Response to Maria Zeitz (RC1)

We want to thank the reviewer for their effort and the very helpful comments on our manuscript. Below find detailed answers to all comments. The reviewer's comments are in blue, our answers in black and in *italic* we show specific modifications to the original manuscript.

General comments:

While this is not the first study on the topic of Greenland Ice Sheet stability, it is to my knowledge the first study which takes the full glacial-interglacial temperature range into account. Even so, while the stability of the Greenland Ice Sheet has been studied with different models and different forcings, many details seem to depend on the type of model, the coupling to the atmospheric conditions and the exact type of forcing. Therefore, it is worthwhile adding another piece of information, in my opinion.

The manuscript is generally well written, clear and easy to understand, however, I have a few methodological and logical concerns.

1) The authors claim that the initial state of the warming branch at a temperature anomaly of -14 K is an LGM-like state and compare it to reconstructions of LGM ice extent. However, some important features do not match LGM, particularly the insolation and the sea level. There is a gap between the attempt of an idealized analysis of the stability landscape and the connection to a realistic paleoclimate, which is not easily bridged. I suggest that the authors use more careful language at this point, so that the idea that the "cold" initial state should be equivalent to the actual LGM configuration of the Greenland Ice Sheet doesn't arise. In addition, a sensitivity analysis with LGM levels of insolation or sea level might be beneficial towards comparing the modeled "cold" state with LGM ice extent reconstructions.

We appreciate your comments regarding the characterization of our "LGM-like" initial state and its comparison with a realistic LGM state. We agree that further clarification would strengthen the paper. For the sake of clarification, the temperature anomaly relative to present-day for our LGM-like state is -12 K of regional summer air temperature, not -14 K.

Your suggestion of a sensitivity analysis with LGM insolation and sea level levels is a valuable addition. We have therefore conducted five new 60-kyr spin-up simulations with a regional summer air temperature anomaly of −12 K, varying insolation and sea level, and revised the manuscript as described below. According to the relative sea level (RSL) reconstruction by Waelbroeck et al. (2002), RSL during the LGM was 120 m below present-day values (±13 m uncertainty). During the subsequent 5 kyr, regional summer air temperatures increased by only ca. 2 K (Buizert et al., 2018), while sea level had risen to 80 m below present-day values. Given this substantial change in boundary conditions with minimal temperature variation, we performed the following experiment permutations under identical temperature forcing as the LGM-like state:

- Present-day (PD) insolation with PD sea level (the simulation presented as the LGM-like state in the original manuscript)
- PD insolation with sea level at -120 m
- PD insolation with sea level at -80 m
- LGM insolation with PD sea level

- LGM insolation with sea level at -120 m
- LGM insolation with sea level at -80 m

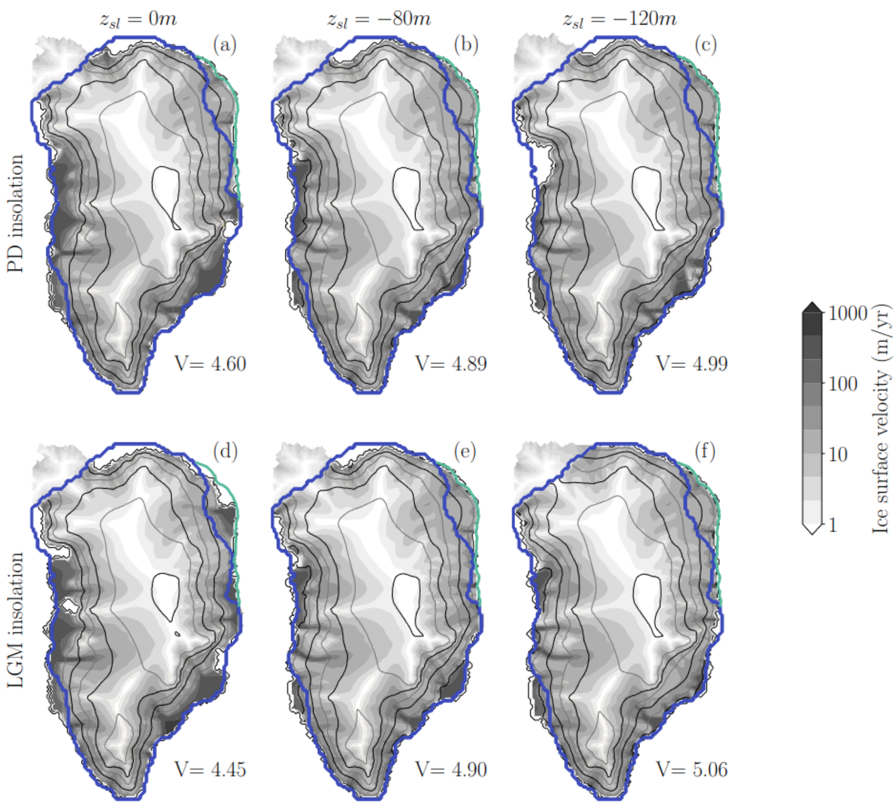

Figure R1.1: Final state of spin-up simulations under −12 K regional summer air temperature anomaly forcing. The upper row shows simulations with present-day insolation; the lower row shows simulations with 20 kyr BP insolation. The first column represents present-day global sea level conditions, while the second and third columns show simulations with global sea levels 80 m and 120 m below present-day values, respectively. Volume above flotation (in millions km³) is displayed at the bottom of each panel. The black contour lines indicate the surface elevation every 500 m starting from 0 and with a thicker line every 1000 m. The blue lines indicate the ice-sheet margin of Lecavalier et al. (2014) and the green line is the maximum extent of grounded ice from the full glacial extent (18-16 kyr BP) in the northeast region by Leger et al. (2024).

Fig. R1.1 displays the different equilibrium cold states corresponding to each of these six cases under a −12 K regional summer air temperature anomaly forcing. The primary difference lies in the amount of floating ice along the western and southeastern margins. The simulation with reduced insolation and lowest sea level (Fig. R1.1e) most closely approximates LGM boundary conditions. Under these conditions, the majority of previously floating ice is grounded, resulting in the highest volume above flotation: 5.06 million km³, equivalent to an anomaly of 4.81 m of sea level equivalent (SLE)—a value consistent with previous studies.

More importantly, the ice-sheet geometry remains similar across all reconstructions and falls within reconstructed paleoclimatic margins, consistent with the original simulation. Notably, the northeastern margin, where the first tipping point occurs, exhibits virtually identical behavior across all simulations.

This allows us to address the reviewer's comment presented later:

"L328ff: As far as I understand the sea level remains constant in these simulations. How would a decrease of sea level to a realistic number during LGM affect the MISI-driven bifurcation?"

Indeed, sea level remains constant throughout the equilibrium diagram to isolate the temperature effect on the ice sheet, as explained at line 156. As noted throughout the manuscript, the MISI-driven bifurcation point value is strongly influenced by how ocean-ice interactions are modeled and calibrated, resulting in considerable uncertainty in its precise value. Similarly, we acknowledge that redoing the bifurcation diagram with different sea-level values could potentially alter the bifurcation point. However, given the consistent northeastern margin configuration observed across our LGM-like state simulations, such changes would likely be minor (and inside the ocean forcing uncertainty range illustrated in Fig. C1) rather than substantial modifications to the overall system behavior.

To verify this assessment, we performed a new warming branch of the stability diagram (focusing only on the temperature range around the first bifurcation point) using a quasi-equilibrium simulation (forcing rate of $3 \cdot 10^{-5}$ $K \cdot yr^{-1}$). This simulation takes the LGM reconstruction from Fig. R1.1f as initial state (LGM insolation and -120 m of sea level). The result is shown in Fig. R1.2, where the main difference compared to the original simulation is a shift in the bifurcation point value (as expected and within the uncertainty range) and a slightly more gradual behavior at the beginning of this transition (around -8.5 K). This second aspect is due to the fact that starting from a slightly larger volume, there is initially a thinning of the ice thickness before MISI is subsequently triggered. The volume difference after crossing the bifurcation point is due to the grounding line remaining slightly more advanced in this simulation than in the original one. These changes in the initial ice-sheet state also generate changes in the bathymetry, which in turn modify the exact equilibrium positions of the grounding line. Nevertheless, we observe that the overall system behavior is preserved. In a transient trajectory from LGM conditions to the present, there would be an increase in temperatures, but also in insolation and global sea level (with the respective contribution from each ice sheet). Therefore, it is reasonable to expect that the exact bifurcation point would lie somewhere between these two idealized states. However, determining the precise transition would be more appropriate for a transient study focused on reconstructing the deglaciation. All in all, figure R.1.2 shows that a decrease of sea level to a realistic number during the LGM does not alter the MISI-driven character of the bifurcation.

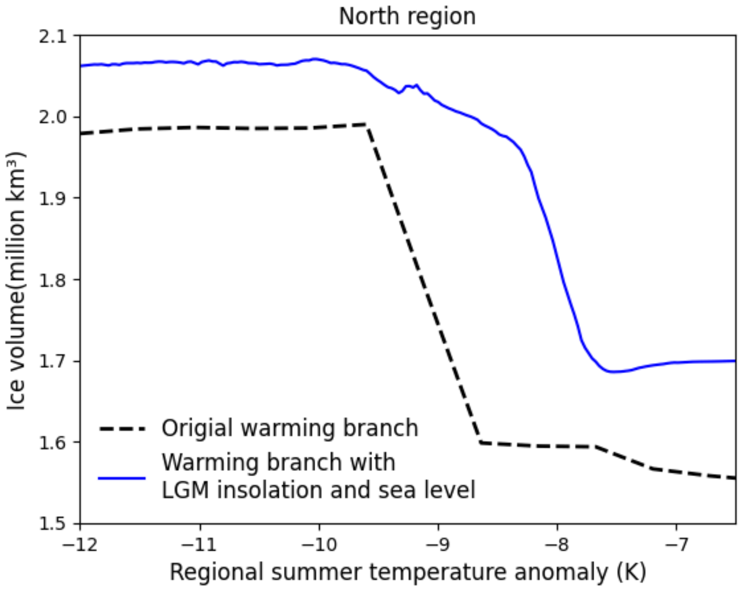

Figure R1.2: Total ice volume in the North region in quasi-equilibrium simulations with a forcing rate of $3{\cdot}10^{-5}$ $K{\cdot}yr^{-1}$. The dashed black line shows the original simulation from the manuscript. The blue line shows a new run with LGM insolation and sea level, using the initial state shown in Fig. R1.1f.

Here is the detailed explanation of the changes in the manuscript related with this comment:

1. We have included LGM sensitivity analysis (Fig. R1.1) as an appendix in the manuscript.

2. We will now talk about “LGM-like state”instead of “LGM state” throughout the text.
3. In Section 2.4 (L156), when first introducing the LGM-like state, we will add a more explicit statement: *"It is important to note that this LGM-like state is an idealized configuration designed to analyze the effect of temperature on ice-sheet stability in isolation from other major paleoclimate forcings such as changes in sea level, insolation, or atmospheric CO2 that are relevant to achieve a realistic LGM state."*

L162: As mentioned above, the comparison to the LGM reconstructions might imply to the reader, that the starting point for the warming branch is indeed an LGM climate. Please clarify this paragraph.

4. In Section 2.4 (L160), when comparing the cold (LGM-like) state with a reconstruction of the LGM our aim was to put in context the initial state of the diagram. Even if lacking important climatic boundary conditions of the LGM, the model achieves a reasonable representation of the state. We will clarify this point in order to avoid confusion.

2) The authors undertake some efforts to distinguish different feedbacks which influence or even trigger the tipping of the Greenland Ice Sheet. The claim that the first tipping is driven by ice ocean interaction and MISI in the north-west is well supported by the data in the following subsection. However, the claim in line 267 “Finally, at ~+1.5 K elevation and albedo feedbacks are triggered and the SMB drops abruptly and becomes negative.” is not supported by a similarly thorough analysis. The data shown in Figure 6 is not sufficient to support the claim of very specific feedbacks at play. Similarly, in line 270 an increase in sub-shelf melting, a reduction of ice shelves and a margin retreat are mentioned, which isn’t visible in Figure 6 either (nor in any other of the presented figures).

I appreciate the attempts to disentangle the different mechanisms at work for the two different tipping points. I would suggest strengthening analysis in section 3.2 and backing it with a more thorough analysis of the data.

We appreciate the reviewer's observation. We focus our detailed analysis on the first tipping point as it represents a novel finding. In contrast, this second tipping point driven by melt and albedo feedbacks has already been characterized in previous literature (Robinson et al., 2012; Höning et al., 2023; Bochow et al., 2023; Petrini et al., 2025; Pattyn et al., 2018; Noël et al., 2021; Boers et al., 2025), and our results align with these established findings.

We agree that Figure 6 does not explicitly show the different feedbacks; it serves to distinguish the forcing mechanisms that trigger them (differentiating between oceanic and atmospheric drivers). Moreover, these statements cannot rely solely on what figures show. What we can affirm is that the abrupt volume loss preceded by a SMB decline (with the decrease in basal melting, which shows that basal processes play no role in the volume reduction) confirms atmospheric processes as the primary driver. Additionally, at ca. +1.5 K, the abrupt transition from positive to negative SMB values can only be possible with atmospheric feedbacks playing a role. According to previous literature, the feedbacks that trigger this tipping point are albedo and elevation feedbacks, both of which are included in our experiments. Based on this, we can conclude that at ca. +1.5 K these atmospheric feedbacks are activated.

It is true that this description is largely based on the findings of previous studies on this topic, therefore we will modify the text to make this clear. Complementarily, and following also a suggestion from referee #2, we have now explored the stability diagram also by considering the oceanic and atmospheric forcings separately. Figure R2.2 (in the referee #2 response) shows the former hysteresis diagram together with the OCN only and ATM only new realizations. It clearly shows that the second tipping point can only be simulated in presence of the atmospheric feedbacks described above.

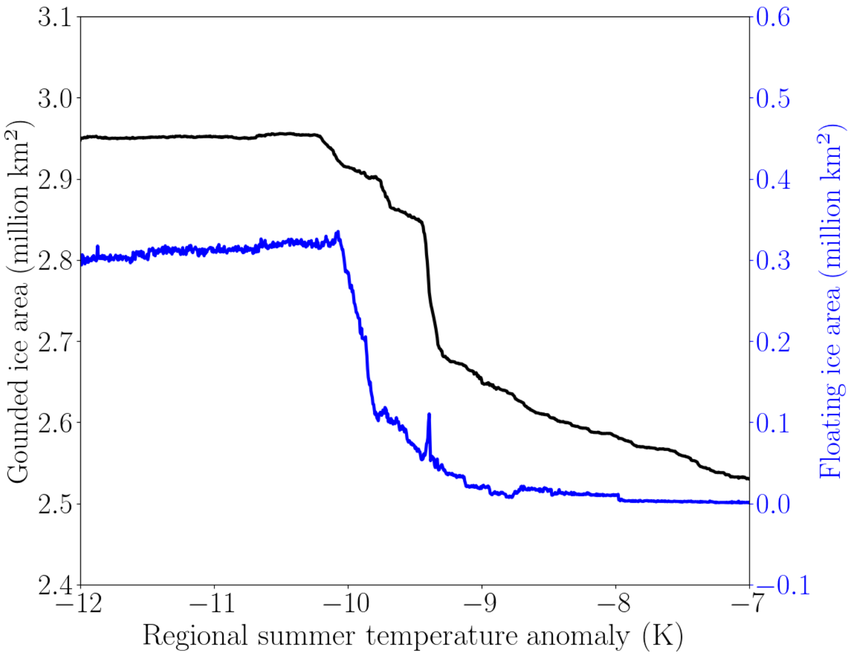

Figure R1.3: Ice area (grounded and floating) in the warming branch of the quasi-equilibrium simulation of $1{\cdot}10^{-5}$K$\cdot$yr^{-1} in the temperature range of the first bifurcation point.

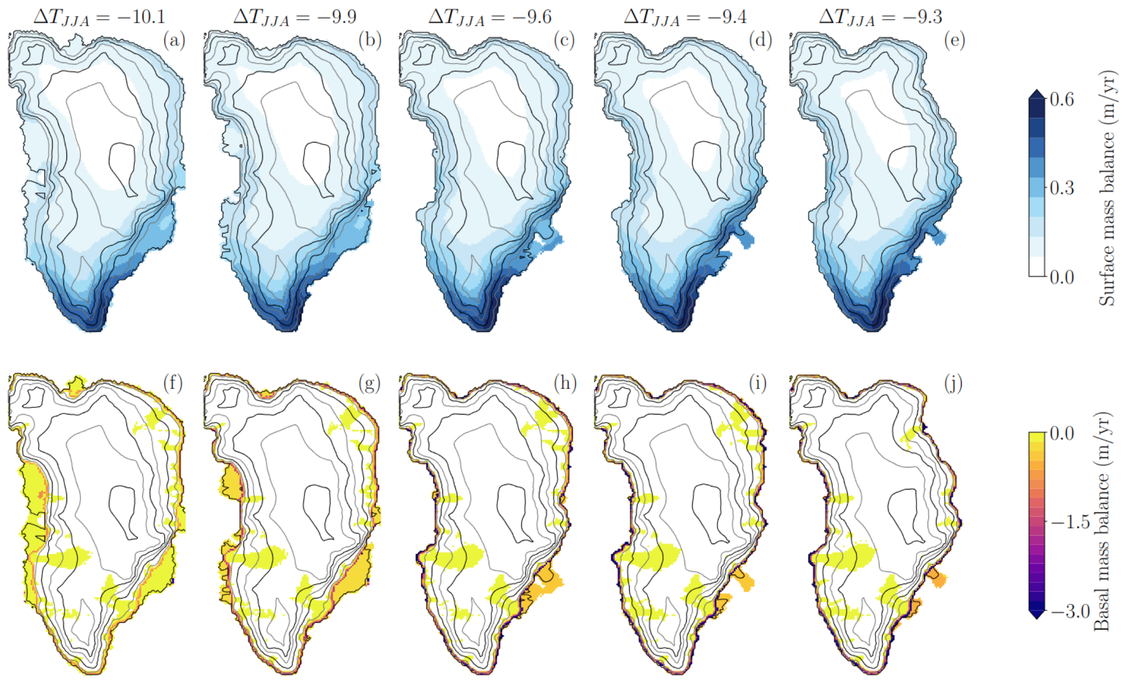

Figure R1.4: a)-e) Surface mass balance and f)-j) basal mass balance at different temperatures from the warming branch of the quasi-equilibrium simulation of $1{\cdot}10^{-5}$K$\cdot$yr^{-1} before the first tipping point. The black contour lines indicate the surface elevation every 500 m starting from 0 and with a thicker line every 1000 m. Note that the higher values in basal melting in f)-j) highlight the grounding line.

Regarding line 270, we will add Figs. R1.3 and R1.4 to an appendix in the manuscript. Fig. R1.3 shows the grounded and floating area in the warming branch of the quasi-equilibrium simulation of $1{\cdot}10^{-5}$K·yr^{-1} (black line in Fig. 3a in the manuscript). There, we can see that after a very small loss in grounded ice, there is a significant loss of floating ice, followed by an abrupt retreat of grounded ice (and the grounding line). In Fig. R1.4 we see the evolution of the ice-sheet surface and basal mass balance before the first tipping point (Fig. R1.4a-e and f-i) and just after (Fig. R1.4e and j). We can see that there is basal melting at the grounding line, at the base of the ice shelves and in some regions under the ice sheet where there is a water layer. As temperatures increase, basal melting also increases both at the grounding line around the ice-sheet margin and below the ice shelves. This causes the retreat of the ice shelves until they remain only in reduced form in the southeast. Finally, the melting becomes so high that the northeast retreats abruptly.

Detailed Comments

L25 and l36: consider citing Solgaard et al. (2012) among the other studies on the topic of Greenland tipping

This is a good suggestion, thank you. We will add Solgaard and Langen (2012) in this two sentences in the introduction:

"Ice-sheet modelling studies furthermore suggest that the GrIS shows multistability and hysteresis with respect to the temperature forcing (Solgaard and Langen, 2012; Robinson et al., 2012; Höning et al., 2023)."

"In addition, the hysteresis behavior of the GrIS has only been studied under temperatures above present-day values (Solgaard and Langen, 2012; Robinson et al., 2012; Bochow et al., 2023; Höning et al., 2023)."

L79: Calving seems to play some role in the crossing of the tipping points. Therefore, I would appreciate if the Von Mises stress criterion for calving would be motivated a little better and ideally discussed.

We chose this approach because it is physically based on the principle that calving is governed by the tensile stress regime at the ice-sheet front and has demonstrated good performance in reproducing observed calving rates across different glaciers (Morlighem et al., 2016; Choi et al., 2018; Goelzer et al., 2017). While other calving criteria exist that represent the calving phenomenon with greater complexity, the computational cost of applying these methods at the full ice-sheet scale makes it preferable to employ simpler criteria such as Von Mises, which nonetheless yields satisfactory results consistent with previous studies of the GrIS using Yelmo (Tabone et al., 2024).

We will add this motivation to the revised manuscript.

L85: please define “present day” with an exact time period

We define the present-day conditions as the average conditions between 1958 and 2001. However, it seems that in one version of the drafts this was mistakenly omitted from the sentence, so we thank the reviewer for noticing it and we have now added it on line 85:

“The temperature and humidity over the ocean around Greenland are imposed as boundary conditions, for which the climatological mean from the years 1958–2001 of the ERA-40 reanalysis (Uppala et al., 2005) to represent present-day conditions.”

L112: Do I understand correctly that the humidity over the ocean is held constant at the boundary conditions and does not increase with increasing temperatures? Does the precipitation within the simulation box adapt to changing temperatures? How much additional effect would be expected from a humidity correction at the boundaries?

No, the humidity at the boundary conditions is not constant. Over the ocean, two variables are prescribed: air temperature T and relative humidity r. Then the specific humidity Q at the boundary is given by:

$Q = Q_{sat}(T) \cdot r$;

where $Q_{sat}(T)$ is the saturation specific humidity, which depends on temperature following the Clausius-Clapeyron equation. Therefore, across the diagram, the temperature at the boundaries changes and the relative humidity is constant, but the specific humidity changes according to the temperature variations. In REMBO, the choice to maintain constant relative humidity rather than specific humidity is based on the fact that specific humidity has a stronger temperature dependence.

The total amount of precipitation is then calculated as a function of the specific humidity, thus accounting for the changing temperature and following the equation:

$$P = (1 + k|\nabla z_s|)\left(\frac{Q}{\tau}\right).$$

where ∇z_s is the gradient of the surface elevation, τ is the water turnover time in the atmosphere and k is an empirical parameter (Robinson et al., 2010). Therefore, even if the relative humidity is constant, the total precipitation changes with temperature. We will clarify this in the manuscript.

We implement a spatially heterogeneous bias correction approach for precipitation to address the systematic biases in the simulated precipitation patterns produced by REMBO, which exhibits enhanced precipitation amounts in the southwestern and northern regions of the domain (Robinson et al., 2010). On the other hand, implementing a boundary humidity adjustment would effectively modify the total amount of available humidity throughout the domain. However, such an approach would fail to address the underlying spatial heterogeneity in the precipitation bias patterns, whose origin lies in the regional character of the model and its coarse spatial resolution.

L138: express the equation also in terms of $\Delta TJJA$. I was confused by the numbers if line 155 and line 159 because it slipped my attention that $\Delta Tocn$ refers to the annual temperature anomaly instead of the "normal" input.

Fixed.

L139: "… for purely floating ice shelves…" The sentence is a bit unclear, as all ice shelves are attached to the ice sheet, by definition. Do you mean in simulation cells with purely floating ice shelves? Does the ice model Yelmo have a mechanism for partially floating and partially grounded ice within a simulation cell?

When modeling the oceanic forcing, we distinguish between the grid cells located at the grounding line and the rest of the ice shelf. We understand that the original sentence may have been confusing (we were using "purely floating ice shelves" to differentiate the corresponding cells from those at the grounding line), so we have rewritten it as follows:

"On the other hand, the sub-shelf basal melting rate B_{sh} is lower than that at the grounding line."

L162: As mentioned above, the comparison to the LGM reconstructions might imply to the reader, that the starting point for the warming branch is indeed an LGM climate. Please clarify this paragraph.

Following general comment 1, we have revised the text to prevent confusion as outlined in our previous response.

L174: I suppose the mass balance is negative at the margins of the former ice sheet, not the equilibrium state reached at ΔT = +4 K.

The mass balance is indeed negative at the margins of the equilibrium state at ΔT_JJA = +4 K. We see that the original formulation may be ambiguous given the minimal ice extent at this temperature anomaly. However, persistent ice caps remain in the easternmost region and at the southern tip under these conditions. Our reference to negative marginal mass balance refers to these residual ice masses. We have revised the sentence at line 174 as follows:

"The mass balance of the remnant ice caps is almost zero in their interior (all incoming accumulation is melted away) and negative at their margins, especially in areas in contact with the ocean (not shown)."

Figure 1: Why is the shaded ice not taken into account for the volume calculation?

We have also simulated the ice evolution on Ellesmere Island due to its influence on the size, shape, and dynamics of the GrIS. However, when calculating the volume of the GrIS, ice on this island is typically not included, both for present-day and past conditions. Therefore, although we simulated ice evolution there as well, we excluded it from our calculations in order to maintain consistency with the domain that is normally included in GrIS volume estimates.

In order to make this clearer in the text, we modify the final sentence in the Figure 1 caption:

"Note that the lightly shaded area in the northwest (over Ellesmere Island) indicates the part of the simulated ice sheet that is not taken into account for the volume calculation in order to be consistent with the usual GrIS domain."

L239: “These results clearly show the impact of atmospheric feedbacks related to elevation and albedo…” which is not so clear to me from the data. Please clarify or add additional information.

See below.

L266ff: This paragraph explains the changes in the surface mass balance, however, all claims in the paragraph are not supported by data which would be available to the reader. This contrasts with the thorough discussion on grounding line shape and ice dynamics on the previous section. Please, support the claims with data or with references.

We will revise these two comments (L239 and L266) in order to clarify that the statements are supported by different studies showing that the albedo and elevation feedbacks generate the presence of this bifurcation point for a warming above present-day values, such as Robinson et al. (2012), Höning et al. (2023), Bochow et al. (2023), Petrini et al. (2025), Pattyn et al. (2018), and Noël et al. (2021). Please see also our answer to general comment # 2.

L298f: I find this sentence surprisingly hard to read. I also kept wondering, why the volume decreases if the tipping point isn’t reached yet. Consider rewriting for clarity.

Volume reductions in the western and eastern regions show linear behavior due to the absence of bifurcation points in these areas. Only the northern region exhibits threshold behavior due to its particular bathymetry. Ice losses in the eastern and western regions result from the warming and the increase in BMB and start earlier due to floating ice at their margins, which experiences a higher BMB. The original formulation of the sentence may have been misleading regarding the absence of tipping points in these regions. We rewrite the sentence in this way in the manuscript in order to make it clearer:

"At approximately -10.5 K, the ocean warming triggers the sub-shelf melting, leading to a gradual volume decline. Volume losses are concentrated in the western and southeastern regions, where the ice at the margin is floating and exhibits a higher sensitivity to oceanic forcing at lower temperatures. While these zones experience nearly linear losses up to approximately +2 K, in the northern region –where the ice is grounded and its thickness is higher– the volume remains nearly constant until the temperature anomaly reaches -9.4 K (in this quasi-equilibrium simulation), when the abrupt ice loss occurs."

L301f: I'm not sure if the relationship between initial melting and the acceleration of ice flow is sufficiently explained. And I didn't quite understand how the spike in calving is related to the previous.

Starting from a quasi-equilibrium state, an increase in basal melting generates an imbalance that causes grounding-line retreat. Given the retrograde bedrock configuration, this results in an increase in cross-sectional area, which generates enhanced ice flux towards the exterior. This triggers a MISI, with consequent grounding line retreat, transition of ice from grounded to floating conditions, progressive thinning that finally leads to calving, resulting in a substantial reduction in ice volume. Therefore, the increased calving represents a consequence of the MISI-driven ice flux enhancement. We have reformulated the text to provide a more detailed explanation.

L328ff: As far as I understand the sea level remains constant in these simulations. How would a decrease of sea level to a realistic number during LGM affect the MISI-driven bifurcation?

Addressed above. See our answer to general comment #2.

L358: Is there any interpretation for the existence of intermediate states? How does this compare to the intermediate states found in Robinson et al. 2012?

Conceptually, these intermediate states differ from those reported by Robinson et al. (2012), where the intermediate branch corresponded to the new equilibrium states reached by starting with intermediate (transient) initial conditions. In their framework, the upper branch (referred to in this work as the warming or retreating branch) represents the equilibrium states at different temperatures starting from an initial state similar to pre-industrial conditions. The lower branch (referred to here as the cooling or regrowth branch) represents equilibrium states at different temperatures when starting from a virtually ice-free initial state. The intermediate branch accounts for equilibrium states reached starting from intermediate initial conditions, with the light red area in Figure 1 of Robinson et al. (2012) indicating the volume range of initial states that, when subjected to different temperatures, ultimately reach equilibrium on the intermediate branch.

In contrast, in our simulations, the intermediate states belong to the cooling (lower) branch, meaning that they are reached starting from an ice-free initial state. The key difference from Robinson et al. (2012) is that their results show abrupt regrowth requiring lower temperatures, whereas our simulations demonstrate regrowth beginning at higher temperatures and proceeding through a two-step process. Nevertheless, it is interesting to highlight that the intermediate state in both studies has a similar shape (covering south and central Greenland) and volume (in relative terms).

This demonstrates that the combination of albedo reduction following considerable ice retreat and low precipitation in northern Greenland makes it more difficult for this region to recover the ice.

Therefore, the existence of these intermediate states in our work is attributed to two factors: first, the stability of an ice-sheet configuration where only southern and central Greenland remain ice-covered (in agreement with Robinson et al., 2012), and second, the capacity of the ice sheet to recover at higher temperatures. To fully understand the reason for this earlier recovery, a comparative analysis between the results of different studies that have examined the hysteresis of the GrIS would be necessary, including factors such as surface mass balance, bedrock and surface elevation, among others. However, we think that such a detailed comparison is beyond the scope of this study.

We will expand the discussion related to these intermediate states in order to make it clearer.

L360ff: I understand that further analysis might be beyond the scope of the study. I would still be curious to hear more about why the initial ice volume is only regained at temperatures of at ΔT = -5 K.

If we understand correctly, with “initial ice volume” the reviewer is referring to the ice volume at ΔT_{jja} = 0 K (3.4 million km³) in the warming branch. As illustrated in the figure below and Fig. R1.5 of the manuscript, the warming branch equilibrium state at ΔT_{jja} = 0 K (hereafter W0) has indeed the same total volume as the cooling branch equilibrium state at ΔT_{jja} = -5K (hereafter C-5). However, these states don't have the same ice distribution. The W0 state shows more ice in the NEGIS area and a complete ice coverage over Scoresby Sund, while the C-5 state has a higher ice thickness along the western margin, resulting in equivalent total volumes despite different geometries.

We attribute the delayed volume recovery (requiring cooling to -5K) to the same physical mechanisms responsible for the existence of hysteresis in the -9K to 0K temperature range (lines L245-260). Specifically: (1) coastal margin irregularities prevent ice expansion in certain coastal regions under the C-5 conditions, and (2) bathymetric peak in the NEGIS region acts as a pinning point in the W0 state, allowing ice accumulation. On the other hand, the lower temperatures in C-5 state allow ice thickening in areas with more regular coastal geometry.

Thus, while these states have similar total volumes, they have a different configuration and the volume equivalence is somehow fortuitous. The need for such pronounced cooling to achieve a similar ice-sheet volume underscores the influence of the non-linear feedback mechanisms previously outlined.

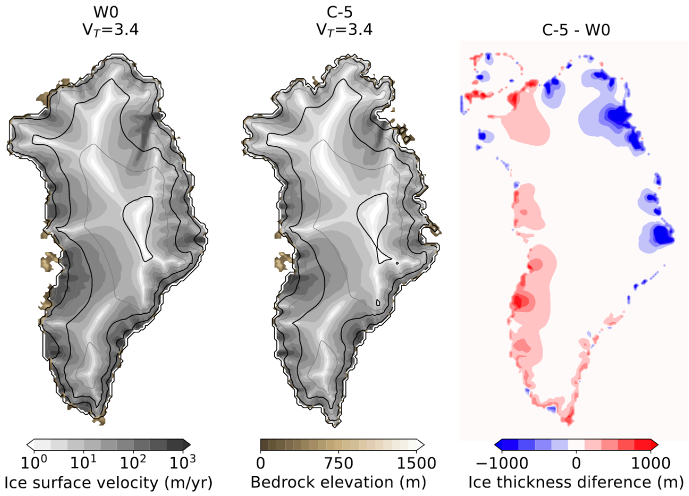

Figure R1.5: From left to right: warming branch equilibrium state for $\Delta T_{jja} = 0$ from the branching-off experiment (W0); cooling branch equilibrium state from the branching-off experiment at $\Delta T_{jja} = -5K$ (C-5); and ice thickness difference between C-5 and W0. V_T is in million km³ and represents the ice-sheet total volume of each state. The black contour lines indicate the surface elevation every 500 m starting from 0 and with a thicker line every 1000 m.

Appendix A: How does the present-day state compare to the equilibrium states at ΔT = 0 K on the cooling and the warming branch? What does it mean for the stability of the present-day state, that it is in the unstable zone between two stable branches?

This is a very good question. The present-day state shown in Appendix A has a volume of 3.16 million km³, which falls between the volume of the equilibrium state in the warming branch (3.37 million km³) and the equilibrium state in the cooling branch (2.81 million km³) for that same temperature forcing (ΔT_{jja} = 0). Moreover, when comparing the ice sheet in these three simulations (Fig. R1.6), the present-day state is in between the two equilibrium branches.

Different studies have pointed out that the GrIS is not in equilibrium neither at present nor during pre-industrial times (Yang et al., 2022). This study suggests that the current state results from the ice sheet having retreated beyond its present margin during the Holocene Thermal Maximum (when temperatures exceeded current levels) and subsequently regrowing. Our results also indicate that the present state of the GrIS is not the product of gradual and slow temperature changes since the LGM, as would be the case for the equilibrium states shown in the hysteresis diagram. Instead, achieving the current state starting at the LGM requires an overshooting of present temperatures followed by subsequent cooling, which would have caused the GrIS to transition between the two equilibrium branches since the LGM as it happens during the last deglaciation. We have also considered this question and we plan to address it in future work, including transient simulations of the last deglaciation (starting from a realistic LGM state, actually). Therefore, we do not elaborate on this topic in detail in the present manuscript.

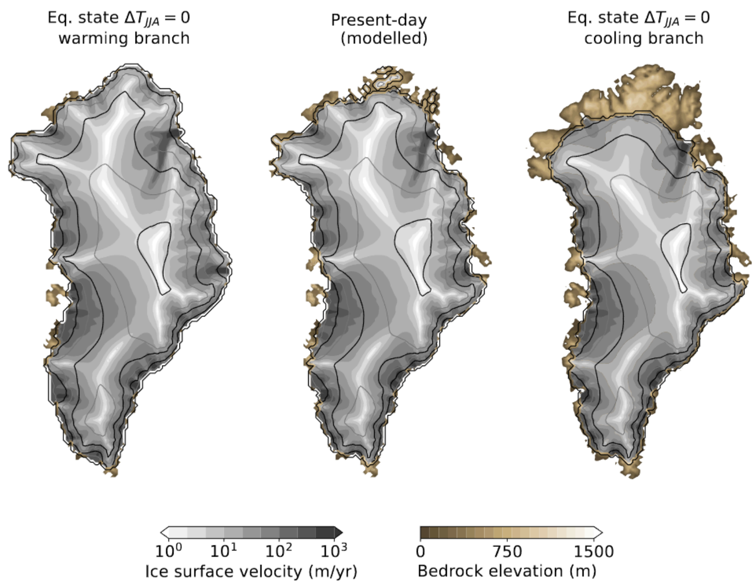

Figure R1.6: From left to right: warming branch equilibrium state from the branching-off experiment; present-day steady-state simulation (described in Appendix A in the manuscript); cooling branch equilibrium state from the branching-off experiment. The black contour lines indicate the surface elevation every 500 m starting from 0 and with a thicker line every 1000 m.

References

Boers, N., Liu, T., Bathiany, S., Ben-Yami, M., Blaschke, L. L., Bochow, N., ... & Smith, T. (2025). Destabilization of Earth system tipping elements. *Nature Geoscience*, 1-12.

Buizert, C., Keisling, B. A., Box, J. E., He, F., Carlson, A. E., Sinclair, G., & DeConto, R. M. (2018). Greenland-wide seasonal temperatures during the last deglaciation. *Geophysical Research Letters*, *45*(4), 1905-1914.

Choi, Y., Morlighem, M., Wood, M., and Bondzio, J. H.: Comparison of four calving laws to model Greenland outlet glaciers, The Cryosphere, 12, 3735–3746, https://doi.org/10.5194/tc-12-3735-2018, 2018.

Goelzer, H., Robinson, A., Seroussi, H., & Van De Wal, R. S. (2017). Recent progress in Greenland ice sheet modelling. *Current climate change reports*, *3*(4), 291-302.

Lecavalier, B. S., Milne, G. A., Simpson, M. J., Wake, L., Huybrechts, P., Tarasov, L., ... & Larsen, N. K. (2014). A model of Greenland ice sheet deglaciation constrained by observations of relative sea level and ice extent. *Quaternary Science Reviews*, *102*, 54-84.

Leger, T. P., Clark, C. D., Huynh, C., Jones, S., Ely, J. C., Bradley, S. L., ... & Hughes, A. L. (2023). A Greenland-wide empirical reconstruction of paleo ice-sheet retreat informed by ice extent markers: PaleoGrIS version 1.0. *Climate of the Past Discussions*, *2023*, 1-97.

Robinson, A., Calov, R., & Ganopolski, A. (2010). An efficient regional energy-moisture balance model for simulation of the Greenland Ice Sheet response to climate change. *The Cryosphere*, *4*(2), 129-144.

Waelbroeck, C., Labeyrie, L., Michel, E., Duplessy, J. C., Mcmanus, J. F., Lambeck, K., ... & Labracherie, M. (2002). Sea-level and deep water temperature changes derived from benthic foraminifera isotopic records. *Quaternary science reviews*, *21*(1-3), 295-305.

Yang, H., Krebs-Kanzow, U., Kleiner, T., Sidorenko, D., Rodehacke, C. B., Shi, X., ... & Lohmann, G. (2022). Impact of paleoclimate on present and future evolution of the Greenland Ice Sheet. *Plos one*, *17*(1), e0259816.

Bochow, N., Poltronieri, A., Robinson, A., Montoya, M., Rypdal, M., & Boers, N. (2023). Overshooting the critical threshold for the Greenland ice sheet. *Nature*, *622*(7983), 528-536.

Petrini, M., Scherrenberg, M. D., Muntjewerf, L., Vizcaino, M., Sellevold, R., Leguy, G. R., ... & Goelzer, H. (2025). A topographically controlled tipping point for complete Greenland ice sheet melt. *The Cryosphere*, *19*(1), 63-81.

Pattyn, F., Ritz, C., Hanna, E., Asay-Davis, X., DeConto, R., Durand, G., ... & Van den Broeke, M. (2018). The Greenland and Antarctic ice sheets under 1.5 C global warming. *Nature climate change*, *8*(12), 1053-1061.

Noël, B., Van Kampenhout, L., Lenaerts, J. T. M., Van de Berg, W. J., & Van Den Broeke, M. R. (2021). A 21st century warming threshold for sustained Greenland ice sheet mass loss. *Geophysical Research Letters*, *48*(5), e2020GL090471.

Höning, D., Willeit, M., Calov, R., Klemann, V., Bagge, M., & Ganopolski, A. (2023). Multistability and transient response of the Greenland ice sheet to anthropogenic CO2 emissions. *Geophysical Research Letters*, *50*(6), e2022GL101827.

Robinson, A., Calov, R., & Ganopolski, A. (2012). Multistability and critical thresholds of the Greenland ice sheet. *Nature Climate Change*, *2*(6), 429-432.

Robinson, A., Calov, R., & Ganopolski, A. (2010). An efficient regional energy-moisture balance model for simulation of the Greenland Ice Sheet response to climate change. *The Cryosphere*, *4*(2), 129-144.