Date: 26 April 2017
Subject: Resubmission of manuscript

Dear Editor,

Please find attached our manuscript entitled "**The influence of ice sheets on temperature during the past 38 million years**" by L.B. Stap, R.S.W. van de Wal, B. de Boer, R. Bintanja and L.J. Lourens for your consideration for publication in Climate of the Past. It is a significant revision of our earlier submitted manuscript "The influence of ice sheets on the climate during the past 38 million years" (Ref. No. cp-2016-109), which was handled by Dr. Guo.

We believe we have addressed all the concerns raised by the reviewers. In our opinion, this has led to a more comprehensive and more detailed manuscript. It is also better structured, which has increased its readability. This revised manuscript complements a point-by-point answer to the comments of the referees, which was uploaded earlier.

Firstly, the title is revised as you requested, although we feel the new title does not entirely do justice to the content, which consists of results of a coupled ice sheet-climate model. As such, it is more than only about temperature. Secondly, we have expanded the introduction section with more discussion of recent work with coupled ice sheet and climate models. This provides more context to our study. Furthermore, the method section now includes the formulas for calculating benthic $\delta^{18}O$ and CO_2, so that the reader does not have to go back to older publications to understand our current study. Most importantly, we have included more comparison to earlier work. A section has been added, comparing our CO_2 results to proxy data records. Also, the discussion of our temperature records and the interrelationships between CO2, sea level and temperature have been improved, most notably by including a figure comparing our results to the data analyses of Foster and Rohling (2013). Lastly, the conclusion section and abstract now reflect the main merits and findings of this study better.

We hope the new manuscript is to the satisfaction of the reviewers and yourself.

Yours Faithfully,

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REPLY TO THE COMMENTS BY THE REVIEWERS

Color coding:
Black – comments by reviewers
Blue – reply by authors

General

Our manuscript has been reviewed by two referees, who have very different opinions. Reviewer #1 is very positive and suggests some minor modifications. Reviewer #2, however, is uncomfortable with our study's target, methodology and analyses, and sees no point in its publication. He/she feels our current model setup lacks important geological processes that might influence the relation between CO_2 and climate over time. This is certainly a valid argument, that is why we do not wish to claim we present a definite CO_2 simulation over the past 38 Myr. We realize this constraint on the implications of our study has not come across well enough before, therefore we will make it clearer in the revised manuscript. Nevertheless, we feel our current model setup represents a complementary approach to more sophisticated models that have been deployed over shorter time scales (mainly snap-shot simulations), which generally also do not take into account these geological processes. Furthermore, it is a step forward from using a standalone ice sheet model. Our study has a two-fold aim: being a first step in the direction of transient coupled simulations of the climate and cryosphere on long time scales, and quantifying the influence of ice sheets on climate variability. Therewith it is an intermediate step that provides valuable information for the research community. In our opinion, this justifies publication of the current results. Nonetheless, we will take into account as much of the suggestions by both reviewers as possible, which in our opinion will significantly improve the quality of the manuscript. Below, we will answer to the comments of the reviewers.

A point where both reviewers agree upon is that the current structure of the paper could be improved. We have therefore decided to follow their suggestions, and structure the revised manuscript in the following manner:

Introduction
The introduction will be expanded with a discussion of studies using more sophisticated climate and ice sheet models on shorter time scales. This will provide a better perspective of the research field and our contribution to it. We will more clearly state the purpose of our current study.

Model
This section will include a more thorough description of the coupled climate-ice sheet model we use, as well as the inverse routine to simulate CO_2. The equations we use to calculate $\delta^{18}O$ and CO_2 will be provided. The setup of the different model runs we perform will be moved to the Results and Discussion sections to improve readability of the paper.

Results and Discussion I: Long-term transient simulations
This section will demonstrate the different CO_2 concentrations we obtain over the past 800 kyr when we integrate the model over the past 5 Myr or 38 Myr. We will introduce the hysteresis runs to explore this difference further. Thereafter, we will more clearly describe how and why we re-tune the model. We will present the main results of our new 38 Myr reference run: $\delta^{18}O$, CO_2, ice-volume-equivalent sea level (total, as well split into contributions from NH and Antarctica) and global mean temperature. The CO_2 will be compared to the proxy data by Beerling and Royer (2011), and based on this, we will address the caveats and shortcomings of our model, and thereby the scope and target of our study.

Results and Discussion II: Ice sheet-climate interaction
This section will remain largely the same as Section 4, but including more discussion.

Summary and Conclusions
The discussion of the results will be moved to the Results and Discussion sections. This section will only contain a brief summary of our experiments, and the conclusions we derive from them.

Reviewer #1

General comments:

The authors present results of model simulations of the past 38 million years using a simple zonally averaged energy balance model coupled to a 1D ice sheet model. The presented results contribute to our understanding of climate - ice sheet interactions on very long timescales and therefore the paper represents a valuable contribution to this research field. The use of a relatively simple model is justified by the very long transient simulations which would be too computationally expensive to perform with more complex models. However, in order for the paper to be suitable for publication in Climate of the Past, some minor issues listed below should be addressed.

We thank the reviewer for considering our work, and we are pleased that he/she agrees with our general approach. We will explain below how we will take the comments into consideration. In our opinion this will improve the quality of the paper, hopefully to the satisfaction of the reviewer.

The model is described only very briefly in the Methodology section. I'm aware that the model is described in more detail in previous publications, but it would be useful to the reader who is not familiar with the model if some more details would be given (e.g. the resolution of the ice sheet model is not even mentioned in the text). How is the model initialized? Additionally, the way the CO2 concentration is derived in the model and applied as forcing is crucial for the simulations performed and should be described in the paper. I would suggest to at least include the equations for $\delta18O$ and CO2.

The reviewer suggests to expand the Methodology section, a point to which reviewer #2 agrees. Therefore, in our new Model section we will include a more thorough explanation of our modelling strategy, including the equations used to calculate $\delta^{18}O$ and CO_2.

The description of the experiments used to show the hysteresis behavior of the model is spread over several sections of the paper, which is very confusing to the reader. First it is mentioned in the Methodology section that using different δ18O stacks gives very different results but no reason for that is given until section 3. Then at the end of Section 2 (Page 4, lines 7-15) the hysteresis experiments are described, but it is difficult to understand why these experiments are needed before knowing what the problem is (which is only outlined in Section 3). I would suggest collecting all of this in one section describing the difference between 5Myr and 38Myr simulations, the experiment setup for diagnosing the reason for the differences, the hysteresis behavior and the retuning procedure.

We will follow the suggestion of the reviewer, and move the description of the model runs to the new Results and Discussion sections. In the new section Results and Discussion I, we will describe the difference between 5 Myr and 38 Myr simulations, the experiment setup for diagnosing the reason for the differences, the hysteresis behavior and the retuning procedure, as the reviewer suggests. This section will end with a comparison of our simulated CO_2 to the proxy data of Beerling and Royer (2011) (see also our reply on a further comment by reviewer #1).

I'm not aware of any other modeling study showing a hysteresis behavior that is caused by the atmosphere model or ocean model when excluding overturning, so it would be interesting to know what is causing this. Because of the relatively short time scale of atmospheric processes, it seems difficult to imagine that the climate model keeps memory of the initial conditions over multimillenial time scales. Could the authors elaborate on this? Are the different hysteresis branches really stable equilibria of the model? Also, does this hysteresis behavior depend on the forcing rate (50 ppm/50 kyr)? What are the initial conditions for these experiments?

The hysteresis runs will be more thoroughly described in the revised manuscript. The questions raised by the reviewer will be addressed.

The model-derived atmospheric CO2 could be compared with available proxy data (e.g. Beerling and Royer, 2011).

The new Results and Discussion I section will contain a comparison of our model results to the proxy data of Beerling and Royer (2011). Based on this comparison, we will address the caveats and shortcomings of our model. This will clarify the purport of our current model results, as well as indicate a route to go forward from here.

It would be interesting to see also the sea level evolution (maybe also the ice volume evolution separately for NH and Antarctica) and possibly global temperature evolution, also to make it easier for the reader to interpret Figures 4 and 5.

The new Results and Discussion II section will start with a figure showing the main results of our new reference run (after re-tuning): $\delta^{18}O$, CO_2, ice-volume-equivalent sea level (total, as well split into contributions from NH and Antarctica) and global mean temperature.

Figure 3 is very hard to read, especially Figure 3b. Maybe Figure 3b could be split in 3 different plots?

To improve readability, Figure 3b will be split as suggested by the reviewer, such that the revised Figure 3 will be composed of 4 subplots (A-D).

The following sentence in the abstract (Lines 8-9) is not clear, at least not until one has read the rest of the paper: 'Firstly, we investigate the relation between global temperature and $_{CO2}$, which changes once the model run has experienced high CO2 concentrations.'

In the revised abstract, we will be clearer on the implication of the analyses of the hysteresis runs.

Reviewer #2

This paper deals with an important issue: the role of ice sheets on the climate evolution since the late Eocene (38 Ma). To achieve this goal, they use simplified climate energy balanced models and also a simplified ice sheet model. Using these tools enables them to simulate very long time spans.

General comment:
Whereas this is an important issue for which there are many unsolved problems as the evolution of Antartica ice-sheets during Oligocene and Miocene and its implication on climate, I feel very uncomfortable with the target, the methodology used and the analysis provided in this paper.

We thank the reviewer for a careful consideration of our work. Unfortunately, he/she is very critical towards our modelling approach. Although the reviewer certainly has some valid points, we still believe our results represent a step forward in our understanding of the influence of ice sheets on long-term climate variability. Below, we will describe why, and which revisions we make to hopefully ease the objections of the reviewer as much as possible.

These authors had first used this tool to investigate the relationship between cryosphere and climate for 1 million year (Lennert, B Stap, 2014) and extend afterwards to 8 million years (Lennert B Stap, 2016, A). In this new paper, they enlarge the period to 38 million years. But for many reasons I will explain below, this extension is not convincing with respect to many features: a first obvious one is the role of tectonics on CO2 that the authors perfectly know because they also recently published a paper concerning this issue (Lennert, B Stap, 2016 B). The tectonics, through many different processes, will affect atmospheric pCO2 (see Godderis for a review). For instance opening and closing sea ways may change climate and CO2, orogenesis (E.G Tibetan Plateau Uplift) and plate motion that will impact silicate weathering. Therefore, the extension to 38 Ma they provide in this paper is not really reliable. They reconstruct the pCO2 as a prognostic variable from their model which is indeed important but as they online derive it from radiative perturbation there are missing many fundamental processes. Consequently, their reconstructions of pCO2

over the 38 million years is not in good agreement with data as the authors recognize but instead of accounting for causes of such a disagreement on geological time scale they tuned the model with different parametrization of the clouds physics. This caveat makes the paper not appropriate for publication. Nevertheless, there are potential interesting sensitivity experiments that are possible with such a tool.

The reviewer mentions a number of geological processes that are not taken into account in our model setup, but could influence CO_2. However, our study does not concern which processes govern the CO_2 concentration in the atmosphere – to address this issue, one would need a carbon cycle model – but what influence CO_2 has on the climate, and how ice sheet variability changes this influence. Nevertheless, changing topography could lead to a different relation between CO_2 and the coupled climate-ice sheet system, e.g. via changing ocean overturning strength and surface elevation. Indeed, in a previous publication (Stap et al., 2016B) we have explored the effect of the latter process. Our model is unable to simulate some of the aspects shown by proxy data, as we will show in a comparison of our results to the data of Beerling and Royer (2011) in the revised manuscript. We therefore do not wish to claim that we provide the final answer to the evolution of CO_2 over the past 38 Myr. However, our modelling results clearly represent a step forward from previous studies using a stand-alone ice-sheet model (De Boer et al., 2010), and provide valuable insights into the influence of ice sheet variability on climate. In the revised manuscript, we will be very clear on the purpose of our current study, as well as the caveats that can be addressed in further research.

Another drawback
is the fact that they avoid in the introduction to give a context of the state of the art of climate cryosphere interaction using sophisticated GCM as De Conto and Pollard (for instance De Conto and Pollard in Nature 2003, Geoscientific Model Development 2012 and Earth and Planetary Science Letters 2015) developed since many years. One of the major results of De Conto et al. study is to be able to reproduce the evolution of ice sheets since Eocene. They pointed out the importance of cryospheric processes (Pollard and De Conto, EPSL, 2015) that are not discussed at all in this manuscript.

The second mayor concern of the reviewer regards the lack of discussion of previous results, in particular the work of Pollard and DeConto in many much-cited publications. This point is easily addressed by expanding the introduction of our study to include this discussion. Here, as well as in the new section Results and Discussion II, we will discuss how our results relate to their work, which generally concerns shorter time scales (mostly snap-shot simulations) but using a more sophisticated model setup. We refrain from quantitative comparisons on short time scales, however, since our intention is not to capture any event in great detail, but to provide the larger picture of the long-term influence of ice sheets on the climate. Our results are also not completely independent of the work of Pollard and DeConto, since the inception CO_2 level of the Antarctic ice sheet is highly dependent on the parametrisation of the mass balance in our model, and is matched to the one found by Pollard and DeConto (~780 ppm). This will be explained in the revised manuscript.

Due to these two major problems I don't believe that at this stage such a paper may be published. Nevertheless I will give more details and comments because there is a large room for improvement if the authors want to resubmit their manuscript.

Detailed comments:

1. Abstract First, the relationships between CO2 temperature and
ice sheets are consistent within the framework of the modeling study but completely
inconsistent with available data concerning CO2 evolution since 38 million years. This
is clearly shown in the paper but not in the abstract itself.

We would argue that our results are actually not as bad as the reviewer states here, as we
will show in a more rigorous comparison to proxy data in the revised manuscript. However,
we will mention the shortcomings of our model also in the abstract.

Second, the authors insist
on very obvious results as for instance it is colder when you get an ice sheet but the
most interesting part of the work is to provide many sensitivity experiments. Indeed, this
approach, conversely to GCM, as for example De Conto and Pollard (Palaeogeography,
Palaeoclimatology, Palaeoecology 2003), allows them to quantify specifically the role
of albedo on one side and elevation on the other side. This is not clearly stated in the
abstract.

We agree with the reviewer that a main merit of our setup is that it lets us attribute the
effect of ice sheets on the climate to two important feedbacks: the ice-albedo feedback and
the surface-height-temperature feedback. As this is a main result of this work, it will be
mentioned in the revised abstract.

Introduction: This section is a bit short. Some references are missing which may
be important. For instance, concerning the Pliocene and Greenland onset, recent
publications of Contoux et al (EPSL, 2015) and for MMCO a publication of Hamon
(Geology, 2012) constrains on Antarctica ice sheet at MMCO and also Hamon (Climate
of the Past, 2013) which depict the role of East Tethys seaway on Antarctica ice sheet
40 million years ago. More importantly, the authors should discuss the interest of their
approach compared to the development of GCM studies as those published by De
Conto and Pollard (EPSL, 2015) which pinpointed the importance to parametrize the
ice sheet with sophisticated models to capture correctly the ice sheet dynamics and
therefore to reproduce the ice sheet evolution through Eocene.

We thank the reviewer very much for pointing our attention at these studies. We will
expand the Introduction section with a discussion of these works, which will give the reader
a better perspective of the field and our contribution to it. See also our reply to an earlier
comment by the reviewer.

Methodology section: First, the authors claimed they used Penthic ïA¸d'O18 isotope
records to infer the temperature of the Ocean, but it is absolutely unclear to me how
they really disentangle the part corresponding to ice-sheet melting and the part due to
bottom sea surface temperature. This first step has to be clarified, since it is used then
to derive through radiative calculation the atmospheric CO2. I strongly believe than
in a first step, the authors should have used the different proxy reconstruction used
for CO2 as published in the literature, which provides different CO2 evolution (Boron

isotopes, Alkenon, leaf stomates,. . .) to validate their simplified coupled model. Such a strategy based on CO2 reconstruction from data allows to test the response of their tool in terms of cryosphere and climate evolution. Instead, they choose to compute the CO2 from the reconstructed SST, derived from their radiative model.

The reviewer is unclear as to how our inverse CO_2 calculation from benthic $\delta^{18}O$ data works, a concern shared by reviewer #1. Therefore, in our new Model section we will include a more thorough explanation of our modelling strategy, including the equations used to calculate $\delta^{18}O$ and CO_2. We would like to stress that CO_2 is not obtained from SST data, but from benthic $\delta^{18}O$ data which is disentangled into contributions from deep-sea temperature and land ice volume in our model.

In a previous publication (Stap et al., 2014), we have validated our coupled model, using CO_2 data from the EPICA Dome C record as input. A reason to refrain from using proxy CO_2 data from earlier times as input is that it is currently to scarce and intermittent. Moreover, there is large inter- as well as intra-proxy disagreement. Instead, we opt to use an inverse routine, and compare the results to available proxy data. In the revised manuscript, an explanation of this choice and a comparison to the data of Beerling and Royer (2011) shall be included.

As you know,
there are many reasons and causes that may affect atmospheric CO2, that cannot be accounted for in this very simple modeling tool, especially when dealing with geological time span (38 million years). For instance, seaway changes - and there are many seaway changes in that period (see Zhang et al. Climate of the Past. 2011) - or the impact of mountain uplift and associated weathering (see Raymo et al. Nature 1992 and C France-Lanord, Nature, 1997). Therefore, the only processes they captured here, attributing Ocean temperature changes to CO2, is obviously missing a lot of important processes that will change the atmospheric CO2 during that period.

We agree with the reviewer that we are missing certain processes that affect the relation between CO_2 and the coupled climate/cryosphere on these long time scales. We therefore do not want to present our CO_2 record as the definite simulation of CO_2. Rather, as we express in the Discussion/Conclusion section, we pave the way for long time scale simulations, identifying interesting phenomena and potential obstacles. One of these is precisely that the processes mentioned by the reviewer are important. As this contradicts certain earlier studies (e.g. Foster and Rohling, 2013), we shall make this clearer in our new section Results and Discussion I.

Moreover, they
use a fixed contribution for the methane in this radiative calculation, (factor 1.3, which is supposed to include the methane radiative perturbation). This value is certainly valid for the last million years, for which data are available, but which is also a very cold period compared to the last 37 million years period they are investigating.

We would like to argue that we do not see a better alternative here. The factor 1.3 is indeed derived over the past 800 kyr, the only period over which we have reliable CH4 and N2O data, as is explained in our publication Stap et al. (2014). We will mention the implication of this modelling choice in the new section Results and Discussion I.

Finally, they
consider the lapse rate also constant through time whereas, this has been also shown
as oversimplified (Svetlana Botsyun et al., Climate of the Past. 2016).

Here again, this is the best we can do at this moment. This point will also be included in the discussion.

These important
caveats in the methodology used here, which are absolutely not discussed, imply, as
the authors themselves pinpoint, very large underestimation of their computed CO2
when compared to different proxies: the CO2 computed from the temperature record
of Zacchos or Raymo, but also those much more accurate and directly obtained from
Antarctica ice core (EPICA).

We are afraid that we have not been able to convey our findings well enough to the reviewer. We do not underestimate a proxy computed from the temperature record of Raymo and Zacchos. The point is that initially we tuned the model to simulate CO_2 over the past 800 kyr in agreement with the EPICA Dome C record. Using the same exact model, however, we lose this agreement if we start our model integration further back in time (38 Myr instead of 5 Myr ago). The disagreement can therefore not be caused by omitted processes, since we in fact use the same model. Hence, we explore the cause of it, by analysing separate hysteresis runs. We will explain this more clearly in the revised manuscript.

The authors claimed that such a mismatch may be overcome
by changing the optical properties of the clouds. This is not really serious for
me, because it is a kind of tuning without really understanding what is the physics of
the problem, but more importantly, they do this tuning for all the time period, whereas
there is a strong bias using only EPICA data, which is associated to a very cold period
compared to the whole period they are studying. Indeed, most of these 38 million years
were much warmer than LGM or present day climate. Therefore, there is no reason for
a constant tuning.

We would like to remind the reviewer that all models used to simulate climate and/or ice sheets are to lesser or stronger degree tuned in some way. We want to be very clear that the cloud optical thickness is such a tuning factor used in our climate model. Indeed, there is no physical preference for its value before or after re-tuning. Precisely therefore we choose this factor to regain agreement with the EPICA Dome C core. Although we agree that physically cloud properties may change in different climates, we think using a variable parameter setting is cumbersome and does not lead to increased understanding of the studied system. However, we will explain the implications of this choice more precisely in the revised manuscript.

This also explains why the underestimation is so large for deep time
(larger for Zacchos than for Raymo). This methodology by itself induces many problems
and leads the authors to explore methodological induced problems, as hysteresis,

rather than to really try to capture the dynamics of the cryosphere, or the evolution of the climate in their result section.

We are not clear on which underestimation the reviewer refers to here. Also, we are confused by the argument that we do not try to capture the dynamics of the cryosphere, or the evolution of the climate. In our opinion, this is thoroughly dealt with in the Section 4, which will be transformed into a new section Results and Discussion II.

Part 3 results: The part concerning hysteresis is not relevant and convincing for me. Hysteresis has been shown to be an important factor to account for instance in glacial/interglacial cycles (see for instance papers from Paillard Nature 2001, Calov, GRL, 2005. Alvarez-solas Nature Geosci, 2010, De Conto and Pollard Nature 2008. . .). Here the analyses of the results which depict a strong correlation with the initial climate is not really explained in terms of physics and for me belongs much more to model caveats and development than to the analyses of results interesting enough to be published.

We realise that ice sheet variability, as well as other feedbacks, may cause hysteresis in models and possibly also in the real world. This hysteresis, which is also to some degree contained in our coupled model, however, is inherently different from the hysteresis we explore in this study. We aim to find out why the simulated CO_2 over the past 5 Myr is so different if we start our simulations further back in time. This is in fact not caused by a feedback, but by the core of the model we use, since it shows up even when all feedback processes are shut off. The hysteresis shown by our model may therefore very well be a model caveat. We, however, think this is all the more reason to be honest about its implications and not sweep it under the carpet, particularly because results of this model have already been published before. Moreover, more sophisticated models have, as far as we know, not been tested for this behaviour. We therefore advise to check for this, before using such models on long-time scales (when computer power permits to do so). This is also in line with our studies objective of identifying interesting phenomena and potential obstacles for transient long-term simulations.

Part 4 discussion: In the discussion section, the summary of the paper is too exhaustive, we really expect a discussion of the results and comparison with the results of other models. For example, these last years, many studies provided by De Conto and Pollard depicted very new results on climate and ice sheets evolution, since the last 40 million years. In this part, we should expect a serious comparison between these results and those provided by the others including the fact that the tools used are different. Therefore, it would be interesting to discuss the result of these two complementary approaches (GCM versus simplified models). Such a discussion will allow the authors to clarify the potential and weaknesses of their method. For instance, simplified tools as used here do not capture important processes that are necessary to simulate ice sheet evolution in GCM. The authors show comment on this point in the discussion section and also highlight on the fact that their tools allow to quantify different forcing factors through the sensitivity experiments.

We realise now that the discussion of our results in relation to previous studies from our own and other research groups is a little bit obscured by the combination with the summary

of our current study in this section. We will therefore move the discussion to the two new Results and Discussion sections in the revised manuscript. In the introduction, and in these Results and Discussion sections, we will also include a discussion of our results with respect to the work of Pollard and DeConto. In the new Summary and Conclusion section we will present our main conclusions.

Conclusion: I strongly believe that there is much room for improvement in this paper. The sections that are devoted to sensitivity experiments (albedo vs topography of the ice sheets) could be a valuable contribution, but at this stage and, accounting for the weaknesses in methodology and construction design of the paper, I think the paper should be rejected. Nevertheless, there are some parts of paper, that, if completely rebuilt could be used and might be a valuable contribution, but in a framework of a completely new and rethought paper.

It is unfortunate that the reviewer thinks the results of our current model setup are not suitable for publication. In our opinion, we provide a complementary approach to snap-shot and short timescale results of more sophisticated models, and make a step forward from just using stand-alone ice sheet models (e.g. De Boer et al., 2010). Even though this does not lead to any definite answer – there is always a way forward in science – it represents a marked improvement with important results and implications, one of them indeed being the attribution of the effect of ice sheets on the climate to albedo and topographic changes. We think it is therefore a meaningful contribution to the research field. Nevertheless, as we pointed out, we will make an effort to make the merit of our approach, and the purpose of our study clearer, mainly by improving the structure of the paper and by adding more discussion.

References:

Beerling, D. J., & Royer, D. L. (2011). Convergent Cenozoic CO_2 history. *Nature Geoscience*, *4*(7), 418-420.

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The influence of ice sheets on temperature during the past 38 million years

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Abstract. Since the inception of the Antarctic ice sheet at the Eocene-Oligocene Transition ($\sim$34 Myr ago), land ice has played a crucial role in Earth's climate. Through feedbacks in the climate system, land ice variability modifies atmospheric temperature changes induced by orbital, topographical and greenhouse gas variations. Quantification of these feedbacks on long time scales has hitherto scarcely been undertaken. In this study, we use a zonally averaged energy balance climate model bi-directionally coupled to a one-dimensional ice sheet model, capturing the ice-albedo and surface-height-temperature feedbacks. Potentially important transient changes in topographic boundary conditions by tectonics and erosion are not taken into account, but briefly discussed. The relative simplicity of the coupled model allows us to perform integrations over the past 38 Myr in a fully transient fashion, using a benthic oxygen isotope record as forcing to inversely simulate CO_2. Firstly, we find that the results of the simulations over the past 5 Myr are dependent on whether the model run is started at 5 or 38 Myr ago. This is because the relation between CO_2 and temperature is subject to hysteresis. When the climate cools from very high CO_2 levels, as in the longer transient 38 Myr run, temperatures in the lower CO_2 range of the past 5 Myr are higher than when the climate is initialized at low temperatures. Consequently, the modeled CO_2 concentrations depend on the initial state. Taking the realistic warm initialization into account we come to a best estimate of CO_2, temperature, ice volume-equivalent sea level and benthic $\delta^{18}O$ over the past 38 Myr. Secondly, we study the influence of ice sheets on the evolution of global temperature and polar amplification by comparing runs with ice sheet-climate interaction switched on and off. By passing only albedo or surface height changes to the climate model, we can distinguish the separate effects of the ice-albedo and surface-height-temperature feedbacks. We find that ice volume variability has a strong enhancing effect on atmospheric temperature changes, particularly in the regions where the ice sheets are located. As a result, polar amplification in the Northern Hemisphere decreases towards

warmer climates as there is little land ice left to melt. Conversely, decay of the Antarctic ice sheet increases polar amplification in the Southern Hemisphere in the high-CO_2 regime. Our results also show that in cooler climates than the pre-industrial, the ice-albedo feedback predominates the surface-height-temperature feedback, while in warmer climates they are more equal in strength.

1 Introduction

The most abundant information source with the highest resolution on Cenozoic global climate change are stacked benthic oxygen isotope ($\delta^{18}O$) records (Lisiecki and Raymo, 2005; Zachos et al., 2008; Cramer et al., 2009), which have been well-studied and statistically analysed (e.g. Mudelsee et al., 2014). The benthic $\delta^{18}O$ signal is known to be comprised of two factors (e.g. Chappell and Shackleton, 1986): 1) the deep-sea temperature, and 2) the volume of land ice on Earth. An additional independent record of either one is therefore required to separate the signal into its individual constituents. Deep-sea temperature records can be reconstructed based on the Mg/Ca proxy (Lear et al., 2000; Sosdian and Rosenthal, 2009; Elderfield et al., 2012), but a global average deep-sea temperature is hard to obtain. Sea level records are also available, but are subject to the same problem of inferring a global mean (Miller et al., 2005; Kominz et al., 2008; Rohling et al., 2014). Studies using sea level records face the additional challenge of converting local sea level to ice volume, which is not straightforward mainly because of dynamic topography (Mitrovica and Milne, 2003; Kendall et al., 2005). Alternatively, calculation of benthic $\delta^{18}O$ can be incorporated in coupled ice sheet-climate models, by using parameterisations of the contribution of deep-sea temperature (Duplessy et al., 2002), and the isotopic content of ice sheets (Cuffey, 2000). Hitherto, studies using this approach have mostly focused on relatively short time intervals surrounding important climatic events, such as the Eocene-Oligocene Transition (33.9 Myr ago; Tigchelaar et al. (2011); Ladant et al. (2014); Wilson et al. (2013)), the Mid-Miocene Climatic Optimum (13.9 Myr ago; Langebroek et al. (2010); Gasson et al. (2016)), and the Pliocene-Pleistocene Transition (2.6 Myr ago, Willeit et al. (2015)), using models of varying complexity. These studies have provided valuable information because they have simulated these key events in great detail. However, they do not to describe climate change on multi-million year time scales, and prove consistency by transiently simulating multiple events using the same set-up. This is mainly due to insufficient computer power. An effort was made to simulate the past 3 Myr using a model of reduced complexity (Berger et al., 1999), but they did not include the Southern Hemisphere and the Antarctic ice sheet in their model. Here, we will use a coupling between an energy balance global climate model and a one-dimensional ice sheet model of all major ice sheets, to perform transient simulations over the past 38 Myr. By focusing on the long term evolution of climate, we provide a complementary approach to snap-shot and short time slice experiments of more complex models.

Our model approach builds on the inverse routine to derive atmospheric temperature from benthic $\delta^{18}O$, that was introduced by Oerlemans (2004). This methodology was consequently developed further to force stand-alone three-dimensional ice sheet models over the past 1 Myr (Bintanja et al., 2005; De Boer et al., 2013), the past 3 Myr (Bintanja and Van de Wal, 2008), and the past 5 Myr (De Boer et al., 2014). With this inverse routine, the past 40 Myr were simulated (De Boer et al., 2010) and further analysed (De Boer et al., 2012), using a one-dimensional ice sheet model that calculates all land ice on Earth. In Stap

et al. (2014), this ice sheet model was coupled to a zonally averaged energy balance climate model (Bintanja, 1997), and run over the past 800 kyr forced by a compiled ice core CO_2 record (Petit et al., 1999; Siegenthaler et al., 2005; Lüthi et al., 2008). Inclusion of a climate model added CO_2 to global temperature and sea level as an integrated component of the simulated system. In addition, it rendered the possibility of investigating ice sheet-climate interactions, specifically the ice-albedo and

5 surface-height-temperature feedbacks. Furthermore, instead of annual mean and globally uniform temperature perturbations to present-day climate, seasonal meridional temperature distributions were used to force the different ice sheets. In a subsequent study, the inverse routine was transformed to yield CO_2 concentrations using the benthic $\delta^{18}O$ as input, making CO_2 a prognostic variable (Stap et al., 2016a). The resulting values were used to force the coupled model over the past 5 Myr. In Stap et al. (2016b), the model was run over the period 38 to 10 Myr ago, and the influence of Antarctic topographic changes

10 on the simulated CO_2 was investigated. In this study, however, changes in topographic boundary conditions are not included, although their effect is briefly discussed.

Here, we will use this coupled ice sheet-climate model forced by benthic $\delta^{18}O$ to transiently simulate the entire past 38 Myr. We recognise that our simulation of CO_2 may be improved in subsequent studies that include geological processes that are still missing in our model setup. For instance tectonics leading to mountain uplift (Kutzbach et al., 1993) and closure of sea

15 ways (Kennett, 1977; Toggweiler and Bjornsson, 2000; Hamon et al., 2013), erosion (Wilson et al., 2012; Gasson et al., 2015; Stap et al., 2016b), and vegetation changes (Knorr et al., 2011; Liakka et al., 2014; Hamon et al., 2012) may have affected the climate system during the past 38 Myr. Here, we will purely focus on the influence of ice sheets on the climate, in particular the relation between CO_2 and temperature, during this time. Earlier studies using more complex stand-alone ice sheet models and coupled ice sheet-climate models have for example determined the CO_2 thresholds for glaciation of Antarctica (DeConto

20 and Pollard, 2003; Langebroek et al., 2010; Ladant et al., 2014; Gasson et al., 2014) and the Northern Hemisphere (DeConto et al., 2008; Hamon et al., 2013). Furthermore, they have investigated the hysteresis in the relation between ice volume and CO_2 (Pollard and DeConto, 2005), as well as the behaviour of the Antarctic ice sheet during the Oligocene (Pollard et al., 2013) and Plio-Pleistocene (Pollard and DeConto, 2009). Our model represents reduced complexity in the model hierarchy (see e.g. Pollard (2010) for a discussion), and adds a long-term transient perspective on the evolution of ice sheets and the

25 climate. It reconciles knowledge on benthic $\delta^{18}O$, CO_2, sea level and temperature. However, it lacks ice-shelf dynamics and sophisticated grounding line parameterisations that have been shown to be important at least on short time scales (Pollard and DeConto, 2012; Pollard et al., 2015). Therefore, this study should be seen as a first step in the direction of simulating multi-million year time spans with coupled ice sheet-climate models, aiming to combine temperature, sea level and CO_2 in one framework. As such, its goal is two-fold. Firstly, as a precursory study, we attempt to identify interesting phenomena

30 and potential obstacles, and set a reference simulation over the past 38 Myr. When results of more sophisticated models are achieved, they can be compared to ours to see which features appear in the full hierarchy of models and which are specific to more comprehensive models including more physics. Secondly, we perform multiple 38-Myr integrations of our coupled model with ice-sheet climate interactions switched on and off. This allows us to quantify the effect of these interactions on global temperature perturbations and polar amplification, distinguishing between the ice-albedo and the surface-height-temperature

35 feedback.

2 Model

We use the same simplified coupled ice sheet-climate model setup as Stap et al. (2016b) (Fig. 1). The climate component is a zonally averaged energy balance climate model (based on North, 1975; Bintanja, 1997) with 5° latitudinal and 0.5 day temporal resolution. It includes a simple ocean model that has 6 vertical layers and mimics meridional ocean circulation with varying

5   strength based on the density difference between the polar and equatorial waters. Sea ice cover is calculated thermodynamically at 1.25° resolution. The climate model provides the local monthly temperature input (T) to the mass balance module of one-dimensional models of the five major Cenozoic ice sheets (Eurasian, North American, Greenland, West Antarctic and East Antarctic ice sheets) (De Boer et al., 2010). Herein, accumulation follows as a temperature-dependent fraction of precipitation (P):

10   $$P = P_0 e^{0.04T - R/R_c}, \tag{1}$$

where P_0 is present-day precipitation, R is ice-sheet radius, and R_c is an ice-sheet dependent critical radius. An insolation-temperature melt formulation is used to calculate ablation:

$$M = [10T + 0.513(1-\alpha)Q + C_{abl}]/100. \tag{2}$$

Here, α is surface albedo, and Q local radiation obtained from Laskar et al. (2004). Ice-sheet dependent tuning factors C_{abl}

15   determine the thresholds for which ablation starts (listed in Stap et al., 2014). The ice sheet models calculate the surface height change and ice sheet extent using the Shallow Ice Approximation with 15 to 25 km resolution depending on the ice-sheet (De Boer et al., 2010). This information is used to update the land ice fraction and surface height profile in the climate model for the next time step. Exchange of variables takes place every 500 model years, constituting the output-timestep of the coupled model. The isotopic content of the ice sheets is calculated using the parameterisation of Cuffey (2000):

20   $$\delta^{18}O_i = \delta^{18}O_{PD} + \beta_T \Delta T + \beta_Z \Delta Z, \tag{3}$$

where β_T and β_Z are ice-sheet dependent parameters (values listed in De Boer et al., 2010), that determine the influence of annual mean temperature (ΔT) and surface height (ΔZ) perturbatons with respect to present day. Present-day isotopic contents ($\delta^{18}O_{PD}$) match the modeled values of an earlier study by Lhomme et al. (2005). The modeled benthic $\delta^{18}O$ values follow from:

25   $$\delta^{18}O = [\delta^{18}O_b]_{PD} - \frac{\overline{\delta^{18}O_i V_i}}{V_o} + \left[\frac{\overline{\delta^{18}O_i V_i}}{V_o}\right]_{PD} + \gamma \Delta T_o, \tag{4}$$

where $[\delta^{18}O_b]_{PD}$ is the observed present-day value of benthic $\delta^{18}O$, and V_o and V_i are the ocean and land ice volume. The final term on the right hand side quantifies the influence of deep-sea temperature change with respect to present day (ΔT_o). Gain factor γ is set to 0.28 ‰K^{-1}, taken from a paleotemperature equation (Duplessy et al., 2002). The coupled model is forced by insolation data (Laskar et al., 2004) and an inverse routine, which yields CO_2 concentrations from the difference

30   between the modeled benthic $\delta^{18}O$ value and an observed value an output-timestep later (Stap et al., 2016a):

$$CO_2 = \overline{CO_2} * \exp[c * \{\delta^{18}O(t) - \delta^{18}O_{obs}(t + 0.5kyr)\}], \tag{5}$$

where $\overline{CO_2}$ is the mean CO_2 concentration of the preceding 15 kyr, and c a strength-determining parameter (Stap et al., 2016a). For the observed benthic $\delta^{18}O$ values we use stacked records (Lisiecki and Raymo, 2005; Zachos et al., 2008), which effectively serve as model input. The radiative forcing anomaly with respect to present day is multiplied by a factor 1.3. to account for non-CO_2 greenhouse gases. This factor is based on analysis of CH_4 and N_2O records over the past 800 kyr (see Stap et al., 2014), and used to account for the lack of knowledge on non-CO_2 greenhouse gases prior to that period. An increase or decrease in the relative contribution of non-CO_2 would need an opposing change in CO_2. The result of the coupled model consists of mutually consistent records of benthic $\delta^{18}O$, atmospheric CO_2, temperature, and ice-volume equivalent sea level.

3   Results and Discussion I: Long-term transient simulations

3.1   Hysteresis

We perform two model runs, one over the past 38 Myr (this run is called '38 Myr'), and one over the past 5 Myr (this run is called '5 Myr'). The 38-Myr run uses the stacked benthic $\delta^{18}O$ record of Zachos et al. (2008) as forcing. As a spin-up the model was initialized with a 1500 ppm CO_2 concentration for 50 kyr, and thereafter run for 2 Myr between 40 and 38 Myr ago using the inverse routine. This run is an extension of the reference run used in Stap et al. (2016b), to include the past 10 Myr. The record of Lisiecki and Raymo (2005) was used to force the 5-Myr run, after initializing the model for 100 kyr with 430 ppm CO_2. This run served as a reference run before in Stap et al. (2016a). When we compare the final 5 Myr of our 38-Myr simulation to our 5-Myr simulation, we notice that the 38-Myr simulation shows much lower CO_2 concentrations (Fig. 2b, green and blue lines). These contradicting results cannot be explained by the use of different forcing records - Zachos et al. (2008) for 38 Myr as opposed to Lisiecki and Raymo (2005) for 5 Myr - as these show similar values during this time (0.02 ‰ average difference, not shown). To explore the difference, we additionally conduct four pairs of experiments with the model in forward mode. In forward mode, we do not use the inverse routine, but force the model by a-priori designed CO_2 scenarios. We initialize the model using: a 450 ppm CO_2 concentration; no land ice; glacio-isostatically relaxed present-day topography. We force the model by changing the CO_2 input in steps of 50 ppm every 50 kyr. In one set of experiments (named 'up'), the CO_2 is first raised from 450 ppm to 1200 ppm, then lowered to 150 ppm, and increased again to 600 ppm. In the other set (named 'down'), the CO_2 is initially dropped from 450 to 150 ppm, then raised to 1200 ppm, and ultimately decreased again to 300 ppm. Insolation is kept at PD level throughout all these equilibrium experiments.

Starting at 450 ppm CO_2, the 'up' and 'down' runs show the same initial global temperature (Fig. 3a). However, in the 'down' run, where the CO_2 progresses stepwise downward first and then upward, the global temperatures at low (< 450 ppm) CO_2 values are approximately a degree lower than those in the 'up' run, where CO_2 is first raised and then lowered. When the 'down' run is integrated over another CO_2 cycle, it shows the same global temperatures as the 'up' run (not shown). This means that once the coupled model has experienced high CO_2 values during its run, as is the case in the 38-Myr run but not in the 5-Myr run, the climates at lower CO_2 are warmer. This has important consequences for the simulated CO_2 concentrations as they have to decline further to obtain similar temperatures, which is what happens in the transient 38-Myr simulation forced

by the inverse routine. The different branches in Fig. 3 are stable equilibria of the model. As long as the model is indeed in equilibrium at every time step, its behaviour does not depend on the forcing rate: using 50 ppm/100 kyr or 100 ppm/100 kyr leads to the same results. This behaviour is a form of hysteresis as results depend on previous conditions of the model. The question now arises what the cause of this hysteresis is. The global temperature difference between the 'up' and 'down' run is 0.94 K at 150 ppm CO_2. When the ice sheet model is uncoupled, and the climate model is directly forced in the same manner but using PD ice sheets, this reduces to 0.69 K (Fig. 3b). Keeping the ocean overturning strength fixed at PD also leads to a small reduction; the difference becomes 0.73 K (Fig. 3c). The combined effect of uncoupled ice and ocean overturning strength is still not sufficient to eliminate the hysteresis (Fig. 3d). Even when in addition sea ice and snow cover are kept constant, a small hysteresis is present (not shown). This means that the hysteresis is inherent to the core of the climate model: the parameterisation of vertical and horizontal energy transfer in the ocean and atmosphere. The factors mentioned above act to enhance this hysteresis.

3.2 Retuning: new reference simulation

Originally, the 5-Myr run was calibrated to an Antarctic ice core proxy-CO_2 record (Petit et al., 1999; Siegenthaler et al., 2005; Lüthi et al., 2008) over the past 800 kyr (Stap et al., 2016a). It shows negligible bias (-3.9 ppm) to that record. As a consequence of the hysteresis described in Sect. 3.1, the 38-Myr run shows much lower values than this proxy record with a -47.7 ppm bias (Fig. 2a, mind that here we show 1-kyr values instead of 40-kyr averages). The simulated CO_2 over the past 5 Myr in the 38-Myr run is also much lower than the hybrid proxy data-model reconstruction by Van de Wal et al. (2011) (Fig. 2b, black line). Therefore, we deduce that the CO_2 record of the 38-Myr run is not realistic over this period. To regain agreement with the ice core record, which we judge to be essential for a transient paleoclimate simulation, we define a new reference run in this study (new REF). In this new reference run we increase the cloud optical thickness parameter τ_{cl} from 3.11 to 3.41. We opt to alter this parameter because it was already used as a tuning parameter in the original climate model (Bintanja, 1997), and in the ice sheet-climate model coupling (Stap et al., 2014). Both values are physically plausible. Increasing τ_{cl} will lower the temperatures calculated by the climate model, such that for the same benthic $\delta^{18}O$ higher simulated CO_2 levels are obtained, in better agreement with the ice core record. However, this will also raise the threshold CO_2 level for the inception of the East Antarctic ice sheet (EAIS). By increasing the ablation threshold parameter C_{abl} in the insolation-temperature-melt calculation (Eq. (2)) of the EAIS (from -30 to -10), this ice sheet glaciates at lower temperatures and therefore at lower CO_2 concentrations. This parameter was used in Stap et al. (2014) to match the CO_2 inception point for Antarctic glacial inception to the one found by DeConto and Pollard (2003) (~ 780 ppm), and is also poorly constrained. Changing this parameter compensates the unintended CO_2 threshold increase. We force the model in the same way as the earlier 38-Myr run over the past 38 Myr using the stacked benthic record of Zachos et al. (2008) as forcing.

The CO_2 concentrations of the new reference simulation are shown in Fig. 2 by red lines. The simulated CO_2 levels right before the Eocene-Oligocene Transition (EOT; $\sim$33.9 Myr ago) and at the Middle-Miocene Climatic Optimum (MMCO; 17 to 15 Myr ago) are similarly high around 650 to 750 ppm, likewise as in the earlier 38-Myr simulation (Fig. 2c). In the time between these events, CO_2 in the new reference run is modestly higher, up to 100 ppm. This is because the deep-sea

temperatures are lower at the same CO_2, and therefore contribute less to the $\delta^{18}O$ anomaly with respect to present day. Compensating for the lower deep-sea temperatures, higher CO_2 increases the $\delta^{18}O$ anomaly, by increasing both deep-sea temperature and the contribution of ice volume, hence raising sea levels (not shown). After the MMCO, when the EAIS has stabilised to near-PD size, the new reference simulation shows higher CO_2 values than the earlier 38-Myr simulation. Over the past million years, the new reference simulation (Fig. 2a, red line) agrees much better with the 5-Myr run (Fig. 2a, green line) and with the ice core CO_2 record (Fig. 2a, cyan line); the bias with respect to this proxy-record is reduced to 13.6 ppm. Even after re-calibration, the simulated new reference CO_2 remains lower than in the 5-Myr run during the Pliocene and early Pleistocene (Fig. 2b, green line), as a consequence of the hysteresis. Although it is more variable than the reconstruction based on a constant Earth System Sensitivity (ESS) by Van de Wal et al. (2011) (Fig. 2b, black line), the long-term means are now similar. It remains debatable whether the shorter 5-Myr run or the new reference simulation is the most veracious over the past 5 Myr. On the one hand, the long simulation carries a longer memory, which would be closer to the state of the actual climate system. On the other hand, it is uncertain how accurately our climate model simulates very warm climates; the climate model is designed and tested for PD and LGM climates (Bintanja and Oerlemans, 1996). This argument favours the shorter 5-Myr run as the more trustworthy result, implying that CO_2 levels over the last 5 Myrs may have been up to 470 ppm.

3.3 Comparison to proxy CO_2 data

A comprehensive quantitative comparison between our CO_2 simulation and proxy data is hindered by scarcity and intermittency of data records. Nevertheless, in Fig. 4 we show the new reference CO_2 results together with proxy data over three periods where data is relatively abundant: the Middle Pliocene Warm Period (MPWP; 3.5 to 2.5 Myr ago), the Middle Miocene (18 to 13 Myr ago) and the Early Oligocene (35 to 30 Myr ago). The data is based on three important proxies (see also Beerling and Royer, 2011) : alkenones (Pagani et al., 1999, 2011; Badger et al., 2013), boron isotopes (Pearson et al., 2009; Foster et al., 2012; Greenop et al., 2014; Martínez-Botí et al., 2015) and stomata (Van der Burgh et al., 1993; Kürschner, 1996; Kürschner et al., 2008; Retallack, 2009). During the MPWP, the simulated CO_2 is more variable than the alkenone data from Badger et al. (2013) (Fig. 4a). This discrepancy between our simulation and the alkenone proxy (Pagani et al., 1999) is persistent throughout the Miocene (not shown). The variability in our simulation is more in line with the boron isotope proxy (Martínez-Botí et al., 2015). However, our simulation shows lower CO_2 values between 3.3 and 2.9 Myr ago. Comparison to the boron isotope proxy over the Miocene is hampered by lack of data. It would be interesting to know if this proxy also shows larger variability during that epoch. During the Middle Miocene, our simulated CO_2 is considerably higher than all proxy data records (Fig. 4b). Contrarily, during the Early Oligocene, it is a little bit lower (Fig. 4c). As mentioned in Sect. 3.2, the CO_2 inception point for Antarctic glacial inception was tuned to the one found by DeConto and Pollard (2003) ($\sim$ 780 ppm). It is therefore in agreement with the range of Antarctic glaciation values found by Gasson et al. (2014), using combinations of an ice dynamical model coupled to seven climate models (Fig. 4c, yellow shading). Using a different tuning, our Middle Miocene values are closer to the observations, but then the Early Oligocene values are also much lower (see Stap et al., 2016b, for an analysis). In short, we are not able to simulate a difference in CO_2 between right before the Eocene-Oligocene Transition and during the Middle Miocene Climatic Optimum. As our CO_2 simulation is obtained using benthic $\delta^{18}O$, this could indicate a discrepancy

between the $\delta^{18}O$ and CO_2 proxies. More likely, however, it is for the largest part due to processes missing in our model set-up. Over the course of millions of years, tectonics, erosion and vegetation have led to topography and albedo changes that affect the climate system. Contrary to the findings of Foster and Rohling (2013), these geological processes may have had a significant impact on the relation between CO_2, sea level and temperature (Gasson et al., 2016). Indeed, in Stap et al. (2016b) we showed that in our model erosion could lead to a changing relation between CO_2 and ice volume over time, bringing $\delta^{18}O$ and CO_2 in line with other proxy indicators arguing for the importance of erosion. Also, we use a uniform lapse rate correction of 6.5 K for height changes in our model, which is a potentially important simplification (Gasson et al., 2014; Botsyun et al., 2016). In the future, our simulation can be improved by including these processes. For now, with this limitation in mind, we will focus on the long-term interaction between ice sheets and the climate in our model.

4 Results and Discussion II: Ice sheet-climate interaction

Figure 5 shows the main results of the new reference run: benthic $\delta^{18}O$, atmospheric CO_2, ice-volume-equivalent sea level and global temperature. In our model, the relation between temperature and ice volume can roughly be divided into three regimes (see also De Boer et al. (2010) and Van de Wal et al. (2011)): 1) at low CO_2 values, strong ice volume variability due to dynamic Northern Hemispheric ice sheets, 2) at intermediate CO_2 values, weaker variability, 3) at high CO_2 values, strong variability due to a dynamic Antarctic ice sheet. This constitutes a sigmoidal temperature-sea level relation. The data-analysis results of Gasson et al. (2012) show a similar shape, but with higher deep-sea temperature anomalies during the warmer-than-PI climates. This was also the case in an earlier study using the same ice sheet model, but with parameterised deep-sea temperatures (De Boer et al., 2010). Our modeled trend in sea surface temperatures is also lower then suggested by proxy data (Herbert et al., 2016) (not shown). The modeled relation between logarithmic CO_2 and sea level is also sigmoidal. This is in very good agreement with the results from Foster and Rohling (2013), who derived a functional relation between these quantities from a geological data perspective (Fig. 6). However, sea level during the stable middle regime is lower in our case. This is coherent with the modeling results of Gasson (2013) (Suppl. Fig. 1). Possibly, recent advances in ice sheet modeling (Pollard et al., 2015) can explain (part of) this difference between models and data. Furthermore, our highest CO_2 levels are slightly lower than the data shows. The modeled CO_2 threshold for Antarctic glaciation is highly dependent on the mass balance parametrisation (Stap et al., 2016b; Gasson, 2013), the climate model used (Gasson et al., 2014), and the Antarctic topography (Stap et al., 2016b; Gasson et al., 2014). In our case, this threshold is distinctly higher than for that for land ice in the Northern Hemisphere. This is mainly a consequence of the higher latitude of the Antarctic continent, and in line with earlier findings (DeConto et al., 2008).

Next, we will investigate the influence of ice sheet-climate interaction on polar amplification, and on the Earth System Sensitivity (ESS). The ESS is defined as the global temperature response to a radiative forcing caused by changing CO_2, taking into account all climate feedbacks (PALAEOSENS Project Members, 2012). This radiative forcing by CO_2 is proportional to the logarithmic change of CO_2 (Myhre et al., 1998). In Fig. 7 (red dots), we therefore show the relation between global temperature anomalies from pre-industrial (PI) and the logarithm of CO_2 divided by a reference PI value of 280 ppm in our

new reference run. Evidently, this relation is not constant, as in warm climates the global temperature increase for a given CO_2 increase is less strong. The slope of a least squares linear regression shows a value of 10.6 K for $\ln(CO_2/CO_{2,\mathrm{ref}})$ values below 0 ($CO_2 < 280$ ppm: coldest climates), and 3.7 K for values above 0.69 ($CO_2 > 560$ ppm: warmest climates), a reduction of 65 %. These values are equivalent to 7.3 K and 2.6 K per CO_2 doubling respectively. Hence, ESS is not constant, in contrast

5 with the implicit assumption in Van de Wal et al. (2011). In fact, in our model ESS is stronger at lower CO_2. This is similar to the findings of Hansen et al. (2013), who performed CO_2 doubling experiments using the simplified atmosphere-ocean model of Russell et al. (1995). However, their ESS decrease is less strong, as it drops from ∼6 per CO_2 doubling from 155 to 310 ppm CO_2 to ∼5.5 K from 620 to 1240 ppm (see their Fig. 7b). They eventually also find increased sensitivity again at very high CO_2 levels (2480 to 9920 ppm), which is outside the range we simulate during our time span.

10 In our ice uncoupled run, the slope of the relation between CO_2 and global temperature reduces by only 46 % from 5.6 K to 3.0 K going from the coldest to the warmest $\ln(CO_2/CO_{2,\mathrm{ref}})$ regime (Fig. 7, blue dots). In this case, the standard error of a linear regression through all data points is reduced by 58 % with respect to the fully coupled run, from 0.0050 K to 0.0021 K. The fact that the relation between $\ln(CO_2/CO_{2,\mathrm{ref}})$ and global temperature is better approximated by a linear fit when land ice is uncoupled means the $\log(CO_2)$-T relation is more linear. Hence, climate sensitivity is more constant. However, even

15 when ice sheets are kept at PD level, the relation between logarithmic CO_2 and global temperature shows a declining slope (Fig. 7, blue dots). Therefore, decreased sensitivity at higher CO_2 is not only determined by reduced ice volume variability. This finding may be compared to the hybrid data-model results for climate sensitivity of Köhler et al. (2015), as well as to the modeled climate sensitivity of Friedrich et al. (2016). Köhler et al. (2015) investigated the relation between the radiative forcing of proxy-data CO_2 (Petit et al., 1999; Siegenthaler et al., 2005; Lüthi et al., 2008; Hönisch et al., 2009) - which is linearly

20 related to logarithmic CO_2 (Myhre et al., 1998) - and modeled global temperature (De Boer et al., 2014, scaled). Friedrich et al. (2016) forced the intermediate complexity climate model LOVECLIM over the past 800 kyr using the ice core record (Petit et al., 1999; Siegenthaler et al., 2005; Lüthi et al., 2008) and a Northern Hemispheric ice sheet reconstruction (Ganopolski and Calov, 2011). The resulting climate sensitivity of these studies is opposite to ours, as they show increased climate sensitivity at higher CO_2 concentrations. These studies, however, consider a smaller range of CO_2. Furthermore, they calculate climate

25 sensitivity in a different way. They do take into account ice volume variations, but compensate for their effect by adding their radiative forcing to the forcing induced by CO_2 variations (see PALAEOSENS Project Members, 2012). Implicitly assumed in their approach is that these radiative forcings have the same effect on temperature, which may not generally be the case (Yoshimori et al., 2011). The difference between the results of our model and these studies could point to contrasting strengths of the fast feedbacks in the climate system, which is material for future investigation. Our findings are in agreement with Ritz

30 et al. (2011), who used a two-dimensional energy balance climate model that showed an increase of climate sensitivity from 3.0 K per CO_2 doubling at PI conditions to 4.3 K at LGM conditions.

The coldest global temperature anomaly in our results is amplified by 79 % (factor 1.79) if land ice changes are incorporated, by 50 % if only albedo is coupled (Fig. 7, black dots), and by 4 % if only surface height is coupled (Fig. 7, orange dots). The warmest anomaly is only increased by 21 % (factor 1.21) when ice is coupled, by 9 % when only albedo is coupled, and by 3 %

35 when only surface height is coupled. This means the surface-height-temperature feedback becomes relatively more important

in warmer climates.

The influence of ice sheets on the climate is strongest in the region where they are situated, leading to increased polar amplification. This is demonstrated by the relations between global temperature and Northern Hemispheric (40 to 80° N, Fig. 8a), and Antarctic temperature (60 to 90° S, Fig. 8b). In the Northern Hemisphere, the minimum local temperature with respect to PI is -2.0 K in the uncoupled case, and -9.5 K in the run with fully coupled land ice. When only the albedo or surface height changes are coupled, the Northern Hemispheric temperature anomaly reaches -6.4 K and -2.8 K low points respectively. Conversely, the amount of land ice lost in warmer climates is relatively small, as only the Greenland ice sheet (~ 7 m.s.l.e.) is left to melt. Consequently, the Northern Hemispheric temperature is then not affected much by not including land ice changes. The remaining polar amplification in the Northern Hemisphere is hence mostly caused by other factors, such as sea ice and snow cover variability. In the Southern Hemisphere, the lowest temperature is similar for the coupled and uncoupled simulations, although it is achieved at a higher global temperature in the uncoupled case. These Southern Hemispheric temperatures are similarly low because the Antarctic ice sheet grows relatively little in size towards colder-than-PI conditions (see also Stap et al., 2014). When Antarctica is allowed to melt in warm climates, however, the local temperature increase becomes much stronger: 11.6 instead of 5.9 K with respect to PI. In these conditions, we find that coupling albedo changes leads to a maximum Antarctic temperature anomaly of only 7.0 K (Fig. 8b, black dots). When only surface height changes are coupled, this anomaly reaches 7.4 K (Fig. 8b, orange dots). This result implies that albedo changes are relatively less important in Antarctica than in the high latitudes of the Northern Hemisphere. The reason is that the Antarctic continent remains snow covered throughout most of the year when the land ice retreats, which reduces the albedo change (see also Stap et al., 2016a). Since temperature changes are strongest in the Southern Hemisphere in warmer-than-PI climates, this explains the increased relative importance of the surface-height-temperature feedback on ESS in these climates. The different response of the northern and southern high latitudes to CO_2 changes challenges the approach of De Boer et al. (2010) and De Boer et al. (2012), who reconstructed a single high-latitude temperature anomaly. Furthermore, their record cannot readily be translated to global conditions by a constant factor (as is done in e.g Martínez-Botí et al., 2015), because the conversion depends on the prevailing climate state. This problem with recalculating high latitude values in terms of global mean changes also holds for other local proxy data like marine, terrestrial or ice core records.

Finally, we compare the relation between global temperature and logarithmic CO_2 in three model runs with uncoupled ice (Fig. 9). In one run the ice sheets are kept at PD condition as before (now called PD ice, blue dots), in another one we use the LGM condition (LGM ice, black dots), and in the last one all ice is removed (no ice, red dots). Naturally, the more ice is present on Earth, the colder the climate becomes, so the LGM ice run is colder than the PD ice run, which in turn is colder than the no ice run. The difference between the PD ice and the no ice run is fairly uniform over the whole CO_2 range. The difference between the LGM ice and the no ice run, however, is larger in cold climates than in warm climates as it shrinks from $\sim$2.8 to $\sim$1.6 K. This is explained by the extra land ice in the LGM ice run cooling the climate and increasing the area on Earth covered by snow and sea ice. As a result of this area increase, the land surface has a higher albedo, which cools the climate further. In cold climates this effect is stronger because the snow- and sea ice-covered area grows more towards the equator, where there

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is more incoming solar radiation. Consequently, the albedo increase is more effective as it leads to absorption of more energy, and thus to a stronger temperature decrease.

5   Summary and conclusions

We have presented mutually consistent transient simulations of atmospheric CO_2 content, temperature, ice volume and benthic $\delta^{18}O$ over the past 38 million years. They were obtained using a coupling between a zonally averaged energy balance climate model and a one-dimensional ice sheet model. As forcing, we have used an inverse routine that yields atmospheric CO_2 from an observed benthic $\delta^{18}O$ record (Zachos et al., 2008). This allowed us to simulate periods before 800 kyr ago, for which ice core records are not available and CO_2 data are uncertain, scarce and intermittent (Beerling and Royer, 2011). Focusing on long-term interactions between land ice and climate, we haven taken a complementary approach to snap-shot and short timescale simulations that have been published before (e.g. Langebroek et al., 2010; Ladant et al., 2014; Gasson et al., 2014; Pollard et al., 2015). Our coupled model results represent an improvement upon the work of De Boer et al. (2010), who used the same ice-sheet model in stand-alone form to simulate the past 40 Myr (De Boer et al., 2010, 2012). The inclusion of a climate model has enabled us to simulate, and force the different ice sheets with, seasonal meridional temperature distributions instead of globally uniform perturbations to present-day climate with a fixed seasonal cycle. Nonetheless, we recognise that our coupled ice sheet-model is relatively simple. It does not include ice-shelf dynamics and sophisticated grounding line parameterisations (Pollard and DeConto, 2012; Pollard et al., 2015) However, more complex models, such as GCMs coupled to three-dimensional thermodynamic ice models, are as of yet not suitable to perform multi-million year integrations because of limited computer power. Our results therefore serve as a reference, to which results of these more sophisticated models can be compared once they are achieved. This facilitates an analysis of which features appear in the full hierarchy of models and which are specific to more comprehensive models including more physics. Furthermore, by comparing our fully coupled simulation to model runs with the ice-albedo feedback, or the surface-height-temperature feedback, or both switched off, we have quantified the effect of ice-sheet climate interactions on Earth System Sensitivity (ESS) and polar amplification on long time scales.

In our model, the results for CO_2 concentrations lower than roughly 450 ppm depend on the transient evolution of CO_2. When during the run the model has previously experienced high CO_2 values, temperatures are higher than when this is not the case. This hysteresis is persistent even in runs without any change in albedo due to snow-, sea ice- or permanent land ice-cover and without changes in ocean overturning strength. However, these factors do enhance it. It is still unknown whether this is an artefact of our model or is also exhibited by other models. We therefore suggest that in the future, climate models should be tested for this behaviour by confronting them with high CO_2 values before simulating cooler climates.

As was already demonstrated in Stap et al. (2016b), our model is unable to capture the difference in CO_2 suggested by proxy data between the time right before the Eocene-Oligocene Transition ($\sim$34 Myr ago) and during the Middle Miocene Climatic Optimum ($\sim$15 Myr ago). This is because the forcing benthic $\delta^{18}O$ values are similar during these times. Our simulation of CO_2 may be improved by extending the model with more aspects of the climate system, moving towards a full Earth System

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Model. Important aspects hitherto neglected in our model are the effects of dynamic vegetation (e.g. Knorr et al., 2011; Liakka et al., 2014; Hamon et al., 2012), and changing topographic boundary conditions as a result of tectonics and erosion (Wilson et al., 2012; Gasson et al., 2015; Stap et al., 2016b). Ultimately, the model could also be coupled to a carbon cycle model, e.g. BICYCLE (Köhler and Fischer, 2004), in order to simulate climate using only insolation data as input.

5   In our model, ice volume changes enhance the modeled effect of CO_2 on temperature via the ice-albedo and the surface-height-temperature feedbacks, particularly in the regions where the ice sheets are located. At low CO_2 values, the Northern Hemispheric ice sheets change in size, causing large fluctuations in the temperature on this hemisphere. The ice-albedo feedback is much stronger than the surface-height-temperature feedback in these conditions (see also Stap et al., 2014). This is reflected in the Northern Hemispheric (40 to 80° N), Antarctic (60 to 90° S), and global temperature profiles. At intermediate
10   CO_2 values, there is only weaker land ice volume variability, as in the Northern Hemisphere there is little land ice left to melt, and in the Southern Hemisphere it is not yet warm enough for deglaciation of Antarctica. Consequently, temperature changes are only minorly enhanced, both locally and globally. At high CO_2 values, the Antarctic ice sheet is more dynamic, so that temperature changes more strongly on the Southern Hemisphere. Here, the impact of the ice-albedo feedback is weaker, since most of the continent remains snow-covered during large parts of the year when the ice sheet retreats. Hence, the surface-height-
15   temperature feedback becomes relatively more important. When the ice sheets are kept constant, temperature perturbations are much less strong and more uniformly distributed over the globe.

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

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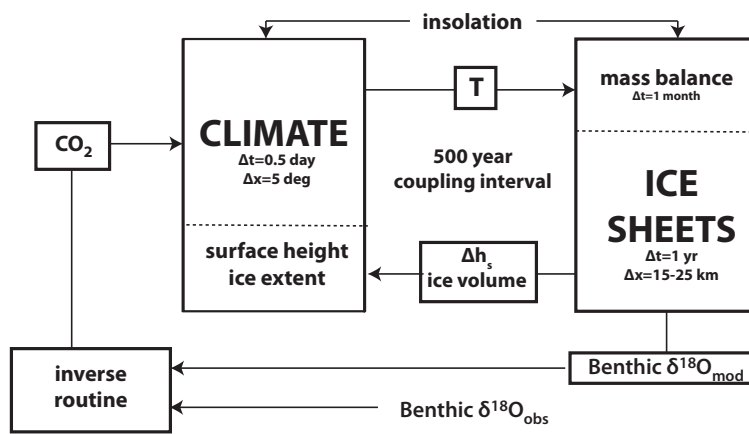
Figure 1. Schematic overview of the coupling of the zonally averaged energy balance climate model and the one-dimensional ice sheet model.
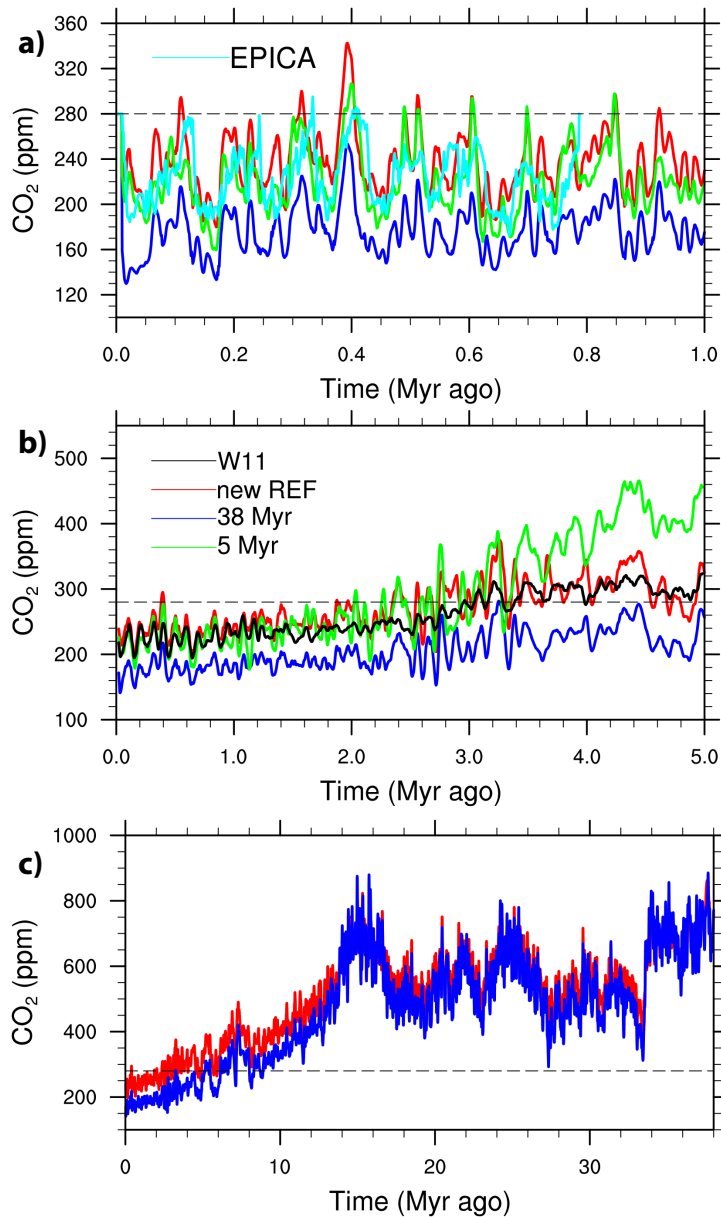
Figure 2. Atmospheric CO_2 over the past 1 Myr (a), the past 5 Myr (b) and the past 38 Myr (c). Shown are the 5-Myr run from Stap et al. (2016a) (green), the extended 38-Myr run from Stap et al. (2016b) (blue), the new reference run with altered cloud optical thickness (red), the hybrid proxy data-model reconstruction by Van de Wal et al. (2011) (W11; black), and the ice core record (Petit et al., 1999; Siegenthaler et al., 2005; Lüthi et al., 2008) (EPICA; cyan). The ice cores stem from Antarctica, the oldest values are from the EPICA Dome C core. Mind the differing y-scales. The dashed line shows the pre-industrial value (280 ppm). In panel (a) we do not show 40-kyr averages, but the 1-kyr output of the model.
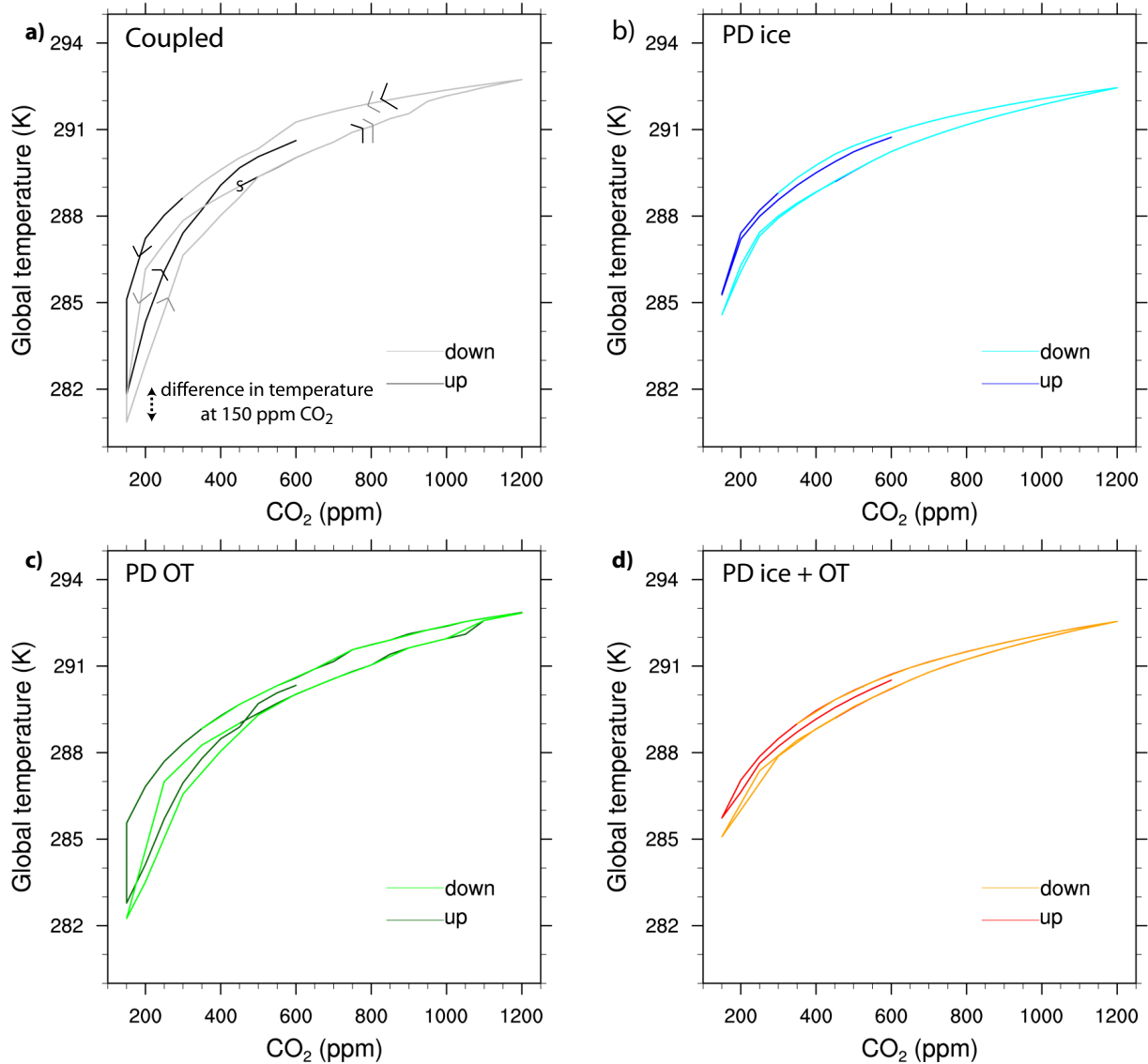
Figure 3. Relation between CO_2 and global temperature in the equilibrium runs. In (a), the fully coupled model output is shown. The startpoint of the simulation at 450 ppm CO_2 is marked by an S, and the consequent evolution for both runs is marked by colored arrows. The black line shows the up run, where CO_2 is increased first, the grey line shows the down run, where CO_2 is decreased first. At high CO_2 levels, the black line is overlaid by the grey line. In (b) the output with uncoupled ice (blue/cyan), in (c) with uncoupled ocean overturning strength (darkgreen/green) and in (d) with both these factors uncoupled (red/orange) are shown. The darker colors (blue, darkgreen, red) show the up runs, the lighter colors (cyan, green, orange) show the down runs. The startpoint and evolution are the same as in (a).
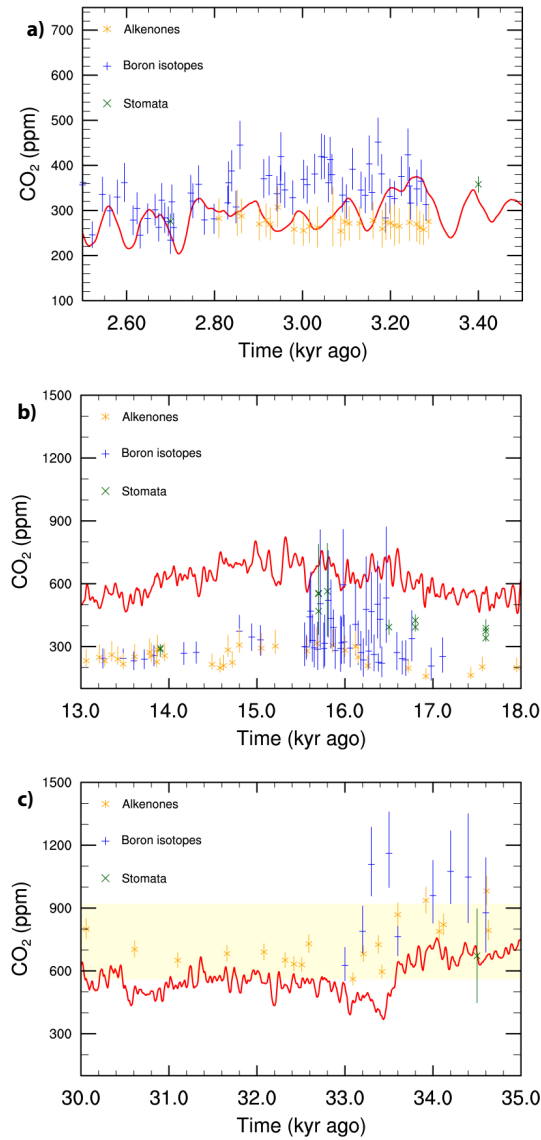
Figure 4. Simulated CO_2 concentrations of the new reference run for the (a) Mid Pliocene Warm Period (3.5 to 2.5 Myr ago), (b) Middle Miocene (18 to 13 Myr ago) and **(b)** Early Oligocene (35 to 30 Myr) ago. Shown are 40-kyr running averages. Proxy-data reconstructions based on alkenones (Pagani et al., 1999, 2011; Badger et al., 2013) are indicated by orange asterisks. Boron-isotope-based data (Pearson et al., 2009; Foster et al., 2012; Greenop et al., 2014; Martínez-Botí et al., 2015) are indicated by blue plusses. Stomata-based data (Van der Burgh et al., 1993; Kürschner, 1996; Kürschner et al., 2008; Retallack, 2009) are indicated by green crosses. Yellow shading indicates the range of Antarctic glaciation values found by Gasson et al. (2014).
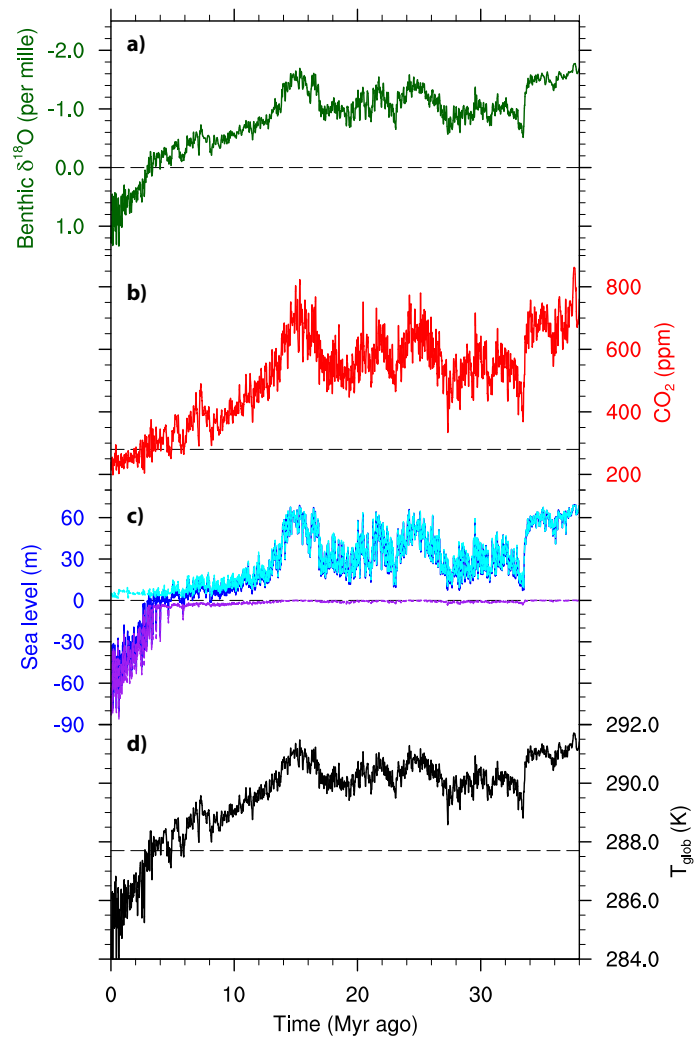
Figure 5. Results of the new reference run: (a) benthic $\delta^{18}O$, (b) atmospheric CO_2, (c) ice-volume-equivalent sea level in meters above present day (blue), and contributions from the Northern Hemispheric (purple) and Antarctic ice sheets (cyan), (d) global mean temperature (T_{glob}). Shown are 40-kyr running averages. Dotted lines represent pre-industrial (PI) values.
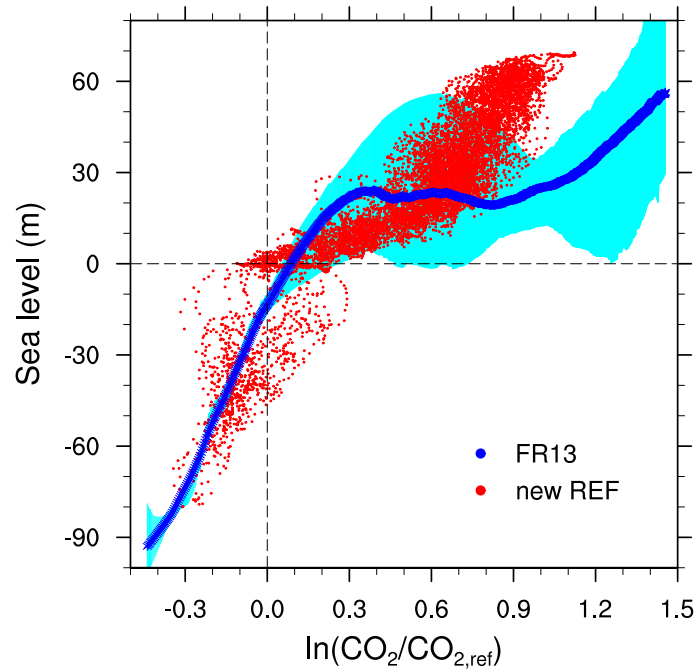
Figure 6. Relation between the logarithm of CO_2 divided by the PI value of 280 ppm, and ice-volume-equivalent sea level anomalies with respect to PI, for the reference simulation (red dots), compared to the median case of the probabilistic data analysis in Foster and Rohling (2013) (FR13, blue dots) with their 95% uncertainty range in cyan.
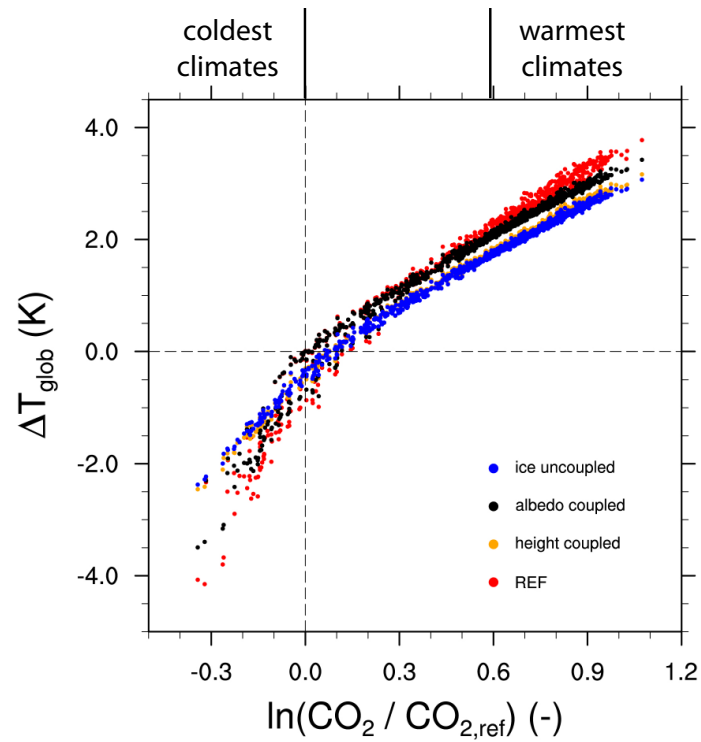
Figure 7. Relation between the logarithm of CO_2 divided by the PI value of $280\,\mathrm{ppm}$, and global temperature anomalies with respect to PI (of the reference run), for the reference simulation (red dots), the simulation with uncoupled ice (blue dots) and the simulation with only surface height (orange dots) or albedo (black dots) coupled.
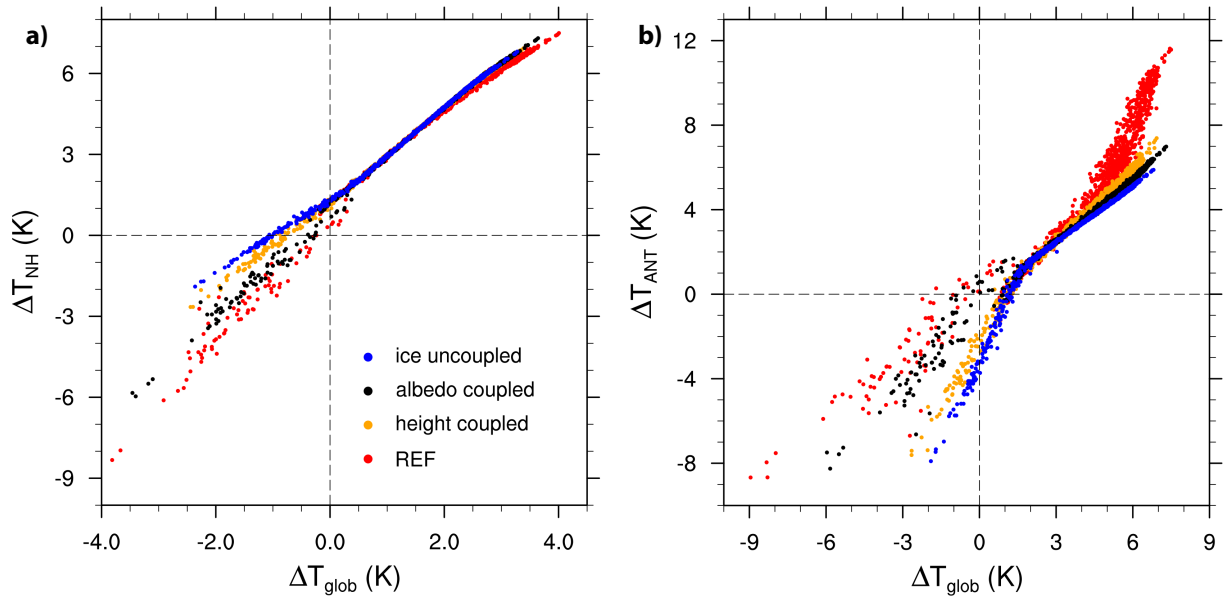
Figure 8. Relation between anomalies with respect to PI (of the reference run) of global temperature, and (a) Northern Hemispheric temperature (40 to 80° N), and (b) Antarctic temperature (60 to 90° S), for the reference simulation (red dots), the simulation with uncoupled ice (blue dots) and the simulation with only surface height (orange dots) or albedo (black dots) coupled.
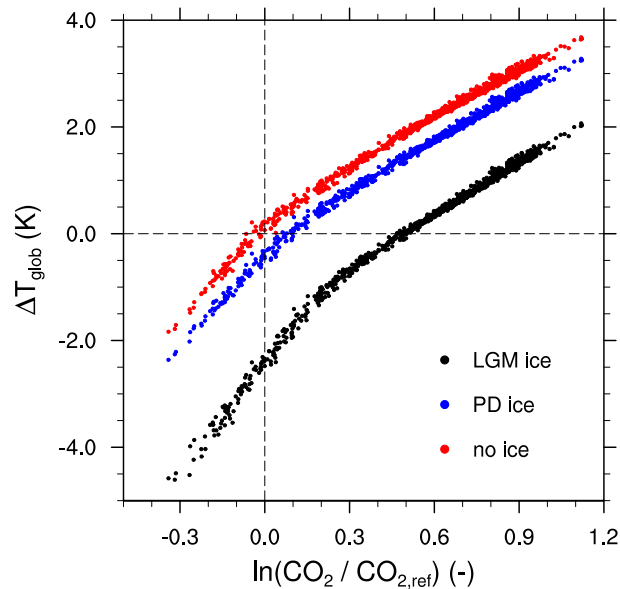

Figure 9. Relation between the logarithm of CO_2 divided by the PI value of 280 ppm, and global temperature anomalies with respect to PI (of the reference run), for the simulation with ice kept at PD level (blue dots), at LGM level (black dots) and the simulation with all land ice removed (red dots).