Millennial-scale climate oscillations triggered by deglacial meltwater discharge in last glacial maximum simulations

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Abstract
Our limited understanding of millennial-scale variability in the context of the last glacial period can be explained by the lack of a reliable modelling framework to study abrupt climate changes under realistic glacial backgrounds. In this article, we describe a new set of long-run Last Glacial Maximum experiments where such climate shifts were triggered by different snapshots of ice-sheet meltwater derived from the early stages of the last deglaciation. Depending on the location and the magnitude of the forcing, we observe three distinct dynamical regimes and highlight a subtle window of opportunity where the climate can sustain oscillations between cold and warm modes. We identify the European-Arctic and Nordic Seas regions as being most sensitive to meltwater discharge in the context of switching to a cold mode, compared to freshwater fluxes from the Laurentide ice sheets. These cold climates follow a consistent pattern in temperature, sea ice and convection, and are largely independent from freshwater release as a result of effective AMOC collapse. Warm modes, on the other hand, show more complexity in their response to the regional pattern of the meltwater input, and within them, we observe significant differences linked to the reorganisation of deep water formation sites and the subpolar gyre. Broadly, the main characteristics of the oscillations, obtained under full-glacial conditions with realistically low meltwater discharge, are comparable to $\delta^{18}O$ records of the last glacial period, although our simplified experiment design prevents detailed conclusions from being drawn on whether these represent actual Dansgaard-Oeschger events.

Plain Language Summary

During the last glacial period (115,000 to 12,000 years before present), the baseline cold climate was continuously disturbed by intense and abrupt climate changes. They completely modified the climate for a few thousand years or so, resulting, for instance, in massive temperature shifts and complete reorganisations of ocean circulation. These abrupt changes have been observed in climate records from the Northern Hemisphere and also can be traced in records from the Southern Hemisphere. Yet, we still do not know what triggers these changes, and often cannot simulate them at the right time under known environmental conditions. In the context of the Last Glacial maximum, a cold period 21,000 years ago with extensive ice over the Northern Hemisphere, this article analyses a new set of climate model simulations that test the effects of freshwater melting from the ice sheets at different periods of the early deglaciation ($\sim$ 21,000 to 18,000 years before present). Under some conditions, the resulting experiments displayed an Atlantic Ocean that oscillates between strong and collapsed basin-wide circulation, causing approximately 10°C of temperature change over Greenland; a behaviour that resembles observed abrupt climate changes.

1 Introduction

The last glacial period was characterised by strong millennial-scale variability (e.g. Bigg & Wadley, 2001; Wolff et al., 2010; Fletcher et al., 2010)., observed through the occurrence of sharp and dramatic shifts in climate state. The best example of such abrupt changes are Dansgaard-Oeschger events (D-O events; Dansgaard et al., 1993). They consist of transitions between cold stadial and warm interstadial climate conditions that occur in cycles as long as six hundred to a few thousand years. Dansgaard-Oeschger events were first identified in $\delta^{18}O$ records of Greenland ice cores (Bond et al., 1993) before also being observed in Antarctica (Blunier & Brook, 2001; Voelker, 2002). Since their discovery, they have been identified in a wide range of different parts of the Earth system, both marine (e.g. Shackleton et al., 2000; Wolff et al., 2010; Dokken et al., 2013; Henry et al., 2016) and terrestrial (e.g. Goñi et al., 2000; Y. J. Wang et al., 2001; X. Wang et al., 2007; Margari et al., 2009; Stockhecke et al., 2016), and can be linked to meridional shifts of the Intertropical Convergence Zone (ITCZ; Peterson & Haug, 2006).

During decades of study, numerous hypotheses have been put forward to understand the underlying mechanisms behind D-O events (a comprehensive list can be found in Li and Born (2019)), and, more generally, millennial scale variability, yet they remain largely unexplained. Nonetheless, at the crossroads of all theories lies the crucial role of the Atlantic Meridional Overturning Circulation (AMOC) (Rahmstorf, 2002; Burckel et al., 2015; Henry et al., 2016). A modification of the thermohaline circulation affects heat and salt redistribution between the tropics and the poles, and consequently has a global-scale impact on the climate (Clark et al., 2002; Rahmstorf, 2002). There is substantial evidence that the AMOC has existed in other configurations (or 'modes') than the one we observe at present times (e.g. Böhm et al., 2015), and that AMOC may thus have the capacity to exist in multiple stable states, as predicted theoretically (Stommel, 1961) and supported by early observations (Broecker et al., 1985) and climate models (Manabe & Stouffer, 1988). Our understanding of abrupt climate change, therefore, relies on uncovering the cause of AMOC mode switches.

The AMOC can be disrupted by freshwater release events in the North Atlantic-Arctic region. They have the power to target vital points of the thermohaline circulation by affecting the ocean density profile at North Atlantic Deep Water (NADW) formation sites (Broecker et al., 1985; Paillard & Labeyriet, 1994; Vidal et al., 1997). In models, freshwater hosing experiments have been widely used to force abrupt climate transitions (e.g. Manabe & Stouffer, 1997; Ganopolski & Rahmstorf, 2001; Kageyama et al., 2010) and observe hysteresis cycles (e.g. Schmittner et al., 2002). They also highlighted the large sensitivity of the climate to the strength and the location of the release, especially in the Greenland, Iceland and Nordic (GIN) Seas (Smith & Gregory, 2009; Roche et al., 2010). Consequently, it is valuable to explore the different sources of freshwater that had the potential to lead to millennial-scale variability.

Iceberg surges during Heinrich events (H events; Heinrich, 1988) recorded by Ice Rafted Debris (IRD) in the North Atlantic (Hemming, 2004), were first candidate to be held responsible for initiating stadial climates. It is now widely accepted that H events are not at the origin of D-O events, they are triggered within stadial states (Barker et al., 2015) and are not recorded at every D-O occurrence (Lynch-Stieglitz, 2017). Instead, we can conceive of them as a likely response to the earlier climate-ocean perturbation or even a positive feedback mechanism for perpetuating/amplifying stadial climates (Ivanovic et al., 2018). Meltwater released from the long term deglaciation of ice sheets was another significant source of freshwater during the last glacial period (Gregoire et al., 2012), although only a few studies have investigated the influence of such 'background' melt (e.g. Ivanovic et al., 2018; Kapsch et al., 2022; Matero et al., 2017), probably because it requires precise constraints on the ice sheet geometry and history of melt/growth (Bethke et al., 2012). Holding the most complete records of ice sheet evolution, both in terms of spatial and temporal resolution, (e.g. Dyke, 2004; Hughes et al., 2016; Briggs et al., 2014; Bradwell et al., 2021), the last deglaciation (and especially its early phase, $\sim$ 21–16 ka BP, thousand years before present) offers the perfect setting to assess the ability of the early phase of continental deglaciation (i.e. the long-term background melt from disintegrating ice sheets) to generate millennial-scale variability in glacial conditions.

The last deglaciation initiated from the Last Glacial Maximum (LGM; $\sim$ 21 ka, (Clark et al., 2009), which corresponds to a maximum in continental ice sheet extent in the Northern Hemisphere (Batchelor et al., 2019) shaped during the preceding glacial period. The upper cell of the AMOC was likely shallower, but it is not known whether it was stronger or weaker than present day (Gebbie, 2014; Lynch-Stieglitz, 2017; Muglia & Schmittner, 2021). A steady increase in Northern Hemisphere summer insolation (Berger, 1978) triggered the long-term demise of the Laurentide and Eurasian ice sheets, with rising concentrations of atmospheric CO_2 positively reinforcing the deglaciation (Gregoire et al., 2015). However, while Southern Hemisphere temperatures gradually rose (Parrenin et al., 2007), the climate in the North remained cold for several thousand years; a period known as Heinrich Stadial 1 (∼18–15 ka BP) (Denton et al., 2006; Roche et al., 2011; Ng et al., 2018). The most recent of Heinrich events, H1 (Hemming, 2004; Stanford et al., 2011), began some two thousand years after the onset of Henirich Stadial 1 (Stern & Lisiecki, 2013; Hodell et al., 2017). In the years of deglaciation that followed the LGM, several millennial scale events were observed (Weber et al., 2014). Two episodes are particularly relevant to our study: the sudden Bølling Warming (∼ 14.5–13 ka BP; Severinghaus & Brook, 1999) concurrent with an intensification of the AMOC (Ng et al., 2018; Du et al., 2020), and the ensuing Younger Dryas, when Northern Hemisphere climate abruptly returned to a stadial state with glacial re-advance (∼ 13–12 ka; Murton et al., 2010; Liu et al., 2012). While not formally identified as D-O events, similarities in climate and ocean evolution between these last deglaciation and D-O oscillations have prompted others to at least draw analogies between them, and to speculate on whether they have a common cause (e.g. Obase & Abe-Ouchi, 2019).

Simulating climate oscillations in glacial conditions has proven to be very challenging, and even more so during the LGM. This is because of the strong feedback between the large ice sheets and wind stress, deep water formation and energy balance (Oka et al., 2012; Ullman et al., 2014; Beghin et al., 2015; Roberts & Valdes, 2017), which act to intensify, or at least stabilise, the AMOC (Oka et al., 2012; Klockmann et al., 2016; Sherriff-Tadano et al., 2018). Most models from both the Paleoclimate Modelling Intercomparison Project Phase 3 (PMIP3; Muglia & Schmittner, 2015) and Phase 4 (PMIP4; Kageyama et al., 2021) tend to simulate a deeper and stronger NADW than inferred from palaeo records, which could explain why few modelling studies have observed millennial scale variability in glacial background (e.g. Klockmann et al., 2018). In order to trigger abrupt climate transitions, freshwater hosing experiments have historically needed to overestimate fluxes as reviewed by Kageyama et al. (2010) (e.g. Liu et al., 2009; Menviel et al., 2011), and have not succeeded in simulating abrupt changes when using 'realistic' fluxes (Bethke et al., 2012; Gregoire et al., 2012; Snoll et al., 2022). Obase and Abe-Ouchi (2019) have arguably come the closest to overcoming this meltwater 'paradox' by simulating the Bølling Warming even with some deglacial meltwater forcing. However, even they require a significantly lower than likely freshwater discharge from the deglaciating ice sheets (e.g. Peltier et al., 2015).

The dispute over what could feasibly cause abrupt climate changes not directly driven by freshwater fluxes led the community to start actively searching for oscillating behaviours in their models. At the same time, the criticism that simulations integrated for only a few hundred or a thousand years should not be considered to have a steady-state or 'spun-up' ocean circulation began to gain traction (Marzocchi & Jansen, 2017; Dentith et al., 2019), prompting modellers to run long simulations with higher-order climate models — made possible by the increase of computational power — in order to examine long-term drifts. It is therefore probably not a coincidence that more and more coupled Atmosphere-Ocean General Circulation Models (AOGCMs) have reported observing AMOC mode oscillations in recent years (e.g. Peltier & Vettoretti, 2014; Brown & Galbraith, 2016; Klockmann et al., 2018; Sherriff-Tadano & Abe-Ouchi, 2020). They have been achieved under a range of different freshwater hosing scenarios (e.g. Cheng et al., 2011), atmospheric CO_2 concentrations (e.g. Zhang et al., 2017) and ice sheet geometries (e.g. Klockmann et al., 2018), although, to our knowledge, only Peltier and Vettoretti (2014) managed to obtain AMOC oscillations under glacial maximum conditions.

To sum-up the combined results from these studies, there seems to exist a window of opportunity (Barker & Knorr, 2021) in each model's inputs (parameter values, boundary conditions and forcings) and background climates where oscillations can establish and sustain (e.g. Peltier & Vettoretti, 2014; Brown & Galbraith, 2016; Klockmann et al., 2018). The ice sheets' layout in particular has a strong influence on the local and global climate, including on the atmospheric circulation (Löfverström et al., 2014; Roberts et al., 2014; Sherriff-Tadano et al., 2021), the gyres (Gregoire et al., 2018), the energy balance (Roberts & Valdes, 2017) and freshwater fluxes (Matero et al., 2017). As a result, the new generation of better constrained and more detailed ice sheet reconstructions such as ICE-6G_C (Peltier et al., 2015; Argus et al., 2014) and GLAC-1D (Tarasov & Peltier, 2002; Tarasov et al., 2012; Briggs et al., 2014; Ivanovic et al., 2016) may prove to be decisive in whether or not the 'right' conditions for triggering abrupt climate changes are obtained.

In this paper, we present our contribution to this initiative in the form of a new set of LGM simulations forced with deglacial meltwater. Inspired by an initial experiment that showed millennial-scale variability under a transient meltwater forcing, we designed our simulations with fixed meltwater inputs in order to be able to describe the oscillations in detail and evaluate the sensitivity of the oscillatory behaviour to meltwater patterns. These inputs were derived from snapshots of the early deglaciation meltwater history (specifically, between 21.5 and 17 ka BP) calculated from GLAC-1D ice sheet reconstruction.

Depending on the freshwater pattern, we observe three different dynamical regimes, including regular and self-sustained climate oscillations. The oscillations are characterised by switches between strong, shallow glacial AMOC and near- or completely- collapsed AMOC modes, a Greenland surface cooling/warming of $\sim 10°$C, and a periodicity of about 1,500 ka. This regime can be sustained for about 10,000 years (the maximum length of the experiments). These are, to our knowledge, the first general circulation model simulations to use ice sheet reconstruction-derived distributions of meltwater to produce strong AMOC oscillations under glacial maximum climate conditions, and to investigate the sensitivity of the oscillations to patterns of meltwater discharged to the ocean.

The non-oscillating clusters of our experiments inform us of the pre-requisite conditions for passing through the oscillating window. Here, we describe the various simulations in detail, their oscillatory or non-oscillatory climate/ocean states, and the different Earth system components involved in the abrupt events. We conclude with a discussion on the relevance of our simulations in the context of known past abrupt climate changes. The design introduced in this study allows us to undertake a relatively systematic set of sensitivity tests of the impact of realistic freshwater distributions (albeit for unrealistic lengths of time) on ocean circulation, with the multi-millennial integrations enabling us to explore the long-term effect of each pattern.

2 Methods

2.1 The model

The simulations introduced in this article were completed using the BRIDGE (Bristol Research Initiative for the Dynamic Global Environment group) version of the HadCM3 atmosphere-ocean general circulation model (GCM) (Valdes et al., 2017). This GCM consists of a 19 layers $\times$ 2.5° $\times$ 3.75° atmosphere model more completely described by Pope et al. (2000), coupled every simulation day with a 20 layers (up to 5,500m deep) $\times$ 1.25° $\times$ 1.25° ocean model, described by Gordon et al. (2000) (Bryan & Cox, 1972; Fofonoff & Millard Jr, 1983; Fofonoff, 1985). This version of HadCM3 includes the MOSES 2.1 land model (P. M. Cox et al., 1999), and the TRIFFID dynamic vegetation model (P. Cox, 2001). HadCM3 has been tested in many different scenarios (I.P.C.C., 2014; Reichler & Kim, 2008), and was optimised for running multi-millennial simulations (Valdes et al., 2017).

2.2 Experimental design

The LGM simulation that makes up the base climate state for all simulations presented here was created following the PMIP4 protocol for 21 ka BP (Kageyama et al.,

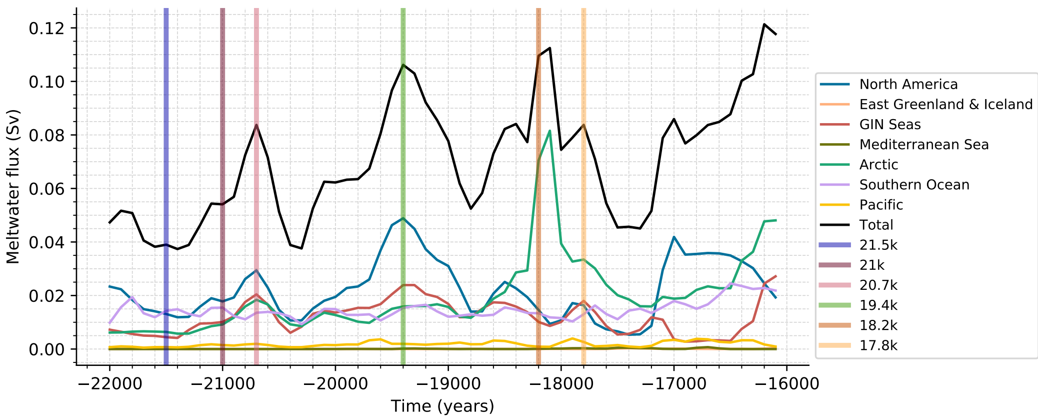

Figure 1. Meltwater discharge history over the early deglaciation and its distribution over key regions (defined in Figure S2*b*). This plot incorporates the 200-years smoothing described in section S2. Vertical bars represent the time steps chosen for calculating each constant meltwater-forcing snapshot (see section 2.2, and Table 1).

2017) using the GLAC-1D ice sheet reconstruction (Tarasov & Peltier, 2002; Tarasov et al., 2012; Briggs et al., 2014; Ivanovic et al., 2016); see section S1 (Bereiter et al., 2015; Loulergue et al., 2008; Schilt et al., 2010). This new HadCM3 LGM simulation was initialised from a chain of existing multi-millennial HadCM3 PMIP3 LGM simulations, which were started from multi-millennial continuations of earlier HadCM3 LGM simulations (Davies-Barnard et al., 2017), giving a pre-PMIP4 LGM spin-up of several thousand years. The new PMIP4 GLAC-1D set-up was integrated for 3,500 years, and the end of this final spin-up phase provides the initial condition for all simulations presented here. We continued the LGM simulation for a further 4,000 years in parallel with our other experiments to provide a reference climate state (*CTRL*) for comparison to the other simulations. There are small, steady drifts in the ocean over the run (Figure S5), but the signal of the trends are dwarfed in comparison to the changes of interest described below, and little is gained for this study by extending *CTRL* further. The starting year of *CTRL* is defined as year 0.

GLAC-1D has rarely been used for LGM simulations compared to the other ice sheet reconstructions (Kageyama et al., 2021) such as ICE-6G_C (Peltier et al., 2015) and the PMIP3 ice sheet (Abe-Ouchi et al., 2015). It was preferred for this study because compared to the alternative reconstructions, it includes more recent constraints on the Eurasian ice sheets (provided by the DATED-1 project; (Hughes et al., 2016), a region that could be crucial for accurately capturing the early deglacial climate history (Ivanovic et al., 2018). A transient meltwater history was derived from GLAC-1D's representation of the deglaciation (section S2). We decided against using the transient meltwater flux, because the added complexity introduced by the temporal variability and possible ocean 'memory' of the preceding [uncertain] meltwater history would have convoluted the physical interpretation of our results. Instead, we examined the triggering of abrupt climate changes using a simpler approach; by selecting six interesting, different, fixed-forcing scenarios (our 'snapshots'), that allow us to investigate the sensitivity of the glacial ocean and surface climate to early deglacial freshwater inputs (Figure 1).

The snapshots were identified for their ability to collectively capture a broad range of possible situations that may have led to changes in ocean circulation and surface climate. The six scenarios correspond to different modes of discharge and are named after the period they were extracted from (see Figure 1); see Figure S2*c* for the spatial dis-

Simulation	Meltwater (Total flux)	Integration length	Category	Salinity Target (PSU)
CTRL	None	4,000 years	reference	35.8334
21.5k	21.5 ka (0.039 Sv)	4,000 years	warm	35.834
21k	21 ka (0.054 Sv)	4,000 years	warm	35.8334
20.7k	20.7 ka (0.084 Sv)	10,000 years	oscillating	35.8225
19.4k	19.4 ka (0.106 Sv)	10,000 years	oscillating	35.7901
18.2k	18.2 ka (0.109 Sv)	10,000 years	slow-recovery	35.7348
17.8k	17.8 ka (0.084 Sv)	10,000 years	oscillating	35.7125

Table 1. Experiments summary. All experiments were designed with LGM boundary conditions, using the LGM GLAC-1D ice sheet extent and associated geographies. Entries in the *Category* and global mean *Salinity Target (PSU)* columns are explained in sections 4 and 2.2, respectively.

tribution of the fluxes. The *21.5k*, *21k* and *20.7k* snapshots were chosen for being close to the LGM, sharing a similar distribution, but with different rates of meltwater discharge. The *19.4k* snapshot hosts a strong Labrador Sea/North Eastern American coast/Gulf of Mexico discharge (shortened to 'North American' discharge hereafter), but has a relatively small meltwater flux to the Arctic. Conversely, *18.2k* and *17.8k* have high Arctic and low North Atlantic discharge, with *18.2k* having the most freshwater entering the Arctic. These six snapshots of the deglacial meltwater history were used as forcing for six new equilibrium-type (i.e. constant-forcing) simulations, started from year 0. The meltwater inputs were kept constant throughout the runs.

Table 1 presents a summary of all experiments. The difference between any of them is the prescribed ice sheet meltwater (or absence of it, in *CTRL*).

The idea of having a continuous fixed meltwater discharge for thousands of years is unrealistic by nature. However, in the most extreme scenario (*18.2k*), the total forcing corresponds to a sea level rise of 102 m in 10,000 years, which, for context, is still less than what has been reconstructed for the whole of the last deglaciation (Lambeck et al., 2014). It therefore remains appropriate to consider our results in light of glacial and deglacial variability in order to understand the effect of the forcing, though we are careful to highlight that this is not a transient simulation of deglacial meltwater. To avoid long-term drifts in mean ocean salinity caused by the long freshwater forcing, we impose a constant global mean salinity target (following the *VFLUX* method of Dentith et al. (2019) commensurate with the starting condition for each 'snapshot' (Table 1). The salinity target conserves water in relation to terrestrial ice volume (applied as relative to the present day) and thus, in the context of these simulations, counteracts global freshening by removing the excess water as a very small proportion of freshwater from every ocean grid cell at every ocean model timestep (one hour). This approach is in keeping with the snapshot/equilibrium experimental design whilst still allowing the ocean to 'feel' the surface forcing. See section S3 for details.

Most simulations were run for 10,000 years; long enough to characterise their climate behaviours. However, like *CTRL*, two simulations (*21.5k* and *20.7k*), were terminated after 4,000 years. At this point, little was changing in those simulations, and since no further time series were required for the analysis, we opted to conserve the computing resource.

Some simulations experienced numerical instability in the stream function off the coast of the Philippines after a few thousand years. This quirk was resolved by smoothing the bathymetry in the region of the instability and restarting the run a few years before the instability arose. More information, including the detail of the smoothing algorithm and its very minor impact on the climate response, is given in section S4.

2.3 Characterising oscillations and defining warm/cold-mode composites

All experiments apart from *CTRL* show some kind of alternation between weaker and stronger AMOC phases, or 'modes'. Changes in the AMOC are correlated to increases and decreases in NGRIP temperatures (North Greenland Ice Core Project, 42.32° W, 75.01° N) and so we will also refer to these phases as 'cold' and 'warm' modes, respectively. We do not use the terms *stadial* and *interstadial* to describe the cold and warm states because of the complicated connotations associated with these terms, but they may be thought of in such a light.

In order to characterise oscillations in the simulations, we applied a filtering algorithm and Fourier analysis to derive the spectrum of the temperature time series at the location of NGRIP, see section S5 for details. If there is a peak in the spectrum, then oscillations can be defined and described, and we we apply a Butterworth low-pass filter to screen-out the frequencies lower than the millennial-scale variability of interest.

We also found it useful, in our analysis, to examine characteristics common to all warm and cold modes in the suite of simulations. Thus, to build a composite of the two modes from the time series of results, we defined quantitative boundaries bespoke to each simulation (it proved ineffective to adopt a consistent definition for all simulations because of their differences). Points below the weak limits (in AMOC strength/NGRIP temperature) were added to the composite cold mode and points above the strong limits were added to the warm modes. This approach is described in section S6, where we demonstrate that the choice of how to define the composite modes does not significantly impact the results, and that to manually set up the weak and strong limits was an easy and robust method to build the composite states.

3 A new weak, shallow AMOC LGM simulation

The *CTRL* run replicates and continues the HadCM3-GLAC-1D LGM simulation presented in Kageyama et al. (2021). The global mean surface temperature is 6.6°C colder than Pre-Industrial (PI). Compared to other PMIP4 simulations, this simulation is in the coolest range, almost 2°C below the average mean temperature, and is colder than any PMIP3 simulations analysed by Kageyama et al. (2021), yet close to the current estimate from global temperature reconstructions ($\sim 6.1°C \pm 0.4°C$ cooler than PI in Tierney et al. (2020), $\sim 7.0 \pm 1.0°C$ cooler than PI in Osman et al. (2021). The global mean ocean surface temperature cools by 3.4°C, which is significantly cooler than the $\sim 1.7 \pm 0.1°C$ cooler than PI inferred by Paul et al. (2021), but again is a good match to the reconstruction by Tierney et al. (2020) ($\sim 3.1 \pm 0.3°C$ cooler than PI). Contrary to most PMIP4 models, HadCM3 produces an AMOC that is both shallower and weaker in a full-glacial background compared to its pre-industrial state, although AMOC is still vigorous with a maximum strength of 20 Sv at 30° N. We also note that the simulation gets slightly cooler when using GLAC-1D compared to ICE-6G_C (Ivanovic et al., 2018), despite similar values for the maximum overturning circulation.

A summary of the *CTRL* equilibrium climate is shown by Figure 2. The thermohaline circulation is fuelled by intense convection in the northeast Atlantic, with deep water formation sites located primarily south of Iceland and west of the British Isles (Figure 2*b*). This creates a corridor in the eastern part of the Atlantic where warm waters can transit to high latitudes, while the western part of the ocean gets covered by winter sea ice and observes a much cooler climate (Figure 2*c*). The winter sea ice layer extends towards the East of the North Atlantic basin, creating a strong North-South gra-

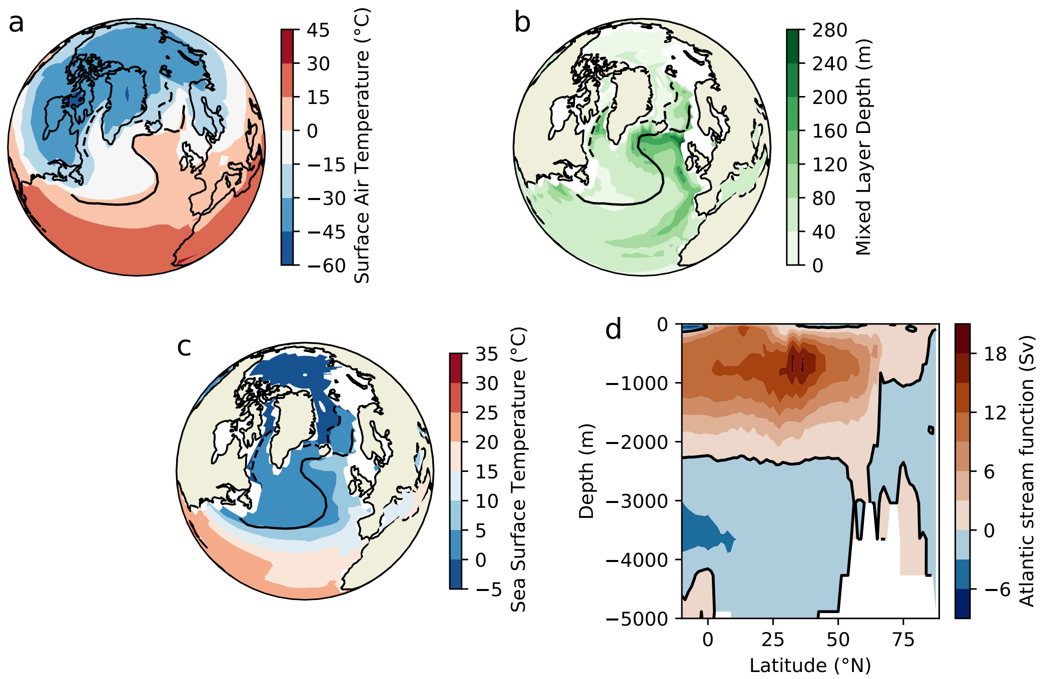

Figure 2. Mean annual conditions between simulation years 3900 and 4000 for *CTRL*. *a.* Surface air temperature. *b.* Mixed layer depth. *c.* Sea surface temperature. *d.* Meridional overturning stream function in the Atlantic basin. Dashed/solid lines indicate the March/September 50% sea-ice extent.

dient in atmospheric temperature in this region (Figure 2*a*). The Arctic Ocean is covered in sea ice all year long. Dense waters sink in the Labrador Sea when the sea ice is less extensive in this region in late autumn.

4 Climate response to the meltwater perturbations

We observe significant differences in the climate response of the six meltwater experiments (Figure 3), best encapsulated by the evolution of the AMOC index. For this study, we define the AMOC index as the maximal value of the overturning circulation in the Atlantic ocean at 26.5° N. This index corresponds to the modern RAPID-array AMOC measurement grid (Smeed et al., 2014) and has been regularly used in palaeo-studies (e.g. Guo et al., 2019).

The meltwater simulations can be assigned to three different regimes, or clusters, according to the AMOC index (Figure 3*b*). In the first grouping, simulations *21.5k* and *21k* returned to the reference state after a short cooling event during the initial years of the forcing, when the ocean adjusts to the introduction of the weak meltwater fluxes. This AMOC decline lasted for approximately 500 years, with the index weakening by as much as 5 Sv for *21k* and 1.5 Sv for *21.5k* approximately halfway through this initial cooling. While *21.5k* hardly recorded a change of polar temperatures (Figure 3*c*), the drop in *21k* AMOC strength drives up to 5°C cooling over Greenland. Neither simulation shows a strong response over Antarctica (Figure 3*d*). Accordingly, *21.5k* and *21k* will be referred to as *warm simulations* (note that with an LGM baseline climate, this term is relative).

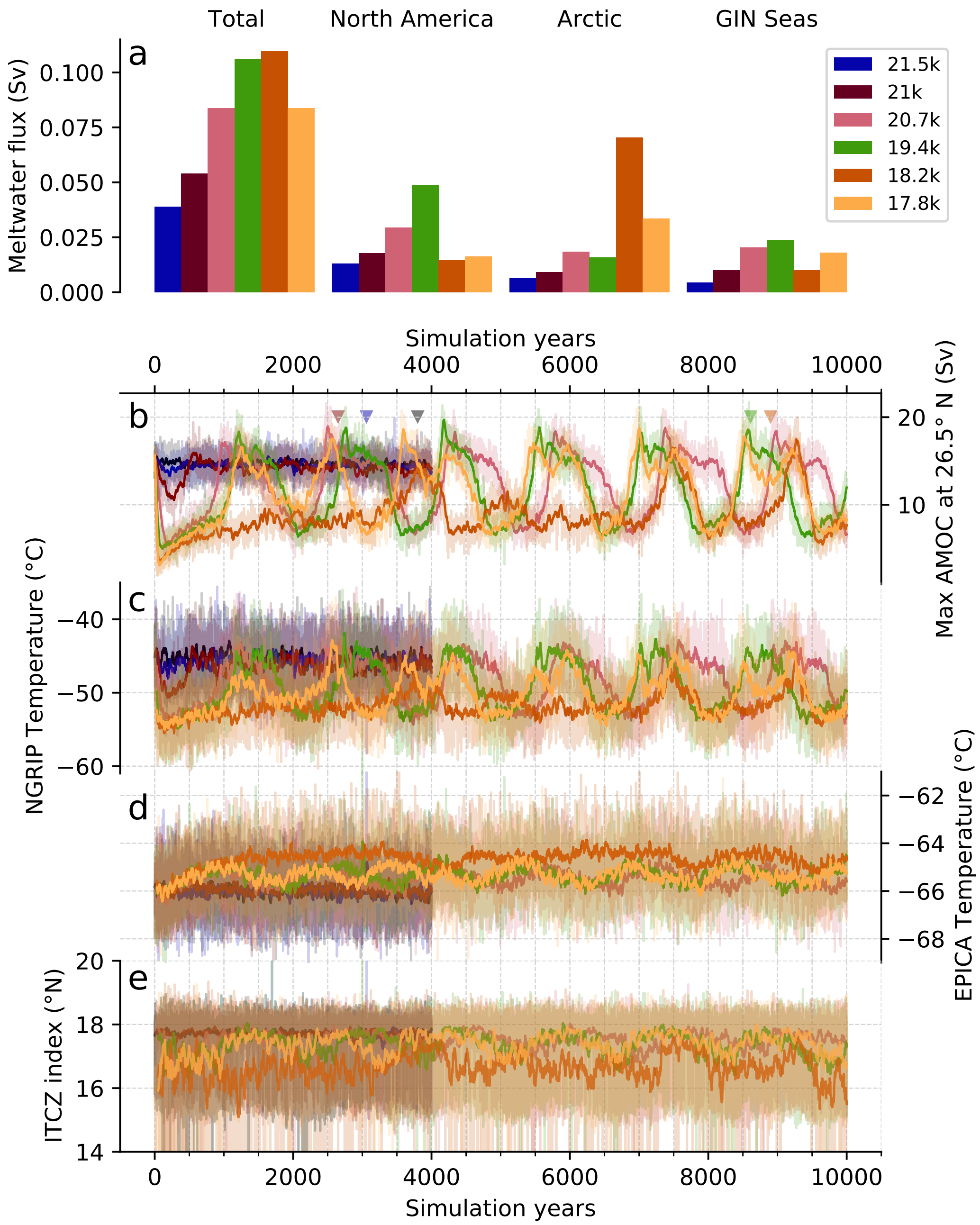

Figure 3. *a.* Snapshot experiments' total meltwater discharge and distributions, summarised (the three main regions are defined in Figure S2. *b.* AMOC index (max Atlantic overturning circulation at 26.5° N). *c.* Greenland Surface Air Temperatures at NGRIP (42.32° W, 75.01° N). *d.* Antarctica Surface Air Temperatures at EPICA Dome C (Concordia Station of the European Project for Ice Coring in Antarctica, 123.21° E, 75.06° S). *e.* Intertropical Conversion Zone index (corresponding to the simulated mean Northern extent of the equatorial rain belt, defined in section S7, inspired by Braconnot et al. (2007) and Singarayer et al. (2017)). Solid lines represent the 30-years running mean and transparent envelopes represent inter-annual variability, except for panel *e*, where the solid line is the 50-year running mean for ease of readability. Arrows indicate the date of the application of localised bathymetric-smoothing (see section S4) for *21k*, *21.5k* and *CTRL* (left to right).

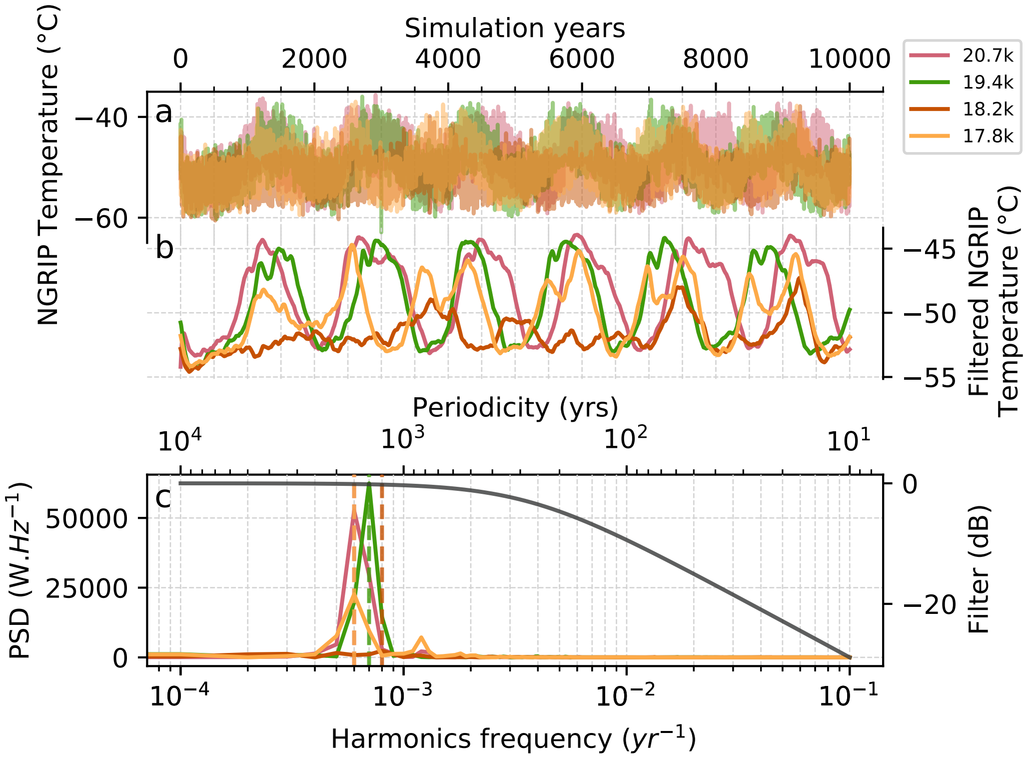

Figure 4. Spectral analysis of simulated surface air temperature through time above NGRIP (42.32° W, 75.01° N). *a.* Unfiltered signal for *oscillating* and *slow-recovery* simulations. *b.* Filtered response, using first class low-pass Butterworth filter. *c.* Power Spectral density (PSD, left hand scale) of the unfiltered signal. Dotted lines indicate the dominant frequency/period for each simulation. Grey line in panel *c* shows the bode diagram of the low-pass filter. Simulations *21.5k* and *21k* are not shown as their Fourier analysis was not conclusive.

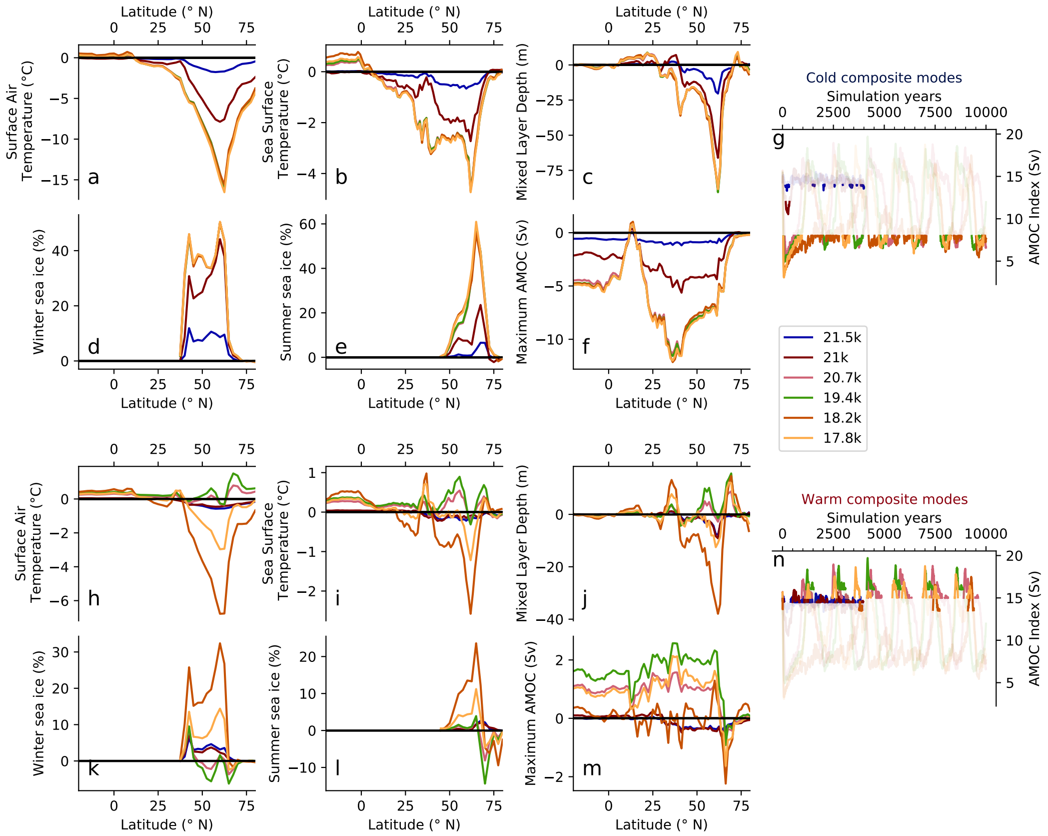

Figure 5. Composite cold and warm modes' mean zonal anomalies between the meltwater simulations and the reference state in the Atlantic (70° W – 10° E). For cold modes (top), panels show the zonally averaged *a.* surface air temperature, *b.* sea surface temperature, *c.* mixed layer depth, *d.* winter sea ice concentration, *e.* summer sea ice and *f.* maximum overturning circulation flow over the water column. For warm modes (bottom), panels show the zonally averaged *h.* surface air temperature, *i.* sea surface temperature, *j.* mixed layer depth, *k.* winter sea ice concentration, *l.* summer sea ice and *m.* maximum overturning circulation flow over the water column. For orientation, an AMOC time series highlighting the periods contributing to the composite cold and warm modes is shown by panels *g* and *n*, respectively.

In a second regime of its own, the AMOC in *18.2k* almost entirely collapsed as soon as meltwater was discharged, causing Northern Hemisphere cooling and a significant shift southward of the ITCZ (Figure 3*e*). Short recovery episodes occur at irregular intervals through the 10,000 years of the simulation, but they cannot be sustained for longer than a few hundred years. This simulation will be labelled a *slow-recovery simulation*.

Finally, through Fourier analysis (Figure 4*c*; section 2.3) we identify three *oscillating simulations* (*20.7k*, *19.4k* and *17.8k*). They switch to a cold state at the onset of the run and continue in a quasi-oscillating regime between cold (near collapsed AMOC, similar to *18.2k*) and warm modes (equivalent or stronger AMOC with respect to *CTRL*). The three *oscillating simulations* behave in a similar way and will be described in more detail in the following sections.

During the cold modes (Figure 5, top), we observe strong consistency between the *oscillating simulations* and the *slow-recovery simulation*, to the point where it becomes almost impossible to distinguish between them. These experiments show a maximum in atmospheric and oceanic cooling around 60° N (Figure 5*a*, *b*), where sea ice cover is now present both in winter and summer after the deep water formation sites vanished (Figure 5*c*–*e*). A second peak of winter sea ice is noticeable around 40° N, but is not as clear in summer sea ice nor mixed layer depth. This second peak corresponds to the closing of the warm water corridor off the western coast of Europe and the spread of winter sea ice in this region. It is around these latitudes that we observe a maximal reduction of the thermohaline circulation by up to 14 Sv (Figure 5*f*). Because of the loss of convection at high latitudes, the upper cell of the AMOC largely dwindles north and south of 20° N. The *warm simulations* follow a similar pattern of disruption, but are not nearly so intense. In *21.5k*, the soft decline of the AMOC is consistent with a shift southward of the convection sites, most likely resulting from a slightly cooler North Atlantic climate (Figure 5*c*–*f*). Summer sea ice extent is very similar to *CTRL* values in *21.5k*. The slowdown of the AMOC is slightly more pronounced in *21k*, with a clear reduction of sea surface temperature by as much as 2°C and of surface air temperature by up to 7.5°C at high latitudes (Figure 5*a*, *b* and *f*). It is remarkable that *21k*'s winter sea ice expansion and mixed layer depth shallowing are comparable to the *oscillating* and *slow-recovery* simulations, demonstrating an increased seasonality compared to the reference *CTRL* state (Figure 5*c*–*d*).

We do not observe such consistency between simulations during their warm modes (Figure 5, bottom). They all show a significant recovery from the cold modes, but it is impossible to underline a single common behaviour. For example, despite a couple of periodic recovery phases of the AMOC in *18.2k* (around 3,500, 7,000 and 9,000 years into the run; Figure 3*b*), the warm modes remain in a relatively cold-climate state (Figure 5 *h*–*i*). At 60° N, sea surface temperatures are down 2°C and surface air temperatures drop by 6.5°C compared to *CTRL*. Shallower mixed layer depths around the same latitude indicate that Iceland/Irminger Basin and Labrador Sea convection sites are still greatly disrupted (Figure 5*j*). This keeps the sea ice edge far south in both summer and winter (Figure 5*k*–*l*). On the other hand, the strong modes of the *warm simulations* are comparable to the reference *CTRL* state, only with a slightly cooler climate (Figure 5*i*–*j*) and a slower overturning circulation above 30° N in the Atlantic (Figure 5*m*). Amongst the *oscillating simulations*, the AMOC index is stronger than in *CTRL* (Figure 5*m*), increasing by as much as 2 Sv in the subpolar region. However, for the most part, these simulations also show disparities in key climate descriptors, at least in terms of amplitude of the anomalies. Surface temperatures and sea ice extent in *18.2k* and *17.8k* bear the closest resemblance across this subset of simulations, whereas temperatures and sea ice extent in *20.7k* and *19.4k* indicate a slightly warmer climate (Figure 5*h*, *i*, *j*, *l*) despite there being stronger ocean convection in *17.8k* (Figure 5*j*). The oscillating simulations all show an increase in winter sea ice around 40° N compared to *CTRL*, corresponding to the narrowing of the warm water corridor along the coast of western Europe (Figure 5*k*). In summary, none of the simulations exactly returned to the *CTRL* reference climate during their respective warm modes, indicating significant regional legacy induced by the imposed meltwater patterns.

5 Influence of the meltwater discharge

Abrupt climate changes are triggered by constant meltwater discharge in our simulations. Yet, we observe a strong non-linearity between the climate response and the location and the influx of freshwater, (Figure 3). For instance, despite a similar total influx of about 0.1 Sv, *19.4k* is tipped into an *oscillating* regime while *18.2k* ends up in a *slow-recovery* state. On the other hand, *20.7k* and *19.4k* display similar oscillating dynamics even though the total flux is around 20% weaker in *20.7k* than *19.4k*. All of this hints at the importance of differences in the meltwater discharge pattern. However, whilst the spatial distribution of freshwater forcing is comparable between *21k* and *20.7k*, only

20.7k manages to generate oscillations, demonstrating that while the spatial distribution of the freshwater flux is an important control on the oceanic response, there is also a sweet spot in the perturbation conditions where the forcing needs to be strong enough to trigger a switch to stadial states, but not so strong that it (semi-)permanently suppresses a recovery as in *18.2k*.

A threshold is reached in the cold climate modes, when the climate cools so far that it becomes insensitive to further meltwater discharge (Figure 5). Independent of the forcing, all simulations produce a similar spatial pattern in temperature, sea ice and deep water formation layout during the weak phases, but only when the surface atmosphere cools by as much as 15°C does the climate cross a tipping point where the cold phases can be sustained for a few hundred years. This corresponds to the vanishing of all oceanic convection north of 40° N (Figure 5*c*), resulting in an almost collapsed AMOC (up to 12 Sv weaker) in the North Atlantic (Figure 5*f*). When all the deep water formation sites have vanished, the response becomes decoupled from the forcing and only smaller regional effects can be induced. This phenomenon resonates with the conclusion of Smith and Gregory (2009).

During warm AMOC modes, the differences in climate response seem to be driven by the regional patterns of discharge. The influence of the different forcings can be tracked by creating a composite of warm modes mixed layer depth anomalies (Figure S10), producing clearly distinct results from the *warm* simulations, the simulations with prevalent North American meltwater inputs (*20.7k*, *19.4k*) and the simulations dominated by Arctic discharge (*18.2k*, *17.8k*). However, this relationship is sometimes counter-intuitive. For instance, despite having a stronger North American forcing in *19.4K* than in *18.2k* and *17.8k*, Labrador sea deep water formation is more impacted in the latter two simulations. In addition, although *18.2k* and *17.8k* have high Arctic discharge, Iceland Sea convection intensifies and the Irminger Basin convection weakens.

The effect of Arctic/GIN Seas discharge is decisive for triggering the shift from warm to cold AMOC states and leads to the strongest modification of the warm modes (Figure 5). The two most disrupted warm stages are found in *18.2k* and *17.8k*, both dominated by Arctic discharge. We infer that this comes from the relative position of other convection sites. For example, when released in the Arctic, freshwater follows the Greenland currents to reach the Irminger/Iceland basins relatively quickly, thus having undergone less dispersal than if taking a longer more circuitous route, particularly compared to meltwater entrained in the Atlantic gyres (Born & Levermann, 2010). Arctic meltwater will thus target the main deep water formation sites without having been significantly mixed into becoming warmer and saltier waters. Meltwater entering the GIN Seas should play a similar role, but the relatively low discharge in this area in our simulations makes it hard to conclude with certainty, and could explain why we do not observe a strong signal in this region (Figure S10).

Meltwater released off the northeast coast of North America have a weaker impact. Simulation *19.4k* has greater North Atlantic discharge, and similar fluxes to the Arctic and GIN Seas compared to *20.7k*. However, this increase in North Atlantic meltwater does not drive any significant ocean or climate response. Here, again, the explanation may relate to the location of the deep water formation sites. Because Labrador Sea convection is not always active, and easily shut down in our simulations, freshwater has to transit all around the subpolar gyre to target the more crucial sites in the eastern North Atlantic. By then, it has been mixed with tropical waters, weakening the forcing. The increased sensitivity to Arctic discharge compared to North American discharge in our simulations ties in with the conclusions of Roche et al. (2010) and Condron and Winsor (2012).

The existence of a sweet spot in the rate and location of meltwater discharge to the ocean for triggering climate transitions also depends on the background climate. Understanding the conditions leading to the creation of such a window of opportunity (Barker & Knorr, 2021) where abrupt climate changes can arise has been widely discussed (e.g. Brown & Galbraith, 2016; Zhang et al., 2017; Klockmann et al., 2018). Among the parameters likely to influence it, the choice of ice sheet reconstruction seems key, even more so in our simulations, as we rely on it both for the background climate state and for the meltwater forcing scenarios. In previous HadCM3-family deglaciation studies (e.g. Ivanovic et al., 2018; Gregoire et al., 2012), different ice sheets reconstructions did not yield as significant climate transitions, in spite of having a comparable baseline climate state (Kageyama et al., 2021) and magnitude of forcing. The higher temporal resolution of GLAC-1D and its treatment of the Eurasian ice sheet makes it a perfect candidate to attain this subtle balance, as hypothesised by Ivanovic et al. (2018). Precisely what about this particular reference state provides such a compliant framework for simulating AMOC oscillations may form the basis of future study, but we can reasonably hypothesise that very high-latitude meltwater is a pre-requisite for obtaining the sweet spot for producing climate variability, since without it, no oscillations are observed.

6 Bimodal warm states linked to reorganisation of deep water formation and subpolar gyre layout

Intriguingly, the *oscillating simulations* frequently undergo a double warm peak during times of strong AMOC (Figure 4*b*). This is particularly clear in the late cycles of *17.8k* (i.e. after 3,000 years), where the first peak in each warm phase is slightly cooler than the second. It indicates two distinct climate states separated by a cooling and then warming transition that is lower in amplitude than the cold-AMOC to warm-AMOC mode shift and which do not resemble the classical two-stage recovery hypothesis described by Renold et al. (2010) and Cheng et al. (2011). Changes in sites of deep water formation and the SubPolar Gyre (SPG) have been at the centre of recent studies of abrupt climate and ocean circulation changes (Li & Born, 2019; Klockmann et al., 2020) and could shed some light on this warm-mode transition stage. Hence, we next examine the relationship between the AMOC index, the longitude of the centre of Mass of the Mixed Layer Depth (MLD COM) and different physical indicators.

There is a linear relationship of roughly $1\,^{\circ}\mathrm{C}\,\mathrm{Sv}^{-1}$ between the AMOC index and the temperature at NGRIP (Figure 6*a*), demonstrating once again that AMOC index and NGRIP temperatures are interchangeable when it comes to identifying climate modes in our simulations. The intensity of the convection sites follow a similar trend (Figure 6*b*), with roughly a 10 metre increase of the maximum mixed layer depth in the North Atlantic for each degree of warming over Greenland. Both these relationships remain consistent irrespective of the specific warm modes that the *oscillating simulations* are in. However, for equivalent maximum AMOC strengths, the location of the deep water formation sites and the geometry of the subpolar gyre are clearly distinct between the two different warm modes. This is manifested in the ESPG index and latitude of the SPG COM, which split into two branches for high AMOC indices (Figure 6*c*–*d*). To further describe this phenomenon, we examine the winter and autumn convection layouts and sea ice formations for the two warm modes (Figure 7).

We follow the dynamics of the warm mode shifts by tracking the arrows in Figure 6, omitting in this analysis the initial spin-up period corresponding to values falling outside of the cycles (for example, the first 300 years in *20.7k*). Starting from a *cold* mode, the recovery of the AMOC is first fuelled by convection in the GIN Seas, with very little signal in the Irminger basin, which is covered in sea ice both in Autumn and Winter (Figure 7*a*, *b*, *e*, *f*). This results in a shift eastwards and northwards of the centre of mass of the convection sites, as indicated by a deeper mixed layer (Figure 6*h*). Although situated in an area of intense sea ice formation, the Labrador Sea deep water formation site is not always reactivated during this mode. As a result, the subpolar gyre, which is initially weak and contracted during the *cold* phase, gets slightly stronger and

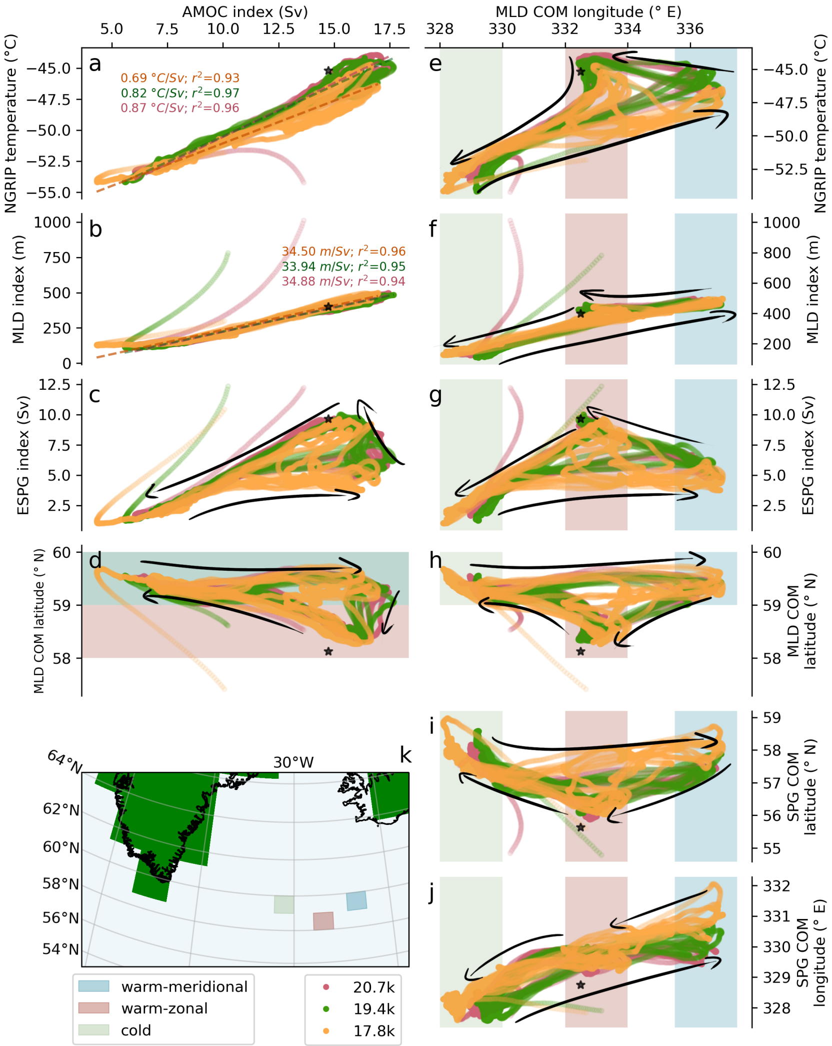

Figure 6. Phase plot of the three *oscillating simulations* showing the relationship between the maximal value of the overturning circulation in the Atlantic ocean at 26.5° N (AMOC index) and *a.* surface air temperature at NGRIP (42.32° W, 75.01° N), *b.* the maximum mixed layer depth in the Northern Hemisphere Atlantic (MLD index), *c.* the mean barotropic stream function in Eastern North Atlantic (ESPG index - see Klockmann et al. (2020) and Figure S9 for zones definition), *d.* the latitude of the centre of mass of the mixed layer depth (MLD COM latitude); and between the longitude of the centre of mass of the mixed layer depth (MLD COM longitude) and *e.* the temperature at NGRIP, *f.* the MLD index, *g.* the ESPG index, *h.* MLD COM latitude, *i.* the latitude of the centre of mass of the barotropic stream function in the subpolar region (SPG COM latitude) and *j.* the longitude of the centre of mass of the barotropic stream function in the subpolar region (SPG COM longitude). The calculation of the centre of mass is detailed in section S8. Colour shading indicates the location of the main areas of activity of the centre of mass of the mixed layer depth during the different modes, as plotted on panel *k.* Arrows indicate the direction of flow (through time) over the phase space; from green (cold) to blue (warm-meridional) to red (warm-zonal) modes. All time series were filtered following the algorithm presented in section S5. The black stars indicate the mean annual values calculated from the last 100-years of *CTRL*.

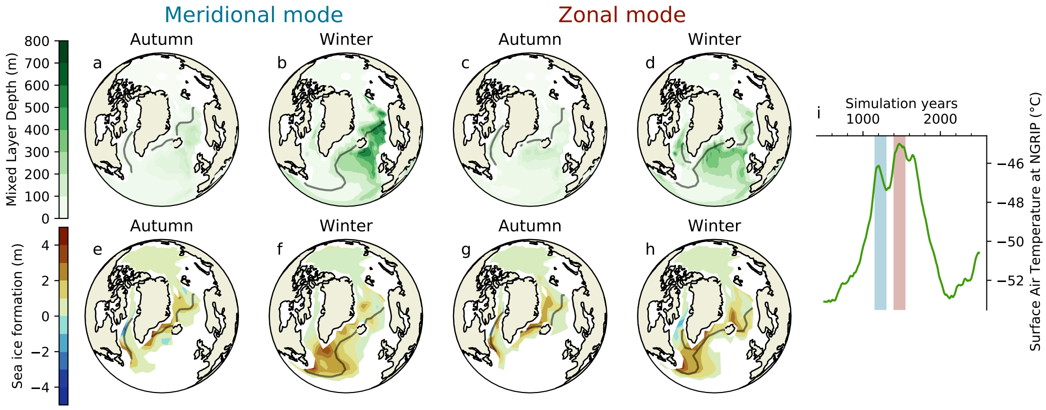

Figure 7. Seasonal Mixed layer depth (panels *a*–*d*) and sea ice formation (panels *e*–*h*) during the meridional (panels *a*, *b*, *e*, *f*) and zonal (panels *c*, *d*, *g*, *h*) warm modes in simulation *19.4k*, defined as the periods indicated in blue and red (respectively) in panel *i*. Autumn (September, October and November) and winter (December, January and February) seasonal means are shown for each variable in each of the two modes. Solid lines indicate the contour for 50% sea-ice concentration. Autumn sea ice formation is defined as the difference between November and September ice depth (in metres) and Winter between February and December. Panel *i* shows the times series of maximum Atlantic overturning circulation at 26° N in the same simulation, identifying the 'meridional' and 'zonal' modes.

extends eastwards (Figure 6*g*, *i*, *j*), entering the *warm-meridional* mode (so labelled for the disposition of the deep water formation sites during this phase). After sustaining a *warm-meridional* state for a few hundred years, the deep water formation layout is disrupted again to return to a state that resembles *CTRL* (comparing Figure 2 to Figure 7*c*–*d*), with convection occurring primarily in the Iceland/Irminger basin and the resumption of Labrador Sea deep water formation in Autumn (Figure 7*c*, *d*, *g*, *h*). The convection is distributed over a larger region and consequently the MLD index slightly dwindles. Conversely, the subpolar gyre intensifies, and moves southward and eastward (Figure 6*i*–*j*). We call this state the *warm-zonal* phase.

Interestingly, we also observe short episodes of AMOC overshoot occurring immediately after the transition from *cold* to *warm-meridional* modes. The overshoots only exist at the onset of *warm-meridional modes* and are associated with stronger convection in the GIN Seas (characterised by a more eastern value of the MLD COM longitude in Figure 7*e*).

Overall, these two *warm* states of strong AMOC are different to the two modes described by Cheng et al. (2011), where there is a transfer of deep water formation across the Atlantic from the the Labrador Sea to the GIN Seas, the latter associated with an AMOC overshot. In all three of *oscillating* simulations, a relatively strong convection is maintained throughout both warm phases of the cycle, and the deep water formation layout only shifts around the Iceland basin.

7 A good example of Dansgaard-Oeschger events?

At first glance, the *oscillating simulations* resemble recorded D-O events. From a purely descriptive point of view, presented in Table 2, we observe a periodicity (defined as the inverse of the dominant frequency in Figure 4) of between 1,540 and 1,930 years (the *18.2k* simulation has a periodicity of 1,290 years, but strictly we do not define this as an *oscillating* simulation). In terms of the duration of D-O cycles, this simulated periodicity is close to the range approximated from palaeo records (about 1,500 years) during times of regular occurrence (Thomas et al., 2009; Lohmann & Ditlevsen, 2019). As an example, we compared our simulated cycles with DO 9-11, which occurred between 44 ka BP and 40 ka BP (Figure 8). This is also very similar to the simulated events of Klockmann et al. (2020) and Armstrong et al. (2022), but two to three times longer than Peltier and Vettoretti (2014). As already discussed, *18.2k* does not qualify as an *oscillating* simulation due to its *slow-recovery* characteristics (section 4). Yet, in Figure 4, we observe two smaller peaks identified by the frequency analysis algorithm, one at $\sim$1,300 years and one at $\sim$3,500 years, which also correspond to typical D-O values (e.g. Kindler et al., 2014).

The temperature ranges between warm and cold climate/AMOC modes are similar, with changes of about 10°C recorded (Huber et al., 2006; Kindler et al., 2014) and simulated (this study) at NGRIP. From the other model studies cited in this study, only Peltier and Vettoretti (2014) obtained a similar amplitude of Greenland temperature change. Both Klockmann et al. (2020) and Armstrong et al. (2022) observed smaller transitions similar more to the 6°C amplitude of *18.2k* oscillations, but still within the lower range of reconstructed D-O events.

Our simulations also show a bipolar see-saw phenomenon (Stocker, 1998; Blunier & Brook, 2001), with a lag between the Northern and Southern Hemisphere temperature changes of around 100 years (Figure 8*a*–*b*), consistent with the suggestions of Blunier and Brook (2001) and Stocker and Johnsen (2003).

Notwithstanding these similarities with recorded D-O events, because all simulations were realised in a maximum glacial background (specifically, the LGM), the analogy between these simulations and D-O events is not straight forward. This is all the more true as we can also identify significant discrepancies between the model results and observations. For example, the simulated cycles do not match the typical shapes of D-O events (Lohmann & Ditlevsen, 2019); an abrupt warming followed by a slow cooling over a few hundred to a few thousand years within the warm phase followed by a sharper cooling to finish off the cycle. Our simulations displayed a gradual transition between the cold modes to the warm modes, and the strong-AMOC phases were maintained rather steadily for ~500 years before undergoing a slow cooling into the weak-AMOC phase of the cycle (Table 2). The warming and cooling rates are less sharp than palaeo-records suggest, with a typical rate of 3°C temperature change per 100 years, about 10 times slower than indicated in Lohmann and Ditlevsen (2019). It has to be noted that despite a potential damping of the warming and cooling rates due to the filtering, the duration of filtered signals matched the unfiltered ones well, indicating that our conclusion on the shape of the simulated cycles are not an artefact of the analytical method. Using the same model, but simulating an older time period, Armstrong et al. (2022) similarly produced an oscillating ocean/climate with warming rates not exceeding 3°C 100yrs^{-1}, in contrast to the ten times faster faster warming rates simulated by Peltier and Vettoretti (2014) and Klockmann et al. (2020) (their cooling rates overlap with ours).

Presently, we cannot categorically conclude whether or not our oscillations relate to D-O events. Differences in the shape of real and simulated D-O cycles may be explained by the quasi-idealised nature of our experiment design, specifically, the glacial maximum and fixed nature of our climate model boundary conditions/forcings (set to 21 ka, PMIP4 LGM protocol with GLAC-1D ice sheet plus the respective meltwater scenarios from the early deglaciation in GLAC-1D). The framework for our simulations was designed to simplify the identification of different behaviours in response to early deglacial meltwater forcing, mainly inspired by the hypotheses formed in conclusion to earlier work by Ivanovic et al. (2018). It provides a solid and systematic foundation for further work to study the mechanisms at play in the climate simulations presented here. However, this framework

Simulation	Periodicity (ka)	Amplitude (°C)	Warming rate (°C 100yrs^{-1})	Cooling rate (°C 100yrs^{-1})
20.7k	1.54	10.9	4.94	-3.98
19.4k	1.93	8.92	5.19	-3.72
18.2k	1.29	5.90	2.84	-3.57
17.8k	1.67	7.65	4.17	-5.28
Peltier and Vettoretti (2014)	$\sim$ 0.8	$\sim$ 10	$\sim$ 50	$\sim$ -5
Klockmann et al. (2020)	1.5 to 2.0	5 to 6	25 to 50	-5 to -35
Armstrong et al. (2022)	$\sim$ 1.5	6.5 to 8.5	$\sim$ 3	$\sim$ -3

Table 2. Typical oscillation descriptors for the *oscillating* and *slow-recovery* simulations defined from the filtered NGRIP temperature time series (Figure 4) and compared to simulations from Peltier and Vettoretti (2014), Klockmann et al. (2020) and Armstrong et al. (2022). Periodicity was identified using the analysis presented in Figure 4*c*. Amplitude was defined as the difference between the warmest and coldest point of the filtered NGRIP temperature time series Figure 4*b*. Warming and cooling rates were defined as the maximum and minimum of the derivative of the filtered NGRIP temperatures time series, respectively. Descriptors of the simulations performed by Peltier and Vettoretti (2014) were estimated from Figure 1 by Vettoretti and Peltier (2018). Descriptors of the simulations performed by Klockmann et al. (2020) were taken from their main text. Descriptors of the simulations performed by Armstrong et al. (2022) were taken from the main text and estimated from Figure 2 by Armstrong et al. (2022).

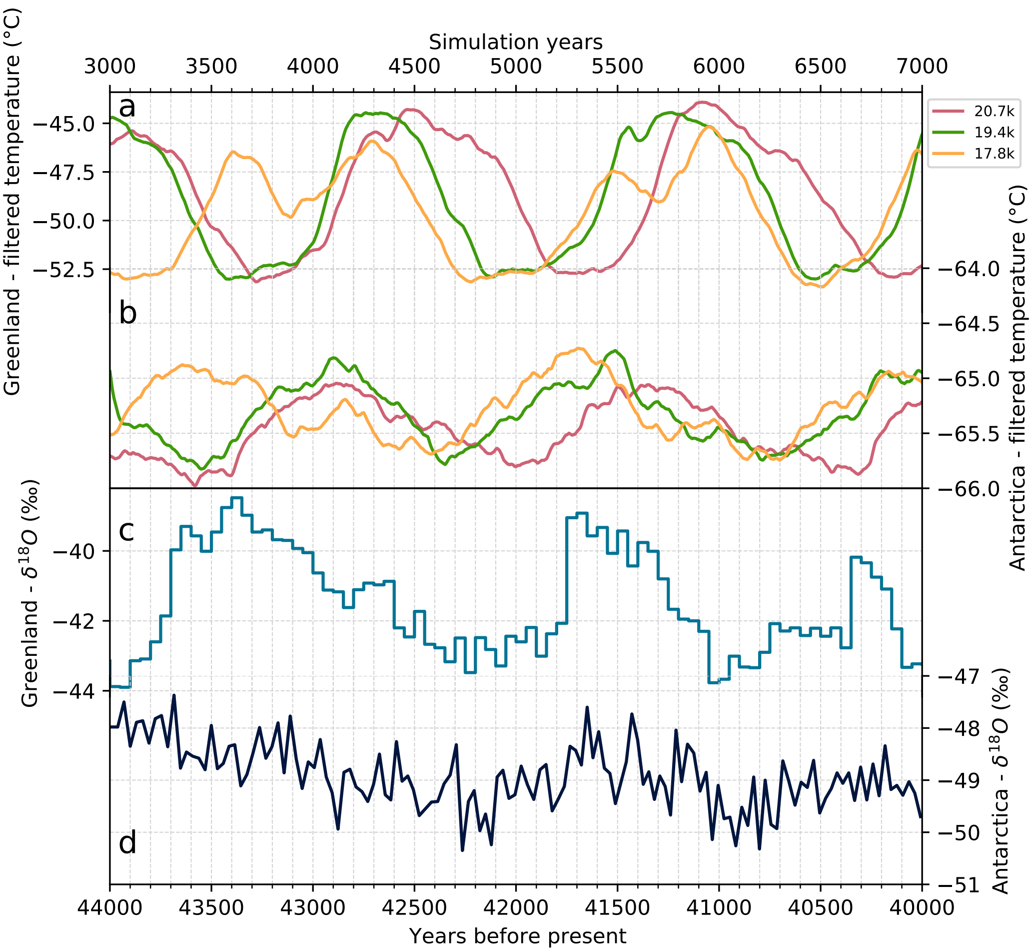

Figure 8. Temporally filtered signals of simulated surface air temperature over *a.* Greenland (NGRIP; 42.32° W, 75.01° N) and *b.* Antarctica (EPICA Dome C; 123.21° E, 75.06° S) for simulation years 3,000 to 7,000. *c.* NGRIP and *d.* EDML (Dronning Maud Land; 0.04° E, 75.00° S) $\delta^{18}O$ records between years 40,000 and 44,000 before present showing DO events 11, 10 and 9, from N.G.R.I.P (2004) and Barbante et al. (2006), respectively.

may interfere with or block some of the complex dynamics of D-O events and damp the abrupt climate changes. We indeed observe the sharpest cooling events at the onset of each simulation, indicating that the initial reorganisation of deep water formation sites in response to a change in freshwater forcing may lead to the strongest and fastest climate disruption. Also, our setup does not include feedbacks between ice sheet melt and temperature changes, which could influence the periodicity of amplitude of changes (Gregoire et al., 2016; Ivanovic et al., 2017). We therefore speculate that implementing a transient meltwater pattern consistent with what is known about past ice sheets during times of abrupt climate change could be a good way to better account for the dynamical interactions and abrupt reorganisations of the earth system, but ultimately, transient coupled climate-ice sheet simulations are needed to fully unlock the challenge of understanding D-O cycles and abrupt deglaical climate change. Finally, the discrepancies between observed D-O cycles and our simulations may also be related to weaknesses in the climate model itself. As concluded by Armstrong et al. (2022), this version of HadCM3 seems to be unable to capture the fast physics component that has been observed in (Vettoretti & Peltier, 2018), and this may be related to the representation of ocean vertical diffusion (Peltier & Vettoretti, 2014).

Furthermore, and specifically to address D-O cycles, we also need to be able to reproduce the phenomenon of oscillating weak-strong AMOC modes outside of a glacial maximum background. Marine Isotope Stage 3 (MIS3, 29–57 ka BP, (Lisiecki & Raymo, 2005), as depicted in Figure 8*c*–*d* for comparison to our results, has been often considered an appropriate candidate for such studies. Stadial conditions then were warmer than at the LGM, and it contains the most regular occurrences of D-O events recorded. Currently, one major challenge for setting off such a suite of simulations is that our oscillations are triggered by meltwater discharge, with a strong dependency on nuanced differences in the rate and location of freshwater inputs to the ocean. Thus, robust and detailed constraints on ice sheet extent are necessary to design an appropriate model experiment.

Apart from Heinrich stadials, MIS3 was not a time when ice sheet melting is thought to have been particularly strong, and there is likely to have been a lower meltwater flux than in our *oscillating simulations* (Hughes et al., 2016; Batchelor et al., 2019). However, DO 15-17 are believed to have been associated with changes of ice sheet extent (Lambeck, 2004), and because their shape remarkably resembles the observed oscillatory cycles of our simulations (Barbante et al., 2006; Rasmussen et al., 2016; Erhardt et al., 2019), they could be good candidates for being triggered by relatively low-levels of Northern Hemisphere ice sheet melt; it is possible that a weaker glacial climate state had a more sensitive ocean to smaller meltwater fluxes than our LGM-based simulations. However, the further back in time we go, the less constrained ice sheet extent (and geometry more broadly) is, which again poses a practical limitation for how well such a climate model experiment could be designed to explore the detail of real past events. In any case, the next step in our investigation of the relationship between our simulations and real DO cycles will be to more deeply understand the physical mechanisms triggered during the model's AMOC and Greenland temperature oscillations.

8 Conclusion

Using snapshots of the meltwater discharge derived from the GLAC-1D ice sheet history of the early last deglaciation, we produced a set of oscillating simulations in the HadCM3 climate model under glacial maximum (PMIP4 LGM) conditions. Switching regularly between cold and warm modes, this behaviour can only be attained in a narrow range of circumstances: if the freshwater forcing is too strong and/or applied in a particularly sensitive part of the ocean, the climate cannot fully recover to a warm mode and stays cold for the majority of the simulation. On the other hand, if the forcing is too weak, the transition to a cold mode is incomplete and the system rebounds to stay permanently in a warm state.

Understanding what can be defined as a 'weak' or 'strong' freshwater flux is not straightforward, as the response to the forcing is non-linear. All simulations are in the range of plausible magnitudes for meltwater discharge over the last deglaciation — although the scenarios are made quasi-idealised by sustaining a constant meltwater discharge over thousands of years rather than following the transient history — but differ significantly in terms of total amplitude, and the geographical distribution of the freshwater fluxes. We identified the release regions closest to the main convection sites, namely the Nordic Seas and the Irminger/Iceland Basins, to be the most sensitive regions to meltwater forcing. One possible inference from this finding is that in spite of being smaller in size compared to the North American ice sheet, Eurasian ice sheet demise punches above its weight in having the potential to trigger abrupt climate changes over the last deglaciation. Conversely, Laurentide ice sheet meltwater is required to be much more substantial to produce similar disruptions to Atlantic Ocean circulation and climate. The impact of the initial climate and ocean state on our results must also be considered. The cold interstadial state obtained with GLAC-1D ice sheets, with a relatively weak AMOC, convection concentrated around Iceland and extensive sea ice in the Western North Atlantic, creates a background that may favour the triggering of abrupt climate changes from Eurasian ice sheet melting.

When reaching a cold mode in our simulations, the ocean and climate response to meltwater forcing becomes decoupled from the direct influence of freshwater input at high latitudes. Because the AMOC is collapsed in these states, and the susceptible North Atlantic deep water formation sites vanished, meltwater perturbations will not propagate any longer, producing only smaller, regional perturbations. Cold modes correspond to an extreme cooling of the North Atlantic, most brutally over former convection regions, with winter sea ice stretching down to Spain. We note a strong consistency in the climate change pattern of all simulations irrespective of whether they belong to the *oscillating*, *slow-recovery*, or, to a lesser extent, *warm* clusters of simulations. This is not true for the warm modes, where the response is very dependent on the meltwater forcing scenario. We observe a range of different AMOC responses, from overshooting to damping, and the resulting climate is very different for each of the simulations. In particular, the simulations dominated by Arctic discharge never manage to completely return to the reference (*CTRL*) state. We also observe a two-phase, bi-modal warm condition in the *oscillating* simulations, with shifts between deep water-formation sites from a more meridional distribution (with primary convection situated in the Nordic seas) to a more zonal configuration comparable to *CTRL*. Resumption of Labrador Sea convection is inconsistent, sometimes being short lived and intense and sometimes being wholly absent, and in these glacial simulations, its connection to the upper cell of the overturning circulation is tenuous.

In summary, while we cannot conclusively determine that the oscillations we simulate are comparable to D-O events, there are resemblances that make such a comparison inviting. D-O events were rarely observed in full-glacial climates and are not believed to be directly forced by changes in ice sheet meltwater discharge. Still, the oscillating simulations presented here offer a valuable framework to analyse further the mechanisms behind the millennial-scale variability in glacial backgrounds. Intriguingly, they provide a set of insightful simulations that are so far unique in that relatively small influxes of freshwater trigger self-sustaining AMOC and Greenland temperature oscillations from glacial *maximum* conditions with a very large LGM North American ice sheet. Finally, the presented simulations pave the way for further study of ice sheet meltwater as a *trigger*, but not a direct *driver*, of abrupt climate changes within the last deglaciation.

9 Data and code availability

The new model data presented here will be available from the University of Leeds Research Data repository upon acceptance of the article. The presented plots realised were produced using the matplotlib python library and some of the colormaps were taken from Crameri et al. (2020). The code is available on github at `https://github.com/Olnavy/ROME2022_paleoceanography_oscillations`, it relies on the packages *pyleaoclim_leeds v1.0* (`https://github.com/Olnavy/pylaeoclim_leeds`) and *mw_protocol v1.0* (`https://github.com/climate-ice/mw_protocol`), also created by the main author.

Acknowledgments

YMR was supported by the Leeds-York-Hull Natural Environment Research Council (NERC) Doctoral Training Partnership (DTP) Panorama under grant NE/S007458/1. LJG was funded by MR/S016961/1 and SST was funded by NE/T007443/1. Simulations were run on the high performance facilities ARC3 and ARC4 at the University of Leeds, UK. All climate and ocean circulation proxy data and model simulations compared to are published elsewhere, with the exception of Armstrong et al. 2022, which is still under review. Thanks are due to CEMAC for technical assistance on the numerical resource, with special thanks to Richard Rigby for assistance with the scripts for managing and processing model output. Thanks to Julia Tindal and Steve Hunter for helpful technical discussions and investigations. Thanks to Kerim Nisancioglu and Chuncheng Guo for insightful discussion about the paper, and to Heather Andres whose presentation at vEGU21 inspired Figure 6.

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