Dear Editor,

We would like to thank the reviewers and yourself for the excellent and constructive feedback on our revised manuscript. Below we provide a point-by-point response to all comments. The original comments are kept in black and are numbered (in red). We provide our responses in green, and show manuscript revisions in *blue italics*. We also provide a "latexdiff" version of the manuscript, where the changes listed in the responses are highlighted.

We would like to emphasise that we have enlarged all the figures mentioned by the reviewers, and changed single-row plots to two-row plots where suggested. Reviewer 1 further notes in his comment 11 that Figure 10 demanded excessive digital enlargement for reading, and mentions in comment 9 that we could use the full width of the page to make the figures larger. However, in the LaTeX template two-column figures are set to a width of 12 cm, although using 18 cm still seems to keep the figure within the same width as the text. We therefore submit an 18 cm-wide Figure 10, but would be happy to submit it in another size if this is not permitted.

We hope that you find our responses satisfactory, and that you will accept our revised version for publication.

Best regards,

Martim Mas e Braga (corresponding author)

Editor's comments

1. Lines 4-6: check the logic of this sentence – statements are not really linked
We revised the sentence slightly, which now reads (L4):
"Ice core data indicate that warmer-than-present temperatures lasted for longer than during other interglacials. However, the response of the Antarctic ice sheets and their contribution to sea level rise remain unclear."

2. Line 10-11: confusing use of 'sea level reconstructions' in two different senses
We slightly rephrased the sentence for clarity. It now reads (L9):
"In the case of a West Antarctic Ice Sheet collapse, which is the most probable scenario according to far-field sea level reconstructions, the range is further reduced to 6.7-8.2 m independently of the choices of external sea-level forcing and millennial-scale climate variability."

3. Line 283: I think the opening statement is based on experiments driven by different proxy records, rather than the interpretation of the proxy records themselves
The statement is indeed unclear, thank you for pointing that out. We clarified it by writing (L293):
"We base this statement on results from experiments forced by different proxy records with significant differences in their structure during the MIS11c peak warming."

4. Line 322: when describing experiment (ii) I suggest emphasising that 410 ka lies after the sudden increase in ocean temperature but before the maximum is reached
Thank you for the suggestion. We added this information in a parenthesis (L334):

"(i.e., just after the sudden increase in ocean temperatures, but before the maximum is reached; cf. Fig 9b)"

5. General: use of 'this' is occasionally ambiguous, e.g. lines 38, 191, 340
We list below the changes made to the mentioned lines. In L340, we note that changes were also made as a response to comment 8 of Reviewer 1.
L37: *"The unusual length of MIS11c and a transition […]"*
L201: *"Although the similarity to the modern AIS configuration has been loosely inferred from sedimentary (Capron et al., 2019) and ice-core (EPICA Community Members, 2004) proxy records, to our knowledge there is no direct evidence to support this claim […]"*
L353: *"The critical warming of 0.4 ºC we found for MIS11c is close to the equilibrium model results […]"*

6. General: please clarify how you convert ice volume to sea level contribution
We added a clarification of our sea level contribution conversion in the methods (L132):

"Sea-level contribution at a given time step is computed in SICOPOLIS as the difference in total ice volume above flotation between the ice sheet at the time step and the spun-up Pre-Industrial ice sheet. When computing ice volume, differences in bedrock elevation between the two ice sheets are accounted for by using a common reference bedrock elevation in all time steps. We further correct for the projection effect on the horizontal grid area."

7. General: please clarify how you define WAIS collapse
We consider WAIS to have collapsed when the Weddell, Ross, and Amundsen seas become interconnected. We clarify this after the first mention of WAIS collapse in L245:
"However, it recovers more quickly than the EDC and DF experiments as soon as the peak warming is over and the climate starts to shift back to PI conditions, without a WAIS collapse (we consider the WAIS to have collapsed when the Weddell, Ross, and Amundsen seas become interconnected; Fig. 5)."

8. Supp. Fig. S2: please clarify whether the forcings shown relate to glacial or interglacial conditions
We added to this information to the caption of Fig. S2:
"Comparison of Pre-Industrial CCSM3 forcings (right) to reference data (left) from RACMO2"

9. Supp. Fig. S13: please clarify what is meant by 'anomalies'
We have added an explanation of anomalies to the caption of Fig. S13, which now reads:
"(a-f) Forcing fields used to construct the climate forcings used in this study. Left fields show PI mean states, while right fields show applied anomalies (i.e., LGM−PI for temperatures, and LGM÷PI for precipitation). Anomalies are multiplied by the GI and then combined with the left-side fields following Eqs. (4) and (5)."

Reviewer 1's comments

Scientific comments

1. L79-82 In the first review I asked to motivate the choice of ensembles and parameter changes a bit better and the authors have added a paragraph that discusses this. However, I feel the argument that these parameters have not been addressed before a bit weak. Many other parameters (e.g. type of basal sliding law, different ocean parameterisation) could also have been chosen, so it would be good to state why you selected these particular parameters. Are they what you consider most important for sea-level rise or are they the most unconstrained? Please clarify this.
We decided to focus on aspects that were external to the model, (i.e., that mainly impact boundary and starting conditions) rather than aspects that are more model dependent. For this same reason, we refrained from calling them "parameters". We chose those aspects expecting that our findings regarding the different choices considered would be of use to a wider community. We add this justification, refining the last paragraph of the introduction (L79):

"The sensitivity of ice volume changes across glacial-interglacial time scales to model parameters was extensively explored by Albrecht et al. (2020). DeConto & Pollard (2016) carried out a large ensemble analysis for the LIG and the Pliocene, where parameters related to ice-shelf loss were constrained according to their ability to simulate target ranges of sea-level contribution. Simpler flow-line models have also been used to evaluate uncertainties in basal conditions (Gladstone et al., 2017) and flow-law parameters (Zeitz et al., 2020). Here, we perform five ensembles of experiments that focus on choices that are external to the numerical model, and could help guiding other modelling efforts on the choice of forcings and boundary conditions. We evaluate the impact on AIS volume and extent during MIS11c of the choice of proxy record (including their differences in signal intensity and structure), the choice of sea level reconstruction, and of uncertainties in assumptions regarding the geometry of the AIS at the start of MIS11c."

In the discussion, we further acknowledge the fact that we do not test for parameters such as the choice of sliding law (see response to comment 7) or different ocean parameterisations, and that they could impact our results (also mentioned in comment 6, and jointly answered in comment 7).

2. L181 Could you briefly say why you picked the EDC record for your ensemble here?
We use the EDC record because our thermal spin-up was also performed using data from this record. We justify our choice for the EDC as the reference ice core towards the end of Sect. 2.1 (L144):

"The EDC ice core was chosen for the thermal spinup and as forcing for the ensemble runs, because it spans the longest period among the three ice cores tested, while still providing a relatively high temporal resolution."

3. Sections 3.3. and 3.4 I think you could combine these two sections and simply state that they are not important for ice-sheet volume differences. Especially regarding section 3.3., volume differences are basically non-existent. At least I cannot see much difference at all in Fig. 4c. Similar arguments apply for Fig. 4e. Also, regarding the difference in floating ice volume at the beginning of the simulation, is that a result of forcing the same initial geometry with different sea-levels or why do they not start from the same value?
We agree that both sections were very short, and have merged them. Yes, the differences in floating ice volume at 425 ka are a result from forcing the same initial geometry with different sea-level reconstructions. As explained in the methods (paragraph starting at L138), they do not start from the same point in Figs. 4d and e because the starting point, under a common geometry, is at 425 ka (see Fig. S12 for an example for CFEN).

4. Caption of Fig. 8: I think that your statement: “Everywhere where Q bm >SMB, ice shelves are thinning” is questionable. I think if you think about a Lagrangian framework this is correct, but if you think about it in an Eulerian framework (more common in ice-sheet modelling), you could have higher basal melt rates than SMB, but still have local thickening because thicker is is being advected from upstream. This is also true if you consider entire ice shelves. You can have a negative budget from SMB and basal melting, but still gain mass, because thicker ice is being advected from upstream. So I recommend deleting this statement.

That is a very good point. We wanted to better guide the reader in terms of the magnitude of each process (because of the different sign conventions for each), but introduced a dubious statement. We have therefore removed it.

5. Paragraph starting at L304: I think it is not surprising that ocean melting does not do as much to the EAIS as it does to the WAIS. WAIS is a marine ice sheet with large shelves providing a lot of buttressing, while the EAIS has only small ice shelves which provide less buttressing and is predominately not marine-based.

We agree with this statement, and never intended to claim otherwise. We include the provided remarks and rephrased the end of the paragraph (L322):

“This observed tipping point at 412 ka also explains why the different ice-sheet configurations all follow the same trend from that moment onwards (Fig. 6), and why the evolution of WAIS and EAIS sea level contributions diverge. As ocean forcing becomes the main driver of ice-sheet retreat, it has a much larger impact on marine-based portions of the ice sheet. Around most of the EAIS (except for the Amery Ice Shelf), ice shelves are small and provide little buttressing. Hence, because most of the EAIS is grounded above sea level, its sub-shelf melting is not high enough to force grounding line retreat as strongly as in the WAIS. As a consequence, ice melt is dominated by surface ablation at the ice-sheet fringes (cf. hatched patterns in Fig. 8).”

6. Paragraph starting L340: I think here it would be good to add a qualifier that these numbers are for your particular ocean melt parameterisation. If you use a different paramerisation (e.g. linear relation), these thresholds would most certainly change as well.

We inserted the suggested qualifier while reformulating this paragraph as a response to comment 7.

7. L349-354 I do not find your different resolution experiments particularly helpful or well thought-through. So you test, 20, 16, and 15 km. This seems like a random choice of model resolutions. Especially the step from 16 to 15 km, is really small. If you want to do this more rigorously, you would have to do a convergence study (see for example Cornford et al. 2016 or Schannwell et al. 2018). I am not suggesting that you should do this, but I think you are definitely overstating your results and should be more cautious with your conclusions. In your Fig S15c, it is hard to tell because it is rather small, but you definitely see differences in sea level, even from 16 to 15 km. So your results are certainly to a degree mesh resolution dependent. From my own experience, if you increase your resolution to 10 km you start resolving ice streams a lot better and you would probably see more differences. I think that your different mesh resolutions are not fine enough to make the claim that results are mesh resolution independent. Rather, from the evidence that I see the contrary is the case. If you cannot run higher resolved simulations with your model because of computational restrictions that is fine, but it should be clearly stated.

In our reply to the first round of revisions, we noted that experiments at higher resolutions (12 and 10 km) were still underway and would be finished in time to be included in the following round of reviews. We have now included them to the supplement. Nevertheless, the reviewer brings a fair point regarding our statement about the sensitivity of our results to grid resolution. Based on the results of all higher resolution runs, and on comments 1 and 6, we rephrased with the following (L366):

"Moreover, AIS minimum extent and the timing of WAIS collapse are robust regardless of model resolution (Fig. S15). A set of simulations performed with several resolutions (from 20 to 10 km) showed virtually the same changes in ice-sheet extent, and modest variations in ice volume, which amount to a spread of 1.2 m s.l.e. in sea level contribution at 405 ka. Alternative sliding laws or sub-shelf melting parameterisations, for example using a linear dependence of sub-shelf melt to ocean thermal forcing, or applying a more physically realistic approach (e.g., Reese et al., 2018) were not ested, and could influence our results. For example, numerical modelling studies in which the WAIS did not collapse during MIS11 were acknowledged to be less sensitive to the ability of ocean temperatures to drive basal melting (e.g., Pollard & DeConto, 2009; Tigchelaar et al., 2019). Finally […]"

8. In you conclusions you state "...WAIS collapse was caused by the duration rather than the intensity of warming...", but in the discussion you say that both conditions have to be met and that even a shorter, but more intense ocean warming may also lead to WAIS collapse.
We understand how it might be confusing when looking at these two statements isolated. We state that both criteria need to be met, and when comparing to other studies we remark that the duration is specific to MIS11c. We then acknowledge that a stronger warming is also able to cause a WAIS collapse – which was most likely the case for LIG. We tried to make this distinction clearer by rewriting the paragraph starting at L353, also taking into account the changes presented above as a response to comment 7:

"The inferred critical warming of intermediate-depth ocean temperatures of 0.4 °C for MIS11c is close to the equilibrium model results in Garbe et al. (2020), but lower than results from Turney et al. (2020) for the AIS retreat during the LIG. While the former study shows a strong WAIS retreat is already possible for an ocean warming of 0.7 °C, the latter identifies a tipping point at 2 °C warming in ocean temperatures. In other interglacials, such as the LIG, WAIS collapse was triggered by ocean warming with a higher intensity and of shorter duration than during MIS11c (Dutton et al., 2015, Turney et al., 2020), since a stronger rate of warming can drive ice retreat at a much faster pace. Thus, WAIS collapse during MIS11c was likely attained because ocean temperatures exceeded a modest threshold for long enough (over 4 kyr)."

Figure comments:

9. The main point that needs improving are Figure sizes, Figs 3, 5, 7, 8 are plotting continent wide grounding lines. But each panel is so small, it is nearly impossible to see any differences. So please make each panel a lot bigger. If it helps, you could change to a 2x2 panel format. You can also use the full width of the page to make them bigger. For example, Figure sizes are much better in the supplement.
We changed all 4x1 figures to a 2x2 format, and increased font size from 16 to 18 pt. Regarding using the full width of the page, please note our comment at the beginning of this letter. If permitted, we would happily increase the other figure widths to full page as well.

10. Figure 8: I am sorry, but the hatching where basal melting is dominating SMB and vice versa is not visible. Even with a 300% zoom it is hard to see. You would probably have to have a zoom-in into the regions you talk about in the text in an additional Figure to see this.

We changed the figure to a 2x2 panel and increased font size to 18 pt. We note that the different patterns of hatching were perfectly visible at 126% zoom on a 22-inch monitor.

11. Please also make Figure 10 a lot bigger. The insets are so small I had to use 300% zoom to see everything. No chance on the printout.
We have enlarged the figure, as noted in the beginning of our response letter.

Technical corrections

12. L98 “controlled by a temporally fixed”
13. L278 “CFEN equivalent run” maybe
14. L304 “close to grounding lines”
We have used all suggestions above.

Reviewer 2’s comments

1. Line 10: add a word to “choices of sea level changes reconstructions”
Based on comment 2 from the Editor, we rephrased this passage as:
“In the case of a West Antarctic Ice Sheet collapse, which is the most probable scenario according to far-field sea level reconstructions, the range is further reduced to 6.7-8.2 m independently of the choices of external sea-level forcing and millennial-scale climate variability.”

2. Line 12: “choice of initial ice sheet configuration”
Added.

3. Line 13: Please reformulate “ reproduce its recorded sea level high stand” into “to match the recorded global sea level high stand”. This is because the sea level proxies record global signal and not only that of the Antarctic ice sheet and the big problem is to disentangle the individual signals.
That is a very good point, we have changed accordingly.

4. Lines 64-69: I would reformulate this paragraph first because atmospheric circulation over Antarctica is not homogeneous and this is what is evidenced by the different ice core records. Please include this statement in this paragraph. Also split the difference in temperature magnitude and the duration of the warmth. It will make this paragraph clearer.
We have reformulated the paragraph incorporating the mentioned suggestions. It now reads (L63):
“Constraints are also scarce for the MIS11c climate, and its heterogeneity is reflected in the ice core records. Reconstructions from different ice cores located in East Antarctica (circles in Fig. 1) show different histories regarding the evolution of atmospheric surface temperature. For example, the Vostok ice core surface air temperature reconstruction (Petit et al., 1999; Bazin et al., 2013) reveals a weak temperature peak (about 1.6°C above PI around 410 ka) compared to those of EPICA Dome C (EDC; over 2.7°C above PI around 406 ka, Jouzel et al., 2007) and Dome Fuji (DF; 2.5°C above PI around 407 ka, Uemura et al., 2018). The latter two ice-core records also present a peak-warming period of much longer duration (ca. 15 kyr compared to 7 kyr at Vostok).”

5. Line 81: “sea level changes reconstructions”
We rephrased the expression as (L86): *“the choice of sea level reconstruction”*

6. Line 86: "terrain-following vertical layers"
Added "vertical" to the phrase.

7. Equation (5): I don't see the exponential in the formula. Please correct it.
We believe the reviewer might have been confused. There is no explicit exponential (i.e., the letter *e*) in the function, but the terms "1-GI(t)" and "GI(t)" are the exponents of P_{PI} and P_{LGM} respectively.

8. In sections 2.3.1 to 2.3.3: please refer to panels in Figure 2.
We added to the sentence before the last in section 2.3.1 (L177): "(orange and black dashed lines in Fig. 2b respectively)"

We already mention Fig. 2c in the last sentence of section 2.3.3 (L196).

9. Figure 4 caption: panel e) description is missing.
Thanks for spotting this, we have rephrased the part referring to the SLEN ensemble:
"Panels d and e show floating and total ice volumes (in 10^6 km^3), respectively, for the SLEN sea-level forcing reconstructions forced by EDC GI"

10. Figure 5: is definitely too small. I would suggest to put two panels by raw in stead of all panels on the same raw. Also I would change the yellow color with orange or something more visible.
We changed it to a two-row figure as also suggested by Reviewer 1 in his comment 9. Regarding the color, we have changed the plotting order, making it significantly easier to see where the grounding lines differ.

11. Section 3.3: so actually SICOPOLIS grounding line scheme is highly insensitive to sea level changes… This is not the case of other ice sheet models. This is an important point also to be somehow discussed at the end of this paper in terms of uncertainties in those simulated volume and scenarios due to the model physics.
We had already mentioned our model insensitivity to sea-level forcing in the paragraph starting at line 360. We aimed to show that, compared to other external forcings, it played a minor role – which is similar in other models. Also, we acknowledge that this could be a limitation of the spatial resolution used (L360):
"Despite differences in the model sensitivity to ocean temperature, our results support those of Tigchelaar et al. (2019) and Albrecht et al. (2020) regarding the minor role that variations in sea level play in driving ice-sheet retreat compared to other external forcings. Although the coarse treatment of the grounding lines could have had an influence on the seeming insensitivity of our experiments to sea-level uncertainties, other models of similar resolution which apply different sub-grid parameterisations to the grounding lines yield similar results (Tigchelaar et al., 2019, Sutter et al., 2019, Albrecht et al., 2020). Hence, while this caveat must be taken into consideration, it does not appear to have influenced our results dramatically."

12. Figure 7 and figure 8: same as for Figure 5, perhaps better in two raws than all panel on a single raw. It is up to you.
We have made the same changes as in the response to comment 10 (i.e., Fig. 5).

13. Line 283: are you sure "height of MIS11c" is a correct English wording? Sounds a bit odd. Perhaps "peak of MIS11c" would be better? Not sure…just asking.
We believe it is correct, but agree that it is not standard. Thus, we have changed it to *"peak of MIS11c"*

14. Line 232 and 354-368: I think that you could simplify quite a lot the paragraph here. The main point is that LR04 is a stack of 57 globally distributed sediment cores with very few cores in the Southern Ocean…Thus LR04 represents a global averaged signal and not a local polar signal, in contrast with all the other ice core records…To me this is an important difference that also explains most of the timing and magnitude difference you see in all your simulations.

We reorganised the paragraph, precisely focusing on why LR04 misses the warming recorded by the ice cores (L375):

"The LR04 reconstruction is composed of a stack of 57 globally-distributed ocean sediment cores (Lisiecki & Raymo, 2005), with a strong deficit over the Southern Ocean. In the Nordic Seas, paleoceanographic records indicate that the ocean was colder than present during MIS11 (Bauch et al., 2000; Kandiano et al., 2016; Doherty & Thibodeau, 2018). Colder ocean temperatures in the Northern Hemisphere explain why LR04 shows oxygen isotopic values similar to the Holocene during MIS11c (Lisiecki & Raymo, 2005) despite the geological evidence that there was a contribution to higher-than-Holocene sea levels from both Greenland and Antarctica (Scherer et al., 1998; Raymo & Mitrovica, 2012). Hence, the inclusion of many Northern Hemisphere records in the LR04 stack explains why it fails to capture the Antarctic warming during MIS11c seen in the ice cores, and the differences in timing compared to them. This also helps explain why the different criteria adopted for changing its scaling procedure had little effect on the results (Fig. 4b). A possible way of circumventing this problem could be to adopt a similar scaling approach to Sutter et al., (2019), who combined the LR04 stack and EDC ice-core temperature records, which, in their study, also led to WAIS collapse during MIS11c."

15. From the supplementary CCSM3 tends to be colder than what is suggested by proxies: don't you think that it might anyway affect your spin-up and thus velocities and thus makes EAIS marine-based basin particularly insensitive to oceanic warming?

During the thermal spin-up we do not use the CCSM3 LGM climate. As stated in the methods (last paragraph of Sect. 2.1), we use the surface temperature from EDC as an anomaly to PI climate. The CCSM3 LGM climate is only used to produce the anomalies that are modulated by the GI during the 425-394 ka experiments.

16. Supplementary figure S14: Another interesting point this that comparing the dD and the derived dT° from your glacial index, it seems that if that glacial index would not overestimate the warmth at DF, simulations using DF would perhaps not lead to a WAIS collapse or perhaps delay it a lot. Just something to keep in mind. Here the fact the your glacial indices overestimate systematically the proxies results from the fact that LGM is too cold in the CCSM3 climate forcing, while PI is acceptable. This is a direct impact of your CCSM3 bias, despite the fact that you say in the supplementary that it would not affect your result. I think it does. So perhaps, just a sentence in the discussion of the main manuscript about this would be appreciated.

The reviewer offers two good points, which we address below.

First, we mention the cold bias of CCSM3, which does not appear to be fully addressed by the lapse-rate correction. We add after the first mention of Fig. S14 (L304):

"[…] (Supplementary Fig. S14). This is most likely due to the cold bias in CCSM3, which persisted despite the lapse-rate correction applied. Nevertheless, Vostok's GI-reconstructed temperature peak matches the peak observed in DF […]"

Second, we agree that if the GI-derived temperatures were not overestimated, the DF experiment would resemble the "extended Vostok-peak warmth" experiment, which further strengthens our point that the duration of warmer-than-present temperatures played a crucial role during MIS11c. We added the following to the discussion (L348):

“Considering that the temperature peak reconstructed by the Vostok GI is the closest to the δD - derived temperature peaks in DF and EDC (Fig. S14), a more prolonged warming as seen in the DF and EDC ice core seems to be a crucial condition for the modelled WAIS drawdown during MIS 11c. For example, if the GI-derived temperature for DF was not overestimated, and had its peak value close to its isotope-derived value, the response would likely resemble the experiment where Vostok-peak conditions were kept constant from 410 ka onwards.”

References mentioned in this letter that were not previously included in the manuscript:

DeConto, Robert M., and Pollard, D. "Contribution of Antarctica to past and future sea-level rise." Nature 531 (2016): 591-597.

Gladstone, Rupert M., et al. "Marine ice sheet model performance depends on basal sliding physics and sub-shelf melting." The Cryosphere 11.1 (2017): 319-329.

Reese, R., Albrecht, T., Mengel, M., Asay-Davis, X., and Winkelmann, R. "Antarctic sub-shelf melt rates via PICO", The Cryosphere, 12, (2018): 1969–1985.

Zeitz, Maria, Levermann, A., and Winkelmann, R. "Sensitivity of ice loss to uncertainty in flow law parameters in an idealized one-dimensional geometry." The Cryosphere 14.10 (2020): 3537-3550.

Sensitivity of the Antarctic ice sheets to the peak warming of Marine Isotope Stage 11

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Abstract. Studying the response of the Antarctic ice sheets during periods when climate conditions were similar to the present can provide important insights into current observed changes and help identify natural drivers of ice sheet retreat. In this context, the Marine Isotope Substage 11c (MIS11c) interglacial offers a suitable scenario, given that during its later portion, orbital parameters were close to our current interglacial. ~~In particular, ice~~ Ice core data indicate that warmer-than-present temperatures lasted for longer than during other interglacials~~, and~~ . However, the response of the Antarctic ice sheets and their contribution to sea level rise remain unclear. We explore the dynamics of the Antarctic ice sheets during this period using a numerical ice-sheet model forced by MIS11c climate conditions derived from climate model outputs scaled by three glaciological and one sedimentary proxy records of ice volume. Our results indicate that the East and West Antarctic ice sheets contributed with 3.2 to 8.2 m to the MIS11c sea level rise. In the case of a West Antarctic Ice Sheet collapse, which is the most probable scenario according to far-field sea level reconstructions, the range is further reduced to 6.7–8.2 m ~~,~~ independently of the choices of ~~sea level reconstructions~~ external sea-level forcing and millennial-scale climate variability. Within this latter range, the main source of uncertainty arises from the sensitivity of the East Antarctic Ice Sheet to a choice of initial ice sheet configuration. We found that the warmer regional climate signal captured by Antarctic ice cores during peak MIS11c is crucial to ~~reproduce its recorded~~ match the recorded global sea level highstand. Furthermore, we show that a modest 0.4 °C oceanic warming at intermediate depths leads to a collapse of the West Antarctic Ice Sheet if sustained for at least 4 thousand years.

1 Introduction

~~Marine Isotope Substage 11c (hereafter MIS11c), was a remarkable interglacial because it lasted~~ Lasting for as much as 30 thousand years (kyr), between 425 and 395 thousand years ago ~~(ka; Lisiecki and Raymo, 2005; Tzedakis et al., 2012), thus making it~~ (ka), Marine Isotope Substage 11c (hereafter MIS11c) was the longest interglacial of the Quaternary (ka; Lisiecki and Raymo, 200 . It also marked the transition from weaker to more pronounced glacial-interglacial cycles (EPICA Community Members, 2004). Its long duration is attributed to a modulation of the precession cycle, resulting in CO_2 levels that were high enough to suppress the cooling of the climate system due to the low eccentricity and thus reduced insolation (Hodell et al., 2000). Moreover, ocean sediment cores (e.g., Hodell et al., 2000) and climate models (e.g., Rachmayani et al., 2017) show that the MIS11c global overturning circulation was at an enhanced state, resulting in asynchronous warming of the southern and northern high latitudes (i.e., they did not reach their warming peak at the same time; Steig and Alley, 2002). However, Dutton et al. (2015) point out that climate modelling experiments with realistic orbital and greenhouse gas forcings fail to fully capture this MIS11c warming despite the fact that orbital parameters were almost identical to Present Day (PD) during its late stage (cf. EPICA Community Members, 2004; Raynaud et al., 2005). Earlier studies (e.g., Milker et al., 2013; Kleinen et al., 2014) have shown that climate models also tend to underestimate climate variations during MIS11c, for which ice core reconstructions show the mean annual atmospheric temperature over Antarctica to have been about 2 °C warmer than Pre-Industrial (PI) values.

A better understanding of the climate dynamics during Quaternary interglacials, especially those that were warmer than today, is critical because they can help assess Earth's natural response to future environmental conditions (Capron et al., 2019). Among these periods, MIS 5e (also referred to as the Eemian, Last Interglacial, or LIG; Shackleton et al., 2003) was originally proposed to be a possible analogue for the future of our current interglacial (Kukla, 1997). More recently, MIS11c has been considered another suitable candidate, since its orbital conditions were closest to PD (Berger and Loutre, 2003; Loutre and Berger, 2003; Raynaud et al., 2005). Furthermore, ice core evidence indicates that Termination V (i.e., the deglaciation that preceded MIS11) was quite similar to the last deglaciation in terms of rates of change in temperature and greenhouse gas concentrations (EPICA Community Members, 2004). The unusual length of ~~this interglacial~~ MIS11c and a transition to stronger glacial-interglacial cycles seen in the subsequent geological record may have been triggered by a reduced stability of the West Antarctic Ice Sheet (WAIS, Fig. 1). The latter may have been due to the cumulative effects of the ice sheet lowering its bed (Holden et al., 2011), which in turn provided a positive climate feedback (Holden et al., 2010). The long duration of MIS11 was also shown to be a key condition to triggering the massive retreat of the Greenland Ice Sheet (GIS; Robinson et al., 2017). Elucidating the response of the Antarctic ice sheets (AIS) to past interglacials can also help identify various triggers of ice sheet retreat. This is because each interglacial has its unique characteristics: for example, while MIS11c was longer than the LIG, the latter was significantly warmer (Lisiecki and Raymo, 2005; Dutton et al., 2015).

The MIS11c history of Antarctica is less constrained than that of Greenland (e.g., Willerslev et al., 2007; Reyes et al., 2014; Dutton et al., 2015; Robinson et al., 2017). Whereas Raymo and Mitrovica (2012) consider that the WAIS had collapsed and that the East Antarctic Ice Sheet (EAIS, Fig. 1) provided a minor contribution based on their estimate of MIS11c global sea levels of 6 to 13 m above PD, studies directly assessing the AIS response have been elusive. For example, sedimentary evidence has been inconclusive regarding the possibility of a collapse of the WAIS during some Quaternary interglacials (Hillenbrand et al., 2002, 2009; Scherer, 2003), and evidence for the instability of marine sectors of the EAIS has only recently been provided (Wilson et al., 2018; Blackburn et al., 2020). Counter-intuitively, the dating of onshore moraines in the Dry Valleys to MIS11c has been used to indirectly support regional ice sheet retreat (Swanger et al., 2017). Swanger et al. (2017) argue that ice sheet retreat in the Ross Embayment provided nearby open-water conditions and therefore a source of moisture and enhanced precipitation, fueling local glacier growth. Previous numerical modelling experiments that encompass MIS11 also lack a consensus regarding AIS volume changes. For example, Sutter et al. (2019) report an increased ice volume variability from MIS11 onwards, caused by stronger atmospheric and oceanic temperature variations, while Tigchelaar et al. (2018) only obtained significant volume changes during the last 800 kyr when increasing their ocean temperatures to values as high as 4 °C.

Conversely, de Boer et al. (2013) report higher sea level contributions during MIS 15e, 13, and 9, and weaker contributions during MIS 11c and 5e. Among the past interglacials, the LIG and Pliocene are considered to be the closest analogues to MIS11c, and studies acknowledge the possibility of a WAIS collapse in both periods (e.g., Hearty et al., 2007; Naish et al., 2009; Pollard and DeConto, 2009). However, Pliocene model results were shown to be highly dependent on the choice of climate and ice-sheet models (de Boer et al., 2015; Dolan et al., 2018).

~~The~~ Constraints are also scarce for the MIS11c climate~~is also loosely constrained~~, and its heterogeneity is reflected in the ice core records. Reconstructions from different ice cores ~~do not fully agree on how Antarctic surface air temperatureevolved during this period~~located in East Antarctica (circles in Fig. 1) show different histories regarding the evolution of atmospheric surface temperature. For example, the Vostok ice core surface air temperature reconstruction (Petit et al., 1999; Bazin et al., 2013) reveals a ~~much shorter and weaker period of peak warming~~ weak temperature peak (about 1.6 °C ~~higher than~~ above PI around 410 ka) ~~than that inferred from the~~compared to those of EPICA Dome C ~~(EDC; Jouzel et al., 2007)~~ (EDC; over 2.7 °C above PI aro and Dome Fuji ~~(DF; Uemura et al., 2018) ice cores~~(DF; 2.5 °C above PI around 407 ka, Uemura et al., 2018). The latter ~~show a~~ two ice-core records also present apeak-warming period of much longer duration (ca. 15 kyr ~~) of warmer-than-present temperatures, peaking at over 2.7 C above PI around 406 ka for EDC, and 2.5 C above PI at about 407 ka for DF (locations are shown in Fig. 1~~compared to 7 kyr at Vostok).

As detailed, many modelling studies have investigated AIS responses over time periods that include MIS11. However, so far none has focused specifically on this period. Given the scarce information for MIS11 and conflicting constraints on how Antarctica responded to this exceptionally long interglacial (Milker et al., 2013; Dutton et al., 2015), we here focus on MIS11c, the peak warming period between 420 and 394 ka. Our aim is to reduce the current uncertainties in the AIS behaviour during MIS11c, addressing the following questions:

1. How did the AIS respond to the warming of MIS11c? More specifically, what are the uncertainties in the AIS minimum configuration, timing and potential sea level contribution?

2. What was the main driver of the changes in the AIS volume? Was it warming duration, peak temperature, changes in precipitation, or changes in the oceanic forcing?

~~For this purpose~~The sensitivity of ice volume changes across glacial-interglacial time scales to model parameters was extensively explored by Albrecht et al. (2020). DeConto and Pollard (2016) carried out a large ensemble analysis for the LIG and the Pliocene, where parameters related to ice-shelf loss were constrained according to their ability to simulate target ranges of sea-level contribution. Simpler flow-line models have also been used to evaluate uncertainties in basal conditions (Gladstone et al., 2017) and flow-law parameters (Zeitz et al., 2020). Here, we perform five ensembles of ~~numerical simulations of the AIS evolution and focus on aspects that remain unaddressed by previous studies~~experiments that focus on choices that are external to the numerical model, and could help guiding other modelling efforts on the choice of forcings and boundary conditions. We evaluate the impact on ~~resulting ice~~ AIS volume and extent during MIS11c of the choice of proxy ~~records~~ record (including their differences in signal intensity and structure), the choice of sea level reconstruction, and of uncertainties in assumptions regarding the geometry of the AIS at the start of MIS11c.

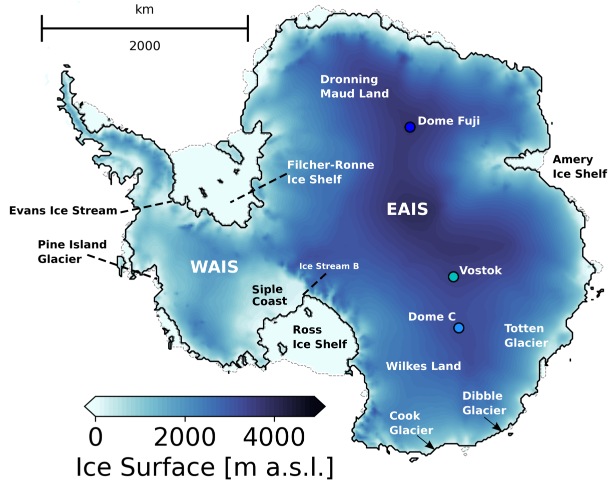

Figure 1. Surface topography of the AIS at the start of our core experiments (425 ka), based on a calibration against Bedmap2 (Fretwell et al., 2013, , see Sect. 2.1). Locations mentioned in the text are showcased, including the drilling sites of the ice (circles) and sediment (red diamonds) cores on and around Antarctica, respectively.

2 Methods

2.1 Ice-sheet model

For our experiments we employ the 3D thermomechanical polythermal ice-sheet model SICOPOLIS (Greve, 1997; Sato and Greve, 2012) with a 20 km horizontal grid resolution and 81 terrain-following vertical layers. It uses the one-layer enthalpy scheme of Greve and Blatter (2016), which is able to correctly track the position of the cold-temperate transition in the thermal structure of a polythermal ice body.

The model combines the Shallow Ice Approximation (SIA) and Shelfy Stream Approximation (SStA) using (c.f. Bernales et al., 2017a, Eq. 1)

$$\mathbf{U} = (1 - w) \cdot \mathbf{u}_{\mathrm{sia}} + \mathbf{u}_{\mathrm{ssta}}, \tag{1}$$

where $\mathbf{U}$ is the resulting hybrid velocity, $\mathbf{u}_{\mathrm{sia}}$ and $\mathbf{u}_{\mathrm{ssta}}$ are the SIA and SStA horizontal velocities, respectively, and w is a weight computed as

$$w(|\mathbf{u}_{\mathrm{ssta}}|) = \frac{2}{\pi}\arctan\left(\frac{|\mathbf{u}_{\mathrm{ssta}}|^2}{u_{\mathrm{ref}}^2}\right), \tag{2}$$

where the reference velocity u_{ref} is set to 30 ma^{-1}, marking the transition between slow and fast ice. This hybrid scheme reduces the contribution from SIA velocities mostly in coastal areas of fast ice flow and heterogeneous topography, where this approximation becomes invalid. Basal sliding is implemented within the computation of SStA velocities as a Weertman-type law (cf. Bernales et al., 2017a, Eqs. 2–6). The amount of sliding is controlled by a temporally fixed, spatially varying map of friction coefficients that was iteratively adjusted during an initial present-day equilibrium run (cf. Pollard and DeConto, 2012b), such that the grounded ice thickness matches the present-day observations from Bedmap2 (Fretwell et al., 2013) as close as possible. Sliding coefficients in sub-ice shelf and ocean areas are set to $10^5\,\mathrm{m\,a^{-1}\,Pa^{-1}}$, representing soft, deformable sediment, in case the grounded ice advances over this region. The initial bedrock, ice base, and ocean floor elevations are also taken from Bedmap2. Enhancement factors for both grounded and floating ice are set to 1, based on sensitivity tests in Bernales et al. (2017b). This choice provides the best match between observed and modelled ice thickness for this hybrid scheme, similar to the findings in Pollard and DeConto (2012a).

Surface mass balance is calculated as the difference between accumulation and surface melting. The latter is computed using a semi-analytical solution of the positive degree day (PDD) model following Calov and Greve (2005). Near-surface air temperatures entering the PDD scheme are adjusted through a lapse rate correction of $8.0\,^{\circ}\mathrm{C\,km^{-1}}$ to account for differences between the modelled ice sheet topography and that used in the climate model from which the air temperatures are taken. For the basal mass balance of ice shelves, we use a calibration scheme of basal melting rates developed in Bernales et al. (2017b) to optimise a parameterisation based on Beckmann and Goosse (2003) and Martin et al. (2011) that assumes a quadratic dependence on ocean thermal forcing (Holland et al., 2008; Pollard and DeConto, 2012a; Favier et al., 2019). This optimised parameterisation is able to respond to variations in the applied Glacial Index (GI, Sect. 2.2) forcing. A more detailed description of this parameterisation is given in Sect. 1 of the supplementary material. In our experiments, we prescribe a time lag of 300 years for the ocean response to GI variations, which is considered the most likely lag in response time of the ocean compared to the atmosphere in the Southern Ocean (Yang and Zhu, 2011). At the grounding line, the basal mass balance of partially floating grid cells is computed as the average melting of the surrounding, fully floating cells, multiplied by a factor between 0 and 1 that depends on the fraction of the cell that is floating. This fraction is computed using an estimate of the sub-grid grounding line position based on an interpolation of the current, modelled bedrock and ice-shelf basal topographies. At the ice shelf fronts, calving events are parameterised through a simple thickness threshold, where ice thinner than 50 m is instantly calved away.

Glacial isostatic adjustment is implemented using a simple elastic lithosphere, relaxing asthenosphere (ELRA) model, with a time lag of 1 kyr and flexural rigidity of $2.0 \times 10^{25}\,\mathrm{N\,m}$, which Konrad et al. (2014) found to best reproduce the results of a fully-coupled ice sheet–self-gravitating viscoelastic solid Earth model. The geothermal heat flux applied at the base of the lithosphere is taken from Maule et al. (2005) and is kept constant. All relevant parameters used in the modelling experiments are listed in Table 1.

Sea-level contribution at a given time step is computed in SICOPOLIS as the difference in total ice volume above flotation between the ice sheet at the time step and the spun-up Pre-Industrial ice sheet. When computing ice volume, differences in bedrock elevation between the two ice sheets are accounted for by using a common reference bedrock elevation in all time steps. We also correct for the projection effect on the horizontal grid area.

All ensembles cover a period from 420 to 394 ka. After the calibration for basal sliding mentioned above, we initialise the AIS by performing a thermal spin-up over a period of 195 kyr from 620 to 425 ka, i.e., apply a transient surface temperature

Table 1. Main parameters used in the experiments.

Parameter	Name	Value	Units
$\mathrm{E}_{\mathrm{grounded}}$	Enhancement factor (grounded ice)	1	
$\mathrm{E}_{\mathrm{floating}}$	Enhancement factor (ice shelves)	1	
n	Glen's Flow Law exponent	3	
p	Weertman's Law p exponent	3	
q	Weertman's Law q exponent	2	
τ	ELRA model time lag	1	kyr
D	ELRA model flexural rigidity	2.0×10^{25}	$\mathrm{N\,m}$
γ_{lr}	Lapse rate correction	8.0	$^{\circ}\mathrm{C\,km^{-1}}$
S_0	Sea water salinity	35	
ρ_{sw}	Sea water density	1028	$\mathrm{kg\,m^{-3}}$
ρ_{ice}	Ice density	910	$\mathrm{kg\,m^{-3}}$
c_{p_0}	Ocean mixed layer specific heat capacity	3974	$\mathrm{J\,kg^{-1}\,K^{-1}}$
γ_T	Thermal change velocity	10^{-4}	$\mathrm{m\,s^{-1}}$
L_i	Latent heat of fusion	3.35×10^{5}	$\mathrm{J\,kg^{-1}\,K^{-1}}$

signal from the EDC ice core (Jouzel et al., 2007) as an anomaly to our PI climate (described in the next section) while keeping the ice sheet geometry constant at our previously calibrated Bedmap2-based configuration. We then let the AIS freely evolve for 5 kyr, between 425 and 420 ka, applying transient GI forcing during the relaxation period (Fig. S12). We chose 425 ka as the starting point for relaxation because it is when the MIS11c oxygen isotope values in the EDC ice core are closest to PI. In summary, we ignore the first 5 kyr (425–420 ka) to avoid a shock from suddenly letting the ice-sheet topography freely evolve at the start of our period of interest. Figure 1 shows the thermally spun-up ice sheet configuration at 425 ka, from which the ~~relaxation~~ simulations start. The EDC ice core was chosen for the thermal spin-up and as forcing for the ensemble runs because it spans the longest period among the three ice cores tested, while still providing a relatively high temporal resolution.

2.2 Climate forcing and core experiments

In an effort to assess similarities and differences in existing paleoclimate reconstructions, and regional differences in the ice-core records, we perform an ensemble of simulations where each member is forced by a GI (Eq. 3) derived from δD from ice cores, or δ^{18}O from the LR04 stack of deep-sea sediment cores (Fig. 2a; Petit et al., 2001; EPICA Community Members, 2004; Lisiecki and Raymo, 2005; Uemura et al., 2018). Since an ensemble of fully coupled climate-ice sheet model runs over 26 kyr is at present computationally challenging, an evaluation of possible scenarios for the peak-temperature response during MIS11c based on the paleoclimate signals from different ice sheet sectors can be a cheaper, yet effective approach. The GI method is a way of weighting the contributions from interglacial (PI) and full glacial (Last Glacial Maximum; LGM) average

Table 2. Ice and sediment cores reference values used in Eq. (3), together with the age (in thousand years before present; ka) from which the reference values were obtained. The respective age models of each core, and their references, are listed.

Record	Type (isotope)	δX_{PI} [‰]	δX_{LGM} [‰]	Age (ka)	Age model	Reference
EDC	Ice (δD)	-397.4	-449.3	24.0	EDC3	EPICA Community Members (2004)
DF	Ice (δD)	-425.3	-469.5	22.8	AICC2012	Uemura et al. (2018)
Vostok	Ice (δD)	-440.9	-488.3	24.4	GT4	Petit et al. (2001)
LR04	Sediment (δ^{18}O)	3.23	4.99	20.0	LR04	Lisiecki and Raymo (2005)

states. It does so by rescaling a variable curve (usually temperature or isotope reconstructions from an ice or sediment record) based on reference PI and LGM values, which consider PI climate as $\mathrm{GI} = 0$ and LGM climate as $\mathrm{GI} = 1$ (Eq. 3):

$$GI(t) = \frac{\delta \mathrm{X}(t) - \delta \mathrm{X}_{\mathrm{PI}}}{\delta \mathrm{X}_{\mathrm{LGM}} - \delta \mathrm{X}_{\mathrm{PI}}} \tag{3}$$

Where t is time, and X is deuterium for the ice cores or ^{18}O for sediment cores. The value for $\delta \mathrm{X}_{\mathrm{PI}}$ was obtained as the average of the last 1000 years before 1850 CE, while $\delta \mathrm{X}_{\mathrm{LGM}}$ was taken as the minimum and maximum value for $\delta \mathrm{D}$ and δ^{18}O, respectively, between 19 and 26.5 ka (cf. Clark et al., 2009; Clason et al., 2014). For our two reference climate states (i.e., PI and LGM), we use the Community Climate System Model version 3 (CCSM3) PI time slice in Rachmayani et al. (2016), and the LGM time slice in Handiani et al. (2013), which used identical model versions and were run on the same platform. A brief assessment of the model biases against PD data is provided (Sects. 2 and 3 of the supplementary material). The atmospheric and ocean temperature (T) fields at time t are reconstructed based on their respective PI and LGM reference fields (T_{PI} and $\mathrm{T}_{\mathrm{LGM}}$ respectively) using (see also Fig. S13):

$$T(t) = T_{\mathrm{PI}} + GI(t) \cdot (T_{\mathrm{LGM}} - T_{\mathrm{PI}}) \tag{4}$$

while precipitation is given by an exponential function to prevent negative values and to ensure a smooth transition between the PI and LGM states:

$$P(t) = {P_{\mathrm{PI}}}^{1-\mathrm{GI}(t)} \cdot {P_{\mathrm{LGM}}}^{\mathrm{GI}(t)} \tag{5}$$

The PI and LGM reference values (including the reference ages for the latter) for the three ice cores and the LR04 stack are summarised in Table 2, together with their respective age models. The ensemble of simulations forced by different GI curves (Climate Forcing ENsemble, CFEN) constitutes our core experiments.

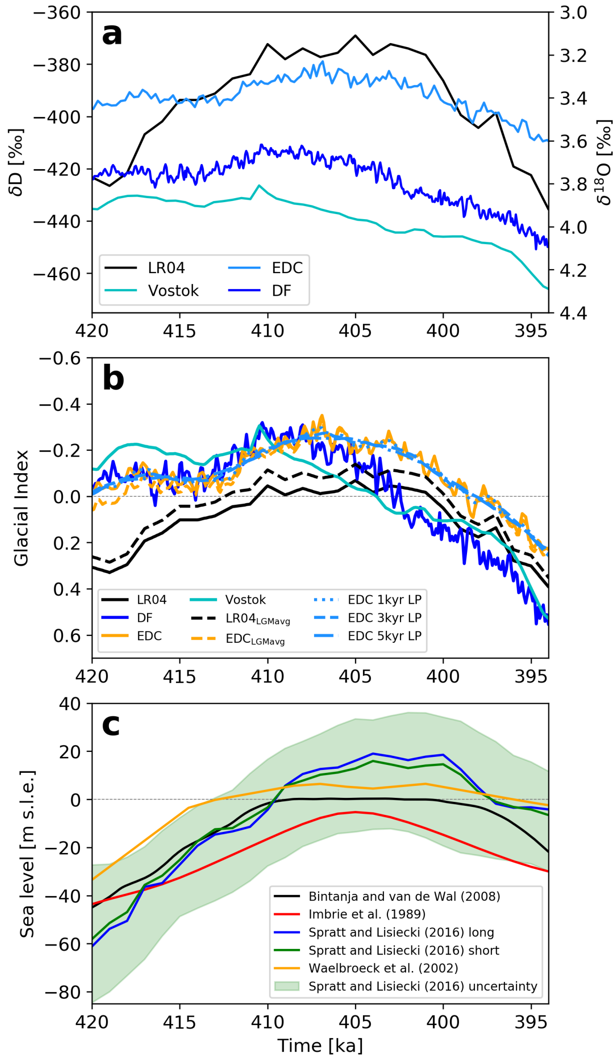

Figure 2. Reconstructions used in this study: (a) LR04 δ^{18}O (black) and Vostok, Dome C (EDC), and Dome Fuji (DF) ice-core δD [‰]; (b) resulting Glacial Indices from the reconstructions in (a) (cf. Sect. 2 and Table 3 for the legends); (c) global mean sea level anomaly relative to PI (meter sea level equivalent, m s.l.e.).

2.3 Sensitivity experiments

2.3.1 Sensitivity to the GI scaling

Because different approaches have been used to transform the isotope curves into a GI, we assess the sensitivity to the choice of the scaling procedure by performing an additional scaling using another reference value for δX_{LGM}. In the new scaling procedure, δX_{LGM} is the average (between 19 ka and 26.5 ka) rather than the peak value. We compare the effects of using these two procedures when applied to the EDC ice core δD and the LR04 stack δ^{18}O records (orange and black dashed lines in Fig. 2b respectively). We call this ensemble the Scaling Sensitivity ENsemble (SSEN)."

2.3.2 Sensitivity to millennial-scale variability

Given the different temporal resolutions of climate records, lower-resolution reconstructions such as LR04 and Vostok might not capture the impact of millennial variability or shorter events, as do EDC and DF (Fig. 2a). Thus, we assess the potential effects of record data resolution and millennial (or shorter) time scale variability by applying 1, 3, and 5 kyr low-pass filters to the EDC ice core GI and forcing our model with the resulting smoothed GI curves (light blue lines in Fig. 2b). We then compare these three simulations to the original EDC-derived ice sheet history, and call this ensemble the Resolution Sensitivity Ensemble (RSEN).

2.3.3 Sensitivity to sea level

Mean sea level plays an important role in determining the flotation of the ice sheet and the stresses at its marine margins. Uncertainties in global mean sea level reconstructions are therefore a significant concern, and several studies have indeed focused on improving their estimates (e.g., Imbrie et al., 1989; Waelbroeck et al., 2002; Bintanja and van de Wal, 2008; Spratt and Lisiecki, 2016, Fig. 2c). We evaluate the effect of using a particular sea level reconstruction on the evolution of the AIS by running an ensemble of simulations with EDC-derived GI, where each member uses a different sea level reconstruction. Sea level curves included in this ensemble are three of the reconstructions presented by Spratt and Lisiecki (2016), termed "long" (i.e., uses records that extend as far back as 798 ka), "short" (uses records that extend at least until 430 ka), and the "upper uncertainty boundary" from their records, because we consider their lower uncertainty boundary to be satisfactorily covered by SPECMAP (Imbrie et al., 1989), which we include. We also include in the analysis the reconstructions from Bintanja and van de Wal (2008) and from Waelbroeck et al. (2002). All these records are presented in Fig. 2c, and we call this ensemble, where we test different sea level reconstructions, the Sea Level Sensitivity Ensemble (SLSEN).

2.3.4 Sensitivity to the choice of initial ice sheet geometry

Similar studies that assess AIS changes over glacial and interglacial cycles often adopt a PI or PD starting geometry (e.g., Sutter et al., 2019; Tigchelaar et al., 2019; Albrecht et al., 2020). We have followed the same approach in our CFEN experiments (see Sect. 2.2). Although ~~this~~ the similarity to the modern AIS configuration has been loosely inferred from sedimentary (Capron et al., 2019) and ice-core (EPICA Community Members, 2004) proxy records, to our knowledge there is no direct evidence to support this claim (e.g., Swanger et al., 2017). Hence, we also perform an ensemble of simulations starting from different ice sheet geometries. This allows for an evaluation of the influence of an initial AIS configuration at 420 ka on its modelled retreat and advance (including possible thresholds), and provides an uncertainty envelope in its potential sea level contribution based on this criterion. We call this the Starting Geometry Sensitivity ENsemble (SGSEN), and its three unique geometries are forced with the ice-core reconstructed climate forcings tested in CFEN.

In order to create a representative range of initial geometries at 420 ka, we use a common starting geometry, but vary the relaxation time. For this purpose, we first create an ancillary geometry by perturbing the thermally spun-up AIS with a constant LGM climate (air temperature and precipitation rates) and no sub ice-shelf melting over a 5 kyr period. The resulting ancillary ice sheet (which has an extent that sits between PI and LGM configurations) is then placed at 420, 425 and 430 ka and runs transiently (following the respective GIs) until 394 ka. This creates a representative range of starting geometries at 420 ka (Fig. 3), and each initial ice sheet geometry is labelled gmt1 to gmt3 (Fig. 3a-c; shortest relaxation is gmt1, longest is gmt3). The gmt1 initial topography is generally more extensive and thinner than the control. Its grounding line advanced at the southern margin of the Filcher-Ronne Ice Shelf and at Siple Coast, but the ice sheet interior is on average 200 m thinner than the control and up to 500 m thinner across particular regions such as the dome areas of the WAIS and Wilkes Land (Dome C). It is, however, about 200 m thicker at its fringes, which results in a gentler surface gradient towards the ice sheet margins. The gmt2 initial topography is less than 100 m thinner than control over the EAIS interior, and about 100 m thicker over the WAIS interior and at the EAIS margins. Finally, the gmt3 initial topography is overall thicker than control, though not by more than 100 m except at the western side of the Antarctic Peninsula and the WAIS margins, where some regions are up to 300 m thicker (Fig. 3c). Table 3 summarises all experiments described in this section.

3 Results

3.1 Climate forcing reconstructions

Considering the four adopted isotope curves (Fig. 2a,b), although similar at first sight, the GI reconstructions are different from one another, and therefore offer a range of modelled ice-sheet responses. The LR04 GI reconstruction is generally colder, showing conditions warmer than PI only for the warmest period of MIS11c (i.e., between ca. 410 ka and 400 ka). Consequently, it does not show a peak warming as strong as the other reconstructions (Fig. 2b). Although the ice cores have similar ranges in GI values and similar overall aspects of the curves (and good covariance between EDC and DF; Uemura et al., 2018), they differ in key aspects. The Vostok reconstruction starts at a warmer state than the others at 420 ka, has a modest peak warming at 410 ka, and then consistently declines towards a colder state (crossing the GI = 0 line at about 404 ka). The EDC reconstruction shows a mildly warmer-than-PI state at 420 ka, which persists until about 412 ka. Subsequently, the peak warming starts and persists (in a slightly warmer state than reconstructed with Vostok after 410 ka) until 397 ka. Its rate of decline after 404 ka is similar to the Vostok and LR04 curves, although it is in a warmer state. Finally, the DF reconstruction is somewhere in-between the other two ice cores (Fig. 2b). It shows quite stable conditions at the start (i.e., no pronounced warming), rising to a rather pronounced warming peak similar in structure to the EDC reconstruction, but peaks at 410 ka, similar to the Vostok curve. Finally, its rate of decline is similar to the other cores and so it crosses PI values (GI = 0) later than the Vostok but earlier than the EDC curves, between 404 ka and 403 ka.

The ice sheet history for MIS11c using the LR04 forcing is clearly different from the others. The ice sheet loses less than a third of its volume compared to the other CFEN members, and becomes smaller than PD for a duration of 9 kyr, while the others are consistently below PD levels (Fig. 4a). It is worth reminding that, in contrast to other members of CFEN, the

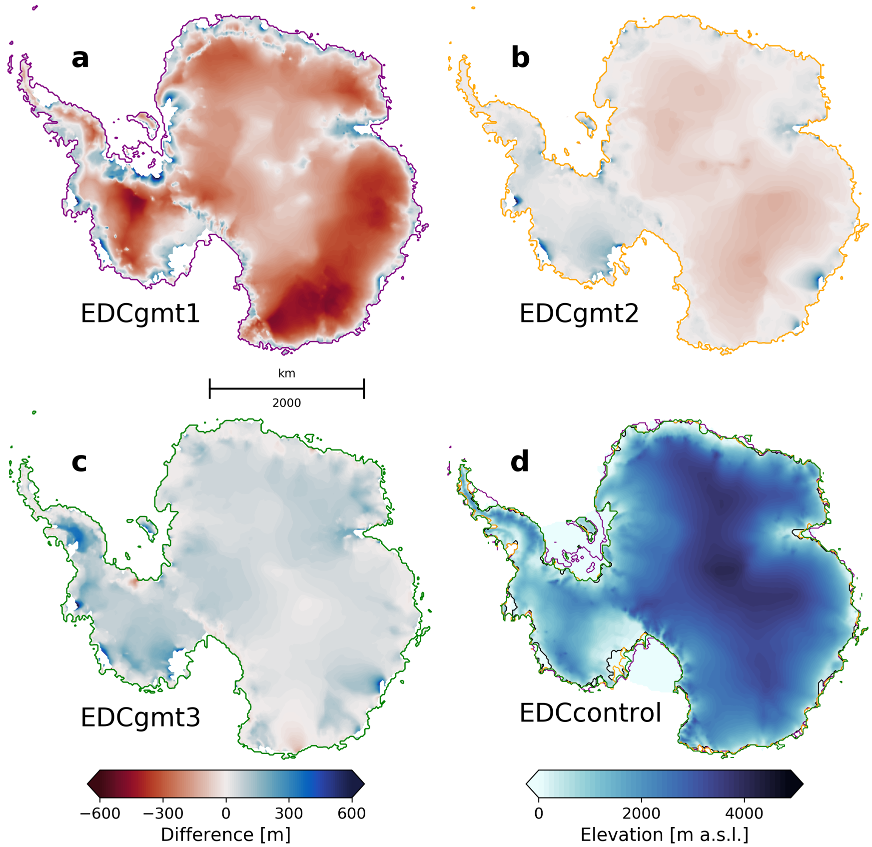

Figure 3. (a-c) Three different starting ice sheet geometries at 420 ka for gmt1–3 using EDC forcing. the EDC CFEN member is used as "control". Color scheme shows differences in surface elevation between each geometry and the control for 420 ka (d). Differences are only shown where the ice is grounded in both geometries, and coloured lines show the respective grounding lines in gmt1-3, also overlain in (d)

LR04 curve starts with colder-than-PI conditions and does not produce a peak warming as strong as the others. It only shows a brief period of warmer-than-PI conditions between 410 and 401 ka (Fig. 2b), resulting in an overall larger AIS (Fig. 5). The ice core CFEN members yield lower ice volumes throughout the entire MIS11c (Fig. 4a), but with important variations. The Vostok-forced experiment, for example, ~~introduces~~ suffers a faster ice loss at the beginning of the simulation period, when it shows a sudden warming. However, it recovers more quickly than the EDC and DF experiments as soon as the peak warming is over and the climate starts to shift back to PI conditions, without a WAIS collapse (we consider the WAIS to have collapsed when the Weddell, Ross, and Amundsen seas become interconnected; Fig. 5).

The members that result in a collapse of the WAIS (forced with the DF and EDC reconstructions) reveal slightly different responses (Fig. 4a). The experiment forced by the EDC reconstruction shows an AIS volume reduction after a sudden warming

Table 3. Summary of performed experiments grouped by ensemble, listing their respective GI forcings, applied sea level reconstruction, and choice of initial geometry. LGMavg denotes that the GI was rescaled using the average LGM value as opposed to the peak value (cf. Sect. 2.3.1 and Table 4). The SGSEN experiments were grouped for better visualisation, but each SGSEN row corresponds to 3 experiments, one starting from each geometry (gmt1–3).

Ensemble	Experiment	GI forcing	Sea level reconstruction	Initial Geometry
CFEN	lr04	LR04	Bintanja and van de Wal (2008)	control
CFEN	edc	EDC	Bintanja and van de Wal (2008)	control
CFEN	df	DF	Bintanja and van de Wal (2008)	control
CFEN	vos	Vostok	Bintanja and van de Wal (2008)	control
SSEN	lr04lgmavg	$\mathrm{LR04}_{\mathrm{LGMavg}}$	Bintanja and van de Wal (2008)	control
SSEN	edclgmavg	$\mathrm{EDC}_{\mathrm{LGMavg}}$	Bintanja and van de Wal (2008)	control
RSEN	lp1bx	EDC (1 kyr low pass, LP)	Bintanja and van de Wal (2008)	control
RSEN	lp3bx	EDC (3 kyr low pass, LP)	Bintanja and van de Wal (2008)	control
RSEN	lp5bx	EDC (5 kyr low pass, LP)	Bintanja and van de Wal (2008)	control
SLSEN	s16l	EDC	Spratt and Lisiecki (2016) long	control
SLSEN	s16s	EDC	Spratt and Lisiecki (2016) short	control
SLSEN	s16u	EDC	Spratt and Lisiecki (2016) upper uncertainty	control
SLSEN	spm	EDC	Imbrie et al. (1989)	control
SLSEN	wae	EDC	Waelbroeck et al. (2002)	control
SGSEN	edcgmt[1-3]	EDC	Bintanja and van de Wal (2008)	gmt1-3
SGSEN	dfgmt[1-3]	DF	Bintanja and van de Wal (2008)	gmt1-3
SGSEN	vosgmt[1-3]	Vostok	Bintanja and van de Wal (2008)	gmt1-3

at around 418 ka, but the WAIS collapse is delayed until 407–406 ka (Fig. 5), following a second short period with an increased warming rate after 412 ka, that leads up to the peak-warming of MIS11c. The DF experiment on the other hand is rather stable until 412 ka, when the climate starts warming towards its peak. Most of the retreat is triggered after the sudden temperature rise at 412 ka, as opposed to when the peak warming occurs.

3.2 Sensitivity to rescaling of the climate forcings

The different δ isotope reference values used for the SSEN experiments are shown in Table 4 (cf. Table 2). Using an LGM-averaged value results in a smaller ice sheet for the LR04 GI, while for the EDC GI it results in a slightly larger AIS than their correspondent CFEN experiments throughout the entire MIS11c (Fig. 4b). The LR04-LGM-averaged run, however, still does not produce AIS retreat as significant as the other experiments, with 3.4% less volume $(1 \cdot 10^6\ \mathrm{km}^3)$ at 402 ka when compared to its original rescaling. The warmer conditions resulting from the GI rescaling are still not enough to compensate for the

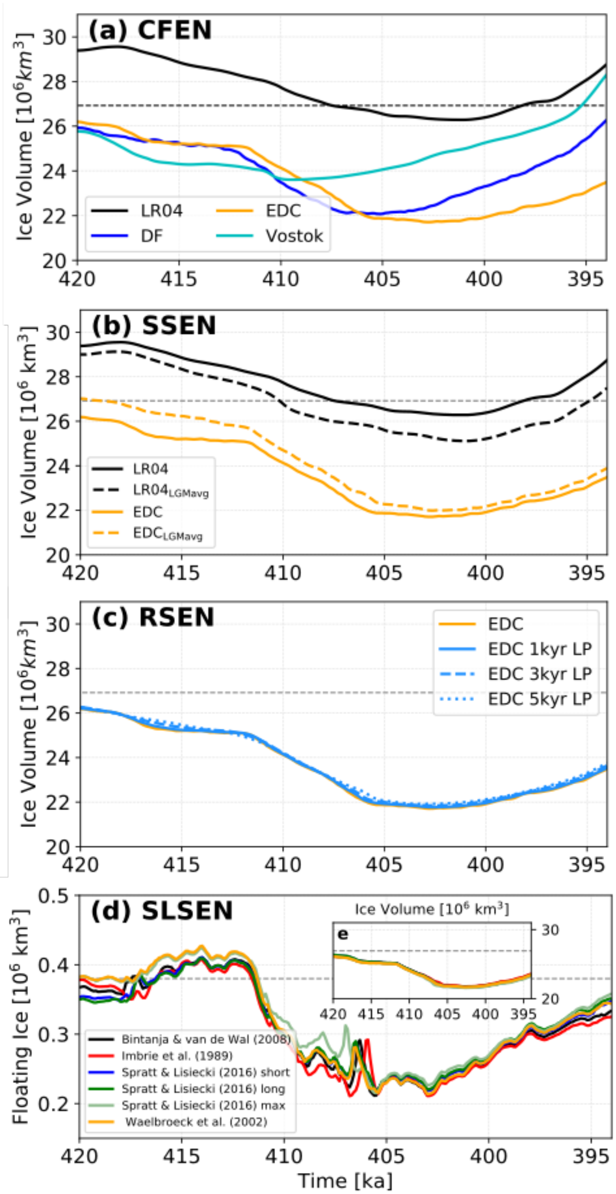

Figure 4. Sensitivity of AIS response (in total ice volume, $10^6 \mathrm{km}^3$) between 420 ka and 394 ka to (a) CFEN GI reconstructions; (b) SSEN rescaled GI reconstructions; (c) RSEN low-pass filtered GI reconstructions~~; (~~. Panels d and e show floating and total ice volumes (in $10^6 \mathrm{km}^3$), respectively, for the SLEN ~~sea level~~ sea-level forcing reconstructions forced by EDC GI (cf. Table 3). Dashed line shows PD ice volume (Fretwell et al., 2013)

initial growth caused by significantly colder-than-PI conditions at 420 ka, and during the preceding relaxation stage. Although differences in ice-sheet volumes exist between the different scaling strategies in the EDC-forced experiments, the resulting ice

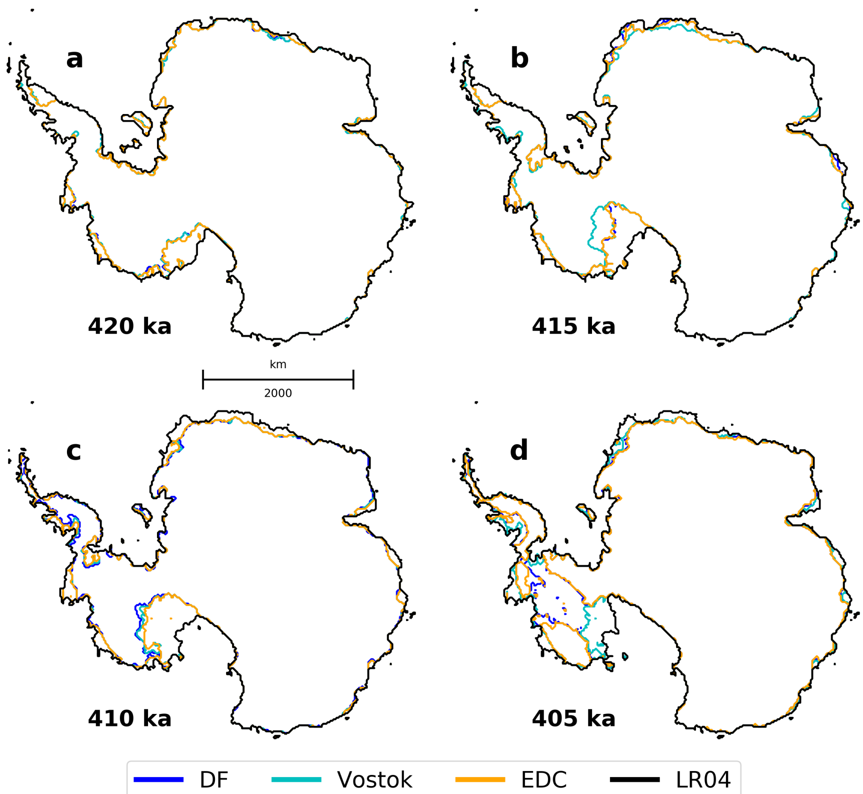

Figure 5. Grounding lines at 420, 415, 410, and 405 ka for the CFEN simulations.

Table 4. Different isotope values adopted for the GI rescaling procedure. *Hol* is the reference value produced by the average over the last 10 kyr (which replaces PI in Eq. 3 for the respective experiments), while *LGMavg* is the reference value obtained from the average between 26 and 19.5 ka (see Sect. 2.3.1).

Record	δX_{PI} [‰]	δX_{Hol} [‰]	δX_{LGM} [‰]	$\delta X_{\mathrm{LGMavg}}$ [‰]
EDC	-397.4	-394.6	-449.3	-442.3
LR04	3.23	3.33	4.99	4.85

sheet histories are quite similar. Despite ice-sheet volume at 402 ka being smaller in the run where the LGM reference is taken as the peak value, the differently scaled ice sheet is only 2.3% larger in volume than the CFEN ice-sheet ($0.5 \cdot 10^6$ km^3).

3.3 Sensitivity to millennial variability and sea level reconstructions

The trajectories of each ensemble member in RSEN agree with one another (Fig. 4c), showing increased delays in the ice sheet retreat in response to the filtering intensity. Also, although it is possible to see slight differences in ice sheet volumes between ensemble members (the volume is larger the more filtered the forcing is), it is negligible compared to the overall changes in volume experienced by the entire ensemble.

3.4 ~~Sensitivity to sea level reconstructions~~

Although the range of global mean sea level reconstructions is wide (nearly reaching 60 m between 405 ka and 400 ka; Fig. 2c), the AIS response in terms of volume is remarkably similar for different sea level curves (Fig. 4e). The differences in sea level have their largest impacts on the volume of floating ice (Fig. 4d). It directly reflects their effect on the flotation of ice, and consequently on the grounding line position. The SLSEN member with the highest sea level rise (i.e., the upper uncertainty boundary of Spratt and Lisiecki, 2016) deviates the most from the other members, especially in the portion of grounded ice being brought to flotation (Fig. 4d). However, the differences are not significant enough to yield substantially distinct ice volume changes (Fig. 4e).

3.4 Sensitivity to the choice of initial ice sheet geometry

Looking at how the four initial geometries (gmt1-3 and the control) evolve under the three different climate forcings from the ice-core derived GI reconstructions (Fig. 6), it becomes clear that all members under the same climate forcing have a tendency to follow the same path despite differing initial ice sheet configurations. The spread in minimum ice-sheet volumes (and consequently implications for WAIS collapse) due to assumptions of starting geometry becomes rather small, between 1 and 3 m s.l.e. at 405 ka among the three different forcings in SGSEN. The different ice sheet configurations also show a similar pacing of retreat after 412 ka, indicating that their corresponding volume by that time did not affect its rate of retreat due to climate warming. In our SGSEN simulations, it appears that the main source of variability between ice sheets with different geometries comes from specific EAIS drainage basins, such as those of Totten, Dibble, and Cook glaciers (Fig. 7 showcases the EDC ensemble; cf. Fig. 1 for geographical locations). The latter two remain thicker in the alternative geometry experiments than in the correspondent CFEN experiment, whereas the former is thinner in gmt3 (Fig. 7c). Some variability can also be observed in the WAIS domain. Parts of Pine Island Glacier appear to resist ice sheet collapse in the thicker-ice-geometry experiments (gmt3) when compared to the ~~CFEN correspondent~~ CFEN-equivalent run (Figs. 7c,d). Given the observed spread, the three ensemble members constrain the range of potential sea level contributions from Antarctica during the MIS11c highstand at 405 ka to 3.2–8.2 m (minimum from Vostok, maximum from EDC). This range can be essentially linked to whether the WAIS has collapsed or not during this period.

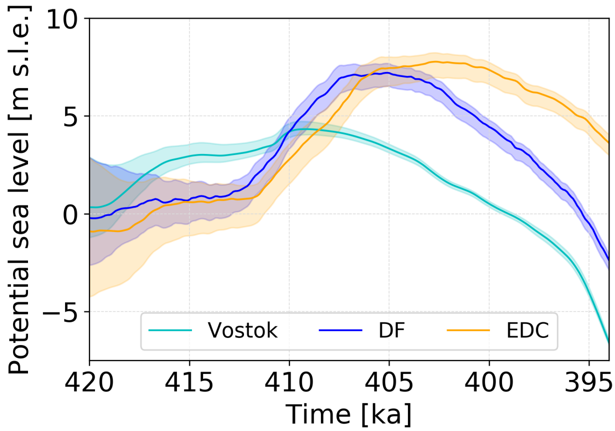

Figure 6. Sensitivity of the AIS response to CFEN GI reconstructions (Vostok, DF, EDC) between 420 and 394 ka with uncertainty bands from four distinct initial ice sheet starting geometries (gmt1–3 and respective CFEN member), expressed in contribution to global mean sea level [m s.l.e.]. Solid lines show the mean of each common-forcing ensemble member, while the color filling shows the spread given by the different starting geometries.

4 Discussion

Our simulations show that during the ~~height~~ peak of MIS11c, the WAIS probably collapsed. We base this statement on results from experiments forced by different proxy records with significant differences in their structure during the MIS11c peak warming. One consisted of a short single peak (Vostok), while others showed a prolonged period of (relatively) warmer conditions (LR04, DF, and EDC). Despite having a warming peak of a similar GI magnitude at 410 ka, the Vostok-forced CFEN member is the only ice core-forced ensemble member that shows no collapse of the WAIS. Although the remaining climate reconstructions all show a longer peak, differences still exist among them. For example, EDC and DF, which are the most similar to each other, start shifting to their warmest conditions at about the same time around 414 ka, but peak at different times. DF peaks at 410 ka, which is 3 kyr earlier than EDC. Regardless of this difference, the simulated WAIS collapse occurs at 407 ka using the DF and at 406 ka using the EDC core forcing, which is closer than their timing of peak warming. Experiments forced by both records also yielded similar ice volumes (Fig. 4a) and extents (Fig. 5). It should be mentioned that the combination of GI and climate-model forcing results in a warmer signal in the surface temperatures at the DF, EDC, and Vostok core sites than obtained directly from their δD records (Supplementary Fig. S14). This is most likely due to the cold bias in CCSM3, which persisted despite the lapse-rate correction applied. Nevertheless, Vostok's GI-reconstructed temperature peak ~~, however,~~ matches the peak observed in DF for its δD-derived curve, and is also close to the warmest temperature reconstructed with the EDC isotopes. Finally, LR04 stands out when compared to the ice cores, and will be discussed in more detail separately.

Although sensitivity experiments show WAIS-collapse results using DF and EDC to be robust, the timing of the events discussed above should be taken with caution for two main reasons. First, we are forcing the entire AIS model with a climate signal from the EAIS, while previous studies have shown that the WAIS could have responded over 2 kyr earlier to changes in

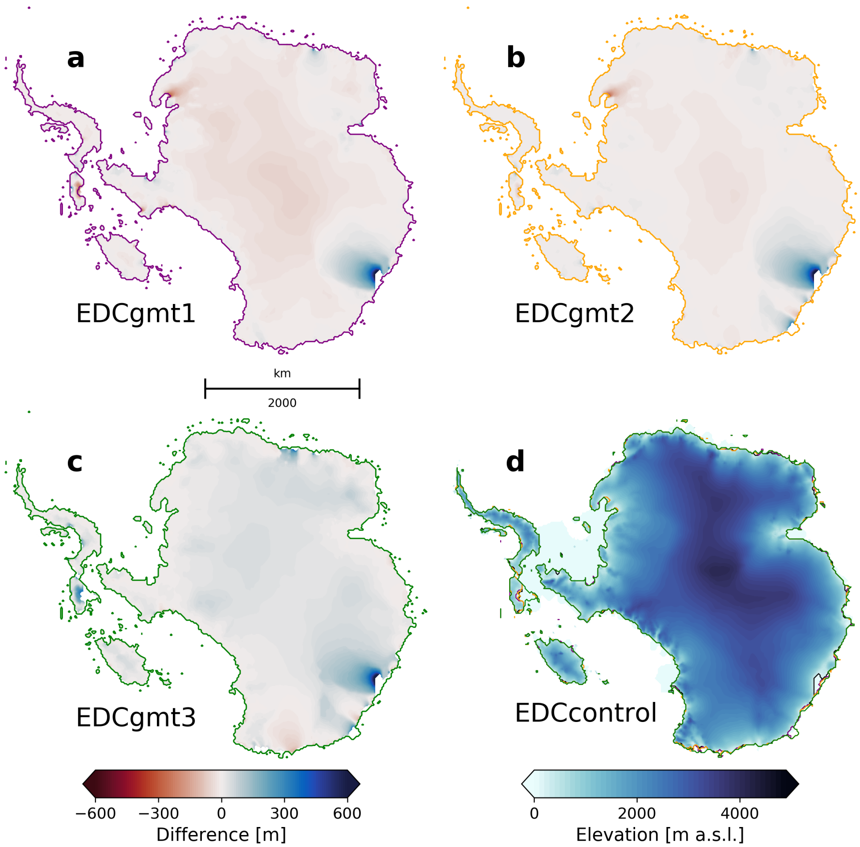

Figure 7. (a-c) Ice sheet geometries at 405 ka for the EDC CFEN member using three different starting geometries at 420 ka (Fig. 3). Color scheme shows differences in surface elevation between each geometry and the control for 405 ka (d). Differences are only shown where the ice is grounded in both geometries, and coloured lines show the respective grounding lines in gmt1-3, also overlain in (d)

climate (WAIS Divide Project Members, 2013). Second, all discrepancies in the timing of the events discussed so far recorded by the ice-core records, especially the peak warming and ice sheet collapse, are within the uncertainty in their respective age models (Parrenin et al., 2007; Bazin et al., 2013). Consequently, these two factors prevent us from establishing an exact timing of these events, which means that the lags in AIS response are the most important to be considered.

In all our CFEN simulations, ice sheet retreat is associated with stronger basal melting close to ~~ice shelf~~ grounding lines, especially at Siple Coast, and in the Ross and Filchner-Ronne ice shelves (Fig. 8). Surface ablation seems to be significant only over the fringes of the EAIS, notably at Dronning Maud Land (DML) and the Amery ice shelf, where surface temperatures reach positive values during summer (Fig. 9a). Nevertheless, they show limited retreat compared to the former two in the WAIS regions. The strong WAIS retreat seen in the EDC and DF-forced runs starting from 412 ka is triggered by an increase

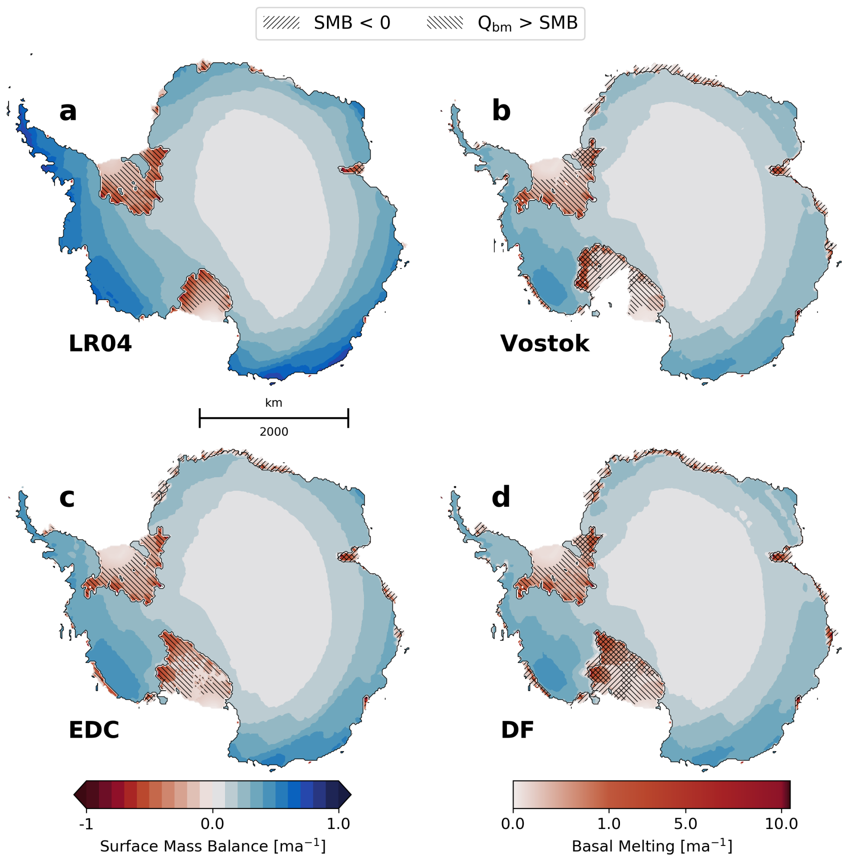

Figure 8. Surface Mass Balance (SMB, ma^{-1}) for the grounded ice and basal melting (Q_{bm}, ma^{-1}) for the ice shelves for the CFEN simulations at 415 ka. Hatched areas show where basal melting dominates over surface mass balance and where surface mass balance is negative (i.e., where surface ablation occurs).~~Everywhere where $Q_{\mathrm{bm}} > \mathrm{SMB}$, ice shelves are thinning.~~

in ocean temperatures at intermediate depths (hereafter defined as the average between 400 and 1000 m depth) under the Ross and Filchner-Ronne ice shelves (Fig. 9b). Although this increase is progressive, it triggers a faster loss of volume by the WAIS compared to the EAIS after 412 ka (Fig. 9c), in contrast with a similar evolution between the ice sheets before then. This observed tipping point at 412 ka also explains why the different ice-sheet configurations all follow the same trend from that moment onwards (Fig. 6), ~~as~~ and why the evolution of WAIS and EAIS sea level contributions diverge. As ocean forcing becomes the main driver of ice-sheet retreat, it has a much larger impact on marine-based portions of the ice sheet. Around most of the EAIS (except for the Amery Ice Shelf), ~~neither ocean temperatures nor ice-shelf melt rates are~~ ice shelves are small and provide little buttressing. Hence, because most of the EAIS is grounded above sea level, its sub-shelf melting is not high enough to force grounding line retreat as strongly as in the ~~aforementioned regions, and ice loss~~ WAIS. As a consequence, ice melt is dominated by surface ablation at the ice-sheet fringes (cf. hatched patterns in Fig. 8).

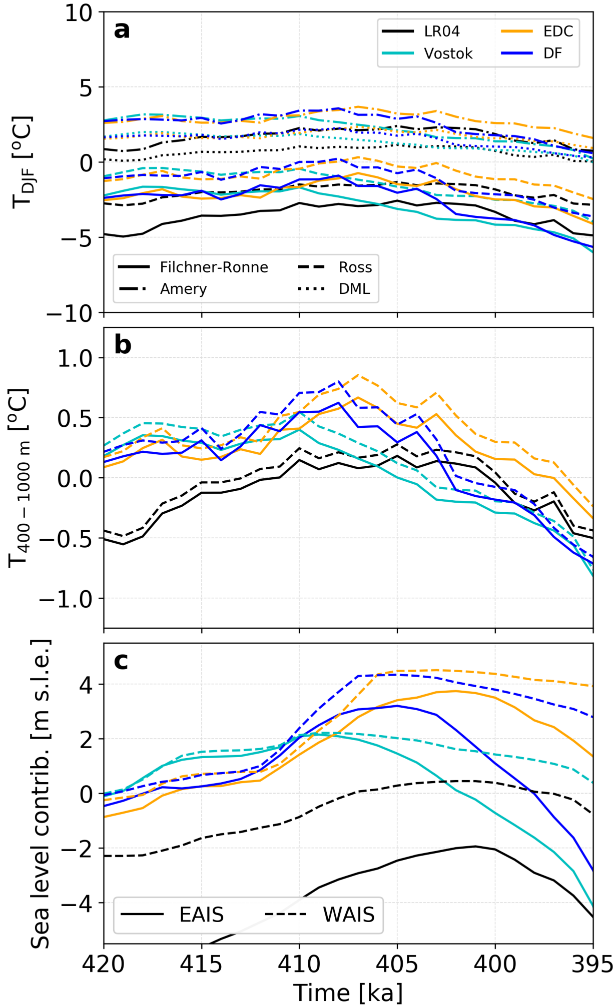

Figure 9. Evolution throughout MIS11 for each CFEN member for (a) Summer surface air temperature [°C] averaged over the main Antarctic ice shelves; (b) ocean temperatures averaged between 400 and 1000 m [°C] for the Filchner-Ronne and Ross ice shelves; (c) sea level contribution by EAIS and WAIS. Colours denote the respective CFEN member, while line styles in panels (a,b) denote each ice shelf, and each ice sheet in panel (c). DML refers to all smaller ice shelves along the Dronning Maud Land margin.

The average intermediate-depth ocean temperatures under the Filcher-Ronne and Ross ice shelves peak between 0.4 and 0.85 °C for the three ice core-forced CFEN members (Fig. 9b). This happens at 410 ka for Vostok, 408 ka for DF, and 407 ka for EDC. Strong WAIS retreat, however, starts before the peak in forcing, supporting the presence of a tipping point at 412 ka. To further test whether this tipping point is the trigger of WAIS collapse, we have performed four additional experiments: *(i)* forced by EDC GI, but keeping the GI constant after 416 ka (i.e., before the threshold found in ocean temperatures), *(ii)* forced by EDC GI, but keeping the GI constant after 410 ka (i.e., just after the sudden increase in ocean temperatures, but before the maximum is reached; cf. Fig. 9b), *(iii)* forced by Vostok GI, where climate forcing is kept constant at its peak condition at 410 ka, and *(iv)* forced by Vostok GI where, after the 410 ka peak, GI is brought back to its 411 ka value (i.e., between the peak and the observed tipping point) and kept constant. Figures 10a,b show that keeping the EDC-derived climate constant at 416 ka conditions prevents the WAIS from collapsing, while keeping it constant at 410 ka conditions delays its collapse by almost 5 kyr compared to the core CFEN run. The Vostok-based simulations (Figs. 10e-h) show that there is indeed a threshold, which is of approximately 0.45 °C for the Filchner-Ronne ice shelf, and 0.54 °C for the Ross ice shelf. However, our results also imply that this threshold must be sustained for at least 4 kyr to cause a collapse (compare red and blue dashed lines in Figs. 10f-h). A short peak at this threshold and subsequent cooling prevents the WAIS from collapsing, compared to keeping it constant at the same peak value (Fig. 10e,f). Comparing these values to PI temperatures averaged over the same extent of the water column, the magnitude of warming necessary to cross this threshold is 0.4 °C. In other words, a warming of this magnitude can be understood as the condition necessary for WAIS collapse (Figs. 10c,d,g,h). Additional experiments where we test for a weakened ocean forcing further confirm this threshold, as a complete collapse of the WAIS is prevented when the temperatures at intermediate depths fail to reach a 0.4 °C warming relative to PI under the Filchner-Ronne and Ross ice shelves (Sect. 4 of the supplementary material). Considering that the temperature peak reconstructed by the Vostok GI is the closest to the δD-derived temperature peaks in DF and EDC (Fig. S14), a more prolonged warming as seen in the DF and EDC ice core seems to be ~~the~~ a crucial condition for the ~~WAIS drawdown, similar to what was suggested by Robinson et al. (2017) for the GIS, while the peak 's intensity could have accelerated or delayed the timing of collapse~~modelled WAIS drawdown during MIS11c. For example, if the GI-derived temperature for DF was not overestimated, and had its peak value close to its isotope-derived value, the response would likely resemble the experiment where Vostok-peak conditions were kept constant from 410 ka onwards.

~~This threshold~~ The inferred critical warming of intermediate-depth ocean temperatures of 0.4 °C for MIS11c is close to the equilibrium model results in Garbe et al. (2020), but lower than ~~the~~ results from Turney et al. (2020) for the AIS retreat during the LIG. While the former study shows a strong WAIS retreat is already possible for an ocean warming of 0.7 °C, the latter ~~identify~~ identifies a tipping point at 2 °C warming in ocean temperatures. ~~It should be noted that a minimum duration of the warming period as a key factor for the WAIS collapse is specific to~~ In other interglacials, such as the LIG, WAIS collapse was triggered by ocean warming with a higher intensity and of shorter duration than during MIS11c ~~. A more intense albeit shorter peak warming could also trigger WAIS collapse~~(Dutton et al., 2015; Turney et al., 2020), since a ~~strong~~ stronger rate of warming can drive ice retreat at a much faster pace~~(Dutton et al., 2015; Turney et al., 2020). Numerical modelling studies in which the WAIS did not collapse during MIS11 were acknowledged to be less sensitive to the ability of ocean temperatures to drive basal melting (e.g., Pollard and DeConto, 2009; Tigchelaar et al., 2019)~~. Thus, WAIS collapse during MIS11c was likely attained because ocean temperatures exceeded a modest threshold for long enough (over 4 kyr).

Despite differences in the model sensitivity to ocean temperature, our results support those of Tigchelaar et al. (2019) and Albrecht et al. (2020) regarding the minor role that variations in sea level play in driving ice-sheet retreat compared to other external forcings. Although the coarse treatment of the grounding lines could have had an influence on the seeming insensitivity of our experiments to ~~sea level~~ sea-level uncertainties, other models of similar resolution which apply different sub-grid

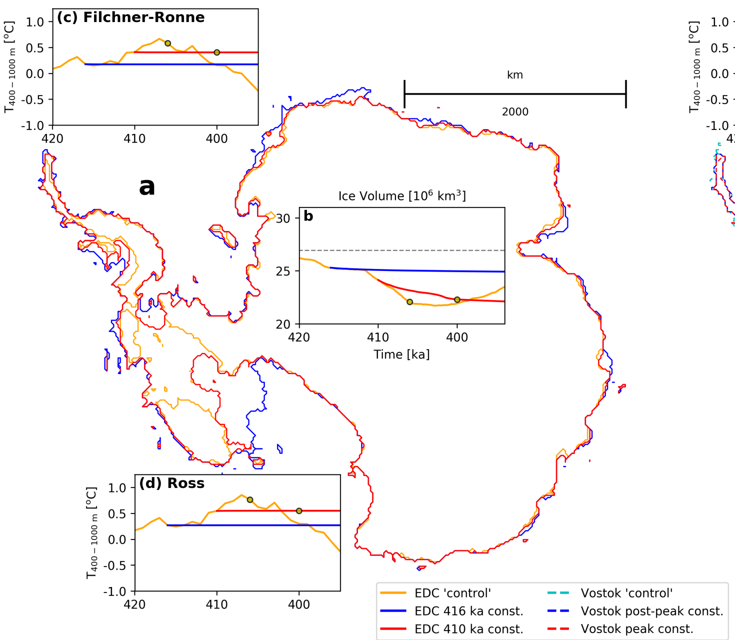

Figure 10. Thresholds for WAIS collapse. (a,e) grounding lines at 405 ka for three EDC-based (solid lines) and three Vostok-based (dashed lines) experiments, respectively (see below for explanation); (b,e) ice volume ($10^6 \mathrm{km}^3$), (c,d; g,h) intermediate-depth (400–1000 m) ocean temperatures [°C] for the Filchner-Ronne and Ross ice shelves, respectively. Time series cover the period between 420 and 395 ka for both EDC (solid lines) and Vostok-based (dashed lines) experiments. Orange line shows the EDC control run, while cyan line shows the Vostok control run. Blue lines show EDC and Vostok simulations where climate was kept constant and the WAIS did not collapse, while the red lines show EDC and Vostok simulations where climate was kept constant and the WAIS collapsed. Yellow circles show the moment when the WAIS breaks down and an open-water connection between the Ross, Weddell and Amundsen seas is established.

parameterisations to the grounding lines yield similar results (Tigchelaar et al., 2019; Sutter et al., 2019; Albrecht et al., 2020) . Hence, while this caveat must be taken into consideration, it does not appear to have influenced our results dramatically.

Moreover, AIS ~~extent,~~ minimum extent and the timing of WAIS collapse ~~, and its contribution to sea level~~ are robust regardless of model resolution (Fig. S15). A set of simulations performed with several resolutions (from 20 to 10 km) showed virtually the same changes in ice-sheet extent, and modest variations in ice volume, which amount to a spread of 1.2 m s.l.e. in sea level contribution at 405 ka. Alternative sliding laws or sub-shelf melting parameterisations, for example using a linear dependence of sub-shelf melt to ocean thermal forcing, or applying a more physically realistic approach (e.g., Reese et al., 2018) were not ested, and could influence our results. For example, numerical modelling studies in which the WAIS did not collapse during MIS11 were acknowledged to be less sensitive to the ability of ocean temperatures to drive basal melting (Pollard and DeConto, 2009; Tigchelaar et al., 2019). Finally, we note that, despite very different approaches in reconstructing transient signals, neither Pollard and DeConto (2009) nor we were able to simulate a collapse of the WAIS using the LR04 stack as climate forcing.

~~We find that the relatively low temporal resolution of~~ The LR04 ~~is not the reason why it did not produce a strong WAIS ice retreat. All RSEN experiments using low-pass-filter-forcing on the EDC GI reconstruction show a similar trajectory compared to the unfiltered forcing. The fact that MIS11c marine records in~~ reconstruction is composed of a stack of 57 globally-distributed ocean sediment cores (Lisiecki and Raymo, 2005), with a strong deficit over the Southern Ocean. In the Nordic Seas, paleoceanographic records indicate that the ocean was colder than present during MIS11 (Bauch et al., 2000; Kandiano et al., 2 . Colder ocean temperatures in the Northern Hemisphere explain why LR04 ~~show~~ shows oxygen isotopic values similar to the Holocene ~~(Lisiecki and Raymo, 2005) despite geological evidence showing~~ during MIS11c (Lisiecki and Raymo, 2005) despite the geological evidence that there was a contribution to higher-than-Holocene sea levels from both Greenland and Antarctica (Scherer et al., 1998; Raymo and Mitrovica, 2012)~~implies that, if true, the ocean must have been colder. Indeed, paleoceanographic records from the Nordic Seas, for example, indicate that they were colder than present during MIS11 (Bauch et al., 2000; Kandiano et al., 2016; Doherty and Thibodeau, 2018). Southern Ocean records remain equivocal about a warming during MIS11 relative to the Holocene (e.g., Droxler et al., 2003)~~. Hence, the inclusion of many Northern Hemisphere records in the LR04 stack ~~could explain~~ explains why it fails to capture the Antarctic warming during MIS11c seen in the ice cores, and the differences in timing compared to them. This also helps explain why the different criteria adopted for changing its scaling procedure had little effect on the results (Fig. 4b). A possible way of circumventing this problem could be to adopt a similar scaling approach to Sutter et al. (2019), who combined the LR04 stack and EDC ice-core temperature records, which, in their study, also led to WAIS collapse during MIS11c.

In East Antarctica, our simulations do not capture the ice sheet retreat into the Wilkes Subglacial Basin recently proposed by Wilson et al. (2018) and Blackburn et al. (2020) for MIS11. Blackburn et al. (2020) suggest this retreat to have been caused by ocean warming, with little to no atmospheric influence. However, further paleoceanographic data are needed to fully understand this retreat (Noble et al., 2020), which so far has not been captured by other model experiments (cf. Wilson et al., 2018, Fig. 2b). As for West Antarctica, far-field sea level reconstructions suggest that a WAIS collapse was the most probable scenario (Raymo and Mitrovica, 2012; Chen et al., 2014) when comparing their results with estimates for the contribution from the GIS. While Robinson et al. (2017) found that Greenland contributed between 3.9 and 7.0 m to sea level rise (having 6.1 m s.l.e. as the most likely value), the AIS contribution cannot be constrained by simply subtracting the GIS's contribution from the global sea level highstand. The suggested asynchronicity between the GIS and AIS minimum extents (Steig and Alley, 2002) and the uncertainties in the age models of the different analysed ice cores (Petit et al., 1999; Parrenin et al., 2007; Bazin et al., 2013) prevent a simple relationship between both ice-sheet records to be established. Based on the ice-core experiments, our interval for the potential sea level contribution of the AIS is 3.2–8.2 m. This wide range is mainly related to whether the WAIS collapses or not. Considering the cases where the WAIS collapsed (i.e., EDC and DF core experiments) as the most probable scenario, our interval for the potential sea level contribution of the AIS is 6.7–8.2 m. In this case, the EAIS contribution is the largest source of uncertainty, being most sensitive to the choice of starting ice geometry. This effect is strongest over Wilkes Land, where the spread in position of the grounding line is wider, and ice thickness is more variable than for other basins

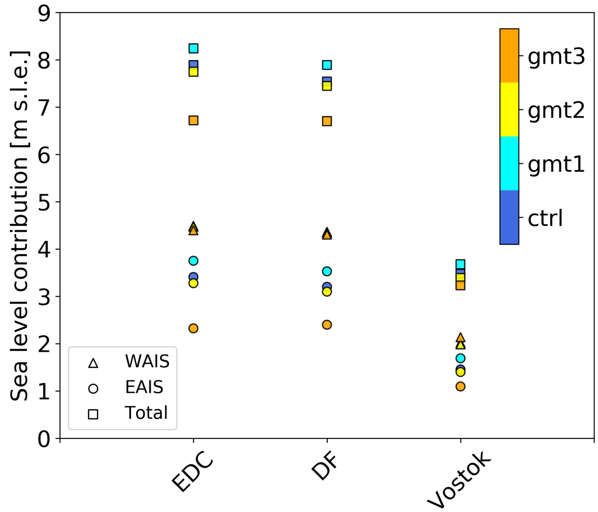

Figure 11. Sea level contribution (in m s.l.e.) of each SGSEN member during the global sea level highstand at 405 ka.

(Fig. 7). While nearby drainage basins, such as those of Totten and Dibble glaciers, become more stable given the larger ice sheet configurations of the alternative geometries (Figs. 3b,c), Cook glacier, emanating from Wilkes Subglacial basin, appears to thin regardless of the choice of initial geometry (Figs. ~~3~~7a-c). Overall, the EAIS contributes 1.1 to 3.7 m s.l.e. at 405 ka (Fig. 11). Conversely, the WAIS was rather insensitive to the choice of starting geometry (yielding 4.3–4.5 m s.l.e. at 405 ka in the case of a collapse, and 2.0–2.1 otherwise) due to the stronger role played by the sub-shelf ocean forcing after 412 ka. There are, however, two stabilising feedbacks which are not incorporated in our model: *(i)* a local sea-level drop caused by a reduced gravitational attraction of a shrinking ice sheet (e.g., Mitrovica et al., 2009), and *(ii)* the observed faster rebound of the crust due to a lower mantle viscosity in some WAIS locations (Barletta et al., 2018). The first effect is probably small based on our model's insensitivity to sea-level changes over these time scales, but we have been unable to robustly test the effect of a faster rebound on AIS response during MIS11c. However, we note that our ELRA model is set up with a relatively short response time of 1 kyr, for which the resulting bedrock uplift is still not able to trigger a stabilizing effect large enough to prevent WAIS collapse.

5 Conclusions

Several studies have been carried out in order to reconstruct past ice changes over the Antarctic continent, but to our knowledge no special focus has been given to Antarctica's response to the peak warming during MIS11c and the driving mechanisms behind it. To fill this gap we evaluated the deglaciation of Antarctica using a numerical ice-sheet model forced by a combination of climate model time-slice-forcing and various transient records through a Glacial Index (GI). The records were obtained from ice cores of the EAIS interior and a stacked record of deep-sea sediment cores taken from far-field regions. We evaluated the sensitivity of our results to *(i)* the scaling of the GI, *(ii)* millennial variability and temporal record resolution, *(iii)* different sea level reconstructions, and *(iv)* initial ice sheet configurations. While sea level, higher-frequency variability, and the GI scaling of the records seemed to play a small role, different responses were seen for both East and West Antarctic Ice Sheets regarding the different applied transient signals, and for the initial ice sheet configurations. Among the applied ice-core reconstructions, the warming captured by the Vostok ice core during MIS11c was not strong enough to cause a collapse of the WAIS, which was attributed to the short duration of its peak. Our results indicate that our modelled WAIS collapse was caused by the duration rather than the intensity of warming, and that it was insensitive to the choice of the starting geometry. The latter proved to be a larger source of uncertainty for the EAIS. Regarding the initial questions posed in the beginning of this study, we now provide short answers to them:

1. **How did the AIS respond to the peak warming of MIS11c? What are the uncertainties in the AIS minimum configuration, its timing and potential sea level contribution?**
   Using transient signals from EAIS ice cores, we found a range in sea level contribution of 3.2 to 8.2 m s.l.e., which mainly reflects whether the WAIS has collapsed or not in our experiments. For the former scenario –which is supported by far-field sea level reconstructions– we find that a WAIS collapse during MIS11c is attained after a prolonged warming period of the ocean of ca. 4 kyr. The resulting AIS contribution in this case is 6.7–8.2 m s.l.e. at 405–402 ka. Uncertainties in these values are primarily due to the choice of climate forcing and ice sheet starting configuration (at 420 ka). While the contribution to sea level rise by the WAIS was consistent among those experiments that yielded its collapse (4.3–4.5 m s.l.e.), the EAIS contribution remained more uncertain because of its sensitivity to the initial geometry of the ice sheet (2.4–3.7 m s.l.e.).

2. **What was the main driver of the changes in the AIS volume? Was it warming duration, peak temperature, changes in precipitation, or changes in the oceanic forcing?**
   We identify a tipping point at ca. 412 ka, beyond which strong WAIS retreat occured in response to the ocean warming. Past this point, retreat leading to WAIS collapse was mostly sensitive to warming duration more than intensity, provided ocean temperatures at intermediate depths become 0.4 °C warmer than PI under the Filchner-Ronne and Ross ice shelves. This threshold should be sustained for at least 4 kyr so that strong WAIS ice retreat is triggered.

Code and data availability. The numerical code for the ice-sheet model SICOPOLIS can be obtained in http://sicopolis.net/. All settings files used for the model runs are available in https://github.com/martimmas/MIS11c_exps. The full model outputs are available upon request to the corresponding author.

Author contributions. MMB, IR and JB designed the study. Experiments were carried out and analyzed by MMB and JB. MMB wrote the manuscript with contributions from all co-authors

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

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