Authors' response

We thank all 3 referees for their insightful and constructive comments. Both the second and third referee found our discussion of "other paleoproxies" lacking. We have therefore revised this section and summarized here why our results could potentially be reconciled with ocean d13C despite a higher terrestrial carbon inventory (we also suggest here that increased CaCO3 weathering leads to higher rather than lower ocean d13C due to the input of isotopically heavy weathering products, and acknowledge that lower SSTs would result in lower ocean d13C due to enhanced fractionation at the air-sea interface caused by the cooling):

1) Decrease in marine productivity means lower d13C as phytoplankton discriminate against 13C during photosynthesis, giving the marine organic carbon reservoir a low d13C. Zimov et al., 2009 for instance suggested that their glacial terrestrial carbon increase scenario is only possible if there was a decline in the marine reservoir of organic carbon.

2) Increasing sea ice means lower d13C due to reduced gas exchange

3) Winds are held fixed in our model. However, if these had been lower, the gas exchange at the air-sea interface might have been reduced further.

4) We vary land-ocean CaCO3 fluxes but not land to ocean organic matter fluxes. If our ensemble incorporated enhanced weathering of organic matter and/or reduced deposition of organic carbon at continental margins due to lower sea levels, d13C could potentially have decreased significantly. Wallmann, 2014 for instance attributes the observed decrease primarily to this process.

5) Enhanced glacial carbonate ion concentrations may have reduced the d13C in foraminera shells without altering mean ocean d13C (Lea et al., 1999).

Caveat: LGM deoxygenation over large areas of the deep ocean. This is used in the literature as evidence of a larger LGM ocean carbon inventory (due to reduced ventilation). To challenge this argument, one might show that deoxygenation is partly caused by an increased LGM ocean primary productivity, which means more consumption of O2 at depth. Yet, we are suggesting that LGM POC export fluxes may have been lower. A potential counter-argument therefore is that: yes, productivity was lower but if we were to include processes such as increasing remineralisation depth due to decreasing ocean temperature, or increasing ballasting, more O2 would be consumed at depth. Hence, one could potentially still have deep ocean deoxygenation.

Our conclusion with regard to reconciliation with ocean d13C and oxygen is that conceptually our results can only be reconciled with carbon isotope and oxygen data if additional processes not included in our study are brought into play.

We also note that the manuscript previously included a section on D14C as well as discussion of transient changes in carbon isotopes. In the revision, we remove these components as: (1) we believe these are less crucial, and (2) to keep with the aim of our study, which was not to do a comprehensive assessment of model results against observations. Because of (2), we also remove from the "other paleoproxies" section our suggestions for further simulations.

Responses to Referee 1

General comments

My main concerns are following.

1) In this study, the burial of land carbon by LGM ice sheets was simulated, which previously has not been done in an LGM equilibrium experiment set-up. However, it's not clear how this process is set up in GENIE. The authors attempt to justify the modelled larger LGM land carbon storage and smaller oceanic carbon storage. However, the arguments still seem weak, since many recent studies have suggested the opposite (for instance, Anderson et al., 2019, Global Biogeochemical Cycles).

2) Simulations were carried out with wide ranges of parameters, and some have captured the dramatic LGM CO_2 drop. The relationship among different model variables, without explaining why these relations occur. After reading the whole manuscript, there were still so many question marks left in my mind. As readers of a scientific paper, we often like to see a question got answered within one paper. Thus, I have to say I personally didn't find it a pleasant reading experience.

In the response letter, the author argued: "We note that this paper was intended as part one of two related papers. This paper describing the relationships between ensemble outputs, and the second, currently being finalised for submission, describing dependencies on ensemble parameters to isolate mechanisms. ... " I would argue, that one only need to read the introduction and methodology of this paper and the analysis of the next paper, since the relations have to be mentioned when they are analysed. I do understand that it's not feasible to put everything in one huge paper. However, I don't agree that the value of this manuscript should be so much dependent on the next one.

My suggestion is to re-organise content of this manuscript and of the one in preparation such that in the first one, the methodology is described in detail, the behaviour of the physical states of the (sub)ensembles are described and EXPLAINED. In the second manuscript, the results concerning terrestrial and ocean carbon cycle are described and explained.

While we appreciate and agree that there are advantages and disadvantages to any way of dividing the material, we prefer the present configuration because the physical and carbon-cycle processes are very strongly connected, indeed the principal utility of our modelling framework is the ability to focus on these connections. We therefore prefer to retain physical and carbon-cycle processes in both papers, focusing first on describing the model output variables, potential relationships between them and whether or not in agreement with observations.

3) Section 3.2.6: Other paleo proxies

I think it's inappropriate to discuss extensively about carbon isotopes, which are not at all accounted for in this model. I got two main messages from this session. First, the authors attempt to argue that their modelled lower LGM oceanic carbon storage is possible, although this is not supported by many previous studies. However, I find the supporting arguments rather weak. See the specific comments below.
Second, the authors briefly summarise some previous work on 13C and 14C, and suggest some further simulations (see the specific comments). I would like to ask the authors the purpose of this text. Does it help achieving your research aim?

Our research aim is to describe the range of possible CO2 responses given our uncertainty-based approach, and to a lesser extent, how this our response space compares with observations. To that end, we have reduced the section on other paleo proxies and also modified some of our supporting statements.

Specific comments

Page 1 Line 30 - Page 3 Line 5:
At the beginning of the introduction, the authors spend more than one page to review the hypotheses for the LGM CO2 drop compared to preindustrial. When reading this, I start to expect that this manuscript is about deepening the understanding of these mechanisms or the interplay among them. However, from Page 3 line 6, the authors state that their aim is to examine the sensitivity of LGM CO2 drop to model parameters. So why mentioning at all those mechanisms? At this stage, there seems no link between the mentioned mechanisms and the model parameters to be varied in GENIE. As stated in the comments to the first submitted manuscript, I appreciate that the authors give an comprehensive review of the mechanisms that governing the LGM atmospheric CO2 drop. However, it becomes excessive when it does not prepare the readers for the central topic of this study.
My suggestions for improvement is following:

1) Shorten the paragraphs on LGM CO2 drop mechanisms. Provide the links, if exist, between the mechanisms and the model parameters varied in this study.
2)The authors do state that "There is rarely any investigation of the impact of alternative assumptions regarding parameter values ..." I still think some words should be spent on what has been done in the

(limited) previous studies.

We feel that an overview of the mechanisms is useful since, to the extent possible, we attempt to link these to the model output variables we describe. We also add the following sentence in the introduction to further clarify the aim of this manuscript: “Knowledge of these relationships can in turn inform analysis, in the future, of the relationship between the ensemble parameters and model outputs, in order to isolate individual LGM CO2 mechanisms”.

With regard to what has been done in previous studies in terms of investigation of the impact of alternative assumptions regarding parameter values, we added the following sentence: “An example of a relevant study is Bouttes et al., 2011, which varied model parameters controlling the importance of iron fertilisation, bring rejection and stratification-dependent diffusion in an ensemble setting, assessing the agreement of the model output with data.”

Page 3 Line 15

It has not yet been implemented in GENIE in an equilibrium set-up, or it has not been implemented in other models?

We have specified that it is the first time it is done in models more generally.

Page 4 Line 29-31

Have the OL1 and VPC been explained in previous GENIE papers? If not, please provide more details about what do these parameters do. e.g., what happens if OL1 is increased.

We clarify that these parameters have been explained in Holden et al., 2013b.

Page 7 Line 10-11

What is effect of bioturbation, to increase vertical diffusivity in the sediment pore water? If so, the sediment equilibrium faster with bioturbation.

The effect is to mix sediment material. Bioturbation is turned off to allow surface sediment composition and CaCO3 burial rates to come to an equilibrium faster. (http://www.seao2.info/cgenie/docs/cGENIE.muffin.Examples.pdf)

Page 8 Line 3-6

Is it the first time that the burial of terrestrial carbon under LGM ice sheet is implemented in GENIE? If yes, please provide here details about how it is done. If not, please provide references.

We include more details in the revised manuscript.

Page 12 Line 9-12

Here salinity rather than sea level is shown and is (mainly) described. So please consider change the title.

We have revised the title.

Page 16 Line 8

There was no "China" 21k before present. "Eastern Asia" instead?

We use Eastern Asia in the revised manuscript.

Page 25 Line 23

I did not find observational data in Fig 11.

Correct. We have revised the text accordingly.

Page 30 Line 16-18

It has been mentioned many times that "We estimate that up to ~60 ppmv of ΔCO_2". Please provide information about how this was done.

We clarify in the revision that this is based on our expert opinion.

Page 33 Line 23-24

I find this argument weak. Indeed the LGM marine primary production is lower in the area of sea ice. However, both modelling and proxy data studies (e.g. Muglia et al., 2018, Earth and Planetary Science Letters, and references therein) have suggested that the LGM primary production in the Southern Ocean is higher compared to pi.

We discuss POC export in more detail on pp29-30 but reference Zimov et al., 2009 here as the authors there argued that the decrease in ocean $\delta^{13}C$ could have been caused by a decrease in marine productivity.

Page 34 Line 1-3

I would expect the opposite: LGM ocean delta_13C is lower when CaCO3 weathering flux is higher. Increased CaCO3 weathering flux results in the uptake of CO2 into the ocean. The atmospheric delta_13C is low (< -6 permil). Thus the uptake of atmospheric CO2 will decrease oceanic delta_13C.

Page 35 Line 5-6

There are other methods to estimate LGM ocean carbon inventory, e.g. based oxygen (see Anderson et al., 2019, Global Biogeochemical Cycles).

Agreed, we discuss this in the final paragraph of section 3.2.6.

Page 35 Line 15-30

Instead of blaming missing processes in the model, why not just start analysing your ensemble results to investigate how oceanic oxygen, carbon storage, and so on, respond to different physical states?

We believe our parameter-space exploration is sufficiently comprehensive to be relatively confident that the discrepancy strongly indicates the importance of processes not represented in the model. Indeed, this is an important result of the extensive computational experiment undertaken and described. As we have explained, our preference is to present the evidence for this in terms of the full analysis of the model states compared to observations in this, already long, paper, and reserve the emulation-based factor analysis for a later paper, which will necessarily be another long paper for a proper exposition. Attempting to add a very short section on the factor analysis to this paper would, in our view, be necessarily incomplete and hence highly unsatisfactory.

Figures

1) When describing model results shown in a figure (or a table), it would be easier for the readers to follow if "what is shown in which figure(table)" is given at the beginning of the text. The authors have all the pictures in mind but the readers don't.

In several places, we have included a statement of the following kind: “As shown in Fig”.

2) There are too many figures. Some figures could be combined. For instance, Fig 1 (a) and (b) can be combined into one figure. And in this way one can clearly compared different ensemble sets. In the case of Fig 1, the authors could also consider using nonlinear vertical axis such that the low frequencies are visible.

Where ENS16 and ENS104 were shown separately, these have been combined into one figure. In the case of Fig. 10 we also noticed that ENS16 was not previously shown so we include it in the revised manuscript.

3) Fig 9

Please use the same colorbar range for (a) and (c), such that it's easier to compare between LGM and pi.

We have modified the figure as suggested.

Responses to Referee 2

Interactive comment on "Coupled climate-carbon cycle simulation of the Last Glacial Maximum atmospheric CO_2 decrease using a large ensemble of modern plausible parameter sets" *by Krista M. S. Kemppinen et al.*

Pearse James Buchanan

February 2, 2019

Institute for Marine and Antarctic Studies, University of Tasmania, Australia.

In their revised manuscript, Kemppinen and coauthors have made a strong attempt to address the many demands of the three reviewers. Overall, I am pleased with the revisions but am still left unsatisfied. I advocate for further revisions before publication.

The introduction is better, although still needs some work. The authors introduce many mechanisms in the Introduction that might affect CO2. I appreciate that the authors are trying to cover a lot of ground in their introduction, but to simply name a mechanism and then not explain to the reader how that mechanism or change works to affect CO2 is confusing and can be misleading to the reader. I realise that it is beyond the scope of this work to present explain how every mechanism in the biosphere/lithosphere can alter CO2, but I think it is equally poor to name all of them and then not give the reader an explanation of why they are important. For instance, the authors list many factors in the 3rd paragraph in just one sentence and fail to provide any conceptual clarity. Possibly, restricting this initial discussion of mechanisms to just a few important ones would be fine, so long as you detail the links to climate. Alternatively, a table/schematic that describes each mechanism you list in the introduction and takes the reader through its links to global CO2 and climate (i.e. its drivers, processes and effect on CO2) would be a helpful addition.

We have clarified the linkages between the processes and CO2 and/or the drivers of the processes.

The methods are clearer.

The results section is clearer and more concise.

The abstract is much clearer, more interesting and conveys the major findings. I would advocate for one more sentence, however, that describes why an increase in TerrC and a decrease in OceanC is associated with a pCO2 drawdown. Something to the tune of "preserving, rather than destroying respired C buried under ice sheets and slower remineralisation rates in soils". I think this will be of interest to the field and improve the popularity (or notoriety?) of the study.

We have modified the abstract to reflect this suggestion.

To this point, I am still confused about what the authors mean when they talk about ice-sheet carbon. My confusion stems from an ignorance of what processes are affecting carbon when the ice sheets grow. The authors state that "The increases in terrestrial biosphere carbon are predominantly due to our choice to preserve rather than destroy carbon in ice sheet areas. However, the ensemble soil respiration also tends to decrease significantly more than net photosynthesis resulting in relatively large increases in non-burial carbon". Great, I follow this. Currently, I understand that carbon is preserved in the soils as ice sheets grow over these areas, and that cooler temperatures globally also tend to increase carbon in soils where vegetation is present.

However, then the authors make statements like "It is the combination of our ice sheet carbon stocks increasing rather than decreasing when exposed to LGM climate, and our choice to preserve rather than destroy this carbon", and I am lost. Why does carbon *increase* under ice sheets? It is entirely possible that carbon is *preserved*, but I do not know what process enables carbon in soil under an ice sheet to increase, and the authors do not offer a process that explains this.

They then go on to say "If most of the carbon that was present in ice sheet areas at the end of the preindustrial runs had been lost to climate forcings, it would not matter much [to what? I assume pCO2] whether the remaining stocks had been destroyed or preserved.". I assume from this statement that the authors mean that carbon tends to accumulate in the terrestrial reservoir regardless of changes under ice sheets because soil respiration is reduced more than photosynthesis under cooler climates, and that this is the most important term. This makes sense when viewed through the lens of the metabolic theory of ecology [by the way it would be good to cite some work of the metabolic theory of ecology (Brown, etc.)]. However, this statement then seems to contradict a previous statement: "If the LGM burial carbon inventories were to be removed, DTERRC would be negative in 13 our of 16 simulations, despite the fact that terrestrial carbon also increases outside of the ice sheet areas in 15 out of 16 simulations." This suggests that burial of carbon from under ice sheets (?) is super important for gains in the terrestrial reservoir, while gains in non-ice sheet areas are not so important.

One of two things are causing confusion and make me sceptical of the results. One, the authors have not provided an explanation of the processes and assumptions of how they treat carbon under ice sheets. More explanation of the processes governing vegetation-soil-burial carbon stocks would be very useful. Such an explanation may be all that is necessary to resolve these apparently conflicting statements that make the results difficult to understand. Second, errors in the treatment of carbon under ice sheets were made in the initial runs, causing carbon to *increase* in non-physical ways, and the authors are trying to hide this, which causes confusion. The authors state in a response to reviewer 3, for instance, that "Our LGM burial carbon estimates include the initial preindustrial carbon inventories, plus carbon accumulated in response to glacial forcings". With all previous information, this statement makes me think that carbon is accumulating under ice-sheets, which is odd. Is this due to some undefined process with physical underpinning, or an unrealistic, non-physical process?

We have clarified the ice-sheet/carbon interaction/link throughout the text. In summary, however, carbon does not get modified further once it is covered by ice.

I also have some concerns with regards to the discussion of $\delta^{13}C$ in the "Other paleo proxies" section. A strong reason for a lower terrestrial C reservoir under glacial conditions is because of the decrease in $\delta^{13}C$ recorded in benthic foraminifera recovered from glacial sediments. The decrease in $\delta^{13}C$ is thought to originate from the land, as terrestrial organic matter with very negative $\delta^{13}C$ signatures found its way into the atmosphere as the terrestrial reservoir diminished.

Carbon in the atmosphere then moved into the ocean, leading to an explanation of atmospheric CO2 drawdown and the lower oceanic δ^{13}C values. If the major result of this study is to hold, then the authors must present a new and plausible explanation of this decrease in the ocean that does not contradict their simulated increase in terrestrial C.

To do this, the authors invoke a reduced marine productivity, low sea surface temperatures, and greater sea ice extent. However, they do not go further to explain why these features of a glacial climate would decrease δ^{13}C in the ocean without requiring the loss of carbon from the land. My reasoning suggests that:

1. Low marine production would actually increase d13C, because less negative d13C would be transferred via organic matter to depth. NOT CONSISTENT.
2. Low SST would make δ^{13}C more positive because fractionation during outgassing of CO_2 (which is what you're invoking) would leave more ^{13}C in the ocean. NOT CONSISTENT.
3. Greater sea ice area would limit production, as you have stated, which would increase δ^{13}C. NOT CONSISTENT.

The authors then mention other studies that could change δ^{13}C over the glaciation and deglaciation (in the subsequent paragraph), and I suppose their aim here is to introduce uncertainty for the causes of the trends in δ^{13}C (and radiocarbon) to challenge the accepted wisdom. However, with the exception of the Lea 1999 study they cite, I am left unconvinced by their argument and therefore sceptical of their result. Once again, the authors seem to list off many studies without talking through in a conceptually clear manner why that response/mechanism could help to explain their results. Moreover, their mention of Hain et al (2011) is misleading, because this study did not discount depleted Δ^{14}C in a glacial ocean, but rather the "extremely" depleted values found in the Pacific by Stott et al (2009) and Marchitto et al (2007). An increase in the carbon reservoir of the ocean was not challenged by Hain et al (2011).

It may be true that Hain et al., 2011 do not negate the presence of an increased ocean carbon reservoir. However, the authors do cast doubt on both the interpretation of the mid-depth D14C anomalies and the existence of an extremely isolated deep reservoir due to constraints of sediment CaCO3 and ocean O2 (and atmospheric CO2).

Major revisions are once again needed. Overall, I am of the opinion that many of my concerns could be allayed if only more effort was put into making the arguments clearer. So, if the authors can (1) improve the introduction by clarifying why a certain process affects CO2 or do not invoke that process (alternatively schematic/table for reader), (2) clarify the processes that allow carbon to increase under ice sheets over their LGM simulations, and (3) improve their discussion of their results against the carbon isotope and paleo proxy data sets, then I advocate publication.

I **strongly suggest** that the manuscript provides a more thorough description of how the model treats carbon under ice sheet growth. This may allay my concerns totally and I believe would focus and strengthen the later discussion of alternative processes to explain the carbon isotopes.

1 General comments

Some other general concerns that I had as I read the article:

- I think that the Results sections might be better if they were split into "Pre-Industrial conditions", "LGM ensemble conditions", and "Carbon cycling through terrestrial, ocean

and lithospheric reservoirs". This would be a nice way to present the main features of the results, which are pretty cut and dry in the first two sections. The real meat of the paper lies in the consequences for carbon cycling, and an interested reader could flick between "LGM ensemble conditions" and "Carbon cycling through terrestrial, ocean and lithospheric reservoirs" sections to see changes in conditions and consequences for carbon, respectively.

We think this is a good suggestion. However, since we also discuss changes not impacted by the physical conditions (weathering) and changes impacted by changes in the carbon stocks ($CaCO_3$ burial), we believe that the present structure works better. Our description of the model results has, however, been modified to make for an easier read.

- My initial suggestion regarding overlay of the red (ENS-16) over the yellow (ENS-104) bars in the histograms still stands. This would reduce a lot of unnecessary replication in the figures. The authors could use a transparency setting to ensure that the red and yellow are easily seen if they have the same number of experiments (frequency), and possibly use a break in the vertical axis to emphasise the lower frequency ENS-16 experiments. I note that reviewer 3 also suggested this.

We have included ENS16 and ENS104 in the same figure.

- Another suggestion RE figures is that you use a bimodal colour scheme to present changes in your spatial plots. Reds for positive, blues for negatives and centre the range around zero so places with no change are clearly seen. It is misleading to readers assessing your results to present unbalanced colour schemes when discussion change.

We have modified the figures to reflect this suggestion.

- Once again, the writing needs some attention. There are too many adjectives, unnecessarily difficult acronyms, long subordinate clauses, and double negatives in some instances. Please make it easy for the reader. My native language is english, so I mostly understand with a repeat reading of sentences, but many scientists are not.

As suggested above, we have reviewed the writing throughout the text.

2 Specific comments

Abstract

Introduction

- Paragraph 3 is one long sentence. Please break it into more sentences to make it easier to read.

We have revised the third paragraph.

- In my opinion, the paragraphs 5 and 6 of the introduction are unnecessary. You cover these points in the methods and abstract. The content could be reiterated in a small paragraph under the main Results heading.
- The final paragraph is unnecessary, but I understand that reviewer 1 thought it was a good idea. Up to you.

We indeed feel that the final paragraph adds clarity. For this reason also we keep the preceding paragraphs: they include which model variables will be evaluated (reviewer 3 suggestion), and since we review mechanisms at the beginning of the introduction, we want to mention that not all are included in our ensemble, soon after this.

Methods

- Page 4, line 16 - What is EFPC? you do not define it. I see you mention it only twice. Just spell it out.

The revision incorporates the suggested edit.

- Page 7, line 23 - You mention that the model requires a detrital flux field that is specific to the ocean component. I strongly suggest you explain what this detrital flux does and how it affects your carbon cycling in the ocean between LGM and PI experiments. It might be explained more thoroughly in Ridgwell and Hargreaves (2007), but I would like clarification here.

We do not explicitly model all sediment material (e.g. opal) and therefore prescribe these components to get the right wt% distribution of $CaCO_3$ in sediments. The detrital flux does not affect the carbon cycling in the ocean between LGM and PI experiments.

- Page 8, line 10 - "If one expects..." Please rephrase this sentence. It took me three times over to make sense of it.
- Page 8, line 24 - "In the latter case,..." Please make it more clear what you mean by this sentence. I logic is not clear.

The revision takes into account the above suggestions.

Preindustrial simulations

- Page 10, line 7 - "deemed not uncontrovesially implausible" is a double negative. Make it easy for the reader. "deemed plausible."

The manuscript has been revised as suggested.

LGM ensemble simulations

- Page 12, lines 9-12 - can you provide changes in mean salinity in units of psu alongside your percentages?

We present percentages rather than psu units because the variable we describe is the FFX parameter (freshwater scaling flux factor), expressed as a percentage increase in LGM salinity. A value of 1.5 is designed to yield a sea level drop of 120 m and concomitant 3% increase in global salinity. We vary FFX between 1 and 2 in the ensemble.

In reviewing this section, we noticed that the sentence "there is, however, a weak positive correlation between %S and ΔCO_2 in ENS_{315} (r=0.17)" ought to be removed as the correlation cannot be deemed significant.

- Page 21, line 4 - You still use PGACF acronym here despite using ENS-315 elsewhere. Also, PGACF is not defined.

We have replaced both instances of PGCAF remaining with ENS104.

- Page 23, line 22 - The increase in terrestrial carbon under ice sheets as a result of LGM climate forcings needs to be clarified both here and up front in the Intro/methods.
- Page 24, line 3 - After reading this sentence many times, I think I now understand what you mean. You mean that if all carbon under ice-sheets were to have been destroyed then TerrC would have decreased under LGM scenarios. Could you make this clearer by being simpler?
- Page 25, line 1 - Why would vegetation carbon increase under an ice sheet?
- Page 25, line 11 - "Here, the increase in terrestrial biosphere carbon both inside and outside of the ice sheet areas, are presumed to reflect the decrease in soil respiration rate due to colder SATs exceeding the decrease in photosynthesis rate due to lower CO_2, SAT and precipitation, as they are mainly driven by soil carbon increases." Again, how could an increase in soil carbon occur under an ice sheet that should not have any vegetation providing a carbon input?
- Figure 11 - It would be helpful to see the outline for where the ice sheets were placed in the simulations.

As suggested above, we have gone through mentions of the ice-sheet/carbon interaction/link in the manuscript and clarified the text throughout.

For the ice sheet outline, we could provide as a supplementary material, the following figure:

The Antarctic, Greenland and LGM ice sheets are identified in Fig. S1 by their surface albedo (0.8), as there is no default ice sheet output. The surface albedo field is from a random LGM simulation, at t=2000 years, and with snow albedo = 0.6 to ensure the ice sheets are distinguishable from snow.

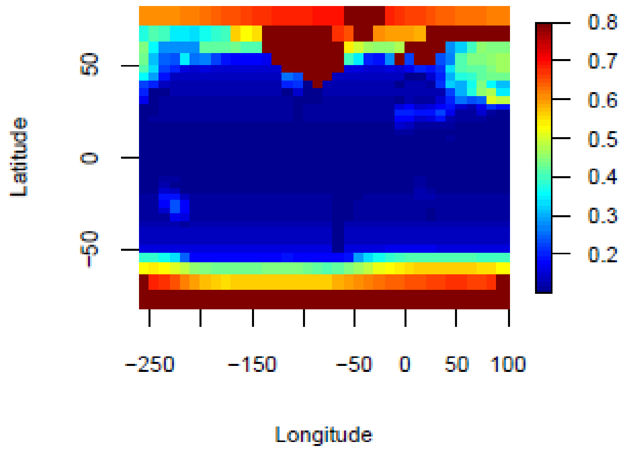

Fig. S1. Surface albedo. LGM ice sheets are shown in brown (albedo of 0.8).

- Page 26, line 11 - "The highly negative LGM terrestrial carbon changes..." This is an example of too many adjectives. "The loss of terrestrial carbon under LGM conditions" is easier to read.

We have modified the sentence in the revision.

- Page 27 - Table 5 is not simple to read. Perhaps you could represent this in a better way? Maybe a figure? It just takes a long time for me to look at it and understand what is going on given the headings and the lists of numbers.
- Page 29, line 6 - Just for your information, the Buchanan 2016 study did more than just make estimates of POC export during a glacial climate, and integrated global climate changes in physical variables including temperature, sea ice, circulation.
- Page 30, line 7 - again PGACF is used.

We have changed the description of the table to make it easier to understand. This should now also be aided by the modifications pertaining to the ice sheet/carbon interaction/link.

Conclusion

- The first paragraph is basically the same as what is in the introduction. Simplify and reduce.

The first paragraph has been revised in the manner suggested above.

- Page 36, line 14 - "equally plausible" of what. Once again, I really advocate for the authors to specify what they mean.

This has now been clarified in the revision.

- And you cannot say "broad agreement" until there is a better argument for why your results could be compatible with a lower $\delta^{13}C$ in the ocean.

This section has been revised.

3 Technical corrections

1. Page 3, line 14 - replace "gets" with "is"
2. Page 5, line 18 - replace "which" with "those that" in both instances.

We have incorporated the technical corrections.

Responses to referee 3

Second review of " Coupled climate-carbon cycle simulation of the Last Glacial Maximum atmospheric **CO2** decrease using a large ensemble of modern plausible parameter sets"

The changes made by the authors result in a much better article, but there are a few changes that should have been made and that seem to have been forgotten, in particular concerning figures (changing the scale, merging all three ensembles on the bar plots). Also the ensemble names have been changed but the old names have been used in a few places. In addition, I have a few technical comments. Once these changes have been made this article is suitable for publication.

In the revision, we have changed, as appropriate, the figure scales and merged the ensembles. Where old names were forgotten, these have also been replaced, and the technical comments below addressed.

Technical comments

p4. l.24 “The ensemble”: could you be more precise? For example, it could be replaced by “the simulation ensemble”?

p.5 l.3 A point is missing after “(Holden et al., 2010b)”

p.5 l. 5 Replace “global average carbonate preindustrial weathering rates” by “global average preindustrial carbonate weathering rates”

p.7 l.10 replace “..” by “.”

p.7 l.18 Define what you call “snowball Earth” (in terms of temperature for example, what threshold you used).

p.7 l.25 I suggest to remind what GWS is: “scaled by GWS (land to ocean bicarbonate flux scaling factor).”

p.9 Figure 1 (and following) As said in the previous stage of review, you could plot all three ensembles (grey, red and yellow) on the same plot.

p.10 l.32. Replace “simulate” by “simulated”

p.12 l.9-10 Put the units in the right place just after the numbers given: “level is 2.84 ± 0.62 and the range is 2 to 4” -> “level is 2.84 ± 0.62 % and the range is 2 to 4 %”

p.12 l.11 Replace “in ENS104 than in both ENS315” by “in ENS104 and in both ENS315”

p.13 l.6 replace “..” by “.”

p.15 Figure 6 Could you have the same scale for panels a and c?

p.16 l.8 Replace “(Bartlein et al., 2011 in Alder and Hostetler, 2015)” by “(Bartlein et al., 2011; Alder and Hostetler, 2015)”

p.17 l.9 replace “..” by “.”

p.20 Figure 9 Could you also have the same scale for panels a and c here?

p.21 l.4 Replace PGACF by ENS104

p.21 l.8 Re-explain here what scenario 1 is when you mention it.

p.26 l.15 no coma after “further”

p.30 l.7 Replace PGACF by ENS104

p.32 l.8 Re-explain what %LOC is

p.34 l.27 replace “2013 in Köhler et al.,” by “2013; Köhler et al.,”

Coupled climate-carbon cycle simulation of the Last Glacial Maximum atmospheric CO_2 decrease using a large ensemble of modern plausible parameter sets

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Abstract. During the Last Glacial Maximum (LGM), atmospheric CO_2 was around 90 ppmv lower than during the preindustrial period. The reasons for this decrease are most often elucidated through factorial experiments testing the impact of individual mechanisms. Due to uncertainty in our understanding of the real system, however, the different models used to conduct the experiments inevitably take on different parameter values, and different structures. In this paper, the objective therefore, is to take an uncertainty-based approach to investigating the LGM CO_2 drop by simulating it with a large ensemble of parameter sets, designed to allow for a wide range of large-scale feedback response strengths. Our aim is not to definitely explain the causes of the CO_2 drop but rather explore the range of possible responses. We find that the LGM CO_2 decrease tends to predominantly be associated with decreasing sea surface temperatures (SSTs), increasing sea ice area, a weakening of the Atlantic Meridional Overturning Circulation (AMOC), a strengthening of the Antarctic Bottom Water (AABW) cell in the Atlantic Ocean, a decreasing ocean biological productivity, an increasing $CaCO_3$ weathering flux, and an increasing deep-sea $CaCO_3$ burial flux. The majority of our simulations also predict an increase in terrestrial carbon, coupled with a decrease in

ocean and increase in lithospheric carbon. We attribute the increase in terrestrial carbon to a slower soil respiration rate, as well as the preservation rather than destruction of carbon by the LGM ice sheets. An initial comparison of these dominant changes with observations and paleo-proxies other than carbon isotope and oxygen data ~~-~~suggests broad agreement. ~~However, a comparison against carbon isotope data would be needed for a more robust assessment.~~However, we advise more detailed comparisons in the future, and also note that, conceptually at least, our results can only be reconciled with carbon isotope and oxygen data if additional processes not included in our model are brought into play.

1 Introduction

Analyses of Antarctic ice core records suggest that the atmospheric CO_2 concentration at the Last Glacial Maximum (LGM), about 21 kyr ago, was around 190 ppmv, well below the preindustrial atmospheric concentration of around 280 ppmv. The most commonly accepted mechanisms to explain the atmospheric CO_2 decrease include lower sea surface temperatures, which increases the ocean CO_2 solubility (Martin et al., 2005; Menviel et al., 2012), increased iron ~~fertilisatio~~fertilization of ~~n~~ocean phytoplankton by aeolian dust input~~-~~ (Bopp et al., 2003; Oka et al., 2011; Jaccard et al., 2013; Ziegler et al., 2013; Martínez-Garcia et al., 2014; Lambert et al., 2015), ~~sea-ice~~ capping of air-sea gas exchange by expanding sea ice (Stephens and Keeling, 2000; Sun and Matsumoto, 2010; Chikamoto et al., 2012) and ocean circulation/stratification changes increasing the net CO_2 flux into the ocean (Adkins et al., 2002; Lynch-Stieglitz et al., 2007; Skinner et al., 2010; Lippold et al., 2012; Gebbie, 2014; Skinner et al., 2014; Tiedemann et al., 2015; de la Fuente et al., 2015; Freeman et al., 2015). These changes may in turn be ~~,~~ due to a range of possible mechanisms such as increased brine rejection (Shin et al., 2003; Bouttes et al., 2010, 2011; Zhang et al., 2013; Ballarotta et al., 2014), a shift in/weakening of the westerly wind belt over the Southern Ocean (Toggweiler et al., 2006; Anderson et al., 2009; Völker and Köhler, 2013), stronger westerly winds over the North Atlantic (Muglia and Schmittner, 2015), and a reduced or reversed buoyancy flux from the atmosphere to the ocean surface in the Southern Ocean (Watson and Garabato, 2006; Ferrari et al., 2014). A process that is conversely assumed to have contributed to increasing atmospheric CO_2 is increasing salinity and ocean total dissolved inorganic carbon (DIC) concentration in response to decreasing sea level (Ciais et al., 2013).

A dominant assumption is also that the terrestrial biosphere carbon inventory was reduced (Crowley et al., 1995; Adams and Faure, 1998; Ciais et al., 2012; Peterson et al., 2014), in line with independent estimates of an ocean carbon inventory that was enhanced by several hundred petagrams (Goodwin and Lauderdale, 2013; Sarnthein et al., 2013; Allen et al., 2015; Skinner et al., 2015; Schmittner and Somes, 2016). The decrease in terrestrial carbon is generally attributed to unfavourable climatic conditions for photosynthesis, and the destruction of organic material by moving ice sheets (e.g. Otto et al., 2002; Prentice et al., 2011; Brovkin et al., 2012; O'ishi and Abe-Ouchi, 2013). The hypothesis that there was an increase in terrestrial carbon has, however, also been put forward (e.g. Zeng, 2003; Zimov, 2006), with some studies additionally suggesting little net change (e.g. Brovkin and Ganopolski, 2015). Processes proposed to be responsible for the terrestrial carbon increase include growth in 'inert' or permafrost carbon, slower 'active' soil respiration rates, continental shelf regrowth, and the preservation rather than destruction of terrestrial biosphere carbon in areas to be covered by the expanding Laurentide and Eurasian ice sheets (Weitemeyer and Buffett, 2006; Franzén and Cropp, 2007; Zeng et al., 2007; Zimov et al., 2009; Zech et al., 2011).

Other mechanisms which may have affected the LGM atmospheric CO_2 change include changes in carbonate weathering rate, through its control on the ocean ALK:DIC ratio and consequently the solubility of CO_2 ~~-~~(Munhoven, 2002; Jones et al., 2002; Foster and Vance, 2006; Vance et al., 2009; Brovkin et al., 2012; Crocket et al., 2012; Lupker et al., 2013; Simmons et al., 2016). The change in carbonate weathering rate would in turn have been caused by lower sea level, exposing previously submerged rock, the presence of a greater amount of glacial flour, which is more susceptible to weathering (Kohfeld and Ridgwell, 2009), or potentially higher soil carbon content. The lower sea level may also have ~~, decreasing dissolved organic carbon inventory due to a more stratified deep ocean (Ma and Tian, 2014),~~ reduced shallow water carbonate deposition~~, which has the opposite impact of increased carbonate weathering rates~~ by decreasing the area of shallow ocean (Opdyke and Walker, 1992; Kleypas et al., 1997; Brovkin et al., 2007), and increased ~~and increased~~ oceanic PO_4 inventory, ~~due to e.g. lower sea level,~~ alleviating the PO_4 limitation on marine production (Tamburini and Follmi, 2009; Wallmann, 2014, 2015). Other potential CO_2 mechanisms include decreasing dissolved organic carbon inventory due to a more stratified deep ocean (Ma and Tian, 2014), and reduced marine bacterial metabolic rate in response to lower ocean temperatures~~, which~~. The lower metabolic rate acts to decrease the return rate of DIC from the remineralisation of organic material, and hence the concentration of CO_2 at the ocean surface ~~-~~(Matsumoto et al., 2007; Roth et al., 2014). The net flux of CO_2 into the ocean may also have increased due to enhanced diatom production caused by the~~,~~ leakage ~~silicic acid leakage, or the leaking out~~ of silicic acid trapped in the Southern Ocean ~~to fuel diatom production, and hence potentially enhancing the uptake of CO_2~~ (Matsumoto et al., 2002, 2014), ~~Si fertilisation or the fertilization of diatom productivity in response to~~ or increased Si inventory, ~~-~~caused by increased input of Si from wind-born dust or enhanced weathering (Harrison, 2000; Tréguer and Pondaven, 2000). ~~and increased oceanic PO_4 inventory, due to e.g. lower sea level, alleviating the PO_4 limitation on marine production (Tamburini and Follmi, 2009; Wallmann, 2014, 2015).~~

Mechanisms put forward to explain the LGM atmospheric CO_2 decrease arise from paleo-data and model studies. The latter most often involve factorial experiments, introducing mechanisms one at a time. There is rarely any investigation of the impact of alternative assumptions regarding parameter values or model structure. An example of a relevant study is Bouttes et al., 2011, which varied model parameters controlling the importance of iron fertilisation, brine rejection and stratification-dependent diffusion in an ensemble setting, assessing the agreement of the model output with data. Here, our aim is conversely to take an uncertainty-based approach to investigating the LGM CO_2 drop by simulating it with a large ensemble of parameter sets designed to allow for a wide range of large-scale feedback response strengths (Holden et al., 2013a). The objective is not to definitely explain the causes of the CO_2 drop but rather explore the range of possible responses. By *responses* we mean physical and biogeochemical changes in the Earth System (e.g. change in global particulate organic carbon export flux) and how these might be linked to ΔCO_2 and to each other, rather than specific mechanisms (e.g. iron fertilisation). Knowledge of these relationships can in turn inform analysis, in the future, of the relationship between the ensemble parameters and model outputs, in order to isolate individual LGM CO_2 mechanisms. In this study, we furthermore seek to simulate the LGM atmospheric CO_2 drop with the simulated CO_2 feeding back to the simulated climate, which is still infrequently done in LGM CO_2 experiments, and the first time it is done with GENIE-1. Moreover, rather than assuming that terrestrial carbon is~~gets~~ destroyed by the LGM ice sheets, we assume that it is~~gets~~ gradually buried. This assumption has not yet been implemented, in GENIE-1 or other models, in an equilibrium set-up.

Despite our ensemble varying many of the parameters thought to contribute to variability in glacial-interglacial atmospheric CO_2, not all sources of uncertainty can be captured, and this is reflected in our simulated ΔCO_2 distribution. We estimate that up to ~60 ppmv of ΔCO_2 could be due to processes not included in our model and error in our process representations (see section 2.4 for details). We thus treat ~~simulations with~~ ΔCO_2 between ~-90 and -30 ppmv as "equally plausible", and focus on describing the physical and biogeochemical changes seen in the~~is~~ subset of simulations with this ΔCO_2. We also do an initial assessment of how the subset mean and/or dominant (in terms of sign) responses compare against observations and paleo-proxies, including temperature, sea ice, precipitation, AMOC & AABW cell strengths, terrestrial carbon, ocean carbon, particulate organic matter export and deep-sea $CaCO_3$ burial.

Finally, to test the robustness of relationships derived from the analysis of the ensemble subset with ΔCO_2 between ~-90 and -30 ppmv, we briefly compare the physical and biogeochemical changes seen therein with the changes seen in the ensemble with no ΔCO_2 filter, and the ensemble with a more negative ΔCO_2 filter (~-90 to -60 ppmv) (section 2.4). In general, the same dominant relationships between ΔCO_2 and the physical and biogeochemical changes are observed as in the subset with ΔCO_2 between ~-90 and -30 ppmv. In the case of the ensemble subset with ΔCO_2 between ~-90 and -60 ppmv, we additionally look at what proportion of the total terrestrial carbon change comes from within the ice sheet areas, and from there draw conclusions for the rest of the ensemble.

The paper is organised as follows. The introduction section, Section 1, is followed by Section 2, which describes the model, the ensemble, the simulation set-up and the ensemble subsets to be analysed. Section 3 is the results and discussion section, which includes a brief evaluation of the preindustrial (control) spin-up simulation to verify reproducibility of Holden et al., 2013a. The majority of the section is devoted to the LGM simulation: namely, diagnosis of the physical and biogeochemical changes (including potential causal relationships) seen in the subset with ΔCO_2 between ~-30 and -90 ppmv, and to a lesser extent, the ensemble with both more and less constrained ΔCO_2. Comparison of the first subset against observations and paleo-proxies is also included. Section 4 provides the key conclusions.

2 Methods

2.1 The model

The GENIE-1 configuration is as described in Holden et al. (2013a). The physical model consists of a three-dimensional frictional geostrophic ocean model (GOLDSTEIN) coupled to a thermodynamic/dynamic sea ice model (Edwards and Marsh, 2005; Marsh et al., 2011) and a two-dimensional Energy-Moisture Balance Model (EMBM). Atmospheric tracers are a sub-component of the EMBM, with a simple module (ATCHEM) used to store the concentration of atmospheric gases and their relevant isotopic properties (Lenton et al., 2007). The model land surface physics and terrestrial carbon cycle are represented by ENTS (Williamson et al., 2006). The ocean biogeochemistry model (BIOGEM) is as described in Ridgwell et al. (2007) but includes a representation of iron cycling (Annan and Hargreaves, 2010), and the biological uptake scheme of Doney et al. (2006). The model sediments are represented by SEDGEM (Ridgwell and Hargreaves, 2007). GENIE-1 also includes a land surface weathering model, ROKGEM (Colbourn, 2011), which redistributes prescribed weathering fluxes according to a fixed river-routing scheme. The model is on a 36 x 36 equal-area horizontal grid, with 16 vertical levels in the ocean.

2.2 The simulation ensemble

The GENIE-1 ensemble consists of 471 parameter sets, varying 29 key model parameters over the ranges in Table 1. It derives from the 471-member ~~EFPC~~ emulator-filtered plausibility constrained ensemble of Holden et al. (2013a), which varies 24 active parameters and 1 dummy parameter (as a check against over-fitting). The parameter values in Holden et al. (2013a) were derived by building emulators of eight preindustrial climate metrics and applying a rejection sampling method known as approximate bayesian computation (ABC) to find parameter sets that the emulators predicted were modern plausible. Two parameters were later added to the ~~EFPC~~ ensemble, in Holden et al. (2013b), to describe the un-modelled response of clouds to global average temperature change (OL1), and the uncertain response of photosynthesis to changing atmospheric CO_2 concentration (VPC). The parameters are as described in Holden et al., 2013b. We add two further parameters here that represent uncertain processes specific to the LGM. The first (FFX) scales ice-sheet meltwater fluxes to account for uncertainty in un-modelled isostatic depression at the ice-bedrock interface due to ice sheet growth, and for assuming a fixed land-sea mask (Holden et al., 2010b). We vary the parameter in the ensemble to capture the uncertainty in the magnitude of the glacial sea level drop and its effects on the carbon cycle.The second (GWS) scales the global average preindustrial carbonate ~~preindustrial~~ weathering rates for the LGM, to account for uncertainty in carbonate weathering and un-modelled shallow water carbonate deposition rate changes. For both FFX and GWS, uniform random values were derived using the generation function runif in R.

Table 1. Ensemble parameters. Ranges are from (a) Holden et al. (2013a), (b) Holden et al. (2013b), and (c) Holden et al. (2010b), with the exception of GWS (see main text). The table also precludes the dummy parameter.

Module	Code	Description	Range	Ref.
EMBM	AHD	Atmospheric heat diffusivity ($m^2\ s^{-1}$)	1118875 to 4368143	a
	AMD	Atmospheric moisture diffusivity ($m^2\ s^{-1}$)	50719 to 2852835	a
	APM	Atlantic-Pacific moisture flux scaling	0.1 to 2.0	a
	OL0	Clear skies OLR reduction ($W\ m^{-2}$)	2.6 to 10.0	a
	OL1	OLR feedback ($W\ m^{-2}\ K^{-1}$)	-0.5 to 0.5	b
GOLDSTEIN	OHD	Isopycnal diffusivity ($m^2\ s^{-1}$)	312 to 5644	a
	OVD	Reference diapycnal diffusivity ($m^2\ s^{-1}$)	0.00002 to 0.0002	a
	OP1	Power law for diapycnal diffusivity depth profile	0.008 to 1.5	a
	ODC	Ocean inverse drag coefficient (days)	0.5 to 5.0	a
	WSF	Wind scale factor	1.0 to 3.0	a
	FFX	Freshwater flux scaling factor	1.0 to 2.0	c
SEA-ICE	SID	Sea ice diffusivity ($m^2\ s^{-1}$)	5671 to 99032	a
ENTS	VFC	Fractional vegetation dependence on vegetation carbon density ($m^2\ kgC^{-1}$)	0.4 to 1.0	a
	VBP	Base rate of photosynthesis ($kgC\ m^{-2}\ yr^{-1}$)	3.0 to 5.5	a
	VRA	Vegetation respiration activation energy ($J\ mol^{-1}$)	24211 to 71926	a
	LLR	Leaf litter rate (yr^{-1})	0.08 to 0.3	a
	SRT	Soil respiration activation temperature (K)	198 to 241	a
	VPC	Photosynthesis half-saturation to CO_2 (ppmv)	30 to 697	b
BIOGEM	PHS	PO_4 half-saturation concentration ($mol\ kg^{-1}$)	5.3e-8 to 9.9e-7	a
	PRP	Initial proportion of POC export as recalcitrant fraction	0.01 to 0.1	a
	PRD	e-folding remineralisation depth of non-recalcitrant POC (m)	106 to 995	a
	RRS	Rain ratio scalar	0.02 to 0.1	a
	TCP	Thermodynamic calcification rate power	0.2 to 2.0	a
	PRC	Initial proportion of $CaCO_3$ export as recalcitrant fraction	0.1 to 1.0	a
	CRD	e-folding remineralisation depth of non-recalcitrant $CaCO_3$ (m)	314 to 2962	a
	FES	Iron solubility	0.001 to 0.01	a
	ASG	Air-sea gas exchange parameter	0.1 to 0.5	a

ROKGEM	GWS	Land-to-ocean bicarbonate flux scaling factor	0.5 to 1.5	n/a

2.3 Experimental set-up of the model

The preindustrial ensemble simulation results were repeated to verify reproducibility of Holden et al. (2013a). The simulations were performed in two stages, each lasting 10 kyr, on the Cambridge High Performance Computing (HPC) Cluster Darwin. The first stage involved spinning up the model with atmospheric CO_2 concentration relaxed to 278 ppmv and a closed biogeochemistry system. This means that there are no sediment-ocean interactions and the model forces the $CaCO_3$ weathering and deep sea sediment burial rates into balance. An initial $CaCO_3$ weathering is initially prescribed but this is subsequently rescaled internally to balance the modelled $CaCO_3$ burial rate and conserve alkalinity. In the second stage, atmospheric CO_2 was allowed to evolve freely, with interacting oceans and sediments, and the $CaCO_3$ weathering rate is set equal to the $CaCO_3$ burial rate diagnosed from the end of stage 1. To allow the sediments to reach equilibrium as fast as possible, no bioturbation was modelled in either stage 1 or stage 2.

Each parameter set was then applied to LGM simulations. The modelled preindustrial equilibrium states were used as initial conditions and the ensemble members were integrated for 10 kyr, with freely evolving CO_2. These 10 kyr simulations are variously referred to here as the "LGM equilibrium simulation" or "stage 3", and the LGM equilibrium state refers to the end of stage 3 (see S1 for more details).~~.~~ After application to stages 2 and 3, the original 471 ensemble members were filtered to 315 ensemble members to exclude those simulations with a stage 2 atmospheric CO_2 concentration outside of the range 268 to 288 ppmv (c.f. Prentice et al., 2001), those that~~which~~ entered a snowball Earth state in stage 3 (global annual SAT between ~-68 and -57 °C), or those that ~~or which~~ showed evidence of numerical instability (c.f. Holden et al., 2013b).

Boundary conditions applied in the LGM simulations included orbital parameters (Berger, 1978) and aeolian dust deposition fields (Mahowald et al., 2006). The atmospheric CO_2 used in the radiative code is internally generated, rather than prescribed but the radiative forcing from dust, and gases other than CO_2 was neglected. The model also requires a detrital flux field to the sediments, containing contributions from opal and material from non-aeolian sources (Ridgwell and Hargreaves, 2007). Weathering fluxes from the preindustrial simulation were applied, scaled by GWS (the land-to-ocean bicarbonate flux scaling factor).

The representation of the ice sheets is as described in Holden et al. (2010b), using the terrestrial ice sheet fraction and orography from the ICE-4G reconstruction of Peltier (1994). Rather than initialising the ensemble with the ice sheet extent and orography at 21 kyr BP, the ice sheets are configured to grow from their preindustrial to LGM extent in 1 kyr, at the beginning of the LGM simulation (i.e. years 0-1 kyr) in order to account for the impact of sea level change on ocean tracers. Following Holden et al. (2010b), only the Laurentide and Eurasian Ice Sheets are allowed to change from their preindustrial form (accounting for ~ 80% of global ice sheet change), and we also route the freshwater to build the ice sheets from the Atlantic, Pacific and Arctic, assuming modern topography, rather than extracting it uniformly.

~~As~~ As the ice sheets grow, grid cells on land are gradually covered by ice and any carbon that remains, or has accumulated, is preserved underneath ("buried"). Once the ice covers a grid cell, there is no more exchange between the land carbon in that grid cell and the atmosphere. Prior to being buried, however, it is subject to the same forcings as carbon in any other grid cell.~~discussed in the introduction, the preindustrial terrestrial carbon is preserved underneath the LGM ice sheets, and allowed to interact with the atmosphere prior to its burial.~~ To determine how sensitive the burial carbon amount (i.e. the amount of carbon that is available for preservation underneath the ice) is to the duration of ice sheet build-up, we test the impact of varying the latter from 1~~000~~ to 10 kyr~~,000 years~~ for one ensemble member (extending the total simulation length to 11 kyr~~,000 years~~). Our assumption is that if the difference is negligle, applying the same ensemble member to a transient simulation of the full glacial cycle (and therefore a more realistic ice sheet build-up history) would not have yielded a dramatically different burial carbon inventory. We find that increasing the ice sheet build-up duration indeed changes the burial carbon amount only marginally: an increase of ~~(~~~34 PgC~~)~~. A limitation, however, is that we do not have a way of testing if ~~other~~ the response of other ensemble members ~~may be more sensitive. If one expects the sign of their response to be the same, more sensitive here means more carbon, not less, buried underneath the ice sheet~~would be equally subdued.~~.~~

2.4 Ensemble subsets

Although our ensemble varies many of the parameters thought to contribute to variability in glacial-interglacial atmospheric CO_2, not all sources of uncertainty can be captured. We estimate, based on our expert opinion, that up to ~60 ppmv of ΔCO_2 could be due to error in our process representations and processes not included in our model, such as changing marine bacterial metabolic rate, wind speed (via its effect on gas transfer) and Si fertilization. This is not a comprehensive assessment, however, as our model also does not include processes such as the effect of changing winds on ocean circulation (Toggweiler et al., 2006), Si leakage (Matsumoto et al., 2002, 2013, 2014), the effect of decreasing SSTs on $CaCO_3$ production (Iglesias-Rodriguez et al., 2002), or changing oceanic PO_4 inventory (Menviel et al., 2012). We focus our analyses on the subset of the ensemble with ΔCO_2 between ~-90 and -30 ppmv (Table 2), treating each ~~ensemble member in this~~value in this range as equally plausible. To test the robustness of diagnosed relationships, we also briefly compare the reponse of this subset with the response of the ensemble with no ΔCO_2 filter, and the response of the ensemble with a more negative ΔCO_2 filter. In ~~the latter case~~this second subset, the upper ΔCO_2 limit is set to ~-60 ppmv, roughly equivalent to allowing for an extra atmospheric CO_2 decrease due to changing marine bacterial metabolic rate, wind speed (via its effect on gas transfer) and Si fertilization, between the best and upper estimate of Kohfeld and Ridgwell (2009). The ΔCO_2 distribution in each subset or ensemble is shown in Fig. 1.

Table 2. Ensemble subsets, including ΔCO_2 and number of members in each.

Ensemble	ΔCO_2 range (ppmv)	Number of members
ENS_{315}	-88 to 74	315
ENS_{104}	-88 to -30	104
ENS_{16}	-88 to -59	16

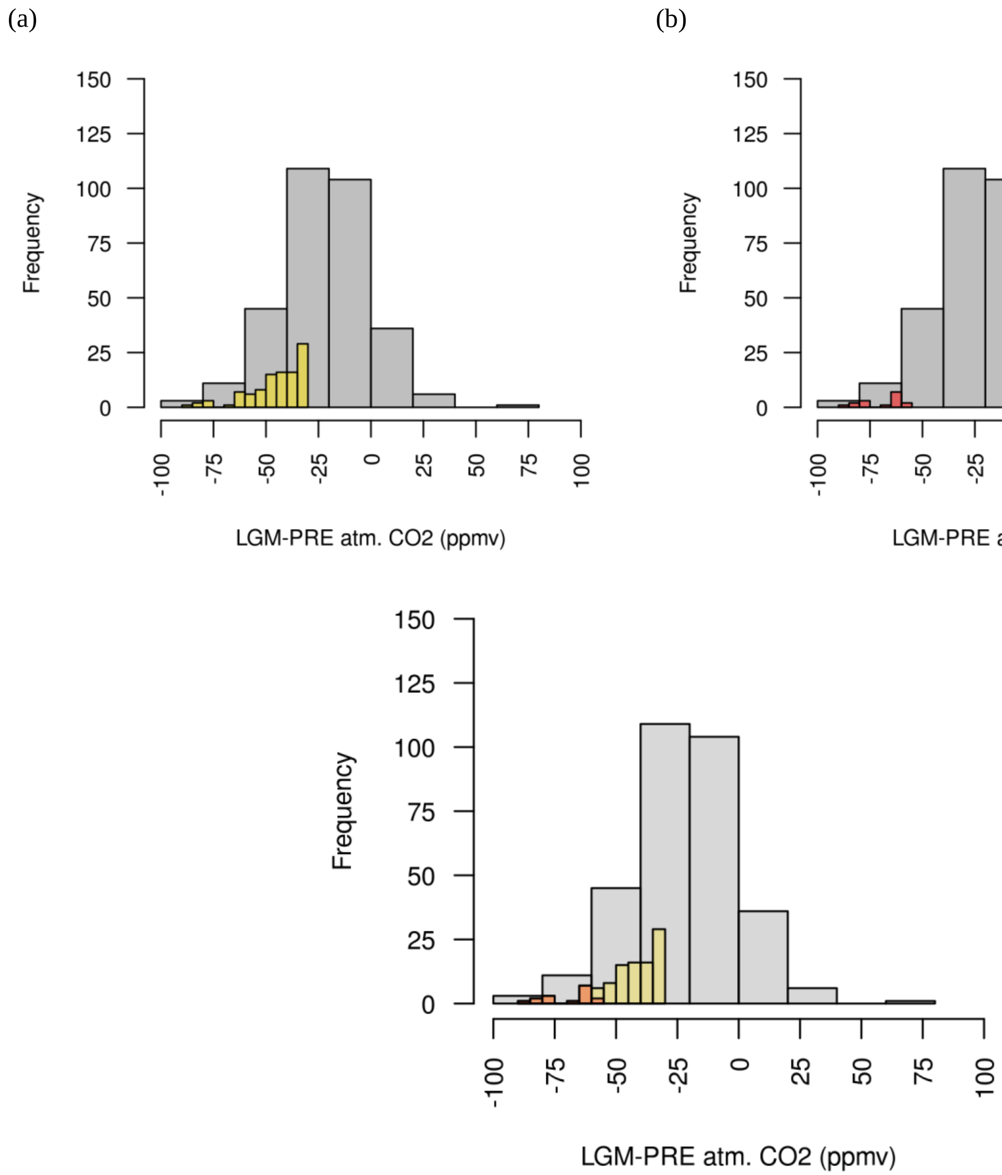


Fig. 1. LGM change in atmospheric CO_2 ~~(a-b)~~ distribution. The ENS_{315} response is shown in grey, the ENS_{104} ensemble response in yellow and the ENS_{16} ensemble response in orange~~purple~~. ~~Unless otherwise specified, the~~ The same colour legend applies to all figures in the manuscript.

3 Results and discussion

3.1 Preindustrial simulations

Comparison of the preindustrial response of ENS_{315} (i.e. the original, non-ΔCO_2 filtered ensemble) against the preindustrial ensemble response of Holden et al., 2013a confirms that the two are very similar. We additionally evaluate ENS_{315} against a few additional preindustrial metrics (see S2) and find responses that can be deemed ~~not uncontroversially im~~plausible, following the design principles for the ensemble, outlined in Holden et al., 2013a.

3.2 LGM simulations

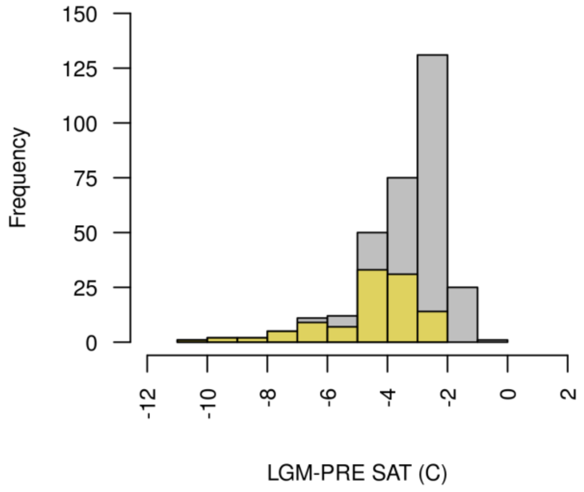
3.2.1 Climate, sea level and ocean circulation

Temperature

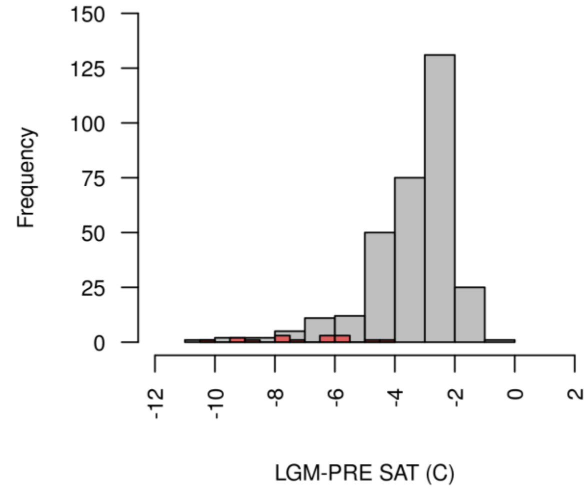
The ENS_{104} mean LGM surface air temperature (SAT) anomaly (ΔSAT) is -4.6 $\pm$ 1.7, and the range is -2.5 to -10.4 °C. The mean is close to the observed ΔSAT of -4 $\pm$ 1 °C (Annan and Hargreaves, 2013), and the range roughly equivalent to the range of previous model-based estimates (Kim et al., 2003; Masson-Delmotte et al., 2006; Schneider von Deimling et al., 2006; Braconnot et al., 2007; Holden et al., 2010a; Brady et al., 2013). The ENS_{104} mean LGM SST anomaly (ΔSST) is -1.8 ± 0.8 °C, and the range is -4.5 and -0.7 °C. The mean is again close to an observational data-constrained model estimate (Schmittner et al. (2011), and within the range of estimates inferred from proxy data (MARGO Project Members 2009 in Masson-Delmotte et al., 2013). There is a positive correlation between ΔSAT and ΔCO_2 (r = 0.75, 0.05 significance level hereforth), most likely reflecting the radiative impact of atmospheric CO_2 on SAT, as well as the effect of changing SAT on ΔCO_2. As suggested above, decreasing SST may contribute to decreasing CO_2 via the CO_2 solubility temperature dependence. Changing SAT may also affect ΔCO_2 via its effects on sea ice, ocean circulation, terrestrial and marine productivity (see below). The positive correlation is reproduced in ENS_{315} (r = 0.74), and as showin in Fig. 2, ΔSAT and ΔSST tend to be less negative in ENS_{315} than in ENS_{104} ~~(Fig. 2)~~. In ENS_{16}, ΔSAT and ΔSST are from the extreme or at least lower end of the ENS_{104} range.

The ENS_{104} mean ΔSAT and ΔSST spatial distributions are shown in Fig. 3. In line with observations (Annan and Hargreaves, 2013), the largest SAT decreases (> 10 °C) are simulated over the Laurentide and Eurasian Ice Sheets. The equator to pole temperature gradient is also broadly reproduced. The largest SST decreases (≥ 4 °C) are found in the North Atlantic and northeast Pacific, with more limited cooling (≤ -2 °C) in the tropics and polar regions, again consistent with observations. The largest SST decreases ought to, however, also be found in the southern hemisphere mid-latitudes whereas the simulated cooling is more moderate.

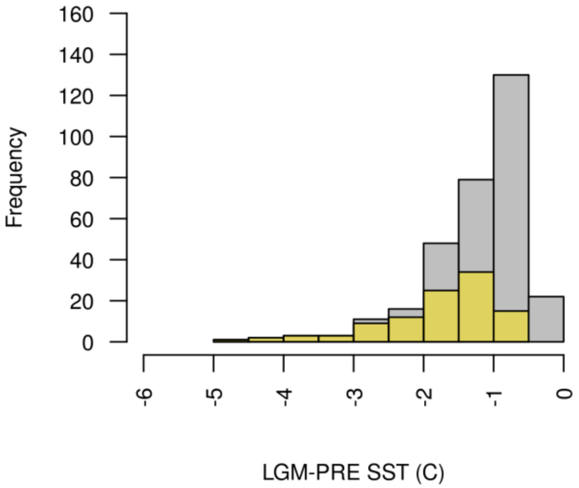
(a)


150
125
100
75
50
25
0
Frequency
-12
-10
-8
-6
-4
-2
0
2
LGM-PRE SAT (C)

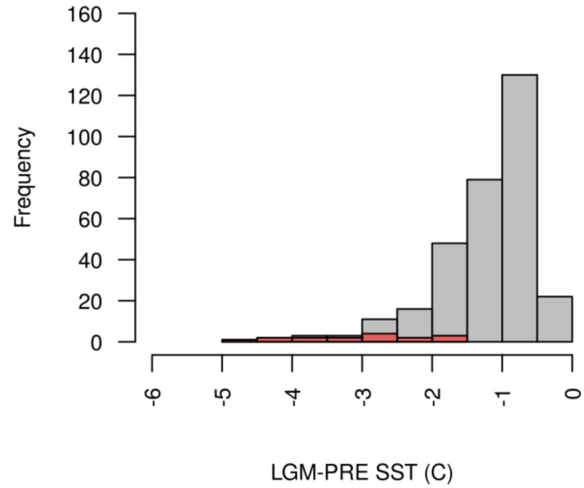
(b)


150
125
100
75
50
25
0
Frequency
-12
-10
-8
-6
-4
-2
0
2
LGM-PRE SAT (C)

(c)


160
140
120
100
80
60
40
20
0
Frequency
-6
-5
-4
-3
-2
-1
0
LGM-PRE SST (C)

(d)


160
140
120
100
80
60
40
20
0
Frequency
-6
-5
-4
-3
-2
-1
0
LGM-PRE SST (C)

(a)

(b)

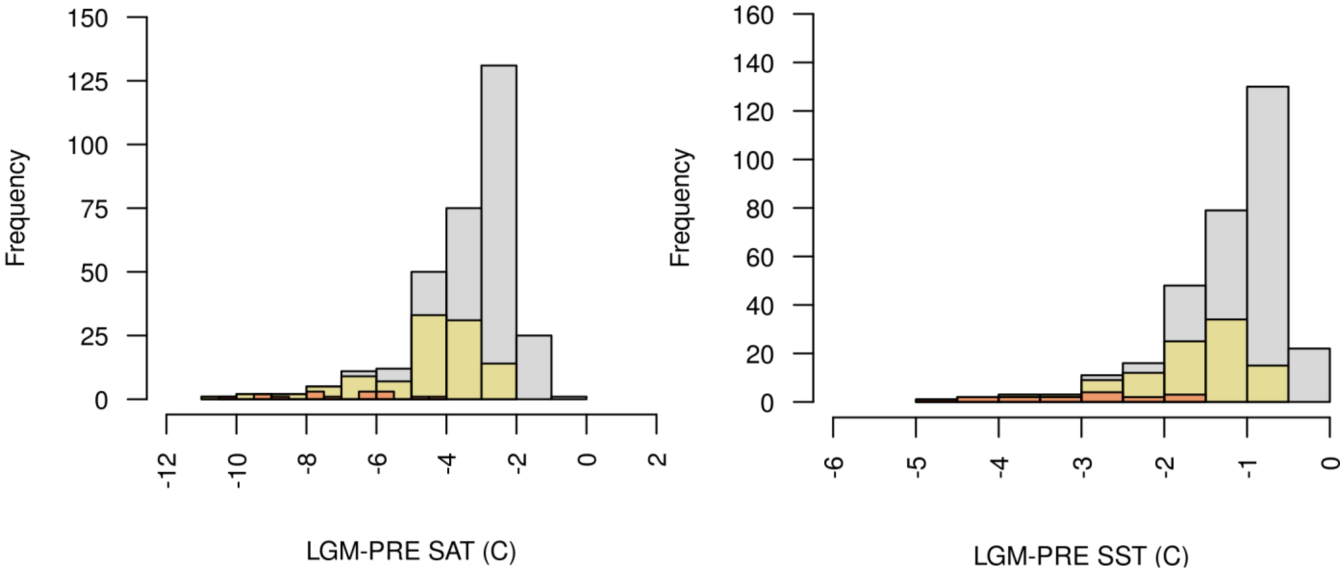
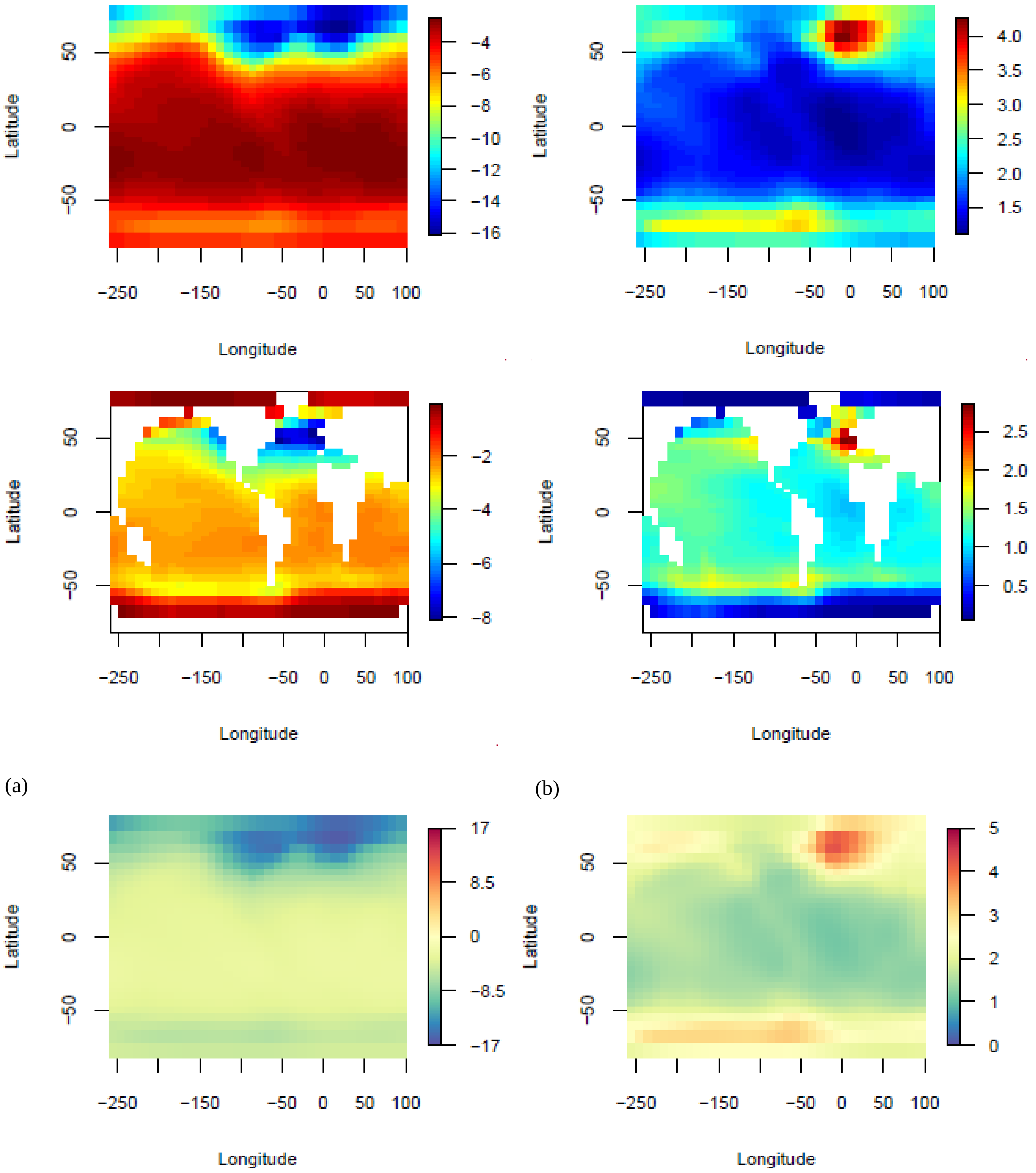


Fig. 2. LGM change in surface air temperature ~~(a-b)~~ and sea surface temperature (a~~c~~-b~~d~~) distributions.


Latitude
Longitude
50
0
−50
−250
−150
−50
0
50
100
−4
−6
−8
−10
−12
−14
−16
4.0
3.5
3.0
2.5
2.0
1.5
−2
−4
−6
−8
2.5
2.0
1.5
1.0
0.5
17
8.5
0
−8.5
−17
5
4
3
2
1
0

(a) (b)

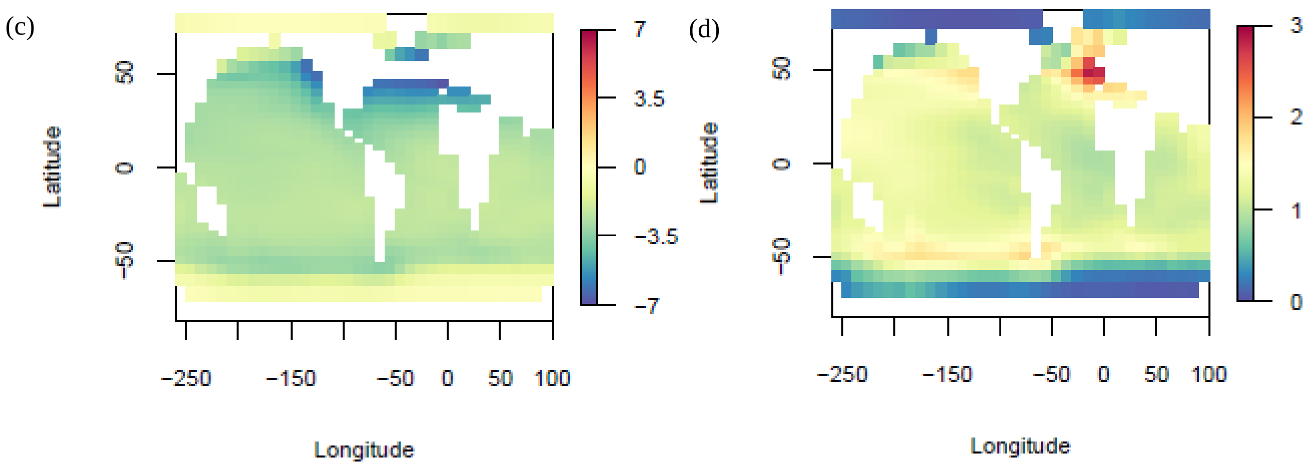


Fig. 3. LGM change in surface air temperature (a-b) and sea surface temperature (c-d) (°C) ENS_{104} mean (left) and standard deviation (right).

S~~ea level~~alinity

The ENS_{104} mean percentage increase in LGM salinity ~~(%S)~~ (and DIC, ALK, PO_4, etc) due to decreasing sea level is 2.84 ± 0.62 %S and the range is 2 to 4 %S. There is no significant relationship between %S and ΔCO_2 in either ENS_{104} or ENS_{315}, and the distribution of %S is similar in ENS_{104} ~~than~~ and in both ENS_{315} and ENS_{16} (Fig. 4). ~~There is, however, a weak positive correlation between %S and ΔCO_2 in ENS_{315} (r = 0.17).~~

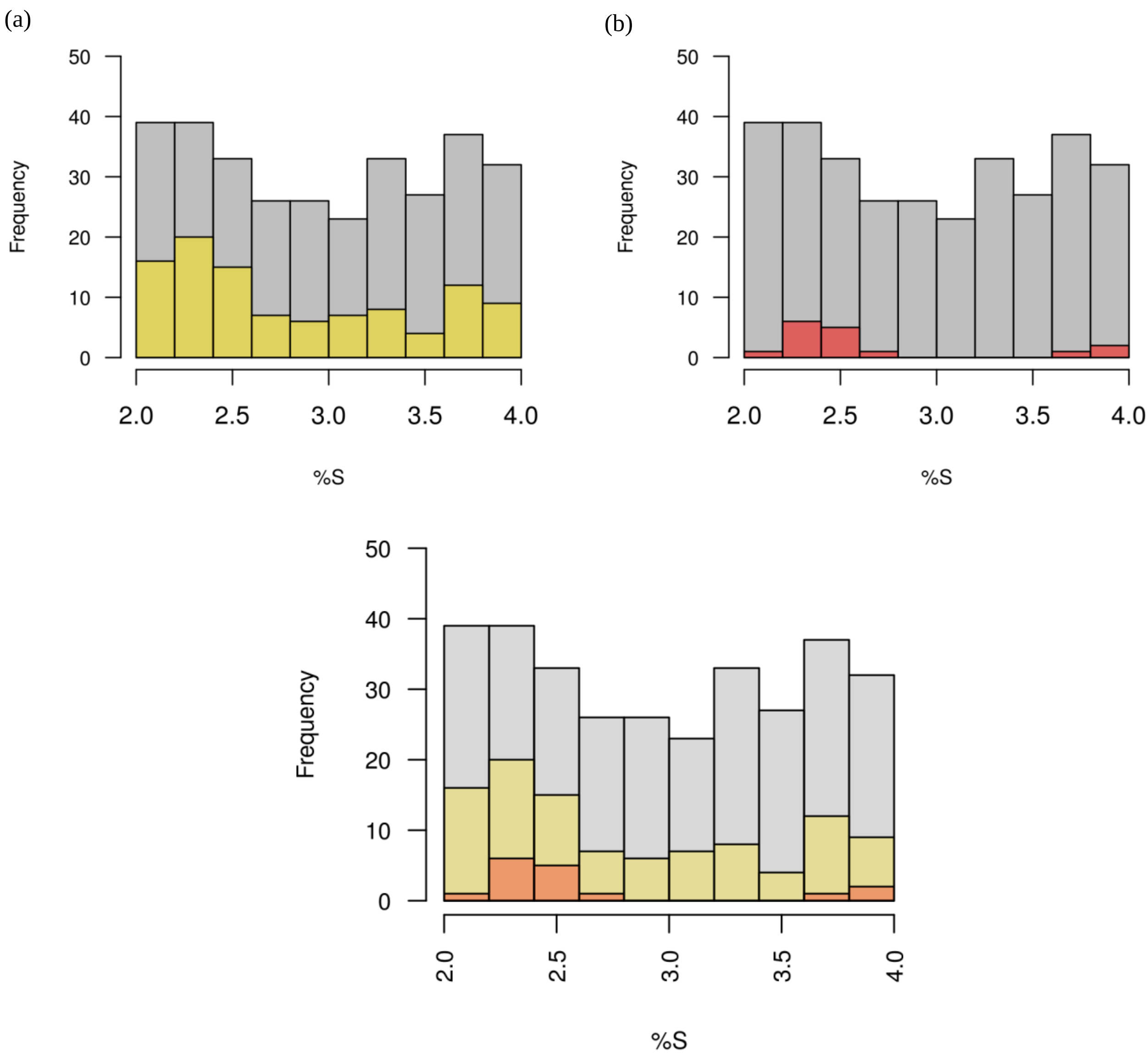


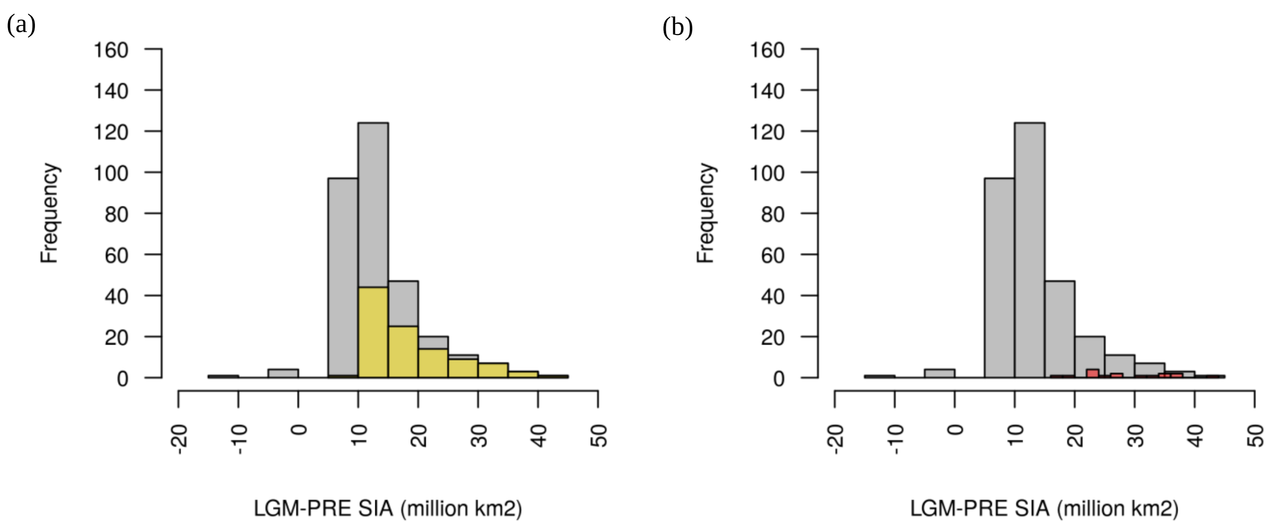
Fig. 4. Percentage increase in LGM salinity due to decreasing sea~~l~~ level ~~(a-b)~~ distribution.

Sea ice

The ENS_{104} mean LGM global annual sea ice area anomaly (ΔSIA) is 18.6 $\pm$ 7.4 million km^2, and the range is 9.9 to 44 million km^2. There is a negative correlation between ΔSIA and ΔSAT (r = -0.97) and between ΔSIA and ΔCO_2 (r = -0.74).~~.~~

The negative correlation between ΔSIA and ΔCO_2 likely reflects the impact of changing atmospheric CO_2 on ΔSIA but may also include a smaller contribution from changing sea ice area to ΔCO_2. Increasing LGM sea ice area ~~may have for instance~~could for instance have capped~~reduced~~ the outgassing of CO_2 from the ocean, particularly in the Southern Ocean ~~via sea ice capping,~~, and also reduced the net ocean-atmosphere CO_2 flux by decreasing the AMOC strength (see below). The negative correlation between ΔSIA and ΔSAT, and between ΔSIA and ΔCO_2, is reproduced in ENS_{315} (r = -0.96 and r = -0.74 respectively). ΔSIA in ENS_{104} also tends to be higher than in ENS_{315}, and smaller than in ENS_{16} (Fig. 5).

As shown in Fig. 6, fractional sea ice cover increases in all regions where sea ice is present in preindustrial simulations, although the largest increases take place in the North Atlantic.



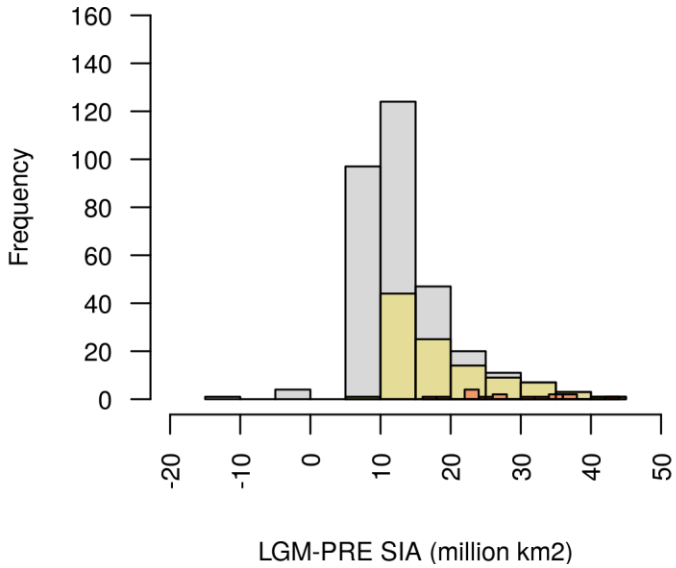
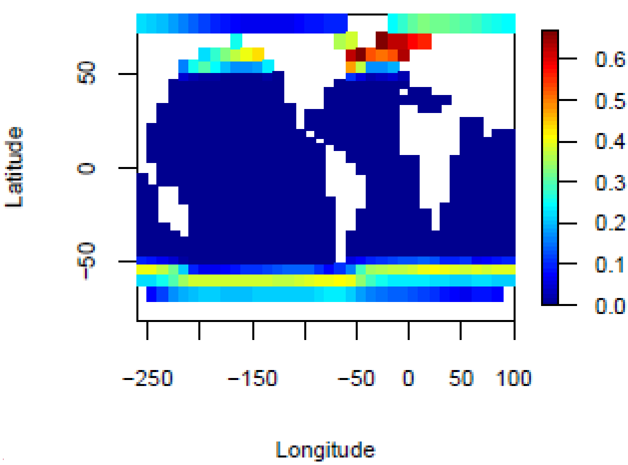


Fig. 5. LGM change in global sea ice area ~~(a-b)~~ distribution.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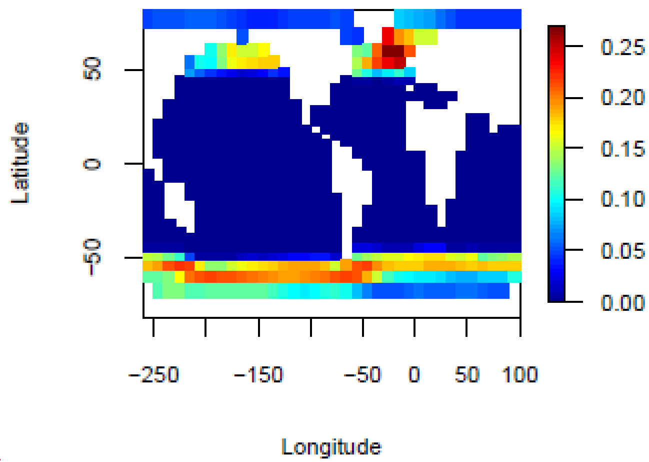
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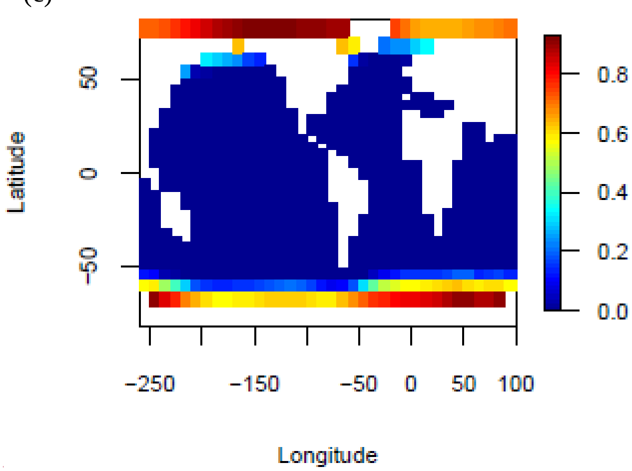
(c) (d)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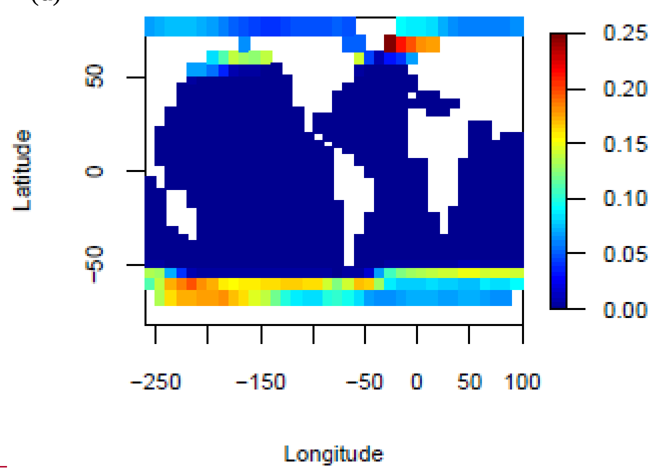
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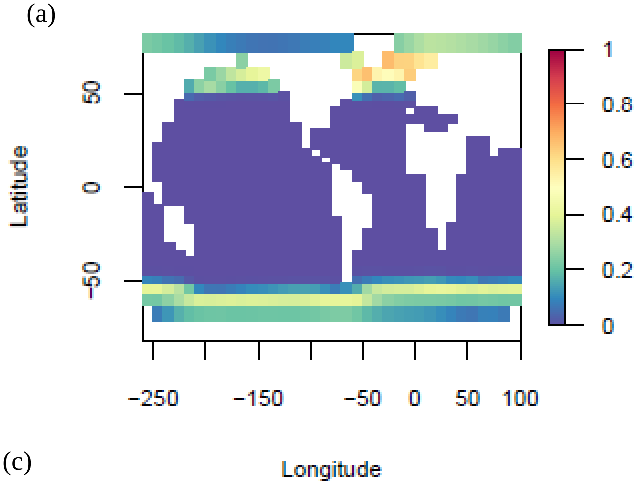
(a) (b)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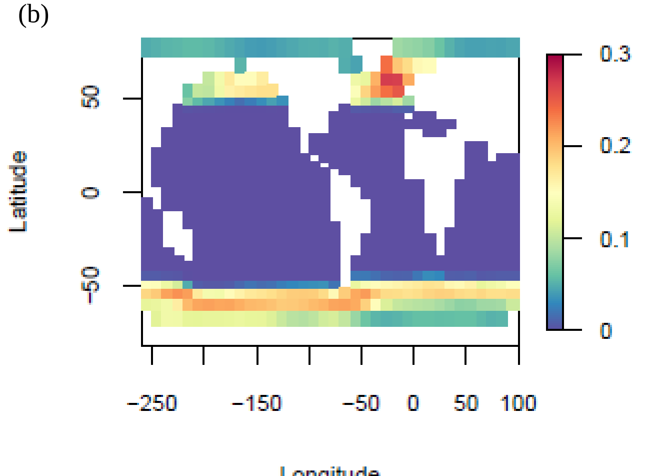
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(c) (d)

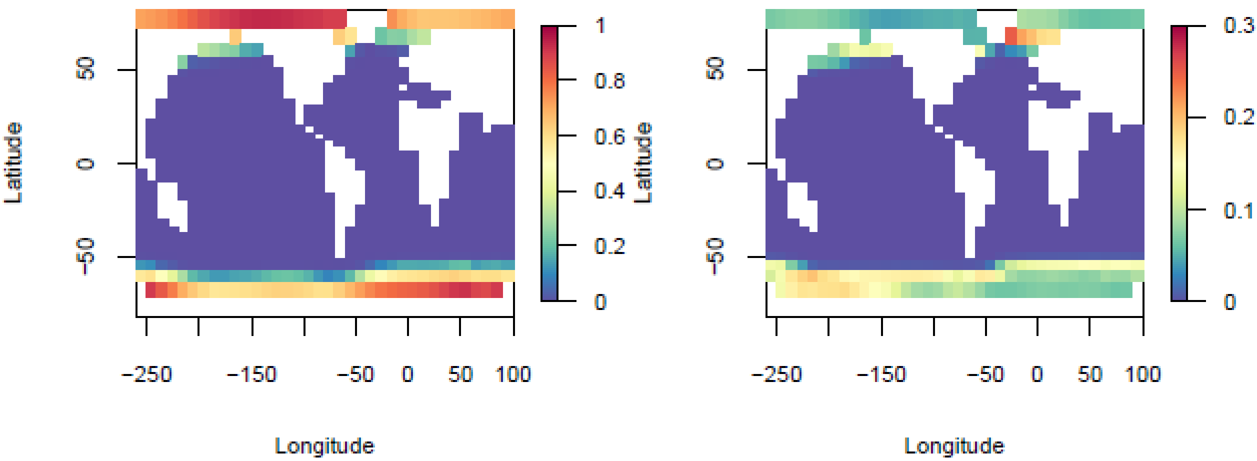


Fig. 6. LGM-PRE (a-b) and PRE (c-d) fractional sea ice cover ENS_{104} means (left) and standard deviations (right).

Precipitation

The ENS_{104} mean spatial distribution of the LGM precipitation rate anomaly (ΔPP) is shown in Fig. 7. The LGM changes are mostly negative but regions of positive ΔPP do exist, notably over Siberia and Australia. The largest LGM precipitation decreases (> 1.5 mm day $^{-1}$, with a maximum of 2.25 mm day $^{-1}$) are found over northern North America, and from around the eastern North Atlantic to northwest Asia, coinciding with the location of the Laurentide and Eurasian Ice Sheets (and the largest increases in fractional sea ice cover) respectively. Relatively large precipitation decreases (> 0.75 mm day $^{-1}$) are also simulated ~~across China~~in eastern Asia, equatorial Africa and other regions of enhanced fractional sea ice cover. Comparison against a pollen-based precipitation reconstruction (Bartlein et al., 2011;~~1 in~~ Alder and Hostetler, 2015) suggest that the simulated precipitation changes over Europe and equatorial Africa are of the right direction, while precipitation changes over western Siberia at least ought to be negative. The sign of the precipitation changes over North America is mostly consistent with observations, which record negative changes over most of the continent. However, positive changes, which are also observed, are not captured. Although not shown here, comparison of the ENS_{104} mean against the ENS_{315}mean suggests that the precipitation patterns in the two are very similar, but the decreases generally tend to be higher in the ENS_{104}mean. The precipitation decreases in ENS_{104}~~the latter~~ conversely tend to be smaller than in ENS_{16}.

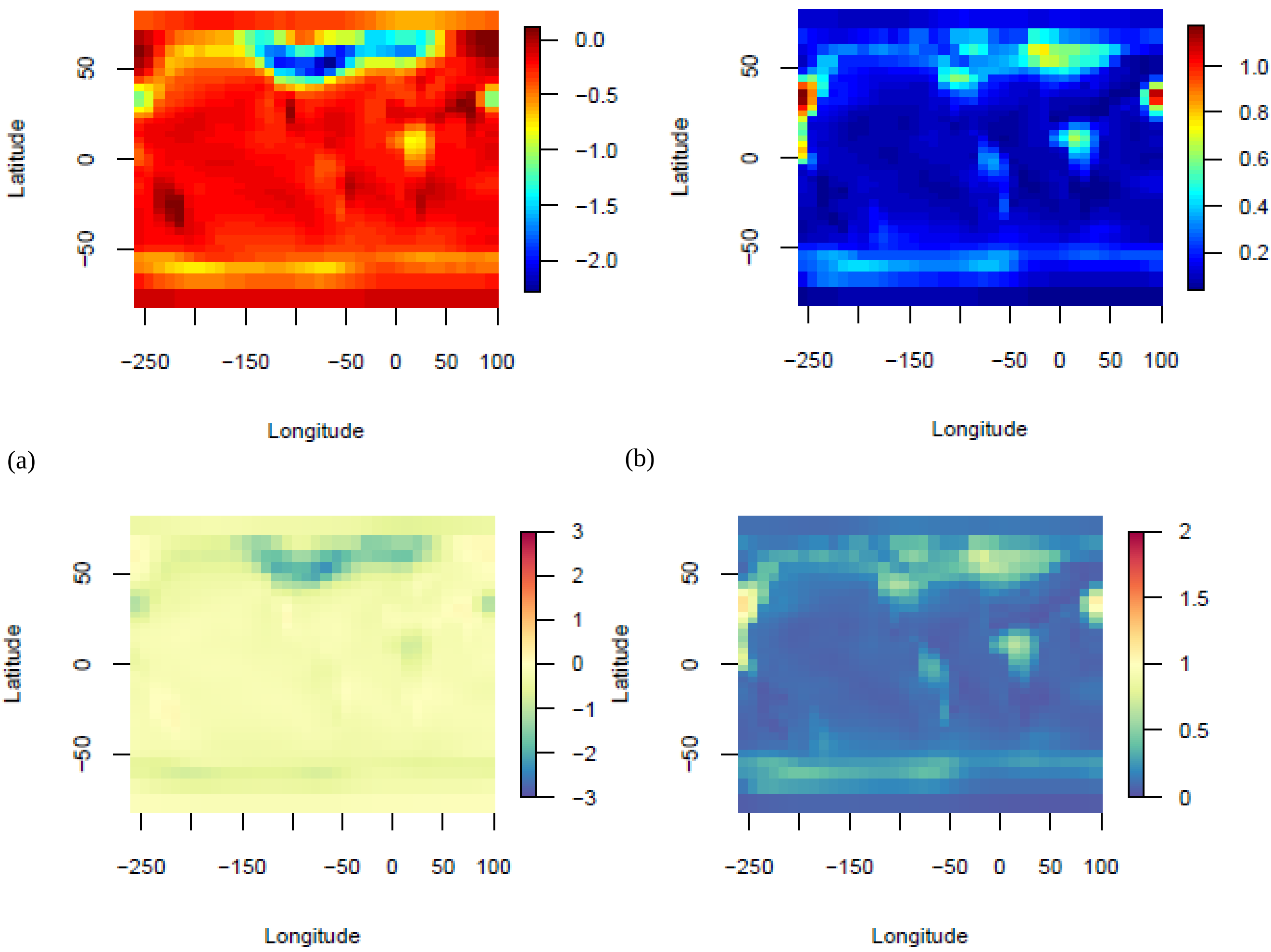


Fig. 7. LGM change in precipitation rate (mm day^{-1}) ENS$_{104}$ mean (a) and standard deviation (b).

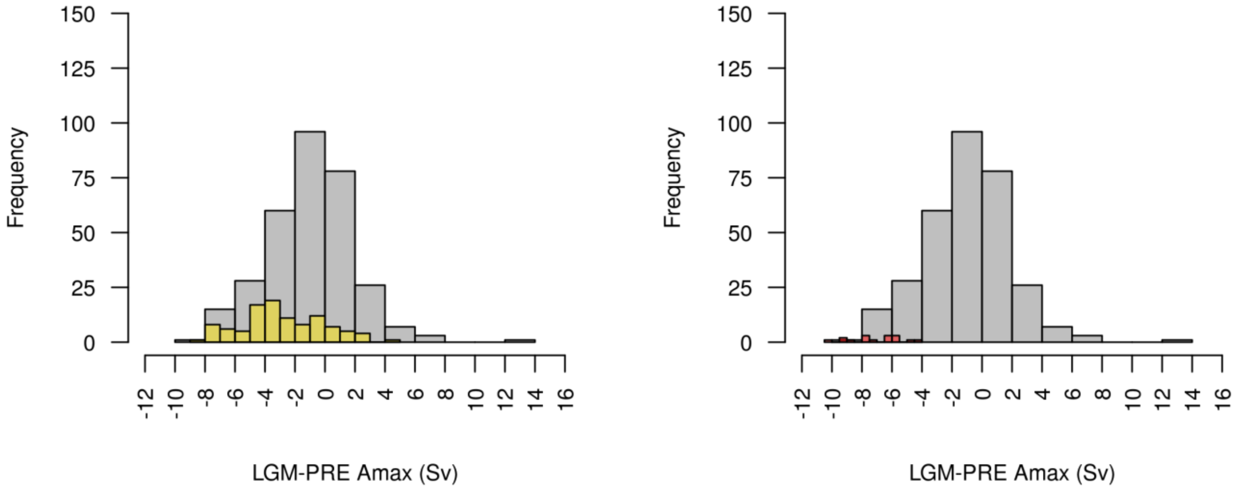
Ocean circulation

The ENS$_{104}$ mean LGM AMOC strength anomaly ($\Delta\psi_{max}$) is -2.8 ± 2.8 Sv, and the range is -8 to 4.7 Sv (Fig. 8). These estimates ~~are~~ lie at the low end of the $\Delta\psi_{max}$ predicted by 9 PMIP2 (Weber et al., 2007) and 8 PMIP3 (Muglia and Schmittner, 2015) coupled model simulations. However, they do not include the more negative $\Delta\psi_{max}$ predicted by Völker and Köhler (2013) for instance. The ENS$_{104}$ mean LGM-PRE AABW cell strength in the Atlantic Ocean ($\Delta\psi_{min}$) is 0.1 ± 1.2 Sv, which here represents an LGM decrease in cell strength as we keep the original (negative) sign for anticlockwise flow of Antarctic water. A negative $\Delta\psi_{min}$ conversely represents an LGM increase in cell strength. The range of $\Delta\psi_{min}$ is -

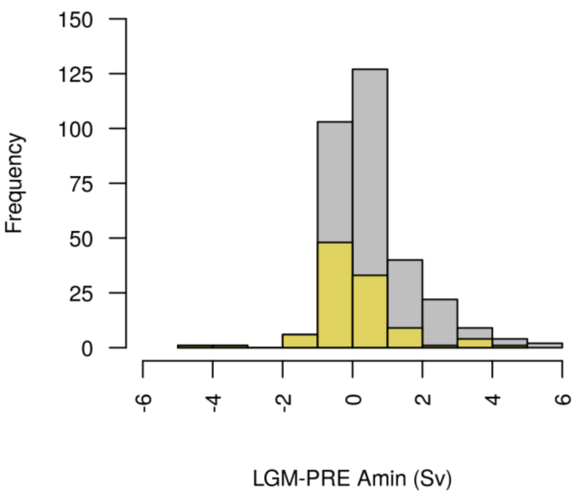
4.3 to 4.3 Sv, roughly comparable to the range of $\Delta\psi_{min}$ predicted in Weber et al. (2007) (see also Muglia and Schmittner, 2015), but excluding the much larger ψ_{min} increase predicted by Kim et al. (2003) for example.~~.~~ As shown in Fig. 9, t~~T~~he northern limits of the ENS_{104} mean LGM AMOC and AABW cells are roughly at the same latitudes as in the preindustrial simulations.~~ (Fig. 9)~~. The maximum depth reached by the ensemble mean AMOC base is also similar to preindustrial. Observations (Lynch-Stieglitz et al., 2007; Lippold et al., 2012; Gebbie, 2014; Böhm et al., 2015), conversely, suggest that the LGM AMOC shoaled to < 2 km, raising its base depth by 2500 and 600 m at the north and south end of the return flow respectively. The LGM AABW, in turn, is thought to have filled the deep Atlantic below 2 km, reaching as far north as 65 °N, which is ca. 25 degrees north of its modern northern limit (Oppo et al., 2015).

Although not shown here, the ENS_{16} ensemble members tend to exhibit a shoaling of the AMOC and enhanced penetration of AABW. With regard to $\Delta\psi_{max}$ and $\Delta\psi_{min}$, one can see from Fig. 8 that these tend to be more negative (i.e. weaker AMOC and stronger AABW) than in ENS_{104}. The $\Delta\psi_{max}$ and $\Delta\psi_{min}$ in ENS_{315} tend to conversely be more positive. In ENS_{104}, we also find a positive relationship between $\Delta\psi_{max}$ and ΔCO_2 (r = 0.57) and a negative relationship between $\Delta\psi_{min}$ and ΔCO_2 (r = -0.42). The relationships are reproduced in ENS_{315} (r = 0.59 and r = -0.36 respectively). We additionally find, in both ENS_{104} and ENS_{315}, negative correlations between $\Delta\psi_{max}$ and $\Delta\psi_{min}$ (r = -0.62 and -0.63), $\Delta\psi_{min}$ and ΔSAT (r = -0.4 and -0.4) and $\Delta\psi_{max}$ and ΔSIA (r = -0.62 and -0.66), as well as positive correlations between $\Delta\psi_{max}$ and ΔSAT (r = 0.68 and 0.66), and $\Delta\psi_{min}$ and ΔSIA (r = 0.37 and 0.42). Based on these relationships, we hypothesise that increasing LGM AABW strength led to an expansion of the AABW cell.~~, with~~ ~~the~~ The latter in turn restricted~~ing~~ the AMOC to lower depths, and reduced~~ing~~ its overturning rate (e.g. Shin et al., 2003). The increase in AABW strength was likely driven by increases in sea ice enhancing brine rejection. Sea ice increases in the North Atlantic may have additionally weakened the AMOC cell by locally reducing deep convection.

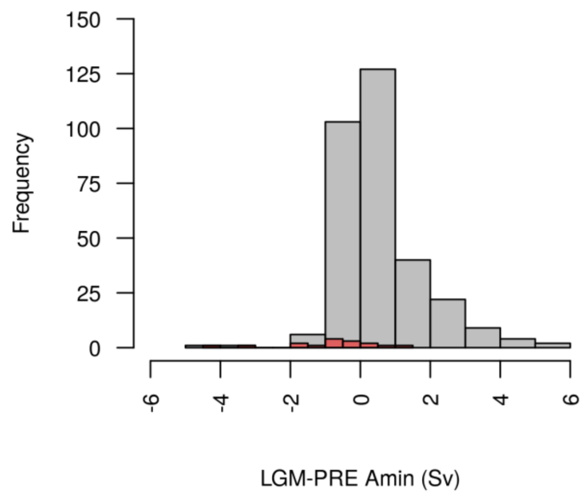
The relationships between $\Delta\psi_{max}$, $\Delta\psi_{min}$ and ΔCO_2 are also consistent with increasing ψ_{min} (decreasing ψ_{max}) contributing to decreasing atmospheric CO_2. The replacement of NADW by AABW in the North Atlantic would, for instance, have ~~for instance~~ led to a dissolution of deep sea sediment $CaCO_3$ due to AABW having a lower bottom water CO_3^{2-}concentration than NADW (see e.g. Yu et al., 2014). The increased $CaCO_3$ dissolution flux would in turn have raised~~s~~ the whole ocean alkalinity, lowering the atmospheric CO_2. Enhanced AABW production would ~~have also~~also have caused the deep ocean to become more stratified, allowing more DIC to accumulate at depth, and promoting further $CaCO_3$ dissolution. A decrease in NADW formation ~~on its own may have moreover~~could have additionally lowered atmospheric CO_2 by reducing~~ the outgassing of~~ CO_2 outgassing at the ocean surface, and reducing the burial rate of deep sea $CaCO_3$ ~~through the~~due to the concomitant increase in deep sea DIC accumulation. Further investigation is, however, required to confirm these causal relationships.


Frequency
LGM-PRE Amax (Sv)
Frequency
LGM-PRE Amax (Sv)

(c)


Frequency
LGM-PRE Amin (Sv)

(d)


Frequency
LGM-PRE Amin (Sv)

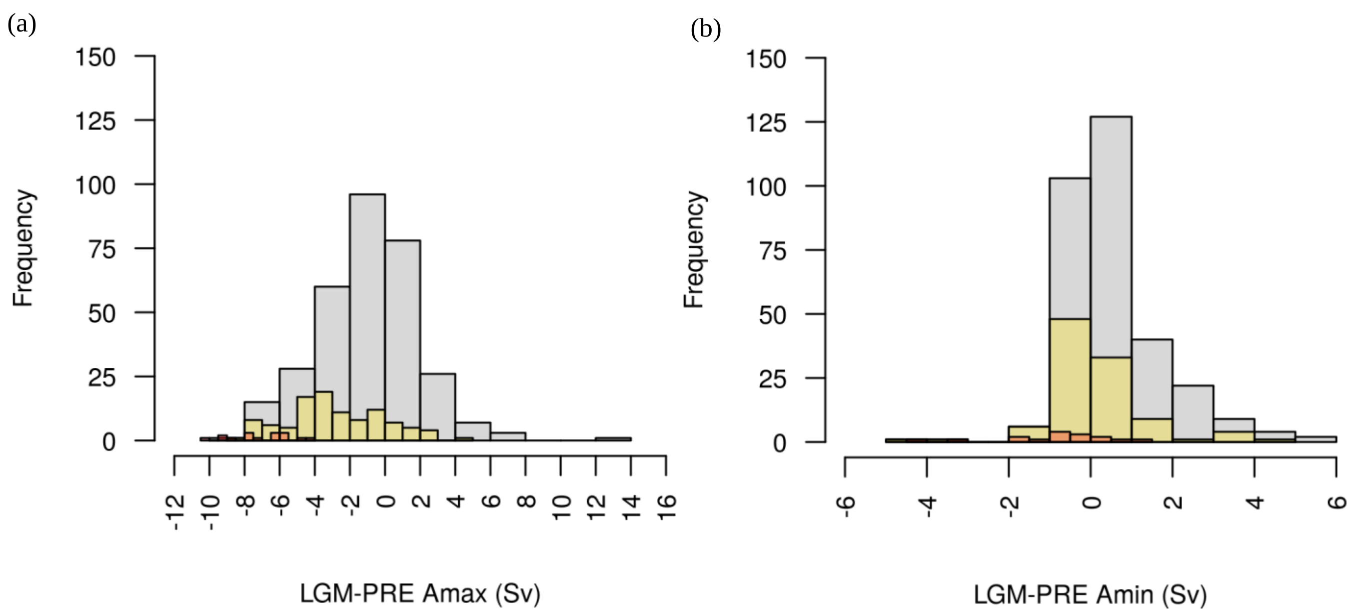
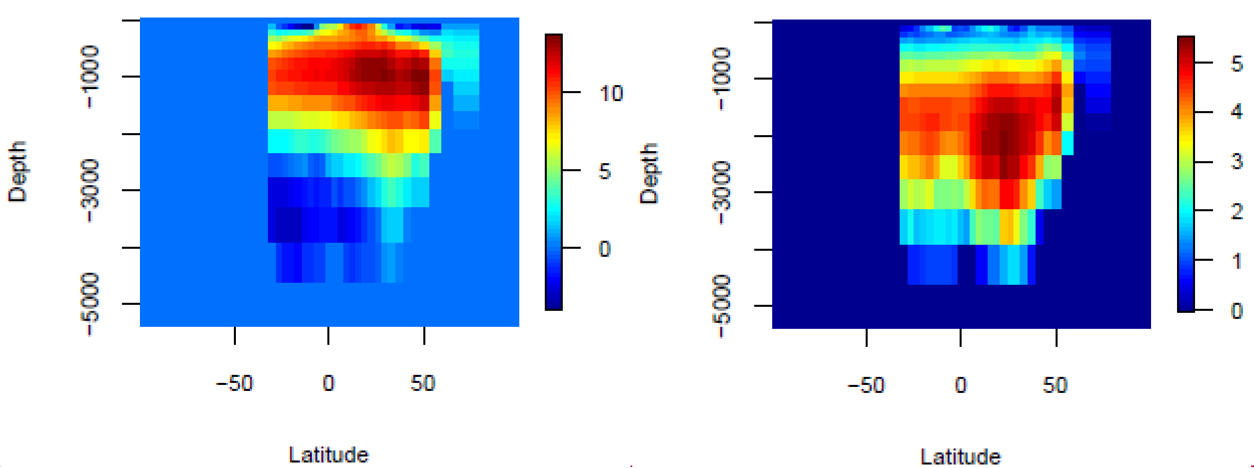


Fig. 8. LGM change in ψ_{max} ~~(a-b)~~ and ψ_{min} (a~~c~~-b~~d~~) distributions.



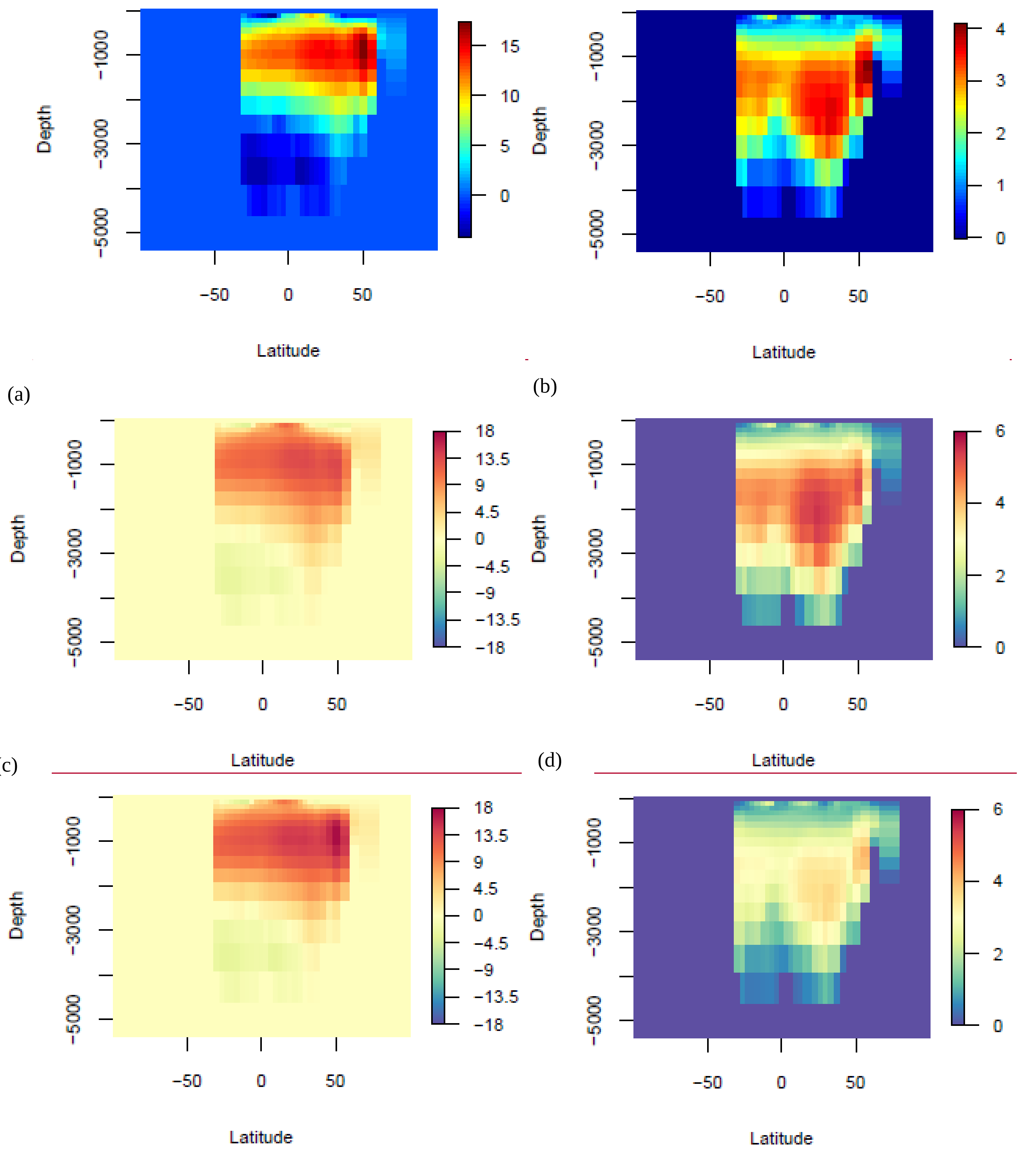
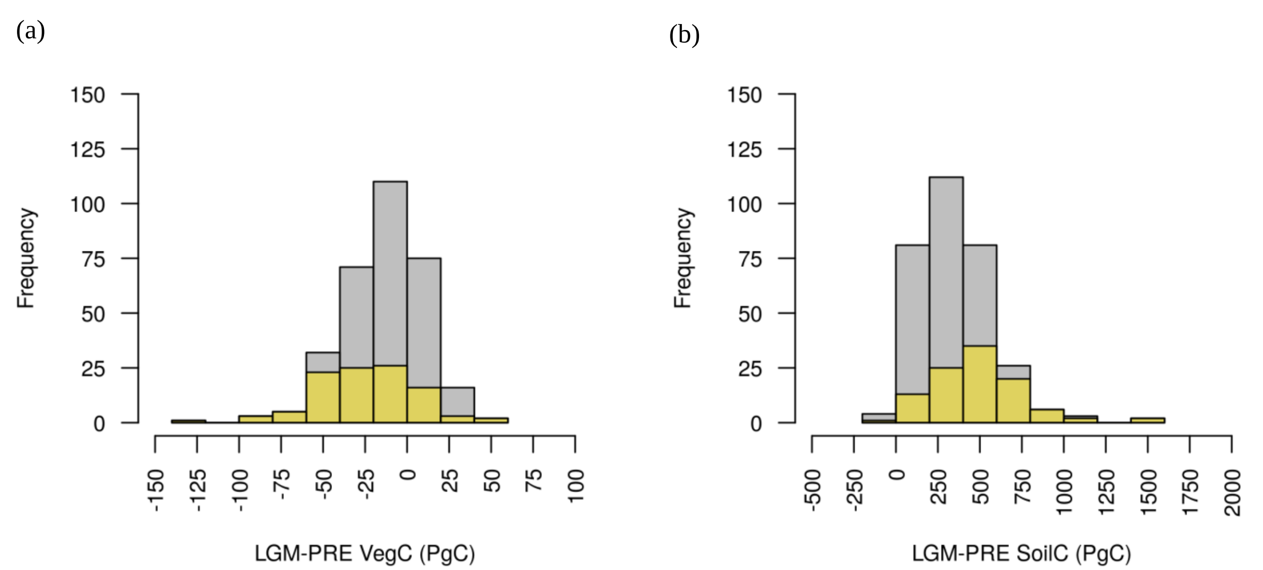


Fig. 9. LGM (a-b) and PRE (c-d) Atlantic overturning stream functions (Sv) ENS_{104} means (left) and standard deviations (right).

3.2.2 Terrestrial biosphere, ocean and lithospheric carbon

As shown in Fig. 10, m~~M~~ost of the ensemble members in ENS_{104}~~the PGACF ensemble~~ predict an LGM increase in terrestrial biosphere (ΔTerrC), and lithospheric[1] (ΔLithC) carbon inventory and a decrease in ocean carbon inventory (ΔOceanC). ~~(Fig. 10).~~ The remaining ensemble members predict one of four other scenarios of carbon partitioning, with the second most common scenario (11% of ensemble members) being increasing terrestrial carbon and decreasing ocean and lithospheric carbon (Table 3). Similar patterns can also be observed in ENS_{315} and ENS_{16}. ~~(Table 3).~~ A likely explanation for scenario 1 (increase in terrestrial biosphere and lithospheric carbon, decrease in ocean carbon) is that the growth of biosphere land carbon ~~on land~~ (see below) causes a flux of CO_2 from the atmosphere to the land, leading to an immediate outgassing of CO_2 from the ocean to remove the atmospheric pCO_2 difference. The CO_2 outgassing also leads to an increase in surface $[CO_3^{2-}]$ and subsequently deep ocean $[CO_3^{2-}]$, which reduces $CaCO_3$ dissolution (and increases lithospheric carbon). The increase in $CaCO_3$ burial in turn decreases $[CO_3^{2-}]$ and increases $[CO_2]$, which is communicated back to the surface, with a resultant increase in atmospheric CO_2 (Kohfeld and Ridgwell, 2009). The above explanation is of course only part of the explanation for this dominant carbon partitioning scenario, with physical mechanisms also expected to play a role, in addition to any changes in ocean productivity and changes in land carbonate weathering (see below).



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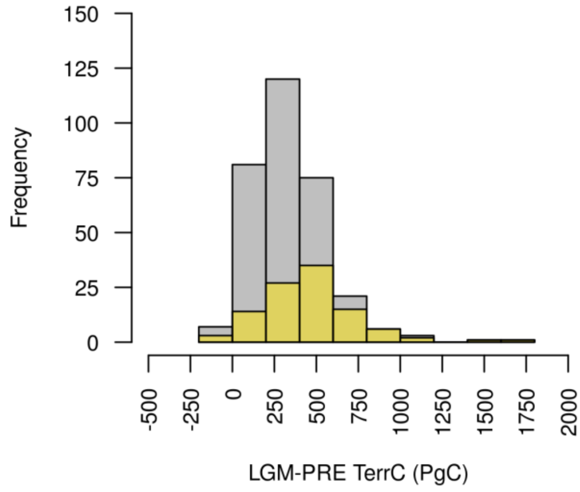
[1] The ΔLithC stems from changes in the deep-sea $CaCO_3$ burial flux and/or $CaCO_3$ weathering/shallow water deposition flux and was initially calculated to ensure that carbon was being conserved over the LGM simulation.

(c)



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75
50
25
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Frequency
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750
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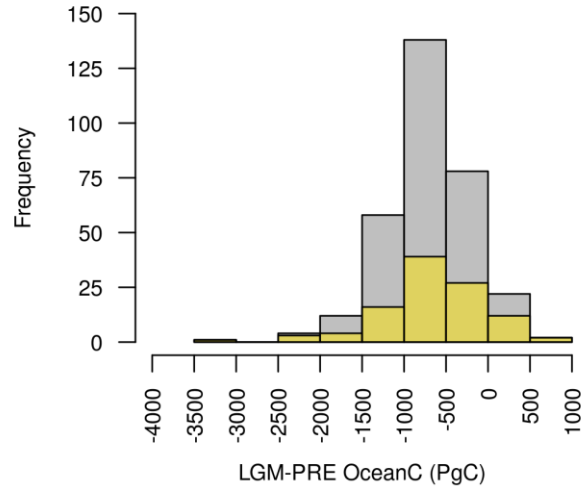
1750
2000
LGM-PRE TerrC (PgC)

(d)



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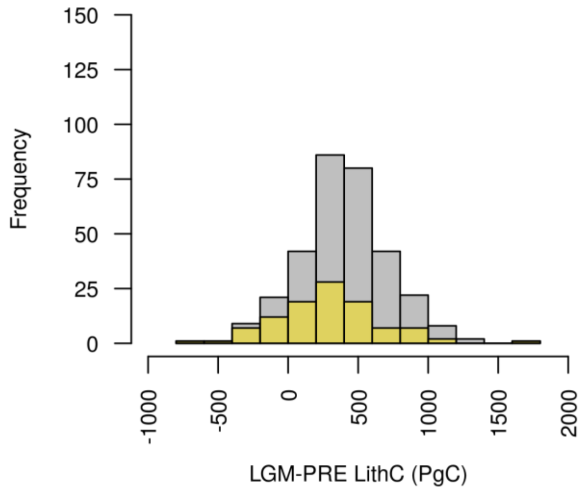
1000
LGM-PRE OceanC (PgC)

(e)



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Frequency
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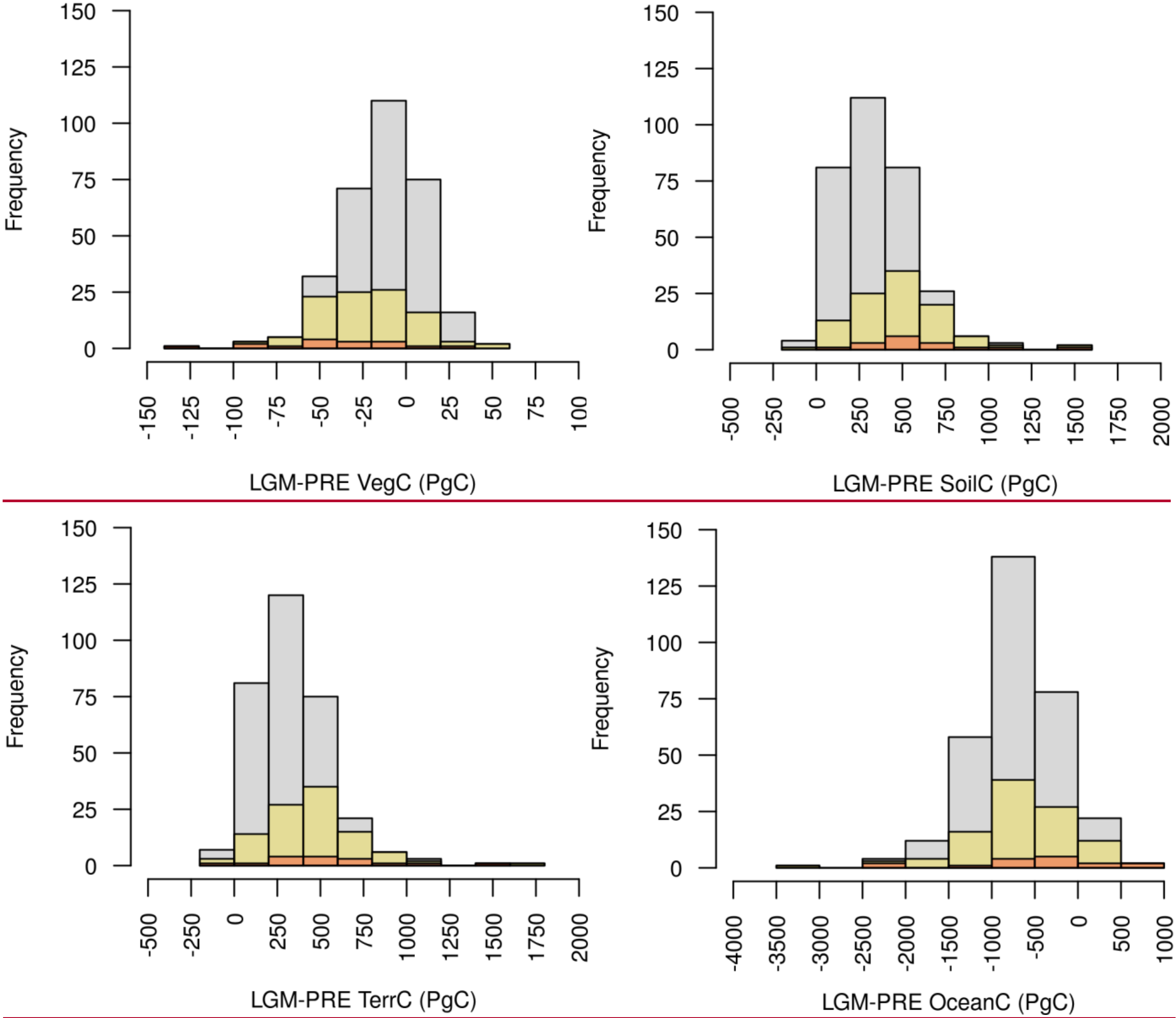
LGM-PRE LithC (PgC)


Frequency
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Frequency
LGM-PRE SoilC (PgC)
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LGM-PRE TerrC (PgC)
Frequency
LGM-PRE OceanC (PgC)

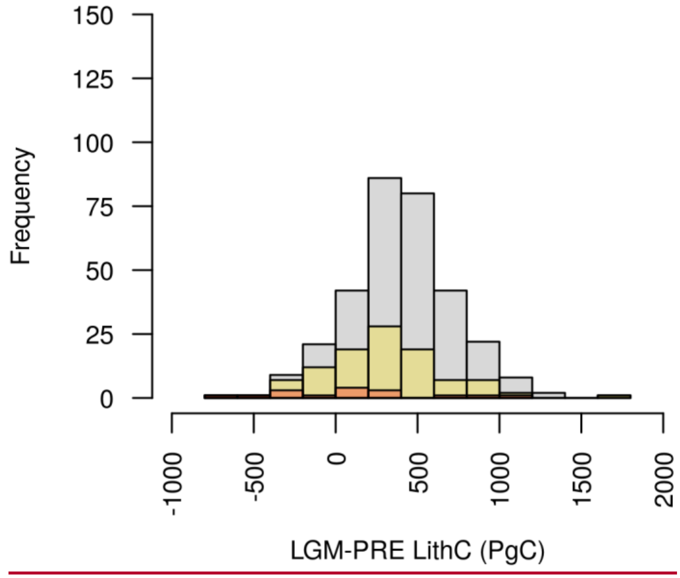


Fig. 10. LGM change in vegetation (a), soil (b), terrestrial (vegetation + soil) (c), ocean (d) and lithospheric (e) carbon inventory distributions.

Table 3. LGM-PRE carbon partitioning scenarios in ENS_{315}, ENS_{104} and ENS_{16}.

			ENS_{315}		ENS_{104}		ENS_{16}	
Scenarios			**Total counts**	**(% of total)**	Total counts	(% of total)	Total counts	(% of total)
1. ΔTerrC(+)	ΔOceanC(-)	ΔLithC(+)	279	(89)	82	(79)	10	(63)
2. ΔTerrC(+)	ΔOceanC(+)	ΔLithC(-)	16	(5)	11	(11)	3	(19)
3. ΔTerrC(+)	ΔOceanC(-)	ΔLithC(-)	11	(3)	8	(8)	2	(13)
4. ΔTerrC(-)	ΔOceanC(+)	ΔLithC(+)	1	(1)	1	(1)	0	(0)
5. ΔTerrC(-)	ΔOceanC(+)	ΔLithC(-)	5	(2)	2	(1)	1	(6)
6. ΔTerrC(-)	ΔOceanC(-)	ΔLithC(+)	1	(1)	0	(0)	0	(0)
7. ΔTerrC(+)	ΔOceanC(+)	ΔLithC(+)	2	(1)	0	(0)	0	(0)

The ENS_{104} mean ΔTerrC, ΔOceanC and ΔLithC, the signs of which are consistent with scenario 1, are reported in Table 4, alongside previous estimates from observational data- and model-based studies. From here we can see that the mean ΔTerrC is only aligned with a handful of estimates and no studies so far report a negative ΔOceanC. Instead, ΔOceanC is estimated to be positive, primarily based on carbon isotope data. The loss of hundreds of petagrams of carbon from the ocean in response to terrestrial carbon growth has, however, been previously proposed (e.g. Zimov et al., 2006). Moreover, if we assume that 90% of the atmospheric CO_2 perturbation caused by the increase in terrestrial biosphere carbon reported in Table 4 gets removed by the ocean and sediments, the change in ocean carbon would be negative, even after adding the remaining carbon to be lost from the atmosphere to the ocean. We discuss what these results would likely mean for carbon isotope data in section 3.2.6.

The positive $\Delta TerrC$ studies in Table 4 attribute the increase in terrestrial biosphere carbon to different factors: Zimov et al., 2009 and Zech et al., 2011 predict large increases in permafrost carbon while Zeng et al., 2003 ignores permafrost. Instead, the author attributes~~ and instead attributes most of the glacial ~~the glacial terrestrial carbon increase to the preservation rather than destruction of carbon in areas to be covered by the LGM ice sheets, ~~ice sheet burial carbon. Another contributor is the~~ lower soil respiration rates (in the active carbon pool) caused by a colder climate, as well as storage of carbon on exposed continental shelves. Here, neither the latter~~is~~ carbon accumulation mechanism, nor that of ~~peat growth (absent also in Zeng, 2003),~~permafrost growth, are included~~ in our model~~. However, ~~As discussed above, permafrost growth is also not represented explicitly but there is an ~~our model does attempt ~~-~~to capture the very slow rates of soil decomposition characteristic of permafrost (Williamson et al., 2006). We attribute the terrestrial carbon increase in our ensemble to a higher soil carbon inventory caused by a decreasing soil respiration rate. As can be seen from Fig. 10, vegetation carbon tends to conversely decrease at the LGM. As in Zeng (2003), we also do not assume that as the ice sheets expand over terrestrial carbon, they destroy it. The lack of this potential loss term means that our $\Delta TerrC$ estimates may be higher than in many previous studies, irrespective of what the response of the terrestrial biosphere is to the LGM climate and CO_2 forcings. O'ishi and Abe-Ouchi (2011) for instance estimated that the LGM climate is responsible for the loss of 502 PgC of terrestrial carbon, while another 388 PgC is removed by the ice sheets. Zeng (2003) conversely proposed that 431 PgC is preserved under the ice sheets at the LGM. This number includes 315 PgC present during the interglacial, and another 116 PgC accumulated in response to the glacial climate forcings, prior to insulation of the terrestrial carbon from the atmosphere by the ice sheet coverage.

~~Similarly to Zeng (2003), the model gradually buries carbon in LGM ice sheet areas. ~~Here, a~~A~~nalysis of ENS_{16} suggests that during the 1000 years of LGM ice sheet build-up, the terrestrial carbon inventory in the areas to be occupied by the ice sheets increases by between 6 and 444 PgC, yielding LGM "ice-sheet or burial" carbon inventories ~~(or "burial" carbon inventories) ~~between 318 and 1341 PgC (Table 5). This increase accounts for less than half of the total LGM change in terrestrial carbon (i.e. $\Delta TerrC$) in the majority of simulations. However, if ~~this "extra" carbon (accumulated in response to climate forcings), and the carbon already present in the ice sheet areas at the end of the preindustrial spin-up,~~this burial carbon were to have been destroyed rather than preserved, $\Delta TerrC$ would be negative in all but 3 simulations, as opposed to positive in all but one simulation (Table 5).

Table 4. $\Delta TerrC$, $\Delta OceanC$ and $\Delta LithC$ (PgC) in this study (ENS_{104} mean, standard deviation and range) and previous studies. PE14 = Peterson et al. (2014), OA13 = O'ishi and Abe-Ouchi (2013), C12 = Ciais et al. (2012), Z03 = Zeng (2003), SA13 = Sarnthein et al. (2013), BG15 = Brovkin and Ganopolski (2015), PR11= Prentice et al. (2011), SK15 = Skinner et al. (2015), SS16 = Schmittner and Somes (2016), GL13 = Goodwin and Lauderdale (2013), A15 = Allen et al. (2015), CR95 = Crowley et al. (1995), AF98 = Adams and Faure (1998), BR12 = Brovkin et al. (2012), ZI09 = Zimov et al. (2009), ZE11 = Zech et al. (2011).

	This study	Previous studies	Ref.	Details
$\Delta TerrC$	467.5±286.5 [-51.6,1603.8]	[-1160, 530]	CR95	Pollen database
		-1500	AF98	Ecological data
		[-694, -550]	PR11	Simulation with LPX
		-600	BR12	Simulation with CLIMBER-2
		-597	OA13	Simulation with MIROC-LPJ
		-511	PE14	Benthic foraminiferal $\delta^{13}C$ records
		-330	C12	Benthic foraminiferal, ice core and terrestrial $\delta^{13}C$ records + simulation with LPJ land ecosystem model
		0	BG15	Simulation with CLIMBER-2 (+ permafrost, peat, glacial burial carbon)
		547	Z03	Simulation with a coupled atmosphere-land-ocean-carbon model
		[200, 400]	ZE11	Soil carbon measurements
		<1000	ZI09	Soil carbon measurements
$\Delta OceanC$	-664±626.9 [-3187.7, 662.4]	[730, 980]	SA13	Ocean radiocarbon records
		687	SK15	Ocean radiocarbon records
		654	A15	Ocean $[CO_3^{2-}]$ reconstructions + benthic foraminiferal $\delta^{13}C$ records
		[570, 970]	GL13	Ocean $[CO_3^{2-}]$ reconstructions
		520	C12	Benthic foraminiferal, ice core and terrestrial $\delta^{13}C$ records + simulation with LPJ land ecosystem model
		[510,670]	SS16	Simulation with MOBI 1.5 coupled to Uvic
$\Delta LithC$	292.5±373.9 [-654.9,1700.9]	n/a	n/a	n/a

Most of the terrestrial carbon increase in areas to be covered by the ice sheets~~ice sheet carbon~~in ENS_{16} is due to soil carbon, with vegetation carbon decreasing in all but one simulation. The range of ~~of carbon inventory changes~~terrestrial carbon increases and the associated burial carbon amounts include~~s the 116 PgC increase in ice sheet carbon, and consequent total burial carbon inventory (431 PgC), estimated by~~ Zeng (2003)'s estimates. ~~It would also allow for an~~Our LGM burial carbon estimates would also accommodate an additional 250-550 PgC (Franzen, 1994) from increased glacial peat accumulation ~~to~~

~~be buried under the ice sheets, as suggested by (~~Zeng, ~~(~~2003). No observational data-based estimates of the LGM burial carbon inventory are available since ~~there is so far only~~only limited evidence exists for organic material being preserved by ice during glaciations (Franzen, 1994 and references in Weitemeyer and Buffett, and Zeng, 2007). Outside of the ice sheets, increases in the terrestrial carbon inventory in ENS_{16} are mostly due to soil carbon, which increases in all simulations. Vegetation carbon, conversely, decreases in the majority of simulations. ~~The~~ Our range of ~~non-ice sheet~~ carbon changes outside of the ice sheet areas include~~s~~ the 198 PgC increase ~~in non-burial non-shelf terrestrial carbon~~ predicted by Zeng (2003) as a result of reduced soil respiration. ~~Here, the increase in terrestrial biosphere carbon both inside and outside of the ice sheet areas, are presumed to reflect the decrease in soil respiration rate due to colder SATs exceeding the decrease in net photosynthesis (i.e. total photosynthesis-respiration) rate due to lower~~ $\cancel{CO_2}$~~, SAT and precipitation, as they are mainly driven by soil carbon increases. In most previous model- and observational data-based studies it is conversely suggested that the reduced photosynthesis effect due to climate and~~ $\cancel{CO_2}$ ~~changes outcompetes the reduced respiration effect on a global scale. In O'ishi and Abe-Ouchi (2013) for example, the terrestrial carbon that would be gained if ice sheet carbon was preserved rather than destroyed (383 PgC) is smaller than the amount of carbon that is lost in response to the LGM climate and~~ $\cancel{CO_2}$ ~~changes (502 PgC).~~

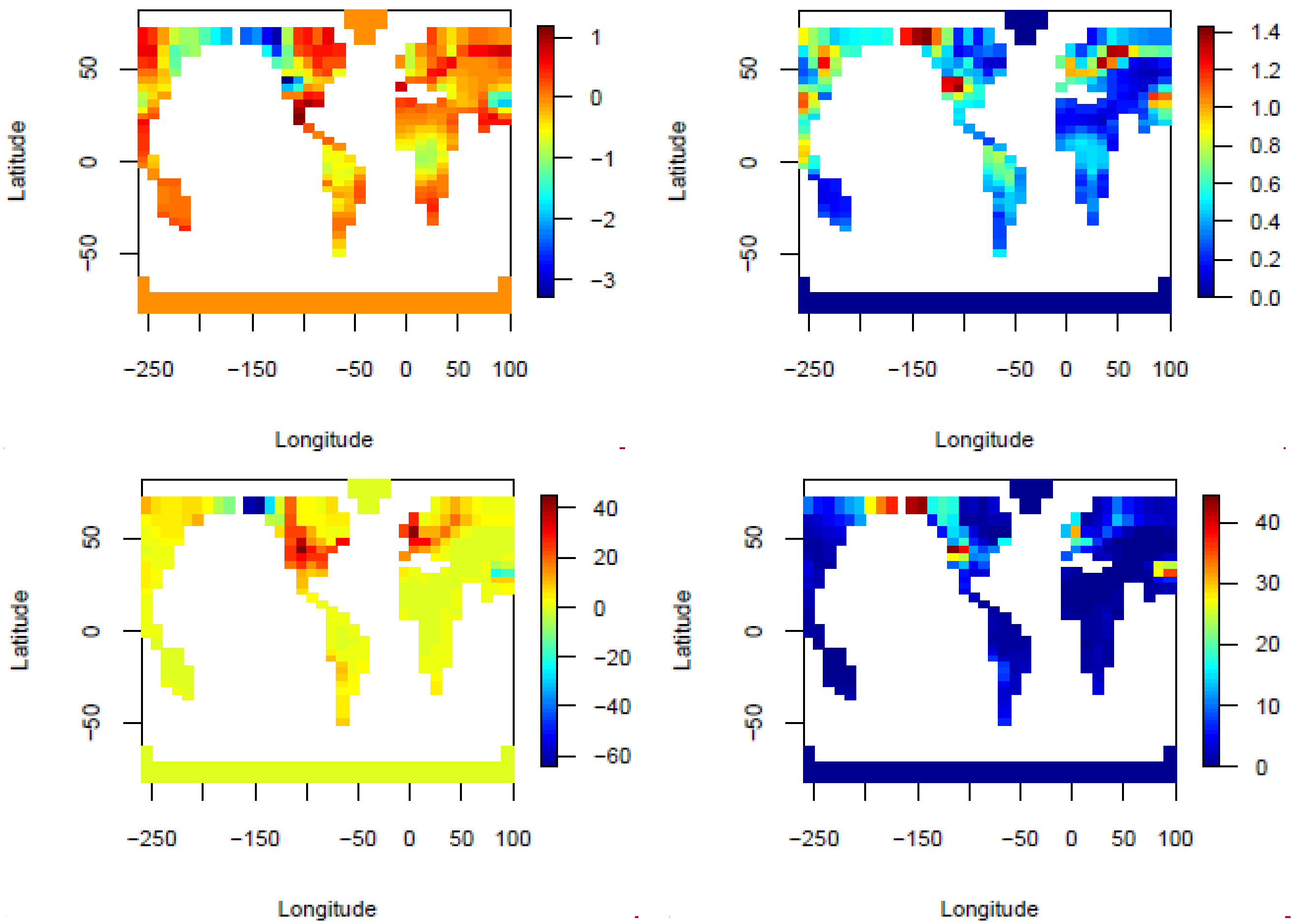
Although not evaluated directly, it is likely that similar ice sheet/non-ice sheet terrestrial carbon proportions than in ENS_{16} are found in ENS_{104} and ENS_{315} because of the similar climate change distributions in all three instances (see earlier sections). Although not shown here, the spatial distribution of $\Delta\mathrm{TerrC}$ in ENS_{16} is also similar to that of the ENS_{104} (and ENS_{315}) mean. The spatial distribution of $\Delta\mathrm{TerrC}$ in ENS_{104} is shown in Fig. 11 ~~and compared against observations~~.

The largest increases in terrestrial carbon ($\geq$ 20 kgC m^{-2}) are found in North America and Europe/western Asia, both within and south of the Laurentide and Eurasian ice sheet margins (Fig. 11). Regions with smaller but still relatively large ($\geq$ 10 kgC m^{-2}) increases include the Andes and Patagonia regions, the southern tip of the African continent, eastern north Siberia and the grid cells just south of the Tibetan plateau. The largest LGM decreases in terrestrial carbon ($\geq$ 10 kgC m^{-2}) conversely tend to be found in northwest North America, Beringia and the Tibetan plateau region. Other regions with relatively large ($\geq$ 5 kgC m^{-2}) decreases include equatorial Africa and the deserts in central Asia. Everywhere else the LGM terrestrial carbon density increases by between 0 and 10 kgC m^{-2}. Comparison against paleoecological reconstruction studies (Crowley et al., 1995) suggests that the simulated terrestrial carbon changes within the Laurentide and Eurasian ice sheet areas are of the wrong sign, except in northwest North America, since these studies assume the complete destruction of vegetation and soils in ice sheet areas. Discrepancies between the ENS_{104} and observations further arise from the rainforest regions, where the ensemble mean predicts terrestrial biosphere carbon density changes between -5 and 10 kgC m^{-2}, well above observed changes of $\sim$ -23 kgC m^{-2}. It is important to note, however, that as suggested in Zeng (2007), the rate of decomposition of soil carbon at the LGM may have been slower than assumed in pollen data-based studies. The largest increases in terrestrial carbon density (~ 40 kgC m^{-2}) produced by the ensemble mean are comparable to those found in areas with permafrost growth (Zimov et al., 2006). However, the peaks are potentially misplaced, being located within and south of the Laurentide and Eurasian ice sheet covered areas, rather than in eastern Siberia and Alaska. Alternatively, terrestrial carbon increases in eastern Siberia and Alaska are simply underestimated in the ensemble mean and large increases in terrestrial carbon indeed took place within the ice sheet areas during glacial periods.

The ~~highly negative~~large LGM decreases in ~~LGM~~ terrestrial carbon ~~changes~~ in northwest North America and adjacent Beringia are likely caused by precipitation decreasing comparatively more than SAT, and causing the decrease in photosynthesis to exceed the decrease in soil respiration. However, it is also noteworthy that, although not shown here, the regions with the largest decreases in terrestrial carbon density, namely northwest North America, Beringia and the Tibetan plateau area, are also the regions with the largest terrestrial carbon densities in the preindustrial ENS$_{315}$ mean~~simulations~~. We ~~further, note that in the preindustrial ENS$_{315}$ mean,~~ further note that the Tibetan soil carbon peak is overestimated in the latter, and the North American soil carbon peak misplaced, compared to observations. We attribute the first discrepancy to the lack of soil weathering in the model and the inclusion of land use effects in the observational data-based estimate (Holden et al., 2013b; Williamson et al., 2006). The second discrepancy is attributed to the lack of explicit representation of permafrost and the absence of moisture control on soil respiration.

Table 5. Ice sheet and non-ice sheet terrestrial carbon stocks in ENS$_{16}$**.** Columns 2 and 5 show the amount of carbon stored in ice sheet areas and under the ice sheets during the preindustrial and LGM periods respectively. Column 3 is the difference between the two inventories. Column 4 is the LGM change in carbon in ice sheet ~~carbon~~ areas expressed as percentage of the total LGM terrestrial carbon change. Column 6 is the LGM change in carbon outside of the ice sheets.

EM	PRE ice sheet	LGM-PRE ice sheet	% LGM-PRE Total land	LGM Burial	LGM-PRE non-ice sheet
442	456	117	51	573	111
873	896	444	29	1341	1089
511	677	262	32	939	567
99	372	33	20	405	130
871	404	149	26	553	425
786	502	131	21	633	486
107	540	86	22	626	310
701	549	161	36	710	283
801	707	275	39	982	423
219	312	6	16	318	-34
694	389	95	30	484	227
623	697	181	36	879	319
522	713	210	28	923	531
863	408	73	16	480	380
478	573	165	41	739	233
837	784	395	33	1179	796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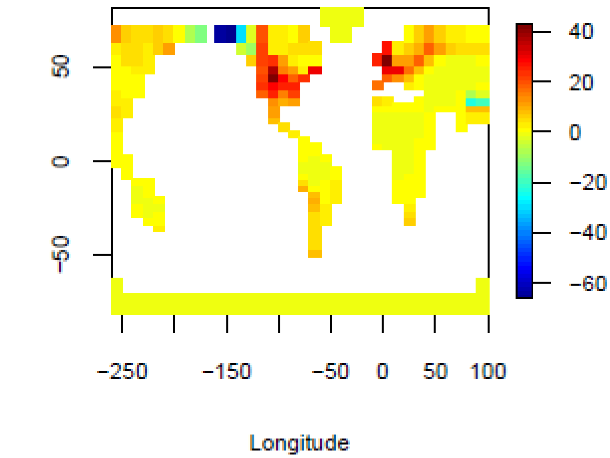
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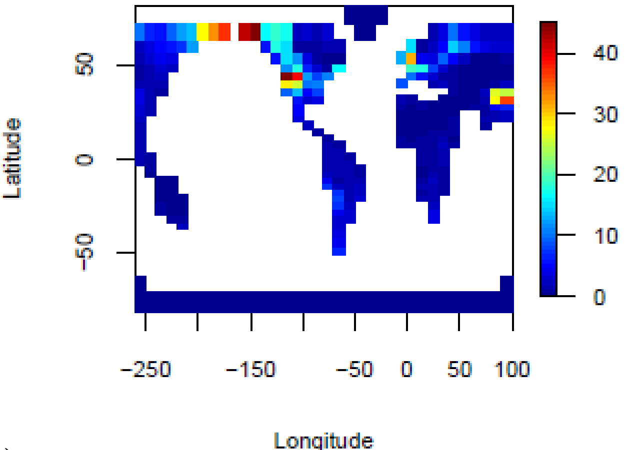
(a)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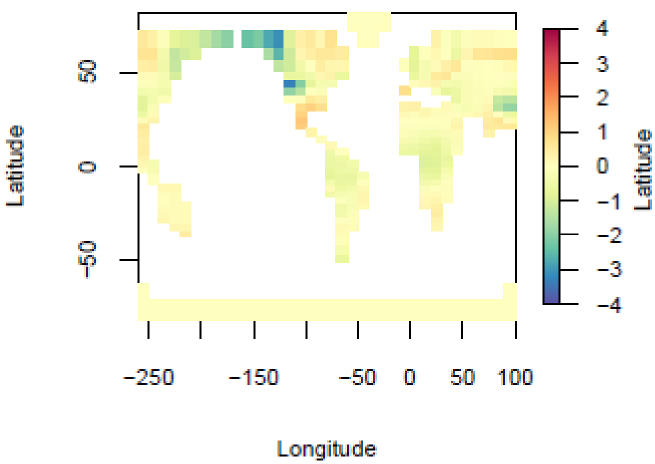
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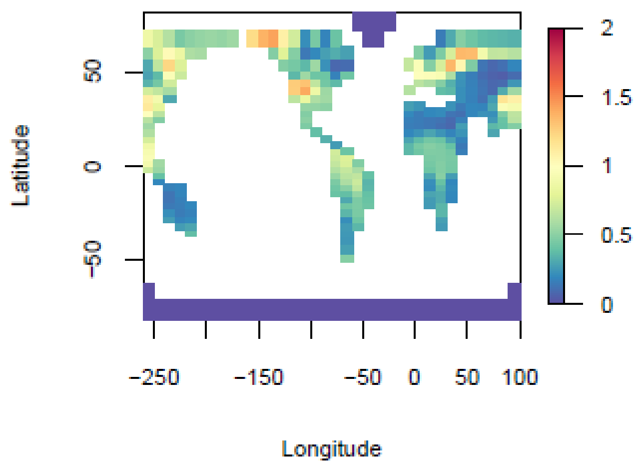
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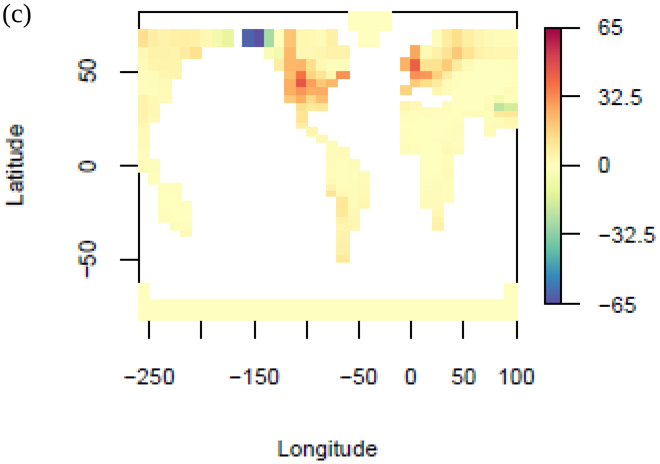
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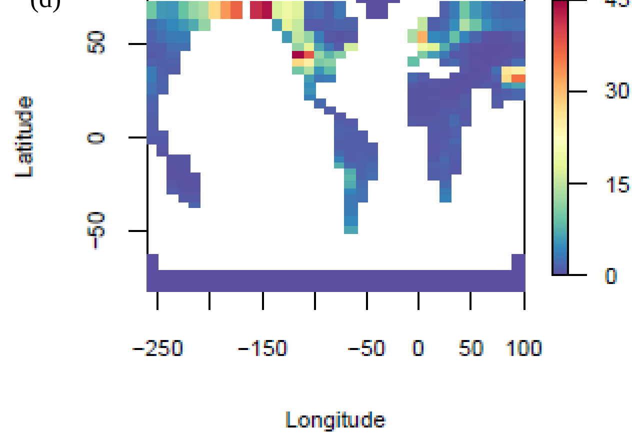
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(f)

(e)

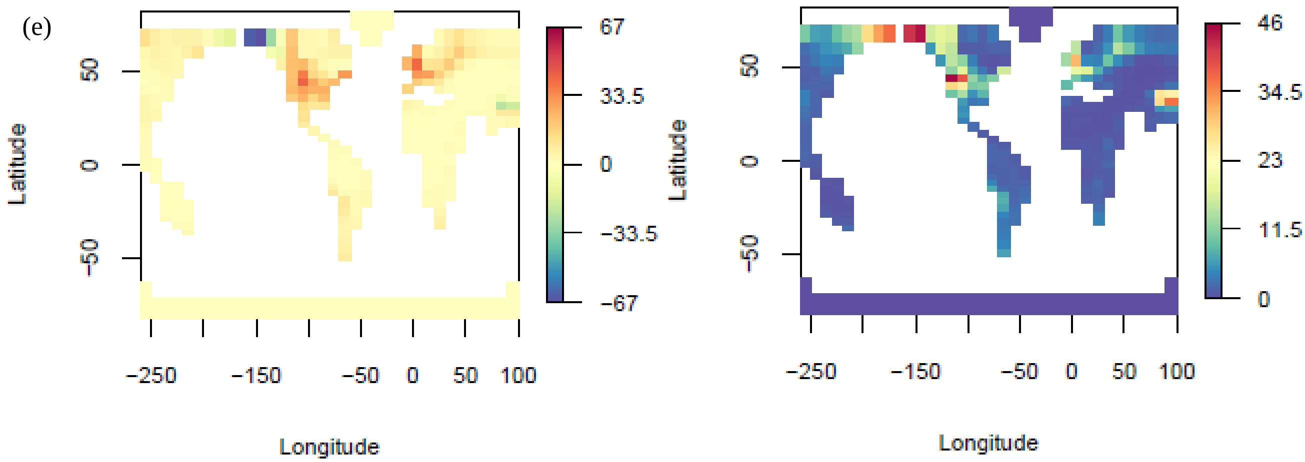


Fig. 11. LGM vegetation (a-b), soil (c-d) and total terrestrial carbon changes (e-f) ENS_{104} mean (left) and standard deviation (right). Units are kgC m^{-2}.

3.2.3 Ocean primary productivity

The ENS_{104} mean LGM total POC export flux anomaly ($\Delta\mathrm{POC}_{\mathrm{exp}}$) is -0.19 $\pm$ 1 PgC yr^{-1} and the range is -2.57 to 2.56 PgC yr^{-1}, roughly consistent with previous model-based estimates (e.g. Brovkin et al., 2002; Bopp et al., 2003; Brovkin et al., 2007; Chikamoto et al., 2012; Palastanga et al., 2013; Schmittner and Somes, 2016; Buchanan et al., 2016)~~ (Fig. 12). ~~. As shown in Fig. 12, t~~T~~he POC flux decreases in ENS_{315} and ENS_{16} tend to be smaller and larger respectively. $\Delta\mathrm{POC}_{\mathrm{exp}}$ is positively correlated with $\Delta\psi_{\mathrm{max}}$ (r = 0.72 and 0.79) and negatively correlated with $\Delta\psi_{\mathrm{min}}$ (r = -0.62 and -0.58) in both ENS_{104} and ENS_{315}. The correlations potentially suggest that decreasing AMOC strength and increasing AABW production lead to decreasing POC export. One possible mechanism is enhanced deep ocean stratification due to increasing AABW formation leading to not only more efficient trapping of DIC at depth (see above), but also nutrients and therefore reduced availability in the euphotic zone. All else held constant, a weaker and shallower AMOC cell would also inhibit the transfer of nutrients from the deep ocean to the surface. A negative correlation can additionally be found between $\Delta\mathrm{POC}_{\mathrm{exp}}$ and $\Delta\mathrm{SIA}$ (r = -0.55 and -0.6), probably because no primary production occurs beneath the sea ice surface.~~ and I~~increasing sea ice area at the LGM therefore leads to decreasing POC export flux. This would also explain the largest ENS_{104} mean decreases in POC export flux, shown in Fig. 13, coinciding with increases in sea ice fraction~~. (Fig. 13)~~.

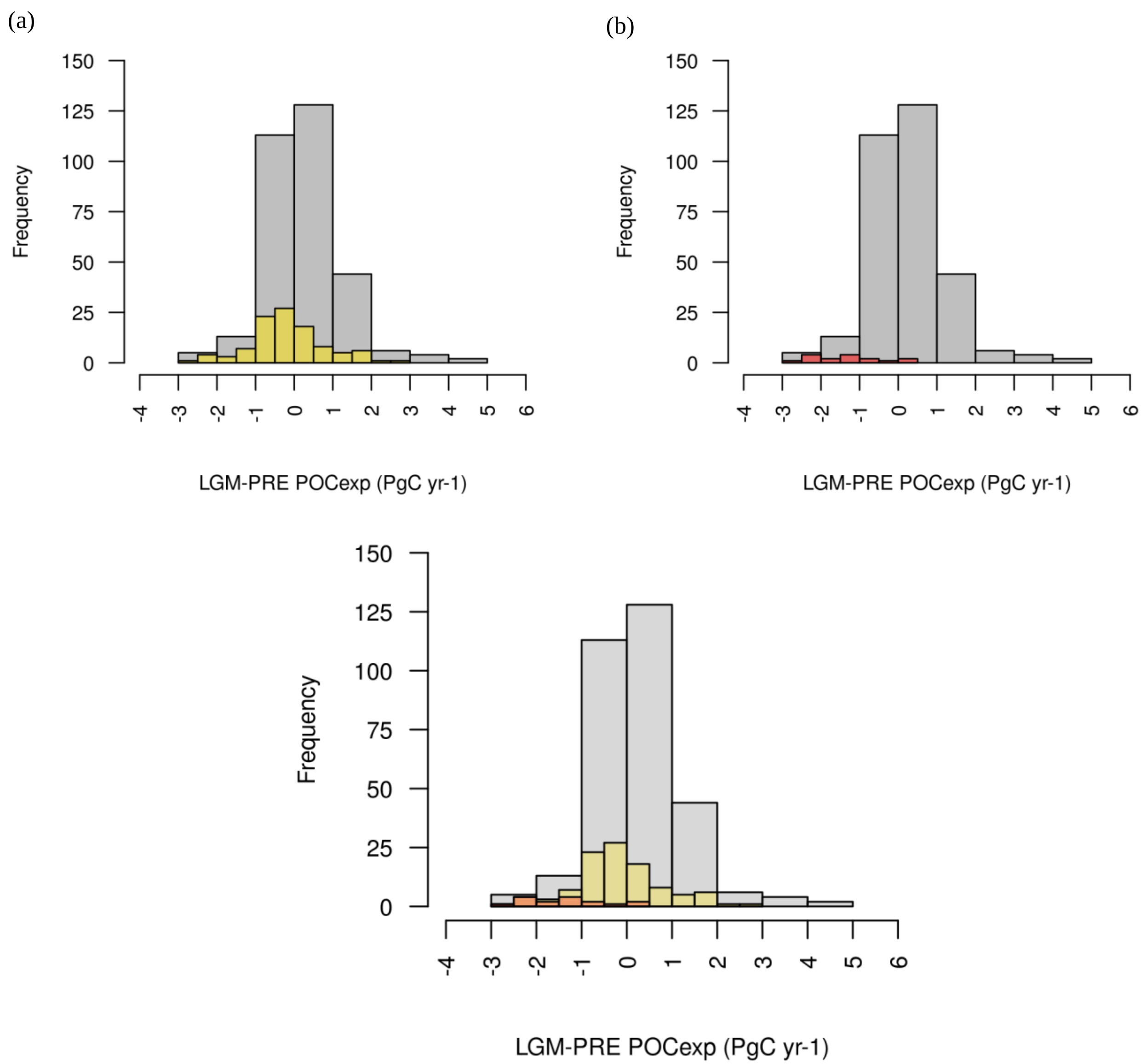


Fig. 12. LGM change in POC export flux ~~(a-b)~~ distributions.

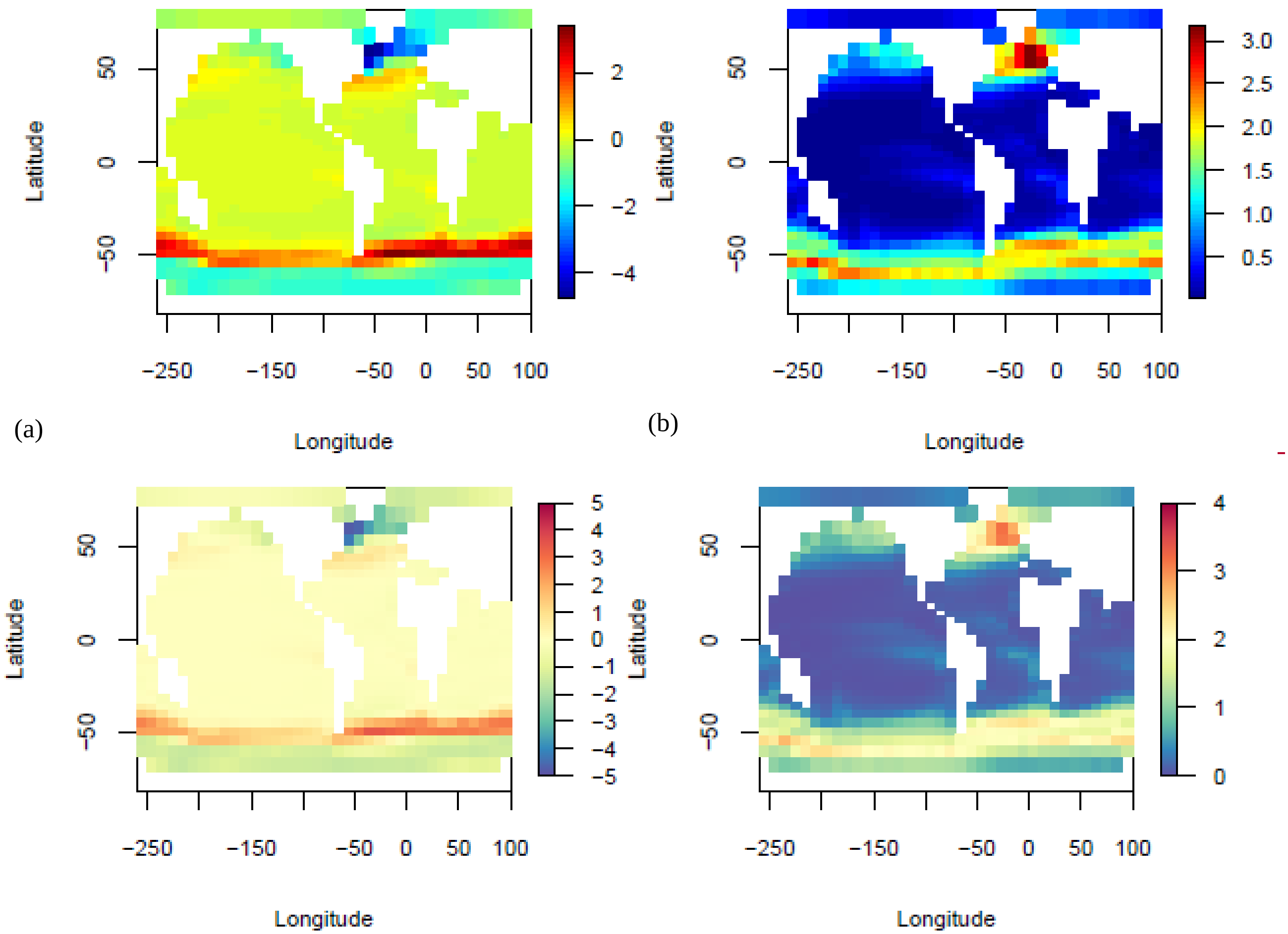


Fig. 13. LGM surface POC export flux change (molC m^{-2} yr^{-1}) ENS_{104} mean (a) and standard deviation (b).

The largest ENS_{104} LGM ~~PGACF ensemble~~ *increases* in POC export flux conversely occur at around 50 °S, roughly in front of the Antarctic sea ice margins. Increases in POC export are also simulated close to the North Pacific and Atlantic sea ice margins, as well as in the eastern equatorial Pacific and the southwest Atlantic upwelling region. The ~~sea ice margin~~increases in POC export flux ~~increases~~ at the sea ice margins are likely caused by the advection of unutilised nutrients from underneath the sea ice. However, they may additionally be ~~caused by~~due to the enhanced iron availability from the increased supply of aeolian dust, particularly in the Southern Ocean and North Pacific since these are strongly limited by iron (Ridgwell et al., 2007). Iron fertilization may also explain the increases in POC export flux in the eastern equatorial Pacific and in the southwest Atlantic upwelling region. Comparison against observations suggests that the ensemble mean POC flux changes immediately north, and south of the Antarctic sea ice margins align with observations of increased and reduced marine productivity in the Subantarctic (~45 to 60 °N) and Southern Ocean respectively (Kohfeld et al., 2005; Kohfeld et al., 2013; Jaccard et al., 2013; Martínez-García et al., 2014). The simulated decreases in export flux in the Arctic and subarctic Atlantic (i.e. above ca. 50 °N), and the increases in export flux immediately south of 50°N are also in agreement with previous reconstructions (Kohfeld et al., 2005; Radi and de Vernal, 2008).The mostly lower LGM export fluxes at the equator and in the South Atlantic are conversely inconsistent with the observational data of Kohfeld et al., (2005). The decreases may be caused by the increases in productivity in HNLC regions reducing the phosphate (the other limiting nutrient in GENIE-1 besides iron) availability for photosynthesis in other regions. They may additionally be due to the model not simulating enhanced nutrient inventories in response to enhanced weathering or reduced shallower water deposition of organic matter. The model also does not vary wind speed which may have resulted in stronger tropical upwelling in the Atlantic at the LGM. The evidence is more ambiguous (or missing) for the Pacific (Jaccard et al., 2010; Kohfeld and Chase, 2011; Kohfeld et al., 2005; Costa et al., 2016) and Indian Oceans (Kohfeld et al., 2005; Singh et al., 2011) and is therefore not discussed in more detail here.

3.2.4 Carbonate weathering and shallow water deposition

The ENS_{104} mean glacial weathering factor (GWS) is 1.16 ± 0.24 (corresponding to a percentage change in the land to ocean bicarbonate flux, ~~(~~%LOC, ~~)~~ of 38.67), and the range is 0.52 to 1.5 (corresponding to a %LOC between -49.33 and 50). As shown in Fig. 14, ~~(Fig. 14).~~ t~~T~~he GWS in ENS_{104} tends to be larger than in ENS_{315} and smaller than in ENS_{16}. There is also a negative correlation between GWS and $\Delta\mathrm{CO}_2$ (r = -0.52) in ENS_{315}, suggesting that increasing the input of bicarbonate to the ocean leads to a decrease in CO_2 by raising the inventories of ALK and DIC in a 2:1 ratio. In ENS_{104}, however, r is below the 0.05 significance level, suggesting that it is less important.

(a) (b)

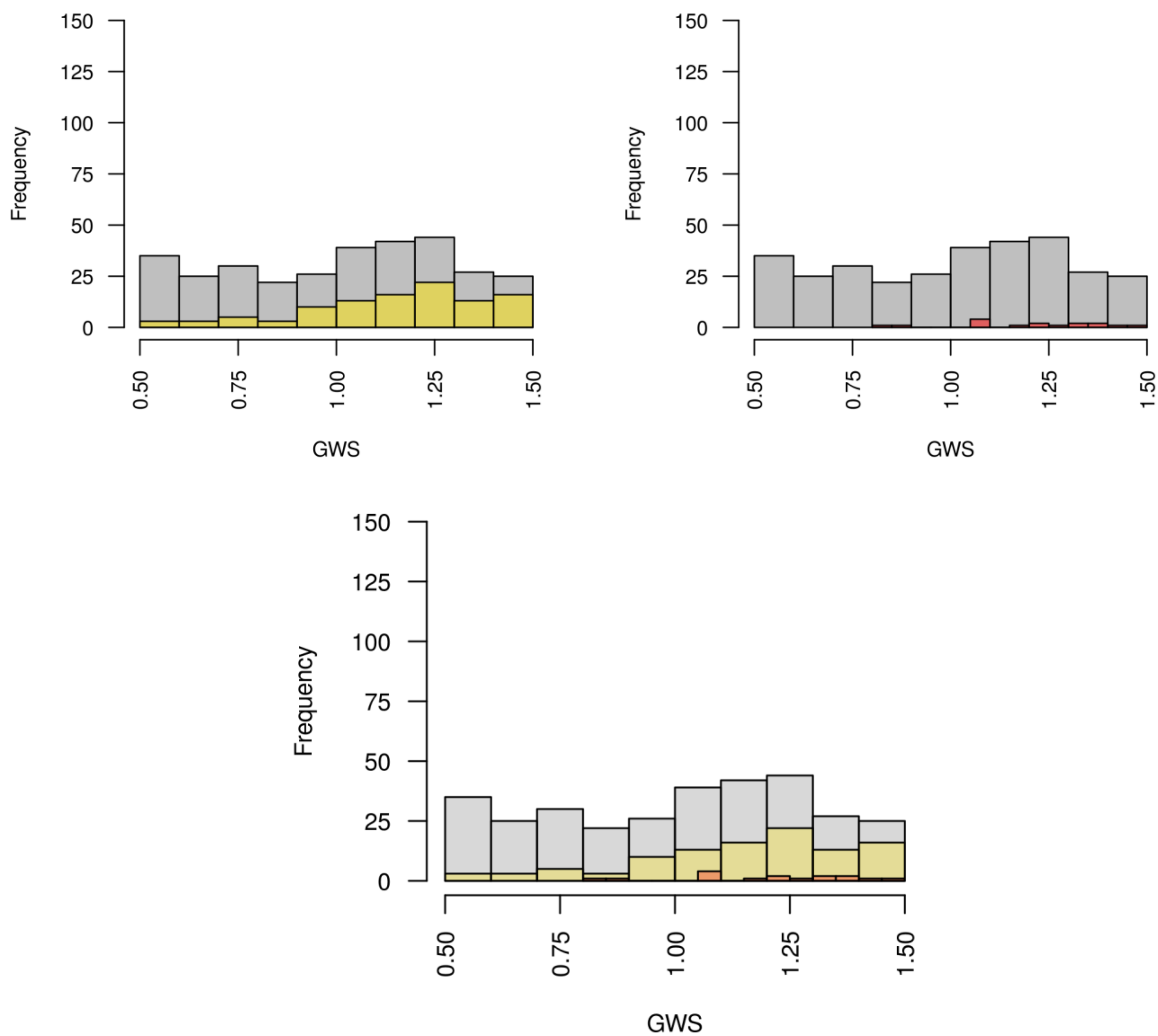
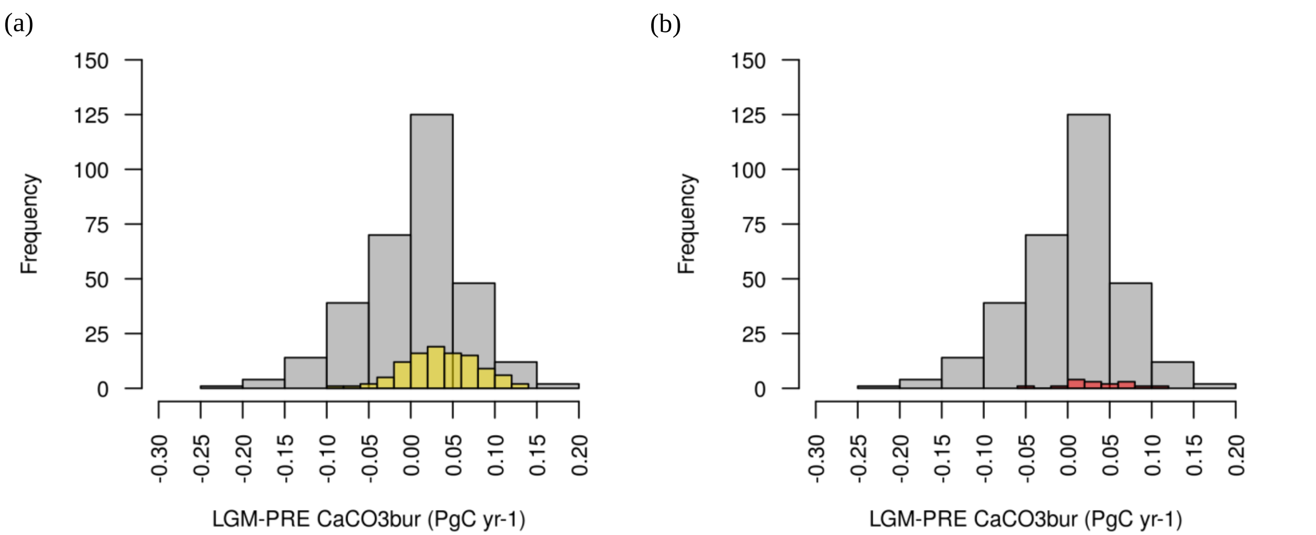


Fig. 14. GWS ~~(a-b)~~ distributions.

3.2.5 Deep-sea carbonate burial

The ENS_{104} mean global deep-sea $CaCO_3$ burial flux anomaly ($\Delta CaCO3_{bur}$) is 0.036 $\pm$ 0.045 PgC yr^{-1} and the range is -0.098 to 0.139 PgC yr^{-1} ~~(Fig. 15).~~. The mean value is ca. 3 times larger than the observed value (Catubig et al., 1998), although the latter still falls within the range of simulated values. As shown in Fig. 15, $\Delta CaCO3_{bur}$ in ENS_{104} tends to be higher than in ENS_{315}, and lower than in ENS_{16}. ~~It~~ The change in global deep-sea $CaCO_3$ burial flux anomaly is strongly determined by GWS, as suggested by the positive correlation (r = 0.88 and 0.9) between the two, in both ENS_{104} and ENS_{315}. Increasing %LOC (the percentage change in the land to ocean bicarbonate flux) should indeed ~~cause the~~enhance the $CaCO_3$ burial flux ~~to increase as~~as increasing ALK means the deep ocean CO_3^{2-} will eventually increase. The latter in turn~~,~~ would cause~~ing~~ the saturation horizon to fall~~ and~~, allowing $CaCO_3$ to accumulate over greater areas (which are now exposed to undersaturated waters) (Sigman and Boyle, 2000). The input of ALK to the surface ocean ~~will also increase~~would also increase the rate of $CaCO_3$ export production (~~which will in turn increase the~~enhancing the sediment deposition flux of $CaCO_3$) since as discussed in Chikamoto et al. (2008), the latter is proportional to the production rate of POC (which is equal to the POC export flux), together with the sea surface saturation state with respect to $CaCO_3$, in GENIE-1. There is indeed also a positive correlation between %LOC and the global change in $CaCO_3$ export (r = 0.27 and 0.4), and between the latter and $\Delta CaCO3_{bur}$ (r = 0.34 and 0.45) in both ENS_{104} and ENS_{315}.



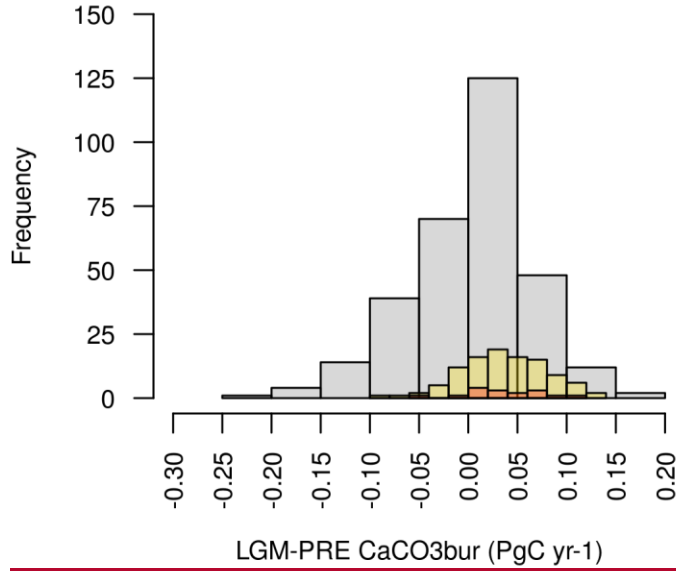


Fig. 15. LGM change in deep sea sediment $CaCO_3$ burial flux ~~(a-b)~~ distributions.

The ENS_{104} mean spatial distribution of $\Delta CaCO3_{bur}$ is shown in Fig. 16. Relatively large increases in burial flux ($\geq 0.5 \times 10^{-5}$ mol cm^{-2} yr^{-1}) can be found at around 50 °S, in the North Pacific, and to a lesser extent the North Atlantic. In other regions, the burial flux is significantly lower or negative, with the largest losses ($\leq$ -0.5 $\times$ 10^{-5} mol cm^{-2} yr^{-1}) occurring the North Atlantic and arctic regions. The only exception is the western North Atlantic, which exhibits a large increase in burial. A comparison of the results against the reconstructions of Catubig et al., 1998 is somewhat difficult as the coverage is poor but overall $CaCO_3$ burial was higher in the North Atlantic and the Pacific, and lower in the tropical and South Atlantic, and the Indian and Southern Ocean.

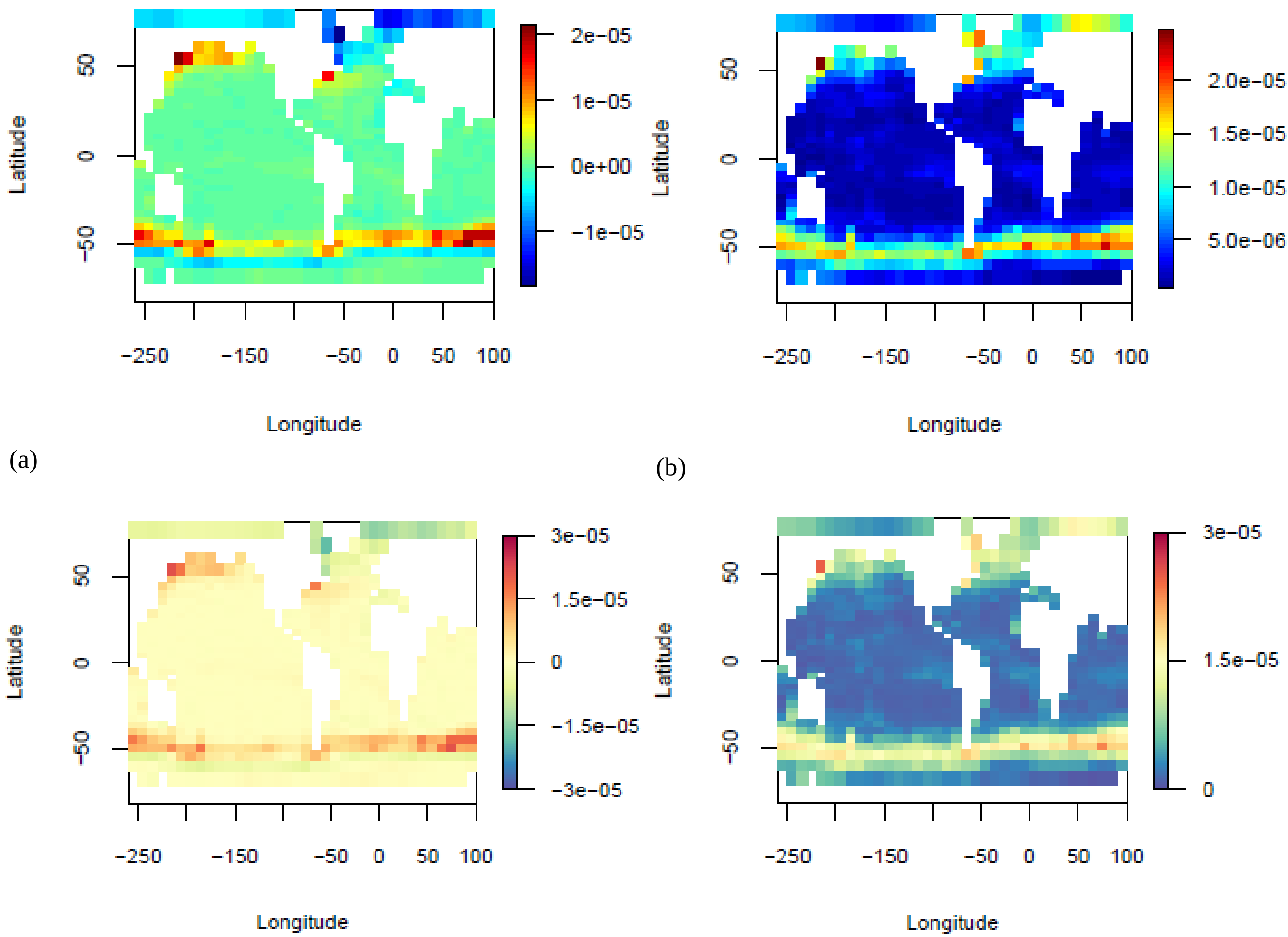


Fig. 16. LGM deep sea $CaCO_3$ burial rate change (mol cm^{-2} yr^{-1}) ENS_{104} mean (a) and standard deviation (b).

3.2.6 Other paleo proxies

~~To further evaluate the sign and magnitude of our simulated LGM changes, and in particular ΔTerrC, a key future test would be to add carbon isotopes into GENIE-1's terrestrial biosphere carbon module and re-run the ENS_{104} LGM simulations with $\Delta^{14}C_{atm\,(DIC)}$ and $\delta^{13}C_{atm\,(DIC)}$ spun up.~~ As shown in Table 4, a frequent argument for a lower glacial terrestrial carbon inventory is the reconstructed mean glacial ocean $\delta^{13}C$ value of ca. 0.35‰ lower than present due to the fact that plants discriminate against ^{13}C during photosynthesis. In our simulations conversely, it follows that the increase in glacial terrestrial carbon inventory would have resulted in an *increase* in ocean $\delta^{13}C$. Decreasing SSTs and increasing $CaCO_3$ weathering would have, moreover, likely raised it further~~thi~~ (in the first instance, by enhancing fractionation at the air-sea interface, and in the second instance, through the input of isotopically heavy weathering products). However, ~~other processes such as reduced marine productivity, lower SSTs (Schmitt et al., 2012) and greater sea ice area (Stephen and Keeling, 2000) may have counteracted at least some of this increase. Increasing LGM $CaCO_3$ weathering flux would have conversely raised ocean $\delta^{13}C$ although our model does not account for changes in the organic carbon weathering flux, which if increased would result in a decrease in $\delta^{13}C$. Wallmann (2014) attribute most of the observed LGM ocean $\delta^{13}C$ decrease to enhanced weathering and reduced deposition of organic carbon at continental margins due to lower sea levels. Menviel et al. (2015) have also shown that weaker surface winds can contribute to a lower ocean $\delta^{13}C$ while these are held fixed in our model. It has further been argued that enhanced glacial carbonate ion concentrations may have reduced the $\delta^{13}C$ in foraminera shells without altering mean ocean $\delta^{13}C$ (Lea et al., 1999).~~ as noted in Zeng (2007), the interpretation of $\delta^{13}C$ value can be complicated by factors such as the impact of enhanced glacial carbonate ion concentrations on $\delta^{13}C$ in foraminera shells (Lea et al., 1999). In addition, there are other processes in our model which may have counteracted at least part of the increase in ocean $\delta^{13}C$. These include reduced marine productivity (e.g. Zimov et al., 2009), as phytoplankton discriminate against ^{13}C during photosynthesis, giving the marine organic carbon reservoir a low $\delta^{13}C$. They also include greater sea ice area, which can lower the ocean $\delta^{13}C$ by reducing the air-sea gas exchange, and therefore the net transfer of ^{13}C into the ocean (Stephen and Keeling, 2000). We, moreover, propose that adding missing processes could have decreased ocean $\delta^{13}C$ even further. These include weaker surface winds (while in our model these are fixed), again through reduced air-sea gas exchange (Menviel et al., 2015), as well as enhanced weathering and reduced deposition of organic carbon at continental margins due to lower sea levels (Wallmann, 2014). More importantly perhaps, however, missing processes would likely be needed to reconcile our lower POC export fluxes with deep ocean oxygen records. These records tend to indicate there was a decrease in LGM deep ocean oxygen concentration (Jaccard et al., 2016). Lower POC export fluxes would conversely have resulted in an increase in deep ocean oxygen due to reduced oxygen consumption at depth. When observed over sufficiently large areas, lower oxygen concentrations can support the presence of an enhanced ocean carbon inventory as deoxygenation can be explained by reduced ocean ventilation (the sole input of oxygen is from the ocean surface) (Wagner and Hendy, 2015). The reduced ventilation is, in turn, assumed to have led to the accumulation of a significant amount of DIC in the ocean interior. Thus, explaining the lower deep ocean oxygen concentrations without having to reduce ocean ventilation as extensively as suggested by previous studies, would as a minimum likely require LGM export production to have increased rather than decreased. However, it may be possible to increase deep ocean oxygen consumption by increasing organic matter at depth but keeping the surface POC export flux constant. This would require adding missing processes such as increasing remineralisation depth with decreasing ocean temperature and increasing ballasting into our model (Kohfeld and Ridgwell, 2009; Menviel et al., 2012).

~~As discussed in Zeng (2003) the glacial increase in terrestrial carbon inventory may also potentially explain transient trends in the glacial-interglacial atmospheric $\delta^{13}C$ record, such as the increase in atmospheric $\delta^{13}C$ over the glaciation, and the decrease of about 0.3 ‰ at the beginning of the deglaciation (Smith et al., 1999). Another feature of the deglacial record, namely the rise in $\delta^{13}C_{atm}$ between ca. 12 and 7 kyr BP, is in turn attributed to increasing SSTs and terrestrial biosphere regrowth on previously ice-covered areas. A more common explanation for the deglacial $\delta^{13}C_{atm}$ variation, which resembles the letter W, is conversely that the beginning of the pattern was caused by the release of old carbon from the deep ocean while the end of the pattern was largely due to terrestrial biosphere regrowth. The middle section of the W, characterised by subdued variation in $\delta^{13}C_{atm}$, is attributed to abrupt climate changes (Schmitt et al., 2012). More recently, however, it has also been suggested that the deglacial decrease in $\delta^{13}C_{atm}$ at ca. 17.5 kyr BP may be caused at least partly by the demise of iron-stimulated Southern Ocean biological productivity (Fischer et al., 2015) and the release of carbon from thawing permafrost (Crichton et al., 2016). A better test, therefore, would be for the ENS_{104} ensemble members to not only simulate LGM-PRE $\delta^{13}C_{atm}$ but also the transient changes in $\delta^{13}C_{atm}$ over the deglaciation. A further test would be to compare the simulated spatial distribution of $\delta^{13}C$ with observations (e.g. Menviel et al., 2017).~~

~~Another frequent but more indirect argument for a lower glacial terrestrial carbon inventory is the deglacial drop in atmospheric $\Delta^{14}C$, which is typically attributed to the release of old ocean carbon accumulated over the previous glacial period. Zeng (2007) and others (e.g. Zech, 2012) conversely propose that it is caused by the release of very old, and therefore ^{14}C-depleted, carbon from the terrestrial biosphere. More recent higher resolution records of deglacial $\Delta^{14}C$ and atmospheric CO_2 (e.g. Durand et al., 2013 in Köhler et al., 2014; Marcott et al., 2014) have, however, so far not been discussed by the enhanced glacial terrestrial carbon inventory studies. Another challenge is that a terrestrial biosphere-induced early deglacial decrease in atmospheric $\Delta^{14}C$ would have also led to a decrease in ocean $\Delta^{14}C$ and this is yet to be reconciled with ocean $\Delta^{14}C$ data. Studies so far have suggested there was a decrease in deep ocean $\Delta^{14}C$ during the glaciation, corresponding to an increase in ventilation age, and an increase in deep ocean $\Delta^{14}C$ at deglaciation, corresponding to a decrease in ventilation age, and also coinciding with the decrease in atmospheric $\Delta^{14}C$ (Hughen et al., 2006; Skinner et al., 2010; Skinner et al., 2015; de la Fuente et al., 2015; Freeman et al., 2015). With a reduced deep ocean ventilation at the LGM, it is assumed that the carbon sequestration capacity~~

~~of the ocean was enhanced, and that the magnitude and spread of these changes resulted in a global increase rather than decrease in ocean carbon.~~

~~A potential limitation, however, with the use of ocean $\Delta^{14}C$ to infer larger LGM ocean carbon inventory pools is the presence of $\Delta^{14}C_{DIC}$ which indicate no change in ventilation age, or conversely an increase (Broecker and Barker, 2007; Broecker and Clark, 2010; Cléroux et al., 2010; Lund et al., 2011), and processes besides water mass aging which may have contributed to decreasing glacial $\Delta^{14}C_{DIC}$ such as decreasing atmospheric ^{14}C production rate, increased weathering of ^{14}C-depleted $CaCO_3$, input of ^{14}C-depleted carbon from the mantle and inaccurate estimation of surface ocean reservoir age (Broecker and Barker, 2007; Lund et al., 2011; Wagner and Hendy, 2015). The feasibility of a large extremely ^{14}C-depleted deep ocean carbon reservoir has also been contested in terms of atmospheric CO_2, deep ocean oxygen (namely the absence of large-scale anoxia) and $CaCO_3$ depth constraints (Hain et al., 2011), and from a dynamical standpoint (Broecker and Clark, 2010). Another strong future test would therefore be for ENS_{104} to simulate spatially resolved LGM ocean $\Delta^{14}C$, as well as the changes in atmospheric $\Delta^{14}C$ over the deglaciation.~~

~~Finally, we have shown that decreasing LGM atmospheric CO_2 in our ensemble tends to be accompanied by decreasing POC export production, which all else held constant would result in an increase in deep ocean oxygenation – a feature which we have not assessed. Proxy records to date conversely tend to indicate that there was a decrease, not an increase, in deep ocean oxygen concentration (Jaccard et al., 2016). When observed over sufficiently large areas, the latter supports the presence of an enhanced ocean carbon inventory as deoxygenation can be explained by reduced ocean ventilation (the sole input of oxygen is from the ocean surface) (Wagner and Hendy, 2015). As discussed above, the reduced ventilation is in turn assumed to have led to the accumulation of a significant amount of DIC in the ocean interior. Thus, explaining the lower deep ocean oxygen concentrations without having to reduce ocean ventilation as extensively as suggested by previous studies would as a minimum likely require LGM export production to have increased rather than decreased (assuming the enhanced POC export production does not automatically result in an increase in glacial ocean carbon inventory). Alternatively, it may be possible to increase deep ocean oxygen consumption by increasing organic matter at depth but keeping the surface POC export flux constant. This would require adding processes such as increasing remineralisation depth with decreasing ocean temperature and increasing ballasting into our model (Kohfeld and Ridgwell, 2009; Menviel et al., 2012). We also note that including missing processes affecting the ocean biological pump such as increased oceanic PO_4 inventory could potentially result in a net increase in LGM POC export, helping lower oxygen concentrations. However, this would then make reconciliation of our positive ΔTerrC with the observed negative LGM ocean $\delta^{13}C$ more difficult.~~

4 Conclusions

We have used an uncertainty-based approach to investigating the LGM atmospheric CO_2 drop by simulating it with a large ensemble of parameter sets and exploring the range of ~~, designed to allow for a wide range of large-scale feedback response strengths. Our objective was not to definitely explain the causes of the CO_2 drop but rather explore the range of possible responses. Our investigation also involved simulating the CO_2 drop with the simulated CO_2 feeding back to the simulated climate, which is still relatively rare in LGM atmospheric CO_2 experiments, and the first time it is done with GENIE-1. Moreover, rather than assuming that terrestrial carbon gets destroyed by the LGM ice sheets, we allowed for its gradual burial. This assumption had not yet been implemented in an equilibrium set-up.~~possible responses. Despite our ensemble varying many of the parameters thought to contribute to variability in glacial-interglacial atmospheric CO_2, ~~not all sources of uncertainty could be captured, and this was reflected in our simulated ΔCO_2 distribution.~~we ~~We~~ estimated that up to ~60 ppmv of ΔCO_2 could be attributed to processes not included in our model and error in our process representations. As a result, we treated ~~simulations with~~ ΔCO_2 between ~-90 and -30 ppmv as ~~“~~equally plausible~~”~~, and focused on describing the responses ~~physical and biogeochemical changes seen in~~of the subset of simulations with this ΔCO_2 ~~subset, as well as their linkages to ΔCO_2.~~ .We found the range of responses to be large, including the presence of five different ways of achieving a plausible ΔCO_2 in terms of the sign of individual carbon reservoir changes. However, several dominant changes could be detected. Namely: the LGM atmospheric CO_2 decrease tended to predominantly be associated with decreasing SSTs, increasing sea ice area, a weakening of the AMOC, a strengthening of the AABW cell in the Atlantic Ocean, a decreasing ocean biological productivity, an increasing $CaCO_3$ weathering flux and an increasing deep-sea $CaCO_3$ burial flux. The majority of our simulations also predicted an increase in terrestrial carbon, coupled with a decrease in ocean and an increase in lithospheric carbon. The increase in terrestrial carbon, which is uncommon in LGM simulations, was attributed to reduced soil respiration in response to the climate forcings, as well as our choice to preserve rather than ~~-~~destroy ~~ice sheet carbon~~carbon that accumulates in ice sheet areas.~~, as well as the fact that the latter increased rather than decreased in response to climate forcings~~. ~~If most of the carbon that was present in ice sheet areas at the end of the preindustrial runs had been lost to climate forcings, it would not matter much whether the remaining stocks had been destroyed or preserved. From the literature, it is not clear how much of the carbon in ice sheet areas is thought to have been lost strictly in response to ice sheet “bulldozing” versus climate impacts.~~ An initial comparison of these dominant changes with observations and paleo-proxies other than carbon isotope and oxygen data suggests broad agreement. However, more detailed comparisons, in future studies, would be useful. We also suggest that, conceptually at least, our results can only be reconciled with carbon isotope and oxygen data if additional processes not included in our model are taken into account.~~An initial comparison of the dominant changes with observations and paleo-proxies suggested broad agreement. However, a comparison against carbon isotope data would be needed to more robustly determine agreement between our model results and empirical data. Another useful future research endeavour would~~

~~be to investigate the relationships between the simulated changes and ensemble parameters, in order to help isolate the individual mechanisms that directly, or indirectly, cause atmospheric CO_2 to change.~~

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