We thank the anonymous reviewer for the positive evaluation of our manuscript and for the helpful comments and suggestions. Below please find a point-to-point response to the comments.

GENERAL COMMENTS:

The results of the study are clearly novel and of great importance to our understanding of H-events and instabilities of the ice sheets surrounding the Atlantic Ocean during the last glacial. The model includes several assumptions which should be better discussed. In particular, the coupling scheme includes period-synchronous 1:10 coupling. Still it is not clear from reading the methods section what this means, and how this choice of coupling method might impact the results. Please elaborate on this.

We expanded the paragraph:

In all simulations used for the analysis, a configuration was used where atmosphere and ocean are coupled with a periodic-synchronous 1:10 coupling (Voss and Sausen, 1996; Mikolajewicz et al., 2007a). The motivation for this is that the atmosphere has no long-term memory (this resides in ocean and ice sheets), but consumes more than 90% of the CPU time of the coupled model system. Periodic-synchronous coupling means that the atmosphere model is only run for one out of ten years. This reduces the computational expense drastically and speeds up the simulations by a factor of three (wallclock time). After each fully coupled year, the ocean is forced for nine years with atmospheric fields, that are obtained by cycling through the previous five fully coupled years. These settings turned out to be a good compromise between having a sufficiently large archive (of atmospheric forcing) to adequately represent the inter-annual variability of the atmospheric forcing and minimizing the delay in the forcing from a large archive. A too small archive leads to a large model drift in the ocean-only phases and thus corrupts the climate of the coupled model, a large archive introduces a large delay. Whereas this coupling technique has only minor effect on the simulated climate response to long-term changes, this technique should not be used for short-term changes or the analysis of short-term variability. With the settings used her the periodic-synchronous coupling introduces a lag of up to 50 years in the atmosphere-ocean system, the average lag of about 25 years is less than 10% of the 300-year averaging window used throughout most of the analysis. Furthermore, most of the changes in the ice sheets driving the atmosphere-ocean system occur on a longer time scale, so the lag can be neglected in the analysis. Parallel to the climate model, the ice sheet model is run for ten years, so ice sheets and ocean are on the same time scale. In the analysis of atmosphere and ocean fields, only the fully coupled years are used. These are also used to obtain the fields for forcing the ice sheet model. Thus, when we speak of a 300 year mean of the ocean 2D fields, only 30 years of data with a 10-year spacing are used. This should yield comparable results to the full data as long as no signals with periods of exact multiples of 10 years are involved (aliasing).

Another key assumption in the model is that the freshwater input is associated with a negative heat input when ice enters the ocean. However, it is not clear from the manuscript how this impacts the results. More detail should be given to this point. How important is it for the response observed?

We only apply the effect to freshwater released by direct ice-ocean interactions (shelf melt, calving), not to surface runoff. The effect mainly consists of an increase in the ice cover in Hudson Bay. We still consider modeling this effect. superior to neglecting it. We provide a few comparison plots of freshwater hosing experiments with and without the latent heat flux as supplement Figure 2 and added

A sensitivity test with our model showed that the inclusion of the latent heat effect of ice discharge increases the ice cover and the cooling in the Labrador Sea, but has minimal consequences outside of this area (Supplement Fig. 2).

The authors choose to study H-events in a transient glacial climate simulation which makes the analysis challenging. It is not clear how this change in boundary conditions impacts the results, and it is not clear why the authors choose a composite of several transient runs in their results description. This should be better explained/discussed in the revised manuscript. Note, however, that including a transient glacial climate is also of interest as it could give clues as to how the H-events change with time. It would be of benefit to the study if such an assessment were to be included.

We added sentences in the experiments section and in the description of the composite analysis:

Experiments:

We chose these simulations as technically quasi-identical subset from various simulations that were performed when working on a model that is able to simulate the last deglaciation. As the simulations consumed considerable resources, we refrained from performing a dedicated ensemble, but made use of the available data.

Composite analysis:

The mechanisms related to the surging of the ice sheet are highly non-linear, leading to variability between individual realizations of the modeled events even under quasi-identical conditions (Soucek and Martinec, 2011). This variability is further amplified by feedbacks in the fully coupled ice sheet–climate model. To reduce the influence of variability and thus obtain more robust results, we perform all further analysis on a composite of all four events.

Note also that the duration and amount of freshwater for the different experiments are surprisingly similar (See table 1). The reasons for this should be discussed.

We expanded the discussion:
The surges show a very similar peak discharge rate. This is most likely set by the geometry of the Hudson strait limiting the flow. Despite this, even the surges occurring at the same time surprisingly dissimilar. ExB and ExC are initialized shortly before the surge, and the only difference between the setups is the routing of the Nile river. Accordingly, the ice sheets are very similar the beginning of the surge (Fig. 1b). Then, however, their evolution diverges due to the extreme non-linearity of the processes involved in the surge with switching between fast sliding and non-sliding basal conditions. The similarity in basic shape and peak discharge as well as the differences between individual realizations resulting from the non-linearities are in perfect agreement with idealized studies (Calov et al., 2010; Soucek and Martinec, 2011) as well as Roberts et al. (2016).

The glacial state is not well defined in the manuscript. More detail should be given to the difference between the background climate states at the time of the H-events simulated in the model.

We added:

The individual pre-surge states show only small deviations from their mean discussed here (Supplement Fig. 1)

And finally, one of my major concerns with this study is the lack of comparison of the results with proxy data. Given that there is a wealth of data highlighting changes in ocean, and atmosphere climate during the glacial and across H-events the model results should be discussed in relation to these. In particular, as the study clearly states its relevance for studying H-events of the past.

We re-wrote large parts of the discussion and included comparisons with proxy data.

During the surge phase, we observe a reduction in temperature and SSS across the North Atlantic (Fig. 5a,c), which is confirmed by proxies from the eastern side of the North Atlantic (e. g. Heinrich, 1988; Bard et al., 2000). During the post-surge phase, a slight ocean cooling remains, but the SSS largely recovers (Fig. 5b,d). On land, pollen records from around the Mediterranean and western Europe (e. g. Tzedakis et al., 2004; Fletcher and Sánchez Goñi, 2008; Fletcher et al., 2010), show a correlation between the Dansgaard-Oescheger cycles and tree types and cover. The pollen indicate relatively cold, arid climates during stadials and relatively warm, humid climates during the interstadials. Among the stadials, the Heinrich Stadials are the coldest and most arid. In the model, we observe colder, and, over large parts of Europe also drier conditions during the surge and post-surge phases (Fig. 10), matching the trend in the proxies. In the surge phase, this is a result of the aforementioned AMOC decrease, in the post-surge phase, the lowered Laurentide Ice Sheet allows the jet stream to expand northwards, and to take a more northerly path across the Atlantic. This is the main mechanism of the topography change experiments of Roberts et al. (2014a), and consistent with simulations by Merz et al. (2015), who showed that changes in the glacial topography triggered anomalies in the stationary wave activity in their atmosphere-land-only model. These anomalies lead to a shift in the eddy-driven jet stream. In sensitivity experiments with glacial and present-day ice sheet configurations under similar climate forcing, they found that the presence of the Laurentide Ice Sheet was the dominant driver for a southward shift and acceleration of the jet stream. The dependence of the jet stream path on ice sheet height is also known from other simulations (Ullman et al., 2014). The slightly different timing of the modeled climate changes with land changes lasting longer than ocean changes calls for high-resolution proxies with reliable cross-dating across the land/ocean interface.

In the tropics, the most prominent change in the model and in proxies is a southward shift of the ITCZ (e. g. Arz et al., 1998; Hessler et al., 2010). Over Africa, we observe this shift in the surge as well as in the post-surge phase. Over the Atlantic, it is 5 a southward shift during the surge phase and a pure increase in the post-surge phase. The shift is also observed in freshwater hosing experiments, where the surface elevation effect generally is not represented. In our experiments as well as in the fresh- water hosing studies, the freshening reduces the Atlantic heat transport and thus causes a dipole anomaly in the sea surface temperatures with cooling in the North (Fig. 5c) and a slight warming in the South (not shown). This is generally associated with the southeast shift of the ITCZ (e. g. Stouffer et al., 2006). Around the Gulf of Mexico, our model yields mixed results in the comparison with proxy data. For all Heinrich events,

Grimm et al. (2006) find an increase in pine tree pollen and indication for an increased lake level in Lake Tulane (central Florida). Our model simulates an increase in precipitation north of Florida, and drying in Florida and the Caribbean during the surge phase (Fig. 10b) and large-scale drying in the post-surge phase (Fig. 10c). Considering that this is a mismatch by only one grid cell in the surge phase, we can safely ignore this. Rühlemann et al. (2004) find a rapid warming of 1–3 K for intermediate depth (1299 resp. 426 m) waters at 17 ka in benthic foraminifera from sediment cores from the Carribean Sill and Angolan coast. Our model shows an sub-surface warming extending from 60 to 400 m at the Caribbean sill reaching 2.4 K in the surge phase and 0.7 K in the post-surge phase, and no change at the Angolan coast (not shown).

In the Greenland ice cores, the Heinrich stadials do not appear colder, but generally longer than the normal Dansgaard- Oeschger stadials (e. g. Bond et al., 1993). Our model simulates an increase in the sea-ice cover (Fig. 6), and colder (Fig. 5) and dryer conditions, during both the surge- and post-surge phase. These effects could plausibly prolong a pre-existing stadial in the ice core records.

The EPICA Community Members (2006) present data from the Byrd (80°S, 110.5°W), EPICA Dronning Moud Land (EDML, 75°S, 0°E), and EPICA Dome C (75°S, 123.3°E) ice cores. They show warming trends during the Dansgaard-Oeschger Stadials reaching values between 0.5 and 3 K, with longer and thus stronger warmings during Heinrich Stadials. This warming is generally ascribed to the bipolar see-saw effect of an AMOC weakening (Stocker and Johnsen , 2003). Our model simulates a general cooling trend most of Antarctica in both phases, with the exception of an 0.1 K warming at EDML during the surge phase. The bipolar see-saw effect could possibly be obtained in a model with a less stable ocean circulation, or with a stronger ice discharge, both leading to a stronger AMOC reduction.

SPECIFIC COMMENTS:

Line 11, page 3: chose either 2D or 3D. Would assume it is both which are relevant here, but only 2D fields can be shown in the paper.

We chose 2D.

Line 14, page 3: it is stated that PDD method is applied. Please give more details of how this is implemented in the model.

We expanded to:

The surface mass balance is computed using downscaled precipitation and temperatures in a Positive Degree Day (PDD, Reeh, 1991) scheme in the ice sheet model. The PDD scheme employs the Calov–Greve integral method (Calov and Greve, 2005) to compute PDDs from monthly mean temperatures and standard deviations. The PDDs are then converted to snow and ice melt. As in Ziemen et al. (2014), the temperature standard deviation for the PDD scheme is computed from 6- hourly atmosphere model output. Extending this method, a minimum sub-monthly standard deviation of 4 K is prescribed. This prevents the standard deviation from falling too low in areas, where ECHAM5 simulates melt and limits the surface temperature to 0°C.

Line 7, page 4: there is a reference to spurious ice over Siberia. What is this and what does it mean? Give more detail with reference to figure e.g. . ..

We added

(Fig. 8 of Ziemen et al. (2014), its re-emergence can be seen in Fig. 2)

Line 8, page 5: The pre-surge AMOC has a strength of 19Sv, whereas the PI AMOC is 15.8 Sv (see Section 3.1). Why is there a difference, and why is the glacial run AMOC stronger? How does this compare with other model studies and with data?

We added:

The stronger AMOC under glacial conditions is a common feature in model simulations (Weber et al., 2007). Proxies are somewhat ambiguous with respect to the strength of the glacial AMOC (Klockmann et al., 2016), but clearly show a shoaling that is missing in our model. In our model, the AMOC strengthening most likely results from wind-stress changes induced by the larger ice sheets having a slightly stronger effect than the reduced greenhouse gas concentration (Klockmann et al., 2016).

Line 11, page 5: it is stated that the Laurentide is connected to Greenland at the glacial in the model, including the entire Greenland shelf. This is an interesting result which should be further discussed. Is this expected from data, are there similar findings by other studies?

We added:

The connection between the Greenland and the Laurentide Ice Sheet during the glacial is well established, as is its widening (e. g. Lecavalier et al., 2014). The exact outlines of the LGM Greenland Ice Sheet are still under debate with growing evidence for advances far onto the shelf (Arndt, 2018). The details in the model results should, however, be taken with a grain of caution, as the resolution of all components is rather low.

Line 15, page 5: it is stated that the Hudson strait Ice stream has a cycle in surging, whereas other ice streams are constant. Explain why. It is clear from the results that the Hudson Bay system is special, with binge-purge type oscillations. Why is this only the case here? Are there other similar systems?

We expanded to clarify:

A large part of the ice discharge into the ocean is channeled in ice streams. Some of them are constantly active (Fig. 2), while most of them, such as the Hudson Strait Ice Stream (24 in Fig. 2), perform surge cycles alternating between active and inactive states (see video supplement). The underlying mechanism for the surges is related to the binge-purge cycles described by MacAyeal (1993). Similar cycles are also observed in a variety of mountain glaciers and have been related to the thermal and precipitation regime of these glaciers (Sevestre and Benn, 2015). In a warm / high precipitation regime, glaciers tend to have a continuously warm base and strong flow often with basal sliding. In cold / low precipitation regimes, glaciers are generally cold based without basal sliding. In an intermediate regime, neither state is be stable, and glaciers perform surge cycles.

Line 30, page 6: it is stated the convection depth changes. However, this is not clear form figure 5. It would be beneficial to add more details in terms of change in e.g. mixed layer depth or similar.

We clarified:

The mixed layer depth in all deep water formation areas is reduced by up to 800 m (compare Fig. 5a, c)

Line 18, page 7: it is stated that the freshwater is drawn down by deep water formation. What is this based on? If this is true, please show it in a figure or similar. Puzzling that freshwater can be drawn down given its low salinity.

We clarified:

During the surge phase, the low salinity anomaly caused by the melting ice is entrained in the NADW. The combination of freshening and cooling leads to a moderated reduction in the density, so the NADW formation remains active, albeit weakened.

Line 15, page 10: it is stated that the jet stream changes. Why is this? Explain.

We expanded:

in the post-surge phase, the lowered Laurentide Ice Sheet allows the jet stream to expand northwards, and to take a more northerly path across the Atlantic. This is the main mechanism of the topography change experiments of Roberts et al. (2014a), and consistent with simulations by Merz et al. (2015), who showed that changes in the glacial topography triggered anomalies in the stationary wave activity in their atmosphere-land-only model. These anomalies lead to a shift in the eddy-driven jet stream. In sensitivity experiments with glacial and present-day ice sheet configurations under similar climate forcing, they found that the presence of the Laurentide Ice Sheet was the dominant driver for a southward shift and acceleration of the jet stream. The dependence of the jet stream path on ice sheet height is also known from other simulations (Ullman et al., 2014)

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