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Climate of the Past Discussions

Deglacial abrupt climate changes: not simply a freshwater problem

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Abstract.

Imposing freshwater flux (FWF) variations in the North Atlantic is an effective method to cause reorganisations of the Atlantic Meridional Overturning Circulation (AMOC) in climate models. Through this approach, models have been able to reproduce the abrupt climate changes of the last glacial period. Such exercises have been useful for gaining insight into a wealth of processes regarding the widespread climatic consequences of AMOC variations. However, an issue that has passed unnoticed is the fact that the timing of the FWF applied in these studies is inconsistent with reconstructions. Here we focus on the deglaciation to show that imposing a FWF that is derived from the sea-level record results in a simulated AMOC evolution in a poor fit with the data, revealing an inconsistency between the generally accepted FWF mechanism and the resulting climatic impacts. Based on these negative results, we propose that the trigger of deglacial abrupt climate changes is not yet fully identified and that mechanisms other than FWF forcing should be explored more than ever.

1 Introduction

Since the discovery of Dansgaard-Oeschger (DO) events (Dansgaard et al., 1993), the mechanisms underlying glacial abrupt climate changes have been the focus of substantial efforts from the paleoclimate research community. Millennial-scale abrupt climate changes appear to be characteristic of glacial rather than interglacial periods, with deglaciations also showing abrupt deviations from smoother orbitally-driven climate behavior. The last deglaciation contains what is thought to be the most recent DO warming, the Bølling-Allerod. Although, the implication of reorganisations of the Atlantic Meridional Overturning Circulation (AMOC) has emerged as a robust paradigm to explain most features of DO events ((Lynch-Stieglitz, 2017) and references therein), there is no consensus yet as to what their ultimate cause was. Several mechanisms have been proposed as potential triggers of these reorganisations. These include external forcings such as variations in the solar radiation (Braun et al., 2005), the tidal effect of the lunar declination (Keeling and Whorf, 2000), coupled ocean-atmosphere oscillatory mechanisms ((Peltier and Vettoretti, 2014)) related to the formation of super polynias (Vettoretti and Peltier, 2016), oscillatory processes derived from the interactions between ice shelves and sea ice (Boers et al., 2018), or climate-related feedbacks favoring the existence of a glacial oscillator (Arzel and England, 2012; Dokken et al., 2013; Banderas et al., 2015; Zhang et al., 2017b).

The fact that millennial-scale abrupt climate changes are characteristic of glacial rather than interglacial periods has led to the widespread assumption that the presence of ice sheets played an important role in their occurrence. Meltwater discharge from ice sheets surrounding the Nordic Seas is often implied as a potential cause of ocean instabilities that could lead to glacial

https://doi.org/10.5194/cp-2019-96
Preprint. Discussion started: 16 September 2019

abrupt climate changes (Schmidt and Hertzberg, 2011). Modelling efforts have thus often resorted to the use of FWF to trigger AMOC variations and thereby simulate millennial-scale variability (Ganopolski and Rahmstorf (2001); Menviel et al. (2014)). Through this approach, it has been possible to reproduce the timing of abrupt climate changes during the last glacial period (LGP) Menviel et al. (2014), as well as during the last deglaciation He et al. (2013); Liu et al. (2009). Such exercises have been very useful for gaining insight into a wealth of processes, including the role of AMOC variations in the hydroclimatic response at low latitudes (Lu et al., 2013; Menviel et al., 2014; Liu et al., 2014; Otto-Bliesner et al., 2014), their subsurface temperature imprint in the North Atlantic ocean (Marcott et al., 2011), the phasing during deglaciation of atmospheric CO2, Antarctic and global temperature (Shakun et al., 2012), as well as the lead of the Southern Hemisphere (SH) relative to the NH (He et al., 2013), the surface mass-balance response of the Laurentide Ice Sheet throughout the deglaciation (Carlson et al., 2012), or the observed deglacial phasing difference in deep water oxygen isotope records (Zhang et al., 2017a).

However, the question as to whether the FWF applied in these studies is realistic or not has never been assessed in detail. For the last deglaciation, Liu et al. (2009) recognised that their model could reproduce the abruptness of the BA onset only if the meltwater flux (MWF) in the North Atlantic in their simulation terminated within centuries before the BA. Ivanovic et al. (2018a) used the ICE-6G reconstruction (Peltier et al., 2015) to explore the effects that a more realistic freshwater forcing had on the oceanic circulation during Heinrich stadial 1. Carlson et al. (2012) furthermore noted that this FWF deviated from that of Peltier (2004b) after 14.5 ka BP, since no MWF associated to MWP-1A was applied in these simulations, and that if such a flux had been added to the North Atlantic, the model would have failed to produce the BA warming. Bethke et al. (2012) showed this was indeed the case in their model when using the ICE-5G reconstruction (Peltier, 2004b) in an attempt to assess whether ice-sheet reconstructions can be used to constrain meltwater for deglacial simulations. This issue has largely passed unnoticed and deserves further attention.

A critical aspect of the last deglaciation thus concerns the timing and source of MWP-1A, the interval of fastest sea-level increase (10-20 m in less than ca. 340-500 years; (Deschamps et al., 2012)) during the last deglaciation (Fairbanks, 1989; Hanebuth et al., 2000; Peltier and Fairbanks, 2006), indicating rapid mass loss from one or more ice sheets to the global oceans. Ice-sheet reconstructions such as ICE-5G (Peltier, 2004b), ICE-6G (Peltier et al., 2015) or GLAC-1D (Tarasov et al., 2012) are based on past sea-level changes and isostatic adjustment of the Earth's crust to ice unloading during deglaciation, and have yielded objectively constrained deglacial meltwater histories. However they are poorly constrained due to their coarse time resolution (between 1 and 0.1 kyr) resulting from the large depth and age uncertainties in individual sea-level proxies, and to uncertainties in isostatic rebound and drainage reconstructions. Two different timings have been proposed for MWP-1A depending on different sea-level proxies. The first scenario, based on Barbados Uranium/Thorium-dated fossil coral records, place it around 14.2 ka BP or earlier (Fairbanks et al., 2005; Peltier and Fairbanks, 2006; Stanford et al., 2006). This implies that MWP-1a coincided with the Older Dryas cooling event that terminated the BA (Stanford et al. (2011) and references therein) precisely at the time when a reconstruction of North Atlantic Deep Water (NADW) strength over Eirik Drift, offshore southern Greenland, indicates a ca. 200 year weakening of the AMOC (Stanford et al., 2006). The second scenario, based on radiocarbon dating from the Sunda Shelf, suggests the MWP-1A started 300-500 years earlier (Hanebuth et al., 2000). This implies that MWP-1A coincided with the BA and the AMOC intensification at that time (McManus et al., 2004; Stanford et al., 2006), which would represent a real conundrum in our understanding of atmosphere-ocean dynamics. Sea-level fingerprinting inferred from an Earth model of glacial isostatic adjustment (Clark and Mix, 2002) and re-analysis of the dating of glacial retreat evidence (Clark et al., 2009) nevertheless support the idea that a large component of MWP-1A originated in the AIS, and coupled climate model studies have shown that if this MWP had been sourced from the AIS, it could have caused the sharp AMOC resumption in the North Atlantic, and the BA warming (Weaver et al., 2003).

Stanford et al. (2011) have more recently presented a method to assess the uncertainties in a compilation of far-field sea-level reconstructions for the last deglaciation. Their approach builds upon the suggestion of Earth models that isostatic sea-level changes at far-field sites (i.e., at low latitudes) would have been minimal during the deglaciation (Milne and Mitrovica, 2008). Thus, they used datasets from these regions together with a Monte Carlo method to establish the confidence intervals for the sea-level history throughout the last deglaciation. Regarding MWP-1A, their results show that sea-level rise started to accelerate around 17 ka BP. MPW-1A was identified as a robust maximum in the rate of sea-level rise, starting around the timing of the Bølling warming (14.6 ka BP), with its maximum post-dating the BA. MWP-1A appears therefore as a broader culmination of the continuous increase in the rates of rise that started at around 17 ka BP. An additional pulse (MWP-1B) is identified just after the end of the Younger Dryas (11.3 ka BP).

Here we focus specifically on the deglaciation to investigate in detail the FWF hypothesis that meltwater injected in the North Atlantic can explain deglacial abrupt climate changes. We simulate the deglaciation history of the AMOC using an Earth system model of intermediate complexity (EMIC) forced by realistic FWF scenarios. The model output is then compared with a reconstruction of the AMOC to assess the realism of each scenario. We start by considering the reconstruction of Peltier (2004b), since it was the reference sea-level data shown in Liu et al. (2009). We then use the more recent ICE-6G and GLAC-1D reconstructions (Peltier et al., 2015; Tarasov et al., 2012) as the same time as we consider the Stanford et al. (2011) reconstruction to incorporate an estimate of uncertainty. Finally, because Stanford et al. (2011) represents global ESL, the source of MWP-1A is not well established. Thus we investigate also whether an important fraction of the FWF input associated with MWP-1A could originate in the AIS, as in Weaver et al. (2003), under the constraint that the total FWF applied is consistent both with NH and the global ESL record.

2 Materials and Methods

The model we used is the EMIC CLIMBER-2, which has been extensively documented (Petoukhov et al., 2000; Ganopolski et al., 2001) and used successfully in many different climatic contexts (e.g. Ganopolski and Brovkin (2017); Alvarez-Solas et al. (2011); Ganopolski and Rahmstorf (2001); Ganopolski et al. (2016)).

The starting point of our study is a climate simulation with CLIMBER-2 with boundary conditions and external forcing representing the Last Glacial Maximum (LGM); namely, an atmospheric CO2 concentration of 200 ppmv, and insolation and ice-sheet distribution prescribed to their values at 22 ka BP. The model was integrated from 22-14 ka BP forced by changes in insolation (Berger, 1978) and atmospheric CO2 concentration, mimicking the experimental setup used by Liu et al. (2009). In addition, FWF forcing was imposed in the North Atlantic following three different scenarios. First, the same FWF used

https://doi.org/10.5194/cp-2019-96
Preprint. Discussion started: 16 September 2019

Climate of the Past Discussions

by Liu et al. (2009) was imposed. An extended version of this simulation to 0 ka BP was subsequently carried out by He et al. (2013); Liu et al. (2014) in what is thereafter referred to as the TraCE-21K experiment; we will thus refer to this run as FT21K. We then considered the NH and global sea-level reconstructions by Peltier (2004b), Tarasov et al. (2012), Peltier et al. (2015) and by Stanford et al. (2011), respectively, to produce a more realistic FWF forcing for the deglaciation (FP2004, FGLAC-1D, FICE-6G and FS2011, respectively). In the latter case, several trajectories corresponding to different percentiles of their estimated sea-level evolution were used. Finally, a new ensemble of simulations with FWF forcing in the Southern Ocean (FSO) was carried out as well. The forcing imposed in this case was specifically chosen so as to guarantee compatibility with both the ice-sheet based reconstructions and Stanford et al. (2011).

The CLIMBER-2 model version used in this study did not include an interactive ice-sheet component. Instead, we fixed the ice-sheet extension and elevation at its LGM values Ganopolski and Rahmstorf (2001). Boundary conditions in terms of topography and the nature of the points (bare rock or ice) were determined from previous offline ice-sheet simulations using of Gremlins in the Northern Hemisphere (Bonelli et al., 2009) and of GRISLI in in the Antarctic Ice Sheet (Philippon et al., 2006). We here focus on the effects that different FWFs can have on the AMOC, thus isolating the interactive effects of the ice sheets is strategically justified.

3 Results

The simulated LGM AMOC strength at the starting point of our simulation at 22 ka is of ca. 22 Sv. In FT21K, under the same FWF forcing as Liu et al. (2009), the CLIMBER-2 model reproduces a very similar evolution to that of the more comprehensive CCSM-3 Atmosphere-Ocean General Circulation Model (AOGCM; Figure 1a). The AMOC starts to decrease gradually at 19.5 ka, shuts down at 17 ka, and abruptly resumes around 14.5 ka, with an overshoot to above 40 Sv, well above its initial value (Figure 1c). As in He et al. (2013); Liu et al. (2009), the shutdown of the AMOC during Heinrich event 1 (HE1) and its subsequent recovery at the BA, as suggested by reconstructions (McManus et al., 2004), is thus well reproduced. This is a clear response to the imposed FW forcing: first an increase up to 0.2 m/kyr during 2 kyr, then a stabilisation for 2 kyr more and a final sudden termination, which Liu et al. (2009) recognise as the trigger of their simulated abrupt BA.

The forcing and boundary conditions used by He et al. (2013); Liu et al. (2009) were claimed to be realistic. Regarding the FWF forcing used, this claim is based on the similarity between the implied equivalent sea level (ESL) variation and the sea-level reconstruction by (Peltier, 2004b) (Figure 1a). However, the critical field here is not the ESL but the actual FW forcing, which is the time derivative of ESL. Comparison between the FWF forcing imposed by He et al. (2013); Liu et al. (2009) with that implied by Peltier (2004b) for the NH clearly illustrates the mismatch between both (Figure 1b). The resulting FWFs are actually almost in opposite phase. Between 20-18 ka the FWF of He et al. (2013); Liu et al. (2009) is zero, while that derived from Peltier (2004b) shows positive, large values. Between 18-15 ka BP, FWF in He et al. (2013); Liu et al. (2009) reaches its peak values while in Peltier (2004b) it is at intermediate ones. At 15 ka BP, FWF in He et al. (2013); Liu et al. (2009) sharply goes to zero while in Peltier (2004b) it increases to its peak values, roughly coinciding with MWP1A.

https://doi.org/10.5194/cp-2019-96
Preprint. Discussion started: 16 September 2019

Climate of the Past Discussions

We next impose (Peltier, 2004b)'s implied FWF in the North Atlantic in the model, in the same location as in FT21K (FP2004, Figure 1). In this case, the timing of the AMOC evolution is completely reversed with respect to FT21K and no longer fits the data. The AMOC shutdown already takes place before 19 ka BP in response to the positive FWF perturbation at that time, thus much earlier than in the paleodata. This shutdown is followed by a slight recovery starting around 17.5 ka BP, at the time of HE1, and by a subsequent shutdown during the BA and MWP-1A, at the time when a recovery is expected according to the data. These results remain robust when considering the more recent GLAC-1D and ICE-6G sea-level reconstructions. When the model is now forced with their derived FWFs (Figure 2a), the AMOC shuts down during the Bølling Allerod in a completely out-of-phase manner when compared to the AMOC reconstruction (Figure 2b).

Our results thus illustrate clearly that when a realistic, SLE-based FWF is imposed in the North Atlantic, the simulated response disagrees with the AMOC reconstruction. Because the ESL is the time-integral of the freshwater forcing, abrupt changes in freshwater forcing can go unnoticed in the ESL.

Liu et al. (2009) argued that the two cases they considered (sudden and gradual FWF shutoff) represent two end members for simulations with more realistic FWF forcing. However, according to Peltier (2004b), Tarasov et al. (2012) or Peltier et al. (2015), this is far from being the case since actually no FWF decrease can be inferred from these sea-level records after 17 ka BP. Instead, FWF either roughly stays constant or strongly increases (Figures 1b and 2a). The same occurs if the more recent reconstruction by Stanford et al. (2011) ESL is used (Figure 3). The latter furthermore includes an estimate of the uncertainty corresponding to different confidence intervals. Figure 3 shows that the ESL and FWF forcing used by Liu et al. (2009) do not agree with the most likely reconstruction for Stanford et al. (2011) either (within the uncertainty range of the latter reconstruction), which again suggests that a gradual FWF increase from 18-14 ka BP is more realistic (Figure 3).

Stanford et al. (2011) corresponds to global ESL. Given its relatively large uncertainty, the question remains as to whether removal of FWF from the North Atlantic just before the BA can be excluded. To address this potential concern, we investigated the AMOC response to the most likely FWF estimates of this reconstruction as well as to the FWF forcing implied by the upper and lower interval limits for the 67%, 95% and 99% confidence levels (Figure 4a). The lower-limit cases (in blue) imply removal of freshwater from the North Atlantic at roughly 21 ka, 19.2 ka, 17.2 ka and 14.5 ka BP; the upper-limit cases (in red), with addition of freshwater in the intermediate time intervals. When imposed into the model, the lower limit cases lead to a more or less abrupt AMOC strengthening (Figure 4b). The upper limit cases, in contrast, lead to a reduction ranging from a gradual decrease to a sudden collapse. The lower 99% case favors a sharp AMOC increase at the BA, at the same time as it fails to reproduce its large reduction during H1. Therefore, none of these curves was able to reproduce the McManus et al. (2004) Pa/Th reconstructions (Figure 4).

The former results highlight an inherent unsolved feature of the evolution of the climate system within glacial climates; namely, how can we reconcile a major meltwater pulse (the MWP-1A) concomitant or slightly lagging an AMOC resumption (the BA)? How can FWF forcing then possibly be the trigger of such a resumption? One possible solution that has been proposed is that part of MWP-1A originated in the AIS. Indeed, the source of MWP-1A is not well established. Previous studies have suggested that a relatively large FWF input into the Southern Ocean would help to boost the AMOC, thereby triggering the BA event (Weaver et al., 2003). Figure 5 shows the simulated AMOC evolution when injecting water both in the North Atlantic and the Southern Ocean. Note that the NH FWF necessary to keep the AMOC on a halted state previous to the Bolling-Allerod is already inconsistent (of greater amplitude) with the global reconstruction. Figure 5 shows that an intermediate AMOC recovery can be promoted by means of antarctic FWF provided the circulation is already stopped. Nevertheless, it should be stressed that the antarctic freshwater quantities for this recovery to be of ca. 10 Sv are about 5 times greater than the global MWP-1a (red curve in Figure 5). Even for a slight recovery of ca. 3 Sv the Antarctic FWF needs to be of a greater amplitude than the global reconstructed meltwater injection during the MWP-1a.

Stanford et al. (2011) represents a constrain on global ESL, part of whose implied FWF could reflect a contribution from the AIS. Thus, we additionally considered an ensemble of scenarios that maintain agreement with the global sea level reconstruction of Stanford et al. (2011), but assume that part of the implied FWF is input into into the Southern Ocean. In other words, all the realisations shown in Figure 6 were forced in such a way that the sum of the North and the South contributions always give the median value of the Stanford et al. (2011) reconstruction. Additionally, FWF cannot be removed abruptly from the North Atlantic as shown by Peltier (2004b), Tarasov et al. (2012) and Peltier et al. (2015). In this case, a relatively large number of simulations matches the data for the time period comprising the LGM and H1: an AMOC reduction generally starts at 19 ka BP; the AMOC shuts down around 18 ka BP. However, in these cases, the AMOC stays weak or halted for the next 10 ka. In the simulations that do not reproduce an AMOC decrease around H1 that is as pronounced as suggested by data, the AMOC is however able to remain active during the BA. These realisations correspond to those whose FWF in the South is the most intense (Figure 6). Among these, one simulation (highlighted in caption of Figure 6) shows a slight AMOC increase. In this case, the AMOC resumption concomitant with enhanced FWF during MWP-1A is possible because a large part of the FWF is input into the Southern Ocean. As seen in previous studies, this decreases the density of the upper water masses in the Southern Ocean and thereby increases the north-south meridional density gradient, boosting an AMOC recovery. However, even for such a modest AMOC recovery (ca. 2 Sv) more than half of the MWP-1A is required to originate in the AIS, which is highly unrealistic, as discussed below.

4 Discussion

Our results show that we are not able to simulate an AMOC response in agreement with paleoceanographic reconstructions by using FWF forcing alone, when it is in agreement with the available ESL reconstructions. The fact that none of the FS2011-based simulations was able to match paleodata questions the idea that NH FWFs were the main cause of the observed abrupt climate changes. Nonetheless, it could yet be argued that a reduction of the FWF into the North Atlantic from 16-14.5 ka BP is in principle still possible within the uncertainty range of the Stanford et al. (2011) ESL reconstruction, since it is possible to conceive a curve that, within this uncertainty, follows the upper limit curves at the beginning of the deglaciation and the lower limit at the end, thus allowing for an AMOC reduction followed by a sudden invigoration at the BA. However, such a curve is likely not a plausible trajectory according to Stanford et al. (2011). The apparent reduction of the FWF previous to the BA corresponding to the 95 and 99% confidence intervals reflects the large uncertainty of the different records used for this period rather than a trajectory of the real sea-level evolution. None of the individual sea-level records analysed by Stanford et al. (2011) supports a negative sea-level rate previous to the BA. Finally, as stated by Stanford et al. (2011): "MWP-1A is of a multi-centennial to millennial nature, and started at around the timing of the Bølling warming (14.6 ka BP). Its character is not so much that of the sharp spike [...], but more like a broader culmination of the continuous increase in the rates of rise that started at around 17 ka BP."

Menviel et al. (2011) concluded that the good correspondence found in their transient deglaciation experiments between model results and paleo-proxy data suggested that most of the millennial-scale variability recorded during the last deglaciation can be explained by freshwater-induced variations in the deep water formation strength in the North Atlantic, Southern Ocean and North Pacific. Based on their results they concluded that the North Atlantic is the key driver of millennial-scale variability during the Last Glacial Termination.

In many studies, a sudden cessation of water injection seems to be a necessary condition for an abrupt recovery of the AMOC at the time of the BA. However such a condition would additionally imply a stagnation (or even a regrowth) of the deglacial retreat of the NH ice sheets just previous to 14.5 ka, which is not suggested by any reconstruction or sea-level record up to now (Liu et al., 2016). Bethke et al. (2012) already posed the question whether ice-sheet reconstructions can be used to constrain meltwater for deglacial simulations. Although they showed that model sensitivity can have an impact on this response, they also failed to reproduce the BA warming using FWF forcing from (Peltier, 2004b) with an idealized routing scheme. Instead, as in our case, they simulated a shutdown of the AMOC with the associated cooling.

Despite the recognition of some of the authors of the disagreement with the sea-level reconstructions (Liu et al., 2009; Carlson et al., 2012) a clear comparison of the FWF forcing with the ESL-implied FWF has never been shown until now. Neither has a simulation been shown for which the FWF forcing derived from the reconstructed ESL is used as the forcing in the model. We have shown here that when such an exercise is conducted, the timing of the abrupt events during deglaciation is completely incompatible with proxies. This has implications for the interpretation of the climate system behavior during deglaciation. For example, Shakun et al. (2012) state: "The second increase (in temperature) occurs during a pronounced interhemispheric seesaw event (Fig. 5), presumably related to a reduction in AMOC strength, as seen in the Pa/Th record and our modelling (Fig. 4f, g)." If in that modelling exercise, the FWF derived from sea-level records were used, their reproduced fit with the evolution of Antarctic temperatures would have presumably failed (as shown here). This disagreement could reduce robustness to their conclusions about the driver of the deglacial CO2 evolution, at least from the modelling perspective.

An unsolved conundrum is the occurrence of a major deglacial pattern (the MWP1-A) concomitant (or slightly lagging) an AMOC resumption (the BA). Our results suggest that a possible way to circumvent this consists of considering that part of the FWF associated to MWP1A originated in the AIS. Although this suggestion has been made before Weaver et al. (2003), an attempt to fit the global ESL record was not performed in that study. Considering Antarctica as a FWF contributor of MWP1-A is convenient in two ways. First, it allows consideration of a milder contribution of the NH water injection around the time of the BA, still fulfilling the constraints of the the ESL record. Second, it has been shown here and in previous studies that such a FWF in the Southern Ocean can facilitate an AMOC recovery provided the circulation was already in a weak or halted state. However, our model failed to simulate an abrupt recovery as suggested by reconstructions (McManus et al., 2004). This result is in agreement with previous studies showing that even when the southern FWF input is much larger than that of the NH,

https://doi.org/10.5194/cp-2019-96
Preprint. Discussion started: 16 September 2019

its climatic effect is overridden by the effect of northern sources (Ivanovic et al., 2018b). The only exception could be if the northern contribution to MWP-1A is insignificant, however this is very unlikely based on geological and modeling evidence (Dyke et al., 2002; Golledge et al., 2014; Gregoire et al., 2012; Liu et al., 2016; Peltier, 2004a; Peltier and Fairbanks, 2006; Tarasov and Peltier, 2005; Tarasov et al., 2012). According to figure 4, the likeliness of such a scenario is even lower, given the amplitude and timing of the inferred FWFs necessary for fulfilling the global constraints on sea level: On one hand, at least half of the MWP1-A had to originate in the AIS. On the other hand, perhaps more importantly because of its unrealism, the AIS would have been largely expanding between 20 and 15 ka BP in order to compensate for the Northern FWF necessary to weaken the AMOC during that period. Such an Antarctic expansion is not suggested by reconstructions, but they even show rather AIS mass loss during this time. Note Golledge et al. (2014) simulated a maximum contribution to MWP-1A from the AIS of ca. 2m provided an abrupt warming in the Southern Ocean is able to trigger destabilization of the AIS. However, Ivanovic et al. (2018b) furthermore showed that even if it were true, the climatic imprint of Antarctic MWF forcing is short and weak. The reason is that the FWF anomaly dissipates rapidly and the Southern Ocean is quickly re-salinised by the Antarctic Circumpolar Current, enhanced sea ice production, and northward Ekman pumping driven by the westerlies.

The fact that North Atlantic FWF is likely not the driver of the abrupt AMOC variations during the deglaciation does not invalidate the value of the results from experiments based on such an approach (Liu et al., 2009; Menviel et al., 2011; He et al., 2013; Menviel et al., 2014). Indeed there is strong evidence that the AMOC did shutdown during this time period and is responsible for abrupt climatic shifts in the North Atlantic and even globally. However, neither conclusions about the validity of the forcing used in those experiments can be reached, nor claims on finding new processes for explaining the abrupt changes are justified.

5 Conclusions

Regardless of the reconstruction used, the FWF required to simulate an AMOC evolution that is consistent with paleoceanographic reconstructions (McManus et al., 2004) is inconsistent with the available ESL reconstructions and vice versa.

No model has so far simulated a realistic AMOC weakening during H1 (or any H-event) using a realistic FWF. By a realistic FWF we mean here a FWF whose integral matches the available sea-level records. Furthermore, no model has so far simulated a realistic AMOC recovery and warming during the BA using a realistic FWF forcing. Considering Antarctica as an active actor contributing to MWP1-A helps both in reconciling the applied FWFs with sea-level records and in favoring an AMOC recovery during the B-A. However, the amplitude of the FWF forcing required is still unrealistic. Considering FWF as the major forcing for deglacial abrupt climate changes would thus require a deeper understanding of the role that the location and distribution of FWFs have on the oceanic response.

All in all, we conclude here that, in spite of the recent major advances on the topic (e.g. Peltier and Vettoretti, 2014; Vettoretti and Peltier, 2016; Boers et al., 2018), the trigger of glacial abrupt climate changes is far from being identified and that mechanisms other than FWF forcing need much closer examination.

https://doi.org/10.5194/cp-2019-96
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Climate of the Past
Discussions

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https://doi.org/10.5194/cp-2019-96
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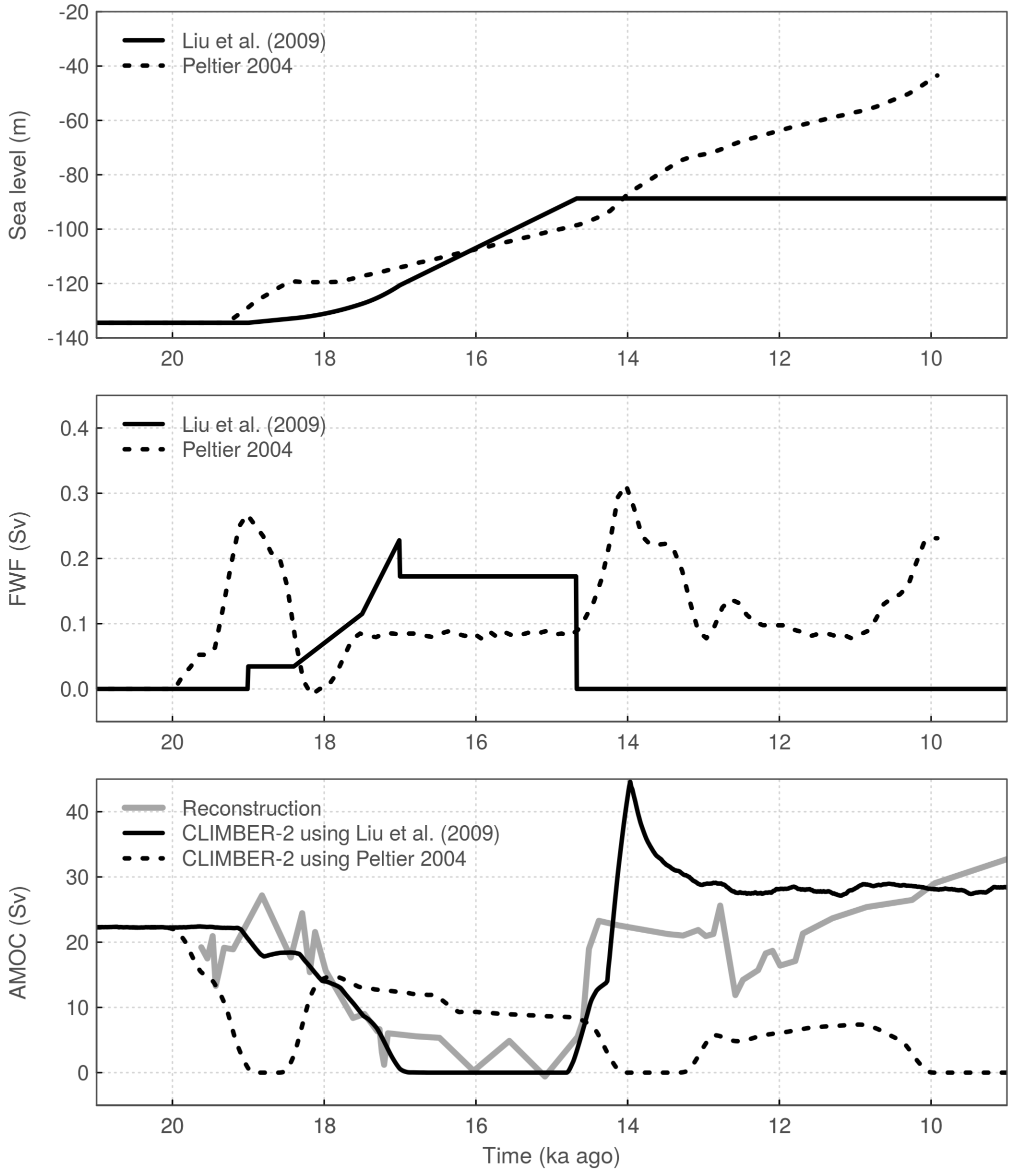

Figure 1. a) Reconstructed ESL (in meters) from Peltier (2004b) (FP2004, dashed line) compared to the ESL used by Liu et al. (2009) (FTR21K, black solid line); b) freshwater fluxes (FWF) in the North Atlantic derived from each of these two approaches (in Sv); c) simulated AMOC evolution with CLIMBER-2 using each of the two forcings.

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Preprint. Discussion started: 16 September 2019

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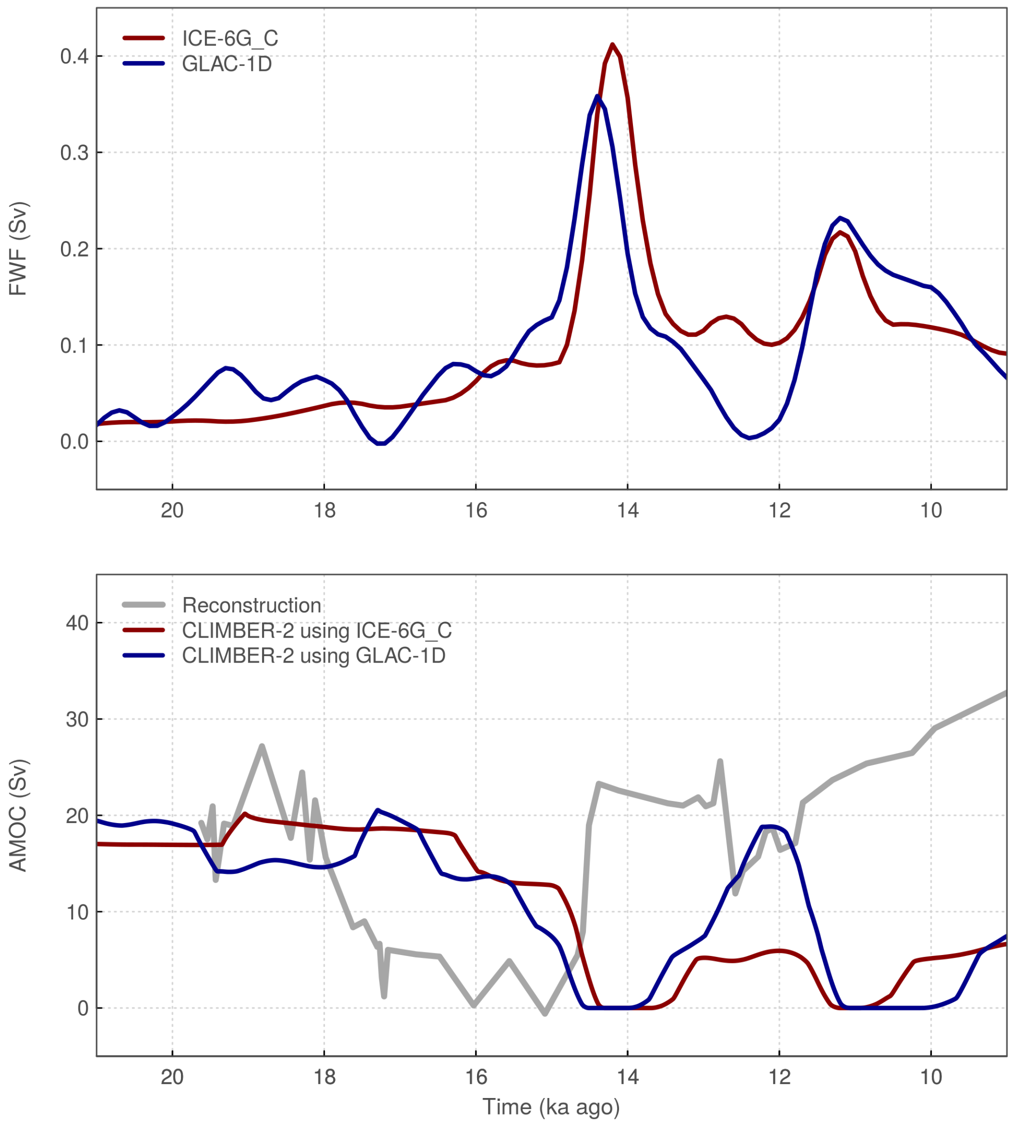

Figure 2. a) Derived freshwater fluxes (FWF) from Tarasov et al. (2012) (blue) and Peltier et al. (2015) (red) (in Sv); b) simulated AMOC evolution with CLIMBER-2 using each of the two forcings.

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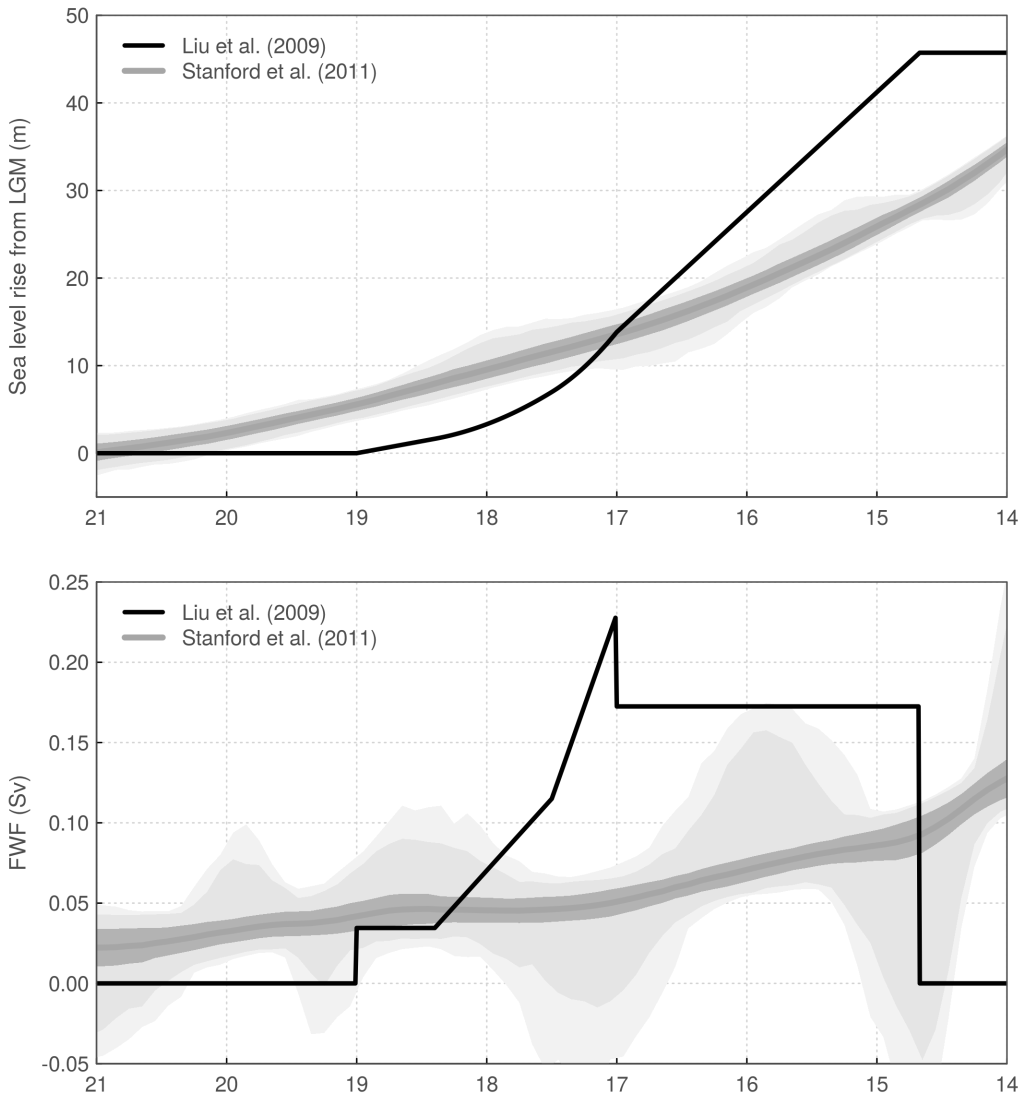

Figure 3. Reconstructed ESL (in meters from the LGM) from Stanford et al. (2011) (grey palette) compared to the ESL used by Liu et al. (2009) (black); b) freshwater fluxes (FWF) derived from each of these two approaches (in Sv).

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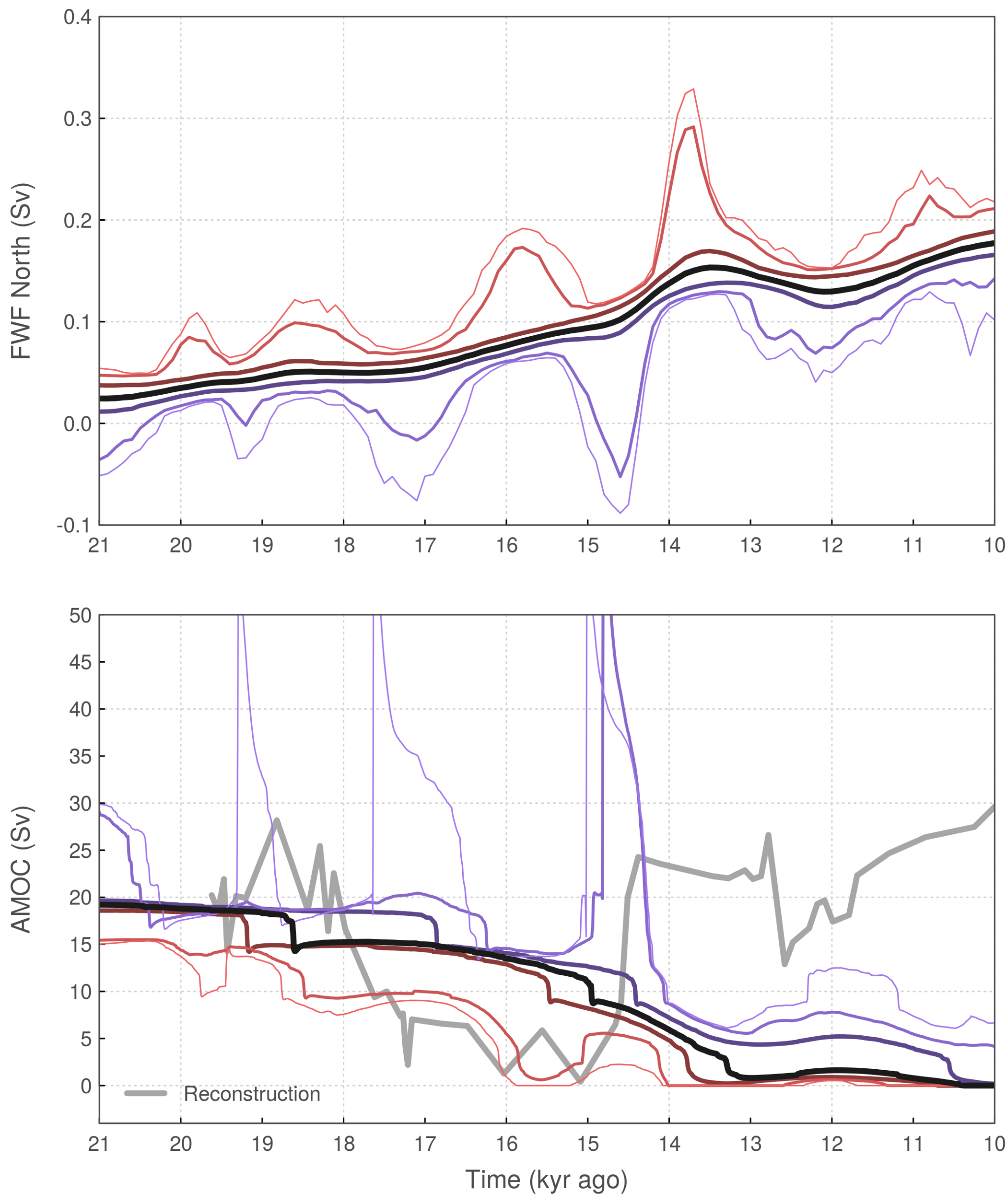

Figure 4. Reconstructed ESL (in meters) from Stanford et al. (2011) (black) compared to the ESL used by Liu et al. (2009) (red); b) the freshwater fluxes (FWF) in the North Atlantic derived from each of these two approaches (in Sv); c) simulated AMOC evolution for with CLIMBER-2 using each of the two forcings FWFs of Standford 2011 and Liu 2009.

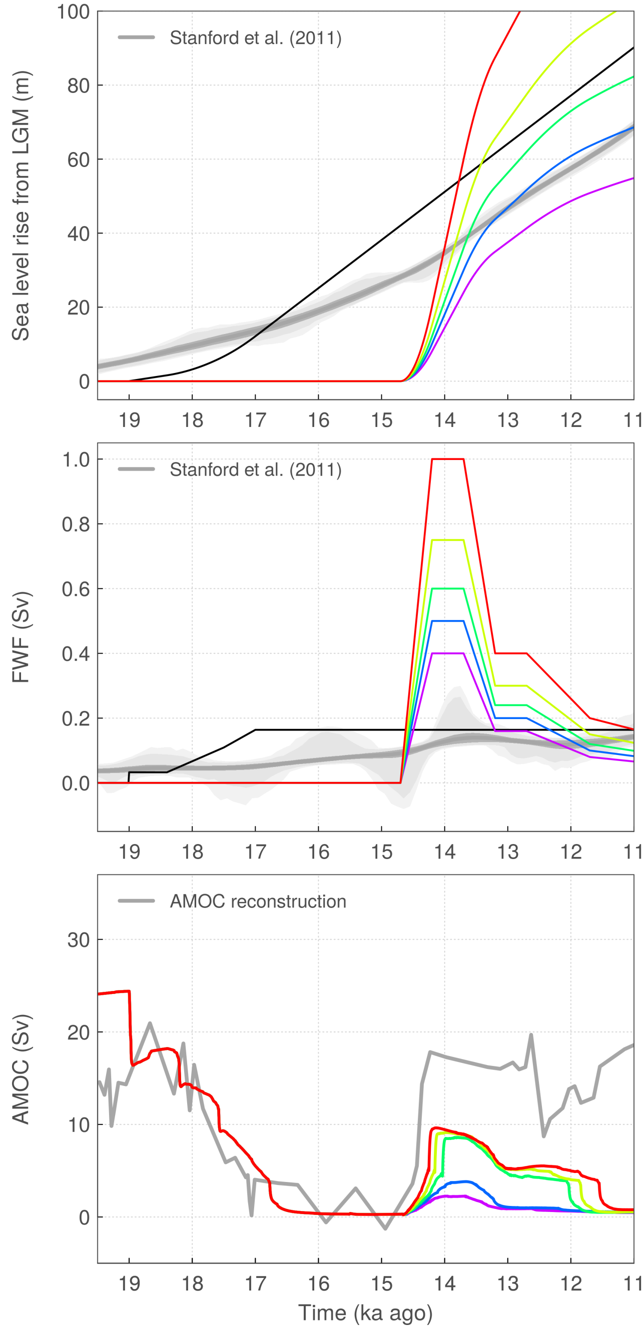

Figure 5. Reconstructed ESL (in meters from the LGM) from Stanford et al. (2011) (grey palette) compared to the ESL used by here. The black line represents the Northern Hemisphere contribution while the color lines illustrate different Antarctic contributions to the sea-level increase; b) freshwater fluxes (FWF) in the North Atlantic (black) and different Antarctic contributions (color) (in Sv); c) simulated AMOC evolution for with CLIMBER-2 using each of the forcings FWFs.

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Preprint. Discussion started: 16 September 2019

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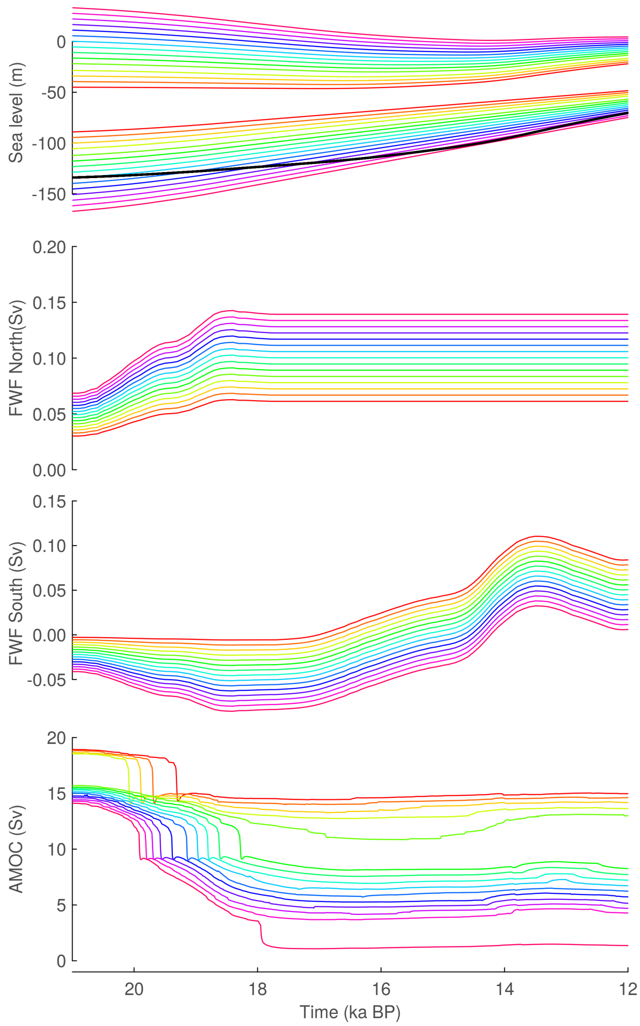

Figure 6. a) Reconstructed contributions to the global ESL evolution from the NH and the SH (in meters; color lines) in the ensembles considered such that the total sum matches Stanford et al. (2011) (black line). The upper part of panel a) shows the Antarctic contributions and the bottom part corresponds to the NH contributions to the global sea-level evolution (black curve); b) FWF in the North Atlantic and c) in the SH (in Sv); d) simulated AMOC evolution. Note that the simulation represented by the lightest green shows a sligth AMOC decrease from 19 ka BP and a slight AMOC intensification during the Bølling Allerod.