Reply to the referee comments by Victor Brovkin

First of all we would like to thank the referee for his very constructive and valid comments. He raises a number of important questions and issues that we address as outlined below:

My main critical remark is about motivation of this study.

Our CPD paper was motivated by the following questions: i) Which empirical data may help to verify/falsify the hypothesis of (WALLMANN, 2014)? ii) To which degree do global ^{14}C data sets assembled by (SARNTHEIN et al., 2015; SARNTHEIN et al., 2013) actually reflect our concepts and ideas on glacial and deglacial ocean circulation and carbon cycling? Iii) Do marine ^{14}C data really form a quantitative proxy of DIC in the glacial deep ocean as proposed by (SARNTHEIN et al., 2013)? To address these questions we are comparing data with model-based tracer distributions.

Why to invest huge intellectual effort into making box model more realistic? Since circulation between model boxes is prescribed by hand, it cannot be more realistic that in a model with explicit ocean dynamics. In my view, the usefulness of box models is limited to conceptual studies.

The modern water fluxes applied to our box model are based on a dynamically consistent circulation field. As explained in Appendix A of our paper, these fluxes were modified to obtain tracer distributions that are consistent with observations in the pre-industrial modern ocean (Fig. R1). The glacial and deglacial changes in ocean circulation were not derived from ocean models but from $\delta^{13}C$ data (Tab. A5) and additional geochemical observations (Fig. R2). Indeed, models with explicit ocean dynamics are superior to any kind of physically unconstrained box model if they generate results that are consistent with observations. However, the physically-constrained models that we are aware of (e.g. (BROVKIN et al., 2012; GANOPOLSKI et al., 2010) are not yet able to reproduce as many tracer and proxy data as our box model. Our paper shows that shelf and sea-level effects help to explain a wide range of findings and we think that the outputs of physically better constrained models may improve in case these effects are included in the model architecture. It is not our intention to promote box modeling *per se* as the method-of-choice. Rather we hope that the concepts and ideas advanced in our paper may stimulate the community and help to further enhance cutting-edge earth system models with explicit ocean dynamics.

What really new can we learn from the improved model study? The novelty should be more clearly presented in conclusions and in the abstract.

For a first time we show model results that are consistent with both the atmospheric pCO_2 record (Figs. 5 and 7) and data on past distribution changes of dissolved oxygen, carbonate, and radiocarbon in the glacial ocean (Figs. 8, 9, 11 and Tabs. C1 and C3). This conformity supports the hypothesis of Wallmann (2014) that the glacial sea-level drop induced a decline in atmospheric pCO_2 and a rise in nutrients, DIC, and alkalinity of the glacial ocean. Also, we first show that the slope of DIC versus radiocarbon observed in the modern deep ocean (SARNTHEIN et al., 2013) is maintained in the glacial ocean (Fig. 13). However, a glacial shift in the intercept now complicates the use of ^{14}C as DIC proxy. The shelf hypothesis was originally developed to explain the deglacial rise in atmospheric pCO_2 (BROECKER, 1982). In contrast, our model analysis now reveals that shelf and sea-level effects are not responsible for this rapid rise but account for a major portion of the slow glacial decline of atmospheric pCO_2 (Figs. 5 and 7). These new results will be presented more clearly in the conclusions of our revised paper.

Another problem with discussion of the box model results is that it is difficult to avoid circular logic... The authors should clearly indicate which results are direct implications of their assumptions and which ones are novel, non-trivial consequences of interactions between biogeochemical components.

To avoid circular reasoning dissolved oxygen, carbonate, and radiocarbon distributions calculated for the glacial ocean were not used to parameterize our model. These distribution patterns and the atmospheric pCO_2 values calculated in the model are non-trivial consequences of interactions between biogeochemical components.

Regarding the POC hypothesis, there are still a few conceptual questions, which require more detailed discussion in introduction. Firstly, why terrestrial plants cannot utilize sediment nutrients after the shelf exposure? Tree roots could be many meters deep.

Plants may indeed utilize sediment nutrients after shelf exposure. This process is not considered in the model since our model does not resolve terrestrial processes. We simply assume that the standing stock of plant and soil POC on the continents is maintained over a glacial cycle. Indeed the uptake of shelf phosphorus by land plants may modulate the weathering flux of phosphorus from exposed shelf areas. However, sensitivity tests show that the decline in depositional area at continental margins has a larger effect on the phosphate content of the glacial ocean than changes in the weathering flux (Tab. 1). The inclusion of plant uptake would thus have only a minor effect on our model results.

Secondly, increased nutrient inventory and utilization reduces the oxygen content. How good is the box model in reproducing oxygen minimum zones at present? Since ocean boxes are huge, do they represent the oxygen limitation in a plausible way?

Our model predicts a general oxygen decline at >2000 m water depth in the glacial ocean (Fig. 8), consistent with observations (JACCARD and GALBRAITH, 2012). However, oxygen minimum zones (OMZs) are not resolved by our model since the entire Indo-Pacific intermediate water at 100 -2000 m water depth is pooled in a single ocean box. The glacial oxygen decline simulated in the model has no significant effect on other model parameters since the oxygen level stays above the threshold values for pelagic denitrification (4 µM) and enhanced phosphate release from sediments (20 µM) in all ocean boxes. Instead, we prescribe a constant rate of pelagic denitrification in our model since we are not able to resolved OMZs. Rates of benthic denitrification are calculated dynamically considering changes in bottom water nitrate and oxygen, POC rain rates to the seabed and depositional areas (BOHLEN et al., 2012). These benthic rates and the burial fluxes of phosphorus and POC may be only moderately affected by the lack of OMZs in our model since the area where OMZs impinge the seafloor amounts to only 1 % of the global seafloor area (BOHLEN et al., 2012).

P 2406, l.15-17. Who are "they"? Sea-level changes? Does weathering add depleted 13C or 14C to the ocean? Changes in 14C are rather a signal of overturning and water masses changes than a result of sea-level changes.

"They" refers to glacial sea-level low-stands. Weathering adds ^{14}C-depleted DIC to the ocean. Our sensitivity tests show that marine ^{14}C values are controlled not only by overturning changes but also by changes in the radiocarbon production rate in the atmosphere, the addition of old carbon via weathering, and the burial of young carbon at the seafloor (section 3.4, Fig. 12, and Appendix D). Sea-level change affects the rates of carbon burial and weathering and thereby marine ^{14}C values.

l.19: What does "reduced deep ocean dynamics" mean: slower overturning? Do we have a proxy for it? The Atlantic overturning was shoaled, but was it slower? How do we know that "transit time" was longer – in the whole ocean? Is it the model outcome or is it the model assumptions? It is written like it is a fact – but is it a model truth or a data truth? If it is an artifact of the model setup, should it be highlighted in the abstract?

In our model, vertical water fluxes across the 2000 m depth horizon indeed are forced to decline under glacial conditions from a modern value of 45 Sv to a LGM value of 31 Sv (p. 2414, l. 27-28 of our paper). This corresponds to an increase in the average residence time of water in the deep ocean (>2000 m) from 470 years in the modern ocean to 680 years during the LGM where the residence time is calculated as ratio of the deep ocean volume (6.65 x 10^{17} m^3 at >2000m) and the global vertical water fluxes across 2000 m. As outlined above, changes in the circulation field were defined using the glacial $\delta^{13}C$ record. The model-based radiocarbon distribution in the glacial deep ocean is broadly consistent with observations (Tab. C3). Thus, the decrease in deep ocean ventilation is not a model artifact but consistent with data. The term "transit time" is based on empirical apparent ^{14}C ventilation ages, but may be somewhat misleading. It will be removed from the abstract.

l. 28: terrestrial biosphere is not accounted for in the study. Can we consider the Holocene dynamics without accounting for the biosphere regrowth and peat accumulation on land?

For many decades it has been assumed that the modern terrestrial carbon pool exceeds the glacial pool by hundreds of Gt because of biosphere regrowth after the glacial termination. This concept was initially developed to explain the low $\delta^{13}C$ values in glacial seawater (SHACKLETON, 1977). However, the latest assessment of terrestrial carbon pools published by the referee indicates that the sum of the modern stocks does not exceed the LGM stock (BROVKIN and GANOPOLSKI, 2015). Moreover, this new view on terrestrial carbon cycling suggests a deglacial decline in total carbon stocks since the carbon release from high latitude areas (melting permafrost soils and soils exposed by the retreat of glacial ice sheets) exceeded the carbon uptake by biosphere regrowth and peat accumulation. Our model explains deglacial and Holocene pCO_2 dynamics and the low glacial $\delta^{13}C$ values by marine processes and sea-level change. We also believe that terrestrial processes may have played a role as well, but we would like to wait with an up-date of our model until a new consensus emerges among the scientific community working on terrestrial carbon cycling.

P. 2407, l.1-110: How could a decline in iron deposition lead to an increase in atmospheric CO2 by 12-13 ppm in few decades? This corresponds to a source of about 30-40 GtC from the ocean. What can cause a sustainable flux of 1-2 GtC/yr from the ocean?

As shown in Fig. 7, about equal portions of the extreme carbon flux may have been triggered by changes in circulation and iron deposition. We admit that iron fertilization in part may have been overweight. However, the residence time of iron in the global ocean amounts to only a few hundred years whereas other nutrients have residence times of several thousand years. Changes in the iron input are thus more likely to cause rapid pCO_2 change than any other marine biogeochemical process. However, the marine iron cycle is poorly understood (DALE et al., 2015). Thus, it is unclear to what extent the deglacial pCO_2 rise may be caused by a decline in dust deposition (iron delivery). In our model we are able to reproduce the deglacial pCO_2 record by a combination of both rapid ventilation pulses in the Southern Ocean and the North Pacific and a decline in nutrient utilization (dust deposition) in the Southern Ocean (Fig. 7). However, further processes such as the deglacial melting of permafrost may also have contributed to the pCO_2 rise because the amplitudes of changes in nutrient utilization and stratification applied in the model are poorly constrained by independent data.

P2409, l.24-29 constraining glacial water fluxes based on d13C: this process should be explained in more details. Was it optimization of parameters?

Water fluxes were varied until the PRE-LGM differences in $\delta^{13}C$ generated by the model were consistent with the PRE-LGM differences recorded in foraminifera (Tab. A5).

P2413: C:P ratio in POM is roughly 100:1, while it is about 50:1 for soils and 1000:1 for wood biomass. Trees on exposed tropical shelves could store much more C than marine sediments, and the net effect on atmospheric CO_2 would be rather neutral

Modern continental margins (shelf and rise) accumulate sedimentary POC at a rate of about 100 - 200 Gt kyr^{-1} (BURDIGE, 2007; DUNNE et al., 2007; HEDGES and KEIL, 1995; WALLMANN et al., 2012). This enormous flux is induced by the high marine productivity of the region and the rapid accumulation of sediments facilitating the burial of marine POC. Trees growing on the shelf would have to accumulate a biomass in the order of 10 000 Gt C to maintain this high carbon flux over the glacial period (ca. 80 kyr), an unlikely scenario since the present global stock of carbon in vegetation is just about 500 GtC. This simple back-of-the-envelop-calculation suggests that the rate of POC accumulation in modern margin sediments is at least an order of magnitude higher than the glacial accumulation rate of POC in trees on the exposed shelf.

Figure 2: numbers are not readable

Figure will be changed accordingly

Fig. 3h, POC weathering: is river POC flux included? If so, is this POC originated from shelves or from internal continental area? Exposed shelf area goes to 0, but POC flux from exposed area is not 0: please explain

Indeed, riverine POC flux is ignored since we do not include soil and plant POC in our model. Our POC weathering flux has two components: i) weathering of POC in exposed shelf sediments and ii) weathering of fossil POC in continental hinterland (WALLMANN, 2014). The latter flux does not go to zero when the exposed shelf area is zero. Both components produce atmospheric CO_2 depleted in ^{13}C and ^{14}C and contribute significantly to the evolution of marine $\delta^{13}C$ values on geological time scales (BERNER, 2004; WALLMANN, 2001; WALLMANN and ALOISI, 2012).

Figure 7: What mechanisms are responsible for the pCO_2 drawdown from 10 to 8 ka? A usual interpretation is that this CO_2 drop is an effect of terrestrial carbon uptake.

The modeled CO_2 flux from the ocean into the atmosphere reaches a minimum at 9 ka. The minimum is accompanied by a maximum in the ocean-to-atmosphere O_2 flux. These changes reflect the recovery of the ocean system from the antecedent ventilation pulse in the Southern Ocean centered at 11.5 ka (Fig. 4d). According to our model, this event removed CO_2 from the ocean interior, enhanced the O_2 content of the deep ocean, and diminished the vertical DIC and O_2 gradients. Hence, the pCO_2 drawdown from 10 to 8 ka is caused by the restoration of the vertical DIC gradient in the model ocean after a massive ventilation pulse in the Southern Ocean.

Figure 10a, black line: the model shows a decrease in carbonate ion concentration during deglaciation. This would lead to a $CaCO_3$ dissolution spike during deglaciation, opposite to what was observed.

Indeed, carbonate ion concentrations regulate the dissolution of pelagic carbonates but are likewise controlled by the rates of carbonate dissolution and burial. Moreover, carbonate burial is not only affected by carbonate dissolution but also by the production and export of carbonate via the biological pump. According to our model, export production reached a deglacial maximum at 16.5 ka (Fig. 6d) which was responsible for the peak in global carbonate burial at that time (Fig. 6f) and the deglacial CO_3^{2-} minimum (Fig. 10a).

Figure 8: Please add a plot of difference LGM to PRE, otherwise it is difficult to see how exactly the P and O2 pools are re-distributed at LGM. Also, add a plot of observations for PRE as averaged on the model resolution from the data. Figure 9: Please add a plot of difference LGM to PRE. Figure 11: plot a difference LGM to PRE. Figure 13: please add arrows of shifts from STD to STD LGM. Is there added value of showing CC and CC-CN LGM versions? Figure 14: a nice conceptual plot, but arrows are very selective. Why there is no direct effect of insolation on climate (temperature) and circulation (via SST/precip pattern)? Also, sea level changes have a direct effect on circulation eg through the Bering Straight.

We will change these plots accordingly.

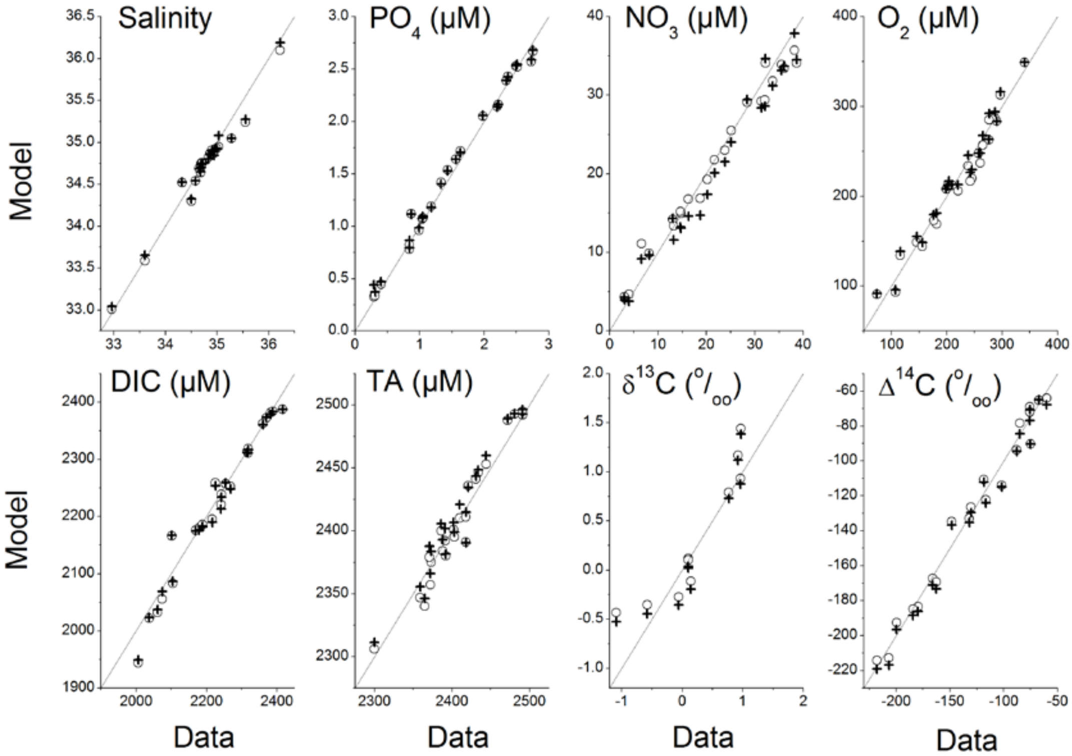

Fig. R1. Tracer concentrations in ocean boxes: Model versus data. Open circles indicate concentrations applied as initial values at 130 ka (Tab. A1 and A2). Crosses are concentrations obtained at the end of simulation STD at 0 ka after the completion of a full glacial cycle. Lines indicate the 1:1 relationship, e.g. the best fit to the data.

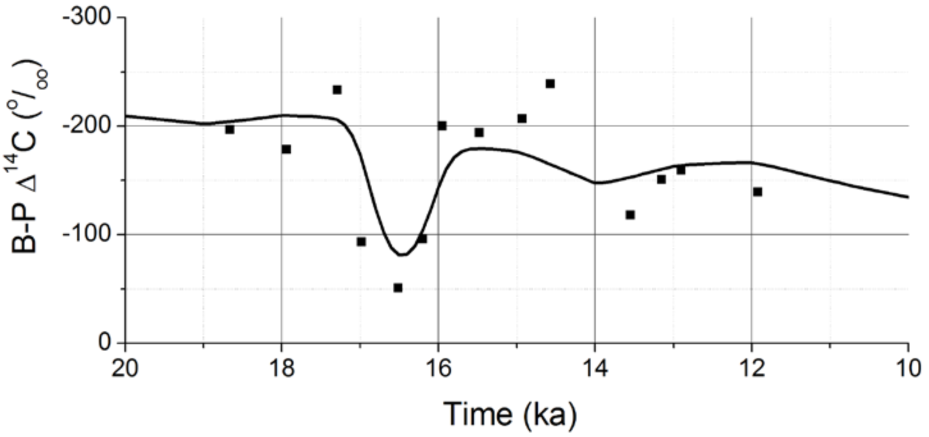

Fig. R2. Deglacial benthic – pelagic radiocarbon record in the North Pacific: Model (line) versus data (squares). The Δ^{14}C-DIC difference between deep water and surface water boxes in the North Pacific is compared to data (Δ^{14}C difference between benthic and pelagic foraminifera, B-P Δ^{14}C) from core MD02-2489 taken at the Alaskan Margin at 3.6 km water depth (RAE et al., 2014). The vertical mixing across 100 m and 2000 m water depth was enhanced by a factor of 10 at 16.5 ka (Fig. 4d of our CPD paper) to reproduce the radiocarbon data.

References

Berner, R. A., 2004. *The Phanerozoic Carbon Cycle: CO_2 and O_2*. Oxford University Press, Oxford.

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 Biochemical Cycles* **26**.

Broecker, W. S., 1982. Glacial to interglacial changes in ocean chemistry. *Prog. Oceanogr.* **11**, 151-197.

Brovkin, V. and Ganopolski, A., 2015. The role of the terrestrial biosphere in CLIMBER-2 simulations of the last 4 glacial CO_2 cycles. *Nova Acta Leopoldina NF* **121**, 43-47.

Brovkin, V., Ganopolski, A., Archer, D., and Munhoven, G., 2012. Glacial CO_2 cycle as a succession of key physical and biogeochemical processes. *Clim. Past* **8**, 251–264.

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.

Dale, A. W., Nickelsen, L., F. Scholz, Hensen, C., Oschlies, A., and Wallmann, K., 2015. A revised global estimate of dissolved iron fluxes from marine sediments. *Global Biogeochem. Cycles* **29**.

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.

Ganopolski, A., Calov, R., and Claussen, M., 2010. Simulation of the last glacial cycle with a coupled climate ice-sheet model of intermediate complexity. *Clim. Past* **6**, 229–244.

Hedges, J. I. and Keil, R. G., 1995. Sedimentary organic matter preservation: an assessment and speculative synthesis. *Marine Chemistry* **49**, 81-115.

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.

Rae, J. W. B., Sarnthein, M., Foster, G. L., Ridgwell, A., Grootes, P. M., and Elliott., T., 2014. Deep water formation in the North Pacific and deglacial CO_2 rise. *Paleoceanography* **29**, 645-667.

Sarnthein, M., Balmer, S., Grootes, P. M., and Mudelsee, M., 2015. Planktic and benthic ^{14}C reservoir ages for three ocean basins, calibrated by a suite of ^{14}C plateaus in the glacial-to-deglacial Suigetsu atmospheric ^{14}C record. *Radiocarbon* **57**, 129-151.

Sarnthein, M., Schneider, B., and Grootes, P. M., 2013. Peak glacial ^{14}C ventilation ages suggest major draw-down of carbon into the abyssal ocean. *Clim. Past* **9**, 2595-2614.

Shackleton, N. J., 1977. Carbon-13 in Uvigerina: Tropical rainforest history in the equatorial Pacific carbonate dissolution cycles. In: Andersen, N. R. and Malahoff, A. Eds.), *The Fate of Fossil Fuel in the Oceans*. Plenum, New York.

Wallmann, K., 2001. Controls on Cretaceous and Cenozoic evolution of seawater composition, atmospheric CO_2 and climate. *Geochimica et Cosmochimica Acta* **65**, 3005-3025.

Wallmann, K., 2014. Is late Quaternary climate change governed by self-sustained oscillations in atmospheric CO_2? *Geochimica et Cosmochimica Acta* **132**, 413-439.

Wallmann, K. and Aloisi, G., 2012. The Global Carbon Cycle: Geological Processes (Chapter 3). In: Knoll, A. H., Canfield, D. E., and Konhauser., K. O. Eds.), *Fundamentals of Geobiology, First Edition*. Blackwell Publishing Ltd.

Wallmann, K., Pinero, E., Burwicz, E., Haeckel, M., Hensen, C., Dale, A., and Ruepke, L., 2012. The global inventory of methane hydrate in marine sediments: A theoretical approach. *Energies* **5**, 2449-2498.