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**10Be ages of flood deposits west of Lake Nipigon, Ontario:
evidence for eastward meltwater drainage during the early
Holocene Epoch**

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1 ¹⁰Be ages of flood deposits west of Lake Nipigon, Ontario: evidence for eastward
2 meltwater drainage during the early Holocene Epoch

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Draft

Abstract

The Nipigon channels, located to the west and northwest of Lake Nipigon, Ontario, are thought to have enabled the eastward drainage of meltwater from glacial Lake Agassiz during the last deglaciation. Here we present the first direct ages of flood deposits in two of these channels using ^{10}Be surface exposure dating. Five ^{10}Be ages of a coarse-grained deposit near the Roaring River in the Kaiashk channel complex indicate deglaciation and cessation of water flow by $\sim 11,070 \pm 430$ yr. To test for inherited nuclides in boulder samples, we also measured the ^{10}Be concentrations of the undersides of two boulders at the Roaring River site. Five ^{10}Be ages of boulders atop a large bedform near Mundell Lake in the Pillar channel complex indicate deglaciation and cessation of water flow by $\sim 10,770 \pm 240$ yr. Two ^{10}Be ages of nearby bedrock are slightly younger ($10,340 \pm 260$ and $9,860 \pm 270$ yr). The ^{10}Be ages from the two sites are statistically indistinguishable and indicate that Laurentide Ice Sheet recession occurred rapidly in the region. We used clast diameters and channel dimensions at the Mundell Lake site to estimate paleo-discharge and evaluate the possibility that meltwater drainage influenced climate conditions. We estimate a large maximum discharge of $119,000\text{--}159,000 \text{ m}^3 \text{ s}^{-1}$ at the site. However, the timing of meltwater discharge at both Roaring River and Mundell Lake is not contemporaneous with abrupt climate events.

Keywords

Surface exposure (^{10}Be) dating, glacial Lake Agassiz, meltwater drainage, spillway, paleo-discharge, early Holocene

1.0 Introduction

It has long been thought that channels in the areas located west of Thunder Bay and west and northwest of Lake Nipigon, Ontario (Fig. 1), served as eastern outlets of glacial Lake Agassiz during the last deglaciation (e.g., Upham 1895; Johnston 1946; Elson 1957, 1967; Zoltai 1965; Teller and Thorleifson 1983, 1987). These channels are incised into the drainage divide between the Lake Agassiz and Lake Superior basins and have spillway sills at progressively lower elevations to the north suggesting successive occupation by meltwater as the Laurentide Ice Sheet (LIS) receded northward (e.g., Teller and Thorleifson 1987). Significant debate has occurred as to the timing of LIS deglaciation in the region and meltwater flow through these channels (e.g., Teller et al. 2005; Lowell et al. 2009), primarily focused on evaluating the hypothesis that eastward meltwater drainage from Lake Agassiz influenced thermohaline circulation in the North Atlantic Ocean and caused abrupt climate change (Broecker et al. 1989).

The channels located west and northwest of Lake Nipigon, Ontario (Fig. 1), are known as the Nipigon channels and host evidence for catastrophic meltwater drainage including deeply incised waterways and water-lain coarse-gravel deposits (e.g., Zoltai 1965; Elson 1957; Teller and Thorleifson 1983, 1987). It is thought that meltwater from Lake Agassiz flowed through the channels into the Lake Nipigon basin and then southward into the Lake Superior basin (e.g., Teller and Thorleifson 1983, 1987; Gary et al. 2011). The ages of the Nipigon channels (interpreted to be <11 cal ka BP) have been constrained using radiocarbon dating of basal lake sediments (Teller et al. 2005) and by correlating strandlines projected from the Lake Agassiz basin to the Nipigon spillways based on the elevations of strandlines and spillways corrected for glacial isostatic adjustment (Johnston 1946; Elson 1967; Teller and Thorleifson 1983, 1987; Teller 2001; Leverington and Teller 2003; Breckenridge 2015). Some of these strandlines subsequently have been dated (Lepper et al. 2011, 2013). Dating spillways in this manner involves significant uncertainties associated with extending waterplanes in the absence of strandlines across long distances, and with poorly known ice-margin positions.

Here we apply ^{10}Be surface exposure dating to provide the first direct ages of flood deposits indicative of meltwater flow within two of the Nipigon channels (i.e., the Kaiashk and the Pillar channel complexes; Fig. 1). We compare the ^{10}Be ages with existing radiocarbon ages of LIS deglaciation in the region, as well as with radiocarbon- and optically stimulated luminescence (OSL)-dated Lake Agassiz strandlines, assuming correlations between these strandlines and the channels are correct. With these data, we provide constraints on the timing of LIS deglaciation west and northwest of Lake Nipigon and the cessation of meltwater flow through the channels. We also provide estimates of paleo-discharge based on clast size and channel morphology data from one channel in the Pillar channel complex (Fig. 1) to assess the possible influence of glacial meltwater on past climate conditions.

2.0 Background

2.1 Prior Research

Five channel complexes located north of the Kaiashk Moraine and west and northwest of Lake Nipigon (Fig. 1) are thought to have drained meltwater to the east (Elson 1957, 1967; Zoltai 1965, 1967; Teller and Thorleifson 1983, 1987). From south to north, these are known as the Kaiashk, Kopka, Pillar, Armstrong and Pikitigushi channel complexes. Each complex contains multiple anastomosing channels leading from a topographically higher region underlain by Archean igneous and metamorphic bedrock to a lower elevation region where Proterozoic diabase overlies the Archean bedrock (Teller and Thorleifson 1983). Thin, sandy till or gravel covers the bedrock in many areas (Zoltai 1965). Some channels have been eroded deeply (<100 m) into the Proterozoic diabase (e.g., Devil's Crater in the Kaiashk channel complex; Fig. 1). Some are shallow channels choked with sand (Zoltai 1965). Closer to the subcontinental drainage divide between Lake Agassiz and Lake Superior, the floors of the channels and nearby areas are covered with large (≥ 1 m diameter) boulders interpreted to have been deposited by flowing water or to be a lag deposit (Elson 1967; Zoltai 1967; Teller and Thorleifson 1983, 1987). Most of the channels are currently dry or host underfit streams or chains of lakes. In general, water flow through the Kaiashk and Kopka channel complexes was eastward and that through the Pillar, Armstrong and Pikitigushi channel complexes was

southward, all draining into the Nipigon basin, then occupied by glacial Lake Kelvin (Zolati 1965) or a high lake level within the Superior and Nipigon basins (i.e., glacial Lake Minong). Lemoine and Teller (1995) and Breckenridge (2007) documented changes in sedimentation in the Lake Nipigon and Lake Superior basins, respectively, including thick varves, and suggest that these register meltwater input from Lake Agassiz.

The timing of deglaciation of the LIS in the area west and northwest of Lake Nipigon and the ages of the channel complexes are poorly constrained. Teller et al. (2005) reported radiocarbon ages of fine vegetative detritus from lakes in the Nipigon channels. The oldest basal age 9320 ± 70 ^{14}C yr BP (10.5 ± 0.2 cal ka BP) is from Lower Vail Lake (Fig. 1) in the Pillar channel complex and directly overlies gravel, suggesting that it is a close minimum-limiting age for cessation of water flow through the channel (Teller et al. 2005). [Lower Vail Lake is in the Pillar channel complex (cf. Thorleifson 1983:41)—not the Kopka channel complex as indicated by Teller et al. (2005)—and is shown as “Vale Lake” in the Ontario and Canada geographic names database. Hereinafter we refer to it as “Vale Lake”]. However, Fisher et al. (2006) questioned the reliability of this age because the type of organic material is unknown and the $\delta^{13}\text{C}$ value of the sample was not reported. Therefore, Fisher et al. (2006) suggested that the most reliable reported minimum-limiting age from Teller et al. (2005) is that of wood from the Devil’s Crater rim core (8155 ± 46 ^{14}C yr BP; 9.1 ± 0.1 cal ka BP).

Efforts have been made to date the channel complexes via correlations with strandlines in the Lake Agassiz basin (e.g., Elson 1957; Teller and Thorleifson 1983; Leverington and Teller 2003; Teller et al. 2005; Breckenridge 2015). Leverington and Teller (2003) proposed that progressive retreat of the LIS northward would have enabled easterly flow from Lake Agassiz across the drainage divide and through successively lower elevation spillways. They suggested that the Kaiashk channel complex was occupied at the end of the upper Campbell stage of Lake Agassiz and that the northernmost Kaiashk and Kopka channel complexes were occupied during the lower Campbell stage. The Pillar channel complex is hypothesized to have been occupied during drainages from the Ojata and Gladstone stages (Leverington and Teller 2003). However, as mentioned above, these correlations have large uncertainties due to the long distances

between strandlines (which end west of the drainage divide between the Lake Agassiz and Lake Superior basins) and poorly known ice-margin positions. Recent work by Breckenridge (2015) mapping strandlines from LiDAR and SRTM digital elevation models projected strandlines associated with the upper (or main) Campbell stage of Lake Agassiz eastwards to the Kashishibog outlet at the head of the Kaiashk channel system (Fig. 1). However, Breckenridge (2015) noted that the projections identify potential outlets, but that there is no direct evidence for Lake Agassiz meltwater flow through the channels.

Beaches associated with the upper and lower Campbell stages are some of the best-dated strandline features in the Lake Agassiz basin. Three maximum-limiting radiocarbon ages for the upper Campbell beach are 9350 ± 100 ^{14}C yr BP (10.6 ± 0.2 cal ka BP; Björck and Keister 1983), 9460 ± 90 ^{14}C yr BP (calibrated 10.7 ± 0.2 cal ka BP; Risberg et al. 1995) and 9460 ± 90 ^{14}C yr BP (10.7 ± 0.2 cal ka BP; Teller et al. 2000) and an average of these ages is 10.6 ± 0.2 cal ka BP (Lepper et al. 2011). Two OSL ages from an undifferentiated Campbell beach ridge west of Fargo, North Dakota are 10.0 ± 0.2 and 10.3 ± 0.2 ka (Lepper et al. 2011). Lepper et al. (2013) provided a mean age (10.5 ± 0.3 ka) from five OSL ages on undifferentiated upper and lower Campbell beach ridges near the southern outlet of Lake Agassiz, with a range of 10.0 ± 0.2 to 10.7 ± 0.2 ka. If the Campbell beaches can be correlated with the Kaiashk and Kopka channel complexes, then these ages would date meltwater drainage through the channels. Strandlines associated with the Ojata and Gladstone stages of Lake Agassiz postdate the Campbell beach ages but are undated. In this study, we add to the chronological constraints on the Nipigon channels by applying ^{10}Be surface exposure dating of flood deposits in the Kaiashk and Pillar channel complexes. The ^{10}Be ages provide minimum ages for exposure of the deposits subsequent to deglaciation of the LIS and cessation of water flow through the channels.

2.2 Geologic Setting of the Sampling Sites

2.2.1 Roaring River Site

The Roaring River occupies one of the southernmost channels in the Kaiashk system (Figs. 1, 2). This channel is cut into Archean granitic bedrock and the surficial

geology of the area is a veneer of sandy glaciolacustrine sediment overlying ground moraine (Mollard 1979). Boulder fields are well exposed in channels that anastomose around slightly higher topography and have been previously described by Teller and Thorleifson (1983). Where recently uncovered due to fire, we observed boulders resting on top of each other suggesting transport by water rather than exposure of a lag deposit following elutriation of fines. We collected samples for ^{10}Be dating at one location exposed by clear-cutting north of the modern Roaring River ($\sim 49.654^\circ\text{N}$, 89.558°W ; Fig. 2). The sample site is a large boulder field with a thin cover of large rounded boulders overlying bedrock. In places, bare bedrock is exposed.

2.2.2 Mundell Lake Site

The Mundell Lake site is located in the Pillar channel complex (Figs. 1, 3). This channel is cut into bedrock at the northern end of Mundell Lake and through a sand and gravel outwash plain overlying bedrock at the southern end of the lake (McQuay 1981). We collected samples for ^{10}Be dating and paleo-hydraulic measurements from a boulder field exposed by clear-cutting on the southeast side of the channel adjacent to Mundell Lake ($\sim 50.221^\circ\text{N}$, 89.164°W ; Fig. 3). In general the area is covered by large (≥ 1 m in diameter) rounded boulders and no bedrock is exposed. We focused ^{10}Be sample collection near the northern edge of the clear-cut where boulders comprise a large bedform. The bedform is an ~ 6 m high asymmetric ridge with a steeper lee side. It appears to be composed entirely of boulders. The long axis of the isolated bedform trends southwest-northeast. Boulders sampled were from the stoss side of the bedform, upstream of the ridge crest, maximizing the likelihood that they were transported by water flow. We also collected samples for ^{10}Be dating from gently sloping bedrock walls on the northeast side of the channel adjacent to Mundell Lake, approximately 1 km upstream from the boulder field.

3.0 Methods

3.1 ^{10}Be dating

We used ^{10}Be dating to determine the most recent time of exposure of boulders and bedrock surfaces (Figs. 1, 2, Table 1). At the Roaring River site in the Kaiashk

channel complex, we collected six samples of boulder surfaces. We also collected samples from the undersides of two boulders (hereinafter referred to as “boulder bottoms”) to test for the presence of ^{10}Be inherited during a prior period of exposure. At the Mundell Lake site in the Pillar channel complex, we collected five samples of boulder surfaces and two samples of bedrock surfaces. All samples were composed of quartz-rich Archean granite. Boulder samples were located in stable geomorphic positions within boulder fields. Samples were removed from flat lying to gently sloping surfaces using a hammer and chisel or the drill-and-blast method of Kelly (2003). In the field, we measured sample locations, sample surface orientations and shielding by the surrounding topography. In the laboratory, we measured sample thicknesses.

We processed ^{10}Be samples in the cosmogenic nuclide laboratories at Lamont-Doherty Earth Observatory and Dartmouth College according to the methods described by Schaefer et al. (2009). All samples were measured at the Center for Acceleratory Mass Spectrometry at Lawrence Livermore National Laboratory (CAMS LLNL). We calculated ^{10}Be ages using the CRONUS-Earth online calculator (Balco et al. 2008) with the Northeastern North American production rate (Balco et al. 2009) and time-variant scaling after Lal (1991) and Stone (2000). Calculating the ^{10}Be ages using other available scaling methods (see Balco et al. 2008) yields ^{10}Be ages less than 3% different from those reported here. ^{10}Be ages are reported in years before the date of collection in 2006 (i.e., “yr”). In contrast, radiocarbon ages discussed (all from prior work) have been recalculated using CALIB 7.0 based on IntCal13 (Reimer et al. 2013) and are reported in thousands of calibrated radiocarbon years before present (i.e., “cal ka BP” with “present” being 1950).

Throughout the time of exposure, there was likely significant cover of the samples by boreal forest vegetation and snow. Cerling and Craig (1994) estimate that vegetation cover may result in ^{10}Be ages ~2–7% younger than the true exposure age. The additional effect of snow cover may have influenced a further reduction of ^{10}Be production in boulder and bedrock surfaces, causing the ^{10}Be ages to be younger than the true exposure age of the flood deposits. We did not correct the ^{10}Be ages for vegetation or snow cover due to the uncertainties associated with assuming the density and duration of cover throughout the time of exposure. Based on field observations of raised quartz veins and

other more resistant surface features, we estimated about 2–3 cm of surface lowering during the exposure period. Therefore, we assumed an erosion rate of $0.0002 \text{ cm yr}^{-1}$ for all samples. Since vegetation and snow cover on the samples, as well as erosion of the rock surfaces, would cause the ^{10}Be ages to be younger than the true exposure age, we suggest that the ^{10}Be ages should be considered minimum ages of deglaciation of the LIS and cessation of meltwater flow through the channels.

3.2 ^{10}Be measurements of boulder bottoms

To test whether ^{10}Be inherited from a prior period of exposure may be present in some of the boulder surface samples, we determined ^{10}Be concentrations of samples from the surfaces and the bottoms of two boulders at the Roaring River site (Tables 1, 2). Samples AF-38 and AF-43 are from the surfaces of two boulders and samples AF-38a and AF-43a are from the bottoms of these boulders. As it was difficult to turn over boulders to obtain samples from the bottom, these samples are from relatively small boulders (<1 m in height). For each pair, we determined the ^{10}Be concentrations of the surface (N_S) and bottom (N_B) samples (both in atoms $\cdot \text{gram}^{-1}$ [at g^{-1}]; Table 2). We then calculated the ^{10}Be production at depth as a fraction of the surface value (P_B/P_S) using the following equation (e.g., Ivy-Ochs 1996):

$$P_B/P_S = e^{-\rho x/\Lambda},$$

where P_B = ^{10}Be production rate at depth x in a boulder (at $\text{g}^{-1} \text{ yr}^{-1}$)

P_S = ^{10}Be production rate at the surface (at $\text{g}^{-1} \text{ yr}^{-1}$)

ρ = density of the rock (g cm^{-3})

x = depth of the sample in a boulder (cm)

Λ = cosmic-ray attenuation length (g cm^{-2}).

We assumed that, if there is no ^{10}Be inherited from a prior period of exposure, then N_B/N_S should equal P_B/P_S . If N_B/N_S is larger than P_B/P_S , then this suggests the presence of ^{10}Be inherited during a prior period of exposure. Since for both samples N_B/N_S is greater than P_B/P_S , we determined an “Expected N_B ” (i.e., one based on the theoretical

calculation of production at depth) using the equation: $\text{Expected } N_B = N_S * (P_B/P_S)$. We then estimated the amount of “Excess ^{10}Be ” by subtracting this “Expected N_B ” from the measured N_B . The “Excess ^{10}Be ” is the amount of ^{10}Be (in at g^{-1}) that may have been produced during a prior period of exposure.

3.3 Paleo-hydraulic estimates

Without field data on high-water marks to estimate energy slopes or detailed cross-sectional areas, we chose to estimate a first-order magnitude reconstruction of paleo-discharge using the continuity equation:

$$Q = Av,$$

where Q = discharge ($\text{m}^3 \text{s}^{-1}$)

A = cross sectional area (m^2)

v = velocity (m s^{-1}).

For calculating velocity (v) we relied on numerous empirically derived formulas including: $v = 0.46 d^{0.5}$ and $v = 0.065 d^{0.5}$ (Williams 1983), $v = 0.49 d^{0.487}$ (Costa 1983), and $v = 0.49 d^{0.381}$ (Koster 1978), where d is the b-axis diameter of a clast. Two of these equations may not be appropriate as Koster’s equation is for clasts <0.3 m diameter in bedforms and Williams’ second listed equation is for initial movement of clasts <1.5 m diameter. Thus, both are expected to underestimate velocity. The first listed equation from Williams (1983) is usually considered an upper limit of flow as it is for clasts between 0.01 and 1.5 m diameter. Most clasts we measured were near or exceeded this upper limit. The Costa (1983) equation may provide the most reasonable solution for velocity as it is for clasts <3.2 m diameter. A similar analysis by Fisher (2004) found that the Costa (1983) equation provided intermediate solutions for velocity.

We calculated paleo-discharges at the Mundell Lake site assuming an average velocity and a maximum velocity. To determine an average velocity, we averaged the velocity calculated by all four methods for each measured boulder. To determine a maximum velocity, we used the average velocity calculated with the Costa (1983) and Williams (1983, second listed) formulae, and the largest boulder observed at the site.

Assuming that the largest boulder was transported, then the discharge calculated using maximum velocity would be the better estimate.

4.0 Results

4.1 ^{10}Be dating

Six ^{10}Be ages of boulders at the Roaring River site range from $10,390 \pm 220$ to $13,170 \pm 650$ yr (Table 1, Figs. 2, 4). The mean age and standard deviation of the samples is $11,420 \pm 940$ yr. Sample AF-39 yielded an anomalously old age (Fig. 4) with a large uncertainty ($13,170 \pm 1,650$ yr). This sample is 1.9σ from the population mean ($n=6$) and is not an outlier according to Chauvenet's criterion (Bevington and Robinson 1992). We suggest that this sample was compromised during processing and measurement and omit it from further discussion. The mean age and standard deviation of the remaining five boulder samples at the Roaring River site is $11,070 \pm 430$ yr. The samples show a relatively normal distribution (Fig. 4) and the dataset has a reduced chi-squared value of 3.02, suggesting that some of the scatter is due to geological uncertainties (e.g., post-depositional movement of boulders, boulder surface erosion, cover by snow, sediment or vegetation, and the presence of nuclides inherited from a prior period of exposure; Balco 2011).

Five ^{10}Be ages of boulders on the large bedform at the Mundell Lake site range from $10,490 \pm 270$ to $11,030 \pm 260$ yr (Table 1, Figs. 3, 4). The mean age and standard deviation of these samples is $10,770 \pm 240$ yr demonstrating a relatively tight sample age distribution (Fig. 4). The reduced chi-squared value of the dataset is 0.83, indicating that the scatter in the dataset is a result of measurement uncertainties (e.g., Balco 2011). Two ^{10}Be ages of bedrock (AF-34 and 35) on the eastern side of the channel now occupied by Mundell Lake are $10,340 \pm 260$ and $9,860 \pm 270$ yr, respectively. The mean age and standard deviation of the bedrock samples is $10,100 \pm 340$ yr and is within the uncertainty of the mean age of the bedform.

4.2 ^{10}Be measurements of boulder bottoms

Two samples of the bottoms of boulders (AF-38a and 43a) at the Roaring River site yield ^{10}Be concentrations of $3.3782 \pm 0.1264 \times 10^4$ and $3.7092 \pm 0.090 \times 10^4$ at g^{-1} ,

respectively (Table 2). Based on the ^{10}Be concentrations of the boulder surfaces and the boulder thicknesses, one would expect ^{10}Be concentrations of $\sim 2.416 \times 10^4$ in sample AF-38a and $\sim 2.883 \times 10^4$ at g^{-1} in sample AF-43a. Therefore, the samples of the bottoms of boulders have an “Excess ^{10}Be ” of $\sim 9,625$ and $8,862$ at g^{-1} , respectively. These “Excess ^{10}Be ” concentrations are equivalent to approximately $\sim 1,700$ (AF-38a) and $1,400$ (AF-43a) years of exposure.

4.3 Paleo-hydraulic estimates

We measured the b-axis diameters of boulders and estimated the channel dimensions at the Mundell Lake site (Table 3). At the Mundell Lake site, the b-axis diameters of boulders adjacent to the bedform ($n=7$) range from 1.2 to 1.8 m, with an average diameter 1.44 m. The b-axis diameters of boulders on the bedform ($n=5$) range from 1.25 to 1.6 m, with an average diameter of 1.38 m. We estimated a former channel width of ~ 600 m based on topographic maps and oblique aerial photographs of the site. We assumed two different values for channel depth (15 and 20 m) from a topographic map because high water marks were not evident and the water depth of Mundell Lake is unknown. Using these depth values, paleo-discharge calculations using average and maximum velocities range from $66,000$ to $88,000 \text{ m}^3\text{s}^{-1}$ and from $119,000$ to $159,000 \text{ m}^3\text{s}^{-1}$, respectively.

5.0 Discussion

5.1 Interpretation of flood deposit ^{10}Be ages

Five ^{10}Be ages of boulders from the Roaring River site yield a mean age of $11,070 \pm 430$ yr. Five ^{10}Be ages of boulders from the Mundell Lake site yield a mean age of $10,770 \pm 240$ yr. There is more scatter in the ^{10}Be ages from the Roaring River site than the Mundell Lake site. Boulders sampled at the Roaring River site were smaller (on average ~ 0.6 m tall) than those at the Mundell Lake site (on average ~ 1 m tall). In addition, many boulders at the Roaring River site were resting directly on bedrock. In contrast, boulders at the Mundell Lake site were resting on other boulders in an imbricate pattern. Based on these differences, we suggest that the interlocking texture of the boulders at the Mundell Lake site led to these samples being more stable, possibly

344 resulting in less scatter in the dataset. Other possible processes that may have influenced
345 scatter in the ^{10}Be ages at the Roaring River site include differential vegetation and/or
346 snow cover, and surface erosion (i.e., spalling) due to forest fires. However, we did not
347 observe any evidence for significant boulder surface erosion at either site.

348 ^{10}Be samples from two boulder bottoms at the Roaring River site suggest that the
349 boulders may contain ^{10}Be inherited from a prior period of exposure. The presence of
350 inherited ^{10}Be in a boulder would influence the ^{10}Be age to be older than the true
351 exposure age and could result in scatter in the dataset. As discussed above, we calculate
352 “Excess ^{10}Be ” concentrations equivalent to approximately ~1,700 and 1,400 years of
353 exposure in samples AF-38a and AF-43a, respectively. However, in order to use the
354 “Excess ^{10}Be ” to quantify how much of N_s may have been produced during a prior period
355 of exposure, one would need to assume that: 1) the amount of inheritance registered at the
356 base of the boulder is the same as what is at the surface; and 2) significant removal of
357 ^{10}Be at the surface has not occurred (e.g., by non-steady state erosion). Therefore, we do
358 not necessarily expect the surfaces of these boulders to contain the same amount of
359 “Excess ^{10}Be ” as the boulder bottoms. Moreover, the calculations described above do not
360 account for ^{10}Be production through the exposed sides of the boulders, which may also
361 contribute to apparent “Excess ^{10}Be ” concentrations in the boulder bottoms. Since the
362 boulders sampled were relatively small (<60 cm in height) and rounded, we suggest that
363 ^{10}Be production through the exposed sides of the boulders was significant. Nonetheless,
364 it is possible that some of the geological scatter in the ^{10}Be ages from the Roaring River
365 site may result from samples that contain ^{10}Be inherited from a prior period of exposure.

366 Based on the tight age distribution and well-preserved landform at the Mundell
367 Lake site, we suggest that the mean age of the bedform ($10,770 \pm 240$ yr) yields a well-
368 established minimum age for deglaciation of the LIS at the north end of the Pillar channel
369 complex and cessation of water flow through the channel now occupied by Mundell
370 Lake. The mean age of the bedform agrees with the radiocarbon minimum-limiting age
371 (10.5 ± 0.2 cal ka BP) from nearby Vale Lake (Fig. 1). The mean age of two bedrock
372 samples at the Mundell Lake site is $10,100 \pm 340$ yr, within the uncertainty of the mean
373 age of the bedform. These samples also provide minimum ages for deglaciation and
374 meltwater flow cessation. However, we suggest that the bedrock samples may have been

influenced by significant vegetation and snow cover due to being flush to the ground and not having the surface roughness of the boulder-landform.

Although the Mundell Lake site is located ~70 km to the north-northeast of the Roaring River site, the mean ages of the boulder deposits at the two sites are statistically indistinguishable, in part due to the scatter in the ^{10}Be ages at the Roaring River site. Therefore, we suggest that deglaciation and cessation of water flow in the Kaiashk and Pillar channel complexes occurred rapidly (likely by $\sim 11.1 \pm 0.4$ to 10.8 ± 0.2 kyr). Rapid ice retreat in this area is consistent with the findings of Lowell et al. (2009) who estimated an ice retreat rate of 161 m yr^{-1} in the region west of Lake Superior.

5.2 Paleo-hydraulic interpretations of the Mundell Lake site

A single large bedform was observed at the Mundell Lake site. This bedform may have been associated with a hydraulic jump, marking the transition from supercritical to subcritical flow where the former channel cross-sectional area widened (Fig. 3). A hydraulic jump would explain the sorting (only boulders) on and possibly within the bedform, as all finer material remained in transport. The paleo-discharges calculated for the Mundell Lake site using average velocity are $66,000\text{--}88,000 \text{ m}^3\text{s}^{-1}$ and using maximum velocity are $119,000\text{--}159,000 \text{ m}^3\text{s}^{-1}$. These values are similar to those determined by Teller and Thorleifson (1983) for spillways associated with the eastern outlet ($100,000\text{--}200,000 \text{ m}^3\text{s}^{-1}$) and the southern outlet ($100,000\text{--}360,000 \text{ m}^3\text{s}^{-1}$) of Lake Agassiz (cf. Fisher 2004).

There have been many paleo-hydraulic reconstructions of floods from large bedforms and boulder deposits associated with deglaciation elsewhere. A full review and discussion of this topic is beyond the scope of this paper but the interested reader is directed to Herget (2005) and Burr et al. (2009). An example of other large bedforms is antidunes in southern British Columbia with wavelengths of 100 to 230 m and heights of 3 to 7 m associated with high magnitude flows from the drainage of glacial Lake Deadman (Carling et al. 2009a). Calculated flow velocities for this site are $\sim 13\text{--}19 \text{ m s}^{-1}$, similar to the maximum values estimated for the Mundell Lake site. Carling et al. (2009a) reviewed the literature on large dune forms and found that, in most cases, large-scale dunes associated with drainage of glacial Lake Missoula (e.g., Baker and Bunker

1985) and the Altai floods (e.g., Herget 2005) have crossbeds composed of much finer gravel than observed at the Mundell Lake site. However, in some cases boulders up to 3 m diameter are present, such as at the Kuray Basin dunes in Siberia, and are assumed to have been transported by water flow (Herget 2005; Carling et al. 2009b). For the Kuray Basin dunes, a flow depth of 20–30 m is associated with a velocity of $\sim 10 \text{ m s}^{-1}$, again, similar to the Mundell Lake site. The bedform at the Mundell Lake site indicates the flow of a large volume of water. Our work, however, does not constrain the duration of flow.

5.3 Comparison of flood features with Lake Agassiz strandlines

As discussed in section 2.1, prior work has inferred ages for the Nipigon channels using correlations between the elevations of strandlines in the Lake Agassiz basin and spillways. Here we can test these correlations using chronological constraints from the Nipigon channels. The mean age of the Roaring River site of $11.1 \pm 0.4 \text{ kyr}$ is interpreted as a minimum-limiting age for deglaciation and water cessation in the channel. Within uncertainties, it only just overlaps with the maximum-limiting radiocarbon ages of upper Campbell beaches (mean of three ages is $10.6 \pm 0.2 \text{ cal ka BP}$; Björck and Keister 1983; Risberg et al. 1995; Teller et al. 1995) and the oldest OSL ages of undifferentiated Campbell beaches (mean of five ages is $10.5 \pm 0.3 \text{ ka}$; Lepper et al. 2013). Thus, the ^{10}Be ages suggest that the channel at the Roaring River site may be older than the upper Campbell beaches. The mean age of the bedform at the Mundell Lake site ($10.8 \pm 0.2 \text{ kyr}$) provides a new constraint on the timing of deglaciation and meltwater cessation through the Pillar channel complex. This age agrees well with the aforementioned radiocarbon and OSL ages of Campbell beaches.

Although it is generally thought that Lake Agassiz sourced the meltwater that formed the Nipigon channels, the relatively old age for the Roaring River site in comparison with the Campbell beach ages, as well as the location of the site 35 km southeast of the proposed Kashishibog outlet for the upper Campbell stage (Breckenridge 2015; Fig. 1), suggest the possibility that meltwater may have originated in a local lake. A local lake to the east of the subcontinental drainage divide near the Roaring River site was mentioned by Mollard and Mollard (1983), and may have formed as ice retreated

northward from the Lac Seul and Kaiashk moraines in conjunction with ice retreating eastwards that ultimately formed the Nipigon moraine (Fig. 1).

5.4 Meltwater drainage and climate conditions

The ^{10}Be ages presented here, as well as the prior radiocarbon ages (i.e., Teller et al. 2005), indicate that LIS deglaciation and meltwater cessation in the Nipigon channels occurred during the early Holocene Epoch. Although we estimate a very large maximum discharge of $119,000\text{--}159,000\text{ m}^3\text{s}^{-1}$ at the Mundell Lake site, we have no information about the duration of meltwater drainage. Moreover, the timing of meltwater passage through the Nipigon channels (here dated at 11.1 ± 0.4 and 10.8 ± 0.2 kyr), attenuated through the Great Lakes basin, and to the Atlantic Ocean does not correspond with any known climate events. Therefore, we have no basis for concluding that meltwater drainage through the Nipigon channels had a significant influence on thermohaline circulation in the North Atlantic Ocean or past climate conditions.

6.0 Conclusion

We present the first direct ages of flood deposits in the Nipigon channels. These ages indicate that deglaciation and cessation of meltwater flow occurred by $\sim 11.1\pm0.4$ in the Kaiashk channel complex and by $\sim 10.8\pm0.2$ kyr in the Pillar channel complex. ^{10}Be concentrations of two boulder bottoms at the Roaring River site are equivalent to approximately $\sim 1,400\text{--}1,700$ years of exposure. The presence of ^{10}Be inherited from prior periods of exposure may have influenced the greater amount of scatter in the dataset from the Roaring River site. However, the agreement of ^{10}Be ages at the Roaring River site with those at the Mundell Lake site, which show a tight distribution, suggests that there is not significant inherited ^{10}Be in the samples. The ages from the Roaring River site appear to be older than prior radiocarbon and OSL ages of the upper Campbell beach, suggesting the possibility that a local source of meltwater may be partially responsible for some features previously explained by meltwater drainage from Lake Agassiz. The ^{10}Be ages from the Mundell Lake site agree well with ages from the Campbell Beach and with a prior radiocarbon age in the area (Teller et al. 2005). We have no basis for concluding that the large meltwater discharge through the Nipigon channels (estimated at a

maximum discharge of 119,000–159,000 m³ s⁻¹ for the Mundell Lake site) influenced past climate conditions.

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Tables

Tables 1-3 see other word file.

Figure captions

Figure 1. Location map showing the Kaiashk, Kopka and Pillar channel complexes, the Roaring River and Mundell Lake sampling sites (white rectangles) and other features mentioned in the text. Arrows represent major flow-paths feeding channels incised into bedrock. The approximate location of the subcontinental drainage divide between the glacial Lake Agassiz and Lake Superior basins is shown as the dashed grey line. TB is Thunder Bay and the other initials are abbreviations for states and provinces.

Figure 2. Maps and photos of the Roaring River sampling site. A) Vertical aerial photograph of the sampling site shown in (B). B) Location of sampled boulders and ^{10}Be ages. C) Low-angle oblique photograph of the study site with a light dusting of snow covering boulders that surround an area of bedrock in the center of the image. D) Ground view of the boulder field and ^{10}Be sample AF-43.

Figure 3. Maps and photos of the Mundell Lake sampling site. A) Vertical aerial photograph of the sampling site shown in (B). B) Location of sampled boulders and bedrock and ^{10}Be ages. The meltwater flow pathway along Mundell Lake is interpreted from (A). The location of the bedform shown in (D) is on the east side of the lake (dashed grey line). The bedform may be a result of a hydraulic jump located where the cross-sectional area of the channel widens and flow velocities may have decreased. C) Ground view of ^{10}Be sample AF-31. D) Low-angle oblique photograph of the sampling

682 site adjacent to Mundell Lake, with a view to the southwest as indicated by the black
683 arrow on (A). White dashed line is the crest of the bedform from which boulders were
684 sampled.

685
686 Figure 4. Probability plots showing ^{10}Be ages from the Roaring River (A) and (B) and
687 Mundell Lake (C) sites. The distribution of ^{10}Be ages from the Roaring River site shown
688 in (A) includes sample AF-39. This sample is omitted in (B). The thin black curves are
689 normal Gaussian distributions of the ^{10}Be ages and uncertainties. The thick black curve
690 is the summed probability of all samples. The statistics shown on the right are for the
691 thick black curve, except for the weighted mean which is for the thin black curves. Green
692 (short-dashed), red (longer-dashed) and black (long-dashed) vertical lines indicate three,
693 two and one standard deviations, respectively, and the blue (solid) vertical line is the
694 mean of the population.

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141x174mm (300 x 300 DPI)

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169x146mm (300 x 300 DPI)

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 161x147mm (300 x 300 DPI)

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203x276mm (300 x 300 DPI)