AGU100 ADVANCING EARTH AND SPACE SCIENCE

Geophysical Research Letters

RESEARCH LETTER
10.1029/2018GL077623

Key Points:
- The simulated impact of Northern Hemisphere sources of Meltwater Pulse 1a overwhelms the climatic influence of Antarctic sources
- This is because Antarctic meltwater is rapidly dissipated and the Southern Ocean is quickly resalinized by Antarctic Circumpolar water
- Therefore, surface climate signals cannot be used to fingerprint the hemispheric source of ice sheet meltwater

Supporting Information:
- Supporting Information S1

Correspondence to:
R. F. Ivanovic,
r.ivanovic@leeds.ac.uk

Citation:
Ivanovic, R. F., Gregoire, L. J., Wickert, A. D., & Burke, A. (2018). Climatic effect of Antarctic meltwater overwhelmed by concurrent Northern Hemispheric melt. *Geophysical Research Letters*, *45*. https://doi.org/10.1029/2018GL077623

Received 20 FEB 2018
Accepted 26 MAY 2018
Accepted article online 4 JUN 2018

Climatic Effect of Antarctic Meltwater Overwhelmed by Concurrent Northern Hemispheric Melt

R. F. Ivanovic[1], L. J. Gregoire[1], A. D. Wickert[2], and A. Burke[3]

[1]School of Earth and Environment, University of Leeds, Leeds, UK, [2]Department of Earth Sciences and Saint Anthony Falls Laboratory, University of Minnesota, Minneapolis, MN, USA, [3]Department of Earth and Environmental Sciences, University of St. Andrews, St. Andrews, UK

Abstract Records indicate that 14,500 years ago, sea level rose by 12–22 m in under 340 years. However, the source of the sea level rise remains contentious, partly due to the competing climatic impact of different hemispheric contributions. Antarctic meltwater could indirectly strengthen the Atlantic Meridional Overturning Circulation (AMOC), causing northern warming, whereas Northern Hemisphere ice sheet meltwater has the opposite effect. This story has recently become more intriguing, due to increasing evidence for sea level contributions from both hemispheres. Using a coupled climate model with freshwater forcing, we demonstrate that the climatic influence of southern-sourced meltwater is overridden by northern sources even when the Antarctic flux is double the North American contribution. This is because the Southern Ocean is quickly resalinized by Antarctic Circumpolar water. These results imply that the pattern of surface climate changes caused by ice sheet melting cannot be used to fingerprint the hemispheric source of the meltwater.

Plain Language Summary The fastest major sea level rise ever recorded took place 14,500 years ago, when sea level rose by 12–22 m in under 340 years. The extra water came from melting ice sheets, which stretched across North America and northern Europe as well as Greenland and Antarctica. We ran a climate model to test the impact of different meltwater contributions from Antarctica and the Northern Hemisphere ice sheets (North America, Greenland, and Eurasia). Our simulations demonstrate that northern meltwater has a much stronger and longer lasting effect on ocean circulation and climate than Southern Hemisphere melt. Consequently, northern melting overrides the impact of southern melting even when the flux of water from North America is only half the magnitude of the Antarctic flux. This means that past climate records cannot be used to identify the contribution of meltwater from different ice sheets: the northern signal can override the southern signal.

1. Introduction

Meltwater Pulse 1a (MWP1a) is the largest rapid sea level rise ever recorded. It is characterized by 12–22 m global mean sea level rise in less than 340 years approximately 14.5 thousand years ago (ka; Deschamps et al., 2012). At the time, the Earth was undergoing a transition from full glacial conditions (at the Last Glacial Maximum, 21 ka) to the current interglacial state. Ice sheets that covered much of North America and northeast Europe were melting away (Dyke, 2004; Hughes et al., 2016), while an expanded Antarctic ice sheet may also have been losing mass (Bentley et al., 2014).

We do not know the precise origin of MWP1a nor exactly how much of it came from the northern ice sheets (mainly North America and Eurasia) or Antarctica. Some favor dominance of a southern source (e.g., Bassett et al., 2005; Carlson, 2009; Clark et al., 1996, 2002; Weaver et al., 2003), but ice sheet modeling suggests Antarctica contributed at most 2 m global mean sea level rise to this event (Golledge et al., 2014), and other lines of research suggest that there was likely a large contribution from the Northern Hemisphere (e.g., Gregoire et al., 2012, 2016; Peltier, 2005; Tarasov & Peltier, 2005; Tarasov et al., 2012). Moreover, a growing body of evidence from far-field sea level records supports a mixed contribution from both hemispheres (Gomez, 2015; J. Liu et al., 2016).

Recently, Ivanovic et al. (2017) investigated the climatic impact of a plausible Northern Hemisphere contribution to MWP1a, demonstrating that the Cordilleran-Laurentide ice-saddle collapse (Gregoire et al., 2012, 2016) would deliver freshwater to the North Atlantic, mostly via the Gulf of Mexico and Beaufort Sea. In the climate model, this causes sites of North Atlantic Deep Water (NADW) formation to become increasingly (vertically) stratified, which weakens the Atlantic Meridional Overturning Circulation (AMOC) by 6 Sv (40%). The associated reduction in poleward heat transport produces widespread Northern Hemisphere cooling of 1–5 °C. However, previous work has suggested that southern meltwater would have the opposite effect by freshening Antarctic Intermediate Water, thus enhancing the Atlantic meridional density gradient and strengthening AMOC (Weaver et al., 2003). This latter hypothesis is nuanced by studies showing a nonlinear response to Antarctic meltwater, depending on its magnitude and rapidity of dispersal from the Southern Ocean (e.g., Menviel et al., 2010; Swingedouw et al., 2009). Therefore, the question was posed as to whether a freshwater contribution from Antarctica would reduce, curtail, or completely override the climatic impact of the northern meltwater (Ivanovic et al., 2017).

The climatic influence of combined northern and southern meltwater events are important to understand because several abrupt Northern Hemisphere climate changes are thought to have taken place around the time of MWP1a, including the Bølling Warming (Buizert et al., 2014) and the Older Dryas cooling (Menviel et al., 2011). Uncertainty in the precise chain of ice-climate interactions makes it difficult to establish the role of MWP1a in these climate changes (Ivanovic et al., 2017). Yet in a warming world with melting ice sheets (McMillan et al., 2014, 2016), it is becoming ever more important to better understand how freshwater fluxes influence ocean circulation and climate. Using the MWP1a case study, this manuscript directly tackles the question of how mixed contributions of meltwater from the Northern and Southern Hemisphere ice sheets impact climate. We do this by running general circulation model (GCM) simulations of four meltwater scenarios (with different volumes and timings), constructed from the latest literature.

2. Methodology

2.1. Model Description

The GCM we use is the BRIDGE version of the UK Met Office's Hadley Centre Coupled Model version 3 (HadCM3; Valdes et al., 2017). It comprises an atmosphere GCM with a horizontal resolution of 2.5° × 3.75° and 19 hybrid-coordinate vertical layers (Pope et al., 2000). It includes the MOSES2.1 (Met Office Surface Exchange Scheme version 2.1) land-surface scheme coupled to the TRIFFID (Top-down Representation of Interactive Foliage and Flora including Dynamics) vegetation model (Cox, 2001). The rigid-lid ocean GCM has a more refined horizontal resolution of 1.25° × 1.25° and 20 vertical layers (Gordon et al., 2000). Because the ocean volume cannot vary, hydrological fluxes (including meltwater pulses) are represented by virtual salinity fluxes. Thus, steric behavior is not simulated, but its effect on freshwater propagation pathways in this study is likely to be small because the fluxes are small compared to ocean currents (<0.1 Sv; Yin et al., 2010) and multicentennial in length.

2.2. Experiment Design

The climate model is based on a 15 ka equilibrium simulation by Singarayer et al. (2011), for which the boundary conditions have since been updated (Argus et al., 2014; Berger, 1978; Loulergue et al., 2008; Lüthi et al., 2008; Peltier et al., 2015; Schilt et al., 2010) according to the PMIP4 (Paleoclimate Intercomparison Project phase 4) last deglaciation protocol (Ivanovic et al., 2016). For reference, the PMIP4 Last Glacial Maximum protocol (Kageyama et al., 2017) is consistent with the last deglaciation protocol at 21 ka, and the updated 15 ka simulation is in the same family of experiments used by Swindles et al. (2018) and Morris et al. (2018). The total spin-up for this equilibrium part of the simulation was 2,250 years, by which time the model had reached near-steady state.

The transient freshwater forcing used in our simulations represents ice sheet melt and is made up of three consecutive stages (Figure 1). The first stage contains only the *background* flux (run for 525 years), which represents the longer-term global deglaciation of the ice sheets calculated by running the ICE-6G_C (VM5a; Argus et al., 2014; Peltier et al., 2015) ice sheet thickness history through a high-resolution drainage network model (Wickert, 2016; Wickert et al., 2013). The second stage contains the *background* flux and the meltwater pulses, which represent the faster changes in sea level at MWP1a. For these, we have constructed four simplified meltwater pulse scenarios (*hiN_loS*, *hiN_leads_loS*, *hiS_loN*, and *hiS_leads_loN*) according to geological constraints, runoff routing and dynamical modeling of ice sheets, sea level fingerprinting, and climate modeling (Golledge et al., 2014; Gregoire et al., 2016; Hughes et al., 2016; Ivanovic et al., 2017; J. Liu et al., 2016; Patton et al., 2017; Weaver et al., 2003). The scenarios are designed to test the climatic impact of a range of feasible MWP1a sources, including differences in the timing of the

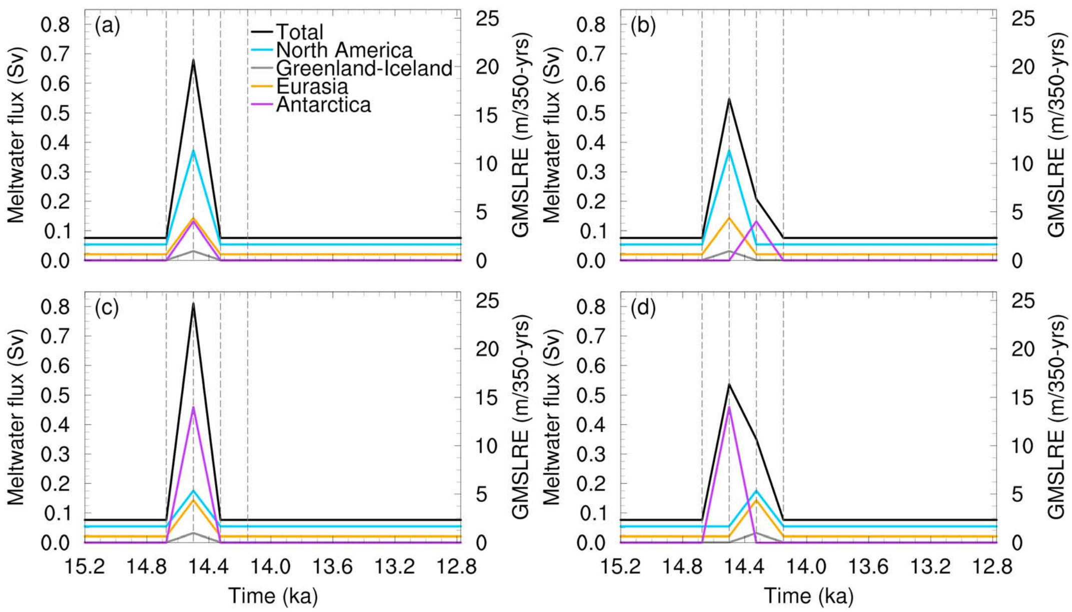

Figure 1. Meltwater Pulse 1a scenarios. (a) *hiN_loS*, (b) *hiN_leads_loS*, (c) *hiS_loN*, and (d) *hiS_leads_loN*. The vertical dashed lines mark changes in the freshwater forcing. See Text S1 for details of how these were constructed.

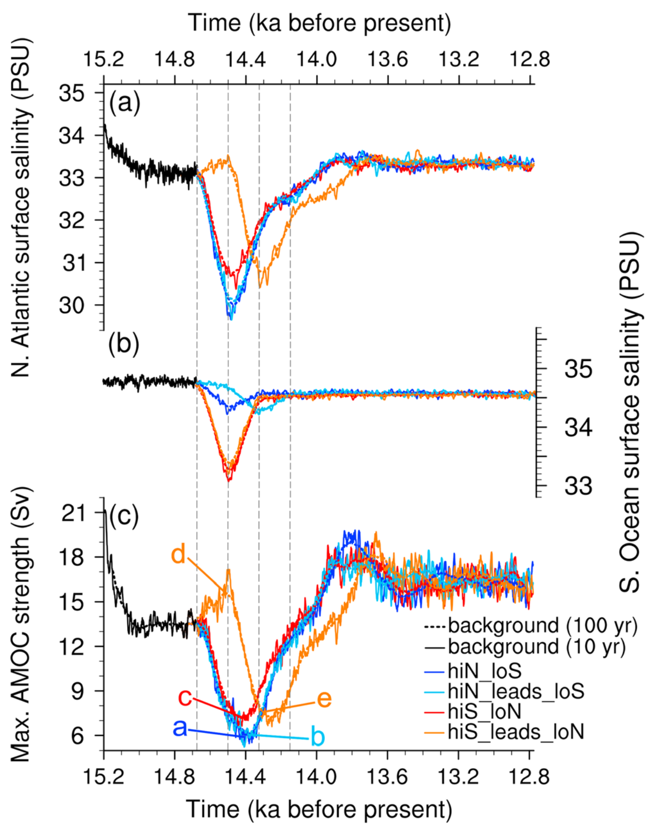

Figure 2. Simulated ocean changes. (a) North Atlantic mean surface salinity (50 to 80°N, 80°W to 20°E); (b) Southern Ocean mean surface salinity (50° to 80°S, 60°W to 0°); (c) maximum Atlantic Meridional Overturning Circulation (AMOC). The vertical dashed lines are the same as Figure 1. Labels a–e correspond to Figure 3 panels.

hemispheric source. They were run for 350–525 years, depending on the scenario. The third stage is a return to only the *background* flux (run for 1,375–1,550 years, depending on the timing of the meltwater pulses), which enables us to examine the model's recovery from the meltwater pulses and their quasi-equilibrium final state. Thus, the total length of the simulations (not including the spin-up period) is 2,425 years. Complete details of the model setup and experiment design, including an explanation of how the *background* and four meltwater pulse fluxes were derived, are given in Text S1.

3. Results

Consistent with the ICE-6G_C (VM5a) ice sheet reconstruction from 15 ka, there is no Antarctic meltwater in the *background* flux, with the largest contributions coming from North America and Eurasia, most of which drains to the Atlantic Ocean (including via the Gulf of Mexico, Arctic Ocean, and Nordic Seas). This *background* ice melt in the initial 525 years of the simulations increases shallow water buoyancy in the mid-high latitude North Atlantic sector (e.g., Figure 2a), reducing NADW formation and slowing AMOC. After ~300 years, the AMOC steadies, having weakened by ~7.5 Sv (36%; black line in Figure 2c), with a maximum strength of 13.5 Sv.

In the *hiN_loS* and *hiN_leads_loS* scenarios, the increase in Northern Hemisphere ice melt further weakens AMOC by 7.5 Sv (another 56% from the *background* flux, with maximum weakening reached ~100 years after peak northern melt; Figure 2c) through its influence on vertical stratification in regions of NADW formation. Because of the inclusion of Eurasian and Greenland-Iceland meltwater in this set of simulations, this is slightly larger than the effect simulated by Ivanovic et al. (2017) with only North American meltwater (6 Sv weakening). However, the impact on climate is very similar: poleward heat transport in the Atlantic is

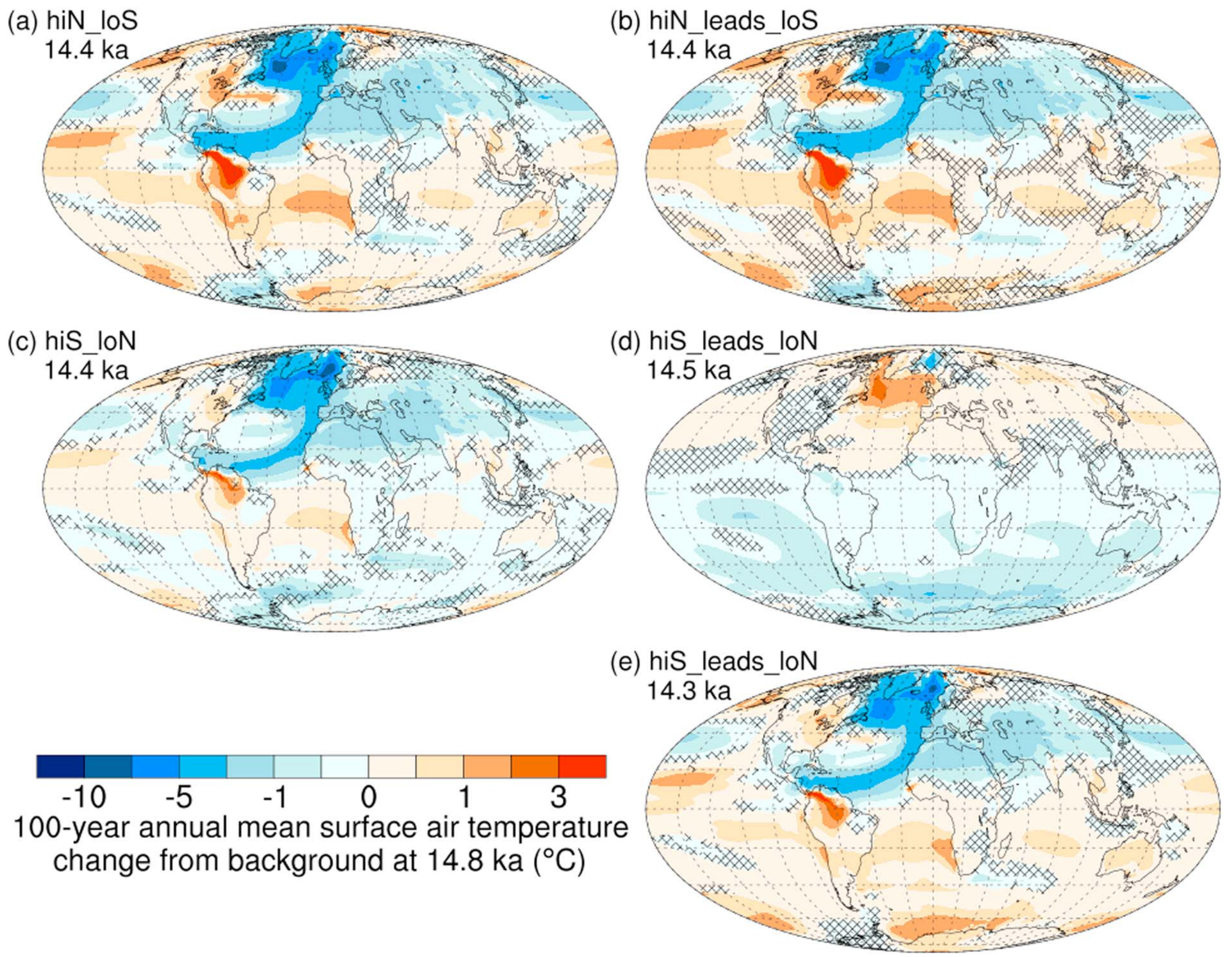

Figure 3. Simulated surface air temperature anomalies from *background* (14.8 ka). Panels correspond to labels a–e in Figure 2. Cross-hatching indicates significance below 99% confidence using a Student's *t* test.

reduced, leading to strong, widespread Northern Hemisphere cooling (with seasonal sea ice feedbacks that insulate the Nordic-Irminger-Labrador Seas in the Boreal winter and increase surface albedo in those regions during Boreal summer) and weak Southern Hemisphere warming (Figure 3). There is winter warming over eastern North America caused by high pressure over the North Atlantic (as shown by Ivanovic et al., 2017). The North Atlantic receives less precipitation (Figure S1) due to lower air temperatures and expanded sea ice (Figure S2), which both act to reduce evaporation, and the east Pacific-Atlantic Intertropical Convergence Zone (ITCZ) is displaced southward toward the anomalously warmer hemisphere. See Ivanovic et al. (2017; section 3) for details of these processes and a discussion of the associated warming and drying that takes place in the Amazon rainforest. Reduced convection in the high latitudes leads to intense Arctic subsurface warming (up to 5 °C and down to 2,200 m; e.g., Figure S3. All presented anomalies are calculated from the 100-year annual running means unless otherwise stated). This reduces proximal sea ice volume, and the Arctic surface air heats up by 0.5–2 °C when AMOC is at a minimum.

The southern melt included in *hiN_loS* and *hiN_leads_loS* has very little lasting impact on ocean circulation or climate. This is particularly evident when compared to the previous study (e.g., Figure 2 in Ivanovic et al., 2017), which had no Antarctic meltwater, yet showed remarkably similar results. It is explained by the quick dissipation of the Southern Ocean freshwater anomaly by the Antarctic Circumpolar Current (Figure 2b); the rate and magnitude of surface freshening/resalinization directly correspond to the rate and magnitude of the Southern Ocean meltwater forcing, with a smaller long-term effect driven by AMOC. Consequently, Antarctic melting produces only a localized, immediate, and short-lived reduction in deep convection near the sites of freshwater input, causing a regional drop in sea surface temperatures that is similarly brief compared to the northern high-latitude climate anomalies (Figure S4). This is in agreement with previous studies using HadCM3 (Richardson et al., 2005) and other models (Ma & Wu, 2010; Menviel et al., 2010; Stouffer et al., 2007; Swingedouw et al., 2009; Trevena et al., 2008), which show a regional cooling of the surface ocean in response to Antarctic meltwater discharge, and that through a combination of rapid mixing by the Antarctic Circumpolar Current, northward Ekman pumping (driven by the westerlies), and enhanced sea ice production, the shallow Southern Ocean is resalinized (and warmed) within a few years or decades of reducing the freshwater flux. Even when the Antarctic melting is delayed by 175 years (*hiN_leads_loS*), the Southern Ocean freshwater pulse begins only ~50 years after the maximum reduction in AMOC strength, when the influence of northern melt remains strong. Thus, the localized and transient nature of the signal combined with the small magnitude of the Antarctic meltwater flux in *hiN_loS* and *hiN_leads_loS* (Figures 1a and 1b) mean that any short-lived effect is easily overridden by the stronger and longer lasting response to northern melt.

For *hiS_loN*, the larger contribution from Antarctica slightly reduces the maximum effect on AMOC (by 1 Sv), but there is little in the surface climate to distinguish this simulation from *hiN_loS* and *hiN_leads_loS* except for ~1 °C Southern Ocean surface cooling that lasts for the duration of the forcing (e.g., Figure S4). Even with the larger contribution from Antarctica, the effects of the southern freshwater forcing are quickly mixed away due to the relatively local and fast influence of Antarctic melt versus the wider-reaching and more prolonged response to reduced NADW formation, as described for *hiN_loS* and *hiN_leads_loS*.

The influence of Antarctic meltwater on ocean circulation and surface climate is most apparent in *hiS_leads_loN*. The Antarctic meltwater increases vertical stratification in the upper water column near sites of input, weakening convection and reducing Antarctic Bottom Water formation by 1–3 Sv (Figure S5). Because the intermediate-deep Southern Ocean is relatively warm, inhibited mixing with underlying water causes regional surface cooling of up to 2 °C (Figure 3) and the northward expansion of sea ice by a few degrees. Another impact of the reduced mixing is subsurface warming of 0.5–2 °C at depths of 200–1,600 m as heat builds up in the intermediate ocean (e.g., Figure S6). This slightly weakens the density-depth gradient and partially (though quickly) damps the effect of surface freshening on vertical mixing, although the impact of the warming on water density is very minor compared to the freshening. Thus, the subsurface warming is of secondary importance for preventing longer term changes in ocean circulation compared to the role of the Antarctic Circumpolar Current in displacing the freshwater. The warming could have a more influential role as a positive feedback mechanism to West Antarctic melting by eroding the basal ice shelf (Golledge et al., 2014; Menviel et al., 2010). However, without a coupled ice sheet in our simulations, there is no enhanced melting, and the imposed Antarctic freshwater flux is vastly insufficient for the associated shallow salinity anomaly to be advected to sites of NADW formation, despite its rapid dispersal. Therefore, in the absence of a northern meltwater pulse (which inhibits NADW formation), Southern Ocean freshening acts to weaken Antarctic Bottom Water formation (Figure S5; Swingedouw et al., 2009), deepening the Southern Ocean and South Atlantic pycnocline, which slightly enhances the subsurface-meridional density gradient (Figure S7; Swingedouw et al., 2009; Weaver et al., 2003), causing AMOC to strengthen by 2 Sv (Figure 2). This produces a brief period of widespread Northern Hemisphere warming (mostly $< +1$ °C, but up to +3 °C in the Labrador Sea-Iceland-United Kingdom sector, Figure 3, where NADW formation strengthens and sea ice cover reduces) and corresponding Southern Hemisphere cooling (<2 °C), which induces a weak northward shift of the ITCZ (Figure S1). The Nordic Seas undergo surface cooling where a deepening of the AMOC and also the strengthening of convection further south cause a small reduction in shallow, northward flow past the Faroe Islands. The 175-year lag in the onset of Northern Hemisphere melt in *hiS_leads_loN* delays its effect on AMOC only by the same amount of time (~175 years) and makes no difference to the maximum AMOC reduction (Figure 2, red versus orange). Therefore, by ~14.3 ka, very similar climate anomalies are observed in *hiS_leads_loN* as in *hiN_loS* and *hiN_leads_loS* ~14.5 ka (e.g., Figure 3).

The climate change in *hiS_loN* combines the effects seen in *hiS_leads_loN* ~14.5 ka with those in *hiN_loS* and *hiN_leads_loS* ~14.4 ka (Figure 3). The AMOC slowdown causes widespread northern cooling of 1–5 °C, with a particularly strong signal in the North Atlantic (that is amplified to >5 °C cooling in regions with enhanced sea ice formation), and winter warming over eastern North America. In parts of the Southern Hemisphere, surface cooling overrides the weak warming induced by northern meltwater, resulting in a much patchier and damped temperature signal compared to the other simulations (Figure 3c). Warming in the southern tropics and subtropics pulls the tropical rain belt south (Figure S1), but the anomaly is not as strong as in *hiN_loS* and *hiN_leads_loS* because of the weaker interhemispheric temperature change.

The mechanisms that govern the model's recovery from northern meltwater sources are relatively straightforward. Reducing the freshwater forcing allows AMOC to strengthen as upper water column buoyancy is lost, stimulating convection. This enables North Atlantic salinity to be replenished from more southerly latitudes, positively feeding back to the recovery. In all simulations, it takes around 1,000 years for AMOC to strengthen and stabilize after peak northern melting (14.5 ka). There is an initial period of most rapid recovery in the first 175 years (+2.5 Sv per 100 years) after maximum weakening, when the forcing is ramped-down to *background* levels of meltwater discharge. After this, a slower strengthening of AMOC takes place (~1.5 Sv per 100 years) as relatively salty, shallow subtropical water continues to be transported north and the meltwater forcing remains constant at *background* levels. Thus, NADW formation is invigorated, providing a positive feedback to AMOC strengthening.

In the simulations, ocean circulation imports 0.2–0.4 Sv of freshwater (net flux) to the Atlantic basin at 32° S, with the lower values occurring during periods with larger northern freshwater forcing and decreased AMOC, and the higher values occurring during weak northern forcing, larger southern meltwater forcing, and stronger AMOC. This net freshwater import negatively feeds back to a change in AMOC strength; when the northern meltwater forcing reduces and AMOC recovers strength, an increasing volume of freshwater is imported from the South Atlantic and carried poleward, where it raises shallow water buoyancy and contributes toward stabilizing the water column. Thus, eventually, AMOC stops strengthening and reaches a new equilibrium state, but only after deeper convection is initiated north of 60° N and Labrador Sea deep-water formation switches on for the first time (~14 ka), leading to a small AMOC "overshoot" of 1.5–2.5 Sv. Consequently, AMOC stabilizes with a maximum strength of 16.5 Sv, which is 3 Sv stronger than before the meltwater pulses. This makes the North Atlantic warmer (+2 °C surface air temperature), which reduces sea ice cover in the Labrador and Nordic Seas, and also increases evaporation. Thus, AMOC is stronger after the meltwater pulses than before them, even though the same *background* forcing is applied. This highlights the implicit complexity of the Atlantic circulation and shows that (modeled) AMOC can exist in multiple stable states. In this case, it was tipped between equilibria by the short, idealized meltwater pulse events. Therefore, to consider AMOC only in terms of *bi*stability, where AMOC is either "on" or "off" (i.e., collapsed or very weak), is an oversimplification (Wunsch, 2010).

4. Discussion and Conclusions

Understanding the competing effects of different MWP1a sources is crucial for knowing why abrupt climate changes took place during the last deglaciation, such as the Bølling Warming and Older Dryas Cooling. Using an intermediate complexity model with a strong hysteresis behavior, Weaver et al. (2003) proposed that a dominantly southern-source of MWP1a could drive the Bølling Warming by enhancing the Atlantic meridional density gradient. However, this scenario has not been reproduced by GCMs and there are a number of reasons why it seems implausible:

First, the new results presented here suggest that even with a large Antarctic contribution, the climatic effects of Southern Hemispheric melt would likely have been overridden by Northern Hemisphere melt unless the northern contribution to MWP1a was insignificant in magnitude or absent entirely. This is currently very unlikely based on geological and modeling evidence (e.g., Dyke, 2004; Golledge et al., 2014; Gregoire et al., 2012; J. Liu et al., 2016; Peltier, 2005; Tarasov & Peltier, 2005; Tarasov et al., 2012), but even if it were true, the climatic imprint of Antarctic meltwater forcing is short and weak, mainly because the associated freshwater anomaly is rapidly dissipated and the Southern Ocean is quickly resalinized by the Antarctic Circumpolar Current, enhanced sea ice production, and northward Ekman pumping driven by the westerlies.

Second, because a large contribution from northern ice sheets is likely, then MWP1a could have caused a strong AMOC slowdown and widespread northern cooling, as shown here and by other work (e.g., Ivanovic et al., 2017; Menviel et al., 2011). This may not be true if AMOC was already in a collapsed state (AMOC was forced to be shut down in the study by Weaver et al., 2003), in which case freshwater from the northern ice sheets cannot be as impactful on NADW (and therefore on climate) as presented here. However, since northern meltwater would still contribute toward maintaining suppressed NADW formation, it is unlikely that Southern Ocean freshening from Antarctic meltwater and any associated reduction in

Antarctic Bottom Water formation could be sufficient to kick-start a collapsed AMOC for the reasons discussed above. Moreover, the likely result of this scenario would not be a rapid reinvigoration of AMOC, but it could provide a mechanism for northern contributions to MWP1a causing only a very small perturbation to North Atlantic convection and climate. While this may help to reconcile comparisons to the Greenland ice core record (Ivanovic et al., 2017, Figure 3), it does not present a viable cause for the Bølling Warming. It is difficult to know precisely the initial condition of the ocean circulation at the start of MWP1a, because uncertainty in the timing of the rapid sea level rise and ice sheet melting means that it is not known whether it began during Heinrich Stadial 1 (e.g., ~16 ka; Scenario 2 proposed by Ivanovic et al., 2017), when AMOC decreased in strength and shoaled (note, however, that it remained active; Bradtmiller et al., 2014), or later, when AMOC was stronger (e.g., Barker et al., 2010; Bradtmiller et al., 2014; Chen et al., 2015; McManus et al., 2004; Scenario 1 proposed by Ivanovic et al., 2017). Furthermore, although there is information on whether AMOC became stronger or weaker through this period of deglaciation (e.g., see previous references), it is also not known exactly what the absolute AMOC strength was in either of these periods. Our relatively weak (13.5 Sv) and shallow (<2,000 m deep) AMOC with abyssal Antarctic Bottom Water in the Atlantic (Figure S8) is a reasonable initial condition for AMOC at the start of MWP1a if the meltwater pulse began during Heinrich Stadial 1 (Scenario 2, above), though the simulated AMOC may be strong enough to be a better representation of conditions after the Bølling Warming (Scenario 1, above), which Ivanovic et al. (2017) propose to be the more likely scenario. Moreover, even in the extreme case that AMOC was collapsed when an Antarctic pulse freshened the Southern Ocean, this freshening probably did not drive the rapid Northern Hemisphere warming ~14.5 ka due to the dominant effects of northern meltwater.

Third, MWP1a is not the only possible explanation for the Bølling Warming. It has been suggested that the abrupt climate event could have been triggered by other mechanisms, for example, due to internal variability of ocean circulation and the transport of salt and heat (Ganopolski & Roche, 2009; Knorr & Lohmann, 2007; Peltier & Vettoretti, 2014), nonlinear responses to gradual trends in climate forcings (such as CO_2; Zhang et al., 2017), or a linear response to the instantaneous cessation of North Atlantic freshwater forcing (Z. Liu et al., 2009). This in turn may have triggered a substantial North American contribution to MWP1a by melting the Laurentide-Cordilleran ice-saddle (Gregoire et al., 2016), presenting a plausible mechanism for Older Dryas cooling or at least the termination of the Bølling Warming and associated AMOC strengthening (Scenario 1 suggested by Ivanovic et al., 2017). Therefore, to fully address the chain of events and underpinning processes, other contemporaneous forcings to MWP1a will need to be considered, such as rising CO_2 and orbital changes (Ivanovic et al., 2016). Nevertheless, the mechanisms underpinning our results are robust, and although the precise location and magnitude of some of the climate anomalies vary between models (as noted by Menviel et al., 2010), the climatic response to southern melt is at least qualitatively consistent between studies.

Another important aspect of our results is the *multi*stability of an active AMOC, which is never off in our experiments. A different state is reached in the quasi-equilibrium period simulated before and after the meltwater pulses under the same boundary conditions and forcings (including the *background* meltwater flux); Labrador Sea deepwater formation was switched on for the first time during the recovery from northern melting. This highlights the importance of simulating climate transitions in the lead-up to any event or time period of interest even if the pretransition and posttransition climate forcings are the same. It testifies to more complex behavior than a *bi*modal on/off state and exposes the inadequacy of the somewhat entrenched and oversimplified paradigm of AMOC *bi*stability, providing a clear indication of how important it is to interpret oceanographic-proxy data in a sophisticated way. Furthermore, the overriding dominance of the impact of northern meltwater makes it very difficult, if not impossible, to use measured climate signals to fingerprint meltwater sources. For example, the differences in ocean circulation are likely beyond our present detection limit using geochemical tracers preserved in sedimentary archives.

In conclusion, the results presented here demonstrate that the simulated climatic impacts of northern meltwater are much stronger and longer lasting than the effects of southerly-sourced freshwater in the context of the MWP1a abrupt sea level change. Melting of the Northern Hemisphere ice sheets reduces Atlantic-Arctic shallow ocean salinity, suppressing NADW formation, which causes northern cooling and southern warming for several hundred years after the northern forcing has finished. During periods of elevated Antarctic melting, convection in the Southern Ocean is reduced and weak sea surface cooling persists in mid-high southern latitudes. However, with an active Antarctic Circumpolar Current in place, the impact of Antarctic freshwater remains constrained to Southern Ocean surface cooling that lasts only as long as the meltwater forcing. Thus, the dominant changes in AMOC and surface climate are driven by the Northern Hemisphere melt. Of even wider relevance, this suggests that in the future, as in the past, century-scale Atlantic ocean circulation may be more susceptible to Greenland melting than to freshwater from Antarctica.

Acknowledgments
R. F. I. is funded by Natural Environment Research Council (NERC) grant NE/K008536/1. Numerical climate model simulations made use of the N8 HPC Centre of Excellence (N8 consortium and EPSRC grant EP/K000225/1) and the University of Leeds Physical Climate Change Research Group high-performance computing resources. We are grateful to Piers Forster for helpful comments, to three anonymous reviewers for very constructive suggestions, and to Janet Sprintall for swift editorial handling. Presented model data are available from the University of Leeds Research Data repository, https://doi.org/10.5518/395.

References

Argus, D. F., Peltier, W. R., Drummond, R., & Moore, A. W. (2014). The Antarctica component of postglacial rebound model ICE-6G_C (VM5a) based on GPS positioning, exposure age dating of ice thicknesses, and relative sea level histories. *Geophysical Journal International*, *198*(1), 537–563. https://doi.org/10.1093/gji/ggu140

Barker, S., Knorr, G., Vautravers, M. J., Diz, P., & Skinner, L. C. (2010). Extreme deepening of the Atlantic overturning circulation during deglaciation. *Nature Geoscience*, *3*(8), 567–571. https://doi.org/10.1038/ngeo921

Bassett, S. E., Milne, G. A., Mitrovica, J. X., & Clark, P. U. (2005). Ice sheet and solid Earth influences on far-field sea-level histories. *Science*, *309*(5736), 925–928. https://doi.org/10.1126/science.1111575

Bentley, M. J., Ó Cofaigh, C., Anderson, J. B., Conway, H., Davies, B., Graham, A. G. C., et al. (2014). A community-based geological reconstruction of Antarctic Ice Sheet deglaciation since the Last Glacial Maximum. *Quaternary Science Reviews*, *100*, 1–9. https://doi.org/10.1016/j.quascirev.2014.06.025

Berger, A. (1978). Long-term variations of daily insolation and Quaternary climatic changes. *Journal of the Atmospheric Sciences*, *35*(12), 2362–2367. https://doi.org/10.1175/1520-0469(1978)035<2362:LTVODI>2.0.CO;2

Bradtmiller, L. I., McManus, J. F., & Robinson, L. F. (2014). 231Pa/230Th evidence for a weakened but persistent Atlantic meridional overturning circulation during Heinrich Stadial 1. *Nature Communications*, *5*(1), 5817. https://doi.org/10.1038/ncomms6817

Buizert, C., Gkinis, V., Severinghaus, J. P., He, F., Lecavalier, B. S., Kindler, P., et al. (2014). Greenland temperature response to climate forcing during the last deglaciation. *Science*, *345*(6201), 1177–1180. https://doi.org/10.1126/science.1254961

Carlson, A. E. (2009). Geochemical constraints on the Laurentide Ice Sheet contribution to Meltwater Pulse 1A. *Quaternary Science Reviews*, *28*(17–18), 1625–1630. https://doi.org/10.1016/j.quascirev.2009.02.011

Chen, T., Robinson, L. F., Burke, A., Southon, J., Spooner, P., Morris, P. J., & Ng, H. C. (2015). Synchronous centennial abrupt events in the ocean and atmosphere during the last deglaciation. *Science*, *349*(6255), 1537–1541. https://doi.org/10.1126/science.aac6159

Clark, P. U., Alley, R. B., Keigwin, L. D., Licciardi, J. M., Johnsen, S. J., & Wang, H. (1996). Origin of the first global meltwater pulse following the Last Glacial Maximum. *Paleoceanography*, *11*, 563–577. https://doi.org/10.1029/96PA01419

Clark, P. U., Mitrovica, J. X., Milne, G. A., & Tamisiea, M. E. (2002). Sea-level fingerprinting as a direct test for the source of global meltwater pulse IA. *Science*, *295*(5564), 2438–2441. https://doi.org/10.1126/science.1068797

Cox, P. M. (2001). Description of the "TRIFFID" dynamic global vegetation model. Hadley Centre Technical Note 24, 17.

Deschamps, P., Durand, N., Bard, E., Hamelin, B., Camoin, G., Thomas, A. L., et al. (2012). Ice-sheet collapse and sea-level rise at the Bolling warming 14,600 years ago. *Nature*, *483*(7391), 559–564. https://doi.org/10.1038/nature10902

Dyke, A. S. (2004). An outline of North American deglaciation with emphasis on central and northern Canada. In J. Ehlers & P. L. Gibbard (Eds.), *Quaternary Glaciations—Extent and Chronology—Part II: North America, Developments in Quaternary Sciences* (Vol. 2, Part II, pp. 373–424). Amsterdam, Netherlands: Elsevier. Retrieved from https://www.lakeheadu.ca/sites/default/files/uploads/53/outlines/2014-15/NECU5311/Dyke_2004_DeglaciationOutline.pdf, https://doi.org/10.1016/S1571-0866(04)80209-4

Ganopolski, A., & Roche, D. M. (2009). On the nature of lead–lag relationships during glacial–interglacial climate transitions. *Quaternary Science Reviews*, *28*(27-28), 3361–3378. https://doi.org/10.1016/j.quascirev.2009.09.019

Golledge, N. R., Menviel, L., Carter, L., Fogwill, C. J., England, M. H., Cortese, G., & Levy, R. H. (2014). Antarctic contribution to Meltwater Pulse 1a from reduced Southern Ocean overturning. *Nature Communications*, *5*(1). https://doi.org/10.1038/ncomms6107

Gomez, N. (2015). Climate science: Small glacier has big effect on sea-level rise. *Nature*, *526*(7574), 510–511. https://doi.org/10.1038/526510a

Gordon, C., Cooper, C., Senior, C. A., Banks, H., Gregory, J. M., Johns, T. C., et al. (2000). The simulation of SST, sea ice extents and ocean heat transports in a version of the Hadley Centre coupled model without flux adjustments. *Climate Dynamics*, *16*(2-3), 147–168. https://doi.org/10.1007/s003820050010

Gregoire, L. J., Otto-Bliesner, B., Valdes, P. J., & Ivanovic, R. F. (2016). Abrupt Bølling warming and ice saddle collapse contributions to the Meltwater Pulse 1a rapid sea level rise. *Geophysical Research Letters*, *43*, 9130–9137. https://doi.org/10.1002/2016GL070356

Gregoire, L. J., Payne, A. J., & Valdes, P. J. (2012). Deglacial rapid sea level rises caused by ice-sheet saddle collapses. *Nature*, *487*(7406), 219–222. https://doi.org/10.1038/nature11257

Hughes, A. L. C., Gyllencreutz, R., Lohne, Ø. S., Mangerud, J., & Svendsen, J. I. (2016). The last Eurasian ice sheets—A chronological database and time-slice reconstruction, DATED-1. *Boreas*, *45*(1), 1–45. https://doi.org/10.1111/bor. 12142

Ivanovic, R. F., Gregoire, L. J., Kageyama, M., Roche, D. M., Valdes, P. J., Burke, A., et al. (2016). Transient climate simulations of the deglaciation 21–9 thousand years before present (version 1)—PMIP4 Core experiment design and boundary conditions. *Geoscientific Model Development*, *9*(7), 2563–2587. https://doi.org/10.5194/gmd-9-2563-2016

Ivanovic, R. F., Gregoire, L. J., Wickert, A. D., Valdes, P. J., & Burke, A. (2017). Collapse of the North American ice saddle 14,500 years ago caused widespread cooling and reduced ocean overturning circulation. *Geophysical Research Letters*, *44*, 383–392. https://doi.org/10.1002/2016GL071849

Kageyama, M., Albani, S., Braconnot, P., Harrison, S. P., Hopcroft, P. O., Ivanovic, R. F., et al. (2017). The PMIP4 contribution to CMIP6 – Part 4: Scientific objectives and experimental design of the PMIP4-CMIP6 Last Glacial Maximum experiments and PMIP4 sensitivity experiments. *Geoscientific Model Development*, *10*(11), 4035–4055. https://doi.org/10.5194/gmd-10-4035-2017

Knorr, G., & Lohmann, G. (2007). Rapid transitions in the Atlantic thermohaline circulation triggered by global warming and meltwater during the last deglaciation. *Geochemistry, Geophysics, Geosystems*, *8*, Q12006. https://doi.org/10.1029/2007GC001604

Liu, J., Milne, G. A., Kopp, R. E., Clark, P. U., & Shennan, I. (2016). Sea-level constraints on the amplitude and source distribution of Meltwater Pulse 1A. *Nature Geoscience*, *9*(2), 130–134. https://doi.org/10.1038/ngeo2616

Liu, Z., Otto-Bliesner, B. L., He, F., Brady, E. C., Tomas, R., Clark, P. U., et al. (2009). Transient simulation of last deglaciation with a new mechanism for Bølling-Allerød warming. *Science*, *325*(5938), 310–314. https://doi.org/10.1126/science.1171041

Loulergue, L., Schilt, A., Spahni, R., Masson-Delmotte, V., Blunier, T., Lemieux, B., et al. (2008). Orbital and millennial-scale features of atmospheric CH_4 over the past 800,000 years. *Nature*, *453*(7193), 383–386. https://doi.org/10.1038/nature06950

Lüthi, D., Le Floch, M., Bereiter, B., Blunier, T., Barnola, J.-M., Siegenthaler, U., et al. (2008). High-resolution carbon dioxide concentration record 650,000–800,000 years before present. *Nature*, *453*(7193), 379–382. https://doi.org/10.1038/nature06949

Ma, H., & Wu, L. (2010). Global teleconnections in response to freshening over the Antarctic Ocean. *Journal of Climate*, *24*(4), 1071–1088. https://doi.org/10.1175/2010JCLI3634.1

McManus, J. F., Francois, R., Gherardi, J.-M., Keigwin, L. D., & Brown-Leger, S. (2004). Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes. *Nature*, *428*(6985), 834–837. https://doi.org/10.1038/nature02494

McMillan, M., Leeson, A., Shepherd, A., Briggs, K., Armitage, T. W. K., Hogg, A., et al. (2016). A high-resolution record of Greenland mass balance. *Geophysical Research Letters*, *43*, 7002–7010. https://doi.org/10.1002/2016GL069666

McMillan, M., Shepherd, A., Sundal, A., Briggs, K., Muir, A., Ridout, A., et al. (2014). Increased ice losses from Antarctica detected by CryoSat-2. *Geophysical Research Letters*, *41*, 3899–3905. https://doi.org/10.1002/2014GL060111

Menviel, L., Timmermann, A., Timm, O. E., & Mouchet, A. (2010). Climate and biogeochemical response to a rapid melting of the West Antarctic Ice Sheet during interglacials and implications for future climate. *Paleoceanography*, *25*, PA4231. https://doi.org/10.1029/2009PA001892

Menviel, L., Timmermann, A., Timm, O. E., & Mouchet, A. (2011). Deconstructing the Last Glacial termination: The role of millennial and orbital-scale forcings. *Quaternary Science Reviews*, *30*(9–10), 1155–1172. https://doi.org/10.1016/j.quascirev.2011.02.005

Morris, P. J., Swindles, G. T., Valdes, P. J., Ivanovic, R. F., Gregoire, L. J., Smith, M. W., et al. (2018). Global peatland initiation driven by regionally asynchronous warming. *Proceedings of the National Academy of Sciences of the United States of America*. https://doi.org/10.1073/pnas.1717838115

Patton, H., Hubbard, A., Andreassen, K., Auriac, A., Whitehouse, P. L., Stroeven, A. P., et al. (2017). Deglaciation of the Eurasian ice sheet complex. *Quaternary Science Reviews*, *169*, 148–172. https://doi.org/10.1016/j.quascirev.2017.05.019

Peltier, W. R. (2005). On the hemispheric origins of Meltwater Pulse 1a. *Quaternary Science Reviews*, *24*(14–15), 1655–1671. https://doi.org/10.1016/j.quascirev.2004.06.023

Peltier, W. R., Argus, D. F., & Drummond, R. (2015). Space geodesy constrains ice age terminal deglaciation: The global ICE-6G_C (VM5a) model. *Journal of Geophysical Research: Solid Earth*, *120*, 450–487. https://doi.org/10.1002/2014JB011176

Peltier, W. R., & Vettoretti, G. (2014). Dansgaard-Oeschger oscillations predicted in a comprehensive model of glacial climate: A "kicked" salt oscillator in the Atlantic. *Geophysical Research Letters*, *41*, 7306–7313. https://doi.org/10.1002/2014GL061413

Pope, V. D., Gallani, M. L., Rowntree, P. R., & Stratton, R. A. (2000). The impact of new physical parametrizations in the Hadley Centre climate model: HadAM3. *Climate Dynamics*, *16*(2-3), 123–146. https://doi.org/10.1007/s003820050009

Richardson, G., Wadley, M. R., Heywood, K. J., Stevens, D. P., & Banks, H. T. (2005). Short-term climate response to a freshwater pulse in the Southern Ocean. *Geophysical Research Letters*, *32*, L03702. https://doi.org/10.1029/2004GL021586

Schilt, A., Baumgartner, M., Schwander, J., Buiron, D., Capron, E., Chappellaz, J., et al. (2010). Atmospheric nitrous oxide during the last 140,000 years. *Earth and Planetary Science Letters*, *300*(1–2), 33–43. https://doi.org/10.1016/j.epsl.2010.09.027

Singarayer, J. S., Valdes, P. J., Friedlingstein, P., Nelson, S., & Beerling, D. J. (2011). Late Holocene methane rise caused by orbitally controlled increase in tropical sources. *Nature*, *470*(7332), 82–85. https://doi.org/10.1038/nature09739

Stouffer, R. J., Seidov, D., & Haupt, B. J. (2007). Climate response to external sources of freshwater: North Atlantic versus the Southern Ocean. *Journal of Climate*, *20*(3), 436–448. https://doi.org/10.1175/JCLI4015.1

Swindles, G. T., Morris, P. J., Whitney, B., Galloway, J. M., Gałka, M., Gallego-Sala, A., et al. (2018). Ecosystem state shifts during long-term development of an Amazonian peatland. *Global Change Biology*, *24*(2), 738–757. https://doi.org/10.1111/gcb.13950

Swingedouw, D., Fichefet, T., Goosse, H., & Loutre, M. F. (2009). Impact of transient freshwater releases in the Southern Ocean on the AMOC and climate. *Climate Dynamics*, *33*(2–3), 365–381. https://doi.org/10.1007/s00382-008-0496-1

Tarasov, L., Dyke, A. S., Neal, R. M., & Peltier, W. R. (2012). A data-calibrated distribution of deglacial chronologies for the North American ice complex from glaciological modeling. *Earth and Planetary Science Letters*, *315–316*, 30–40. https://doi.org/10.1016/j.epsl.2011.09.010

Tarasov, L., & Peltier, W. R. (2005). Arctic freshwater forcing of the Younger Dryas cold reversal. *Nature*, *435*(7042), 662–665. https://doi.org/10.1038/nature03617

Trevena, J., Sijp, W. P., & England, M. H. (2008). Stability of Antarctic bottom water formation to freshwater fluxes and implications for global climate. *Journal of Climate*, *21*(13), 3310–3326. https://doi.org/10.1175/2007JCLI2212.1

Valdes, P. J., Armstrong, E., Badger, M. P. S., Bradshaw, C. D., Bragg, F., Crucifix, M., et al. (2017). The BRIDGE HadCM3 family of climate models: HadCM3@Bristol v1.0. *Geoscientific Model Development*, *10*(10), 3715–3743. https://doi.org/10.5194/gmd-10-3715-2017

Weaver, A. J., Saenko, O. A., Clark, P. U., & Mitrovica, J. X. (2003). Meltwater Pulse 1A from Antarctica as a trigger of the Bølling-Allerød warm interval. *Science*, *299*(5613), 1709–1713. https://doi.org/10.1126/science.1081002

Wickert, A. D. (2016). Reconstruction of North American drainage basins and river discharge since the Last Glacial Maximum. *Earth Surface Dynamics*, *4*(4), 831–869. https://doi.org/10.5194/esurf-4-831-2016

Wickert, A. D., Mitrovica, J. X., Williams, C., & Anderson, R. S. (2013). Gradual demise of a thin southern Laurentide ice sheet recorded by Mississippi drainage. *Nature*, *502*(7473), 668–671. https://doi.org/10.1038/nature12609

Wunsch, C. (2010). Towards understanding the Paleocean. *Quaternary Science Reviews*, *29*(17), 1960–1967. https://doi.org/10.1016/j.quascirev.2010.05.020

Yin, J., Stouffer, R. J., Spelman, M. J., & Griffies, S. M. (2010). Evaluating the uncertainty induced by the virtual salt flux assumption in climate simulations and future projections. *Journal of Climate*, *23*(1), 80–96. https://doi.org/10.1175/2009JCLI3084.1

Zhang, X., Knorr, G., Lohmann, G., & Barker, S. (2017). Abrupt North Atlantic circulation changes in response to gradual CO2 forcing in a glacial climate state. *Nature Geoscience*, *10*(7), 518–523. https://doi.org/10.1038/ngeo2974