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Discussions

Interactive comment on "Glacial marine carbon cycle sensitivities to Atlantic ocean circulation reorganization by coupled climate model simulations" *by* M. O. Chikamoto et al.

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Reply to Referee 2

Thank you very much for your helpful suggestions and comments. We modified the manuscript according to all of your comments. *Iteric font* shows Referee2 comments and roman font shows our reply.

In general I felt that the experiments merit a better analysis and discussion. Sections 3 to 5 are very descriptive and a bit listy, and would benefit from a more *careful guiding of the reader to the main findings of this study. Furthermore I felt that some of the most interesting questions that this study could address were not properly discussed. For example, the discussion of circulation changes and carbon storage at section 5.1 is probably the most exciting aspect of this GCM study and this section does not do it justice. For instance, the topical question of the balance between increased AABW formation vs. Southern Ocean stratification in controlling pCO2 is not properly discussed. I also wanted to know more about the causes of the circulation changes; AABW changes do not have to be driven by NADW changes. The discussion of preformed nutrient contents should be expanded to consider potential changes in productivity in source regions, for instance as a result of Fe fertilisation (see Hain et al. 2010).*

1) Effect of the balance between establish of AABW formation and Southern Ocean stratification to the atmospheric CO_2

In the glacial experiment LGb, the maximum of AABW flow increases by 2Sv to 12 Sv compared to PIb, suggesting the enhancement of AABW formation. Since AABW carries carbon-rich water compared to NADW, the dominant mode of AABW contributes to the carbon storage in the abyssal ocean (Fig. 7c). However, as shown by the vertical uniformed changes in δ13C and DIC south of 40S, the Southern Ocean stratification would not be enhanced. Indeed, in our model, the dense water is produced near sea-ice margin via brine rejection, which accompanies to the density-induced deep convection. These features would contribute to the CO_2 release from the ocean to the atmosphere through mixture between surface and deep water, thereby offsetting to the effect of carbon storage by the AABW formation. We added these descriptions in Subsection 5.1.

2) Switching mechanisms of deep-water formation site from the NA to the SO

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The Southern Ocean cooling and the growth of sea ice cause much deep-water formation in the Southern Ocean. This AABW-origin water expands to the deep Atlantic Ocean, compared to the NADW-origin water. The establishment of AABW formation reduces the heat transport to the North Atlantic Ocean, thereby lowering SST in the North Atlantic as well. This seems to prevent the deep-water convection through the expanded sea ice. One of the coauthors in this paper, Ayako Abe-Ouchi, is in preparation for this mechanism using multiple sensitivity studies in MIROC, therefore we don't argue the switching mechanism of deepwater formation more details in this paper. Rather we focus on the change in water chemical property due to the switching of deep and bottom water formation from the North Atlantic to the Southern Ocean. The enhancement of AABW formation reduces δ13C at intermediate depth in the South Atlantic (Fig. 6d), which is not found in a case of NADW slow-down due to the freshwater horsing [Taguliague et al. 2009]. The difference in δ13C response between our and Taguliague study may reflect to the difference of AMOC features, such as the switching of AABW from NADW or the NADW/AABW weakening. We evaluated the enhanced intrusion of AABW by comparing δ13C anomaly with proxy data in subsection 3.2.

3) Potential change of productivity in associated with Fe fertilization.

We have evaluated iron fertilization in LGb simulation and shown enhanced productivity in the Southern Ocean when glacial dust deposition was added in MIROC experiment LGb [Oka et al., 2011]. We added the possibility of source change and its effect to the preformed nutrient change in Subsection 4.4.

I also felt the discussion of $p\mathrm{CO}_2$ changes due to sea-ice, another major aspect of this study, was lacking. I wanted to know more about "capping" of CO_2-rich waters by sea-ice, a topical question which has had relatively little testing. How is permeability parameterised? And biological productivity? And how sensitive are some of these parameters, and thus why are $p$$CO_2$ changes (for LG-is) less dramatic than those of Stephens and Keeling.

According to your suggestion, we discussed the difference of sea-ice effect between Stephens and Keeling (2000) and our study. In Stephens and Keeling (2000), the atmospheric CO_2 drastically decreases by over 10 ppmv when sea ice area increases by 96 % south of the Antarctic Polar Front (APF). On the other hand, our simulation shows the area of annual-mean sea ice coverage reaches to 2.0 $\times$ 10^{13} m^2 and the fraction of sea ice area from free sea-ice area reaches to 85 % for the ocean area south of 60°S (2.3 $\times 10^{13}$ m^2). In this case, the atmospheric CO_2 drops by 6 ppmv, which is consistent with the pCO_2 response in the 85-% sea-ice case of Stephens and Keeling (2000). The fraction of sea-ice coverage to ice-free area south of APF would affect the atmospheric CO_2 change. In LGb-is, primary productivity in the Southern Ocean reduces by 56 %, but global net primary production decreases by 4 % only. This is because a reduction in primary production below the sea ice area is offset by an enhancement of production in the sea-ice margin through a meridional transport of unutilized nutrient. We described these points in Subsection 4.3.

1274, 18: The fact that the solubility experiments don't include CO_2 solubility change due to whole ocean salinity increase is an important caveat, so be more clear with this discussion, and in the initial description of the factorial solubility experiments. This does seem a significant shortcoming of the solubility factorial experiment, though it is hard to think of a way to include this effect without potentially changing circulation - does circulation show a significant change when a uniform salinity increase is applied?

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And how is the 10 ppm increase (line 19) assessed?

As you suggested, the entire salinity change in association with the sea level drops affect the whole ocean solubility, density-driven ocean circulation, and isopycnal-mixing, which would change atmospheric CO_2. We added this possibility in Section 4.4. About the salinity increase effect, we conducted an idealized simulation with additional 1-psu salinity: we give PIb temperature, salinity and ocean dynamics but LGb SST, LGb SSS with an additional 1-psu. In this revised version, we reran this idealized experiment and obtained atmospheric CO_2 of 252 ppmv. This is 6 ppmv higher than the result in PIb-sl (246 ppmv). The atmospheric $p\mathrm{CO}_2$ change due to sea level-driven solubility effect is +6 ppmv.

However it is interesting that the PI-sl and LG-sl experiments give values in such close agreement: given that PI-sl has LG SSS but PI interior sal, and that LG-sl has PI SSS, but LG interior sal, it seems that the effects of salinity on air-sea gas exchange vs. whole ocean CO_2 solubility are roughly equal. This should be discussed.

Our results suggest that effects of temperature and salinity on air-sea gas exchange are roughly equal even when ocean interior CO_2 solubility changes from present-day to LGM, as you suggested. In addition, salinity and temperature in the interior domains are used for the tracer mixing of isopycnal layers in our offline model. Therefore the close results between PI-sl and LG-sl mean that the effects of whole ocean CO_2 solubility and isopycnal mixing on atmospheric $p\mathrm{CO}_2$ are small when we do not consider the salinity increase due to sea-level drops. We added this description in Subsection 4.1.

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C1645 The same argument applies to the circulation experiments, where the switch from PI to LG circulation states will include +10 ppm pCO$_2$ due to increased salinity. This masks the true effect of 'circulation', in terms of changing water mass dominance and movement. For instance when comparing experiments LGb-oc (PIb circ) to LGb (LGb circ), the switch to LGb circulation will also include +10 ppm pCO$_2$ increase due to higher salinity. As the pCO$_2$ difference between the experiments is 264 ?> 260 (-4 ppm), yet the salinity increase should have raised pCO$_2$ by +10 ppm, the effect of this circulation change is really -14 ppm. Investigating the effect of circulation change on pCO$_2$ is a major aim of this paper, so this issue should be considered for each circulation state, and these results discussed.

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1276, 16: as mentioned previously, when the effect of whole ocean salinity change is accounted for, the circulation effects shown in Fig 7 go 10 ppm pCO$_2$ lower. The pCO$_2$ change due to LGb circ becomes -14 for LG conditions, and -3 for PI conditions.

As described above, we estimate the effect of sea level-driven salinity increase by applying the surface salinity addition of 1 PSU, not by applying the salinity addition for the entire ocean. Since our rough estimate corresponds to the surface salinity-increased sea surface solubility only, the exact effect of the whole salinity increase on atmospheric CO$_2$ can be hardly evaluated. According to your suggestion, we noticed that our estimate excludes the CO$_2$ change due to the whole-salinity increase, which would have the potential to amplify the ocean circulation effect in Subsection 4.4.

1276, 24: this analysis of surface DIC, ALK, pH and CO$_2$ is interesting, but I don't think the most interesting aspects of it are fully discussed, and the current interpretation seems at odds with the other experiments. The fact that the high surface *ocean pCO_2 at high latitudes under LGb circulation does not lead to significantly increased atmospheric CO_2 is attributed to sea ice cover. However if glacial sea ice is the key factor in preventing high latitude CO_2 escape, then we would expect that reduced (PI) sea ice cover with LGb circ, as in experiment LGb-in, should cause an increase in atmospheric pCO_2. Instead, the opposite effect is observed, with reduced sea ice causing a decrease in atm pCO_2. This highlights the main point that these figures seem to make: that surface pCO_2 in these zonal averages is a relatively unimportant predictor of atm pCO_2 change. The most important factor controlling atm pCO_2 seems to be the reduced surface DIC (and increased deep DIC storage), not the increased surface pCO_2.*

In Subsection 4.4, the simultaneous changes in DIC and alkalinity and the change in accumulated carbon were shown. The sea ice effects on atmospheric CO_2 are different between the North Atlantic and the Southern Ocean. In LGb-in, the northern sea ice extent inhibits the oceanic carbon uptake by the outcrop of cold water, contributing to an atmospheric pCO_2 rise. In contrast, in LGb-is, the southern sea ice prevents CO_2 release from the ocean, thereby reducing atmospheric CO_2. Newly added figure 13 e and f shows that the reduced accumulated carbon in the NA is contributed by the northern sea-ice extent, whereas the increased carbon is partly caused by the southern sea-ice extent. At low latitudes, the simultaneous changes in DIC and alkalinity does not affect the ocean surface CO_2 concentration. Because of high occupation of ocean area at low-mid latitudes, these changes contribute to a small sensitivity of atmospheric CO_2, as well as at polar region. We modified our description for this analysis (Subsection 4.4).

13C and 14C provide some of our best constraints on LGM circulation and carbon cycling. These tracers are included in the model (1265, 21) and should be plotted and compared to the data to allow evaluation of the LGb state. We added anomalous δ13C and Δ14C in Figs. 5 and 6.

1263, 17: be more clear - is it the increased density, or increased volume, or both?

Boutes et al. (2011) focus on both increased density and increased volume. In their paper, if the very density glacial water is produced via brine rejection and transported without mixing with the surrounding waters, the large volume of carbon-enriched deep water can expand to the deep Atlantic Ocean, and therefore prevent CO_2 from the atmosphere.

1266, 10: would be good to discuss why this warm bias might occur

The current version in our model overestimates the incoming shortwave radiation in the Southern Ocean in associated with small cloud covers and then tends to have a warming bias (Watanabe et al., 2010). We added this sentence in Subsection 2.2.

1266, 26: is it the cooling itself, or brine-rejection, or P-E changes, that leads to in- creased AABW formation?

In the Southern Ocean in LGb, the density change due to the salinity increase via brine rejection accounts for 92 % of the surface density increase. Therefore the brine-rejection enhances the AABW formation much more than the cooling effect. The

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Southern Ocean is characterized by a near-freezing temperature even in the present day, which allows a small temperature change under glacial conditions. This partly contributes to a small impact of cooling on water density. In LGb, both precipitation and evaporation decrease in the Southern Ocean, and the difference between precipitation and evaporation results in a 0.2 mm/day decrease. This change is quite smaller than the surface freshwater change of +1.9 mm/day due to the sea ice melting during the summer. Therefore, the P-E change does not lead to an increase in AABW formation.

1267, 5: considering a closed system carbon cycle, the difference between these experiments is interesting in its own right - 400 GtC is 2/3 the PI atmosphere. How big a change in CO_2 is seen if the cooling of 'b' is applied to an 'a' experiment with non- restoring CO_2?

&

1272, 10: would be good to also have the equivalent basin DIC inventories quoted for LGb, as for the LGa experiment. This would also clarify the statement that 'carbon inventories are reduced': if the total DIC inventories are reduced in each basin, then (given a closed ocn-atm system) atmospheric CO_2 must have risen?

We reran the LGb simulation but using the initial state of PIa carbon distribution and inventory, and then obtained the atmospheric CO_2 reduction of 42 ppmv. This change is larger than that in LGb, suggesting that how much the initial carbon is pooled in the ocean significantly affects the atmospheric CO_2 sensitivity. We described the effect of background climate-carbon state in Subsection 5.1.

1267,11: how long are factorial experiments run for?

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We simulated 2000 years for each factorial experiments.

1269, 13: this is confusing - surely slower settling should give shallower regeneration, and result in higher nutrient contents in the upper ocean? Is this a far-field effect?

We modified the description of nutrient distribution. Slower settling particles contribute to more POC remineralization within the euphotic zone and maintains nutrient in the shallower ocean compared to the distribution using the standard velocity (Plattner et al., 2001).

1269, 24: any idea why the fluxes are larger than observations?

We overestimate primary production around Antarctic, which partly contributes to larger flux compared to observations. Since the productivity is not limited to water temperature (Oschlies and Garcon, 1999), primary production is maintained even in cold water environment.

1270, 12: good that preformed nutrient content is assessed; would be good to include more in discussion of LG results.
&
1273, 10: I don't think that these changes in preformed nutrient fraction or the ?20 ppm changes in pCO_2 should really be classed as "considerable". Also, although I like seeing the changes in preformed nutrient fraction, I don't think these should be

classed as strictly "biological". They could be achieved with fixed productivity regimes, but by forming more deep water through the inefficient Southern Ocean pathway relative to the efficient North Atlantic (e.g. Toggweiler et al. 2006).

As following your suggestion, we modified the analysis of preformed nutrient change due to ocean circulation change. We discussed the preformed nutrient change in Section 4.4 and newly shown in Figure 12. To put the potential of an ocean dynamics perspective on excess nutrient mechanism in the Southern Ocean (Toggweiler et al., 2006), we described this perspective in Subsection 5.1.

Section 3.2 is not well structured - jumps between different parameters and experiments.

We described LGa and LGb separately to assess more clearly the biogeochemical responses to the two types of glacial climate. In addition, the description ofδ13C and Δ14C response to the glacial climate were inserted in Section 3.2 in order to evaluate the glacial simulation using proxy records.

1271, 23: how does this compare to pore water estimates (Adkins et al. 2002)?

As shown in Subsection 5.1, the deep-water temperature in the Southern Ocean drops by 5 °C and the salinity increases by 1.7 psu, when considering salinity increase due to sea-level drops. However this change is not enough to reach the deep ocean proxy records of 37 psu (Adkins et al. 2002).

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1272, 7: I don't understand the statement about DIC 'delivery' - why does forming more low DIC water in the Atlantic mean that Pacific water DIC should increase? As an alternative, perhaps NPIW formation is more sluggish, allowing more DIC to accumulate in this region?

In both LGa and LGb, the winter mixed layer depth deepens in the northeastern Pacific compared to the preindustrial states, which would reflect to a rather increase in NPIW formation. In both cases, the nitrate concentration increases at deep depth in the entire Pacific Ocean. This may suggest that the nutrient accumulation in the Pacific Ocean is responsible for an anomalous meridional transport of unutilized nutrient of Antarctic waters where the sea ice extent prevents biological production.

1272, 19: why do nutrients increase in upwelling zones? In the Pacific this seems likely to be related to higher nutrient concentrations in the subsurface, and is likely to be linked to the increased intermediate depth DIC.

In LGb, unutilized nutrient remains around Antarctica where the sea ice extent prevents biological production. This anomalous unutilized nutrient advects northward via AAIW and then contributes to the nutrient increases in the Pacific Ocean.

1272, 24: to what region does this export flux production apply?

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Global export production is shown.

1272, 27: by what processes does reduced NADW lower surface nutrients? Could this not be due to reduced mixing of nutrients trapped in dense AABW, or less nutrients in better ventilated intermediate waters?

Over the sea ice area, the reduction in ventilation and mixing in the North Atlantic inhibits the nutrient transport from the deep ocean. This nutrient-deficit water advects southward along the shallow meridional overturning, contributing to the nutrient reduction at intermediate depth in the Atlantic Ocean.

1274, 4: true that the response is non-linear. However this is not in itself an explanation of why LGb is more sensitive. These variations in sensitivity under different boundary conditions are an interesting feature of this study and deserve more discussion.

The climate factors of SST and sea ice extent changes larger in LGb from PIb than in LGa from PIa, which cause the larger individual $p\mathrm{CO}_2$ responses in LGb than in LGa. Explained in Section 4.

1275, 12: the phrase "exports more carbon" is, again, not clear to me. I think you mean that DIC in NADW is reduced, AND as NADW is more dominant, whole ocean DIC is therefore reduced (and $p\mathrm{CO}_2$ is higher).

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We modified the sentence as follow: Since NADW is characterized by DIC-poor water compared to AABW, the more dominant mode of NADW reduces DIC preservation over the North Atlantic Ocean (Fig. 10b and f).

1275, 18: are these effects really totally equivalent? How is this assessed?

They evaluated the effect of ocean carbon uptake in the North Atlantic Ocean by switching from the control to the glacial NA SST, and then NA contributes to about 40% of the total atmospheric CO_2 change. However this estimate focuses on the solubility, which is inappropriate to this section of sea-ice effect. Therefore we removed this explaination.

1279, 2: simultaneous DIC and ALK changes are to be expected. Also surface pCO_2 seems to increase, not decrease.

In the Atlantic Ocean surface pH decreases by 0.004 and ocean surface pCO_2 decreases by 2 ppm in LGb from LGb-oc.

1279, 16: this paragraph could be shortened. It would be better to briefly mention these factors and focus on things that the model results presented here can address. If carbonate compensation is to be discussed, then an estimate of the magnitude of this effect should be included.

We shorten this sentence to emphasize our results.

1280, 17: quantify "does not regulate atm pCO_2 *much". This is a topical and interesting discussion and could be expanded.*

We simply evaluate the effect of glacial wind speed change to the atmospheric CO_2 through gas exchange between the atmosphere and ocean. Southern Ocean overturning is affected by wind, so that the wind-driven circulation change would affect the atmospheric CO_2 as well through mixture of the surface carbon-poor waters with carbon-rich deep waters. However it is difficult to quantify the impacts of wind-driven ocean circulation anomalies on atmospheric CO_2 because of the complicated relationship between deep-water formation and deep upwelling. We added the description in Subsection 5.3.

1281, 5: again, what vertical DIC gradient?

The vertical gradient of DIC between subsurface and deep ocean is 38 mmol/m^3 in the South Atlantic Ocean.

1281, 8-11: does not seem to be a major conclusion of this study.

Removed.

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We modified the sentences following all of the other minor comments.

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