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APPENDIX

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Plio-Pleistocene megafloods as a mechanism for Greenlandic megacanyon formation

MODEL SET-UP

When running the global and regional climate models for pre-glaciated Greenland, we make no estimation of the total volume of sediment lost to glacial erosion since the inception of the ice sheet, assuming instead that most of the bed topography pre-dates glaciation as has been argued by other studies (e.g. Cooper et al., 2016), so the effects of glacial erosion are ignored.

This assumption is further supported by measured and inferred rates of glacial erosion which are orders of magnitude smaller than the changes due to isostatic adjustment ($1\text{--}10\text{ m Ma}^{-1}$, Egholm et al. 2017).

Figure S1. Schematic describing the modelling approach used in this study.

Figure S2. Deglaciaded bedrock (left panel) and change relatively to modern (right pattern) used as a boundary condition for running the regional climate model. Bedrock elevation colorscale is the same in Fig. 1 of the main text.

Parameter	Units	Value	Reference Value
Temperature lapse rate	°K per km	5	0
Thickness calving threshold	m	250	50
Stress balance	n/a	sia+ssa	sia
SIA enhancement factor	n/a	1.25	1
Glen exponent in ice flow law for SSA	n/a	3.25	3
Pseudo plastic basal resistance exponent	n/a	0.6	0.25
Till effective pressure fraction of overburden	n/a	0.02	0.02
Bed deformation model	n/a	lc	none
Till parameters: upper value of till friction angle	°	40	15
Till parameters: lower value of till friction angle	°	5	5
Till parameters: elevation at which upper value of till friction angle is used	m	700	1000
Till parameters: elevation at which lower value of till friction angle is used	m	-700	-1000

Table S1. Non-standard parameter values used to run PISM experiments. These parameters were chosen because they provide a good fit to the modern-day ice sheet (see Nielsen et al. 2018).

Additional information about these parameters is available in the PISM documentation.

Asynchronous coupling of climate model to ice sheet model

The climate model output is downscaled to the resolution of the ice-sheet model using a lapse-rate correction (Abe-Ouchi and Blatter 1993). This correction translates the modelled temperature and precipitation fields from the elevation of the preglacial landscape to the elevation of the ice sheet surface. Previous studies have shown that a lapse-rate correction of 5 °K/km is appropriate for modelling northern hemisphere ice sheets on glacial-interglacial timescales (Abe-Ouchi et al. 2007), so we use this value in our simulations. We conducted sensitivity tests and found that using a lapse-rate correction for precipitation (e.g. Ritz et al. 2001) and/or using different values for the lapse-rate correction resulted in different thresholds for ice-sheet stability, but do not change the validity of the mechanism we propose. Therefore for the purposes of this study we present results from the simplest case (temperature lapse-rate 5 °K/km, no lapse rate for precipitation).

Ice sheet model

We note that on the timescales considered here, changes in boundary conditions are poorly constrained, and a simple modelling approach is ideal to study processes and mechanisms that may be modified by, but are ultimately independent of, other changes to the climate system. As has been documented by many previous studies, a strong hysteresis effect controls the growth and decay of ice sheets due in large part to positive climatic feedbacks related to albedo and elevation (modulated by glacial isostatic adjustment) as an ice sheet grows (Weertman, 1961; Abe-Ouchi and Blatter, 1993). In order to initiate retreat, at least one of two main mechanisms must be active: negative surface mass balance or dynamic thinning and retreat. A warming climate can lead to a change in the accumulation area ratio which is

unfavorable for the ice sheet, leading to surface melting and mass wastage. In addition, grounded ice retreating into overdeepened basins, if destabilized, can lead to rapid ice sheet disintegration via the marine ice sheet instability (Weertman, 1974; Schoof, 2007). PISM captures both of these processes, which is an advantage over simpler models. To treat ice-sheet/ocean interactions, we use the simplest calving parameterization in PISM, ocean-kill, which removes any floating ice past the continental shelf break (Fig. 1), limiting the ability of the ice-sheet model to advance onto the continental shelf. The use of this simple calving law may influence some of the dynamics and timing of ice-sheet evolution, and particularly retreat, but are unlikely to change the general result described in the paper. In particular, a more complex calving law would permit the formation of larger ice-shelves, especially in Northwest Greenland, which can buttress and thicken inland ice there. We would expect this process to amplify the feedback processes that result in the formation of proglacial ice-dammed lakes, and as such, our estimates of the rate and magnitude of mass-loss via the mechanism we describe are probably conservative.

For the purposes of this study, we emphasize that while the modelled timing and extent of Greenland inception and retreat phases are sensitive to the climate forcing applied as well as parameter choices for the ice-sheet model and its coupling to the atmosphere, the mechanism for generating outburst floods is independent of these choices. At the ice sheet's peak volume, much of the interior of northern Greenland is grounded 1km below sea level. Rapid warming into an interglacial period, as is characteristic for glacial-interglacial transitions during the Plio-Pleistocene, leads to the destabilization and collapse of the northern margin before the bedrock has time to adjust. This leads to the potential for substantial terrestrial water storage, which would then be vulnerable to drainage via known processes (i.e. ice dam disintegration).

Relatively little is known about Greenland's climate prior to the onset of major glaciations there and what information is available paints a picture strikingly different from the Greenlandic climate we know from modern observations and ice-core records, with flora and fauna indicating reduced sea ice extent and much warmer temperatures than today (Funder et al., 2001; Rybczynski et al., 2013). However, it is clear from marine and terrestrial archives that the ice sheet has also waxed and waned through the Pleistocene, and it is the consequences of that fundamental observation that we wish to highlight here (Flesche Kleiven et al., 2002; Schaefer et al., 2016), with the potential to cause repeated flood events.

Calculation of erosion rates

To estimate the amount of erosion caused by megafloods, we use two approaches. First, we use modern studies of bedrock erosion caused by catastrophic floods to constrain the order of magnitude of landscape change we would expect from our modeled flood volumes. Second, we use a parameterization that predicts spillway erosion rate:

$$\frac{dz_s}{dt} = -(\rho g S)^a * k_e * (z_l - z_s)^a$$

Where the vertical erosion rate is a function of the density of water (ρ), acceleration due to gravity (g), channel slope (S), rock erodibility (k_e), lake surface height (z_l), sill height (z_s), and the erosion law exponent ($a = 1.5$) (Garcia-Castellanos & O'Connor 2018). We consider the range of possible values for uncertain parameters to provide a rough estimate of erosion rates for our modeled flood volumes.

Our back-of-the-envelope calculations suggest that for the rock types found in North Greenland (lithified sandstones) erosion rates in response to these outburst floods would be 0.5–5

m/yr. The maximum depth of the Petermann canyon is ~800 meters, which places 160–1600 years of flood-style erosion by outburst flooding. This is consistent with the jökulhaup model for catastrophic ice-dammed floods, which suggests that Lake Missoula went through many fill-drain cycles during the last deglaciation (Waitt 1985). The flood volumes we reconstruct correspond to ~0.05 m of sea level equivalent (s.l.e.), during periods when the ice sheet lost at least ~8 m s.l.e. in volume (Fig. 2). These outburst floods would have happened late in the deglaciation when North Greenland was mostly ice free, but this mass balance analysis suggests that a single deglaciation would have provided enough melt during this phase to support up to 10s of outburst flood events. Thus, our back of the envelope 160–1600 years of flood-style erosion could be accomplished over more than one and up to tens of interglacials.

Observations of modern catastrophic floods provide another constraint on the total erosion we would expect from these floods. Wilson et al. (2019) looked at a 2015 glacial lake outburst flood (GLOF) in Chile with peak discharge of $5 * 10^{-3}$ Sv and resulted in a maximum of 38 meters of erosion in a sediment-laden alluvial plain. Cook et al. (2018) examined GLOFs in the Himalaya with a peak discharge of $2.3 * 10^{-3}$ Sv which caused 1–10 meters of erosion into sediment. Finally, Lamb & Fonstad (2010) studied an outburst flood in Texas with a peak discharge of $1.4 * 10^{-3}$ Sv which caused 10s of meters of incision in a bedrock canyon. While the discharge estimates for these cases are low compared to what we estimate, all of the studies looked at erosion over smaller spatial scales (1–10 km) than Petermann canyon (~750 km).

Description of advance and retreat of the ice sheet in model simulations

In many individual deglaciations, the Petermann canyon provides a transient outlet to the ocean (Fig. 3). In some scenarios, a pathway is also available via the northeast, near the modern

outlets of Nioghalvfjærdsfjord and Zakariæ. These deglaciation scenarios depict a previously undescribed mechanism with implications for landscape evolution. Following glacial periods when the ice sheet bed has isostatically adjusted to overlying ice load, rapid retreat reveals an overdeepened subaerial basin. Such a basin could have collected runoff and meltwater on land, and an outburst flood could have been triggered by the disintegration of remnant icecaps near the margin or bedrock rebound; subglacial transport of meltwater is likely, but due to the steep surface slopes of the retreated ice sheet relative to the bedrock topography, subglacial hydropotential paths would closely follow surface potential (Fig. 3). The mechanism for these outburst floods would then have been similar to those described for the Laurentide ice sheet during the last deglaciation (Teller et al., 2002; Clarke et al., 2004). To understand the potential of a large subaerial lake in central Greenland to cause outburst floods, we mapped the volume and hydropotential pathways during the two deglaciation scenarios described above (Fig. 3). The volume of the lake below sea level peaks between 2,000 to 4,000 km³ during deglaciations. If the sill elevation is considered, the volume of the lake peaks between 15,000 and 18,000 km³ (Fig. 2). The duration of time these basins exist as subaerial lakes is ~1–3 kyr, a timescale that is dictated by the viscoelastic response of the mantle to the reduced ice load. Though these ice-free depressions form an ice-dammed basin, the earth model does not account for the load caused by the accumulation of water. The lakes we simulate can be hundreds of meters deep in places where the glacial ice sheet reached its maximum thickness (~3 km), meaning this process would slow (not stabilize) the disappearance of these transient lake basins.

The subsequent regrowth phase is slightly different than the retreat phase, as the ice sheet gradually expands from inception points over Camp Century, the northeast coast, and southern Greenland (Fig. 1). Aridity in the northwest, influenced by the Greenland's topography and

simulated storm tracks, restricts glacial inception there, and ice must instead advance from other regions to cover the area.

Other examples of canyon formation via repeated proglacial lake outburst floods

In the Channeled Scablands of western Washington state, repeated outbursts of glacial lake Missoula carved canyons kilometers across and several hundred meters deep. The estimated discharge rate during the formation of these features is 0.1 to $3 \times 10^6 \text{ m}^3 \text{ s}^{-1}$, or 0.1 to 3 Sverdrups (Sv) (Larsen and Lamb, 2016). Although the routing of other meltwater floods associated with the retreating Laurentide ice sheet is less certain (e.g. Condon and Winsor, 2012), the volume of proglacial Lake Agassiz as estimated from shorelines is 9500 km^3 preceding the Younger Dryas, and $163,000 \text{ km}^3$ preceding the 8.2ka event; if released over a period of one year (as it often assumed in modelling studies), these volumes correspond, respectively, to rates of 0.3 and 5.2 Sv (Teller et al., 2002; Clarke et al., 2004).

Implications for abrupt climate change

These outburst floods are a potential driver of abrupt climate change through influence on ocean circulation before glaciation of the northern hemisphere reached magnitudes similar to those observed during the last deglaciation. The potential for relatively low-volume yet potent outburst floods may be easier to integrate with evidence for abrupt climate changes during periods when sea-level variations are minimal.

Many studies have linked iceberg discharge and meltwater floods led to rapid and severe changes in global climate during the LGM (Alley et al., 1993). Though older climate archives generally lack the resolution to study similarly abrupt climate changes, some authors have argued that abrupt climate changes also occurred during past glacial periods in the Plio-Pleistocene (Naafs et al., 2013). These are often attributed to Dansgaard-Oeschger-style

oscillations in the subpolar North Atlantic. Although the exact mechanism for these cycles is unknown, it is likely that variations in freshwater input to the North Atlantic played a role. For instance, Bolton et al. (2010) showed that stable isotope measurements on planktonic foraminifera from the subpolar North Atlantic reveal high-frequency variability in surface conditions there during the early Pleistocene (2.6 — 2.4 Ma). This is in keeping with results from the Mediterranean Sea during Marine Isotope Stage (MIS) 100 (~2.5 Ma), where sea surface conditions responded in synchronization with variations in ice-rafted debris in the North Atlantic (Becker et al., 2006). Thus, it is clear that abrupt climate changes influenced global climate prior to the LGM, yet potential mechanisms for this variability are lacking.

The impact of meltwater outburst floods on global climate is highly dependent on their routing. Condon and Winsor (2012) used a high-resolution ocean circulation model to study the impact of a 5 Sv meltwater pulse sustained over one year, meant to represent a Lake Agassiz outburst flood around the time of the Younger Dryas. They found that routing the water through the St. Lawrence Valley (into the North Atlantic) had little impact on the strength of the AMOC circulation, while water routed through the Mackenzie Valley (into the Arctic Ocean) resulted in a weakened AMOC. The size of the meltwater pulse indicated by our modelling is an order of magnitude smaller than the one used in that study, which is unsurprising given the size of the GrIS relative to the last glacial period Laurentide; however, the northerly outlets where the water is routed are ideally located to be directly entrapped in the East Greenland Current and advected to the North Atlantic with minimal dispersion (Fig 1). Studies of the potential impact of these floods have not been performed despite their obvious potential importance for Atlantic Meridional Overturning Circulation and global climate (Ganopolski and Rahmstorf, 2001).

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