"Nonlinear response of the Antarctic Ice Sheet to late-Quaternary sea level and climate forcing"

Author response to reviewer comments

The authors thank Johannes Sutter and an anonymous reviewer for their thorough review and helpful comments. Below is a line-by-line response to their feedback, with reviewer comments in italics.

Reply to Reviewer Comment 1 (Anonymous)

General comments:

1. Although this is a modelling study, the authors could reach a wider audience and better justify these experiments with more discussion of the experiments in the context of the geologic record. Much of the current debate on the relative roles of external forcings in driving Antarctic Ice Sheet changes is from surface-exposure chronologies that appear to show ice thinning of glaciers that occurs synchronously with changes in some external forcings, but not others (see Goehring et al., 2019 for a recent example). The sensitivity experiments are suited for testing these proposed mechanisms and assumptions, and this justification can be included in the introduction. It is not necessary to compare the model to every record, but a general indication of how the experiments compare to LGM reconstructions (e.g. Bentley et al., 2014) would also be of interest to many. Are the model experiments more consistent with reconstructions in some areas than others? This LGM comparison could be briefly mentioned in Section 3.2.

Thank you for this excellent suggestion. With regards to contextualizing the motivation for these simulations with the deglaciation record, we will add a few sentences about this to the introduction. As for a comparison against the LGM reconstructions, this was discussed in more detail in Tigchelaar et al. (2018). There we wrote that "During glacial maxima, the AIS grounding line extends to the continental shelf break almost everywhere (Fig. 2b). The simulated local LGM ice volume maximum occurs from 23–20 ka, with the grounding line position at this time generally in close agreement with reconstructions. The last deglaciation begins in the Bellingshausen sector before 16 ka, followed by the Amundsen sector around 13 ka, and the Weddell, Ross and Amery sectors around 10 ka and beyond (SI Fig. 2), a pattern in general agreement with time slice reconstructions of grounding line position (Bentley et al., 2014). However, our retreat in the Ross sector occurs at least ~2kyr earlier, and further work will focus on this discrepancy. In the Ross sector, there is an "overshoot" of Siple Coast grounding lines at ~6 to 4ka, and a subsequent re-advance to the modern positions by 0ka. Similar but more localized behaviour occurs in the Weddell sector (SI Fig. 2). This feature was also described in Maris et al. (2014), and is probably due to time-lagged isostatic bedrock rebound and shallowing grounding lines allowing re-advance in the late Holocene (Bradley et al., 2015)."

As mentioned there, the challenge to achieve good simulations of the last deglacial retreat remains a major focus of our and others' modeling efforts. We and others have

applied large ensembles of model parameter sets, and automated scoring algorithms (Briggs et al., 2013, 2014; Pollard et at 2016, 2017), comparing with several diverse types of paleo data, reviewed for instance in RAISED (2014). This data includes grounding-line positions vs. time, and cosmogenically-derived thinning in inland marginal areas. We will add a brief discussion with some of these comparisons to Sect. 3.2 of the manuscript.

2. Another aspect that the authors could improve on is the clarification of caveats and model limitations, which may impact the relative and synergistic effects of the external forcings. There are two key limitations that require more detailed explanations: the sub-ice shelf melt parameterization and the sea level forcing.

2a. The relationship between ocean temperature and ice shelf depth to basal melting/freezing of ice shelves is complex and sub-ice shelf melt parameterization is an active area of research within the ice sheet model community. This is well-outlined in the review paper of Pattyn et al. (2017). The previous paper of Tigchelaar et al. (2018) offers a more detailed discussion of some of these uncertainties with respect to interglacial ocean temperatures, which is worth reiterating in this paper as well since this analysis specifically investigates the individual and combined effect of the ocean forcing. The current discussion seems too brief and there is little information offered in either paper of the parameterization used for sub-ice shelf melting/freezing (see specific comments below).

This is a very valid criticism and one that was shared by the other reviewer. As detailed below, we will expand the discussion of the basal melt parameterization in the Methods section. We will also be more explicit about the shortcomings of the ocean temperature forcing used, and the implications for conclusions about relative importance of external drivers. Specifically, we will elaborate on this in Sect. 3.1 (climate forcing), 3.4.3 (ocean-only run), 3.4.4 (combined forcings), and the Discussion.

For the Discussion, we suggest the following updated text:
“Our finding that ocean temperature forcing plays a limited role in driving changes in Antarctic ice volume contrasts with previous modeling studies of past and future AIS evolution (Golledge et al., 2015; DeConto and Pollard, 2016; Sutter et al., 2016), as well as observations of sustained sub-shelf ice loss in response to ongoing ocean warming at e.g., Pine Island Glacier (Jacobs et al., 2011; Pritchard et al., 2012). This is not surprising, given that the LOVECLIM-simulated ocean temperature anomalies are small (Figs. 3i,l), and ice sheet models typically need ocean warming of 2-5 ºC to initiate interglacial WAIS collapse (Pollard and DeConto, 2009; DeConto and Pollard, 2016; Sutter et al., 2016; Tigchelaar et al., 2018). Absent paleo-reconstructions of near-Antarctic sub-surface ocean temperatures it is difficult to assess how realistic our LOVECLIM simulation is, though critical processes such as Antarctic Bottom Water formation are known to be poorly represented in low-resolution climate models (e.g., Snow et al., 2015), and previous studies have found LOVECLIM in particular to have more muted late-Quaternary temperature variability than other models (Lowry et al., 2019).

In addition, many regional oceanographic processes can affect the circum-Antarctic ocean environment beyond large-scale climate forcing. For example, the blocking effects of sea ice formation (Hellmer et al., 2012), the role of winds in pushing warm waters onto the continental shelf (Thoma et al., 2008; Steig et al., 2012), and the complex geometry of ice shelf cavities (Jacobs et al., 2011; De Rydt et al., 2014) have all been found to be important in observational and modeling studies of current and future oceanic melting of the WAIS ice shelves (Joughin et al., 2014). For that reason, using 400m-depth Southern Ocean temperatures as the sole driver for sub-shelf melt may miss important near-Antarctic dynamics. Furthermore, melt water fluxes from the AIS have been found to lead to cooling of surface waters and warming at intermediate depth (Menviel et al., 2010; Weber et al., 2014), a feedback mechanism that could increase ice sheet loss (Golledge et al., 2014, 2019). These processes can only really be captured in fully coupled ocean-atmosphere-ice sheet simulations at high resolution, something that is currently not feasible for the long timescales of late-Quaternary climate evolution. However, it should be possible to run shorter simulations – of e.g., the Last Interglacial or Marine Isotope Stage 11 – using such a setup, and perform a similar set of sensitivity experiments as done here. This would likely reveal additional nonlinearities as ice sheet and forcing are allowed to evolve together. The accumulation forcing for instance is similarly impacted by low climate model resolution and lack of ice-climate feedbacks. Time-evolving changes in orography and albedo can substantially alter atmospheric circulation patterns and associated rainfall (Steig et al., 2000; Maris et al., 2014; Steig et al., 2015)."

2b. Some discussion of eustatic versus relative sea level change is also warranted as relative sea level changes depend on deformational, gravitational, and rotational effects. The experiments are likely sensitive to model parameters used in the bedrock deformation relation, such as the term for the flexural rigidity of the lithosphere. It should be noted that this term has spatial variability in reality, and is quite different between East and West Antarctica. Does the ice sheet model account for this? If a single value is used, different ice sheet sectors in the model may have higher or lower relative sea level change than is realistic. This may increase or decrease the synergistic effects of the combined forcings as well. The solid Earth response has been explored in other ice sheet models with more complex bed deformation schemes (e.g. Kingslake et al., 2018), with quite distinct ice sheet responses to external forcings with different mantle viscosity values. It is not clear if the model accounts for the gravitational or rotational components of sea level change, as in Gomez et al. (2010) and (2013). These latter components should also be discussed because gravitationally-consistent sea level change can stabilize grounding lines during periods of ice sheet retreat. This relates to the authors' conclusion that sea level forcing must be accounted for in ice sheet projections.

In this work, as in many previous comparable studies, we use a standard ELRA (Elastic Lithosphere Relaxing Asthenosphere) model for bed depression and rebound under the varying ice load. Other much more complex and comprehensive full Earth models are available, and have previously been used in shorter time-scale runs (e.g., Gomez et al., 2013, 2015, 2018; Pollard et al., 2017) but coupling with our full-Earth model setup

would be computationally prohibitive for the 400-kyr time scales of the simulations here. These previous studies have addressed the sensitivity of Antarctic results to ELRA vs. full-Earth models in shorter experiments (last deglacial since 20 ka, and future ~3 kyr). There are two feedbacks introduced by the full-Earth models, due to (i) ice-ocean gravitational interaction, and (ii) low-viscosity mantle zone over regions of West Antarctica. Both of these are potentially negative feedbacks for ice retreat, as they cause shallower ocean depths at rapidly retreating grounding lines and thus smaller ice flux from the interior.

However, these studies have found that the last-deglacial and future retreats in Antarctic basins are not strongly affected, at least for standard Earth structures using the full-Earth model compared to ELRA; also a weak low-viscosity mantle zone only has significant effects for very strong forcing such as future business-as-usual emission scenarios, and not for slower past glacial-interglacial forcing (Gomez et al., 2013, 2015; Pollard et al., 2017). Consequently, we suggest that full-Earth coupling would have only small effects on the paleoclimatic results here. However, it should be addressed in future work, when coupling with full-Earth models in our system becomes computationally feasible for near 1-Myr run lengths. Developments to allow 3D variations in Earth structure, and also to allow much longer run lengths, are in progress (Gomez et al., 2018).

We will address this in the paper by adding the following sentences to the last paragraph of Methods Section 2.2.2: “The PSU-ISM uses a standard Elastic Lithosphere Relaxing Asthenosphere model for bed depression and rebound under the varying ice load, and therefore does not include deformational, gravitational, and rotational contributions to local sea level change. Such contributions would potentially act as negative feedbacks for ice retreat, and cause spatial variability between the East and West Antarctic ice sheets (Gomez et al., 2015). However, previous work that includes full-Earth coupling suggest this would likely only have small effects for our timescales (Gomez et al., 2013, 2015; Pollard et al., 2017), and it is currently not computationally feasible to run a full-Earth model for our 400 kyr simulations (though work is in progress to improve this, e.g., Gomez et al., 2018).”

We will also add a sentence to the second paragraph of the discussion: “Such future studies should make sure to include the deformational and gravitational components of future sea level rise through coupling with a full-Earth model (Gomez et al., 2018).”

Specific comments:

1. Page 5, Lines 12-17: Please show the equation for the sub-ice shelf melt parameterization. Based on the reference provided, I assume that it is Eq. 17 in Pollard and DeConto (2012). If not, please clarify. If so, what is the value used for the transfer factor (KT) and what is it based on? Are different K values used in different basins? How sensitive are ice shelf melt/freeze rates to the value of this parameter in relation to the ocean temperature anomalies? Are modelled melt/freeze rates reasonable with present-day climate forcing?

The parameterization of oceanic basal melting under ice shelves uses Eq. 17 in Pollard and DeConto (2012), but has been consolidated with no ad-hoc variations in coefficients (i.e., K=3 everywhere), as described in Pollard et al. (2015, Supplemental Information). As in several other models, the melt rate is proportional (with a single coefficient) to the square of the temperature difference between the base of the ice and the closest grid point at 400 m depth in an ocean dataset. This yields reasonable patterns of modern sub-ice shelf melt, as shown in the Appendix of Pollard et al. (2017), which are generally within ongoing estimates of empirical uncertainties and rapidly changing decadal trends (Depoorter et al., 2013., Rignot et al, 2013). As a design principle, a spatially uniform coefficient is used rather than tuning it point-by-point to match modern estimated maps, because ocean circulation and the associated best-fit coefficients could well change drastically as cavity geometries underneath ice shelves vary over the course of past or future long-term simulations. We will include the equation, parameter values, and a brief discussion of comparison against observations in the manuscript.

2. Page 5, Lines 25-29: It should be noted that the MICI parameterization is still a topic of debate, and it may not be necessary to reproduce Antarctic sea level contributions of past warm periods (see Edwards et al., 2019).

This is true, and we will make a note of this in the paper. More pertinently, the MICI parameterization is not triggered in our simulations, because interglacial climate anomalies simulated by LOVECLIM are only marginally warmer than present-day temperatures (Fig. 3j). We will add the following to the manuscript: “It is worth noting that these parameterizations – whose validity continues to be debated (Edwards et al., 2019) – do not get triggered by our simulated late-Quaternary climate anomalies (Sect. 3.1).”

3. Page 6, Line 24-25: With the caveat that 34.5 psu may not be appropriate for the ocean salinity at the ice-ocean interface.

The value used (34.5 psu) is selected to represent depths of ~400 m – typical depths of Circumpolar Deep Water (CDW) – but would admittedly be lower in the presence of ice melt. The value for the ocean salinity used in the basal melting parameterization only enters in setting the freezing point and has a very small effect on the basal melt rates, especially compared to uncertainties in water temperatures in the ocean dataset. We will make a note of this in the manuscript.

4. Page 7, Line 25: Can the authors clarify the purpose of including the EOF1 plots in Figure 3?

We include the EOF1 plots because in order to retrieve the local amount of change captured by this mode, one needs to multiply the spatial pattern (EOF) with the time series (PC). We will add a sentence in the first paragraph of Sect. 3.1 to clarify to the reader how these figures can be used together to infer spatio-temporal variability in the climate data: “The full amplitude of this first mode at each point in space can be derived

by multiplying the EOF1 map with the PC1 time series." We will add a similar note to the caption of Fig. 3.

5. Page 8, Line 12: The benthic Southern Ocean temperature reconstruction of Elderfield et al. (2012) could also be included as an alternative to the SST reconstruction.

Thank you for this suggestion. Neither the benthic record nor the SST reconstructions will fully capture the processes driving temperature changes at intermediate water depth, but including them both will give a sense of the magnitude of the differences from surface to bottom. We will include the Elderfield reconstruction in Fig. 3l.

6. Page 9, Line 31: Remove the comma before "because"

This will be done.

7. Page 10, Line 1: Remove the comma before "because"

This will be done.

8. Page 12, Lines 15-26: Meltwater fluxes may partly explain the low ocean temperature anomalies in the climate model, but I would also add that some of this may be specific to LOVECLIM. The authors previously mention the model overestimates present-day minimum sea ice extent and that this may contribute to the underestimation of ocean temperatures (Page 7, Lines 3-4). Glacial ocean temperature anomalies are much more negative in CCSM3 along the Antarctic coasts than in LOVECLIM (see Lowry et al., 2018). Other climate models may also show more positive ocean temperature anomalies during interglacial periods than LOVECLIM.

Thank you. We will update this part of the discussion to emphasize that both LOVECLIM itself and the chosen model setup (low-resolution EMIC not coupled to ice-sheet model) contribute to errors in estimating the sub-shelf ocean temperature anomalies.

9. Figure 1: A darker colour for CO2 and obliquity would make the axes easier to read.

This will be done.

10. Figure 3: Why do the temperature/accumulation composites include only East Antarctic ice cores? Please clarify the spatial domain for the PC1 lines in panels j-l.

The temperature and accumulation time series were chosen for their length, to better allow for comparison against the long simulation with the climate model. All chosen EAIS cores have temperature/accumulation data for at least 150 ka. It would be possible to extend the composite by adding records from e.g. WAIS Divide (67 ka) and Siple Dome (57 ka), but we don't think this would assist with comparison of overall

glacial-interglacial behavior, and – when presented in a composite time series – would misleadingly create the appearance that WAIS data were incorporated for all or most of the 408 ka presented in the figure.

The PC1 lines in Figs. 3j-l correspond to the EOF1 maps in Figs. 3g-i. As noted above, we will add some clarification in the manuscript on how the reader can combine these two pieces of information.

11. Figure 7: See above comment for Figure 1.

The color will be updated per the reviewer's suggestion.

12. Figure 8: Move the legend outside the plot so that it is not overlying the mass balance curves.

This will be done.

Reply to Reviewer Comment 2 (Johannes Sutter)

1. Methods and ocean forcing.
While the method section is clearly written, I think that a more transparent discussion of the strengths and weaknesses of the forcing approach would improve the manuscript and help the reader to put the results into perspective with the literature in the field. The fact that a transient model run spanning several hundred thousand years is used to force the ice sheet model is very impressive, but it should be clearly stated that this is at the expense of resolution which is very coarse. It is for example well known that global ocean models have a hard time resolving circumantarctic circulation which mostly leads to inaccurate representations of warming during interglacials and therefore a muted response of the Antarctic Ice Sheet (e.g. Sutter et al., 2016). I would imagine that the representation of variability of circumantarctic ocean temperatures is even worse in EMICs. I guess this is one of the reasons why ice sheet volume remains relatively high for most interglacials in the manuscript presented here, as well as in Tigchelaar et al. (2018). Throughout the manuscript (Methods, Results, Discussion), it should be re-iterated that with the ocean forcing used in this manuscript, the impact of ocean temperatures on transient interglacial ice sheet dynamics in the late Quaternary cannot be accurately assessed. Upon reading the manuscript, I had the impression that the results shown here imply that ocean temperature forcing in Interglacials or deglaciation phases is not important for ice sheet retreat which would contradict the current literature on how the Antarctic Ice Sheet responds to warmer worlds or in glacial terminations. I am sure that this is not the intended take away message but the chance of misinterpretation for someone not familiar with the field is high.

Thank you for this comment. We are well aware that the simulated ocean temperature anomalies in LOVECLIM are likely too low, that certain ocean processes (such as Antarctic Bottom Water formation and sub-shelf circulation) are not well captured in EMICs and GCMs, and that the lack of dynamic coupling between ocean and ice sheet precludes inclusion of important feedback mechanisms (e.g., freshwater forcing from meltwater fluxes). It is therefore by no means our intention to have the reader take away that ocean forcing is not important for AIS evolution. We will be more clear about these methodological shortcomings throughout the manuscript to remedy this.

2. Ice sheet model
The description of the ice sheet model is rather compact, which is probably due to the fact, that it has been discussed at length in the cited literature. However, a quick reference as how ice shelf mass balance is treated would be helpful (calving, basal melt parameterization). Providing an assessment of how well the basal ice shelf melt pattern matches the present day observed melt rate (e.g. Depoorter et al., 2013, Rignot et al., 2013) would be helpful as well. How is the model tuned, and how does it perform against present day and paleo benchmarks? Also it would be worth mentioning how the resolution (40 km) used here could affect the results compared to higher resolution studies.

The ice-sheet model's design involves the use of local parameterizations for important fine-scale processes, that cannot be resolved explicitly by the model grid. In particular, the ice flux across grounding lines is determined by the boundary-layer treatment of Schoof (2007), which captures grounding-line migration well that would otherwise require much finer resolution around the grounding-line zone. Other such features are the sub-grid interpolation of grounding-line location, and fractional areal cover at the edge of ice shelves (Pollard et al., 2012). These features yield model results that are quite independent of horizontal resolution, both in short tests and in long-term runs, for grid sizes between 5 and 40 km (Pollard et al., 2015, Supplemental Information). We will make a mention of this where we describe ice sheet model resolution in Sect. 2.2.

The parameterization of oceanic basal melting under ice shelves uses Eq. 17 in Pollard and DeConto (2012), but has been consolidated with no ad-hoc variations in coefficients (i.e., K=3 everywhere), as described in Pollard et al. (2015, Supplemental Information). As in several other models, the melt rate is proportional (with a single coefficient) to the square of the temperature difference between the base of the ice and the closest grid point at 400 m depth in an ocean dataset. This yields reasonable patterns of modern sub-ice shelf melt, as shown in the Appendix of Pollard et al. (2017), which are generally within ongoing estimates of empirical uncertainties and rapidly changing decadal trends (Depoorter et al., 2013., Rignot et al, 2013). As a design principle, a spatially uniform coefficient is used rather than tuning it point-by-point to match modern estimated maps, because ocean circulation and the associated best-fit coefficients could well change drastically as cavity geometries underneath ice shelves vary over the course of past or future long-term simulations. We will include this equation, parameter values, and a brief discussion of comparison against observations in the manuscript.

As for paleo-benchmarks, previous simulations with this ice sheet model driven by parameterized climates produce realistic expanded grounding line extents and marginal thicknesses at LGM (Mackintosh et al., 2011; Briggs et al., 2013, 2014; Pollard et al., 2016, 2017). The global sea-level fall corresponding to the expanded ice volume is on the low side of the range of estimates (5 to 8 m, Pollard et al., 2016). A greater challenge is achieving good simulations of the last deglacial retreat through time to the present (~20 to 10 ka). This has been a major focus in our and others modeling efforts. We and others have applied large ensembles of model parameter sets, and automated scoring algorithms (Briggs et al., 2013, 2014; Pollard et at 2016, 2017), comparing with several diverse types of paleo data, reviewed for instance in Bentley et al. (2014). This data includes grounding-line positions vs. time, and cosmogenically-derived thinning in inland marginal areas. The general picture is well simulated, including thin streaming ice over much of the major West Antarctic embayments and adjoining ranges (e.g., Stone et al., 2003; Ackert et al., 2007, 2013, Goehring et al., 2019), but with smaller-scale regional disparities (Johnson et al., 2008, 2014, 2017). In general our model captures the LGM state and subsequent deglacial retreat well, to within the general level of uncertainties within the paleo data, and also the modern state of grounded and floating ice (Pollard and DeConto, 2012; Pollard et al., 2016). We will add a brief summary of these comparisons to the Methods section of the paper.

3. Representation of ocean temperatures
The authors mention in section 2.2.2, that LOVECLIM Southern Ocean temperatures are generally too cold. As you use an anomaly forcing to prevent bias propagation it would be interesting how big the glacial and interglacial temperature anomalies (e.g. at 400 m depth) close the ice shelves are.

As can be seen in Figs. 3i & 3l, the dominant EOF (which explains about 86% of variance) has a glacial-interglacial amplitude of about 0.4 ºC. Fig. R1 shows the zonal mean ocean temperature anomaly at 65 ºS. It has a glacial-interglacial amplitude of 0.6 ºC, with only limited warming during interglacials. To our knowledge there are no near-Antarctic paleo-reconstructions of sub-surface ocean temperatures that could be used to validate this amplitude. We find that simulated Southern Ocean SST variability compares well to e.g., the Ho et al. (2012) record from 54 ºS (both an amplitude of ~8ºC), but LOVECLIM deep-water temperature anomalies are much lower than those reconstructed from the Elderfield et al. (2012) record. We will add these values to our discussion of simulated ocean temperatures in Sect. 3.1.

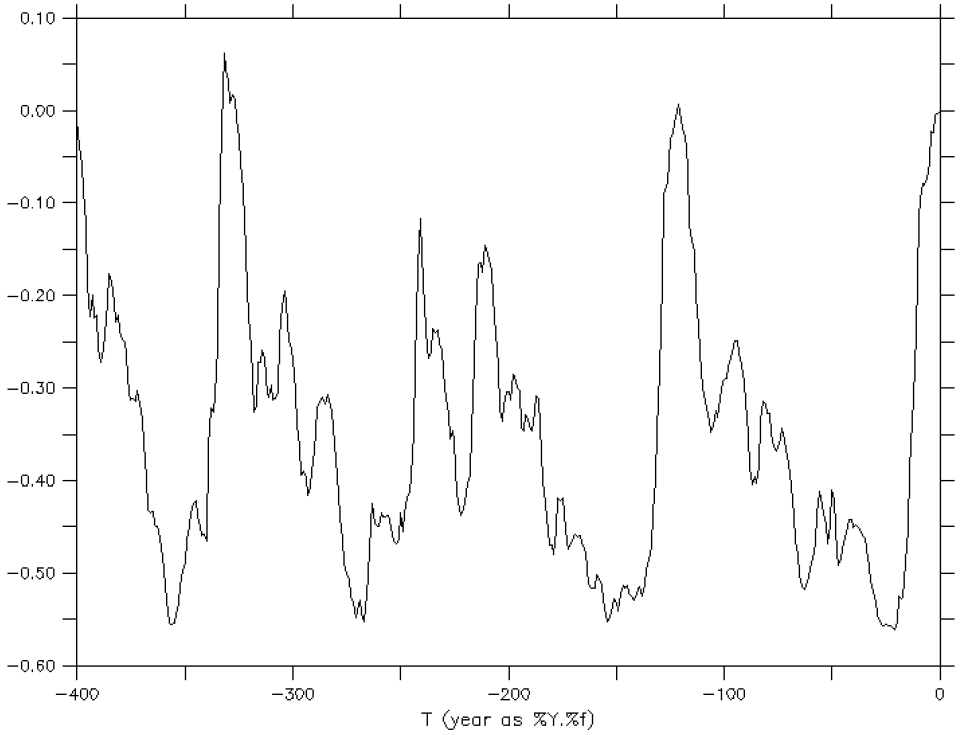

Figure R1 – Zonal mean 400m ocean temperature anomaly at 65 ºS

4. Presentation of Results
The presentation and discussion of the results is currently written in a very affirmative manner which sometimes ignores the biases introduced by the experimental setup. For example 3.4.4 suggests that temporal ocean temperature changes are not relevant for ice volume changes. While this is true for the setup used here, it is not the case in multiple publications on the matter (e.g. Golledge et al., 2015, 2017, 2019, DeConto & Pollard 2016, Sutter et al. 2016,2019, Albrecht et al., 2019 TCD). The authors state that they will discuss the validity of the results in the Discussion (p.11 L 10-11), but I have

the impression that a serious debate about the shortcomings of the approach and therefore the scope of the results is lacking.

Based on this comment, we will make sure to be more explicit about methodological shortcomings in Sects. 3.1, 3.4.3, 3.4.4, and the Discussion. Specific responses and suggested edits are given below.

Specific comments:

Title: as the authors focus on the AIS evolution during the last 408 ka I would rename the title to "Nonlinear response of the Antarctic ice sheet to ***late-****Quaternary sea level and climate forcing" and use late-Quaternary instead of Quaternary throughout the manuscript (already done in the header of section 3.1).*

This is a good suggestion. We will update the title as suggested, and change occurrences of Quaternary to late-Quaternary in the entire manuscript.

Check Antarctic Ice Sheet (AIS) throughout the manuscript. Usually it is written in capital initial letters. Furthermore, while you introduce the abbreviation on page 1 L 17 you mostly don't use it later on.

We will capitalize all instance of Antarctic Ice Sheet, and use the abbreviation AIS where appropriate.

P2 L10: The importance for what?

For ice sheet stability. We will add this to this sentence.

P2 L18-19: I think this is mostly true for glacials but less so for interglacials. While e.g. Konrad et al. (2014) show that a sea level drop due to changes in the gravitational pull during ice loss in interglacials can stabilize the grounding line, the overall rise in sea level during interglacials doesn't play a large role in grounding line retreat as it is mostly limited to just a few meters.

That is correct, but refers particularly to ice sheet evolution 'beyond' present-day sea level. That is not at odds though with the statement that sea level is an important pacemaker for late-Quaternary ice sheet evolution, because that refers to the entire range of ice sheet configurations in between 'full' glacial and interglacial.

P2 L22: maybe rephrase to: "particularly leading to a growth of the East Antarctic Ice Sheet (EAIS)"

This will be changed.

P3 L11-14: inconsistent use of Section versus Sect.

This will be changed so that ‘Section’ if used at the beginning of the sentence, and ‘Sect.’ elsewhere.

P4 L1: maybe rephrase to “Each land grid cell ...”

This will be done.

P4 L19: rephrase to “While the climate model run closely follows ..., here the longwave radiative effect of CO2 was amplified ...”

This will be done.

P4 L34: is the duration of the experiments the reason for the 40km resolution? Then rephrase: Due to limited computational resources and long timescale of the simulations we had to use a relatively coarse resolution of 40 km.

Yes, the reason for this choice of resolution is mostly that we are running multiple simulations of long duration, but also that – as mentioned above – the results are not substantially different if a resolution of e.g., 20 km is used. We will make the suggested edit.

P5 L1: rephrase to “Present day climate forcing is obtained from the [...] interpolated to the ice model grid.”

This will be done.

Do you use the ALBMAP v1 bedrock topography or BEDMAP2 for the initial ice sheet configuration?
For the initial ice sheet configuration we use the BEDMAP2 bedrock topography, as in Pollard. et al. (2015). We will clarify this in Sect. 2.2.

P5 L12-30: I would expect the description of the parameterization of the basal shelf melt calculation and calving to be in the ice sheet model section and not in the climate forcing section.

Because the climate forcing and mass balance calculations are so closely related, it seems more sensible to keep these together. We will rename this section to “Present-day mass balance and climate forcing.”

P6 L6-8: Here I do not understand whether the climate forcing “jumps” every thousand years to a new set of climate anomalies (i.e. the ISM is forced with the same climate anomalies for 1000 years) or whether the transition is smooth. Please clarify.

It is the former – ice sheet model forced with the same anomalies for 1000 years – though during these 1000 years the climate forcing within the ice sheet model will still evolve in response to simulated changes in elevation. We will clarify this sentence by

changing it to: “The climate forcing in the ice sheet model is updated every 1000 calendar years. Climate anomalies are calculated with respect to the LOVECLIM climatology over the last 200 model years (representing 1000 calendar years) and are bilinearly interpolated to the ice sheet model grid.”

P6 L 8-10: maybe rephrase to:
The atmospheric temperature Ta is modified by a lapse rate correction of γ=0.008°Cm-1 to account for surface elevation differences between the reference ice sheet geometry (zobs; Le Brocq et al., 2010) and both the simulated elevation at time t (z(t)), as well as for differences with respect to the LOVECLIM orography (zLC).

Thank you for the suggestion, we will make this change.

P7 L1-2: As you force the ISM with ocean temperature anomalies I guess the glacial-interglacial variability is more relevant for the ice sheet’s evolution than the present day bias. Please add a sentence which quantifies the ocean warming e.g. in MIS5e and MIS11 and the cooling e.g. during the LGM relative to the PI control climate state (LOVECLIM 1000 year average?).

Per main comment 3 above, we will add a sentence quantifying glacial-interglacial temperature anomalies to Sect. 3.1, where we discuss late-Quaternary climate evolution. We think this fits better there than in Sect. 2.2.2, which specifically discusses the modeling setup.

P7 L 11-12: remove sentence “The bottom half of Fig. 2 ...”

This will be done.

P7 L26: it would help the reader if CO2 is plotted in Fig. 3j as well to make the pacing more evident.

Thank you for the suggestion. This will be done.

P7 L31: Maybe I overlooked this but how do you create the ice core composite? If I understand it correctly you use Dome Fuji, EDC, Vostok, TALDICE and EDML. Only Dome Fuji, EDC, Vostok cover the whole 408 ka.

The composite was simply constructed as the average of all available ice core temperature/accumulation records for each time in the past, as done in Parrenin et al. (2013), Supplemental Information. We will clarify this in the manuscript.

P8 L8: Again, you use only one coastal ice core (TALDICE) and 4 interior ice cores. For the latter, the lapse rate correction would be stronger in glacials (e.g. Pollard et al. 2009 and Sutter et al. 2019 TCD Fig. 10). But the biggest discrepancies shown in 3l and 3j occur in Interglacials with too cold ocean and surface temperatures.

As the two time series are on different scales the discrepancies still exist during glacials, but the reviewer is correct in observing that this does not explain the underestimation of interglacial warming by the climate model. We will change this sentence to “This could partially be due to the fact that the LOVECLIM simulation does not include the lapse rate response to the evolving ice sheet height, but also points at an underestimation of polar temperature change in the climate model, especially during interglacials (Tigchelaar et al. 2018).”

P8 L17-18: I could imagine that the underestimation of ocean temperature variability in interglacials is the main reason why the ocean forcing is the weakest driver of interglacial ice volume changes in your simulations. This has important implications for your conclusions as this is a methodological bias and not necessarily reflects the actual response of the AIS e.g. in MIS5e and MIS11.

Per main comment 3 above, we will add a few sentences to this paragraph quantifying the glacial-interglacial temperature anomalies to show these are low compared to reconstructed temperature changes. We will also point out that these modeled anomalies are much lower than the thresholds found to be necessary to simulate ice sheet collapse.

P8, L23-26: Actually MIS7 shows the lowest surface temperature warming in Antarctic ice cores, how come that for this period the AIS volume is higher than in the other interglacials in your simulations?

I’m assuming the reviewer means to ask why the AIS volume is lower during this period. As shown in Fig. 3j, MIS7 also has low annual mean surface temperature warming in our climate model simulation. In Tigchelaar et al. (2018) we explained that the reason for the ice sheet loss around 210 ka is the strong summer insolation forcing during this time (Fig. 1a) that drives high summer surface melt and subsequent ice sheet retreat. This actually does not occur at MIS7 (which is a Northern Hemisphere driven termination, Termination III), but during Termination III-a, and illustrates the north-south asynchronicity that can be caused by precessional forcing.

P9, L12: rephrase to: Figure 5 shows where the individual forcing components have the largest effect on the Antarctic Ice Sheet.

This will be done.

P9, L14-15: what is the reason for this thickening? Reduced surface melt? Retreat caused by hydrofracturing?

As shown in Figs. 4c and 5c, there is a small increase in floating ice volume as well as a small outward expansion of the grounding line during glacial times, so retreat is not the cause for this thickening. Rather, as can be seen in Fig. 8 and is described in Sect. 3.4.2, glacial thickening is primarily caused by a cooling-driven reduction in surface melt and calving rates.

P9, L16: again I expect this to be caused by the forcing setup and that it is not representative during Interglacials.

We will add “Due to the small magnitude of the simulated ocean temperature change” to the beginning of this sentence.

P9, L20: maybe rephrase to: Combined forcing leads to a more pronounced grounding line advance during glacials than in simulations with single forcing.

We will change this to “Combinations of external forcings lead to a more pronounced grounding line advance during glacials than in simulations with one single forcing.”

P9, L27: rephrase to: Figure 7 depicts the response of grounded ice volume to the respective forcing in the different sensitivity runs.

This will be done.

P9, L27: rephrase to: It is important to note that the impact of the sea level forcing in isolation leads to the conversion of grounded into floating ice (during Terminations???).

We will change this sentence to “As noted before, the impact of sea level forcing in isolation is to convert grounded into floating ice during periods of sea level rise, and the other way around during sea level drops.”

P9, L30: Do surface melt rates really increase in glacials?? The maximum elevation change of ice shelves during glacials would be ca. 120 m.

Yes, as can be seen in Fig. 8b, surface melt rates increase in this simulation where only sea level is forced to change. The elevation change of 120 m may seem small, but many areas on the perimeter of the AIS have present-day seasonal maximum temperatures that are close to freezing. Only a small increase in annual mean temperature will therefore be needed to generate more Positive Degree Days in the surface melt scheme.

P10, L5: quantify “fairly consistently”.

We agree this is ambiguous wording and will remove it from this sentence.

P10, L6: Wording. (Now, ...)

We will replace ‘Now,’ with ‘In this case’.

P10, section 3.4.3. This section needs to be expanded, discussing the reasons why the ocean forcing plays a negligible role in the simulations (see main remarks).

We will add the following sentence to this paragraph: “It is important to note here that this small response to ocean temperature forcing is more likely a function of the low amplitude of the LOVECLIM-simulated ocean temperature forcing than it is indicative of low sensitivity of the AIS to changing ocean conditions, as will be discussed further below.”

P10, L23: rephrase to: Our sensitivity runs show that the simulated response of the AIS to late Quaternary external drivers [...]

This will be done.

Section 3.4.4. and the Discussion requires a more detailed disentanglement of what the authors deem to be realistic responses of the AIS to late Quaternary climate and boundary conditions and what they think is due to methodological biases.

In Sect. 3.4.4. we will change the last sentence to: “Kusahara et al. (2015) also found oceanic melt rates to increase during the Last Glacial Maximum in response to grounding line migration, lending support to these findings. However, as shown in Tigchelaar et al. (2018), the low sensitivity of the modeled AIS to interglacial ocean conditions is likely a result of the low amplitude and resolution of the LOVECLIM ocean temperature forcing and lack of ice-ocean feedbacks in our modeling setup.”

In the discussion, we will replace the existing discussion of ocean temperature forcing with the following paragraphs:
“Our finding that ocean temperature forcing plays a limited role in driving changes in Antarctic ice volume contrasts with previous modeling studies of past and future AIS evolution (Golledge et al., 2015; DeConto and Pollard, 2016; Sutter et al., 2016), as well as observations of sustained sub-shelf ice loss in response to ongoing ocean warming at e.g., Pine Island Glacier (Jacobs et al., 2011; Pritchard et al., 2012). This is not surprising, given that the LOVECLIM-simulated ocean temperature anomalies are small (Figs. 3i,l), and ice sheet models typically need ocean warming of 2-5 ºC to initiate interglacial WAIS collapse (Pollard and DeConto, 2009; DeConto and Pollard, 2016; Sutter et al., 2016; Tigchelaar et al., 2018). Absent paleo-reconstructions of near-Antarctic sub-surface ocean temperatures it is difficult to assess how realistic our LOVECLIM simulation is, though critical processes such as Antarctic Bottom Water formation are known to be poorly represented in low-resolution climate models (e.g., Snow et al., 2015), and previous studies have found LOVECLIM in particular to have more muted late-Quaternary temperature variability than other models (Lowry et al., 2019).
In addition, many regional oceanographic processes can affect the circum-Antarctic ocean environment beyond large-scale climate forcing. For example, the blocking effects of sea ice formation (Hellmer et al., 2012), the role of winds in pushing warm waters onto the continental shelf (Thoma et al., 2008; Steig et al., 2012), and the complex geometry of ice shelf cavities (Jacobs et al., 2011; De Rydt et al., 2014) have all been found to be important in observational and modeling studies of current and future oceanic melting of the WAIS ice shelves (Joughin et al., 2014). For that reason,

using 400m-depth Southern Ocean temperatures as the sole driver for sub-shelf melt may miss important near-Antarctic dynamics. Furthermore, melt water fluxes from the AIS have been found to lead to cooling of surface waters and warming at intermediate depth (Menviel et al., 2010; Weber et al., 2014), a feedback mechanism that could increase ice sheet loss (Golledge et al., 2014, 2019). These processes can only really be captured in fully coupled ocean-atmosphere-ice sheet simulations at high resolution, something that is currently not feasible for the long timescales of late-Quaternary climate evolution. However, it should be possible to run shorter simulations – of e.g., the Last Interglacial or Marine Isotope Stage 11 – using such a setup, and perform a similar set of sensitivity experiments as done here. This would likely reveal additional nonlinearities as ice sheet and forcing are allowed to evolve together. The accumulation forcing for instance is similarly impacted by low climate model resolution and lack of ice-climate feedbacks. Time-evolving changes in orography and albedo can substantially alter atmospheric circulation patterns and associated rainfall (Steig et al., 2000; Maris et al., 2014; Steig et al., 2015)."

P11, L30: I do not understand this sentence. Increased ice loss due to sea level rise induced warming? This needs to more explicit, warming due to sea level driven ice sheet retreat and therefore surface lowering?

We simply meant to suggest that future studies should not only include greenhouse gas-induced warming, but also include ice sheet retreat due to rising sea levels (from either the Greenland or Antarctic Ice Sheet). To clarify, we will change this sentence to "Further research should therefore explore whether, given the current configuration of grounding line and bedrock, rising sea levels as a result of global warming would further increase or stabilize ice loss.

P12, L7-8: What is meant by "manually offset"?

What we meant to suggest here was a series of ice sheet model sensitivity experiments in which the phase relationship between sea level and climate forcing is artificially changed to capture the uncertainties in sea level and greenhouse gas dating. To be more clear in the manuscript, we will change this sentence to "… though future sensitivity runs with the ice sheet model could include artificial shifts in the phase relationship between the sea level and climate forcing to explore associated nonlinearities."

P12, L10: rephrase, e.g. : In particular, as the ice sheet grows ice sheet areas with higher precipitation expand leading to a positive feedback while at the same time, the ice margin advances into warmer ocean waters which leads to a negative feedback.

We will change this sentence to: "In particular, as the ice sheet grows the ice sheet expands into areas of higher precipitation, leading to a positive feedback, while at the same time the ice margin advances into warmer ocean waters, causing a negative feedback."

*P12, L15-18: This is the only place in the manuscript which states that LOVECLIM ocean temperature variability is too low and that this could be causal to the muted response during interglacials. Unfortunately, this sentence is right away relativized in the next sentence, citing **one** publication, while a wealth of publications identified ocean warming to be the main driver of ice loss in late Quaternary interglacials (e.g. Golledge et al., 2015,2017,2019, DeConto & Pollard 2016, Sutter et al., 2016).*

As detailed above, we will expand on this discussion in the revised manuscript to better highlight the shortcomings of our methodology. However, it should be noted that in these Golledge, DeConto and Sutter studies, ocean warming is a main driver of ice loss because an ocean forcing is chosen that leads to ice loss, not because high-resolution ocean simulations, circumantarctic temperature reconstructions or a coupled modeling setup dictate a precise level of ocean forcing.

P12, L28-29: replace sentence "Previous modeling studies have failed to elucidate how these different external drivers interact in driving large glacial ice sheet growth and interglacial sea level highstands." E.g. With "In contrast to previous studies, here we focus on the interaction of different external forcings driving Antarctic Ice Sheet changes". There are previous studies who discuss individual forcing components (e.g. Pollard et al. 2009, de Boer et al. 2013), just not as comprehensive as done here.

We will make the suggested change.

P12, L 33-35: I am not fully convinced that this is the case, or at least that this study shows that, as the effect of ocean temperature changes in interglacial ice sheet retreat are not adequately captured in the simulations presented here.

In the revised manuscript we will change this concluding sentence to further highlight the uncertainty around ocean forcing: "Our modeling setup likely underestimates the role of oceanic forcing, which remains largely unbound by the geologic record and needs to be further explored in a coupled climate-ice sheet modeling framework that can account for critical circumantarctic oceanographic processes."

Figures 1,4,5,6,8: move labels a),b),c) out of the figures panels.

This will be done.

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Nonlinear response of the Antarctic ~~ice sheet~~ Ice Sheet to ~~Quaternary~~ late-Quaternary sea level and climate forcing

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Abstract. Antarctic ice volume has varied substantially during the ~~Quaternary~~late-Quaternary, with reconstructions suggesting a glacial ice sheet extending to the continental shelf break, and interglacial sea level highstands of several meters. Throughout this period, changes in the Antarctic ~~ice sheet~~ Ice Sheet were driven by changes in atmospheric and oceanic conditions and global sea level, yet so far, modeling studies have not addressed which of these environmental forcings dominate, and how they interact in the dynamical ice sheet response. Here we force an Antarctic ~~ice sheet~~ Ice Sheet model with global sea level reconstructions and transient, spatially explicit boundary conditions from a 408 ka climate model simulation, not only in concert with each other but, for the first time, also separately. We find that together, these forcings drive glacial-interglacial ice volume changes of 12-14 m SLE, in line with reconstructions and previous modeling studies. None of the individual drivers – atmospheric temperature and precipitation, ocean temperatures, sea level – single-handedly explains the full ice sheet response. In fact, the sum of the individual ice volume changes amounts to less than half of the full ice volume response, indicating the existence of strong nonlinearities and forcing synergy. Both sea level and atmospheric forcing are necessary to create full glacial ice sheet growth, whereas the contribution of ocean melt changes is found to be more a function of ice sheet geometry than climatic change. Our results highlight the importance of accurately representing the relative timing of forcings of past ice sheet simulations, and underscore the need for developing coupled climate-ice sheet modeling frameworks that properly capture key feedbacks.

1 Introduction

~~Though mass loss~~ At a time when the future of the Antarctic ~~ice sheet~~ Ice Sheet (AIS) ~~has accelerated in recent decades (IMBIE team, 2018), future melt rates in a warming climate remain~~ is both critical and highly uncertain (Joughin and Alley, 2011; Church et al., 2013; DeConto and Pollard, 2016), exploring its past behavior can lend insight to its sensitivity to external

forcing. Records show that during the ~~Quaternary (i.e., the past 2.6~~ late-Quaternary (roughly the past one million years), the AIS contributed to both glacial sea level drops of more than 10 m (RAISED Consortium et al., 2014), as well as rapid deglacial sea level rise (Carlson and Clark, 2012) and interglacial sea level highstands of several meters (Dutton et al., 2015). Throughout this period, Antarctic mass balance changes were driven by a wide spectrum of external forcings~~—~~: changes in atmospheric temperatures, accumulation rates, oceanic conditions, and sea level (Tigchelaar et al., 2018)~~— making the Quaternary an interesting test-bed for constraining future AIS behavior~~. There is substantial spatial variability in the sensitivity of the AIS to these forcings, as indicated for example by surface-exposure chronologies since the Last Glacial Maximum (LGM) (e.g., RAISED Consortium et al., 2014; Hillenbrand et al., 2014; Spector et al., 2017; Goehring et al., 2019). So far however, the relative contributions of ~~these~~ different external drivers of past AIS variability and their synergies have not been quantified in modeling studies. Here we address how different forcing agents interact during the last four glacial cycles using a set of experiments with an ~~Antarctic ice sheet~~ AIS model.

Unlike the Greenland ~~ice sheet, the Antarctic ice sheet~~ Ice Sheet, the AIS has large marine-based margins. The ice shelves surrounding the AIS have a buttressing effect, and therefore play an important role in determining its stability. Disintegration of ice shelves can lead to rapid discharge from and acceleration of the grounded ice sheet, in particular when the bed deepens towards the ice sheet interior (a process referred to as 'marine ice sheet instability') (Schoof, 2007; Joughin and Alley, 2011). The importance of the Antarctic marine margins ~~mean~~ for ice sheet stability means that both the marine and the atmospheric environment contribute to ice volume changes. The accelerated mass loss of Pine Island Glacier over the past few decades for instance, has been attributed to enhanced sub-shelf melting in response to warming oceans and changing ocean circulation (Jacobs et al., 2011; Pritchard et al., 2012). The 2002 collapse of the Larsen B ice shelf on the other hand, is thought to be the result of preconditioning by a warming atmosphere (van den Broeke, 2005). During the glacial cycles of the ~~Late Quaternary~~late-Quaternary, changes in eustatic sea level (Fig. 1C) further impacted Antarctic ice shelves: changes in the ice flux at the grounding line turn grounded ice into floating ice during sea level rise, and floating ice into grounded ice during sea level drops (Schoof, 2007). Previous modeling studies of ~~Quaternary~~ late-Quaternary AIS evolution have identified global sea level as an important pacemaker, especially for the West Antarctic ~~ice sheet~~ Ice Sheet (WAIS) (Ritz et al., 2001; Huybrechts, 2002; Pollard and DeConto, 2009). Finally, changes in temperature and circulation patterns drive changes in accumulation rates that can affect both the marine margins and interior ice sheet. Future projections of AIS evolution suggest that in a warming world, accumulation rates will increase as a result of increased atmospheric moisture content, particularly leading to growth of ~~in particular~~ the East Antarctic ~~ice sheet~~ Ice Sheet (EAIS) (Huybrechts et al., 2004; Frieler et al., 2015; Medley and Thomas, 2019).

~~Quaternary~~ Late-Quaternary climate change is ultimately caused by variations in earth's axial tilt and orbit around the sun (Milankovitch, 1941), i.e., precession, eccentricity and obliquity (Fig. 1A,B). These lead to changes in incoming solar radiation that cause a global climate and carbon cycle response that make changes in atmospheric greenhouse gas concentrations – primarily CO_2 (Fig. 1C) – an additional driver of long-term climate variability (Shackleton, 2000). Different climate variables respond differently to each of these forcings, resulting in a rich spectrum of Southern Hemisphere climate variability in both reconstructions (e.g., Steig et al., 2000; Gersonde et al., 2005; Cortese et al., 2007; Jouzel et al., 2007; Ho et al., 2012) and

simulations (e.g., Huybers and Denton, 2008; Menviel et al., 2008; Timmermann et al., 2009; He et al., 2013; Timmermann et al., 2014). ~~Until now however~~

Up to this point, Antarctic modeling studies have not considered how these various forcings interact in driving ice volume changes. Previous studies have either used heavily parameterized climate forcing (Ritz et al., 2001; Huybrechts, 2002; Pollard and DeConto, 2009) or simplified climate and ice sheet configurations (de Boer et al., 2013; Stap et al., 2014); have focused on equilibrium simulations of specific time periods (Golledge et al., 2012); or applied indexed interpolations of extreme climate states (Maris et al., 2015). All of these studies assume that Southern Hemisphere climate variables vary in pace with either Antarctic temperature reconstructions (Petit et al., 1999) or the benthic oxygen isotope record (Lisiecki and Raymo, 2005). ~~These modeling~~ They studies thus ignore the spatial and temporal heterogeneity of ~~Quaternary~~ late-Quaternary climate variability, and preclude a better understanding of how different drivers interact.

The aim of this study is to better understand the individual and combined roles of sea level and climate variability in driving ~~Antarctic ice sheet~~ AIS evolution during the ~~Late Quaternary~~late-Quaternary. To that end we have forced a state-of-the-art ~~Antarctic ice sheet~~ AIS model with spatially-varying and time-evolving atmospheric temperature, precipitation and ocean temperature fields from a climate model simulation over the last four glacial cycles, as well as changes in eustatic sea level from Northern Hemisphere ice sheets. This work builds on Tigchelaar et al. (2018), which used a similar modeling setup but did not isolate individual drivers. We conduct a number of sensitivity experiments to explore the separate role and synergy of individual forcings and mechanisms contributing to past ice sheet variability. Looking at individual forcings allows us to identify which are important, which need modeling improvement, and how they might interact nonlinearly in future Antarctic change. These simulations can also be used to aid in interpretation of the rich AIS deglaciation record (RAISED Consortium et al., 2014; Hillenbrand et al., 2014; Spector et al., 2017; Goehring et al., 2019).

Section 2 provides a detailed overview of our climate and ice sheet modeling setup. In Sect. 3 the main results are presented, with Sect. 3.1 discussing ~~Quaternary~~ late-Quaternary climate variability, Sect. 3.2 and 3.3 describing the ice sheet response to all and individual drivers, and Sect. 3.4 discussing the responsible mechanisms. ~~Sect.~~Section 4 summarizes our results and discusses their implications.

2 Methods

The ~~Late Quaternary~~ late-Quaternary orbital and greenhouse gas forcing shown in Fig. 1 is used to drive a transient simulation with an Earth system model of intermediate complexity (EMIC) over the last four glacial cycles (Sect. 2.1). Climate anomalies from this simulation, together with time-varying global sea level (Fig. 1c), are then used as boundary conditions for various sensitivity experiments (Sect. 2.2.3) with the Penn State University ice sheet model (PSU-ISM; Sect. 2.2.1) according to the equations outlined in Sect. 2.2.2. Figure 2 shows a schematic illustration of this modeling setup.

2.1 Climate model

Our ice sheet model experiments are driven with transient climate anomalies spanning the last four glacial cycles (408 ka to present) (Timmermann et al., 2014; Friedrich et al., 2016; Timmermann and Friedrich, 2016), derived from a simulation with the EMIC LOVECLIM (Goosse et al., 2010), which consists of coupled atmospheric, ocean-sea ice and vegetation components. The atmospheric component of LOVECLIM, ECBILT, is a spectral T21 ($\sim$5.625°$\times$5.625°), three-level model based on the quasi-geostrophic equations, extended by estimates of the ageostrophic terms (Opsteegh et al., 1998). The model contains a full hydrological cycle and includes physical parameterizations of diabatic processes (radiative fluxes, sensible and latent heat fluxes) in the thermodynamic equation.

CLIO, the ocean sea-ice component, is a 3°$\times$3° primitive equation ocean general circulation model with twenty vertical levels, coupled to a thermodynamic-dynamic sea ice model (Goosse and Fichefet, 1999). It uses parameterizations to compute mixing along isopycnals, the effect of mesoscale eddies on diapycnal transport and downsloping currents at the bottom of continental shelves. Finally VECODE is a terrestrial vegetation model that consists of two plant functional types and non-vegetated desert zones (Brovkin et al., 1997). Each land grid cell is assumed to be partially covered by these three land cover types, based on annual mean temperature and rainfall amount and variability.

For the transient climate model simulation, LOVECLIM was forced with time-evolving orbital parameters (Berger, 1978) and reconstructed atmospheric greenhouse gas concentrations (CO_2, CH_4 and N_2O) (Lüthi et al., 2008). The corresponding orbital forcing, annual mean and seasonal insolation changes and CO_2 time series are shown in Fig. 1. In addition, Northern Hemisphere ice sheet conditions were obtained from a transient experiment conducted with the Climate and Biosphere Model, version 2 (CLIMBER-2), coupled to the Northern Hemisphere Simulation Code for Polythermal Ice Sheets (SICOPOLIS) ice sheet model (Ganopolski and Calov, 2011). Orography, albedo and ice mask variations from this simulation are interpolated onto the LOVECLIM grid, where in the presence of land ice, the grid point albedo is set to 0.7 and the vegetation mask is modified. The orography, albedo and ice mask of the ~~Antarctic ice sheet~~ AIS remain constant throughout the simulation. Similarly, time-evolving Antarctic melt water fluxes are not fed back into LOVECLIM. The implications of this lack of ice sheet-climate coupling will be explored in the Discussion.

The orbital, greenhouse gas and ice sheet conditions are applied with a boundary acceleration factor of five (Timm and Timmermann, 2007; Timmermann et al., 2014). The acceleration technique is based on the assumption of relatively fast equilibration of surface variables to slow external drivers; it thus mostly affects the representation of deep ocean currents (Timm and Timmermann, 2007), but not of surface and thermocline processes that matter for our experiments. This means that 200 model years correspond to 1000 calendar years. The LOVECLIM simulation is conducted using ~~Last Glacial Maximum (LGM)~~ LGM ocean bathymetry (Roche et al., 2007) in order to avoid the internally generated Atlantic meridional overturning circulation oscillations described in Friedrich et al. (2010). While the climate model run ~~follows closely~~ closely follows the methodology of Timmermann et al. (2014), ~~in the current simulation the~~ here the longwave radiative effect of CO_2 was amplified by a factor of 1.97, based on model-proxy comparisons using 63 globally-distributed SST-reconstructions (Friedrich et al., 2016). The resulting net climate sensitivity amounts to $\sim$4 °C per CO_2 doubling (Timmermann and Friedrich, 2016) and yields a more

realistic glacial-interglacial amplitude in surface temperatures compared to paleo-proxy data. ~~Our climate modeling strategy is illustrated in the top half of Fig. 2.~~

2.2 Ice sheet model

The 408 ka climate anomalies from LOVECLIM are used to force a number of sensitivity experiments with the PSU-ISM (Fig. 2; Pollard and DeConto, 2009, 2012a; DeConto and Pollard, 2016; Pollard et al., 2016). ~~This model~~ Previous simulations with this ice sheet model driven by parameterized climates produce a realistic LGM state and subsequent deglacial retreat (Mackintosh et al., 2011; Briggs et al., 2013, 2014; Pollard et al., 2016, 2017), to within general levels of uncertainty within the paleo-data, and also the modern state of grounded and floating ice (Pollard and DeConto, 2012a, b; Pollard et al., 2016).

The model is based on a combination of the scaled shallow ice and shallow shelf approximations, and calculates ice velocity across the grounding line using an ice flux parameterization (Schoof, 2007). Basal sliding on unfrozen beds is calculated using a standard drag law, with the basal sliding coefficients derived from a simple inverse method (Pollard and DeConto, 2012b). Bedrock deformation is modeled as an elastic lithospheric plate above local isostatic relaxation; the ~~equilibrium bedrock~~ initial and equilibrium bedrock topography and ice-load state is taken to be modern observed (Bedmap2; Fretwell et al., 2013). The model includes vertical diffusion of heat and storage in bedrock below the ice, which is heated from below by a uniform geothermal heat flux for the EAIS and WAIS (Pollard and DeConto, 2012a).

The model is discretized on a polar stereographic grid~~, which for our long sensitivity experiments has a~~. Due to limited computational resources and the long timescale of the simulations we used a relatively coarse resolution of 40 km~~resolution~~, though note that previous studies with this model have shown that results are quite independent of horizontal resolution (Pollard et al., 2015, Supplemental Information).

2.2.1 Present-day mass balance and climate forcing

~~For modern climate , surface input fields are~~ Present-day surface climate forcing – specifically, annual mean atmospheric temperature $T_{\mathrm{a}}^{\mathrm{obs}}$ (Fig. 3a; van de Berg et al., 2006) and accumulation P^{obs} (Fig. 3b; Comiso, 2000) – is obtained from the ALBMAP v1 database at 5 km resolution (Le Brocq et al., 2010) ~~. First, annual mean atmospheric temperature $T_{\mathrm{a}}^{\mathrm{obs}}$ (Fig. 3a; van de Berg et and accumulation P^{obs} (Fig. 3b; Comiso, 2000) are interpolated onto~~ and interpolated to the ice model grid. ~~Then a~~ A lapse rate correction of $\gamma = 0.008\ ^{\circ}\mathrm{C\,m^{-1}}$ is applied to the atmospheric temperature to correct for differences between observed (z^{obs}; Le Brocq et al., 2010) and model (z) surface elevation. The seasonal cycle in atmospheric temperature is parameterized as a sinusoidal cycle with a range of 20 °C at sea level, increasing linearly with elevation to 30 °C at 3000 m and above (Pollard and DeConto, 2012a), giving $T_{\mathrm{a}}^{\mathrm{obs}}(\tau)$. Surface melt rates are calculated using a positive degree-day (PDD) scheme (Reeh, 1991) that uses different coefficients for ice ($8\ \mathrm{kg\,m^{-2}\,^{\circ}C^{-1}}$) and snow ($3\ \mathrm{kg\,m^{-2}\,^{\circ}C^{-1}}$) and allows for seasonal refreezing as well as diurnal and synoptic variability (Pollard and DeConto, 2012a). Present-day accumulation rates in the model do not contain a seasonal cycle, but are split into rain and snow based on monthly temperatures. In the ocean, the model interpolates modern annual mean 400 m-depth ocean temperatures $T_{\mathrm{o}}^{\mathrm{obs}}$ from the World Ocean Atlas (Locarnini et al., 2010) onto the ice

sheet model grid (Fig. 3c). In areas outside the range of the Locarnini et al. (2010) dataset, ocean temperatures are propagated underneath the ice shelves using a nearest neighbor interpolation.

An important component of AIS modeling is the treatment of ice shelf processes and the ice-ocean interface. While in reality melting at the ice shelf-ocean interface is a function of ocean temperature, salinity and circulation in the ice shelf cavity (Jacobs et al., 1992), most ice sheet models used for long-term simulations make use of parameterizations based on sub-surface ocean temperatures alone. This ice model follows the parameterization developed by Martin et al. (2011) for the PISM-PIK model, where oceanic melt is a function of the difference between ocean temperature and the depth-varying freezing temperature of ocean water~~. Unlike in Martin et al. (2011), the melt rate~~ , with a quadratic dependency on this temperature difference ~~is quadratic (Holland et al., 2008; Pollard and DeConto, 2012a). For modern conditions, the model interpolates annual mean 400 -depth ocean temperatures T_o^{obs} from the World Ocean Atlas (Locarnini et al., 2010) onto the ice sheet model grid (Fig. 3c). In areas outside the range of the Locarnini et al. (2010) dataset, ocean temperatures are propagated underneath the~~ (Holland et al., 2008; Pollard and DeConto, 2012a):

$$\text{OMB} = \frac{K K_T \rho_w c_w}{\rho_i L_f} |T_o - T_f| (T_o - T_f) \tag{1}$$

Here K_T is the transfer coefficient for sub-ice oceanic melting (15.77 $\mathrm{m\,a^{-1}\,K^{-1}}$), ρ_w is the ocean water density (1028 $\mathrm{kg\,m^{-3}}$), ρ_i is the ice ~~shelves using a nearest neighbor interpolation. In addition,~~ density (910 $\mathrm{kg\,m^{-3}}$), c_w is the specific heat of ocean water (4218 $\mathrm{J\,kg^{-1}\,K^{-1}}$), and L_f is the latent heat of fusion (0.335×10^6 $\mathrm{J\,kg^{-1}}$). T_o is the specified ocean temperature, and T_f is the ocean freezing point at ice-base depth, assuming a salinity of 34.5 psu. The salinity – chosen to represent values at typical depths of Circumpolar Deep Water, not considering ice melt – has a very small effect on basal melt rates and is therefore kept constant. As described in Pollard et al. (2015, Supplemental Information), K is a spatially uniform constant (K=3). This yields reasonable patterns of modern sub-ice shelf melt (as shown in Pollard et al., 2017, Appendix), which are generally within ongoing estimates of empirical uncertainties and rapidly changing decadal trends (Depoorter et al., 2013; Rignot et al., 2013). Melt rates at vertical ice faces in direct contact with the ocean are calculated by multiplying the area of each vertical face with the oceanic melt rates at that grid point.

Calving rates at the ice shelf edge are parameterized based on the large-scale stress field, represented by the horizontal divergence of the ice shelf (Pollard and DeConto, 2012a; Nick et al., 2013). In recent years a new set of parameterizations was introduced to the ice sheet model representing sub-grid scale processes that have been hypothesized to significantly increase the sensitivity of the ~~Antarctic ice sheet~~ AIS to climatic forcing (Pollard et al., 2015). These parameterizations include increased calving due to hydrofracturing by surface melt and rainfall draining into crevasses (Nick et al., 2013), as well as structural failure at the grounding line when the vertical face of ice cliffs is too tall ('cliff failure') (Bassis and Walker, 2012). Combined, these two mechanisms have the potential to significantly reduce ice shelf extent and buttressing in warm climates (DeConto and Pollard, 2016; Bell et al., 2018). It is worth noting that these parameterizations – whose validity continues to be debated (Edwards et al., 2019) – do not get triggered by our simulated late-Quaternary climate anomalies (Sect. 3.1).

The mass balance terms in this study are calculated from a file written at run time that stores accumulation (snow+rain), ablation (abl), oceanic melt (ocn), melting at vertical ocean faces (face) and calving (calv), averaged over the entire ice sheet

area. The ablation term (abl) here represents the combined contributions of evaporation at the surface, melting at the base of the grounded ice sheet, and percolation of rain, surface melt water and frictional melt water to the base of the ice sheet, minus refreezing in the ice column. Evaporation and basal melting of grounded ice are both very minor, and surface melt dominates the percolation term; therefore we refer to the ablation term below as ~~'~~'surface melt'.

2.2.2 Climate and sea level forcing over the last 408 ka

Instead of parameterizing the paleo-climate forcing of the ~~Late Quaternary~~late-Quaternary, as done in previous studies, we force the Penn State ice sheet model with climate anomalies from the 408 ka transient experiment described in Sect. 2.1 (Tigchelaar et al., 2018). The climate forcing in the ice sheet model is updated every 1000 calendar years. Climate anomalies are calculated with respect to the LOVECLIM climatology over the last 200 model years (representing 1000 calendar years) and are bilinearly interpolated to the ice sheet model grid~~, then applied and updated every 1000 ice sheet model years. For~~. The atmospheric temperature T_a ~~,~~is modified by a lapse rate correction of $\gamma = 0.008\ ^\circ\mathrm{C\,m^{-1}}$ ~~is applied to correct for differences between LOVECLIM orography z^LC and present-day Antarctic elevation (z^obs; Le Brocq et al., 2010), in addition to differences between present-day elevation and~~ to account for surface elevation differences between the reference ice sheet geometry (z^obs; Le Brocq et al., 2010) and the simulated elevation at time ~~t~~t ($z(t)$). Subsequently, monthly temperature anomalies are added to the present-day temperature field (Fig. 3a; Sect. 2.2.1):

~~$$T_\mathrm{a}(t,\tau) = \quad T_\mathrm{a}^\mathrm{obs}(\tau) - \gamma \times \left(z(t) - z^\mathrm{obs}\right) + \left(T_\mathrm{a}^\mathrm{LC}(t,\tau) - T_\mathrm{a}^\mathrm{LC}(0,\tau)\right) + \gamma \times \left(z^\mathrm{LC} - z^\mathrm{obs}\right),$$~~

$$T_\mathrm{a}(t,\tau) = T_\mathrm{a}^\mathrm{obs}(\tau) - \gamma\left[z(t) - z^\mathrm{obs}\right] + \left[T_\mathrm{a}^\mathrm{LC}(t,\tau) - T_\mathrm{a}^\mathrm{LC}(0,\tau)\right], \tag{2}$$

where t indicates time in years, τ represents month of year, γ is the lapse rate and superscripts 'obs' and 'LC' indicate present-day and LOVECLIM ~~climatologies~~ variables respectively.

Because the ice sheet model does not include a seasonal cycle for present-day precipitation, precipitation anomalies are calculated with respect to annual mean precipitation. Instead of adding the anomalies to the present-day field, as done for atmospheric temperature, present-day precipitation (P^obs) is multiplied with the ratio of monthly LOVECLIM precipitation at time t ($P^\mathrm{LC}(t,\tau)$) to present-day LOVECLIM precipitation ($P^\mathrm{LC}(0)$):

$$P(t,\tau) = P^\mathrm{obs} \times \left[\frac{P^\mathrm{LC}(t,\tau)}{P^\mathrm{LC}(0)}\right]. \tag{3}$$

This is done to ensure that precipitation rates do not go below zero. Annual mean ocean temperature anomalies from the 400 m depth level in LOVECLIM are added to the ice model field as

$$T_\mathrm{o}(t) = T_\mathrm{o}^\mathrm{obs} + \left[T_\mathrm{o}^\mathrm{LC}(t) - T_\mathrm{o}^\mathrm{LC}(0)\right]. \tag{4}$$

The ocean temperature is set not to decrease below -2.18 °C, which is the freezing temperature of sea water with a salinity of 34.5 psu at 400 m depth (Beckmann and Goosse, 2003).

Figures 3d-f show the differences between LOVECLIM and observed present-day climate. Modeled atmospheric temperatures over the Antarctic interior are too high, even when corrected for differences in observed surface elevation and the T21 spectral representation of Antarctic orography in LOVECLIM (Fig. 3d). Present-day Antarctic precipitation is characterized by a temperature-driven low accumulation regime (<50 $\mathrm{mm\,a^{-1}}$) over the Antarctic interior and much higher precipitation rates in coastal areas (>1000 $\mathrm{mm\,a^{-1}}$) as a result of cyclonic activity and topographic uplift (Bromwich, 1988). LOVECLIM does not capture the complex coastal topography of Antarctica well, and therefore underestimates coastal precipitation, distributing it over the ice sheet interior instead (Fig. 3e; Maris et al., 2012). Sub-surface ocean temperatures in LOVECLIM are generally too low in the Southern Ocean, except below the shelves, where they are higher than in the World Ocean Atlas climatology. The lower LOVECLIM temperatures might be related to the fact that for present-day climate, minimum sea ice extent is overestimated (Roche et al., 2012). It should also be noted however that the observed climatology in the Southern Ocean is based on a relatively low number of observations, especially close to the Antarctic continent (Locarnini et al., 2010). In any case, LOVECLIM climate anomalies rather than the full fields are applied to the ice sheet model to avoid the propagation of LOVECLIM biases into the ice sheet evolution. As will be discussed in Sect. 3.1, in spite of present-day biases, LOVECLIM ~~simulates the Quaternary~~ generally simulates the late-Quaternary climate evolution well.

In addition to climate anomalies, the ice sheet model is forced with time-evolving eustatic sea level. Sea level variations are derived from Spratt and Lisiecki (2016) and are plotted in Fig. 1c. While the climate fields are updated every 1000 years, sea level evolves continuously. The ~~bottom half of Fig. 2 illustrates how the climate anomalies and sea level are used to drive the ice sheet model~~ PSU-ISM uses a standard Elastic Lithosphere Relaxing Asthenosphere model for bed depression and rebound under the varying ice load, and therefore does not include deformational, gravitational, and rotational contributions to local sea level change. Such contributions would potentially act as negative feedbacks for ice retreat, and cause spatial variability between the East and West Antarctic Ice Sheets (Gomez et al., 2015). However, previous work that includes full-Earth coupling suggest this would likely only have small effects on our timescales (Gomez et al., 2013, 2015; Pollard et al., 2017), and it is currently not computationally feasible to run a full-Earth model for our 400 kyr simulations (though work is in progress to improve this, e.g .

2.2.3 Sensitivity experiments

The main ice sheet model simulation is run for 408 ka and includes all drivers described in Sect. 2.2.2 (experiment 'all'). In order to isolate the effects of these individual external forcings on ~~Antarctic ice sheet~~ AIS variability and their interaction, we performed a series of sensitivity experiments that include only one or multiple drivers. The individual drivers are either the atmospheric forcing described by Eqs. (2) and (3), the ocean temperature forcing of Eq. (4) or the sea level variations from Spratt and Lisiecki (2016) (experiments 'atm', 'ocn' and 'sl', respectively). In addition to these singular forcing experiments, the model is forced with combinations of two of these three forcings (experiments 'sl+atm', 'sl+ocn' and 'atm+ocn'). These

experiments are designed to quantify the synergistic response of the ~~Antarctic ice sheet~~ AIS to a variety of acting forcings. All sensitivity experiments are summarized in Table 1.

3 Results

3.1 ~~Late Quaternary~~ Late-Quaternary climate forcing

The spatial and temporal evolution of atmospheric temperature, precipitation and sub-surface ocean temperatures are characterized by the first principal component (PC1) and the corresponding spatial pattern (EOF1) as shown in Fig. 3g-l. The full amplitude of this first mode at each point in space can be derived by multiplying the EOF1 map with the PC1 time series. As can be seen in Fig. 3j, annual mean surface temperatures over Antarctica are predominantly paced by changes in atmospheric CO_2 (Fig. 1c). Timmermann et al. (2014) showed that obliquity also contributes to annual mean temperature changes, by affecting annual mean insolation (Fig. 1b) and modulating the strength of the Southern Hemisphere westerlies. The dominant pattern of annual mean temperature changes is homogeneous, with a glacial-interglacial amplitude of ~4-8 °C (Fig. 3g). ~~When compared~~ We compare this to a composite of temperature reconstructions~~from ice cores (Parrenin et al., 2013), the~~ , calculated as the average of available long-term temperature records for each time in the past (Parrenin et al., 2013). The temporal evolution of ~~the LOVECLIM PC1~~ simulated Antarctic air temperature is very similar, but the amplitude is underestimated by a factor of 1.5~~-2, partially~~ -2. This could partially be due to the fact that the LOVECLIM simulation does not include the lapse rate response to the evolving ice sheet height, but also points at an underestimation of polar temperature change in the climate model, especially during interglacials (Tigchelaar et al., 2018). In addition to annual mean temperatures, surface ablation rates are sensitive to changes in seasonal insolation (Huybers and Denton, 2008; Huybers, 2009; Tigchelaar et al., 2018), which is precessionally driven and shown in Fig. 1b.

Precipitation changes display a temporal evolution very similar to that of the atmospheric temperature PC1 (Fig. 3k), confirming that temperature is the dominant driver of precipitation over Antarctica. When compared to a composite of ice core accumulation reconstructions (Steig et al., 2000; Bazin et al., 2013; Vallelonga et al., 2013), LOVECLIM is shown to overestimate precipitation rates during early glacial times. Steig et al. (2000) describe how when the ~~Antarctic ice sheet~~ AIS is expanding, the coastal ice core locations switch from a cyclonic-driven precipitation regime to one driven by temperature with increasing distance from the ice edge. The local precipitation evolution captured by the ice cores thus differs from the large-scale evolution captured by the principal component analysis. This ice sheet-climate feedback is not included in our LOVECLIM simulations.

The temporal evolution of sub-surface ocean temperatures in LOVECLIM (Fig. 3l) is similar to that of surface (not shown) and atmospheric temperatures (Fig. 3j). To our knowledge no reconstructions of intermediate water temperatures in the Southern Ocean exist, so we compare against a long sea surface temperature (SST) record from 54 °S (Ho et al., 2012) ~~. Both in our model simulation as well as in reconstructions, Southern Ocean SST variability is closely related to changes in sea ice area and production, explaining why there is substantial precessional variability in these time series (Timmermann et al., 2009).~~ and a deep-sea temperature record from 41 °S (Elderfield et al., 2012). The LOVECLIM-simulated glacial-interglacial SST anomaly

at the Ho et al. (2012) core location is about 8 °C (not shown), which is similar to the reconstructed amplitude of temperature variability – though the Last Interglacial warming is less pronounced in the climate model. At depth however, the simulated glacial-interglacial amplitude is about three times smaller than the Elderfield et al. (2012) reconstruction (not shown). The LOVECLIM-simulated ocean temperature anomalies close to the Antarctic continent are ~~very small, with the exception of the Weddell sector. The effect~~ also small: zonal mean glacial-interglacial temperature anomalies at 65 °S and 400 m depth are about 0.6 °C, smaller than in some other models (Lowry et al., 2019) and with minimal interglacial warming (Tigchelaar et al., 2018). This means the magnitude of ocean forcing in our simulations is much smaller than thresholds for interglacial ice sheet collapse found in previous sensitivity studies (e.g. Sutter et al., 2016; Tigchelaar et al., 2018). The implications of this possible underestimation of ocean forcing on ice sheet evolution will be discussed further below.

3.2 Ice volume response to external forcing

Figure 4 shows the simulated response of Antarctic total, grounded, and floating ice volume to the individual and combined ~~Late Quaternary~~ late-Quaternary forcings over the last four glacial cycles. With all forcings combined ('all'), the glacial-interglacial difference in ice volume is $\sim 8 \times 10^6$ km^3, or 12-14 m sea level equivalent (SLE) depending on the glacial stage (Fig. 4). During glacial periods, floating ice volume is reduced by about half the present day value ($\sim 7 \times 10^5$ km^3) (Fig. 4c). In our simulations, previous interglacials only contribute 1-2 m to global sea level ($\sim 1 \times 10^6$ km^3), with the deepest interglacial occurring at 210 ka (Termination IIIa). Tigchelaar et al. (2018) showed that local changes in summer insolation play an important role in amplifying interglacial ice loss.

The dominant spatial pattern of ice sheet thickness variability in the 'all' simulation, along with minimum (210 ka) and maximum (18 ka) grounding line extent, are shown in Fig. 5a. At its maximum extent, the grounding line reaches to the continental shelf break everywhere. The simulated minimum grounding line extent over the last 408 ka is very similar to present-day, with further retreat mostly of the Ross and Weddell ice shelves in West Antarctica, and the West and Shackleton ice shelves in East Antarctica. Changes in ice sheet thickness are most pronounced in those regions where the grounded ice sheet expands, in particular the Ross and Weddell sectors, Amundsen Sea and Amery shelf. In the interior of the AIS, thickness changes are generally smaller, but mostly of the same sign.

Generally speaking, our complete-forcing simulation captures the main features of the Antarctic LGM and subsequent deglaciation well, to within the general level of uncertainties within the paleo-data (Pollard and DeConto, 2012a, b; Pollard et al., 2016; Tig . The simulated LGM grounding line position is in close agreement with reconstructions, as is the sequencing of regional deglaciation (Bellingshausen, followed by Amundsen, followed by Weddell, Ross and Amery) (RAISED Consortium et al., 2014) . However, in our simulation retreat in the Ross sector occurs at least ~2 ka earlier than reconstructions suggest, and there is an 'overshoot' of Siple Coast grounding lines at ~6-4 ka. Addressing these discrepancies is the focus of ongoing work, including the large-ensemble simulations of e.g., Briggs and Tarasov (2013); Briggs et al. (2014); Pollard et al. (2017).

3.3 Nonlinear response to climate and sea level forcing

Not one of the individual drivers of ~~Quaternary~~ late-Quaternary AIS variability – sea level, atmospheric temperature and precipitation, ocean temperatures – single-handedly explains the full ice volume evolution (Fig. 4a). Moreover, all of the individual forcings combined only account for less than half of the total ice volume changes, suggesting that they do not add linearly. The largest contribution in terms of both total and grounded ice volume comes from the atmospheric forcing, which explains about a third of glacial ice volume gain, and the entirety of interglacial ice volume loss (Fig. 4a,b). The case is different for floating ice: here sea level changes are responsible for most of the variability, as a lowering sea level converts floating into grounded ice (Fig. 4c; Schoof, 2007). Interestingly, for the floating ice volume, the sum of the individual simulations is not only smaller than, but also often not of the same sign as the floating ice volume changes in the 'all' simulation. When the ice sheet model is forced with two out of three forcings, sea level and atmospheric forcing together almost entirely explain the changes in both grounded and floating ice volume (Fig. 6).

Figure 5 shows where ~~on the Antarctic continent the~~ the individual drivers have the largest effect on the AIS. The sea level forcing drives expansion of the grounding line in the Amundsen, Ross, and Weddell Sea sectors, with small corresponding elevation changes (Fig. 5b). Atmospheric cooling leads to grounding line expansion primarily in the Amundsen, Weddell, and Amery regions, and also leads to thickening of most of the ice shelves (Fig. 5c). During interglacials, the atmospheric forcing causes retreat primarily of the West and Shackleton ice shelves. ~~The~~ Due to the small magnitude of the simulated ocean temperature change, the oceanic forcing alone affects Antarctic ice volume only minimally. In fact, the dominant spatial pattern of ice thickness variability for the 'ocn' simulation only explains $\sim$10% of the variance, and is not driven by external forcing, but rather displays internally generated ice sheet variability in the Siple Dome region (Fig. 5d) with a period of $\sim$10 ka (not shown).

~~When more than one external forcings are combined, the grounding line is able to expand further than with just the single forcings~~ Combinations of external forcings lead to a more pronounced grounding line advance during glacials than in simulations with one single forcing (Fig. 5e-g). As noted above, sea level and atmospheric forcing combined (Fig. 5e) explain most of the grounding line and elevation changes simulated in the full run (Fig. 5a). A combination of sea level and ocean forcing (Fig. 5f) leads to grounding line expansion and ice sheet growth in the Weddell Sea sector, while atmospheric and ocean forcing combined (Fig. 5g) mostly cause ice sheet growth in the Ross Sea.

3.4 Mechanisms explaining ice volume changes

3.4.1 Sea level forcing

Figure 7 ~~plots changes in grounded and ice volume changes~~ depicts the response of grounded ice volume to the respective forcing in the different sensitivity runs~~against their respective forcing functions.Again, it is evident that the role of~~ , with corresponding mass balance changes shown in Fig. 8. As noted before, the impact of sea level forcing in isolation is to ~~directly turn~~ convert grounded into floating ice during periods of sea level rise, and the other way around during sea level drops (Fig. 7a). ~~The corresponding mass balance changes are shown in Fig. 8. With sea level as the only driver, changes in~~ Changes in mass

balance rates are a feedback to ~~the~~ these changes in ice sheet configuration. Ice-sheet integrated surface melt rates (Fig. 8b) increase during glacial periods of sea level drop ~~,~~ because the edges of the ice sheet – where all surface melt occurs – are lower in elevation, with associated higher temperatures. Calving rates (Fig. 8d) similarly increase during periods of low sea level, because the grounding line is positioned more ~~equator-ward~~ equatorward (Fig. 5), increasing ice shelf divergence (Tigchelaar et al., 2018). On the other hand, ice-sheet integrated oceanic melt rates (Fig. 8c) decrease when sea level drops ~~,~~ because the ice-ocean interface area is reduced. These mass balance feedbacks mostly cancel out in the net mass balance (Fig. 8e), explaining why total ice volume changes under isolated sea level forcing are small (Fig. 4).

3.4.2 Atmospheric forcing

When atmospheric forcing is applied in isolation, grounded ice volume ~~fairly consistently~~ increases with decreasing surface temperature, whereas floating ice volume plateaus for ice-sheet averaged temperatures lower than ∼-34°C (Fig. 7b). ~~Now,~~ In this case the mass balance response (Fig. 8) is a combination of both forcing and feedback. Surface melt rates (Fig. 8b) most directly follow the climatic forcing. As detailed in Tigchelaar et al. (2018), periods of high CO_2 and high summer insolation (Fig. 1) are marked by peaks in summer melt rates that also drive increases in calving rates (Fig. 8d). During these periods, the AIS retreats to areas that have lower accumulation rates (Fig. 3b), amplifying the forcing. In cold periods, a reduction in surface melt and calving leads to a small ~~expansion of floating ice volume~~ outward expansion of the grounding line (Fig. ~~4~~5c), which ~~allows~~ causes the floating ice shelves to sit in climatologically warmer waters (Fig. 3c), increasing glacial ocean melt rates (Fig. 8c). ~~At the same time, the glacial ice sheet extends to areas of climatologically high snow fall.~~ The changes in ~~calving rates and~~ ocean melt rates ~~largely balance , so that peaks in accumulation rates drive much of the ice sheet growth~~ almost but not entirely balance the surface melt and calving rate changes, making the mass balance slightly positive during glacial periods (Fig. 8e).

3.4.3 Ocean temperature forcing

Out of the three individual drivers, the ocean temperature forcing by itself leads to the least change in grounded and floating AIS volume, as seen in Figs. 4 and 5d. LOVECLIM-modeled ocean temperature ~~changes are fairly small in amplitude~~ anomalies are small (Fig. 3i), and lead to ~~small~~ minor increases in grounded and floating ice thickness during glacial periods (Fig. 7c). The accompanying mass balance changes are similarly small (Fig. 8). Glacial expansion of floating ice area (Fig. 4c) brings ice shelves into areas with climatologically higher precipitation rates (Fig. 3b), leading to higher glacial accumulation rates (Fig. 8a). The reduced oceanic melt and increased accumulation are balanced by higher calving rates (Fig. 8d). It is important to note here that this small response to ocean temperature forcing is more likely a function of the low amplitude of the LOVECLIM-simulated ocean temperature forcing than it is indicative of low sensitivity of the AIS to changing ocean conditions, as will be discussed further below.

3.4.4 Combined forcings

Our sensitivity runs show that the simulated response of the ~~Antarctic ice sheet to Quaternary~~ AIS to late-Quaternary external drivers is a nonlinear superposition of a) a direct mass balance response to climate variations, b) sea-level induced conversion between grounded and floating ice, and c) areal expansion or contraction against climatological gradients. As shown in Fig. 7d-f, when all forcings combine, sea level is the dominant pace maker of both grounded and floating ice volume. Because sea level and atmospheric temperature vary in concert throughout the ~~Quaternary~~ late-Quaternary (Fig. 1c, Fig. 3j), grounded ice volume also increases with lowering temperatures, while floating ice volume now decouples from atmospheric temperatures (Fig. 7e). The joint sea level and atmospheric forcing amplify each other in the total ice volume response, because the combination of shelf-to-sheet conversion (Fig. 4c) and reduced calving rates (Fig. 8d) allow the grounding line to migrate ~~equator-ward~~ equatorward during glacial times (Fig. 5). This increases the ice sheet area – and thus the ice-sheet integrated accumulation rate (Fig. 8a) – leading to a net positive mass balance (Fig. 8e), and ice sheet growth.

With all forcings combined, the simulated ice sheet response is completely decoupled from the oceanic temperature forcing (Fig. 4f). During glacial periods, the grounding line is closer to warmer Circumpolar Deep Water (Fig. 3c), so that periods of high total ice volume are associated with high ocean temperatures beneath the ice shelves. This is also seen in Fig. 8c, where ice-sheet averaged oceanic melt rates in the 'all' simulation more closely follow those of the 'atm' run than the 'ocn' run. The spatial gradients in ocean temperature are thus larger and more important than temporal (glacial-interglacial) temperature variations (Fig. 3). The main exception to this is the Ross sector, where decreasing ocean temperatures allow for further ice expansion and grounding line migration during glacial times (compare e.g., Fig. 5c & g). ~~Though a similar dependence of glacial~~ Kusahara et al. (2015) also found oceanic melt rates ~~on grounding line position rather than climate forcing was also found by Kusahara et al. (2015), we will briefly discuss the validity of these results below~~to increase during the Last Glacial Maximum in response to grounding line migration, lending support to these findings. However, as shown in Tigchelaar et al. (2018), the low sensitivity of the modeled AIS to interglacial ocean conditions is likely a result of the low amplitude and resolution of the LOVECLIM ocean temperature forcing and lack of ice-ocean feedbacks in our modeling setup.

4 Discussion & Conclusions

Here we presented results from simulations of Antarctic ~~ice sheet~~ Ice Sheet evolution over the past 408 ka. In contrast to previous work which primarily used parameterized forcing, climate anomalies (atmospheric temperature, precipitation and subsurface ocean temperatures) were directly derived from a transient simulation with the EMIC LOVECLIM. The simulated AIS has a glacial-interglacial amplitude of 12–14 m SLE, with the glacial grounding line extending almost entirely to the continental shelf break, and past interglacials showing limited retreat of 1–2 m SLE. Sensitivity experiments where atmospheric, oceanic and sea level forcing were applied in isolation or in pairs, showed that the combined effect of individual forcings is strongly nonlinear. Each of the individual forcings explains less than a third of the full response, and the sum of the individual forcing simulations is less than half of the glacial-interglacial amplitude with all forcings applied jointly. In our simulations, sea level and atmospheric forcing together explain most of the full response, both in terms of amplitude and pacing.

~~While during glacial periods the dynamics of a lowering sea level and a cooling climate amplify each other in AIS growth, interglacial ice volume loss is almost entirely driven by climate forcing alone (Fig. 4).Tigchelaar et al. (2018) showed that maximum interglacial ice loss occurs when high CO_2 concentrations coincide with high Southern Hemisphere summer insolation (Fig. 1). These precessionally-forced periods of warm summers are typically out of phase with eustatic sea level forcing, which is predominantly paced by warm Northern Hemisphere summers (Raymo et al., 2006). Our simulations therefore do not fully explore the response of the AIS to combined climate warming and rising sea levels, as they would co-occur in future climate change. So far,~~ Our finding that ocean temperature forcing plays a limited role in driving changes in Antarctic ice volume contrasts with previous modeling studies of past and future AIS evolution (Golledge et al., 2015; DeConto and Pollard, 2016; Sutter et al., 2 , as well as observations of sustained sub-shelf ice loss in response to ongoing ocean warming at e.g., Pine Island Glacier (Jacobs et al., 2011; Pritchard et al., 2012). This is not surprising, given that the LOVECLIM-simulated ocean temperature anomalies are small (Figs. 3i,l), and ice sheet models typically need ocean warming of 2-5 °C to initiate interglacial WAIS collapse (Pollard and DeConto, 2009; DeConto and Pollard, 2016; Sutter et al., 2016; Tigchelaar et al., 2018). In the absence of paleo-reconstructions of near-Antarctic sub-surface ocean temperatures it is difficult to assess how realistic our LOVECLIM simulation is, though critical processes such as Antarctic Bottom Water formation are known to be poorly represented in low-resolution climate models (e.g., Snow et al., 2015), and previous studies have found LOVECLIM in particular to have more muted late-Quaternary temperature variability than other models (Lowry et al., 2019).

In addition, many regional oceanographic processes can affect the circum-Antarctic ocean environment beyond large-scale climate forcing. For example, the blocking effects of sea ice formation (Hellmer et al., 2012), the role of winds in pushing warm waters onto the continental shelf (Thoma et al., 2008; Steig et al., 2012), and the complex geometry of ice shelf cavities (Jacobs et al., 2011; De Rydt et al., 2014) have all been found to be important in observational and modeling studies of ~~future Antartic ice sheet evolution (e.g., Joughin and Alley, 2011; Scambos et al., 2017; DeConto and Pollard, 2016) have not included changes in eustatic sea level. Further research should therefore explore whether, given the current configuration of grounding line and bedrock (Joughin and Alley, 2011; Joughin et al., 2014), rising sea levels would have the potential to increase ice loss due to warming.~~current and future oceanic melting of the WAIS ice shelves (Joughin et al., 2014). For that reason, using 400 m-depth Southern Ocean temperatures as the sole driver for sub-shelf melt may miss important near-Antarctic dynamics. Furthermore, melt water fluxes from the AIS have been found to lead to cooling of surface waters and warming at intermediate depth (Menviel et al., 2010; Weber et al., 2014), a feedback mechanism that could increase ice sheet loss (Golledge et al., 2014, 2019) .

These processes can only really be captured in fully coupled ocean-atmosphere-ice sheet simulations at high resolution, something that is currently not feasible for the long timescales of late-Quaternary climate evolution. However, it should be possible to run shorter simulations – of e.g., the Last Interglacial or Marine Isotope Stage 11 – using such a setup, and perform a similar set of sensitivity experiments as done here. This would likely reveal additional nonlinearities as ice sheet and forcing are allowed to evolve together. The accumulation forcing for instance is similarly impacted by low climate model resolution and lack of ice-climate feedbacks. Time-evolving changes in orography and albedo can substantially alter atmospheric circulation patterns and associated rainfall (Steig et al., 2000; Maris et al., 2014; Steig et al., 2015).

The strongly nonlinear response of the AIS to different external forcing agents ~~also~~ underscores the importance of ~~supplying~~ driving the ice sheet model with accurately dated sea level and climate forcing. Previous modeling studies of past AIS evolution (Ritz et al., 2001; Huybrechts, 2002; Pollard and DeConto, 2009) have mostly bypassed this issue by assuming that both the sea level and climate forcing vary in concert with either Antarctic temperature reconstructions or the benthic δ^{18}O record. However, ~~as shown in Fig. 1c,~~ global sea level and global climate (i.e., CO_2) do not always vary ~~entirely in phase . Meanwhile, Tigchelaar et al. (2018) highlighted the importance of local summer insolation in addition to~~ in phase (Fig. 1c), and local climate conditions (e.g., through local insolation changes) can deviate substantially from global climate variability (Tigchelaar et al., 2018). At the same time, there are significant uncertainties in the timescales of especially Antarctic climate and CO_2 reconstructions (Lüthi et al., 2008; Bazin et al., 2013). Repeating the LOVECLIM climate simulations with a proper uncertainty range in atmospheric greenhouse gas concentrations would be computationally unfeasible, though future sensitivity runs with the ice sheet model could ~~manually offset the~~ include artificial shifts in the phase relationship between the sea level and climate forcing to explore associated nonlinearities.

~~Finally, our results clearly highlight the interplay between spatial gradients in Southern Hemisphere climate and the temporal evolution of the Antarctic ice sheet. In particular, as the ice sheet grows, it sits in areas of higher precipitation and warmer ocean water, acting as a positive and negative feedback respectively. However, in this modeling setup we are not able to account for feedbacks between ice sheet configuration and climate . In the case of precipitation, changes in orography and albedo could substantially alter atmospheric circulation patterns and associated rainfall (Steig et al., 2000; Maris et al., 2014; Steig et al., 2015) .~~

~~As for ocean temperatures, the limited direct~~ In our simulations, the sea level and climate forcing amplify each other during glacial AIS growth, while interglacial ice volume loss is almost exclusively driven by climate forcing (Fig. 4). Tigchelaar et al. (2018) showed that maximum interglacial ice loss occurs when high CO_2 concentrations coincide with high Southern Hemisphere summer insolation (Fig. 1). These precessionally-forced periods of warm summers are typically out of phase with eustatic sea level forcing, which is predominantly paced by warm Northern Hemisphere summers (Raymo et al., 2006). Our simulations therefore do not fully explore the response of the AIS to ~~ocean temperature changes could be due to the small glacial-interglacial amplitude simulated by LOVECLIM, though a study by Kusahara et al. (2015) with a much more advanced and high-resolution ocean-ice sheet model also found glacial oceanic melt rate changes to respond to ice sheet geometry instead of climatic change. In reality, melting at the ice-ocean interface depends on much more than sub-surface ocean temperatures alone. The blocking effects of sea ice formation (Hellmer et al., 2012), the role of winds in pushing warm waters onto the continental shelf (Thoma et al., 2008; Steig et al., 2012), and the complex geometry of ice shelf cavities (Jacobs et al., 2011; De Rydt et al., 2014) have all been found to be important in observational and~~ combined climate warming and rising sea levels, as they would co-occur in future climate change. So far, most modeling studies of ~~current and future oceanic melting of the WAIS ice shelves (Joughin et al., 2014). Additionally, melt water fluxes from the AIS have been found to lead to cooling of surface waters and warming at intermediate depth (Menviel et al., 2010; Weber et al., 2014), a feedback that could increase ice sheet loss (Golledge et al., 2014, 2019). These processes can only be captured in fully coupled ocean-atmosphere-ice sheet simulations at high resolution, something that is currently not feasible for the long timescales of Quaternary climate evolution~~future Antartic

ice sheet evolution (e.g., Joughin and Alley, 2011; Scambos et al., 2017; DeConto and Pollard, 2016) have not included changes in eustatic sea level. Further research should therefore explore whether – given the current configuration of grounding line and bedrock (Joughin and Alley, 2011; Joughin et al., 2014) – rising sea levels as a result of global warming would further increase or stabilize ice loss. Such future studies should make sure to include the deformational and gravitational components of future sea level rise through coupling with a full-Earth model (Gomez et al., 2018).

In response to changes in atmospheric and oceanic conditions and global sea level, Antarctic ice volume has varied by tens of m SLE throughout the ~~Quaternary~~late-Quaternary, and is expected to decrease in the future. ~~Previous modeling studieshave failed to elucidate how these different external drivers interact in driving large glacial ice sheet growth and interglacial sea level highstands~~In contrast to previous modeling studies, here we focus on the interaction of different external forcings driving Antarctic ice volume changes. Our sensitivity experiments with an Antarctic ~~ice sheet~~ Ice Sheet model over the last four glacial cycles show that the glacial-interglacial ice sheet response to environmental forcing is strongly nonlinear. Both atmospheric cooling and a transformation of dynamic regime by lowering sea level ~~and atmospheric cooling~~ are necessary to generate full glacial ice sheet growth. Our ~~results suggest that the contribution of future sea level rise to Antarctic ice loss, which has so far remained unexplored,~~ modeling setup likely underestimates the role of oceanic forcing, which remains largely unbound by the geologic record and needs to be ~~incorporated in future modeling, while further underscoring the need for~~ further explored in a coupled climate-ice sheet modeling framework that can account for critical circum-Antarctic oceanographic processes.

Author contributions. MT, AT, and DP designed the ice sheet model experiments. TF ran the climate model simulations. DP developed the ice sheet model, and MH developed the ice sheet-climate coupling. MT ran the ice sheet model and analyzed the results; MT, AT, and DP contributed to the interpretation of the results. MT wrote the first draft of the paper; AT, TF, MH, and DP contributed substantially to its final version.

Data availability. Our ice sheet simulations are publicly available at https://climatedata.ibs.re.kr/grav/data/psu-love/.

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

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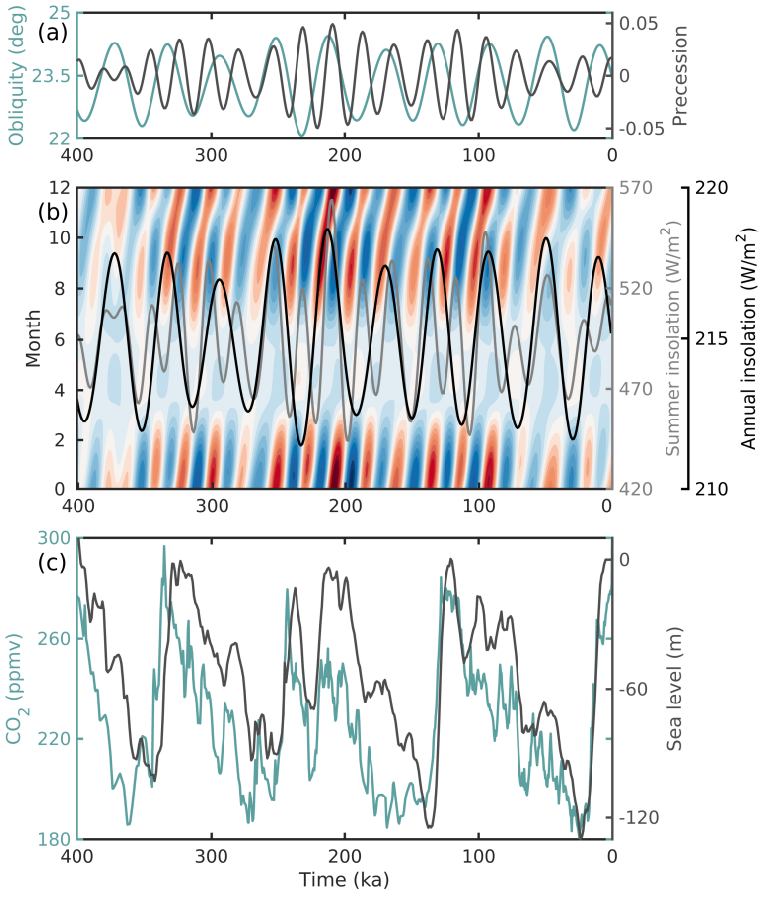

Figure 1. Climate drivers over the last 400 ka – (a) precession (grey) and obliquity (teal) (Laskar et al., 2004); (b) monthly insolation anomalies (colors, contours ranging from ± 65 W m^{-2}), annual mean insolation (black) and summer insolation (grey) at 65 °S as a result of the orbital forcing in (a) (Laskar et al., 2004); (c) atmospheric CO_2 concentration (teal; Lüthi et al., 2008) and global sea level (m) (grey; Spratt and Lisiecki, 2016).

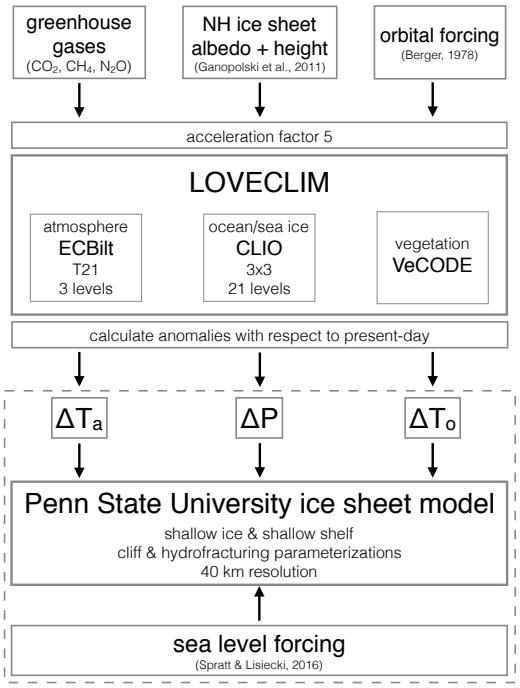

Figure 2. Schematic illustrating the modeling setup as described in Sect. 2.

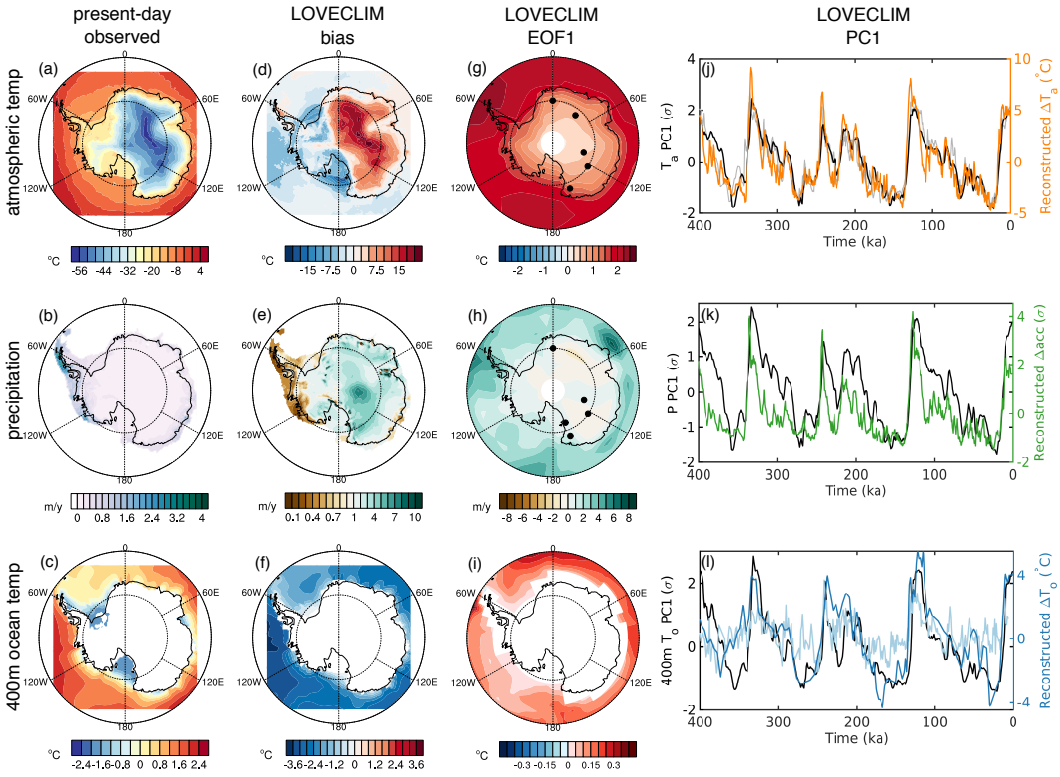

Figure 3. Climate forcing on the ice model grid – (left) Present-day climate conditions (Locarnini et al., 2010; Le Brocq et al., 2010), (second from left) LOVECLIM bias with respect to present-day climate, (third from left) first EOF and (right) first ~~PC1~~ PC in LOVECLIM for (top) annual mean atmospheric temperature, (middle) annual mean accumulation (observed) and precipitation (LOVECLIM) and (bottom) annual mean ocean temperature at 400 m depth. Multiply the EOF pattern with the PC time series to obtain the full amplitude of the dominant mode of variability at each location. For annual mean temperature the LOVECLIM temperatures were ~~first~~ adjusted to observed elevations (Le Brocq et al., 2010) using a lapse-rate correction of 0.008 °C m^{-1}. The atmospheric and ocean temperature biases are plotted as LOVECLIM–observed, while the precipitation bias is plotted as LOVECLIM/observed. In addition, (j) shows a composite of reconstructed temperature anomalies from ice cores (°C, orange; locations indicated by black dots in (g); Parrenin et al., 2013) and the normalized CO_2 record (grey; Lüthi et al., 2008), (k) shows a composite of reconstructed accumulation anomalies from ice cores (σ, green; locations indicated by black dots in (h); Steig et al., 2000; Bazin et al., 2013; Vallelonga et al., 2013) and (l) shows reconstructed deep-sea temperature anomalies at 171 °W, 41 °S (°C, light blue; Elderfield et al., 2012) and SST anomalies at 80 °W, 54 °S ~~(°C, blue; Ho et al., 2012)~~(°C, dark blue; Ho et al., 2012).

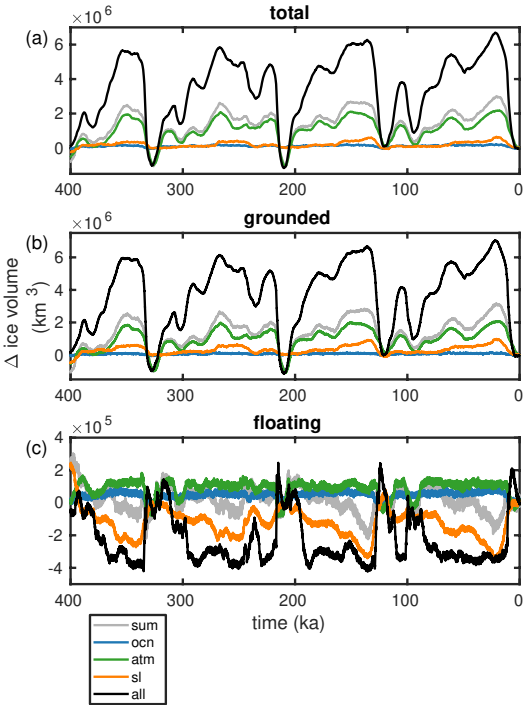

Figure 4. Ice sheet evolution over the last 400 ka for experiments 'ocn' (blue), 'atm' (green), 'sl' (orange), and 'all' (black) (Table 1) – (a) total ice sheet volume; (b) grounded ice sheet volume; and (c) floating ice sheet volume. The grey line is the sum of the individual runs 'ocn', 'atm', and 'sl'.

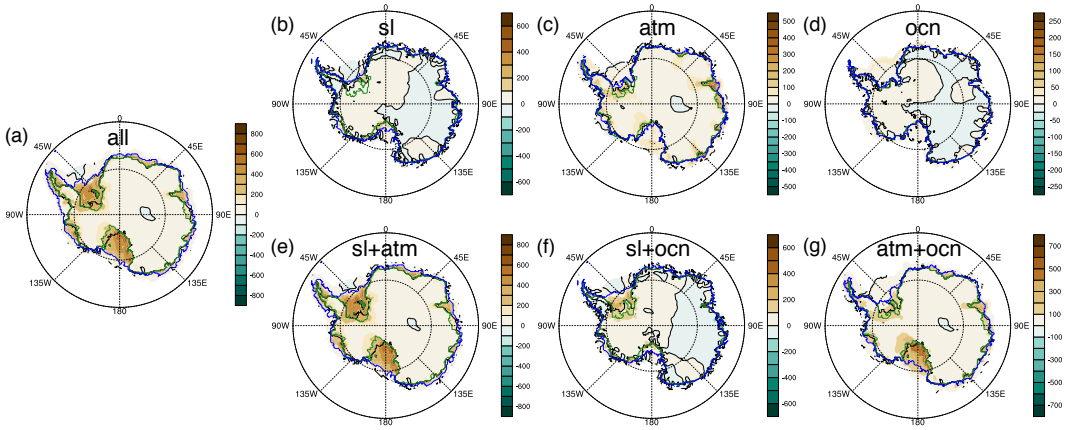

Figure 5. Dominant spatial pattern (first EOF) of ice sheet thickness variability (m) and minimum (green), maximum (blue) and present-day (black) grounding line extent for (a) 'all', minimum at 331 ka, maximum at 18 ka, 75.8% of variance explained; (b) 'sl', minimum at 121 ka, maximum at 18 ka, 39.4% of variance explained; (c) 'atm', minimum at 331 ka, maximum at 350 ka, 50.7% of variance explained; (d) 'ocn', minimum at 7 ka, maximum at 156 ka, 9.6% of variance explained; (e) 'sl+atm', minimum at 331 ka, maximum at 20 ka, 76.6% of variance explained; (f) 'sl+ocn', minimum at 122 ka, maximum at 140 ka, 50.5% of variance explained; (g) 'atm+ocn', minimum at 330 ka, maximum at 354 ka, 63.1% of variance explained.

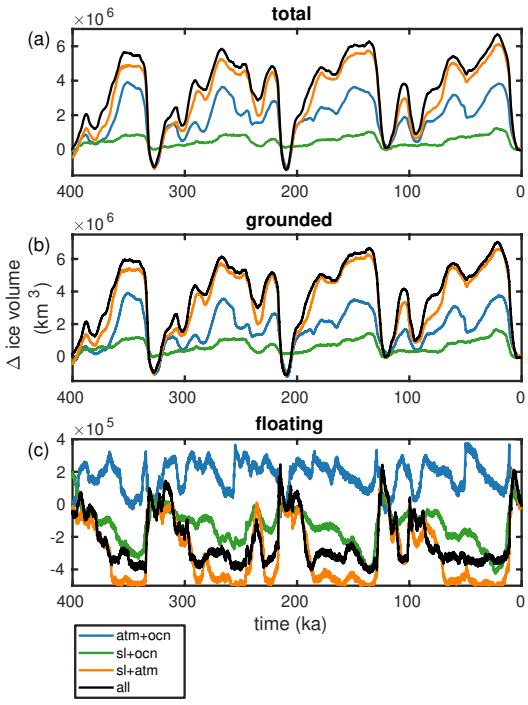

Figure 6. Ice sheet evolution over the last 400 ka for experiments 'atm+ocn' (blue), 'sl+ocn' (green), 'sl+atm' (orange), and 'all' (black) (Table 1) – (a) total ice sheet volume; (b) grounded ice sheet volume; and (c) floating ice sheet volume.

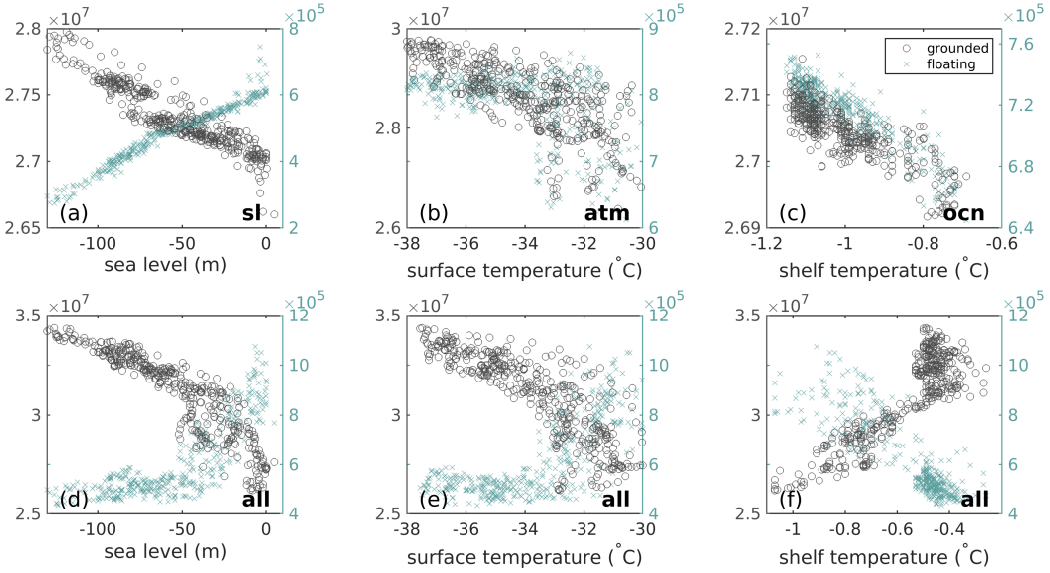

Figure 7. Ice sheet averaged forcing terms against floating (teal crosses) and grounded (grey circles) ice volume (km^3) – sea level in (a) 'sl' and (d) 'all'; atmospheric surface temperature in (b) 'atm' and (e) 'all'; and temperature beneath the ice shelves in (c) 'ocn' and (f) 'all'.

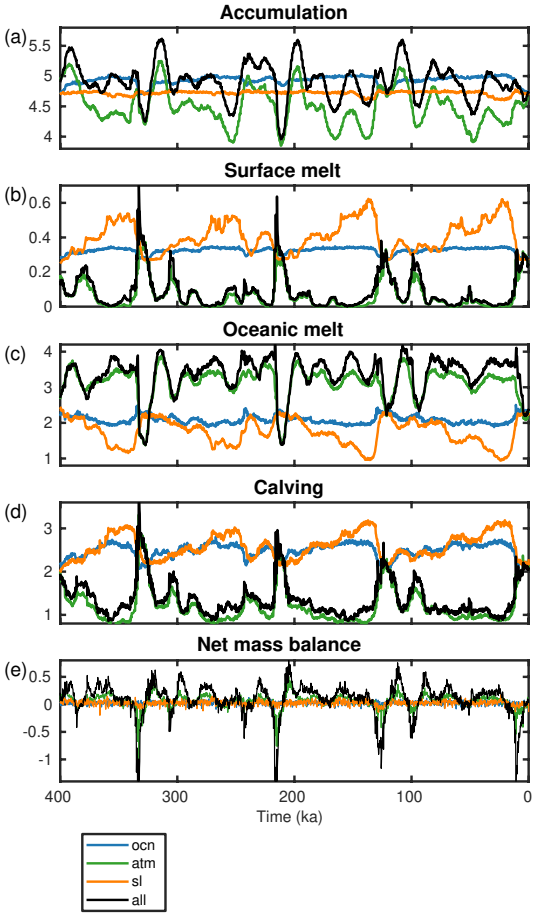

Figure 8. Ice sheet integrated mass balance terms (10^3 Gt y^{-1}) for experiments 'ocn' (blue), 'atm' (green), 'sl' (orange) and 'all' (black) – (a) accumulation, (b) surface melt, (c) oceanic melt, (d) calving, and (e) net mass balance.

Table 1. Overview of the sensitivity experiments described in Sect. 2.2.3

experiment	**description**
all	all forcings (Eqs. (2), (3), (4) & Spratt and Lisiecki (2016))
atm	only atmospheric forcing (Eqs. (2) & (3))
ocn	only ocean temperature forcing (Eq. (4))
sl	only sea level forcing (Spratt and Lisiecki (2016))
sl+atm	sea level and atmospheric forcing (Eqs. (2), (3) & Spratt and Lisiecki (2016))
sl+ocn	sea level and ocean temperature forcing (Eqs. (4) & Spratt and Lisiecki (2016))
atm+ocn	atmospheric and ocean temperature forcing (Eqs. (2), (3), (4))