

## Response to Reviewer #1:

Please see response to the referee's comments in blue font.

Guo and colleagues present an equilibrium simulation of the Marine Isotope Stage 3 (MIS3) with a fully coupled climate model. The simulated climate is very stable and representative of an interstadial climate state with a strong AMOC and relatively high temperatures over Greenland. A stadial climate with a weaker AMOC and lower Greenland temperatures cannot be simulated, not even with typical stadial CO<sub>2</sub> concentrations. Sensitivity studies with even lower CO<sub>2</sub> concentrations and flat ice sheets support the hypothesis that the NorESM model is very far away from a potential threshold where the climate changes from interstadial to stadial conditions.

The topic of the paper – MIS3 climate state and variability – fits well into the scope of *Climate of the Past* and is very relevant for the community. There are few fully coupled MIS3 simulations to date and the presented simulations is therefore very valuable as it adds more data points to the parameter space of glacial forcings and thus helps to understand (1) MIS3 climate variability and (2) the model dependence of glacial climate states.

I recommend the article for publication after some suggested revisions: I believe, the study could be put more into context with existing MIS3 simulations and reconstructions (see general comments below), and a few issues require clarification before publication (see specific comments below).

We thank the reviewer for his/her thorough assessment and constructive comments on our manuscript. We respond to the reviewer's comments below point by point.

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General Comments: \_\_\_\_\_

The presented MIS3 simulation could be more embedded into the existing literature, both in terms of simulations and existing proxies. Throughout the text, especially in Sect 3.2, the analysis is very descriptive and there are very few comparisons with the existing MIS3 simulations in terms of surface temperature response, sea-ice patterns or AMOC state. The authors mention Barron&Pollard (2002), Van Meerbeeck et al (2009) and Brandefeld et al (2011) in the introduction. It is true that there are not so many coupled simulations with MIS3 boundary conditions, but there are some more MIS3 control simulations available that have been published as reference simulations for hosing experiments, e.g. Xiao Zhang et al (GRL, 2014) or Kawamura et al (Science Advances, 2017, here the information is somewhat hidden in the Supplementary Information).

Often the authors compare their simulations to existing simulations and proxies from the LGM. This is an obvious choice, since there are more simulations and reconstructions available for the LGM than for MIS3. But then these comparisons can be a bit confusing/misleading, since we would not expect the climate state of MIS3 and LGM to be the same. I therefore suggest that the authors go carefully through their manuscript again and check in each case what they want to obtain from the LGM comparison. Can some insight be gained from the LGM /MIS3 differences? If possible it would also be good to have a few more comparisons with existing MIS3 reconstructions, there is e.g. a recent study by Sessford et al (Paleoceanography, 2018) on water masses and sea-ice in the Denmark strait.

I believe, a more thorough comparison with the existing simulations and MIS3 proxies can help to highlight where the presented MIS3 simulation provides new insight and thus make the study more interesting and relevant.

> Thanks for the insightful comments and very relevant references. Following the reviewer's suggestion, we have expanded the discussion with a more in-depth comparison of our results with published MIS3 simulations and proxy records; these include the modelling studies of e.g. Merkel et al., 2010; Xiao Zhang et al., 2014; Kawamura et al., 2017,

and proxy studies of, e.g. Böhm et al. 2015; Sessford et al., 2018. There are certainly more MIS3 proxy based studies, however we choose to focus on comparison to the Nordic Seas region.

As for the reviewer's comment on comparison with previous LGM simulations, we do agree that the two periods are distinct, e.g. MIS3 features reduced ice volume and a relatively warm climate compared to the LGM. We have carefully revised the manuscript, and removed the discussion on the comparison of SST and AMOC strength with LGM PMIP studies. We retained the comparison of near surface temperature - considering the close proximity of the two periods in time, and the much larger number of LGM proxy/modelling studies, we think it is useful to make the comparison (e.g. Van Meerbeeck et al. 2009), with caveats in mind of course. We've therefore added the following to the updated manuscript: "Compared to the amount of MIS3 studies, there is a rich literature on both the simulation and reconstruction of the LGM climate. With both similarities as well as apparent differences with regard to the external forcing and the climate, it can be useful to compare the climate of the two periods..."

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Specific Comments:

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p.1, ll.12-15: '[. . .] questioning the potential for unforced abrupt transitions [...]' In the text you phrase that conclusion quite carefully and refer to the model dependence of MIS3 climate (in)stability. In the abstract, the formulation is perhaps a bit too general, the model dependency should appear here, too.

> Following the reviewer's comment, we have edited the abstract making it consistent with the conclusion and including the reference to model dependency.

p.3 - Model description: Would it not be easier to directly describe NorESM1-F rather than describing first how NorESM1-M differs from CCSM4 and then to describe how NorESM1-F differs from NorESM1-M?

> NorESM family of models is based on CCSM4, and NorESM1-M is the first documented version of the family. NorESM1-F is developed based on the NorESM1-M version. To clarify, we've rewritten the first sentence in Section 2.1 as "The NorESM family is based on the Community Climate System Model version 4..."

p.4, ll.28 – p.5, ll.12: It is not quite clear to me, how the exact MIS3 ice sheets are obtained. Are they assembled from different sources? Why not take them all from the same reconstructions? And why is the Barents Sea so problematic? According to its mean depth it should be open also with 70m lower sea-level, no? Are there conflicting reconstructions?

> The global ice sheet data used in our simulation was kindly provided by Lev Tarasov; the data documenting the individual ice sheets were either published separately (as cited in the manuscript) or unpublished (e.g. the Eurasian ice sheet). Regarding Barents Sea, intuitively it should be open with a 70 m reduction of sea level; however, it is likely that there could be grounded ice that actually closes the sea - geological evidence on the existence of grounded ice is sparse and not firm though, as mentioned in the manuscript. Given that the reconstructions are uncertain and the potential impact is significant, we have chosen to discuss this in more detail in the manuscript.

p.6, ll.20: based on what do you decide that the trend is small? Is there also a threshold value such as for deep ocean salinity in the next sentence?

p.6, ll.30: is the sea-ice drift acceptably small?

> There is not any 'hard' threshold to evaluate the model drift, as far as we are aware of; this is true for both modern as well as paleoclimate simulations. We believe that the trends of the evaluated metrics are acceptably small, although we do think that a longer integration would certainly be an advantage. Unfortunately, we are limited by computational time to run the model longer. However, in the revised manuscript we have removed the sentence "For the NorESM MIS3

simulation, we deem the aforementioned global mean ocean cooling trend to be small.”

p.7, ll.8-12: It is interesting though, that the final simulated MIS3 AMOC is still stronger and deeper than at PI, and the AABW cell is also weaker than at PI, even though AABW is saltier and more ventilated. I’ll come back to this issue in a later comment.

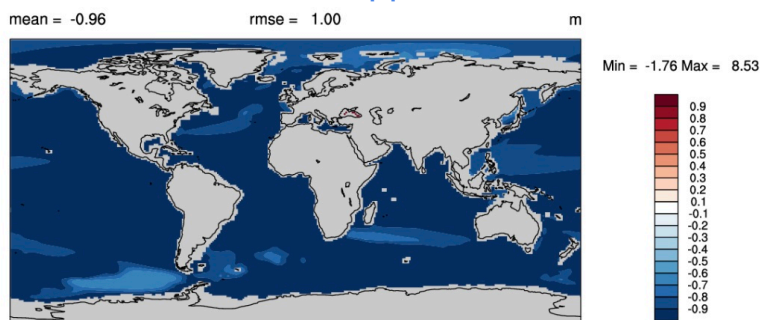
> We respond to the reviewer's comment regarding AMOC below.

p.7, ll.19-22: Are there no MIS3 studies available for the comparisons? (see also general comments)

> There are indeed MIS3 studies available for comparison. As discussed above, we have added the following to the revised manuscript: "For comparison, The CCSM3 MIS3 simulation (with 35 ka boundary conditions) by Merkel et al. (2010) reported a cooling of 3.4 °C, whereas using a different version of CCSM3 configured with 38 ka boundary conditions, Zhang et al. (2014b) reported a cooling of 3.5 °C. A stadial simulation with 44 ka boundary conditions (Brandefelt et al., 2011) shows a much cooler climate, e.g., 5.5 °C compared to the recent past."

p.8, ll.22: The warming in subpolar gyre seems to be more of a dipole. Ist the NAC shift a north-south shift? Can it be seen in the barotropic stream function in Fig. 11?

> The NAC shifts during MIS3 compared to that in PI can be seen from the sea level anomaly map shown below (MIS3 minus PI). The shift is not apparent from the barotropic stream function which is the vertical integration of volume transport for the whole water column, whereas NAC is located in the upper few hundred meters.



p.9, ll.22: Why is there more runoff into South China Sea, when precipitation is decreased according to Fig. 8?

> This is mainly because a new river routing map is generated in our MIS3 configuration due to the change of land-sea mask, leading to large catchments and several arguably 'artificial' rivers flowing into the South China Sea. We've added the new river routing map in the supplementary material, and have also updated the original text accordingly as "The fresh surface water in the South China Sea during MIS3 is due to increased runoff, that is related to the newly generated river routing in this region owing to the change of land/sea mask."

p.9/10, AMOC and hydrography section: I find this section somewhat confusing for many reasons.

(1) I think, the LGM comparisons are not very helpful here (see also general comments), as the MIS3 AMOC is expected to be very different from the LGM AMOC. A comparison with the LGM AMOC and hydrography would be more helpful in the discussion, when speculating about reasons for a stable or unstable AMOC. If available, MIS3 comparisons would be more helpful here. From Böhm et al (2015) it should at least be possible to get a qualitative picture of the distribution of northern and southern sourced waters from the eNd measurements.

> Following the reviewer's comment, we have removed the LGM comparison here (see also our response to the reviewer's general comments). Instead we focus on comparison to MIS3.

We have added the following MIS3 reference to the revised manuscript: "Zhang et al. (2014b) reported a similar strengthening of AMOC during MIS3 (38 ka boundary conditions), e.g., 15.4 Sv which is 1.5 Sv stronger than their PI control simulation, and is much weaker than our simulated strength of AMOC at MIS3. Zhang et al. (2014b) also simulated a shallower upper cell of AMOC, in contrast to the NorESM simulation."

We have also included a comparison with Böhm et al (2015): "The AMOC during interstadials is accompanied by active deep water formation in the North Atlantic, with persistent contributions from the northern sourced water."

(2) I am surprised, that the North Atlantic salinity does not increase more than the global average of 0.6 g/kg. If the mechanism that makes the MIS3 AMOC stronger than the PI AMOC is the same in NorESM than on Muglia&Schmittner (2015) and Klockmann et al (2016/18), I would have expected a much larger salinity increase both at the surface and in the deep North Atlantic.

> Intuitively, one would expect a larger increase of Atlantic salinity, given a stronger AMOC and closure of the Bering Strait. However, it is possible that the MIS3 circulation can move salt across basins, for example from the Atlantic to the Pacific, causing the salinity increase in the North Atlantic < 0.6 g/kg. One possibility is that the enhanced AMOC leads to a larger salt export at the southern basin boundary, if the change in surface salt import from Indian Ocean does not fully compensate for the change in NADW related export. Klockmann et al. (2016) also showed less increase of salinity (relative to the global addition of salt due to sea level lowering) in the North Atlantic although with a stronger AMOC at LGM (c.f. their Fig. 5f). While the effect of closing the Bering Strait is clearly visible in the spatial SSS changes (Fig. 9) – showing e.g. a freshening in the Bering Sea and tendency to more saline Canadian Arctic and western North Atlantic – it is not unlikely our model underestimates this effect as a consequence of the model's routing of freshwater on land. The river routing map (added to the supplementary material) shows large catchments for the Canadian Arctic and North Atlantic and only a small catchment for northeast Pacific. Likely the lack of Atlantic salinity response can be mitigated by manually correcting/tuning the routing map (e.g. based on geological evidence) to route more freshwater into the North Pacific instead of Arctic, something we will consider in future studies. Another potential factor may be changes in the moisture transport across Central America. Given the size of the current paper, we think a more detailed

investigation of the hydrological cycle covering the above aspects should be better done elsewhere.

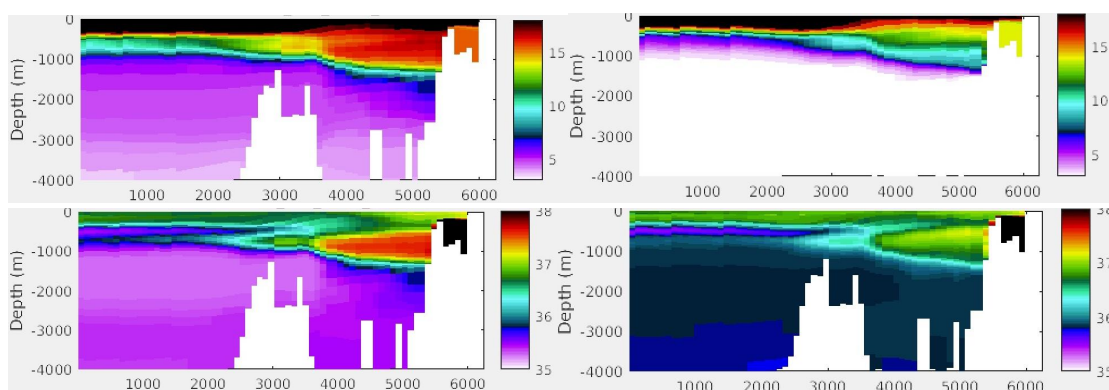
(3) If more warm NADW is present below 3000m, where does the very cold anomaly in the deep North Atlantic come from? Is that Overflow water then?

> First of all, the global ocean is colder during MIS3 than PI (e.g. 1.7 deg C). Also, AABW, although reduced in volume in the deep Atlantic, is colder and contributes to the cold anomaly therein.

It is a possibility that overflow water contributes to the cold anomaly in the deep North Atlantic since large parts of the Nordic Seas are not ice covered in the MIS3 simulation allowing dense water formation in high latitudes to continue.

(4) Can the anomaly at 500-800 m really be attributed to the Mediterranean Outflow? I would expect the outflow at depths around 1100 m.

> yes; the figure below shows a cross-section from the western Atlantic to the Mediterranean along the same latitude: left column - PI; right column - MIS3; top row - temperature; bottom row - salinity. One can see that the change of Mediterranean outflow during MIS3 contributes to the temperature/salinity anomalies at 500-800 m.



A weaker signature of Mediterranean outflow in the MIS3 simulation is plausible as evaporation over the Mediterranean basin is expected to be



reduced under the colder climate i.e. less outflow/inflow is necessary to maintain the freshwater balance of the Mediterranean Sea.

(5) What is ideal age? Is it the time since the water mass was in contact with the surface?

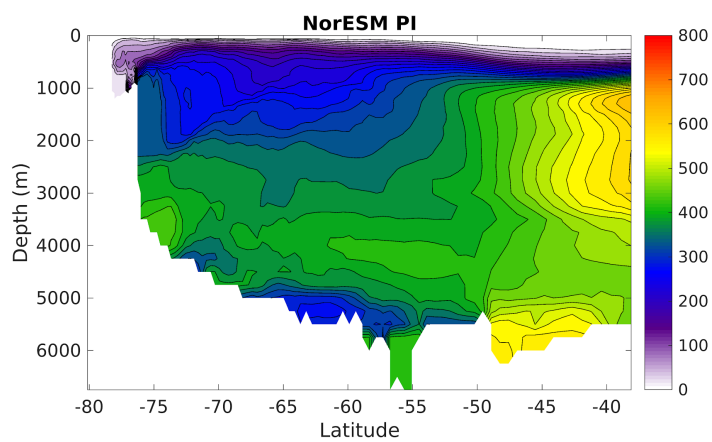
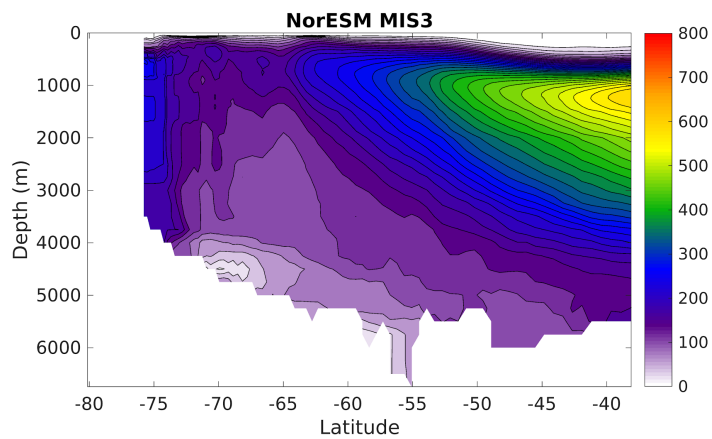
> Yes - it is the time since the water mass last made contact with the surface. We have included this in the revised manuscript.

(6) If the AABW formation is determined by increased sea-ice formation and brine- rejection, how can it be so well ventilated? I understand from the Ferrari (2014) paper, that AABW was very poorly ventilated because it was upwelled under the ice with little exchange with the atmosphere, and that this is one reason for the glacial CO<sub>2</sub> draw- down.

> The reviewer is right. The increased sea ice formation and the associated brine rejection contribute to the formation and salinification of AABW, but one cannot directly link it to enhanced ventilation.

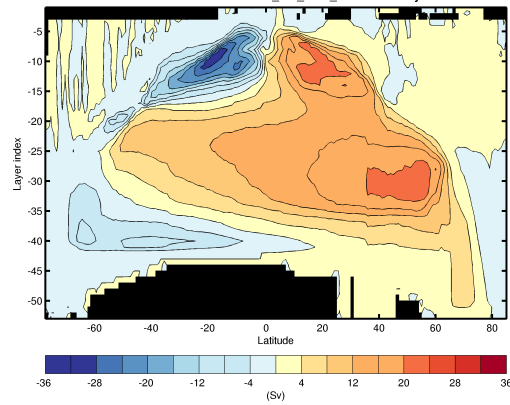
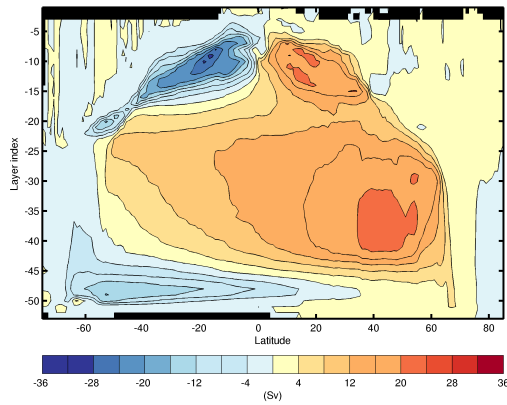
The model does show an enhanced ventilation in the Southern Ocean during MIS3, e.g., the figure below shows the zonal mean ideal age in the Southern Ocean for the MIS3 and PI experiments. It is evident from the figure that AABW is more ventilated during MIS3 compared to PI. The strengthening of ventilation can be caused by processes such as changes in air-sea heat/salt flux, wind etc.

We've therefore modified the text accordingly, and included the figure below into the supplementary material.

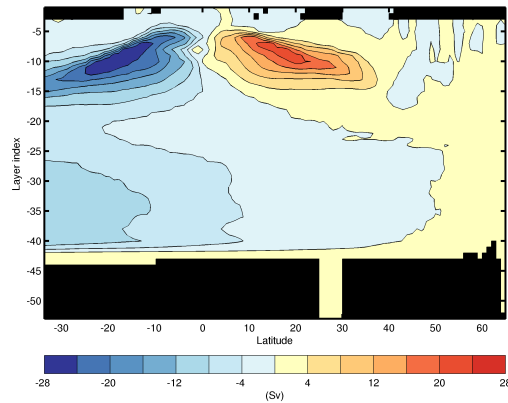
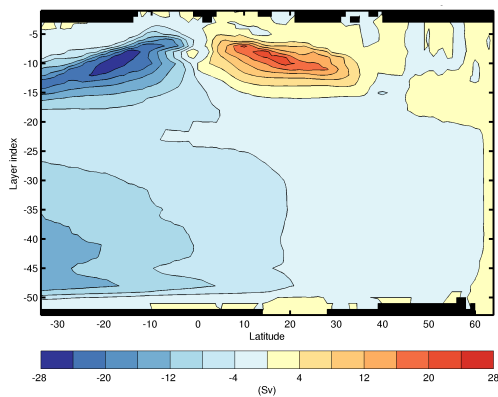


(7) If AABW ventilation and formation increases but the lower overturning cell weakens with less AABW reaching the North Atlantic, where does the AABW go? Is there more AABW in the Pacific?

> Yes. The lower overturning cell associated with AABW in the Atlantic is weakened, but globally it is strengthened - see below the global MOC in isopycnal space for MIS3 (left panel below) and PI (right panel below).



In the Pacific and Indian Ocean, it is also clear that MIS3 (left panel below) features a stronger deep cell associated with AABW compared to that in PI (right panel below).



p.13, ll.11-14: Same as comment (6) above: I understand from the Ferrari (2014) paper, that AABW was very poorly ventilated because it was upwelled under the ice with little exchange with the atmosphere, and that this is one reason for the glacial CO<sub>2</sub> draw-down. So how does that fit with a well ventilated AABW?

> Please see our response to comment (6) above.

p.13, ll.27-31: Same as comment (2) above: If the mechanism that makes the MIS3 AMOC stronger than the PI AMOC is the same in NorESM than on Muglia&Schmittner (2015) and Klockmann et al (2016/18), I would have expected a much larger salinity increase both at the surface in the subpolar gyre and in the deep North Atlantic.

> Please see our response to comment (2) above.

p.14, ll.12-13: Xiao Zhang et al (2014) on the other hand find that MIS3 is close to disequilibrium in their simulation.

> Yes, we agree that the MIS3 simulation by Xiao Zhang et al (2014) is close to disequilibrium, although this study explores the AMOC instability with the freshwater approach. We have added the following into the updated text: “In addition, with freshwater flux as the external forcing, the MIS3 simulation reported by Zhang et al. (2014b) is close to disequilibrium and model bi-stability.”

p.15, ll.1-11: The sensitivity simulations appear quite short, especially the ones with the reduced ice sheets. What determined the length of the simulations? I would argue that the simulations are not in equilibrium, yet. Especially the simulation with 140 ppm could well still be declining – e.g. some of the simulations in Klockmann et al (2018) had about 1500 years of spin up, before the state transitions occurred. I would disagree with the statement that the AMOC strength is unaffected. The responses are small, but both ice sheet reductions and the very low CO<sub>2</sub> lead to an AMOC weakening. Whether the weakening is strong enough to produce a stadial climate state is then another question.

> We agree with the reviewer that the sensitivity experiments are not long enough for quasi-equilibrium states. Computing resource is one concern. We terminated the sensitivity experiments with reduced ice sheet, as we did not see any sign/trend of further AMOC reduction and growth of sea ice for these two experiments. We also agree that the 140 ppm experiment, still with a declining trend, has the potential to reach a weak mode of AMOC. However, given that 140 ppm is already a very low level (e.g. compared to that in Zhang et al., 2017 & Klockmann et al., 2018) and that the sensitivity experiment has run for > 1000 years, it indicates that our simulated MIS3 climate is far away from the bifurcation point. We have added the following to the discussion: “..., despite that the simulations are limited in length (200 to 1000 years) and we therefore cannot fully exclude that further equilibration could bring the climate into a more unstable regime.”

It is true that we should not claim that the strength of AMOC is not affected in the sensitivity experiments. We have modified this in the updated manuscript, e.g. "... AMOC is only slightly reduced" for the ice sheet sensitivity experiments, whereas for the CO2 sensitivity experiments, "the AMOC, although weakened by several Sv, still remains strong".

Figures: I would say there are already almost too many figures. But still I would like to ask for a figure showing also the deep water formation sites on the Northern hemisphere. I think they could be very helpful for understanding the AMOC stability. I personally would find that more informative than e.g. the insolation in Fig. 1.

> Following the reviewer (and another reviewer)' comment, we have moved some text/figures to the supplementary material to enhance the readability of the manuscript. We have also added a map showing both the MIS3 and PI mixed layer depth (an indication of deep water formation region) into the supplementary material.

\_\_\_\_\_ Technical comments: \_\_\_\_\_

p.1, ll.8: remove parentheses around 'by ~13%'

> removed.

p.1, ll.21-23: reformulate sentence for clarity

> We have rephrased the sentence as "Correlated with the rapid warming of Greenland temperature (up to 15  $^{\circ}$ C within a few decades during the stadial-to-interstadial transition), the North Atlantic and Nordic Seas are subject to abrupt climate transitions as interpreted from a number of marine sediment cores...".

p.3, ll.32: Add a reference for HAMOCC

> added, e.g. Maier-Reimer (1993), Maier-Reimer et al. (2005)

p.5, ll.5 and throughout the text, for the convenience of the reader, take care to distinguish between ice sheets/land ice and sea ice. Sometimes only ice is used.

> We've searched through the manuscript to make sure that different types of ice are distinguished.

p.5, ll.10: MSI3 should be MIS3

> corrected.

p.14, ll.24: Remove parentheses around citation

> removed.

Figure 11: solid contours should indicate negative values and dashed contours positive. They are mixed up in the caption.

> revised.

Figure 14: This may be a matter of taste; I find it more appropriate to have non diverging colourmaps for the absolute T and S sections in (a) and (c).

> We would prefer to stay with the current colour maps in this figure, as we would like to have a consistent colour map with previous NorESM evaluations, e.g., Bentsen et al. (2013), Guo et al. (2019).

References: Xiao Zhang et al (2014), Instability of the Atlantic overturning circulation during Marine Isotope Stage 3, Geophysical Research Letters, <https://doi.org/10.1002/2014GL060321>

Kawamura et al (2017), State dependence of climatic instability over the past 720,000 years from Antarctic ice cores and climate modeling, Science Advances, DOI: 10.1126/sciadv.1600446 + Supplementary Information

Sessford et al (2018), High-Resolution Benthic Mg/Ca Temperature Record of the Intermediate Water in the Denmark Strait Across D-O Stadial-Interstadial Cycles, <https://doi.org/10.1029/2018PA003370>

## Response to Reviewer #2:

We respond to the referee's comments in blue font below.

This is a study that aims to improve the understanding of the climate of Marine Isotope Stage 3 (MIS3) when the millennial-time scale climate variability occurred most frequently during the last glacial period. The authors perform simulations of MIS3 with a comprehensive state-of-the-art climate model and compare the climate and the oceanic circulation, especially the Atlantic Meridional Overturning Circulation (AMOC) with those of the Preindustrial (PI) and the Last Glacial Maximum (LGM). The authors further show the sensitivity of the MIS3 climate and the AMOC to modifications in the boundary conditions, such as the Laurentide ice sheet and CO<sub>2</sub>. The model does not exhibit a threshold-type behaviour of the AMOC under low Laurentide ice sheet and low CO<sub>2</sub> level. This is interesting, and it offers the community a chance to improve our understanding of the AMOC and discrepancies among models. This paper should be published because of two reasons. First, it produces important information of the climate of MIS3, which is still rare compared to that of the LGM. Second, it assesses the sensitivity of the MIS3 climate to modifications in the boundary conditions, which is unique compared to previous studies assessing the sensitivity under the PI and the LGM conditions. Nevertheless, I also feel that the important points of this study are still unclear and the manuscript is too long. Below are some suggestions to improve the manuscript.

We thank the reviewer for his/her thorough assessment and constructive comments on our manuscript. We respond to the reviewer's general and specific comments below point by point.

—General comments—

1. The author should focus more on the results of the sensitivity experiments and perform more analysis in these experiments, since they are the most important and interesting point of this study. In particular, what happens to surface salinity and density over the North Atlantic Deep Water (NADW) formation region and Antarctic bottom water



formation (AABW) region when the Laurentide ice sheet and CO<sub>2</sub> are modified? Since previous studies have shown that changes in surface salinity and density are very important in understanding the changes in the strength and the threshold of the AMOC (e.g. Montoya and Levermann 2008, Oka et al. 2012, Sun et al. 2016, Buizert and Schmittner 2015, Sherriff-Tadano et al. 2018, Klockmann et al. 2018, Galbraith and de Lavergne 2018), analysis on this point is very important. This analysis will also give very useful information in comparing the results of your model with other climate models. Further analysis on the depth of the AMOC, as well as sea ice cover over the North Atlantic and Southern Ocean should be conducted (e.g. Klockmann et al. 2016, Kawamura et al. 2017, Galbraith and de Lavergne 2018).

> We agree with the reviewer that the sensitivity experiments related to external forcing should be of high interest to many. However, in this work, we would prefer not to weigh the sensitivity experiments over the results of the MIS3 control simulation. There is a rich literature on LGM simulations, whereas there are few MIS3 simulations with a state-of-the-art climate model. Besides, most of the previous studies on AMOC bi-stability/abrupt climate change in the last glacial are configured with boundary conditions of either LGM or PI, thereby with a deviation/bias already in the control experiment. We therefore believe that a comprehensive assessment of the simulated MIS3 climate would be necessary and could serve as a useful reference and basis for dedicated, future MIS3 simulation studies with the same model that focus on climate sensitivity. However, we agree with the reviewer that the original manuscript can be more compact, and we have accordingly moved certain results to the supplementary material (see our response to the next comment).

In addition, in response to the reviewer's suggestion on more detailed analysis of the sensitivity experiments, we have added a section in the supplementary material showing the response of SSS, winter sea ice, and AMOC in depth-latitude space. As we have mentioned in the first draft of manuscript, NorESM is in a relatively stable state and stays far away from the threshold for state transitions; as a consequence, the

response of the climate system in the sensitivity experiments are relatively small, as reflected in the changes of metrics mentioned above. Specifically, as the changes in e.g. SSS, sea ice, and AMOC geometry are highly related to the strength of AMOC, which is only weakly reduced in the sensitivity experiments; therefore, significant changes in SSS, sea ice, and AMOC geometry etc. would not be expected. We have also added some discussions in the manuscript.

2. The manuscript is too descriptive and long, which makes the reader difficult to understand the important message of this study. In particular, sections 3.1, 3.2, and 4 are too descriptive. I do understand that these sections show important results, however, they do not give new results and rather follows several previous studies. Unless you compare these simulation results with proxies and other climate models in detail, you should move some part of this section to the Supplementary section. This will help shorten the manuscript.

> As also mentioned in our response to the previous comment, we have moved quite a bit of results in sections 3.1, 3.2, and 4 to the supplementary material; these include the time series of sea ice, several atmospheric diagnosis, the whole section of "Modes of variability", and the discussion on the "stadial" experiment. We believe that the results are presented in a more succinct manner in the updated manuscript.

—Specific comments—

#### 1 Introduction

The authors should state the significance of this study compared to previous studies more clearly in the last three paragraphs. These points are vague in the manuscript. I understand that the simulation of MIS3 is important since most previous studies conducted simulations of the LGM when they explore the glacial climate. However, in the manuscript, the significance of this study compared to previous MIS3 modelling studies is vague. This point should be clarified.

> Following the reviewer' comment, we have rephrased the text in Introduction to highlight the significance of our study.

## 2 Methods

P4 L30-32: I couldn't quite understand this sentence. Do you just mean that the shape of the ice sheet is prescribed in the model?

> Yes, exactly; we have clarified this in the updated manuscript: "NorESM1-F does not have a dynamic land ice component, and the assumed ice sheet extent and elevation during MIS3 compared to present day are prescribed."

P5 L10: MSI3 → MIS3 3 Results

> corrected.

### 3.1 Model spin-up

I agree that the model has almost reached a quasi-equilibrium state. However, I feel that this section is too long, which makes the reader tired. Please consider reducing the amount of this section. (See also Comment 2).

> Following the reviewer' suggestion, we have moved the text/discussion on sea ice to the supplementary material. We think that the rest of metrics are important for the evaluation of model drift, and therefore would prefer to keep them.

P7 L11: Where is the location of the open ocean convection over the Southern Ocean in the MIS3 experiment?

> In the Weddell Sea region and the Pacific and Indian sectors of the Southern Ocean. We have included a figure of austral winter mixed layer depth in the supplementary material.

### 3.2 Simulated MIS3 climate

This section is too descriptive and long. Please consider reducing the amount of this section by moving some part of it to the Supplementary. (See also Comment 2).

> We have done so. Please see our response to the reviewer's general comment 2.

P7 L33-P8 L1: The strengthening of the surface easterly wind stress over the Irminger Sea is also caused by the expansion of the Laurentide ice sheet (Sherriff-Tadano et al. 2018).

> yes, good point; reference cited.

P8 L30-P9 L3: Please mention that the lowering of CO<sub>2</sub> is important in causing the expansion of sea ice and in decreasing the sea surface temperature.

> We have added the following to the beginning of the subsection: "The reduced level of CO<sub>2</sub> during MIS3 is important in lowering SST and in causing the expansion of sea ice."

P9 L27: You may remove the first sentence, which is already mentioned in the Introduction.

> sentence removed.

P9L30-L31: Did you try to say 'The deeper overturning stream function is associated with contracted and weakened AABW'?

> The original statement is a bit misleading; we have rephrased it as "The lower overturning cell associated with AABW is contracted and weakened."

P10L21-L25: Kobayashi et al. (2015) also report similar response in their LGM simulation. The decrease in ideal age of the water is attributed to enhanced open ocean convections over the Southern Ocean. You may cite this paper as well.

> We thank the reviewer for the suggestion of this study. We have referred to this paper in the revised manuscript: "Kobayashi et al. (2015)

reported similar response of LGM water mass age in the Southern Ocean owing to enhanced open ocean convections."

P11L18-L20: Merkel et al. (2010) also shows similar results in their MIS3 interstadial simulation. You may cite this study as well.

P11L32-L33: Is this difference statistically significant?

P12L2-L3: Is this difference statistically significant?

> The study by Merkel et al. (2010) is very relevant and has been cited. Following the reviewer's second general comment, we have moved the section of ENSO/NAM to the supplementary material.

4 MIS3 simulation forced by stadial conditions Please consider reducing the amount of this section by moving some part of it to the Supplementary. (See also Comment 2).

> We have done so. Please see our response to the reviewer's general comment 2.

## 5 Discussion

### 5.1 Simulated AMOC in MIS3

P13L13: I rather use 'bottom water formation' than 'open ocean convection'.

> modified.

P13L26-L30: As far as I know, Montoya and Levermann (2008) first showed the potential role of surface winds over the North Atlantic in intensifying the AMOC, Oka et al. (2012) showed that the LGM surface wind enhanced the AMOC with one model, Muglia and Schmittner (2015) confirmed the study of Oka et al. (2012) by performing analysis with PMIP3 climate models, and Sherriff-Tadano et al. (2018) investigated the processes by which surface winds anomaly induced by the ice sheets enhanced the AMOC. These studies should also be cited in this sentence.

> We thank the reviewer for pointing to a detailed list of very relevant references. We have cited them in the updated manuscript.

P13L31: Hu et al. (2015) investigated the impact of the closure of Bering Strait on the AMOC. This study should also be cited in this sentence.

> added; this reference is indeed very relevant.

## 5.2 MIS3 sensitivity to CO2 and ice sheet size

Results presented in this section are really interesting! As mentioned in Comment 1, I strongly encourage the authors to perform more analysis on these sensitivity experiments (surface salinity, density and sea ice cover over the NADW and AABW formation region, and the depth of the AMOC). Based on these analysis, you may further discuss the possible cause of differences among previous modelling studies. (Also, if possible, it may be interesting to discuss changes in surface air temperature and precipitation in the half-size Laurentide ice sheet. This analysis can provide an uncertainty of the simulated temperature and precipitation anomalies arising from the uncertainty in the shape of the MIS3 ice sheet. Just a suggestion.)

> Please see our response to the reviewer' general comment 1 regarding further analysis on the sensitivity experiments.

Discussion on the changes in temperature/precipitation in the modified ice sheet experiments would definitely be interesting to certain readership, as the reviewer suggested. Meanwhile, we would prefer to keep our focus on the AMOC bi-stability in this section and illustrate the relative insensitivity of NorESM MIS3 climate to external forcing. We would therefore not include this discussion in the manuscript, but rather leave it for future studies or for model intercomparison activities.

P15L13: What do you mean by 'ice inhibiting convection'?

> We mean "... Norwegian Sea are covered by sea ice that inhibits convection through its insulating effect."

## Figures

Fig.2: Can you put labels on the contours?

> Yes, we have put labels on the contour lines 1000 m and 2000 m.

Fig.11: Can you add a figure showing the anomaly? It's difficult to understand the difference between MIS3 and PI from these figures.

> We tried to plot an anomaly map on the stream function, e.g. see the figure below. Comparing the two different ways of presenting, we think the original figure is relatively more straightforward and intuitive in comparing the stream functions during the two periods.

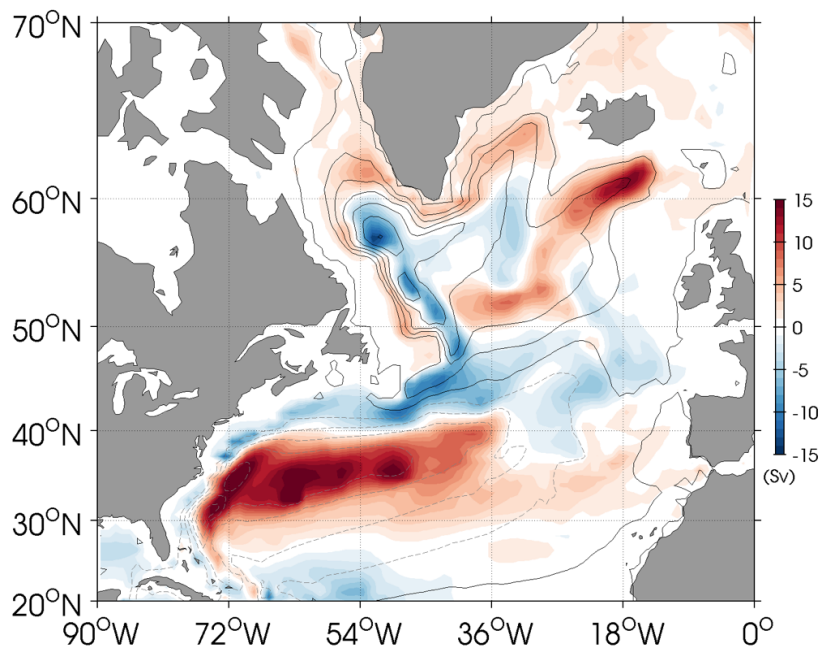


Figure above: MIS3 subtropical and subpolar gyre stream functions (Sv; contours) and difference with PI (Sv; shading)

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## Response to Reviewer #3:

We respond to the referee's comments in blue font below.

In the manuscript cp-2018-165 entitled “Equilibrium simulations of Marine Isotope Stage 3 climate” by Guo et al., the authors compared the simulated climate mean state of Marine Isotope Stage 3 (MIS3) and preindustrial (PI) era using the Norwegian Earth System Model (NorESM). They found a cooler climate in MIS3 relative to PI conditions with a thicker and more expanded sea ice. The AMOC strengthen by 13% with reduced AABW reaching the North Atlantic. Moreover the AABW production actually increases due to the increased sea ice cover in the southern oceans in association to the cooler MIS3 climate. They also show a reduced ENSO and NAM variability. Finally, by doing a few sensitivity simulations by reducing CO<sub>2</sub> concentration or ice sheet height in the North America, they suggest that abrupt transitions of climate from interstadial to stadial state is not likely, and raised the question whether abrupt climate transition would be possible without changes of external forcings. I found this manuscript is well written and easy to follow. The results are interesting to the readership of the Climate Past community. Thus I would like to recommend this manuscript to be accepted after some revision:

We thank the reviewer for his/her overall positive comments on our manuscript. We respond to the reviewer's comments below point by point.

### Comments:

1. The simulations are primarily focus on the PI and MIS3 climate background, thus it is not surprising that the climate states are stable for both conditions. One question the authors did not specifically clearly state is the initial condition of these runs. It seems that both PI and MIS3 runs state from the same ocean initial state, except an increase of the mean salinity for MIS3 run. Is this true? If so, how will this affect the model sensitivity when icesheet height or CO<sub>2</sub> concentration changes?

> What the reviewer commented is true. We state in the manuscript that, for the MIS3 baseline experiment: "As for the PI experiment, the ocean model is initialised with modern temperature and salinity (Steele et al., 2001) with the above mentioned salinity increment applied."; for the sensitivity experiments: "All the sensitivity experiments are branched off and initialised from the MIS3 interstadial simulation, and all other parameters are kept fixed."

The addition of extra salt to the global ocean in the MIS3 simulations must have some effects - although we expect them to be small - on the modelled ocean and climate background state, and therefore the model sensitivity to, e.g. ice sheet height and CO<sub>2</sub> levels. As far as we are aware of, there has not been any study looking into this problem, which merits more investigations. This is highly relevant to any glacial simulations, especially to LGM experiments where PMIP protocols define a global salinity addition of +1 psu.

2. The authors tested the model response of CO<sub>2</sub> reduction by 15 ppmv. The question here is whether the MIS3 stadial climate is caused by CO<sub>2</sub> reduction or by changes of the AMOC? It seems that the authors assumed that the CO<sub>2</sub> reduction is the cause, is it true?

> We did not assume that the CO<sub>2</sub> reduction is the cause of the stadial climate state. We support the wide-accepted view that changes in AMOC (e.g. strong or weak/off mode) are behind the interstadial/stadial climate states. As described in the Introduction, changes in AMOC can be invoked by changes in CO<sub>2</sub> concentrations, as well as other factors such as changes in the size of Laurentide Ice Sheet.

The reason why we do such a sensitivity experiment (15 ppmv lower CO<sub>2</sub>), as also discussed later in the text, is partly motivated by some previous studies (e.g. Zhang et al. 2017; Klockmann et al. 2018) that do indeed show a transition of AMOC mode upon relatively small change of CO<sub>2</sub> level. Another motivation is that if the model is already close to the threshold of mode change, which we did not know beforehand, a small

change of external forcing is able to kick the system into a different state (see also our response to the comment below).

3. Although the experiments done by the authors don't show significant AMOC changes, it seems this is not enough to question the possible multi-equilibria of AMOC, especially the experimental design may not serve the purpose of the authors. A better test is to check whether AMOC has multi-equilibria in the NorESM under glacial condition. If yes, then the authors can test whether an abrupt transition of the AMOC is possible with the absence of the external forcing change. It may be important to test the small changes of the external forcings and whether this small changes can bring the climate state to a critical point in which even smaller changes in freshwater forcing is capable to collapse the AMOC.

> We agree with the reviewer that given the experiments performed, we cannot question the multi-equilibria of the AMOC; rather, the experiments suggest that our simulated MIS3 climate stays far away from the bifurcation/tipping point, and is in contrast to some previous studies that show 'sweet spot' within a certain range of external forcing, therefore addressing model dependence in studying model bi-stability.

We also agree that a more thorough and systematic design of model experiments are needed in exploring the multi-equilibria and hysteresis behaviour of the model, forced with a wider range of external forcing. Once we are close to the 'threshold', a small change of forcing is expected to be able to tip the system from one state to another - or even with self-sustained climate transitions. We realise that such studies are certainly meaningful, but are beyond the scope of the current study, and are worth further investigations in the future.

## Response to Reviewer #4:

We respond to the referee's comments in blue font below.

The manuscript “Equilibrium simulations of Marine Isotope Stage 3 Climate” by Guo and colleagues well present a new MIS3 simulation in their model NorESM1-F. They employed a latest (new) ice sheet configuration without a large Fennoscandian ice sheet to conduct their MIS3-38ka simulation. By several attempts exploring the tipping point/bifurcation in their 38ka simulation, the authors claim that their 38ka simulation in NorESM1-F is too stable to reach a tipping point that was commonly used to explain the millennial-scale variability during the MIS3 in previous modeling studies.

We thank the reviewer for his/her constructive comments on our manuscript. We respond to the reviewer's comments below point by point.

The authors first described their 38ka simulation results in very detail (although it can be more compact), in accompany with a comparison with previous LGM simulations. The LGM is a good reference (LGM) to compare with, but there is lack of detailed discussion of their differences. The 38ka simulation was integrated for 2500 model years. The author argue that their simulation is almost in a quasi-equilibrium since the salinity trend in the Atlantic is less than 0.06 g/kg. How is it defined because it remains possible that the deep ocean is not in a quasi-equilibrium state if there exists a robust salinity increase in the AABW formation region. This is also my concern for their sensitivity simulation with lower CO<sub>2</sub> levels. It is very likely that polar regions need a much longer time scale (> 1000 model years) to cool down, producing the cold enough bottom/deep water masses and so the glacial ocean structure and circulation.

> Following the reviewer (and another reviewer)'s comment on the detailed description of the simulation results, we have significantly

reduced the length of the manuscript by moving some texts/figures/sections to the supplementary material; these include the time series of sea ice, several atmospheric diagnosis, the whole section of "Modes of variability", and part of the "stadial" experiment. The updated paper structure reads more compact than the previous version.

Regarding the reviewer's comment on comparison with previous LGM simulations, we actually tried to avoid a detailed discussion, as the two periods are distinct with each other in certain important aspects, e.g. MIS3 features reduced ice volume and a relatively warm climate compared to the LGM. In the updated manuscript, we've added more discussions with the existing MIS3 studies which were not adequately addressed in the previous version.

As for the length of model simulation, we are aware that the deep ocean, in particular the deep Pacific, cannot be fully ventilated. However, the ideal age in the deep Pacific by the end of model integration (2500 years) is less than 1600 years, indicating a relatively well ventilated ocean. There is not any 'hard' threshold to define the model state of quasi-equilibrium in the paleoclimate simulations. The only proposed threshold for the glacial simulations was proposed by Zhang et al. (2013), e.g. a salinity trend  $< 0.006$  g/kg per century in the deep Atlantic can be considered as quasi-equilibrium. As for the lower CO<sub>2</sub> experiments, one integrated for 800 years and the other 1200 years, we agree that both runs can be extended. The experiments, with their current length of integration, do suggest that our simulated MIS3 climate stays far away from the bifurcation/tipping point, and is in contrast with previous studies that show 'sweet spot' within a certain range of external forcing, therefore addressing model dependence in studying model bi-stability.

The authors also assess the simulated climate mode variability in their simulations. It makes the manuscript more comprehensive. However, I do not find a clear connection to the following investigation of the AMOC bistability? why does the "stadial" condition under 40ka boundary not include freshwater forcing in the North Atlantic? I fully understand the

authors' purpose, but since H4 did feature a robust freshwater input, it would be more comparable and reasonable to force a stadial climate with the North Atlantic freshwater forcing under 40ka-38ka boundary conditions. If the authors would like to investigate the changes in climate variability in stadial conditions, I would suggest conducting the hosing under 38ka boundary condition. This will largely reduce the difficulty in the discussion of differences between interstadial and stadial runs. The present "stadial" climate can be included in the section regarding exploration of AMOC bistability.

> Following the reviewer' comment, we have moved the section on climate variability and the majority of the "stadial" experiment to the supplementary material, and the rest of the discussion on the "stadial" experiment to the AMOC bi-stability section.

We set up the "stadial" experiment to test if a non-Henrich stadial can be simulated with NorESM, as motivated by some previous studies (e.g. Zhang et al. 2017; Klockmann et al. 2018) that show a transition of AMOC mode and therefore a stadial-like climate upon relatively small change of CO<sub>2</sub> level. As the reviewer pointed out, freshwater flux is indeed present at H4, though with large uncertainties in magnitude and location, and should be applied for a 'realistic' stadial climate simulation. This is our ongoing work that we intend to write up in a follow up study; we use the hosing experiments with 38 ka BP boundary conditions to study the transition process from H4 to GI8.

In the discussion part, the author design several sensitivity experiments to explore the potential nonlinear behavior of their 38ka climate. The experiments are reasonable and clear, which can provide end-members of climate responses to glacial-interglacial variations regarding ice volume and pCO<sub>2</sub>. It is a promising try here although probably these runs (especially lower CO<sub>2</sub> runs) are not in quasi-equilibrium, therefore the runs are not as conclusive as the authors argued. In addition, one conceptual mistake in the manuscript is that spontaneous oscillation does not share the same definition with the AMOC bistability, but rather a hopf bifurcation feature. The authors shall go through the manuscript carefully to distinguish their difference.

Overall I find the manuscript is interesting and well within the scope of Clim. Past. It can be accepted for publication after some modifications in the structure as well as refinements of model results description and discussion.

> We thank the reviewer for the positive comments on our manuscript. As we responded above, we agree that extending the sensitivity experiments, together with a wider forcing range, could be more beneficial to the simulation results and conclusions.

We went through the manuscript to double check the conceptual issue pointed out by the reviewer (e.g. see the second and third paragraphs of Section 4.3).

Finally, following all the reviewers' comment, we've made efforts to restructure the manuscript and present the results in a more succinct way.

# Equilibrium simulations of Marine Isotope Stage 3 climate

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## Abstract.

An equilibrium simulation of Marine Isotope Stage 3 (MIS3) climate with boundary conditions characteristic of Greenland Interstadial 8 (GI-8; 38 ka BP) is carried out with the Norwegian Earth System Model (NorESM). A computationally efficient configuration of the model enables long integrations at relatively high resolution, with the simulations reaching a quasi-equilibrium state after 2500 years. We assess the characteristics of the simulated large-scale atmosphere and ocean circulation, precipitation, ocean hydrography, sea ice distribution, and internal variability. The simulated MIS3 interstadial near surface air temperature is 2.9°C cooler than the pre-industrial (PI). The Atlantic Meridional Overturning Circulation (AMOC) is deeper and intensified (by ~ 13%). There is a decrease in the volume of Antarctic Bottom Water (AABW) reaching the Atlantic. ~~However~~ At the same time, there is an increase in ventilation of the Southern Ocean, associated with a significant expansion of Antarctic sea ice and concomitant intensified brine rejection, invigorating ocean convection. In the central Arctic, sea ice is ~ 2 m thicker, with an expansion of sea ice in the Nordic Seas during winter. ~~Simulated MIS3 inter-annual variability of the El Niño-Southern Oscillation (ENSO) and the Arctic Oscillation are weaker compared to the pre-industrial.~~ Attempts at triggering a non-linear transition to a cold stadial climate state, by varying atmospheric CO<sub>2</sub> concentrations and Laurentide Ice Sheet height, suggest that the simulated MIS3 interstadial state in the NorESM is relatively stable, thus questioning ~~the potential for underscoring the role of model dependency, and questioning the existence of~~ unforced abrupt transitions in Greenland climate ~~during the last glacial in the absence of interactive ice sheet-meltwater dynamics.~~

## 1 Introduction

Marine Isotope Stage 3 (MIS3), a period about 60 – 30 ka BP (thousand years before present) during the last glacial, was characterised by millennial-scale abrupt climate transitions. These events are known as Dansgaard-Oeschger (D-O) events, as revealed by the Greenland oxygen isotope ice core records (Dansgaard et al., 1993). A D-O event consists of an abrupt transition from a cold stadial climate state to a relatively warm interstadial climate state, followed by a gradual return to cold stadial conditions (Huber et al., 2006). Correlated with the rapid ~~change of Greenland temperature, warming of~~ warming on Greenland (up to 15 °C within a few decades during the stadial-to-interstadial transition), the North Atlantic and Nordic Seas are subject to abrupt climate transitions as interpreted from a number of marine sediment cores (~~Bond et al., 1997; Rasmussen and Thomsen, 2004; Dokken et al.~~ Bond et al., 1997; Rasmussen and Thomsen, 2004; Dokken et al., 2013; Sadatzki et al., 2019). Towards the end of every few



stadial periods, the marine sediments show evidence of massive calving of the Laurentide Ice Sheet, with large numbers of icebergs transversing and melting in the North Atlantic. These events are known as Heinrich events (Heinrich, 1988). The freshwater from these melting icebergs is thought to have weakened the Atlantic Meridional Overturning Circulation (AMOC), possibly causing further cooling of the Northern Hemisphere (e.g. Broecker, 1994).

5 While there has been significant advances in our understanding of the dynamics behind D-O events in recent years, the key mechanisms triggering these abrupt climate transitions ~~are still under debate and~~ remain illusive. A leading hypothesis is related to a switch between strong, weak, and off modes of the AMOC (Rahmstorf, 2002; Böhm et al., 2015; Henry et al., 2016). This has the potential to significantly alter the circulation and northward heat transport in the Atlantic Ocean. Model based studies have shown that changes in the mode of the AMOC can be triggered by, e.g., freshwater input from melting ice sheets  
10 (Ganopolski and Rahmstorf, 2001), variations in the size of the Laurentide Ice Sheet (Zhang et al., 2014a), as well as changes in atmospheric CO<sub>2</sub> (Klockmann et al., 2018). Another theory for explaining the abrupt warming of Greenland invokes atmospheric circulation changes triggered by transitions in sea ice cover: e.g., Li et al. (2005, 2010) showed that shifts in Greenland precipitation and temperature are consistent with the climate response induced by sea ice growth and retreat, in particular over the Nordic Seas. The sea ice acts as a lid, insulating the ocean from the atmosphere and reducing the amount of heat released.  
15 Proxy data from sediment cores in the Nordic Seas (Rasmussen and Thomsen, 2004; Dokken et al., 2013; Ezat et al., 2014) suggest that the warm Atlantic inflow can be separated from the sea surface by a halocline, and slowly accumulate heat in the subsurface and intermediate/deep waters during MIS3 stadials. Eventually the warming below the halocline destabilises the water column, and brings warm Atlantic water to the surface, tipping the Nordic Seas into an ice-free state that can lead to a rapid warming as seen in the Greenland ice cores (~~e.g., Dokken et al., 2013~~)(e.g., Dokken et al., 2013; Sadatzki et al., 2019).

20 Although prescribed external forcing are often introduced into the model to trigger D-O events (e.g., Ganopolski and Rahmstorf, 2001; Menviel et al., 2014), several model simulations in recent years have been reported to be able to spontaneously reproduce rapid cold-to-warm transitions that resemble D-O events. The underlying mechanisms for these self-sustained oscillations in the models involve, for instance, the "kicked" salt oscillator acting in the Atlantic (Peltier and Vettoretti, 2014) and a similar type of "thermohaline oscillator" (Brown and Galbraith, 2016), stochastic atmospheric forcing (Kleppin et al., 2015),  
25 and the formation of North Atlantic super polynyas and subsequent heat release (Vettoretti and Peltier, 2016).

Most studies of D-O events apply a coupled model configured with last glacial maximum (LGM), or pre-industrial (PI) boundary conditions. Very few model studies apply MIS3 boundary conditions. These MIS3 studies include models with varying resolution and complexities, including an atmospheric general circulation model forced with fixed SSTs (Barron and Pollard, 2002), a coupled model of intermediate complexity (Van Meerbeek et al., 2009), ~~and a general circulation models~~  
30 with relatively coarse resolution (Merkel et al., 2010; Zhang et al., 2014b), and higher-resolution atmospheric and oceanic general circulation model (Brandefelt et al., 2011)general circulation models (Brandefelt et al., 2011; Kawamura et al., 2017). Van Meerbeek et al. (2009) include a stadial and an interstadial simulation and study the model response to changes in orbital forcing and greenhouse gases (GHG), typical of MIS3 conditions; ~~they~~. They conclude that neither orbital, nor GHG forcing can explain the D-O like variability. Rather, a freshwater input to the surface ocean must be invoked to move the system from  
35 an (equilibrium) interstadial state to a (perturbed) stadial state. While a strong AMOC was simulated by Van Meerbeek et al.

(2009), Brandefelt et al. (2011) found an AMOC slowdown of approximately 50% in their MIS3 Greenland Stadial 12 (GS-12; 44 ka BP) configuration. The large reduction in AMOC, and relatively cold North Atlantic SST, is simulated without freshwater forcing. Brandefelt et al. (2011) suggest that the MIS3 equilibrium state is highly ~~model-dependent~~model dependent. Note however, that a dramatic AMOC weakening during non-Heinrich stadials is not supported by geological reconstructions (e.g.,  
5 Böhm et al., 2015).

In this work, we present a MIS3 interstadial equilibrium simulation employing a new version of the Norwegian Earth System Model (NorESM) designed for multi-millennial and ensemble studies (~~?~~(Guo et al., 2019)). The state-of-the-art Earth system model features a 2° horizontal resolution for the atmosphere and land components, and a 1° horizontal resolution for the ocean and sea ice components. A faster model throughput of approximately 90 model years per day can be achieved with  
10 existing hardware, compared to <40 model years per day for the CMIP5 version of NorESM with comparable resolution and hardware. As documented by ~~?~~Guo et al. (2019), this NorESM version shows improved skill compared to the CMIP5 version in simulating AMOC and sea ice thickness/extent, both important quantities when discussing MIS3 climate and the dynamics of D-O events.

We configure the model with boundary conditions characteristic of 38 ka BP, immediately following the onset of Greenland  
15 Interstadial 8 (GI-8). Spontaneous occurrence of D-O like abrupt climate transitions are not simulated in the model during the 2500 year integration using MIS3 boundary conditions. Instead, the experiment serves as a baseline simulation for evaluating equilibrium interstadial climate states during MIS3. A satisfactory quasi-equilibrium state is reached before the end of the long integration, and we assess the ~~basie~~-interstadial climate state of the atmosphere, ocean, and sea ice as represented by the model.

~~With a realistic model configuration using~~The significance of this study is that the MIS3 baseline simulation is configured  
20 with realistic MIS3 boundary conditions, ~~a long equilibrated integration, and a model with and integrated for multi millennia~~  
with a climate model featuring relatively high resolution and ~~encouraging physical performance~~reasonable performance in  
simulating AMOC and sea ice. This experiment is then used as a baseline for glacial sensitivity studies. Given the small  
number of existing MIS3 model studies, we aim to improve our understanding of the MIS3 climate~~and provide a baseline~~  
~~for further sensitivity studies exploring the dynamics,~~ especially the baseline climate and the sensitivity of D-O ~~events~~event  
25 related variability to external forcing within a MIS3 configuration. The structure of the paper is arranged as follows: in Section  
2, we give a brief overview of the NorESM, including details of the ~~new~~-version used in this study, followed by a description of  
the MIS3 experimental configuration; in Section 3, we assess the equilibrium state of the MIS3 long integration with NorESM,  
followed by details of the simulated mean MIS3 interstadial state of the atmosphere, ocean, ~~sea ice, and internal variability of~~  
~~the model. A parallel experiment with typical boundary conditions of Greenland stadials is presented in Section 4, followed by~~  
30 ~~a discussion and sea ice. Discussions~~ on the simulated AMOC and model response to changes in GHG and ice sheet height ~~in~~  
~~Section 5~~are presented in Section 4. The main conclusions are summarized in Section ~~6~~5.

## 2 Methods

### 2.1 The climate model: NorESM

The ~~version of NorESM used in this study~~ [NorESM family](#) is based on the Community Climate System Model version 4 (CCSM4; Gent et al., 2011) but differs from the latter in several aspects (Bentsen et al., 2013): NorESM uses an isopycnic coordinate ocean model that originates from the Miami Isopycnic Coordinate Ocean Model (MICOM; Bleck and Smith, 1990; Bleck et al., 1992). The atmospheric component Community Atmosphere Model 4 (CAM4) has the option to use a modified chemistry-aerosol-cloud-radiation schemes developed for the Oslo version of CAM (e.g. CAM4-Oslo). The HAMBURG Ocean Carbon Cycle (~~HAMOCC~~) ([HAMOCC; Maier-Reimer, 1993; Maier-Reimer et al., 2005](#)) model is adapted to the isopycnic model framework of NorESM and incorporated as the ocean biogeochemistry component of the model.

10 The basic evaluation and validation of the Climate Model Intercomparison Project Phase 5 (CMIP5) version of NorESM (NorESM1-M) is documented by Bentsen et al. (2013). Here we use a recently developed, computationally efficient variant of NorESM1-M: NorESM1-F (~~?~~) ([Guo et al., 2019](#)), designed for multi-millennial and ensemble simulations while retaining the resolution (2° atmosphere/land, 1° ocean/sea ice) and overall quality of NorESM1-M.

15 Compared to NorESM1-M, the model complexity of NorESM1-F is reduced by replacing CAM4-Oslo, that uses emissions of aerosols and explicitly simulates their life cycles, with the standard CAM4 that uses prescribed aerosol concentrations. The coupling frequency between atmosphere-sea ice and atmosphere-land is reduced from half-hourly to hourly, allowing the use of an hourly base time step for the sea ice and land components matching the radiative time step of the model; the dynamic sub-cycling of the sea ice is reduced from 120 to 80 sub-cycles. The last two changes result in a model speed-up of ~30% with a relatively small effect on the modeled climate. In addition, recent code developments in the ocean, atmosphere, and  
20 biogeochemistry components ~~since NorESM1-M~~ have been implemented as documented in detail by ~~?~~ [Guo et al. \(2019\)](#). Of these code developments in NorESM1-F, the updated ocean physics ~~in NorESM1-F~~ lead to improvements over NorESM1-M in simulating the strength of the AMOC and the distribution of sea ice, both are important metrics in simulating past and future climates.

### 2.2 Experimental setup

25 In the MIS3 setup (Table 1), the solar constant is kept fixed at the pre-industrial value (1360.9 W m<sup>-2</sup>), and the orbital parameters are set to values corresponding to 38 ka BP (Berger, 1978). In the Northern Hemisphere (NH), the chosen MIS3 time-slice shows enhanced insolation in spring (April-May-June) relative to present (Fig. 1), followed by reduced summer insolation (July-August-September). In the Southern Hemisphere (SH), changes in insolation are less pronounced, with stronger fall (August-September-October) insolation and weaker winter (November-December-January) insolation.

30 Concentrations of greenhouse gases (GHG) are set according to ice core measurements of typical interstadial conditions following Schilt et al. (2010): CO<sub>2</sub>, CH<sub>4</sub>, N<sub>2</sub>O concentrations are specified as 215 ppm, 550 ppb, and 260 ppb, respectively, which are identical to the MIS3 setup by Van Meerbeek et al. (2009). Chlorofluorocarbon (CFCs) levels are set to zero.

The ocean bathymetry is adapted based on an estimated sea level lowering of 70 m below present day (Waelbroeck et al., 2002). As a consequence, many shallow ocean grid points on the shelf turn into land, thereby modifying the land-sea mask (Fig. 2). Most of the modifications occur in the northern high latitudes, e.g., the East Siberian Shelf, Laptev Shelf, and the Bering Strait (which is closed).

5 The configuration of global ice sheet extent and height (Fig. 2) is derived from a data-constrained ice sheet model for 38 ka BP consisting of the Antarctic (Briggs et al., 2014), Greenland (Tarasov and Richard Peltier, 2002), North American (Tarasov et al., 2012), and Eurasian (Lev Tarasov, personal communication) ice sheets. NorESM1-F does not ~~have a dedicated~~ include a dynamic land ice component, and ~~an the~~ assumed ice sheet ~~state during extent and elevation for MIS3 compared to present day is taken into account by modifying the static land topography~~ are fixed. The altitude of the land surface is kept at pre-  
10 industrial values outside the ice sheet areas. In the areas covered by land ice, the maximum value of the original pre-industrial topography and MIS3 reconstructed ice sheet topography is used; this procedure prevents jumps to high topography adjacent to the ice sheet margin. The resulting MIS3 Laurentide Ice Sheet reaches altitudes higher than 2500 m, but is significantly smaller, both in terms of ice extent and height, when compared to the LGM ice sheets such as ICE-6G (Peltier et al., 2015). The southeastern margin of the Laurentide Ice Sheet is further north and the surface is 300-900 m lower than during LGM.  
15 Similarly, the Eurasian Ice Sheet is smaller, but there is a significant amount of land ice over Fennoscandia.

For the configuration of ice-land-sea mask in the Barents Sea, care needs to be taken as this region is an important pathway allowing warm and saline Atlantic Water traveling north, therefore opening or closing it has significant consequences for Arctic ocean circulation and climate (Smedsrud et al., 2013). ~~There Unfortunately, there~~ is a lack of reliable geological evidence for the existence of land ice in the Barents Sea before the LGM. However, sparse evidence from the Barents Sea-Svalbard region  
20 suggest there was little or no land ice here during MIS3 (MANGERUD et al., 1998; Mangerud et al., 2008; Ólafur Ingólfsson and Landvik, 2013). Therefore, the Barents Sea is kept open-free of land ice with a reduced water depth (by 70 m) in the ~~MIS3~~ MIS3 configuration of the model. The Canadian Archipelago is covered by land ice, blocking the passage of water between Baffin Bay and the Arctic. In Antarctica, the Ross and Weddell Seas are covered by grounded ice rather than floating ice shelves as today.

25 The land surface vegetation type in the MIS3 configuration is set equal to the pre-industrial values, and the extra land points caused by sea level lowering are assigned as tundra (20% grass + 80% bare ground).

With the adjusted MIS3 land-sea mask and surface topography, a new river routing map is produced ~~-(Supplementary Fig. S1)~~. For the ice free land surfaces, the river routing corresponds to the PI simulation. Where there is new land, due to the lower MIS3 sea level, the river mouths-outlets are extended to the ocean. For the ice covered areas, a new map is generated based on  
30 the land ice topography, routing the water from the land ice along the steepest gradient, either directly to the ocean, or to the nearest river if the ice margin terminates on land.

Salt equivalent to  $0.6 \text{ g kg}^{-1}$  is uniformly added to the ocean, to account for the large amount of freshwater stored as ice on land. The vertical coordinate in MICOM is potential density with reference pressure of 2000 dbar, and is adapted from a present day range of  $28.202\text{-}37.800 \text{ kg m}^{-3}$  to  $28.672\text{-}38.270 \text{ kg m}^{-3}$  below the mixed layer, in order to account for the change in  
35 salinity and thus density. As for the PI experiment, the ocean model is initialised with modern temperature and salinity (Steele

et al., 2001) with the above mentioned salinity increment applied. Note that no specific protocols on ocean initialisation are defined for glacial simulations, e.g., within the Paleoclimate Modelling Intercomparison Project (PMIP), ~~and groups choose~~ the groups chose to initialise their models either from present day conditions, or from a glacial ocean state. As illustrated by Zhang et al. (2013), equilibration time for the deep ocean can be significantly reduced (50% in their model) if a glacial, rather than present day ocean state is utilised for initialisation. Note, however, that MIS3 is not a full glacial state, and has a climate in between LGM and PI. Model equilibration for the NorESM MIS3 experiment will be addressed and discussed in Section 3.1.

### 3 Results

#### 3.1 Model spin-up

The MIS3 experiment was run for 2500 years and the PI experiment for 2000 years. When comparing the two ~~simulations,~~ the model years between 1800 and 2000 are averaged.

### 4 Results

#### 3.1 Model spin-up

Both the PI and MIS3 experiments reach a quasi-equilibrium climate state after the multi-millennial integration, as indicated by the time series shown in Fig. 3. The differences between the global mean MIS3 and PI climate are summarised in Table 2.

In the following results, the statistical significance of the calculated trends are tested using the Student's *t*-test with the number of degrees of freedom, accounting for autocorrelation, calculated following Bretherton et al. (1999). Trends with *p* values < 0.05 are considered to be statistically significant.

The MIS3 experiment exhibits a small negative TOA radiation balance ( $-0.16 \text{ W m}^{-2}$  averaged between model years 1801-2000 and  $-0.08 \text{ W m}^{-2}$  between 2301-2500; Fig. 3a). This results in a negative ocean heat flux at the surface, and cooling of the global ocean (Fig. 3d). The cooling trend is  $-0.05 \text{ }^\circ\text{C}$  per century over the model years 1801-2000, and decreases to  $-0.02 \text{ }^\circ\text{C}$  per century over the model years 2301-2500, both are statistically significant. Averaged between 1800 and 2000 model years, the global mean MIS3 ocean temperature is  $1.7 \text{ }^\circ\text{C}$  colder than that of the PI experiment.

At the ocean surface, SSS in the MIS3 experiment exhibits negligible drift over the model years 1801-2000 (Fig. 3b,c), whereas SST shows a small statistically significant cooling trend of  $0.04 \text{ }^\circ\text{C}$  per century. For MIS3, the global mean SST and SSS are  $1.2 \text{ }^\circ\text{C}$  colder and  $0.3 \text{ g kg}^{-1}$  saltier, respectively, compared to the PI experiment. Simulated time evolution of sea ice extent in the Northern and Southern Hemispheres show a small drift towards the end of the evolution (Supplementary Section 1).

While the simulated MIS3 surface properties reach a quasi-equilibrium state, the ocean interior experiences a multi millennial cooling trend (Fig. 3d). The slow adjustment of the deep ocean is considered to be important for the evolution of ocean

stratification and overturning circulation (Zhang et al., 2013; Marzocchi and Jansen, 2017). As a consequence, care should be taken when evaluating surface climatology, and deep ocean equilibration should be assessed. For the NorESM-MIS3 simulation, we deem the aforementioned global mean ocean cooling trend to be small. For the deep ocean, the salinity trend in the Atlantic is found to be smaller than  $0.006 \text{ g kg}^{-1}$  per century during the model years 1800-2000 (not shown), which is the threshold proposed by Zhang et al. (2013) for an ocean state of quasi-equilibrium.

Both experiments exhibit an increasing AMOC at the beginning of the model integration, followed by a gradual equilibration to a weaker state (Fig. 3e). The simulated pre-industrial AMOC at  $26.5^\circ\text{N}$  is 20.9 Sv, which is close to present day observations ( $\sim 18$  Sv; RAPID data from [www.rapid.ac.uk/rapidmoc](http://www.rapid.ac.uk/rapidmoc)). For the first few hundred years of the MIS3 integration, the AMOC is significantly stronger than in the PI (by up to 10 Sv), but the difference between the two experiments is gradually reduced as the integration continues. Averaged between model years 1801-2000, the MIS3 simulated AMOC is 22.8 Sv, only 1.9 Sv greater than in the PI experiment.

~~Simulated time evolution of sea ice area, averaged over the northern and southern hemispheres, is shown in Fig. ???. Both experiments show a small drift at the end of the integration. In the NH, the MIS3 experiment shows a larger minimum sea ice area in September, and a smaller maximum sea ice area in March, compared to the PI experiment. The simulated smaller NH March sea ice area at MIS3, is caused by the large region in the peripheral areas of the Arctic Ocean being defined as land in MIS3, when sea level is lowered by 70 m (see Fig. 2). In the PI experiment, these areas are defined as ocean, which are ice covered during winter, resulting in a larger area of sea ice cover, even if the climate is warmer than at MIS3. As will be shown later, sea ice in MIS3 has a larger extent in the NH when excluding the above offset due to changes in the land-sea mask. In the SH, MIS3 features larger sea ice area compared to PI in both seasons, even though the above land-sea effect is at play (mostly in the Weddell Sea and the Ross Sea). In austral winter (September), sea ice area continues to increase throughout the multi-millennial model integration. However, the trend is diminished towards the end of the simulation. In austral summer (March), MIS3 sea ice area is comparable to PI and slowly increases until a sudden jump is detected close to model year 1600. After model year 1800 the simulated sea ice shows little drift in both experiments. The rapid increase in sea ice in the MIS3 experiment mainly arises from changes in the Weddell and Ross Sea areas of the model.~~

The weakening of the AMOC, after the initial overshoot, occurs concurrently with a shoaling of North Atlantic Deep Water (NADW) and more intrusion of the Antarctic Bottom Water (AABW) as a manifestation of an adjustment of the deep ocean. Previous studies related ~~relate~~ this behaviour to an expansion of Antarctic sea ice in a colder climate (Ferrari et al., 2014; Jansen, 2017): e.g., as Antarctic sea ice grows (Fig. ??? [Supplementary Fig. S2](#)), more brine is rejected, leading to more open ocean convection, favoring the formation of AABW. These processes will be further discussed in Section 4.1.

### 3.1 Simulated MIS3 climate

#### Atmospheric surface temperature, circulation, and precipitation

Simulated annual mean surface air temperature change with respect to PI is shown in Fig. 4. Significant cooling occurs at high latitudes in both hemispheres. This is particularly clear above the MIS3 Laurentide Ice Sheet, where cooling reaches  $25^\circ\text{C}$ , as

well as in the Barents Sea and parts of the Nordic Seas, where sea ice expands (c.f. Fig. 9). The MIS3 experiment also exhibits noticeable cooling over the Kuroshio extension.

5 Simulated global mean near surface air temperature during MIS3 is 2.9 °C cooler than the PI (Table 2). For comparison, the ~~cooling is~~ CCSM3 MIS3 simulation (with 35 ka boundary conditions) by Merkel et al. (2010) showed a cooling of 3.4 °C, whereas using a different version of CCSM3 configured with 38 ka boundary conditions, Zhang et al. (2014b) reported a cooling of 3.5 °C. A stadial simulation with 44 ka boundary conditions (Brandefelt et al., 2011) shows a much cooler climate, e.g., 5.5 °C compared to the recent past.

10 In contrast to the limited number of MIS3 studies, there is a rich literature on the LGM climate. With both similarities as well as apparent differences with regard to the external forcing and the climate, it can be useful to compare the climate of the two periods: our simulated MIS3 cooling is smaller than the reconstructed LGM global mean cooling of  $4.0 \pm 0.8$  °C (Annan and Hargreaves, 2013) and the PMIP2 (Braconnot et al., 2007) and PMIP3 (five models; Braconnot and Kageyama, 2015) simulated LGM cooling range of 3.6-5.7 °C and 4.4-5 °C, respectively.

15 The elevated surface of the MIS3 Laurentide Ice Sheet modifies the atmospheric stationary waves, rendering an enhanced, meandering wave pattern in the vicinity of the North American continent (Fig. 5); the displayed 500-mb geopotential height in winter shows enhanced troughs in the northwestern Pacific and eastern Canada, and enhanced ridges in western Canada. A stronger, more zonal, and northward-shifted (by  $\sim 4$  degrees) subpolar jet above the North Atlantic is revealed by the 200-mb zonal wind (not shown). At the ocean surface, a deepened Aleutian low and the associated development of cyclonic surface wind is found in winter (not shown). The cyclonic wind anomaly advects warm air to Alaska and contributes to the reduced cooling as seen in Fig. 4. In the Atlantic, a southwestward migration of the Icelandic low and Azores high leads to broader, stronger, and more southerly located westerlies over the North Atlantic (Fig. 6). The ice sheet induced wind anomalies over the North Atlantic are common features also seen in the PMIP3 LGM simulations (Muglia and Schmittner, 2015). The zonal mean NH westerlies (surface zonal wind stress) increase by  $\sim 20\%$  relative to PI in NorESM, and shift equatorward by  $\sim 4$  degrees. Furthermore, ~~the expansion of sea ice at MIS3 (see Fig. 9) induces a significant increase in the~~ surface wind stress is significantly increased just off the edge of the sea ice in the Nordic Seas ~~and (see Fig. 9) and in the~~ Irminger Sea, possibly caused by the expanded Laurentide Ice Sheet (e.g., Sherriff-Tadano et al., 2018). In the Labrador Sea, a strong northerly wind anomaly is also induced by the nearby Laurentide Ice Sheet.

20 In the tropics, the northeasterly trade winds are strengthened in the NH, while in the SH the southeasterly trade winds are relatively unchanged. In the Southern Ocean, the westerlies are strengthened in the Pacific Ocean sector and weakened in the Indian Ocean sector. The zonal mean of the westerly wind stress in the Southern Ocean shows a slight strengthening during MIS3 ( $\sim 4\%$ ), with nearly no latitudinal shift.

25 ~~Annual mean global precipitation decreases by  $0.18 \text{ mm day}^{-1}$  as a consequence of the colder and more arid climate~~ Simulated patterns of precipitation during MIS3. ~~Geographically, during DJF, a significant decrease in precipitation is seen in the North Pacific, the western North Atlantic, the Barents Sea and in the Nordic Seas (Fig. ??). Greater precipitation is seen in the eastern and subtropical Pacific and in the eastern North Atlantic. Precipitation in the western Labrador Sea also increases, due to the reduced sea ice cover (see Fig. 9). In the tropics, precipitation in the African and South American monsoon~~

regions, as well as in the western Pacific warm pool, is remarkably reduced, and a southward shift of the ITCZ occurs. In contrast, during JJA, western and northern Europe as well as the North Pacific features more precipitation; the Indian monsoon region experiences less precipitation, and in the tropics, the ITCZ moves south in the Pacific and north in the Indian Ocean. are presented in Supplementary Section 2.

## 5 Ocean circulation and sea ice

### Ocean circulation and sea surface features

The reduced level of CO<sub>2</sub> during MIS3 is important in lowering SST and in causing the expansion of sea ice. Simulated global mean MIS3 SST is 1.2 °C colder with respect to the PI. The ~~simulated cooling is comparable to the compiled MARGO reconstruction that estimates a SST cooling of 1.9±1.8 °C for the LGM (MARGO Project Members, 2009).~~ The geographical distribution of the cooling (Fig. 7a) reflects the change in surface air temperature as shown in Fig. 4. The cooling is relatively modest (1-2 °C) in the tropical and subtropical oceans, and increases towards higher latitudes, in particular in the North Pacific, the Barents Sea, the Nordic Sea, and the Southern Ocean. In contrast, the central North Atlantic exhibits less cooling, and even exhibits warming near the center of the subpolar gyre and at the western rim of the Labrador Sea. While the "warm blob" in the subpolar gyre can be attributed to a shift of the North Atlantic Current at MIS3 (as suggested from the surface velocity and sea level fields; not shown), the relatively weak cooling in the NA subpolar region is likely caused by a stronger AMOC (Fig. 8) and a stronger and slightly more contracted subpolar gyre (Fig. ~~??S4~~) bringing more ocean heat from the tropics to this region (Fig. ~~??b~~) (e.g., the S5) (Atlantic ocean heat transport increases by 15% at 40°N). The overall pattern of weak cooling in the North Atlantic during MIS3 can be compared to the recent twentieth century global warming, with a North Atlantic cooling anomaly, suggested to be caused by a reduction in the AMOC (Rahmstorf et al., 2015). The simulated warming along the western rim of the Labrador Sea (1-2 °C), apart from the contribution from a strengthened subpolar gyre (21% stronger compared to PI), is ~~more directly~~ related to the locally reduced MIS3 sea ice cover (Fig. 9a).

The cooling in the North Pacific is associated with a reduction of northward ocean heat transport in this region (Fig. ~~??bS5~~), e.g., ocean heat transport is 23 % smaller at 30°N during MIS3. In addition, the cooling is accompanied by expanded winter sea ice cover in the northwestern Pacific (Fig. 9a) and a southward shift of the North Pacific subpolar gyre (Fig. ~~??S4~~).

As northward ocean heat transport in the North Pacific decreases, southward ocean heat transport in the South Pacific and Indian Ocean increases (e.g., 13 % increase at 30°S). Further to the south, the MIS3 simulation shows enhanced meridional ocean heat transport across the Antarctic Circumpolar Current (Fig. ~~??bS5~~). However, significant cooling is simulated in this region, and is associated with an equatorward expansion of sea ice (particularly in the western Indian Ocean sector; Fig. 9), concomitant with a northward shift of the Antarctic Circumpolar Current (not shown).

For the surface salinity (Fig. 7b), MIS3 is more saline as a result of the addition of 0.6 g kg<sup>-1</sup> salt into the global ocean due to expansion of ice on land. Exceptions are the North Pacific subpolar area, South Pacific subtropical area, South China Sea, eastern Atlantic, and off the Eurasian shelf. Strong salinity increases are found in Baffin Bay, the western Labrador Sea, the North American sector of the Arctic, and within and east of the Weddell Sea. The salinity decrease in the North Pacific and



salinity increase in the Arctic and in Baffin Bay, can be partially attributed to the closure of the Bering Strait (~~Hu et al., 2010~~) (Hu et al., 2010, 2015) at MIS3. The Bering Strait acts as a passage of freshwater from the Pacific to the Arctic Ocean, with an impact on salinity in Baffin Bay and the North Atlantic via the Canadian Arctic Archipelago. The Bering Strait, with a depth of  $\sim 50$  m at PI (and at present), is closed due to the lower sea level at MIS3, while the passage through the Canadian Arctic Archipelago is blocked by the presence of expanded MIS3 ice sheets (Fig. 2). A net volume transport of 1.3 Sv low salinity water, through the Bering Strait in the PI control run (present day observations are  $\sim 0.8$  Sv; Woodgate et al., 2005) is thus removed in the MIS3 run and contributes to the pattern described above.

Dramatic freshening takes place in the Eurasian sector of the Arctic at MIS3. Here, the Arctic river ~~mouths-outlets~~ extend further out to the open ocean, relative to the present day locations, due to the change of land-sea mask (Fig. 2). Such changes in the locations of river ~~mouths-outlets~~, and thus freshwater input, lead to a decrease in salinity, including the region southwest of Norway, where a large glacial river is generated due to the presence of the nearby Fennoscandian Ice Sheet ~~-(see Fig. S1)~~. In addition, the volume transport of saline water, of North Atlantic origin, into the Arctic via the Barents Sea Opening (from the southern tip of Svalbard to the northern tip of Norway), is reduced from 2.8 Sv at PI to 0.5 Sv at MIS3, contributing to the salinity decrease in the Eurasian sector of the Arctic.

The fresh surface water in the South China Sea ~~is due to increased runoff during MIS3, whereas in the~~ MIS3 is caused by increased runoff from the new river routing in this region owing to the change of land-sea mask. In the Southwest Pacific, surface freshening is due to a southward shift of the ITCZ and an overall decrease of evaporation minus precipitation in the region. Off the coast of Antarctica, enhanced formation of sea ice (Fig. 9), and the associated brine release, leads to an increase in surface salinity; the effect is especially pronounced in the Weddell Sea region.

## 20 AMOC and Atlantic hydrography

~~The AMOC redistributes heat and freshwater and plays a crucial role in the global climate.~~ Our NorESM experiments show a strengthened AMOC at MIS3 (27.5 Sv) relative to the PI (24.3 Sv) (Fig. 8). The depth of the AMOC maximum for MIS3 is unchanged and located at 800 m. The vertical extent of the AMOC is deepened from 3500 m in the PI to 4200 m at  $26^\circ$  N (~~present day RAPID observations reveal a depth of  $\sim 4300$  m; Smeed et al., 2016~~). ~~The deeper overturning stream function (present day RAPID observations give a depth of  $\sim 4300$  m; Smeed et al., 2016). The deep overturning cell associated with AABW is contracted and weakened. For comparison, LGM simulations contributing to PMIP2 feature either deeper/stronger or shallower/weaker AMOC (Weber et al., 2007), whereas most PMIP3 LGM simulations exhibit a deeper/stronger AMOC relative to present day (Muglia and Schmittner, 2015) Zhang et al. (2014b) reported a similar strengthening of AMOC during MIS3 (38 ka boundary conditions), e.g., 15.4 Sv which is 1.5 Sv stronger than their PI control simulation, but is much weaker than our simulated strength of AMOC at MIS3. In contrast to the NorESM simulation, Zhang et al. (2014b) also simulated a shoaling of the AMOC.~~

Together with the changes to the AMOC, the deep Atlantic ocean exhibits changes in the distribution of water masses. The zonal mean Atlantic (including the Nordic Seas and the Atlantic sector of the Southern Ocean ~~and the Nordic Seas~~) temperature (Fig. 10a,b) shows that cooling occurs nearly over the entire water column in the Atlantic basin, with the strongest cooling

detected near the bottom of the thermocline ( $>4$  °C) and in the deep ocean below 3500 m ( $1.5-3$  °C). The larger temperature decrease in the deep North Atlantic compared to the South Atlantic suggests a more homogeneous deep water distribution in the Atlantic basin with a smaller inter-hemispheric gradient during MIS3 (Fig. 10b).

With more vigorous deep water formation in the NH (associated with a stronger AMOC) and ~~also~~ in the SH (as discussed later) during MIS3, ~~a general~~ enhanced upward motion of sea water is expected away from ~~the~~ sinking regions (Munk, 1966); ~~this~~. This leads to a thermocline that is displaced upwards with a sharper vertical gradient, contributing to the cold anomaly near the base of the thermocline ~~as~~ seen in Fig. 10b. Similar upward displacement of the thermocline and an associated cold anomaly are also seen in the Pacific Ocean (not shown). We further note that the ~~especially~~ cold anomaly centred around  $30^{\circ}\text{N}$ , 500-800 m depth in the Atlantic (Fig. 10b) is primarily caused by ~~the~~ reduced warm Mediterranean outflow ~~therein~~ during MIS3 (not shown).

The Atlantic zonal mean salinity anomaly (Fig. 10c,d) shows an overall increase in salinity, except near the bottom of the pycnocline/thermocline, where the waters are subject to enhanced upwelling as discussed above. Greater freshening is also observed in the saline Mediterranean outflow region where it is reduced during MIS3. For the deeper layers, there is a north-south asymmetry, with more saline bottom water in the South Atlantic ( $\sim 0.4$  g  $\text{kg}^{-1}$ ) compared to the deep North Atlantic. Geological reconstructions of the glacial deep Atlantic hydrography from pore water measurements ~~revealed~~ suggest a reversed north-south salinity gradient at the LGM (Adkins et al., 2002), e.g., with AABW being more saline than NADW. In our MIS3 simulation, while the gradient is effectively lowered, the reduction is not sufficient to cause a salinity reversal.

~~The~~ In the cold deep ocean, the salinity effect dominates the change of density ~~in the cold deep ocean~~, manifested by a larger increase of potential density in the Atlantic sector of the Southern Ocean ( $0.6-0.8$   $\text{kg m}^{-3}$ ) ~~than compared to~~ in the Atlantic ( $0.5-0.6$   $\text{kg m}^{-3}$ ) (not shown). A comparison of ~~ideal age distribution shows that ideal age in the South Atlantic and the Atlantic sector of the Southern Ocean is reduced by 200-400~~ the ideal age of the water mass (defined as the time since the water mass last made contact with the surface) shows that the Southern Ocean zonal mean ideal age during MIS3 is relatively homogeneous in the vertical, and is younger by  $\sim 300$  years compared to that in the PI experiment (not shown). In the Southern Ocean, deep water masses are  $\sim 400$  years younger, and the entire water column shows a nearly homogeneous young water mass ( $<10$  years Fig. S7), indicating ~~much~~ enhanced ventilation of AABW. Kobayashi et al. (2015) reported a similar response of simulated water mass age for the LGM in the Southern Ocean owing to enhanced open ocean convections.

## Sea ice

As documented by ~~?~~ Guo et al. (2019), the PI simulation of sea ice agrees well with observations, both in terms of ~~both~~ thickness and extent. The simulated MIS3 sea ice extent and the difference in thickness relative to the PI are shown in Fig. 9. In the NH, MIS3 sea ice is thicker by  $\sim 2$  m in the central Arctic in both March and September. In March, sea ice extends further equatorward in the Pacific (reaching  $40^{\circ}\text{N}$ ; not shown), and is associated with a cooling in the North Pacific (Figs. 4, 7a) and a southward migration of the Pacific subpolar gyre (Fig. ~~??~~ S4).

In the Atlantic at MIS3, there is more winter sea ice south of Newfoundland and in the northeastern Labrador Sea (Fig. 9a). However, there is less sea ice in the western Labrador Sea, which is due to the strong northerly katabatic wind induced by

the presence of the adjacent Laurentide Ice Sheet (Fig. 6). The Nordic Seas are partly ice-covered, with sea ice present off the coast of Norway in winter. However, the central part of the Nordic Seas is ice free even in winter, due to the intrusion of warm Atlantic water across the Iceland-Scotland ridge. Note that the presence of winter sea ice, in particular in the region southwest of Norway, is not solely governed by the surface climate and inflow of Atlantic water; the simulated nearby river runoff from the Fennoscandian Ice Sheet contributes roughly 0.05 Sv of freshwater input (see also the SSS field in Fig. 7) to the region (about a fourth of the Amazon River discharge), ~~which favors~~ favoring the formation of winter sea ice (Fig. 9a).

In September, the simulated MIS3 sea ice retreats and nearly coincides with PI sea ice extent in the Pacific side (not shown) and in the Labrador Sea (Fig. 9b). Greater sea ice extent is found along the coast of South Greenland and in the Nordic Seas. The Barents Sea is fully ice-covered also in summer at MIS3, in contrast to the PI experiment where this region is seasonally ice-free.

In the SH, MIS3 shows extended Antarctic sea ice cover in both seasons. The seasonal cycle is large in both MIS3 and PI experiments, with the total sea ice area varying by a factor of 3 and 4 between March and September, respectively (Fig. ~~??S2~~; Table 2). Furthermore, sea ice is thicker during MIS3, especially in the Weddell Sea region.

### Modes of variability

~~In this section,~~ we briefly evaluate the simulated change of two important climate internal variabilities: the El Niño-Southern Oscillation (ENSO) and the Northern Annular Mode (NAM).

~~The tropical Pacific cools nearly uniformly during MIS3, with a small change in the zonal SST gradient in the eastern Pacific cold tongue and western Pacific warm pool region (Fig. 7a). The amplitude of the SST change in the NINO3.4 region (170°W-120°W, 5°S-5°N) during MIS3 is about 1.2 °C relative to PI, with a weak seasonal cycle (Fig. ??a). As a measure of ENSO, the standard deviation ( $\sigma$ ) of the detrended monthly SST anomalies in the NINO3.4 region is smaller during MIS3 (0.45 °C) relative to PI (0.58 °C). The reduction is across all months and is greater in boreal autumn and winter (Fig. ??b), leading to a slightly weakened annual cycle of NINO3.4 SST variability. Similar reductions of ENSO variability were reported in CCSM3 and CCSM4 LGM simulations (Otto-Bliesner et al., 2006; Brady et al., 2013), but the results show disagreements among the PMIP2 LGM model simulations (Zheng et al., 2008). The results are included in the Supplementary Section 3.~~

MIS3 shows small and negative skewness across most of the year (Fig. ??e); the annual cycle is smaller during MIS3, with the largest discrepancy in boreal summer compared to PI. For the frequency of ENSO events, the NINO3.4 index exhibits most power over 2-5 years for both MIS3 and PI experiments, with the former showing more power in the lower and the latter in the higher end of the range.

Fig. ?? shows the composite anomalies of DJF SST during El Niño years for the MIS3 and PI experiments. An El Niño year is defined here as a year with the NINO3.4 SST anomalies greater than  $1.5\sigma$  for three consecutive months, with at least one DJF months. The SST anomalies in the tropical Pacific are weaker during MIS3 and have a smaller westward extent. The maximum SST anomalies, centered around 120°W, are  $\sim 1.2$  °C at MIS3, compared to  $\sim 1.5$  °C in PI. Stronger negative SST anomalies at MIS3 are seen in the subtropical Pacific in both hemispheres. In the southern Indian Ocean, stronger positive SST

anomalies are simulated during MIS3, whereas the anomalies are weaker in the eastern and central Indian Ocean relative to PI.

The NAM (also known as the Arctic Oscillation) is defined here as the first empirical orthogonal function (EOF) of the NH (20-90° N) DJF sea level pressure (SLP) anomalies. The NAM-explained winter SLP variance is reduced during MIS3 (27%) relative to PI (30%), with the centre of action over the Arctic slightly weaker and slightly eastward-shifted (Fig. ??a,b). The shape of the EOF pattern is more asymmetric over the Arctic at MIS3. With the presence of large ice sheets during MIS3, the simulated centre of action is weakened relative to PI in the North Pacific and eastern North Atlantic, resulting in a gradient from the mid-latitude to the pole that is weaker during MIS3.

#### 4 MIS3 simulation forced by stadial conditions

The NorESM MIS3 simulation presented above is representative of an interstadial climate, i.e. a relatively warm period during the last glacial; in agreement with paleo-reconstructions, this includes, Greenland temperatures only 5–8 °C colder than PI (Huber et al., 2006), a strong AMOC (Henry et al., 2016), and enhanced sea ice cover in the Nordic Seas (Rasmussen and Thomsen, 2004; ). In particular, a high resolution reconstruction of MIS3 sea ice, based on analysis of biomarkers from a marine sediment core in the south Nordic Seas (Sadatzki et al., manuscript submitted to Science Advances), finds near-perennial sea ice cover during cold stadials, and ice-free periods during warm interstadials. This supports our modeled sea ice distribution and suggests a simulated MIS3 interstadial climate state. However, questions remain as to the state of the cold stadial climates of MIS3, which is characterized by much colder Greenland temperatures, a completely ice-covered Nordic Seas, and reduced AMOC compared to the present day (based on the same references as above). These changes are reinforced during stadials including Heinrich events, when the AMOC shows a further weakening owing to the massive release of icebergs into the North Atlantic (e.g., Henry et al., 2016).

It is unclear if the baseline climate during MIS3 is a stadial or interstadial state, nor is it clear that there is indeed a baseline climate, as the climate states can be inherently oscillatory (Peltier and Vettoretti, 2014; Brown and Galbraith, 2016; Klockmann et al., 2018). To examine the possibility of simulating a cold MIS3 stadial climate with NorESM, an additional experiment is branched off from the MIS3 interstadial control experiment after 1700 model years. The MIS3 "stadial" experiment is run for 800 years with 40 ka orbital forcing and reduced GHG levels (e.g. CO<sub>2</sub> ~ 200 ppm, CH<sub>4</sub> ~ 450 ppb, N<sub>2</sub>O ~ 220 ppb) according to Schilt et al. (2010). The major difference compared to the interstadial experiment is that in the "stadial" experiment the CO<sub>2</sub> is lowered by 15 ppmv.

In the MIS3 "stadial" experiment, the global near surface temperature cools by 0.4°C relative to the MIS3 interstadial experiment (Fig. ??). The greatest cooling occurs along the newly formed sea ice margin in the Nordic Seas (~3.8°C) and southwest of the Bering Sea (~1.8°C). The change in South Greenland is ~1.3°C. The zonal mean cooling is ~1°C in the NH high latitudes and is reduced to ~0.3°C in the tropical and subtropical regions (figure not shown); in the Southern Ocean, the zonal mean cooling is ~0.7°C, decreasing to ~0.4°C over Antarctica.

The surface cooling in the Nordic Seas in the "stadial" experiment is reflected in the growth of sea ice in the region, e.g. winter sea ice slightly expands southwards in the Nordic Seas (Fig. ??). Similar increase of sea ice is also simulated in the northwestern Pacific, whereas in the Labrador Sea and in Antarctica, sea ice distribution is nearly identical in the two experiments. Furthermore, there is a negligible change to the AMOC in the "stadial" experiment. Together, these results indicate that given the changes to the GHG levels that is typical of a stadial state, a cold Greenland climate with a weak AMOC cannot be reproduced in our MIS3 setup.

## 4 Discussion

### 4.1 Simulated AMOC in MIS3

Abrupt climate changes such as D-O events have been shown to involve changes in both the geometry and strength of the AMOC, as indicated by a number of marine proxy reconstructions (e.g., see the review by Lynch-Stieglitz, 2017) and numerical simulations (e.g., Peltier and Vettoretti, 2014; Brown and Galbraith, 2016). A In this study, a stronger and deeper AMOC during is simulated during a typical MIS3 interstadial is simulated as compared to the PI (Fig. 8). Given its crucial role in the MIS3 climate climate of MIS3, we further discuss the simulated AMOC and the associated distribution of NADW and AABW in this section.

Previous studies have argued that the increased AABW ventilation and production production of AABW during glacial times is driven by expanded Antarctic sea ice and enhanced brine rejection during ice formation (Shin et al., 2003; Ferrari et al., 2014) :-the sea ice formation in the Southern Ocean (Shin et al., 2003; Ferrari et al., 2014). The brine induces a negative buoyancy flux, increasing open ocean convection and leading to enhanced formation of highly saline AABW (Fig. 10d). Jansen (2017) shows that the changes in the deep ocean stratification and circulation can be interpreted as a direct consequence of atmospheric cooling during glacial times, which induces Antarctic sea ice growth and initiates the processes described above.

The simulated enhanced ventilation of AABW during is enhanced in our MIS3 simulation compared to the PI (Fig. S7). However, in the Atlantic the volume of AABW is not comparable with studies that of the LGM, during which benthic foraminiferal  $\delta^{13}\text{C}$  data suggests that AABW dominated the water column in the Atlantic below up to a depth of  $\sim 2$  km (Curry and Oppo, 2005), together with a shallower NADW cell production. However, measurements of  $^{231}\text{Pa}/^{230}\text{Th}$ , in combination with  $^{143}\text{Nd}/^{144}\text{Nd}$  ( $\epsilon_{\text{Nd}}$ ), indicate that a strong AMOC existed at the LGM despite of a shallow upper cell (Böhm et al., 2015). Böhm et al. (2015) also show that an active AMOC, neither weaker nor shallower than the present day, prevailed over the last glacial cycle, including the D-O interstadials (the exceptions are Heinrich stadials exhibiting a weaker and shallower NADW cell); The AMOC during interstadials is accompanied by active deep water formation in the North Atlantic, with persistent contributions from northern sourced water. Our simulated MIS3 interstadial AMOC agrees with these reconstructions based on chemical tracers, and the difference with the AMOC at LGM can be partly (if not fully) attributed to less Antarctic sea ice formation due to the MIS3 climate being milder than that at the LGM.

While deep water production in the Southern Ocean has the potential to displace and reduce the strength of the NADW cell production, competing effects are at play in the North Atlantic. For example, the altered surface westerlies in the North

Atlantic caused by the elevated Laurentide Ice Sheet (Figs. 2,6) are shown to be able to induce a deeper and stronger AMOC by transporting more salt northward within an intensified gyre circulation, favouring deep ocean convection (Muglia and Schmittner, 2015; Klockmann et al., 2016; Sherriff-Tadano et al., 2018; e.g., Montoya and Levermann, 2008; Oka et al., 2012; Muglia and Schmittner, 2015; Klockmann et al., 2016; Sherriff-Tadano et al., 2018); the closure of Bering Strait leads to an increase of surface salinity in the North Atlantic thereby invigorating deep ocean convection and strengthening the AMOC (Hu et al., 2010, 2015). To isolate and assess the relative impact of these different processes requires a suite of dedicated sensitivity studies which is beyond the scope of this paper, but it is worth mentioning that the processes that take place in both hemispheres act together to create the AMOC shown in Fig. 8.

During the last glacial, sea level lowering and the removal of shallow continental shelves (Fig. 2) result in enhanced tidal dissipation in the open ocean (Egbert et al., 2004), implying enhanced deep ocean mixing and a strengthened AMOC. Schmittner et al. (2015) demonstrated that such tidal effects can dominate over surface buoyancy effects and lead to a strengthening of AMOC by  $\sim 40\%$ . Such tidal effects are not considered in the current study, but once included, the AMOC is expected to strengthen and deepen, potentially displacing the lower AABW cell in the Atlantic.

#### 4.2 MIS3 simulation forced by typical stadial conditions

The NorESM MIS3 simulation presented above is representative of an interstadial climate, i.e. a relatively warm period during the last glacial; in agreement with paleo reconstructions, this includes, Greenland temperatures only 5–8 °C colder than PI (Huber et al., 2006), a strong and active AMOC (Böhm et al., 2015; Henry et al., 2016), and enhanced sea ice-cover in the Nordic Seas (Rasmussen and Thomsen, 2004; Dokken et al., 2013). In particular, using a high-resolution benthic Mg/Ca temperature record in the Denmark Strait, Sessford et al. (2018) inferred that during the baseline interstadial mode, ocean circulation and the associated water mass properties are similar to the present day; furthermore, a high resolution reconstruction of MIS3 sea ice, based on analysis of biomarkers from a marine sediment core in the south Nordic Seas (Sadatzki et al., 2019), finds near-perennial sea ice cover during cold stadials, and ice free periods during warm interstadials. Both proxy studies support our modeled ocean and sea ice state as simulated for the MIS3 interstadial climate state. However, questions remain as to the state of the cold stadial climates of MIS3, which is characterized by much colder Greenland temperatures, a near-completely ice-covered Nordic Seas, and reduced AMOC compared to the present day (based on the same references as above). These changes are reinforced during stadials including Heinrich-events, when the AMOC shows a further weakening, owing to the massive release of icebergs into the North Atlantic (e.g., Henry et al., 2016).

It is unclear if the baseline climate during MIS3 should be a stadial or interstadial state, nor is it clear that there is indeed a baseline climate, as the climate states can be inherently oscillatory (Peltier and Vettoretti, 2014; Brown and Galbraith, 2016; Klockmann et al., 2016). To examine the possibility of simulating a cold MIS3 stadial climate with NorESM, an additional experiment is branched off and initialised from the MIS3 interstadial control experiment after 1700 model years. The MIS3 "stadial" experiment is run for 800 years with 40 ka orbital forcing and reduced GHG levels (e.g. CO<sub>2</sub> - 200 ppm, CH<sub>4</sub> - 450 ppb, N<sub>2</sub>O - 220 ppb) according to Schilt et al. (2010). The major difference compared to the interstadial experiment is that in the "stadial" experiment, the CO<sub>2</sub> is lowered by 15 ppmv.

In the MIS3 "stadial" experiment, the global near surface temperature cools by 0.4°C relative to the MIS3 interstadial experiment (Fig. S11). Winter sea ice slightly expands southwards in the Nordic Seas and in the northwestern Pacific, whereas in the Labrador Sea, sea ice distribution is nearly identical in the two experiments. Furthermore, there is a negligible change to the AMOC. Together, these results indicate that given the changes to the GHG levels that are typical of a stadial state, a cold Greenland climate with a weak AMOC cannot be reproduced in our MIS3 setup without additional freshwater forcing.

### 4.3 MIS3 sensitivity to CO<sub>2</sub> and ice sheet size

With the multi-millennial long integration of the MIS3 simulation presented in this work, only one stable climate state is found, and the model reaches a quasi-equilibrium with a small drift towards the end of the integration. In this section, we explore the potential for model bi-stability associated with the transition between the warm interstadial and cold stadial climate states of MIS3. We do so by perturbing the model with changes in atmospheric CO<sub>2</sub> concentrations and the size of the Laurentide Ice Sheet.

Our NorESM MIS3 simulations agree with that of Van Meerbeeck et al. (2009) using the LOVECLIM model: given MIS3 boundary conditions, the interstadial climate is the equilibrium climate state, whereas the stadial climate is a perturbed state. A CO<sub>2</sub> change of 15 ppmv (typical of stadial conditions) is not sufficient to induce transitions between the two states (Section 4.2). However, Zhang et al. (2017) reported that a gradual change of CO<sub>2</sub> concentrations by 15 ppmv in their experiment with intermediate glacial conditions can be sufficient to trigger transitions between weak/strong AMOC modes and therefore stadial/interstadial states. In contrast, Klockmann et al. (2016) examined the effect of CO<sub>2</sub> changes on the AMOC strength and geometry in an LGM setup, and found a weakening and shoaling of AMOC with decreasing CO<sub>2</sub> (from 284 to 149 ppmv), but without transitioning into a weak AMOC mode. They argued that the presence of LGM ice sheets could enhance the AMOC by impacting the surface wind stress in their model, and thus help maintain the stability of the AMOC. Interestingly, Klockmann et al. (2018) later reported that with a PI ice sheet configuration, the AMOC exhibits ~~bi-stability with D-O like~~ oscillatory behaviour at a CO<sub>2</sub> level of 206 ppmv, above (below) which a strong (weak) AMOC mode persists. In addition, the MIS3 simulation reported by Zhang et al. (2014b) is close to disequilibrium and model bi-stability, as indicated by a series of equilibrium freshwater perturbation experiments.

The studies by Klockmann et al. (2016, 2018); Zhang et al. (2017) on the sensitivity of climate states to CO<sub>2</sub> levels and ice sheet sizes, together with the studies of ~~(e.g., Zhang et al., 2014a; Brown and Galbraith, 2016; Galbraith and de Lavergne, 2018)~~, e.g. Zhang et al. (2014a); Brown and Galbraith (2016); Galbraith and de Lavergne (2018), point to the possibility that there could exist a certain range of CO<sub>2</sub> levels and ice sheet sizes in which the model is subject to a mode transition and even excitation of self-oscillatory behaviour in North Atlantic climate. The existence and range of such a "window" of CO<sub>2</sub> levels and ice sheet heights, however, is expected to be model-dependent. In order to explore the potential for model ~~bi-stability climate~~ transition with NorESM and the existence of a cold stadial state given MIS3 boundary conditions, we investigate the model response to large variations of CO<sub>2</sub> and ice sheet height, using the interstadial simulation as the baseline experiment. Another motivation for seeking a cold climate is that, as discussed by Peltier and Vettoretti (2014), a cold Heinrich stadial-like state

is required to "kick" the system into a self-oscillatory behaviour. Even without the self-oscillation, a cold stadial state would serve as a useful baseline experiment for investigating potential triggers of the abrupt cold-to-warm D-O transitions.

It is not within the scope of this paper to perform a ~~thorough complete~~ examination of the model response to every combination of CO<sub>2</sub> and ice sheet changes. Rather, we report the model response in the more extreme cases, with a primary focus on the response of the AMOC and sea ice. For the CO<sub>2</sub> sensitivity experiments, we reduce the values from 215 ppmv to 180 ppmv and 140 ppmv, ~~respectively~~. For the ice sheet sensitivity experiments, we reduce the height of the Laurentide Ice Sheet by 50% and 100% (the latter is equal to the pre-industrial orography), while keeping the ice mask unchanged. All the sensitivity experiments are branched off ~~and initialised~~ from the MIS3 interstadial simulation, and all other parameters are kept fixed.

~~Contrary~~ ~~Contrary~~ to the studies cited above, the NorESM MIS3 experiments exhibit surprising stability without any significant changes in Greenland temperature, sea ice, or AMOC (Fig. 11). For the low CO<sub>2</sub> experiments, ~~there is a slight expansion of~~ winter sea ice ~~expands slightly~~ in the Nordic Seas (Fig. S11), without any notable changes in the strength and ~~regions of convection(not shown)locations of convection~~. As a consequence, the AMOC, ~~although weakened, still~~ remains strong (Fig. 11). For the experiments with a reduced Laurentide ice sheet, surface wind stress fields are altered (mainly ~~shifted shifting~~ northwards in the North Atlantic; not shown), whereas the strength of AMOC is ~~not affected only slightly reduced~~.

~~Further analysis of key relevant metrics, e.g. spatial distributions of SSS, winter sea ice extent, and AMOC geometry for the sensitivity experiments are included in the Supplementary Section 4. As the changes are highly related to the strength of the AMOC, which is only weakly reduced in the sensitivity experiments, changes in these metrics are also relatively small.~~

The results of the sensitivity experiments to CO<sub>2</sub> and ice sheet height underscore the question of model dependence on the background climate in simulating AMOC transition/bi-stability, ~~despite the fact that the simulations are limited in length (200 to 1000 years) and we therefore cannot fully exclude that further equilibration could bring the climate into a more unstable regime~~. In our simulation, where the Labrador ~~Sea~~ and the Norwegian ~~Sea-Seas~~ are the major convection sites, a significant change of ocean circulations ~~would not be is not~~ expected unless the Labrador ~~Sea~~ and the Norwegian ~~Sea-Seas~~ are covered by ~~ice-inhibiting convection~~ ~~sea ice, thereby inhibiting convection through its insulating effect~~. However, the NorESM MIS3 and PI experiments both appear to be in a stable regime with a strong AMOC and strong convection in the Labrador and Norwegian Seas. As a consequence, the model state is relatively distant from a potential threshold, including a bi-stability of the AMOC and sea ice. In addition, the NorESM1-F model features a ~~relatively~~ low climate sensitivity (the equilibrium climate sensitivity in response to a doubling of atmospheric CO<sub>2</sub> is 2.3 °C) which plays a role in the ~~relatively~~-weak response of the MIS3 climate to a further CO<sub>2</sub> decrease.

To trigger a cold stadial-like climate state in the NorESM, other mechanism including enhanced ~~ice-berg~~ ~~iceberg~~ calving and freshwater input to the North Atlantic from the Laurentide, Greenland and Fennoscandian ice sheets should be considered. There is a rich literature on applying freshwater flux of different magnitudes and locations to study climate response and transitions (e.g. Stouffer et al., 2006; Roche et al., 2010). ~~While beyond the scope of this study, the~~ ~~The~~ prospects of perturbing the MIS3 interstadial climate with freshwater fluxes will be explored in detail in a companion study.



## 5 Conclusions

In this paper, we present an equilibrium simulation of Marine Isotope Stage 3 forced by 38 ka BP boundary conditions, with a recently developed version of the NorESM featuring a horizontal resolution of  $2^\circ$  ~~atmosphere/land-in the atmosphere~~ and  $1^\circ$  ~~ocean/sea-ice in the ocean~~. The fast performance of the model allows the experiments to be integrated for 2500 years. The boundary conditions are notably different from the pre-industrial ~~mainly due to changes~~ in orbital forcing, greenhouse gases, and the height and extent of the global ice sheets.

The reported simulation, with its current length of integration, does not produce spontaneous transitions between colder stadial and warmer interstadial climate states. Rather, we obtain a MIS3 background climate state with a state-of-the-art climate model that can serve as a baseline for investigating mechanisms behind D-O ~~event-events~~ by discriminating different factors that can invoke abrupt transitions, e.g., freshwater input, changes in GHG concentrations, ice sheet size, orbital forcing, and ocean diapycnal mixing.

Despite a small drift due to ~~the ocean cooling as the model is integrated~~ ~~ocean cooling~~, the model reaches a quasi-equilibrium state in terms of both surface ~~metries-properties~~ and deep ocean hydrography. We analyze the large-scale features of the mean climate states and the model internal variabilities, and compare the results to previous ~~simulation studies of both~~ ~~studies of~~ MIS3 and LGM climates. The major findings are as follows:

- ~~Simulated Globally, the simulated~~ MIS3 interstadial climate is ~~globally~~  $2.9^\circ\text{C}$  cooler ~~relative to than~~ the PI, with amplified cooling in the high latitudes, especially above the ice sheets and near the edges of the newly formed sea ice. The presence of the Laurentide Ice Sheet amplifies the atmospheric stationary waves, leading to an enhanced and northward-shifted jet stream, ~~and with~~ stronger and southward-shifted wind stress at the ocean surface. ~~Global mean precipitation during MIS3 is  $0.18\text{ mm day}^{-1}$  lower in the colder MIS3 climate, with both seasonal and geographical changes including a southward shift of the ITCZ.~~
- The global mean SST at MIS3 is  $1.2^\circ\text{C}$  colder than that at PI, with a pattern of modest cooling in the tropics and enhanced cooling at high latitudes. The North Atlantic subpolar region is characterised by less cooling owing to enhanced AMOC and ocean heat transport. Greater cooling is simulated in the North Pacific associated with the expansion of sea ice and southward shift of the subpolar gyre.
- Despite the uniform addition of salt ~~into the global ocean~~ (by  $0.6\text{ g kg}^{-1}$ ) ~~into the global ocean~~, the distribution of SSS exhibits an inhomogeneous pattern of both salinity increase (e.g., in the central Arctic, the Baffin Bay, and the Weddell Sea region) and decrease (e.g., off the Eurasian shelf, in the North Pacific, and western Pacific and the South China Sea). The closure of the Bering Strait, the increase in sea ice formation, and the change of glacial river routing all play important roles in determining the SSS pattern.
- The upper cell of the AMOC is deepened and intensified under the influence of competing factors from both hemispheres: ~~e.g.,~~ the cutoff of freshwater input due to the closed Bering Strait, and the strengthened surface wind stress in the NH subpolar region, both tend to invigorate ~~the~~ AMOC. In the Southern Ocean, expansion of Antarctic sea ice stimulates

AABW production by enhanced salt rejection and deep water production ~~during sea ice formation~~. The results are supported by marine proxy records indicating an AMOC comparable to the present day during the last glacial (except during Heinrich stadials). The enhanced deep ocean ventilation in the Atlantic sector of the Southern Ocean leads to reduced (but not reversed) deep ~~ocean~~-north-south salinity gradients in the Atlantic Ocean. The Atlantic displays pronounced cooling below 3000 m in both hemispheres and near the base of the thermocline, the latter due to stronger upwelling of deep water as a result of enhanced deep water formation in both hemispheres. Reduced Mediterranean outflow during MIS3 contributes to the notable cooling and freshening observed around ~~30N~~30°N, 500-800 m depth.

- Sea ice is notably thicker and greater in extent during MIS3 in both hemispheres and seasons. Arctic sea ice is about 2 meters thicker and extends further equatorward in the Pacific during winter. The Nordic Seas are partly ice-covered in boreal summer; in winter, sea ice extent is greater, but includes an opening in the south due to the intrusion of warm Atlantic Water. In the Southern Hemisphere, Antarctic sea ice is thicker (mainly in the western Indian Ocean sector) and extends further north.
- ~~Simulated ENSO variability is weakened in the MIS3 compared to the PI simulation. For the Arctic Oscillation, simulated centres of action over the Arctic and North Pacific are both weakened, with the latter much reduced due to the presence of the elevated Laurentide Ice Sheet.~~
- A sensitivity experiment with boundary conditions typical of MIS3 stadial conditions does not reproduce the cold temperatures observed on Greenland, indicating that the interstadial climate in NorESM is relatively stable, and forms the baseline climate during MIS3. Further sensitivity experiments including large changes in atmospheric CO<sub>2</sub> levels and Laurentide Ice Sheet heights, aimed at perturbing the system into a cold stadial-like climate, show that the model is not subject to any bi-stability of the AMOC or sea ice. This underscores the role of model ~~dependene~~dependency in studying abrupt climate transitions during MIS3 ~~or in other geological periods in general~~and potentially in the future.

*Code and data availability.* The model code can be obtained upon request. Instructions on how to obtain a copy are given at <https://wiki.met.no/noresm/gitbestpractice>. The full set of model data will be made publicly available through the Norwegian Research Data Archive at <https://archive.norstore.no> upon publication.

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

*Acknowledgements.* This work was funded by the Ice2Ice project that has received funding from the European Research Council under the European Community's Seventh Framework Programme (FP7/2007-2013)/ERC grant agreement No. 610055. We acknowledge Peter Langen, Christian Rodehacke, and Will Roberts for the discussion of configuring MIS3 boundary conditions during a workshop in May

2016. We thank Lev Tarasov for providing the ice sheet data to force the model. We also thank Anne-Katrine Faber for commenting on part of the manuscript. The simulations were performed on resources provided by UNINETT Sigma2 - the National Infrastructure for High Performance Computing and Data Storage in Norway (nn4659k, ns4659k).

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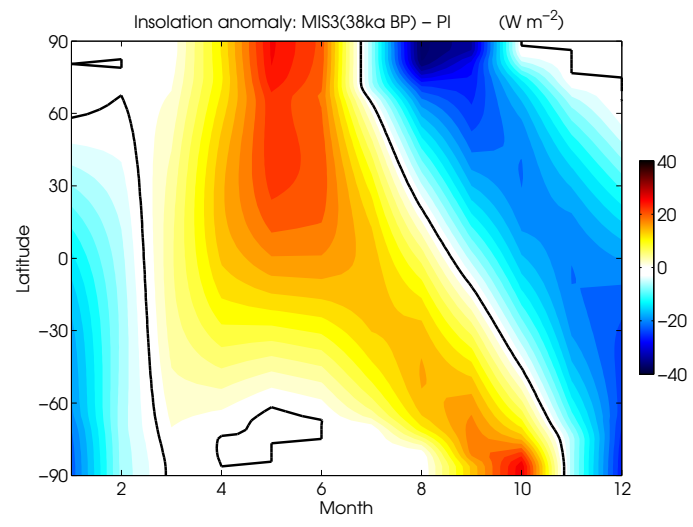
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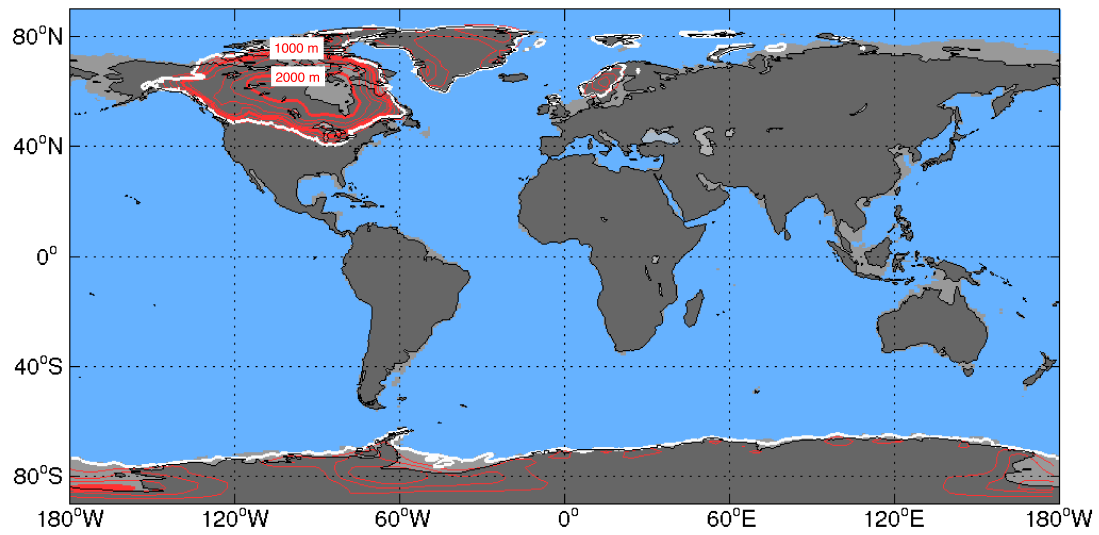
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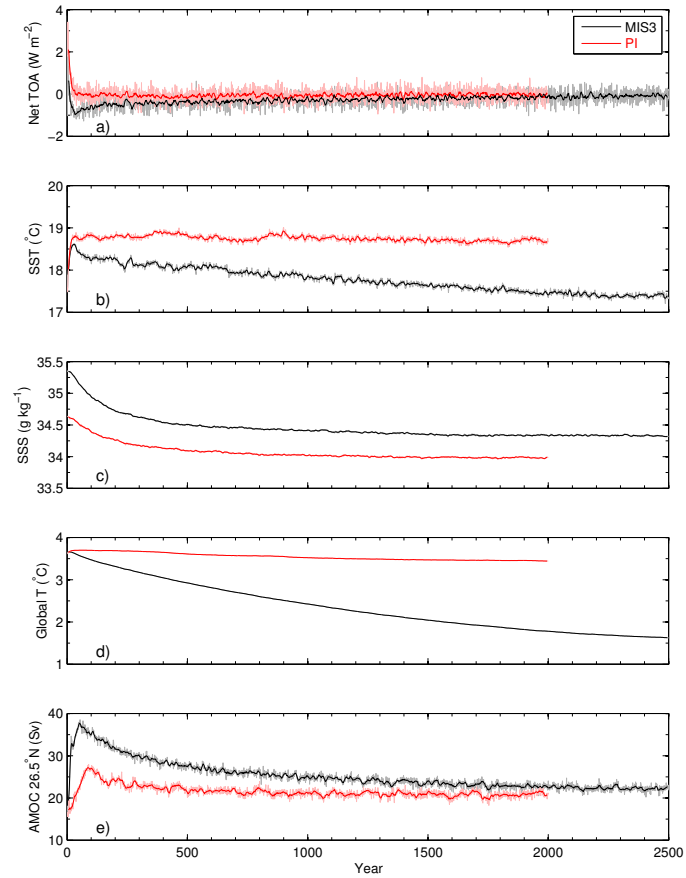
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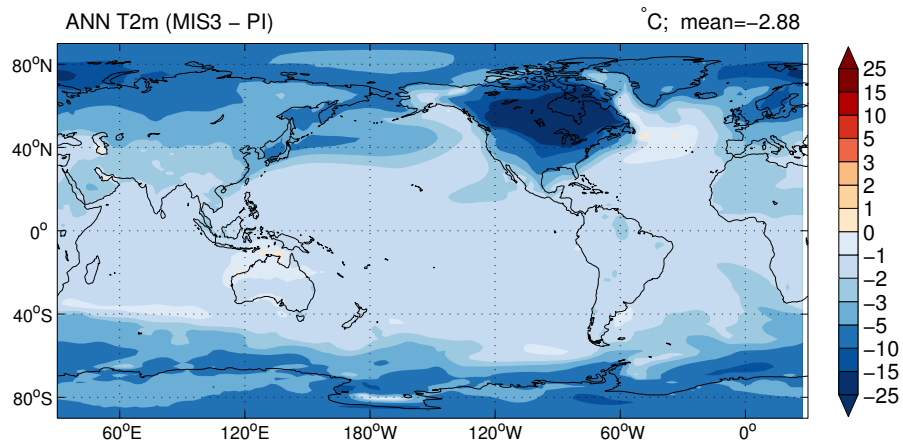
**Figure 1.** The insolation anomalies of MIS3 at 38 ka BP relative to present day ( $W m^{-2}$ ). The thick contour is the zero isoline.



**Figure 2.** Land-sea mask (light grey/blue shading; dark grey are modern land), ice sheet extent (white line) for the MIS3 experiment, and the difference of ice sheet orography relative to PI (red contours with an interval of 250 m; the 1000 and 2000-m isolines are highlighted with bold red lines).

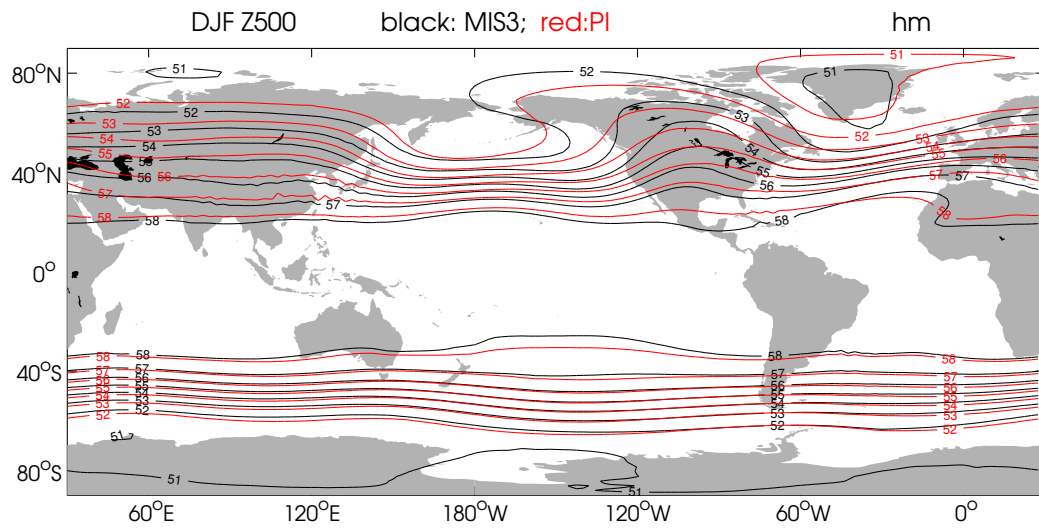


**Figure 3.** Time series of a) TOA radiation balance, b) SST, c) SSS, d) global mean ocean temperature, and e) AMOC strength at 26.5° N for the MIS3 (black) and PI (red) experiments. Light colors denote annual mean values, and dark colors denote 10-year running mean values.

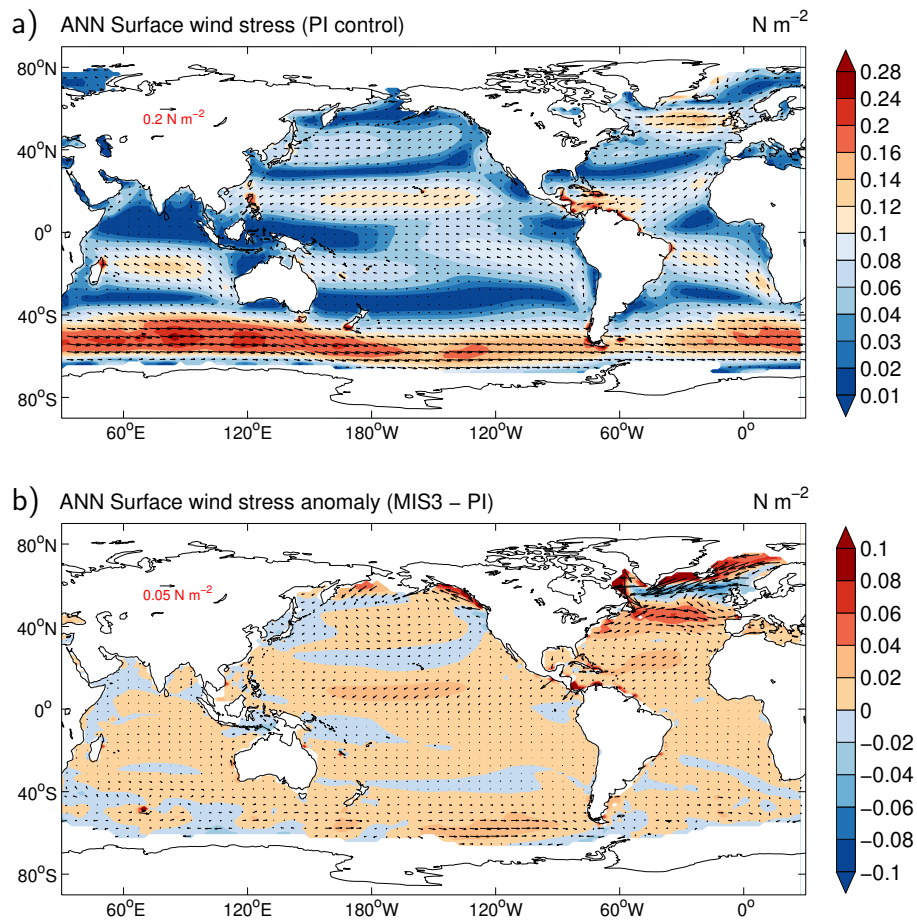


**Figure 4.** Simulated MIS3 minus PI annual mean near surface air temperature (°C).

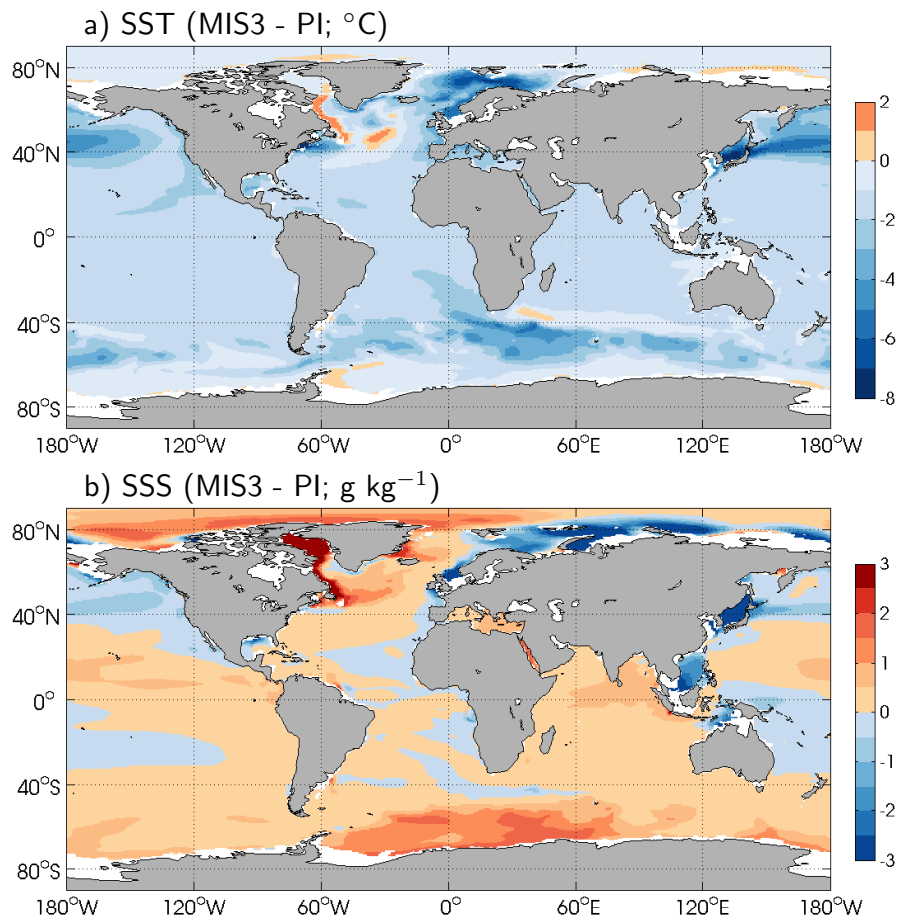
Time-series of a) Northern Hemisphere and b) Southern Hemisphere sea-ice area for the MIS3 (black) and PI (red) experiments. The data shown are 10-year running mean values.



**Figure 5.** Simulated MIS3 and PI DJF 500-mb geopotential height (hm). The black and red contours are for the MIS3 and PI experiments, respectively.



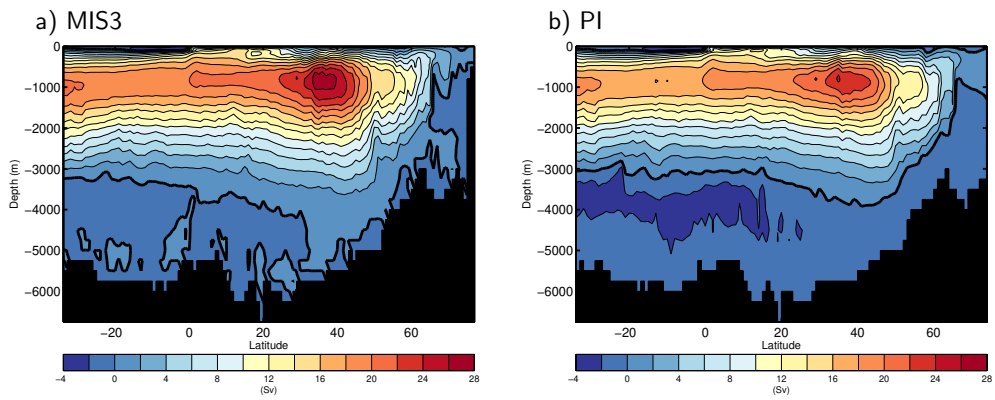
**Figure 6.** Simulated annual surface wind stress over the ocean for a) PI and b) MIS3 minus PI ( $\text{N m}^{-2}$ ).



**Figure 7.** Simulated ~~difference of~~ MIS3 a) SST and b) SSS anomalies relative to PI.

*Simulated seasonal (DJF/JJA) total precipitation for a,b) PI and c,d) MIS3 minus PI ( $\text{mm day}^{-1}$ ).*



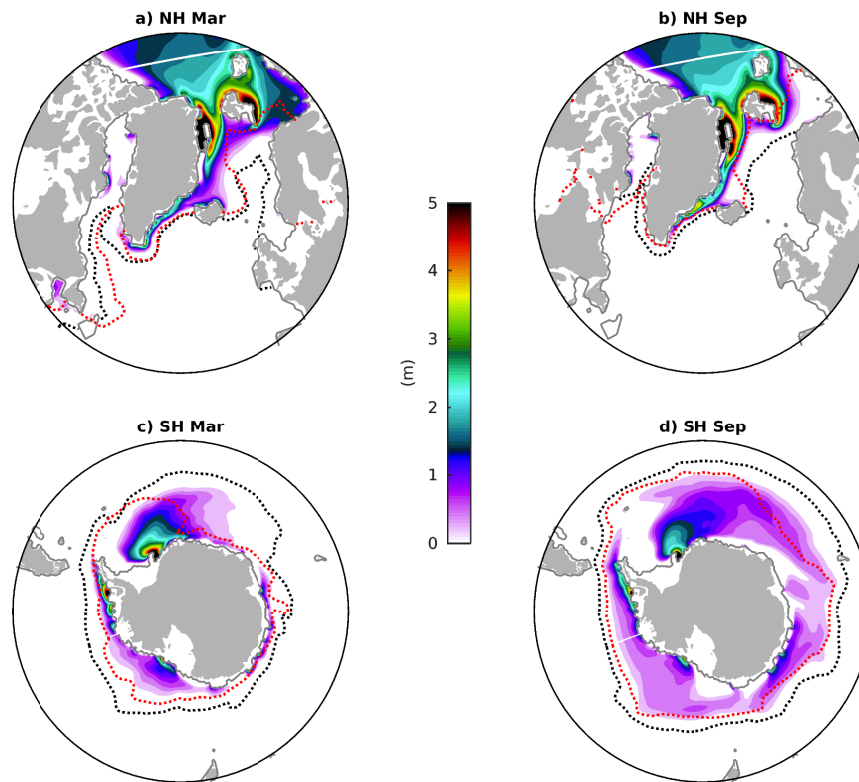


**Figure 8.** Stream functions of AMOC Atlantic Meridional Overturning Circulation for a) MIS3 and b) PI simulations. The thick black line denotes the zero contour line.

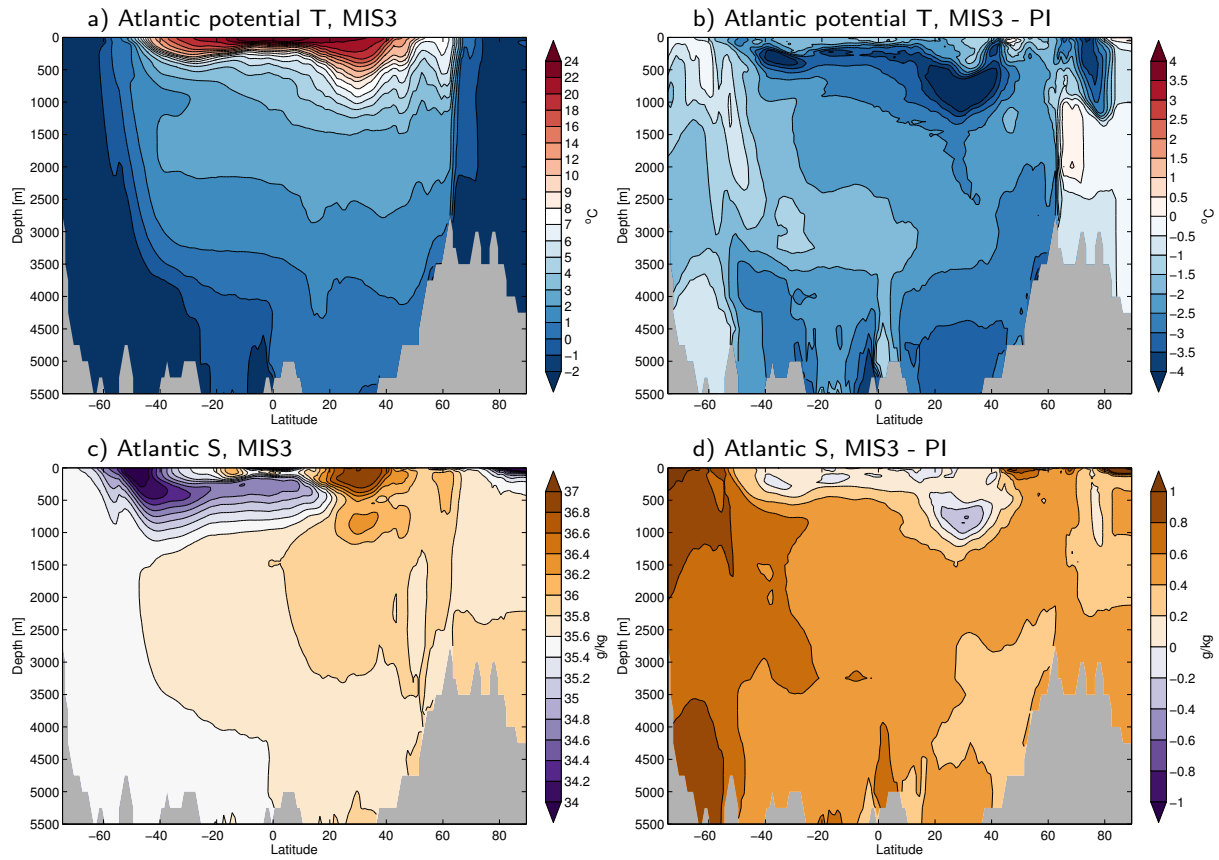
Barotropic stream functions of ocean circulation in the North Pacific and North Atlantic region for MIS3 (black) and PI (red) simulations.

Contour intervals are 10 Sv, with solid and dashed colors denoting positive and negative values, respectively.

Simulated MIS3 (black) and PI (red) northward heat transport for a) global atmosphere, ocean, and total, and for b) global ocean, Atlantic Ocean, and Pacific and Indian Ocean. The ocean heat transport is calculated directly from the ocean model, and the atmospheric heat transport is calculated by meridional integration of the difference between the zonal integration of the net TOA and surface heat flux.



**Figure 9.** Simulated difference of MIS3 - PI sea ice thickness with PI anomalies (shading; m) for a) NH March, b) NH September, c) SH March, and d) SH September. Black and red dashed lines denote the 15% sea ice concentration for MIS3 and PI, respectively. The grey line is marks the MIS3 coast line.



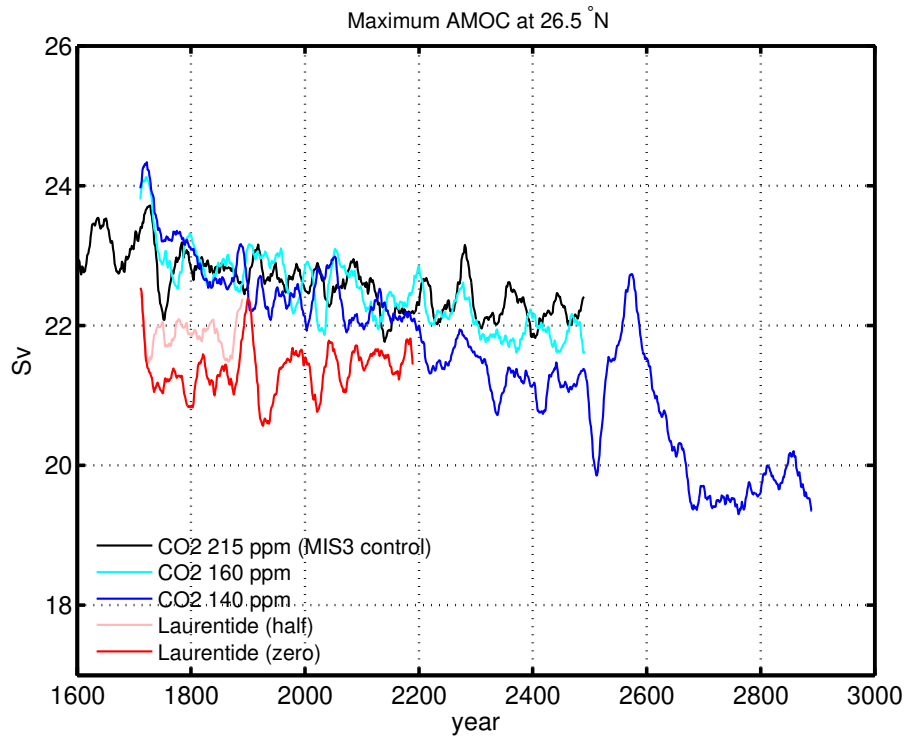
**Figure 10.** Atlantic zonal mean **distribution of** a) potential temperature, **and** c) salinity for the MIS3 experiment, and b,d) **the differences anomalies** relative to PI.

Monthly interannual a) SST, b) standard deviation of SST anomalies, c) skewness of SST anomalies in the NINO3.4 region, and d) power spectra of the NINO3.4 index for the MIS3 and PI, respectively.

Composite DJF SST anomalies during El Niño years for a) PI and b) MIS3.

a,b) Leading empirical orthogonal function (EOF) of the winter (DJF) mean sea-level pressure (SLP) anomalies over the NH (20-90° N) for a) MIS3 and b) PI. The SLP patterns are obtained by regression of anomalies on the leading principal component time series. The contour intervals in both panels are 1 hPa, with the zero line omitted.

Simulated near-surface temperature anomaly of MIS3 "stadial" experiment relative to MIS3 interstadial control run (model output averaged between years 2401-2500). The black and magenta lines indicate the 15% March sea ice concentration for the MIS3 interstadial and "stadial" experiments, respectively.



**Figure 11.** Time series of AMOC at 26.5° N for the low experiments with reduced CO<sub>2</sub> experiments levels and reduced Laurentide Ice Sheet experiments heights. The CO<sub>2</sub> and ice sheet sensitivity experiments are branched off from the MIS3 control run from at year 1700 and year 2500, respectively, and the latter are shifted forward by 800 years in the figure.

**Table 1.** Forcings and boundary conditions for the MIS3 and PI simulations.

exp.	MIS3	PI
Orbital parameters		
Eccentricity	0.013676	0.016708
Obliquity	23.268°	23.441°
Perihelion - 180°	205.94°	102.72°
Trace gases		
CO <sub>2</sub>	215 ppm	285 ppm
CH <sub>4</sub>	550 ppb	792 ppb
N <sub>2</sub> O	260 ppb	276 ppb
CFC	0	12.5 ppt
Solar constant	1360.9 W m <sup>-2</sup>	1360.9 W m <sup>-2</sup>
Ice sheets	data-constrained 38 ka BP	Modern
Vegetation	PI + tundra (new land points)	Modern

**Table 2.** Global mean values for the MIS3 and PI experiments (both averaged between years 1801-2000).

exp.	MIS3	PI
TOA radiation balance	-0.16 W m <sup>-2</sup>	-0.04 W m <sup>-2</sup>
T2m	11.6 °C	14.5 °C
Precipitation	2.66 mm day <sup>-1</sup>	2.84 mm day <sup>-1</sup>
SST	17.5 °C	18.7 °C
SSS	34.3 g kg <sup>-1</sup>	34.0 g kg <sup>-1</sup>
NINO3.4 $\sigma$	0.45 °C	0.58 °C
Global ocean T	1.8 °C	3.5 °C
Global ocean S	35.3 g kg <sup>-1</sup>	34.7 g kg <sup>-1</sup>
Ocean Transports		
Maximum AMOC	27.5 Sv	24.3 Sv
AMOC at 26.5° N	22.8 Sv	20.9 Sv
North Atlantic subtropical gyre	63.8 Sv	53.6 Sv
North Atlantic subpolar gyre	60.7 Sv	50.3 Sv
Florida Strait	16.2 Sv	15.1 Sv
Bering Strait	closed	1.3 Sv
Barents Sea Opening	0.5 Sv	2.8 Sv
Drake Passage	137.9 Sv	114.2 Sv
Sea ice area		
NH Mar	11.1 × 10 <sup>6</sup> km <sup>2</sup>	13.5 × 10 <sup>6</sup> km <sup>2</sup>
NH Sep	6.3 × 10 <sup>6</sup> km <sup>2</sup>	5.2 × 10 <sup>6</sup> km <sup>2</sup>
SH Mar	7.3 × 10 <sup>6</sup> km <sup>2</sup>	4.4 × 10 <sup>6</sup> km <sup>2</sup>
SH Sep	21.5 × 10 <sup>6</sup> km <sup>2</sup>	16.4 × 10 <sup>6</sup> km <sup>2</sup>