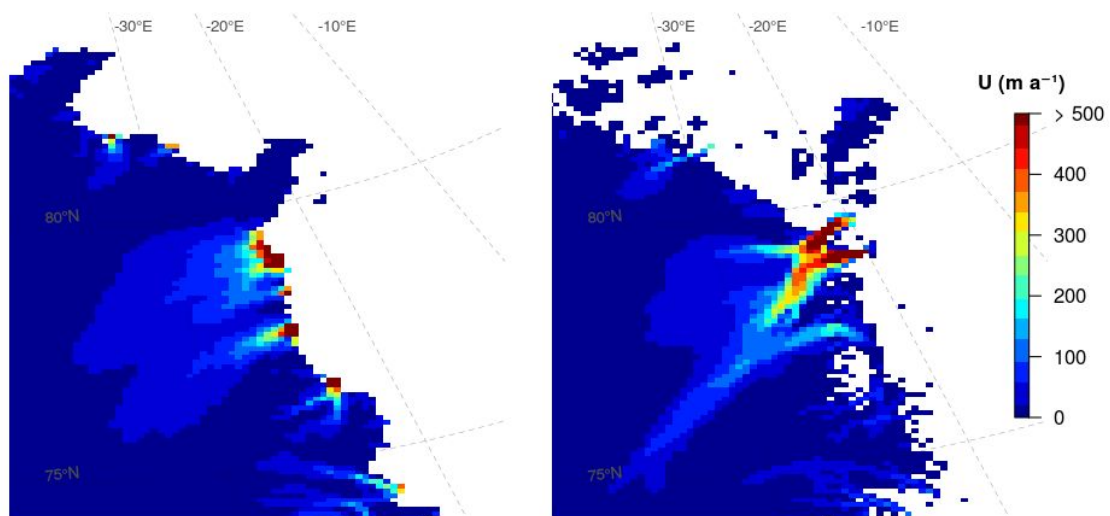


New Fig. 7 in the MS. Simulated GrIS extent at the LGM for different oceanic forcings compared to other glacial reconstructions. Colored lines follow the color scale of new Fig. 3 in the MS (old Fig. 1). Solid black line refers to the maximum glacial extent simulated by Lecavalier et al. (2014), calibrated to match the minimum LGM configuration (Funder et al., 2011) in the northeast. Dashed black line represents the expected maximum glacial extent at the northeast sector as inferred from various geological data (Evans et al., 2009; Arndt et al., 2015, 2017; Winkelmann et al., 2010).

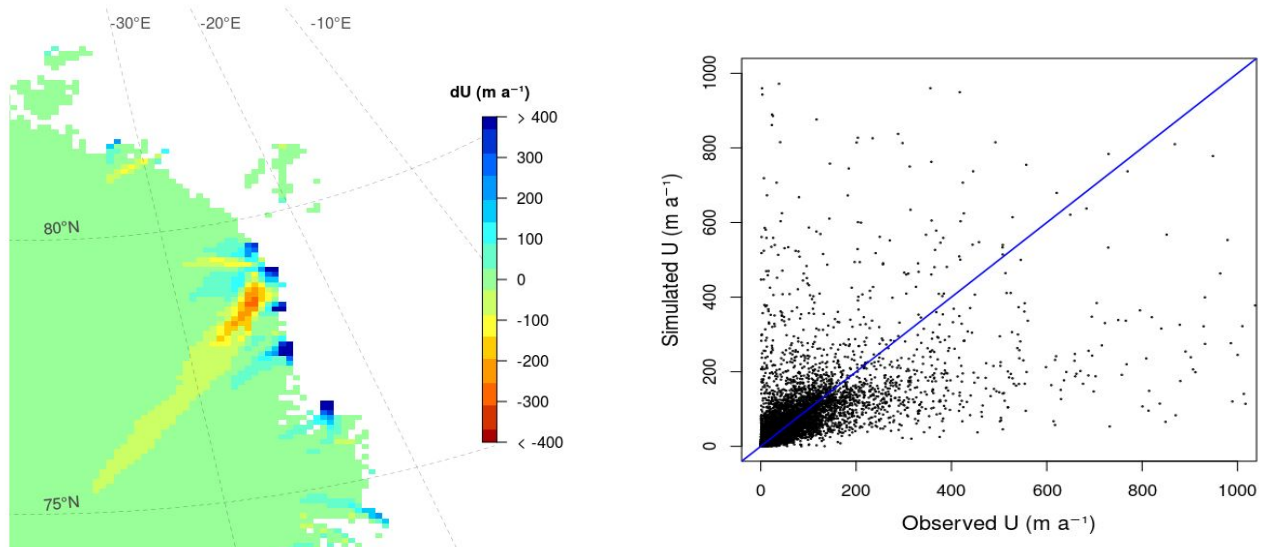




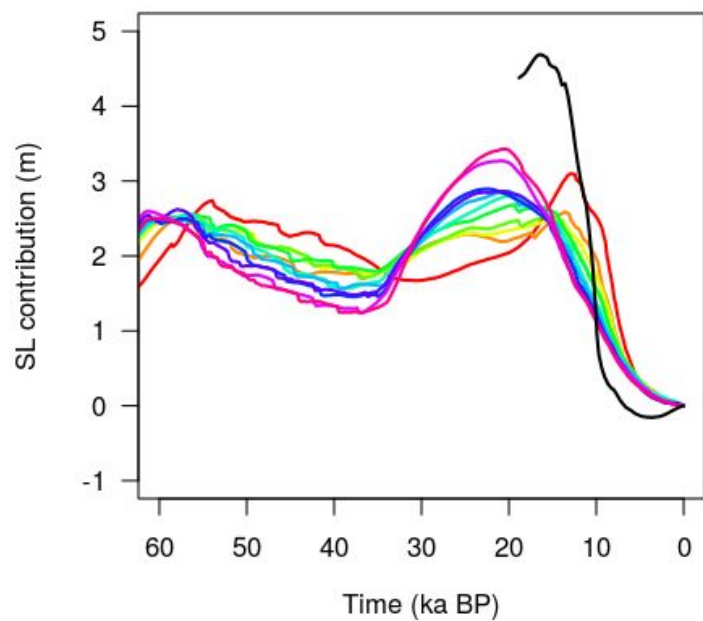
New Fig. S1 in the MS. Simulated minus observed GrIS surface elevation (left panel) and GrIS ice velocity (right panel) for the PD. Green and black lines on the left represent simulated and observed GrIS extents, respectively. Surface elevation data are taken from Schaffer et al. (2016); ice velocity observations from Joughin et al. (2018). Both maps are produced for the  $\kappa = 8 \text{ m a}^{-1} \text{ K}^{-1}$  experiment. However, the choice of another oceanic sensitivity  $\kappa$  would have little effect on the simulated-observed discrepancy.



New Fig. S2 in the MS. Present-day simulated (left panel) and observed (right panel) velocities for the NEGIS sector. Observed data are taken from Joughin et al. (2018).



New Fig. S3 in the MS. Simulated-observed present-day velocities for the NEGIS sector (left panel) and its scatterplot (right panel). Blue line refers to the perfect match between model and data.



New Fig. S4 in the MS. Evolution of the GrIS sea-level contribution for the last 60 kyr. Colored curves refer to the color scale of new Fig. 3 in the MS (old Fig. 1). The black curve refers to the GrIS sea level contribution for the last deglaciation modelled by Lecavalier et al. (2014).