

Reply to the comments by the anonymous Referee #2

I miss a discussion with a critical assessment of the model and its results.

We now discuss the limitations of our model in the revised version of our paper. The major limitations are i) very low spatial resolution, ii) water fluxes between model boxes that are not derived from internal model dynamics, iii) terrestrial inventories of POC in vegetation and soil that are kept constant over the model period. We now discuss to what extent our results are affected by these model limitations.

1a) The authors are asked to perform a simulation where atmospheric d13C is not prescribed but evolves freely for the standard setup and for the setup with constant circulation...

The atmospheric $\delta^{13}\text{C}$ record is affected by vertical mixing processes in the Southern Ocean and the distribution of SSTs and vertical $\delta^{13}\text{C}$ gradients within this region (KÖHLER et al., 2005; SCHMITT et al., 2012; TSCHUMI et al., 2011). Sediment records show an anti-phased pattern in export production and vertical mixing between the region south of the Antarctic Polar Front and the Subantarctic Ocean which may have a strong effect on atmospheric $\delta^{13}\text{C}$ (ANDERSON et al., 2014). Moreover, terrestrial processes may affect the deglacial and Holocene $\delta^{13}\text{C}$ record (SCHMITT et al., 2012). None of these processes are resolved by our model. The entire global surface ocean at $>30^\circ\text{S}$ is pooled in a single box (Fig. 2) while changes in terrestrial POC pools are ignored. In summary, these limitations hinder us to simulate the carbon isotopic composition of atmospheric pCO_2 with our box model.

1a) continued: By prescribing atm. d13C, the ocean d13C signature is also forced through air-sea gas exchange; the comparison of measured and "simulated" d13C-DIC in table A5 appears therefore not very relevant...

To avoid circular reasoning, we indeed excluded surface water $\delta^{13}\text{C}$ values from our model-data comparison (Tab. A5) because they are affected by the prescribed atmospheric $\delta^{13}\text{C}$ values. Sensitivity tests showed that the $\delta^{13}\text{C}$ values of intermediate, deep and bottom water boxes employed to derive the glacial circulation field (Tab. A5) are not significantly affected by the prescribed atmospheric $\delta^{13}\text{C}$ values since the inventory of ^{13}C residing in the global ocean exceeds the atmospheric inventory by almost two orders of magnitude.

1b) 2409, line 27: LGM water fluxes were tuned to match LGM d13C distribution. It is not clear to me how this tuning was performed...

The tuning was done using full transient runs with all forcings applied.

2a) ... (i) to which extent this 24-box model can meaningfully simulate oxygen concentrations (ii) to which extent oxygen in the water column is a proxy for oxygen in the sediment layer. Finally it is controversially discussed whether and to which extent such a fractionation between the burial of C and P indeed occurs (Anderson et al., 2001) and it would be worthwhile to mention this.

Our model predicts for the glacial ocean a general oxygen decline at >2000 m water depth (Fig. 8), consistent with analytical data (JACCARD and GALBRAITH, 2012). Moreover, we use transfer functions derived from global compilations of in-situ benthic fluxes to define benthic denitrification rates and turnover rates of P in marine sediments (BOHLEN et al., 2012; WALLMANN, 2010). Analyses of these benthic flux data suggest that rates of benthic denitrification are influenced by both the rain rate of POC to the seabed and the oxygen content of ambient bottom waters since these parameters affect the oxygen concentration and redox state of surface sediments (bottom row of Tab. B2). In contrast, benthic P fluxes are only affected by bottom water oxygen if the concentrations fall below the threshold value of 20 μM (WALLMANN, 2010), expressed as a Monod term in our model formulation (Tab. B2). In our model runs the minimum oxygen concentration (69 μM in North Pacific intermediate water at 21 ka, p. 2416 of our paper) clearly exceeds this threshold value. Hence, our calculated P burial rates do not reflect any changes in bottom water oxygen.

2b) There are alternative mechanistic formulations to describe the diagenetic processes and sediment models that describe the transport, dissolution and burial of biogenic particles as well as the transport of solutes within the active sediment layers (Heinze et al., 1999; Gehlen et al., 2006) and such models are applied to study ocean sediment interactions on the glacial interglacial time scales, including the possible role of changes in burial, a whole ocean nutrient increase, iron fertilization, or ocean circulation (e.g., (Brovkin et al., 2012; Brovkin et al., 2007; Matsumoto et al., 2014; Menviel et al., 2012; Roth et al., 2014; Tschumi et al., 2011; Lambert et al., 2015) and other).

The findings of this study should be compared to the findings of these and similar earlier studies that discussed burial-nutrient-pCO₂ feedbacks and glacial-interglacial CO₂ variations

We will compare our model results with the outcomes of other earth system models considering sedimentary processes (GEHLEN et al., 2006; HEINZE et al., 1999). However, comparison is difficult since none of these previous models has ever simulated the effects of sea-level change on the burial of POC and P and on benthic denitrification at continental margins.

2c) I am confused about the role of POM weathering. On page 2424, line 20 it is stated that POC weathering increases due to expansion of shelves and the increase appears to be displayed in Figure 3h that shows global rate of POC weathering calculated from Figure 3h, I estimate that this additional weathering leads to an input of about 2000 – 3000 GtC and presumably similar amounts of P and N. Is this realistic? This amount corresponds roughly to the carbon stored in today's soils.

Our standard model run suggests that the shelf (0 – 100m water depths) trapped a total of 4650 Gt POC over the last glacial cycle (130 – 0 ka) while 7870 Gt POC accumulated on the continental slope (100 – 2000m water depth). The POC weathering rate displayed in Fig. 3h is the sum of POC weathering on the exposed shelf and in the continental hinterland (WALLMANN, 2014). Accordingly, shelf weathering released a total of 1940 GtC over the last 130 ka. This is less than 50 % of the POC accumulating on the shelf over the last glacial cycle and consistent with models simulating POC weathering in exposed sedimentary rocks (BOLTON et al., 2006; WALLMANN, 2014). Thus we regard our weathering rates as realistic. Note that the POC burial rates applied in our model (Tab. B1) are conservative, that is lower than most previous estimates (BURDIGE, 2007; DUNNE et al., 2007). The standing stock of POC in marine sediments exceeds the global stock in soils since POC is buried more efficiently in margin sediments (Tab. B1) than in most terrestrial soils. POC burial in margin sediments is promoted by high sedimentation rates and the lack of oxygen in these water-saturated deposits.

2c) continued: I note that the model does not account for the growth of plants on exposed shelves that would supply such a flux by photosynthesis. How does an increase in the amount of POC weathering during glacial times compare with land area covered by ice and the generally smaller productivity and carbon pools during glacial times compared to interglacials?

The latest assessment of terrestrial POC stocks concluded that the modern total stock is close to the LGM stock since the storage in permafrost and below ice largely compensated for the glacial decline in plant and soil carbon (BROVKIN and GANOPOLSKI, 2015). Thus, the overall change in the terrestrial POC stock is probably small. Changes in terrestrial carbon stocks are probably outpaced by much larger changes in POC weathering (Fig. 3h) and POC burial (Fig. 6e) on the continental shelf.

d) page 2413, line 2 it is stated: "Neglecting the glacial increase in the weathering of P-bearing solids is raising the LGM pCO₂ value by 50 ppmv." In other words, about 50 ppm of the glacial-interglacial CO₂ difference are attributable to an increase in P weathering. What is the evidence for this increase?

Most of the P released during chemical weathering originates from apatite, a mineral equally occurring in all rock types (sedimentary, magmatic and metamorphic). Thus, we assume that the P weathering rate is proportional to the total weathering rate, that is the sum of carbonate, POC and silicate weathering (WALLMANN, 2014). During the glacial, total weathering increased due to the weathering of exposed shelf CaCO₃ and POC (MUNHOVEN, 2002; WALLMANN, 2014). This rise led to the increase in P weathering simulated in the model. However, our sensitivity tests show that the overall pCO₂ change induced by chemical weathering of silicate, POC, CaCO₃ and P is small because the glacial CO₂ draw-down induced by P and CaCO₃ weathering was compensated by the CO₂ release induced by POC weathering (Tab. 1, section 3.1).

3) The burial mechanism applied leads to a large increase in atm. CO₂ over the Holocene. How realistic is this given that the model does not include changes in terrestrial carbon storage and implied carbonate compensation and transient changes in the lysocline (Broecker et al., 2001)?

We will acknowledge in our section on model limitations that the Holocene pCO₂ record is probably affected by changes in terrestrial POC pools that are not yet considered in our model.

4) It would be useful to clearly outline which data are used for model tuning and which data are used for the a posteriori evaluation of results.

We will explain our database more clearly in the revised paper.

5) Given the cost-efficiency of a 24-box model I miss a comprehensive variation of model parameters to assess how uncertainties in model parameters affect results

The model includes about 100 parameter values. Thus, it is not possible to perform a comprehensive variation of model parameters and systematic sensitivity test. However, we switched on and off a number of key parameters to study the response of the model system (Tab. 1).

6) Figure 8, 9, and 11 should be changed. The contouring suggests higher model resolution than provided by a 24 box model and is misleading. It would be more appropriate to show the colors on the model grid given in Figure 2 without any interpolation.

We will add grid points to the figures. Moreover, we will add a table to Appendix C listing all LGM tracer concentrations for all boxes to augment the PRE data listed in Tabs. A1 and A2.

References

- Anderson, R. F., Barker, S., Fleisher, M., Gersonde, R., Goldstein, S. L., Kuhn, G., Mortyn, P. G., Pahnke, K., and Sachs, J. P., 2014. Biological response to millennial variability of dust and nutrient supply in the Subantarctic South Atlantic Ocean. *Phil. Trans. R. Soc. A* **372**.
- Bohlen, L., Dale, A., and Wallmann, K., 2012. Simple transfer functions for calculating benthic fixed nitrogen losses and C:N:P regeneration ratios in global biogeochemical models *Global Biogeochemical Cycles* **26**.
- Bolton, E. W., Berner, R. A., and Petsch, S. T., 2006. The weathering of sedimentary organic matter as a control on atmospheric O₂: II. Theoretical modeling. *American Journal of Science* **306**, 575-615.
- Brovkin, V. and Ganopolski, A., 2015. The role of the terrestrial biosphere in CLIMBER-2 simulations of the last 4 glacial CO₂ cycles. *Nova Acta Leopoldina NF* **121**, 43-47.
- Burdige, D. J., 2007. Preservation of organic matter in marine sediments: Controls, mechanisms, and an imbalance in sediment organic carbon budgets? *Chem. Rev.* **107**, 467-485.
- Dunne, J. P., Sarmiento, J. L., and Gnanadesikan, A., 2007. A synthesis of global particle export from the surface ocean and cycling through the ocean interior and on the seafloor. *Global Biogeochem. Cycles* **21**, doi:10.1029/2006GB002907.

- Gehlen, M., Bopp, L., Emprin, N., Aumont, O., Heinze, C., and Ragueneau, O., 2006. Reconciling surface ocean productivity, export fluxes and sediment composition in a global biogeochemical ocean model. *Biogeosciences* **3**, 521-537.
- Heinze, C., Maier-Reimer, E., Winguth, A. M. E., and Archer, D., 1999. A global oceanic sediment model for long-term climate studies. *Global Biogeochemical Cycles* **13**, 221-250.
- Jaccard, S. L. and Galbraith, E. D., 2012. Large climate-driven changes of oceanic oxygen concentrations during the last deglaciation. *Nature Geoscience* **5**, 151-156.
- Köhler, P., Fischer, H., Munhoven, G., and Zeebe, R. E., 2005. Quantitative interpretation of atmospheric carbon records over the last glacial termination. *Global Biogeochemical Cycles* **19**.
- Munhoven, G., 2002. Glacial-interglacial changes of continental weathering: estimates of the related CO₂ and HCO₃⁻ flux variations and their uncertainties. *Global and Planetary Change* **33**, 155-176.
- Schmitt, J., Schneider, R., Elsig, J., Leuenberger, D., Laurantou, A., Chappellaz, J., Köhler, P., Joos, F., Stocker, T. F., Leuenberger, M., and Fischer, H., 2012. Carbon isotope constraints on the deglacial CO₂ rise from ice cores. *Science* **336**, 711-714.
- Tschumi, T., Joos, F., Gehlen, M., and Heinze, C., 2011. Deep ocean ventilation, carbon isotopes, marine sedimentation and the deglacial CO₂ rise. *Clim. Past* **7**, 771–800.
- Wallmann, K., 2010. Phosphorus imbalance in the global ocean? *Global Biogeochemical Cycles* **24**, doi:10.1029/2009GB003643.
- Wallmann, K., 2014. Is late Quaternary climate change governed by self-sustained oscillations in atmospheric CO₂? *Geochimica et Cosmochimica Acta* **132**, 413-439.