Dear Editor and Reviewers,

Thank you for your thoughtful comments and the time and effort you put into our paper "Disentangling the effect of ocean temperatures and isotopic content on the oxygen – isotope signals in the North Atlantic Ocean during Heinrich Event 1 using a global climate model".

Before addressing all your comments point by point, we want to clarify a few points that were made by (almost) all of you.

1) We added the description how the $\delta^{18}O_{calcite}$ is computed.

2) We added a paragraph where we compare the hosing experiment – LS 0.2Sv - conducted by Roche et al. (2014) to our ICE-300 experiment (lines 57-267):

We want to address the different impact of explicitly computing icebergs compared to mimicking icebergs using hosing experiments. Therefore, we compare our ICE-300 experiment to the results of the 0.2 Sv experiment presented by Roche et al. (2014). This can easily be done since we share the same initial conditions and the calving sites in the present paper correspond to the Labrador Sea forcing experiment of Roche et al. (2014).

Generating icebergs at the margin of the Laurentide ice sheet causes a slightly stronger decrease of the AMOC than applying a freshwater flux of 0.2 Sv in the Labrador Sea for 300. Yet, the recovery of the AMOC differs between the two experiments. In the ICE-300 experiment the AMOC picks up immediately at the end of the iceberg forcing (Fig. 1b, green line), whereas the results of Roche et al. (2014, their figure 2b, green line) display a gradual recovery of the AMOC starting only 200 years after the forcing stopped.

3) We have not added any more cores because we only chose four marine sediment cores to begin with because the four marine sediment cores are very well-dated and we have both planktonic $\delta^{18}O$ records and SST reconstruction and / or IRD counts. Moreover, they have a high temporal resolution, a time step of ~200 years.

4) We have completely revised the discussion on the data – model comparison and it now states (lines 360-467):

We compared our model results to planktonic $\delta^{18}O_{calcite}$ data from four marine sediment cores distributed over a wide spatial area of the North Atlantic and Greenland – Iceland – Norwegian (GIN) Seas (Fig. 7f). Before looking in detail at the four cores to investigate whether or not the simulated patterns can be confirmed by the data, several important remarks need to be made. First, while the sea level rise due to the released icebergs during HS1 is accounted for, we do not simulate the background sea level rise starting at 19 ka BP after the onset of the LGM (Lambeck and Chappell, 2001; Lambeck et al., 2014). Therefore, the changing background sea level causes lighter $\delta^{18}O_{calcite}$ values in the proxy data, but does not affect the simulated values. Second, in the model we investigate the impact of sea surface temperature and $\delta^{18}O_{seawater}$ on $\delta^{18}O_{calcite}$ only and ignore other factors that impact the foraminifera, such as seasonality and depth habitat, which influence the $\delta^{18}O_{calcite}$ recorded in foraminifera. Third, since the oceanic model has a resolution of 3°x3° in longitude and latitude, it is unlikely that the modeled $\delta^{18}O_{calcite}$ signal within one grid cell fits perfectly to the recorded $\delta^{18}O_{calcite}$ at one specific site. Therefore, our aim is to compare the

modeled patterns to data, rather than comparing the values precisely computed at the corresponding grid cell to data. Further, it is important to note that the regions we selected to analyze the pattern of simulated $\delta^{18}O_{calcite}$ were not chosen to comprise the marine sediment core sites, instead they are of two types: (i) areas where simulated changes in sea surface temperature (central North Atlantic) or $\delta^{18}O_{seawater}$ (Bafin Bay, Nordic Seas) are maximum, and (ii) areas where neither the change in SST nor in $\delta^{18}O_{seawater}$ dominates (subtropical North Atlantic, northeast North Atlantic, Fig. 7f). Finally, our experiments were performed under fixed LGM equilibrium conditions, whereas the paleoclimatic data also displays changes in the boundary conditions such as changing orbital parameters.

Despite all the aforementioned difficulties, we attempt a first step in trying to better understand the $\delta^{18}O_{calcite}$ pattern recorded in marine sediment cores and its evolution during a Heinrich event by using the simulated $\delta^{18}O_{calcite}$.

We chose well-dated marine sediment cores on which there exist planktonic $\delta^{18}O$ records and SST reconstructions and / or IRD counts at relatively high resolution (time step ~200 y on average for $\delta^{18}O$ and SST records). Two cores (NA87-22, CH69-K09) are situated within the Ruddiman Belt, one core (ENAM93-21) is located in the Norwegian Sea and the fourth core (KNR31 GPC-5) is located far south of Greenland (33°N, Fig. 7f). The four cores are representative of different environments. The cores situated within the Ruddiman belt (NA87-22, CH69-K09) are characterized by high iceberg melt fluxes. Core ENAM93-21 is located in the Norwegian Sea, away from the Ruddiman belt, and therefore is hardly reached by the icebergs coming from the Laurentide ice sheet. Finally, marine sediment core KNR31-GPC5 is representative of areas that are affected by the advection of meltwater rather than by icebergs directly since it is situated far south of the Ruddiman belt.

In cores NA87-22 and CH69-K09 the $\delta^{18}O_{calcite}$ was measured on two planktonic foraminifer species: *Globigerina bulloides* and *Neogloboquadrina pachyderma* (sinistral coiling). *N. pachyderma* s. is expected to prefer a deeper living habitat than *G. bulloides* and depending on the stratification of the water column the two species display similar (well – mixed) or different signals (stratified water column; Kohlfeld et al., 1996; Simstich et al., 2003). SST reconstructions derived from planktonic foraminifer counts and IRD data of core NA87-22 and CH89-K09 are also displayed for comparison with simulated SSTs and IMF. In core ENAM93-21 and KNR31-GPC5, $\delta^{18}O_{calcite}$ was measured on only one planktonic foraminifer species. Additional IRD data for core ENAM93-21 is also displayed (Fig. 8).

The $\delta^{18}O_{calcite}$ signal obtained from *N. pachyderma* s. in core ENAM93-21 displays a gradual decrease of 2.5 ‰ over the HS1 interval (Fig. 8a). The IRD values decrease and then seem to stabilize over HS1. This could be due to the cold conditions at the core location, which prevent icebergs (originating from the Fennoscandian ice sheet) from melting. The ENAM93-21 core is located within the Northeast North Atlantic region defined in Fig. 7f. Yet, the pattern of $\delta^{18}O_{calcite}$ as recorded by *N. pachyderma* s. resembles the modeled $\delta^{18}O_{calcite}$ simulated in the Nordic Sea region (Fig. 7c, 8a), which is north of ENAM93-21. The 2.5‰ decrease in the measured $\delta^{18}O_{calcite}$ is indeed seen in the ICE-600 and ICE-900 experiments, the ICE-300 displaying a decrease of only 1‰ (Fig. 7c). Moreover, the IMF values computed by *i*LOVECLIM in the Nordic Seas region display low values similar to the proxy data.

The IRD pattern from core NA87-22 (Fig. 8b) exhibits an abrupt increase in iceberg melting at ~17.5 ka, which is accompanied by a decrease in $\delta^{18}O_{calcite}$. NA87-22 $\delta^{18}O_{calcite}$ obtained from *G. bulloides* decreases suddenly (-2 ‰) with the arrival of icebergs and stays low throughout HS1, whereas the $\delta^{18}O_{calcite}$ obtained from *N. pachyderma* s. displays a gradual decrease (-1 ‰) until the end of HS1 (Fig. 8b). This non-uniform response in *G. bulloides* and *N. pachyderma* s. could be due to the deeper living habitat of *N. pachyderma* s. in a stratified water column. This core is situated in the northeast North Atlantic region, yet, the $\delta^{18}O_{calcite}$ as obtained from the *G. bulloides* resembles more the surface $\delta^{18}O_{calcite}$ pattern computed in the Baffin Bay region (Fig. 7a, 7f). In the Baffin Bay region the model displays an immediate decrease in $\delta^{18}O_{calcite}$ with the intrusion of iceberg melt water. There the ICE-300 experiment displays a magnitude of change similar to that seen in NA87-22. In the northeast North Atlantic region iLOVECLIM simulates an increase followed by a stable phase in $\delta^{18}O_{calcite}$ because the SST strongly decreases (-6°C), thereby counteracting the decrease in $\delta^{18}O_{seawater.}$ In contrast, the SST of NA87-22 displays only a moderate cooling (-1°C), which is in the same range as the cooling simulated in Baffin Bay. iLOVECLIM overestimates the cooling in the northeast North Atlantic region in comparison with paleoclimatic data because the simulated sea ice edge is too far North (not shown).

In core CH69-K09 both species display a strong 2‰ decrease in $\delta^{18}O_{calcite}$ with the arrival of icebergs at ~16 ka (Fig. 8c). Remarkably, the reconstructed annual mean SST decreases much earlier (at ~17.5 ka) without affecting the $\delta^{18}O_{calcite}$ (Fig. 8c). The lack of response in the recorded $\delta^{18}O_{calcite}$ might be due to a decrease in $\delta^{18}O_{seawater}$ resulting from the arrival of cold meltwater at 17.5 ka, noting that the lack of simultaneous increase in IRD indicates that this cold meltwater was not accompanied by melting icebergs. Another possibility is that the isotopic data reflects properties of the subsurface, since the habitat depth of *G. bulloides* and *N. pachyderma* s. is often the pycnocline, whereas the reconstructed SST is derived from statistical relations between planktonic foraminifera abundances and Atlas SST, and thus corresponds to the surface conditions.

CH69-K09 is situated close to the central North Atlantic region as displayed in Fig. 7f. Yet, in the central North Atlantic the modeled $\delta^{18}O_{calcite}$ displays an increase during the first 300 years of iceberg discharge due the strong decrease in SST whereas observed SST already decreased before the iceberg discharge reaches the site causing a $\delta^{18}O_{calcite}$ pattern quite different from the one simulated by iLOVECLIM. Instead, the recorded *G. bulloides* and *N. pachyderma* s. $\delta^{18}O_{calcite}$ pattern resembles the one of the Baffin Bay region.

In core KNR31-GPC5 *G. ruber* $\delta^{18}O_{calcite}$ displays a short-lived increase of about 0.5 ‰ at ~18 ka and then a decrease of 1.5 ‰ followed by another increase of ~0.7 ‰ (Fig. 8d). This pattern and magnitude is simulated by the ICE-600 and ICE-900 experiment in the subtropical North Atlantic.

It is important to note that the $\delta^{18}O_{calcite}$ as recorded in ENAM93-21, NA87-22 and CH69-K09 displays lighter values at the end of HS1 than at the beginning. This overall decrease in $\delta^{18}O_{calcite}$ is due to the background freshwater flux coming from other ice sheets, which is not incorporated in our experiment set-up.

Overall, the comparison of four marine sediment cores with our model results shows that the $\delta^{18}O_{calcite}$ patterns found in paleoclimatic data are also seen in simulated $\delta^{18}O_{calcite}$ curves, but not

always at the same location. The rate of change of the recorded $\delta^{18}O_{calcite}$ is best simulated/reproduced by the ICE-300 and ICE-600 experiments, yet only the ICE-300 model run displays an SST increase at the end of the iceberg discharge as observed in cores NA87-22 and CH66-K09 SST records.

It would be interesting to extend this investigation to more sediment cores, preferably also at sites closer to the calving locations. This modelling approach offers the possibility to analyze the timing of the Heinrich events at the different locations, its dependence on the evolution of sea surface temperatures and $\delta^{18}O_{seawater.}$ Moreover, it would be of great value to extend this research to other Heinrich events and to investigate if the patterns observed for HS1 are the same for all the Heinrich events, or if they vary and how they depend on the calving locations (e.g. Laurentide Ice Sheet, Fennoscandian Ice Sheet or Barents Ice Sheet). Even though the model used has been proven to simulate the observed $\delta^{18}O_{calcite}$ pattern satisfyingly at different time periods (Caley and Roche 2013; Roche et al., 2014), as well as the iceberg distribution under pre-industrial conditions (Jongma et al., 2009; Bügelmayer et al., 2015a), the coarse resolution of our ocean and atmospheric model components and corresponding parameterizations might influence the simulated $\delta^{18}O_{calcite}$ evolution. Therefore, it would be interesting to repeat this study with a higher resolved global climate model.

5) We have changed figures 5 – 10.
    a. Figures 5/6/7 are now as suggested by reviewer#3, that means that now ICE-300 / ICE-600 / ICE-900 are directly compared within one figure. Instead of 3 figures, we have now 2 (figure 5 displays the anomaly of the iceberg melt flux, the convection layer depth and the sea surface salinity of ICE-300, -600, -900 with the control set-up, figure 6 displays the sea surface temperature, $\delta^{18}O_{seawater}$ and $\delta^{18}O_{calcite}$);
    b. Figure 9 was re-done: all the experiments are within one panel with different colors and the former figure 8 is included (it is now figure 7)
    c. Figure 10 was improved by flipping the x-axis so that it now fits to figure 7