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Catastrophic drainage from the northwestern outlet of glacial Lake Agassiz during the Younger Dryas

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Key Points:

- Peak discharge modelled for the northwestern outlet of glacial Lake Agassiz
- Magnitude of drainage estimated at $1.8\text{-}2.5 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ (1.8-2.5 Sv)
- Northwestern outlet of glacial Lake Agassiz provides a viable link for catastrophic meltwater to the Arctic Ocean during the Younger Dryas

Abstract

Catastrophic meltwater drainage from glacial Lake Agassiz has been hypothesised as a trigger for large-scale ocean circulation change initiating the Younger Dryas cold reversal. Here we quantify the flood discharge that formed the northwestern outlet of Lake Agassiz using a one-dimensional step-backwater model and a zero-dimension gradual-incision model. Applying these two independent models, we estimate a peak discharge range of $1.8\text{--}2.5 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ and a flood volume of $\sim 21,000 \text{ km}^3$. Such a discharge can only be derived from Lake Agassiz rather than one of the two smaller regional glacial lakes: Churchill or Meadow. When coupled with existing ice margin chronologies, these results demonstrate that the northwestern outlet of Lake Agassiz provides a viable link for catastrophic meltwater to drain to the Arctic Ocean over a 5-10 month period during the Younger Dryas, though it is unclear whether this was near its beginning.

Plain Language Summary

The Younger Dryas was a short-lived interval of cooling that interrupted warming during the last deglaciation. The cause of this rapid change in climate is debated. One suggestion is the drainage of meltwater from glacial Lake Agassiz, a large ice-dammed lake in central North America, into the surrounding oceans may have affected ocean circulation, contributing to this climatic event. Here we model the discharge of water from the northwestern outlet of this lake. We estimate the discharge and demonstrate its connection to Lake Agassiz and the Arctic Ocean during the Younger Dryas.

1 Introduction

Retreat of the Laurentide Ice Sheet (LIS) during the last deglaciation resulted in large proglacial lakes forming on its receding southern and western margins. The largest of these was glacial Lake Agassiz, which occupied >1.5 million km^2 during its existence (Teller & Leverington, 2004). Catastrophic freshwater drainage of this lake at the start of the Moorhead Phase has been hypothesised as the trigger for large-scale ocean circulation and global climate change (Broecker et al., 1989; Licciardi et al., 1999; Clark et al., 2001; Teller et al., 2002; Teller & Leverington, 2004), resulting in the Younger Dryas (YD) cooling (12.9-11.7 ka BP). However, the event

magnitude and the drainage pathway remain unclear and disputed (Teller et al., 2005; Lowell et al., 2005, 2009; Fisher et al., 2020; Leydet et al., 2018).

Geomorphic evidence for the hypothesised northwestern drainage of Lake Agassiz is evident in western Canada, notably the Clearwater Lower Athabasca Spillway (CLAS: Figure 1). This >250 km long and 2-3 km wide incised channel and associated large-scale bedforms, record high magnitude discharge. The initial interpretation of the CLAS as the northwestern outlet of Lake Agassiz was based on strandlines and glaciolacustrine deposits at the head of the spillway (Figure 1a). These features were linked to a lake high-stand prior to the drop to the Moorhead low-water phase synchronous with the start of the YD (Fisher & Smith, 1994; Murton et al., 2010; Breckenridge, 2015). Due to uncertainties in ice margin configuration, subsequent studies have proposed the waterbody responsible for forming the CLAS could be the much smaller regional impoundments: glacial lakes Meadow or Churchill (Christiansen 1979; Schreiner, 1983; Fisher et al., 2009; Fisher, 2020; Figure 1b,c). However, a lack of quantitative comparisons of lake volumes and the unknown extent of CLAS incision means the formation of this spillway and magnitude of the freshwater flux to the Arctic Ocean remain unresolved.

Given the spillway's dimensions are likely to have been dynamically adjusting during drainage, we reconstruct the peak discharge via two models: a one-dimensional step-backwater model and a zero-dimensional gradual-incision model. The use of two independent methods allows us to: (1) estimate flood magnitudes consistent with the geomorphic and sedimentologic observations; and (2) estimate the most plausible lake extent and flood volume released from each of three previously proposed glacial lakes Agassiz, Churchill and Meadow. We then discuss the significance of this freshwater discharge in the context of Arctic Ocean circulation.

2 Study Area

The CLAS incises calcareous shale, limestone (Waterways Formation) and sandstone (McMurray Formation) along the majority of its reach, although crystalline basement and sandstones (Mannville Group) occur at its head (Figure S1). The head of the spillway is characterised by a series of eroded uplands >438 metres above sea level (masl) dissected by anastomosing bedrock channels (Figure 1). Downstream of these uplands, the CLAS extends 130 km to the west and is occupied by the underfit Clearwater River. At the modern-day confluence with the Athabasca River, the CLAS bends $\sim 90^\circ$, continuing north ~ 60 km before widening and bifurcating at the Fort Hills. North of the Fort Hills, the spillway terminates at the former Lake McConnell delta (Rhine & Smith, 1988; Smith & Fisher, 1993).

The CLAS is flanked by ten localised scour zones comprising streamlined bedrock (Figure 1d). Catastrophic discharge deposits are preserved on strath terraces or within scour zones, and these include coarse boulder lags, sand and cobble-gravel, with massive to cross-bedded gravel units and local bedrock rip-up clasts (Figure S2a-d). In the lower Athabasca valley, north of Fort Hills, these deposits directly overlie bedrock or deglacial Athabasca River gravels (Smith & Fisher, 1993; Froese et al., 2013), and their sedimentology suggests a single flood event (see Young, 2018; Woywitka, 2019; supplementary methods S1). Lying above or incised into the flood deposits are thin tabular, planar cross-bed sets of coarse gravel and sand representative of sustained lower flow following the catastrophic phase. Where the CLAS bifurcates in a zone stretching ~ 4 km east of the Fort Hills, streamlined and imbricated boulder-gravels, capped with aeolian sand, have been mapped as composite ridge-to-rhomboidal bedforms (Figure S2e-f; Woywitka et al., 2017). Their coarse grain-size and the presence of rip-up clasts indicate a high-magnitude flood.

At the head of the spillway, two principal water-surface planes constrain our flow modelling (Figure 1). The high-water plane cannot exceed the drainage divide at a threshold elevation of 490 masl. The lowermost water plane at 438 masl is associated with sustained post-flood low flow and is demarcated by wave-cut scarps (Fisher & Smith, 1994). This latter elevation is concordant with the elevation of the spillway's bedrock sill and the minimum elevation of channels connecting to the head of the CLAS (Figure S3). The two principal water planes define a maximum drop in lake level of 52 m. Following incision of the spillway, the water surface stabilized at 438 masl. It has

been suggested that volume and extent of the lake responsible for this drop in water level is unclear due to uncertainties in the ice margin configuration forming the lakes' eastern margin. Studies have proposed the waterbody could be the product of evacuation of Lake Agassiz extending into the upper Churchill River Valley (Fisher & Smith, 1994; Murton et al., 2010; Breckenridge, 2015) or much smaller regional impoundments: glacial lakes Meadow or Churchill (Christiansen 1979; Schreiner, 1983; Fisher et al., 2009; Fisher, 2020; Figure 1b,c) separated from Lake Agassiz by ice or topography further to the northeast.

4 Methods

4.1 Step-backwater modelling

Outburst flood discharges are simulated with a 1D step-backwater model (1D model) using HEC-RAS software (Hydrologic Engineering Center, 2001). Steady-state simulations were completed at $0.1 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ increments in a 'mixed flow regime' for two ~20 km long reaches of the CLAS, with the highest density of strath terraces (see Figure 2 for reach locations). Flow energy losses are accounted for using coefficients for flow expansion and contraction (0.1 and 0.3, respectively) and roughness (Manning's n , 0.03-0.05). Field observations suggest the CLAS has experienced significant post-flood incision. To allow for this, we reconstruct the palaeo-channel topography based on the elevation of the strath terraces on which flood gravels and bedforms are deposited (see supplementary methods S2 and Figure S4).

Due to a paucity of flood bars or other depositional landforms within the CLAS, direct constraints on the peak flow's maximum depth are limited. Previous researchers have assumed channel-full conditions in the absence of palaeo-flow indicators (O'Connor & Baker, 1992; Denlinger & O'Connell, 2010). However, time-transgressive spillway evolution models show that this assumption may grossly overestimate peak flow (e.g., Kehew & Lord, 1987; Larsen & Lamb, 2016). There is no observational support for a channel-filling peak flow. Rather, the spillway likely incised progressively, and peak flow level occupied just a fraction of the cross-section. We numerically simulate flood discharge constrained by shear stress estimates, dune morphology, and a channel-full maximum end-member model to investigate these scenarios. Due to the indirect calculation of peak flow, we attribute order-of-magnitude accuracy to the simulations and aim to constrain the discharge of the CLAS accordingly (see Carling, 1996; Carling et al., 2003).

4.1.1 Discharge predictions from channel-full model

The discharge predicted by the channel-full model (CF; Figure 2, 4) is determined iteratively by running the 1D model until the water surface reaches the spillway's maximum capacity—defined as the flood height beyond which flow would have dispersed widely.

4.1.2 Discharge predictions from threshold shear-stress models

Discharges predicted by our threshold shear-stress models are based on the mean intermediate diameter (b -axis) of 100 individual clasts from the coarsest fraction of exposed flood gravels (sampled at five locations; Figure 1). The median intermediate diameter, D_{50} and D_{max} (diameter of the five largest sediment particles at each bar), are calculated for each sample (see supplementary methods S2).

Discharges are predicted by simulating shear stresses that best match thresholds for initial sediment motion (calculated following empirical relations of Shields, 1936; Komar, 1987 (IM*; Figure 2, 4); Ferguson, 1994 (IM**; Figure 2, 4)) and plucking (calculated following methods of Lamb et al., 2015 (SP; Figure 2, 4)). By applying entrainment relations for both bedrock and alluvial bed conditions, we represent the prevailing bedrock detachment and entrainment conditions during and immediately prior to peak discharge, respectively.

4.1.3 Discharge predictions from bedform elevation and geometry models

Raised gravel bedforms (120-160 m long, 3 m high), previously mapped as composite ridge-to-rhomboidal features (Woywitka et al., 2017), are visible where the CLAS widens and bifurcates (Figure 1). Rhomboidal bedforms are rare but useful for estimating flow conditions (Chang & Simons, 1970; Allen, 1982; Carling et al., 2009; Durrant et al., 2017). We use the maximum elevation of these bedforms from the LiDAR data to constrain the minimum floodstage (BE; Figure 2, 4). We restrict our numerical simulations to the downstream portion of Reach 2, where the bedforms occur. We extrapolate the maximum crest elevation and the gradient of the bedforms in Reach 2 to provide a minimum control on peak discharge.

Peak discharges are also estimated from bedform morphology. Previous studies have shown analogous bedform lengths scale primarily with water depth (Durrant et al., 2017). Applying these empirical relations, we estimate flow depth from mean bedform length (132 m, $n=55$). Flow depth is then iteratively simulated (BG; Figure 2, 4) until mean velocity in the 1D model best matches values calculated from bedform geometry (see supplementary methods S2).

4.2 0D gradual-incision modelling

We obtained an independent estimate of peak discharge at the head of the CLAS (see Figure 1 for cross-section location and S3 for supplementary methods), using *spillover*, a zero-dimensional gradual-incision model (0D model) for overtopping outburst floods (code available at <https://github.com/danigeos/spillover>; Garcia-Castellanos et al., 2009). Modelled incision is assumed to deepen the outlet sill below the lake's level at a rate proportional to basal shear stress and substrate erodibility. Critical-flow conditions are assumed at the sill to calculate the shear stress. At the head of the spillway ~50 m of sandstone (Mannville Group) overlies crystalline basement (granite). Erodibility was parameterised to represent the different lithologies: 7.8×10^{-3} and $7.0 \times 10^{-4} \text{ m yr}^{-1} \text{ Pa}^{-1.5}$ for cemented sandstone and granite, respectively, and we applied an uncertainty of $\pm 10\%$ (Garcia-Castellanos and O'Connor, 2018). Manville sandstone is readily eroded, having an unconfined compressive strength of 39 (± 14) MPa (Haug, 2007) in contrast to granite, 100–200 MPa (Goudie, 2006). Based on these parameters, the model calculates the competition between the erosion of the sill and the lake level lowering (sensitive to lake area and hypsometry).

We employ the lake extent of Fisher and Smith (1994) to provide the basis for Lake Agassiz's volumetric calculation. This reconstruction yielded both a lake outburst volume and a lake level drop specific to the northwestern drainage route. No previous volume estimates for lakes Meadow and Churchill have been made. To provide volume estimates for these lakes, 1-arc-second (30 m) SRTM imagery was 'flooded' at the previously proposed water stage elevation (490 m asl for both lakes; Fisher and Smith, 1994). Isostasy was included in our reconstruction of the former lake bathymetry by generating rebound surfaces (palaeo_surface_DEMs) at 11.5 and 12.9 ka BP from a glacioisostatic model (Lambeck et al., 2017). Ice margin positions associated with lake development were derived from Dalton et al. (2020) and minor alterations to the ice margin were

made to accommodate Meadow and Churchill's extents, following previous authors (see Figure 1b,c).

5 Comparison of flood simulation results

Comparison of peak discharge estimates derived from the 1D and 0D models (Figure 4) demonstrates that flood flow did not fill the channel from the palaeochannel bed. The estimates of $5.0\text{--}4.5 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ derived from channel-full 1D modelling far exceed all peak discharge estimates derived from 0D modelling given reasonable erodibility values (Figure 3). This indicates the spillway was incised progressively. Furthermore, the shear stress predicted by the channel-full model far exceeds the threshold for suspension for most boulders (shear stress threshold for suspension of D_{50} is 1368 Pa; following Niño et al., 2003).

By comparison, 1D peak discharge estimates from threshold shear-stress, bedform elevation and geometry models are substantially lower, with an 25th to 75th percentile of $1.8\text{--}2.5 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ (1.8–2.5 Sv). When compared with peak discharge estimates from our 0D models (assuming ~50 m of incision through sandstone), it is clear that the same peak discharge range can be achieved only with an outflow from Lake Agassiz of at least 21,000 km³. By comparison, peak discharge estimates (using the same erodibilities) for the areas of lakes Churchill and Meadow are considerably lower ($2.5\text{--}3.9 \times 10^4$ and $6.1 \times 10^3 \text{ m}^3 \text{ s}^{-1}$, respectively, Figure 3, 4). Obtaining discharges equivalent to Lake Agassiz from these smaller lakes requires erodibility values of $>6.1 \times 10^{-2}$ and $2.3 \times 10^{-1} \text{ m yr}^{-1} \text{ Pa}^{-1.5}$, which are unrealistically high for cemented sandstone (García-Castellanos & O'Connor, 2018). Therefore, these smaller lakes are unlikely sources. Further, our results show that without the underlying granitic basement at the head of the CLAS (~50 m below the Agassiz lake shoreline), the volumes associated with catastrophic drainage would have continued to increase, suggesting this granite provides an important limit to the flood flows following 52 m of incision (Figure 3 and S5).

Based on a 1D modelled peak discharge range of $1.8\text{--}2.5 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ and, assuming a simple triangular flood hydrograph (Alho et al., 2010), the 21,000 km³ outflow from Lake Agassiz would have taken 6–9 months. Additionally, it is probable that continuous drainage of a substantial portion of the LIS into Lake Agassiz prolonged the high discharge (see Licciardi et al, 1999). These estimates therefore provide a minimum flood duration.

6 Discussion

6.1 Flood magnitude and dynamics

Our simulations indicate the catastrophic drainage event that formed the CLAS achieved a peak discharge of $1.8\text{--}2.5 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ (1.8–2.5 Sv; Figure 4). Based on the comparison of the 1D and 0D modelling, we propose that the drainage event through the northwestern outlet exceeded 20,000 km³ from Lake Agassiz and was routed down the Mackenzie River to the Arctic Ocean. Our results indicate Agassiz as the source rather than the much smaller outflows from lakes Meadow or Churchill.

Our model simulations show the CLAS was eroded by a peak floodstage that was only a fraction of the final spillway cross-sectional depth. This is consistent with investigations of other spillways advocating less than channel-full scenarios (Kehew & Lord, 1987; Lapotre et al., 2016; Larsen & Lamb et al., 2016).

The morphology and sedimentology of the CLAS are similar to other spillways in the Interior Plains (Kehew & Lord, 1987; Kehew 1982; Jarret & Malde 1987; Kehew 1993; Clayton & Knox 2008; Norris et al., 2019) and to the eastern outlet of Lake Agassiz (Teller & Thorleifson, 1983). But the CLAS differs from the variety of depositional sediments and landforms seen in the Bonneville spillway. We think the reasons are twofold. First, a flood capable of eroding the CLAS would be expected to leave a discrete sedimentological signature in depositional environments along the spillway. Minimal sedimentation suggests there may have been insufficient sediment availability to build such deposits. This scenario, in turn, indicates a relationship between the abundance of coarse-grained, easily erodible sediment and the formation of depositional landforms within spillway systems (Norris et al., 2019). Second, the lack of a continuous large-scale scour zone probably reflects the constraints on the flow due to the pre-existing topography along the flood route (Froese et al., 2013). The pattern of ten alternating scour zones on either side of the CLAS supports this notion (Figure 1).

6.2 Flood chronology

Regional chronology throughout the Churchill and Athabasca valleys has been constrained by radiocarbon and optically stimulated luminescence ages (e.g. Dyke et al., 2003; Fisher et al., 2009; Munyikwa et al., 2011; 2017; Woywitka, 2019; Dalton et al., 2020). Recent compilation of these chronometers, with the addition of cosmogenic exposure dating at the head of the CLAS suggest deglaciation by $\sim 13 \pm 1.1$ ka BP (Norris, 2020). Based on this age range, a YD timing (12.9-11.7 ka BP) for CLAS incision is plausible, and also compatible with the ~ 10.2 - 9.6 ^{14}C ka BP (~ 12.0 - 10.9 ka BP) radiocarbon dates obtained from post catastrophic-stage organic material deposited within distal fluvial-deltaic sediments north of the Fort Hills (Figure 1) (Froese et al., 2013; Young, 2018). More broadly, the timing of CLAS incision is consistent with recent synthesis of chronologies within the Agassiz basin, which suggests the onset of the lake's Moorhead low-water phase during the YD but not necessarily at the start of the YD (Fisher et al., 2008; Young et al., 2020).

An influx of fresh water into the Arctic Ocean during the YD is consistent with simulations by Tarasov & Peltier (2005), who propose that cooling was a consequence of disruption of the Atlantic Meridional Overturning Circulation (AMOC) by freshwater routed through the Arctic Ocean. Catastrophic drainage to the Arctic Ocean during the YD is suggested from freshening of the Arctic Ocean at ~ 13.0 ka BP revealed by $\delta^{18}\text{O}$ -depleted planktonic foraminifera (Keigwin et al., 2018) and the discovery of flood gravels and a large-scale unconformity dated to ~ 13.0 ka BP at the mouth of the Mackenzie River (Murton et al., 2010). It is probable, given the uncertainty in the age estimations, the CLAS is the source of these high magnitude floods.

6.3 Implications for ocean circulation

Much attention has been directed at assessing the freshwater influx needed to disrupt the AMOC and diminish northward heat transport during the YD (Broecker et al., 1989; Barber et al., 1999; Condrón & Winsor, 2012). Although it has been proposed that YD cooling is a consequence of perturbations to the AMOC (e.g. Keigwin et al., 1991; McManus et al., 2004; Tarasov & Peltier, 2005; Stouffer et al., 2006), the magnitude and location of freshwater injection is debated. Numerical modelling has demonstrated the AMOC is more sensitive to a freshwater input via the Mackenzie River than the Mississippi or St Lawrence River (Condrón & Winsor, 2012). Tarasov & Peltier (2005) model a peak input of 0.1-0.2 Sv to the Arctic Ocean during the YD. A freshwater

perturbation of this magnitude is supported by multi-model simulations, which display a consistent AMOC weakening with a 0.1 Sv input to the northern North Atlantic (Stouffer et al., 2006). When simulating the effect of outburst floods on the break-up of Arctic sea ice, Condron et al. (2020) found that a slowdown of the AMOC and reduction in the northward heat transport could be triggered by a meltwater flux of $\sim 13,000 \text{ km}^3$ over three months combined with a sea ice flux of 0.4 Sv through Fram Strait. Significantly, our estimate of a 1.8-2.5 Sv freshwater flux to the Arctic Ocean is five to ten times that required to weaken the AMOC. However, the question remains as to the duration of the freshwater injection needed to sustain disruption of the AMOC for >1000 yrs. Additional complexity arises because the CLAS discharge to the Arctic Ocean must propagate through a series of lakes (i.e. Lake McConnell and Mackenzie) of which relatively little is known. Despite these uncertainties, our model simulations suggest that a $21,000 \text{ km}^3$ discharge from Lake Agassiz would have sustained flow for at least six to nine months. This discharge alone is unlikely to cause AMOC perturbation on a millennial timescale. However, when combined with a portion of LIS meltwater and precipitation runoff through the newly opened CLAS, sustained through the Moorhead low water phase, we argue that the duration of freshwater input to the Arctic Ocean would have been substantially longer (Licciardi et al., 1999). Recent work has attributed discharge through the Moorhead low to the NW Outlet through the CLAS, until the southern outlet was re-occupied marking the onset of the Emerson Phase (Breckenridge, 2015), estimated between 11,570 and 11,290 BP (Young et al., 2020). In sum, this work demonstrates the NW Outlet was active during the YD, and contributed a peak discharge of 1.8-2.5 Sv. At present we cannot say with certainty this occurred at the start of the YD.

7. Conclusion

We reconstruct the magnitude of catastrophic drainage from the northwestern outlet of Lake Agassiz. The lack of preserved channel cross-sectional profiles and high-water stage indicators limit the precise quantification of the discharge that formed the outlet. Our 1D model attributes the outlet morphology to a peak discharge of 1.8-2.5 Sv, which combined with our 0D model suggests a lake discharge of $\sim 21,000 \text{ km}^3$. Such a discharge must have been sourced from Lake Agassiz rather than one of the proposed smaller regional glacial lakes: Churchill or Meadow. Coupled with chronological constraints on CLAS incision, our results indicate the northwestern drainage of Lake Agassiz. Our best chronological estimates place the NW Outlet within the YD, but not necessarily

at the beginning (~12,900 BP), but provide a viable link for catastrophic meltwater to the Arctic Ocean during the Younger Dryas.

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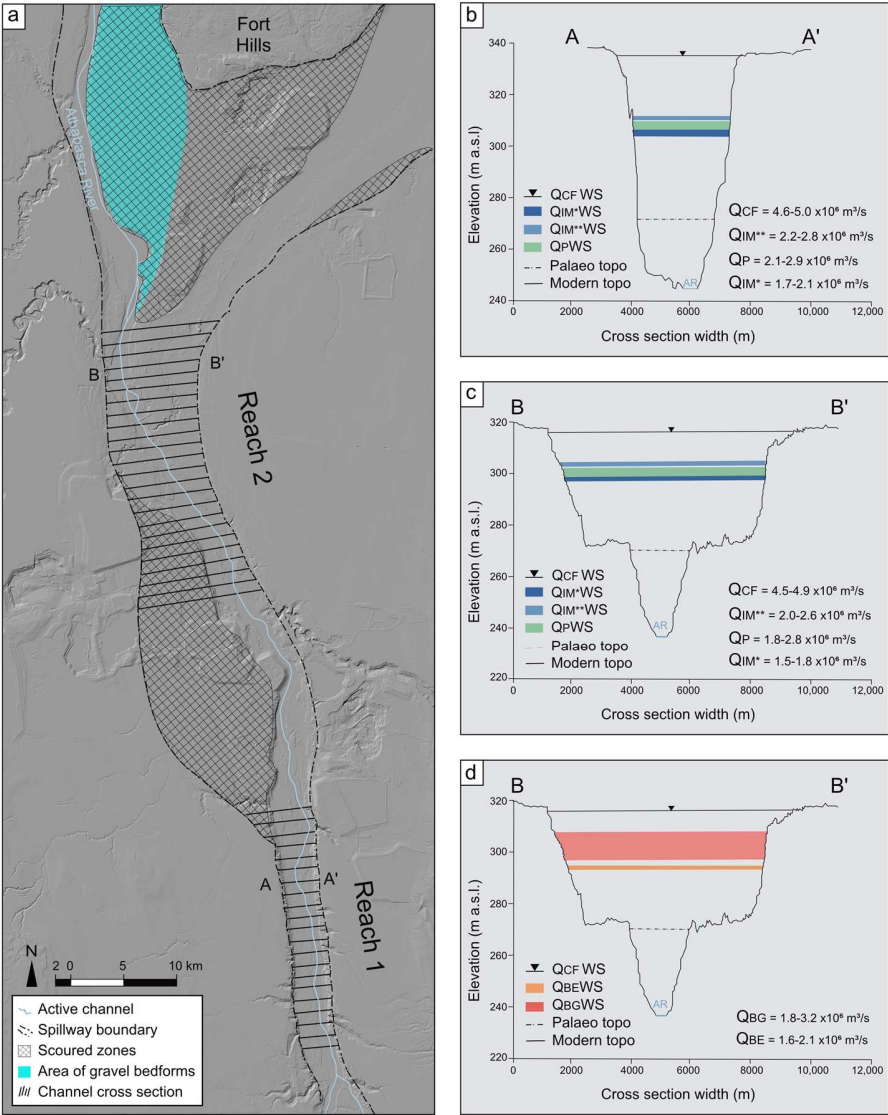
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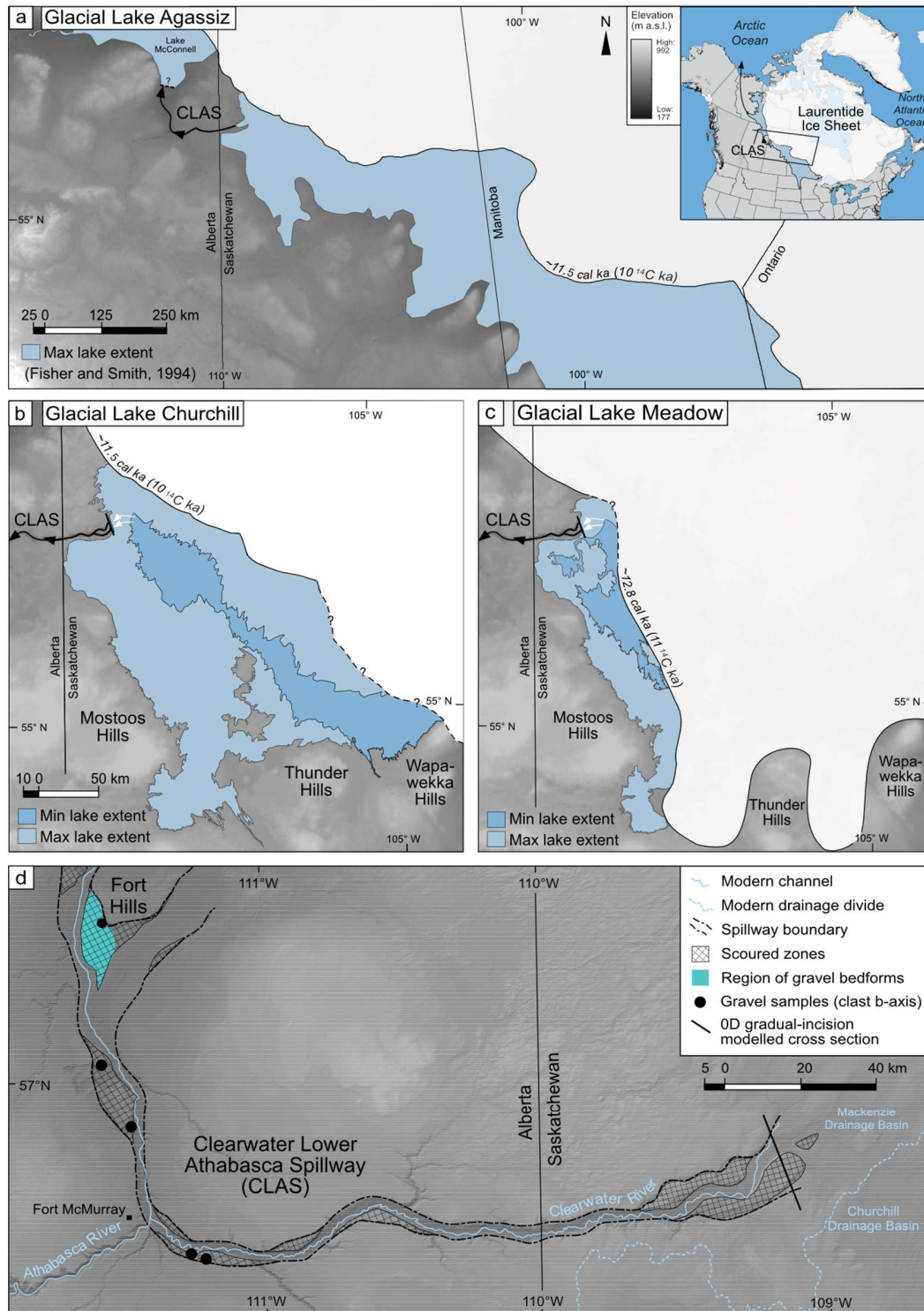
Figure 1. Location of proposed glacial lakes within Churchill River Valley and the CLAS. Ice margins drawn from Dalton et al. (2020). (a) Maximum extent of Lake Agassiz (Fisher and Smith, 1994; Dyke et al., 2004). The proposed northwestern drainage route is shown (black arrow). (b) Maximum and minimum extent of Lake Churchill. Dashed ice sheet margin demarcates alteration by Fisher et al. (2009) to allow the lake's formation. (c) Maximum and minimum extent of Lake Meadow. Dashes indicate the area where the ~12.8 ka BP ice margin has been adjusted to match Christiansen (1979). (d) map of geomorphic features within the CLAS.

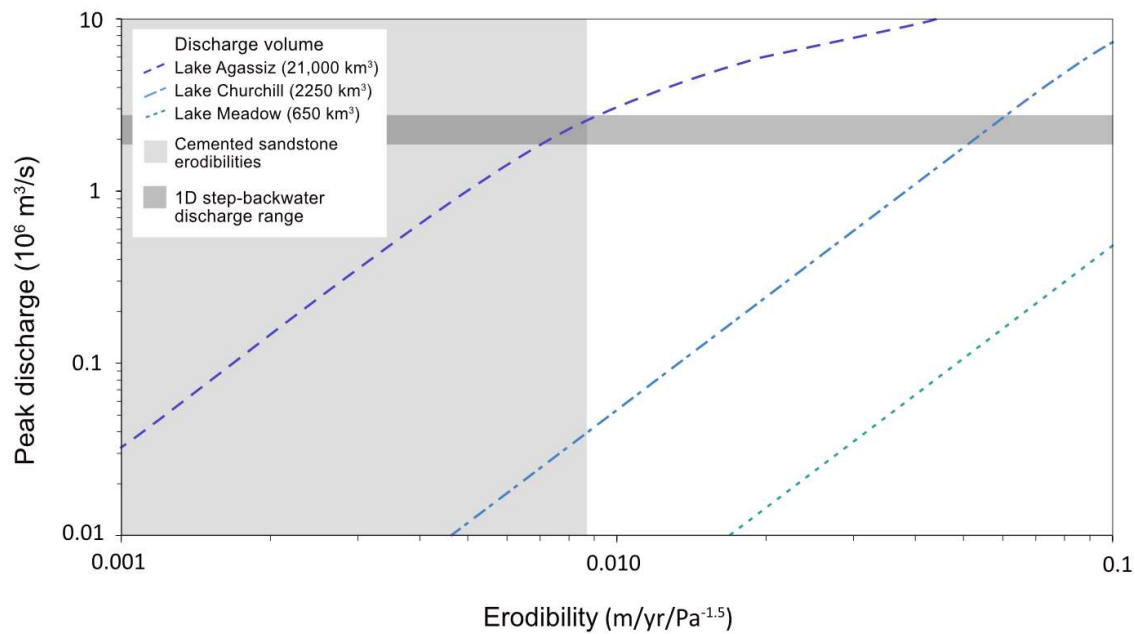
Figure 2. Modelled peak flood discharges and water surfaces (WS) in the CLAS, now occupied by the Athabasca River (AR). (a) 1D modelled reach cross-sections. (b) Peak discharge estimates for Reach 1 and (c, d) Reach 2. Q_{CF} , Q_P , Q_{IM^*} and $Q_{IM^{**}}$ refer to channel-full, sediment plucking, initial motion; Shields (1936) and initial motion; Ferguson (1984) models, Q_{BE} and Q_{BG} refer to bedform elevation and geometry models.

Figure 3. 0D modelled peak discharge from a range of erodibility estimates. The peak discharge range derived from the 1D model (dark grey) can be reproduced with reasonable erodibility values (light gray area) only when the source lake matches the Lake Agassiz dimensions.

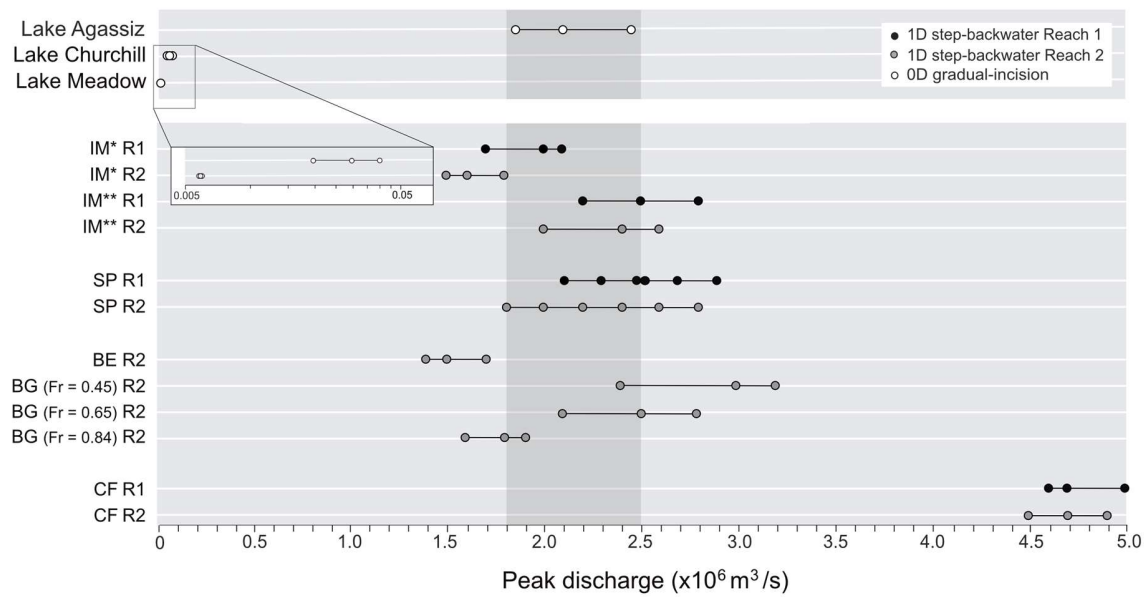
Figure 4. Peak discharge estimates for the CLAS derived from 0D gradual-incision and 1D step-backwater modelling. Peak discharge ranges from 0D models are obtained using a erodibility range of $0.0078 \pm 10\% \text{ m yr}^{-1} \text{ Pa}^{-1.5}$ (consistent with cemented sandstone at the head of the CLAS). Peak discharge ranges for 1D models are based on a Manning's n range of 0.03-0.05. IM^* , IM^{**} , SP , BE , BG and CF refer to initial motion; Shields (1936) and initial motion; Ferguson (1984), sediment plucking, bedform elevation and bedform geometry and channel-full models for reaches 1 and 2 (see Figure 1 for reach locations). Vertical grey shading refers to the most plausible peak discharge range (25th to 75th percentile) $1.8 \text{ to } 2.5 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ derived from 1D modelling.







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