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^{10}Be ages of late Pleistocene deglaciation and Neoglaciation in the north-central Brooks Range, Arctic Alaska

MICHAEL E. BADDING,[1] JASON P. BRINER[1]* and DARRELL S. KAUFMAN[2]
[1]Department of Geology, State University of New York at Buffalo, Buffalo, NY 14260, USA
[2]School of Earth Sciences & Environmental Sustainability, Northern Arizona University, Flagstaff, AZ 86011, USA

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ABSTRACT: We present a chronology of late Pleistocene deglaciation and Neoglaciation for two valleys in the north-central Brooks Range, Alaska, using cosmogenic ^{10}Be exposure dating. The two valleys show evidence of ice retreat from the northern range front before ~16–15 ka, and into individual cirques by ~14 ka. There is no evidence for a standstill or re-advance during the Lateglacial period, indicating that a glacier advance during the Younger Dryas, if any, was less extensive than during the Neoglaciation. The maximum glacier expansion during the Neoglacial is delimited by moraines in two cirques separated by about 200 km and dated to 4.6 ± 0.5 and 2.7 ± 0.2 cal ka BP. Both moraine ages agree with previously published lichen-inferred ages, and confirm that glaciers in the Brooks Range experienced multiple advances of similar magnitude throughout the late Holocene. The similar extent of glaciers during the middle Holocene and the Little Ice Age may imply that the effect of decreasing summer insolation was surpassed by increasing aridity to limit glacier growth as Neoglaciation progressed. Copyright © 2012 John Wiley & Sons, Ltd.

KEYWORDS: Alaska; Brooks Range; Neoglaciation; ^{10}Be dating.

Introduction

Arctic environments are affected by cryosphere–albedo feedback mechanisms that result in climate amplification (Overpeck *et al.*, 1997; Holland and Bitz, 2003; Serreze and Barry, 2011), and the Arctic is expected to undergo significant change during the 21st century (Miller *et al.*, 2010; Sharp *et al.*, 2011; Jacob *et al.*, 2012). Glaciers are one component of the Arctic environment, and have experienced pronounced retreat due to recent climate warming (Meier *et al.*, 2003; Gardner *et al.*, 2011; Fisher *et al.*, 2012). Reconstructions of past glacier fluctuations can provide insights into Arctic climate change. In particular, alpine glaciers respond sensitively to climate change, and when well-dated records of former glacier change are available from around the globe, they can be used to reconstruct spatial patterns of climate change. However, there are a limited number of well-dated alpine glacier chronologies in the Arctic.

The Brooks Range, Arctic Alaska (Fig. 1), affords an opportunity to investigate the timing of alpine glacier retreat because unlike most land area in the Arctic, the Brooks Range was glaciated by alpine glaciers rather than covered by ice sheets. Determining the timing of deglaciation in the Brooks Range provides a means to investigate Arctic climate patterns during the last termination, and to compare these with glacier records from elsewhere in the northern hemisphere. For example, it is unknown whether glaciers in the Brooks Range responded to the abrupt climate changes that define the Lateglacial period in the North Atlantic region (Alley *et al.*, 1993; Denton *et al.*, 2010; Harrison *et al.*, 2010). Although glacier advance during the Younger Dryas period (YD; 12.9–11.7 ka) may be prevalent in some areas around the North Atlantic region (Ingólfsson *et al.*, 1997; Lohne *et al.*, 2012), evidence elsewhere is less definitive (e.g. Gosse *et al.*, 1995; Mangerud and Landvik, 2007; Kelly *et al.*, 2008).

The Brooks Range contains ~1000 modern glaciers, which have been retreating significantly in recent decades (Molina, 2007). It is unknown when glaciers in the Brooks Range expanded following the Holocene warm period (Kaufman *et al.*, 2004), and when they achieved their maximum extent during the Holocene. Previous studies suggest that numerous mid-Holocene moraines are preserved outboard of moraines deposited during the Little Ice Age (LIA; ~1200–1900 AD; Calkin and Ellis, 1980; Ellis and Calkin, 1984; Solomina and Calkin, 2003; Sikorski *et al.*, 2009). Pre-LIA late Holocene moraines are absent in front of many glaciers throughout the northern hemisphere, probably because most glaciers seem to have been driven by the decrease in northern latitude summer insolation throughout the Holocene and they culminated in their greatest extent during the LIA (e.g. Barclay *et al.*, 2009; Briner *et al.*, 2009). Although the mechanism behind Arctic summer cooling throughout the Holocene is well understood on millennial timescales, dating the multiple glacier fluctuations throughout the late Holocene provides additional information about centennial-scale climate changes in the Arctic.

The goals of this study are to: (i) develop a chronology of final deglaciation during the last termination; and (ii) produce numerical ages on outermost late Holocene moraines. We present 17 cosmogenic ^{10}Be exposure ages (hereafter termed ^{10}Be ages) on deglacial and Holocene glacial landforms in the north-central Brooks Range. We report ^{10}Be ages from the headwaters of two valleys to determine when the upper catchments became deglaciated. In addition, we ^{10}Be-dated two pre-LIA moraine crests with previously published lichenometric ages to determine when Brooks Range glaciers attained their maximum Neoglacial configuration.

Setting and glacial history

The Brooks Range is the northernmost mountain range in Alaska, trending east–west and separating the North Slope to the north from the Yukon River basin to the south (Fig. 1). The range stretches ~1000 km from the Chukchi Sea on the west coast to the Alaska–Yukon border in the east, and lies entirely north of the Arctic Circle. The Brooks Range reaches over 2700 m above sea level (a.s.l.), with ~1000 glaciers (mostly

*Correspondence: J. P. BRINER, as above.
E-mail: jbriner@buffalo.edu

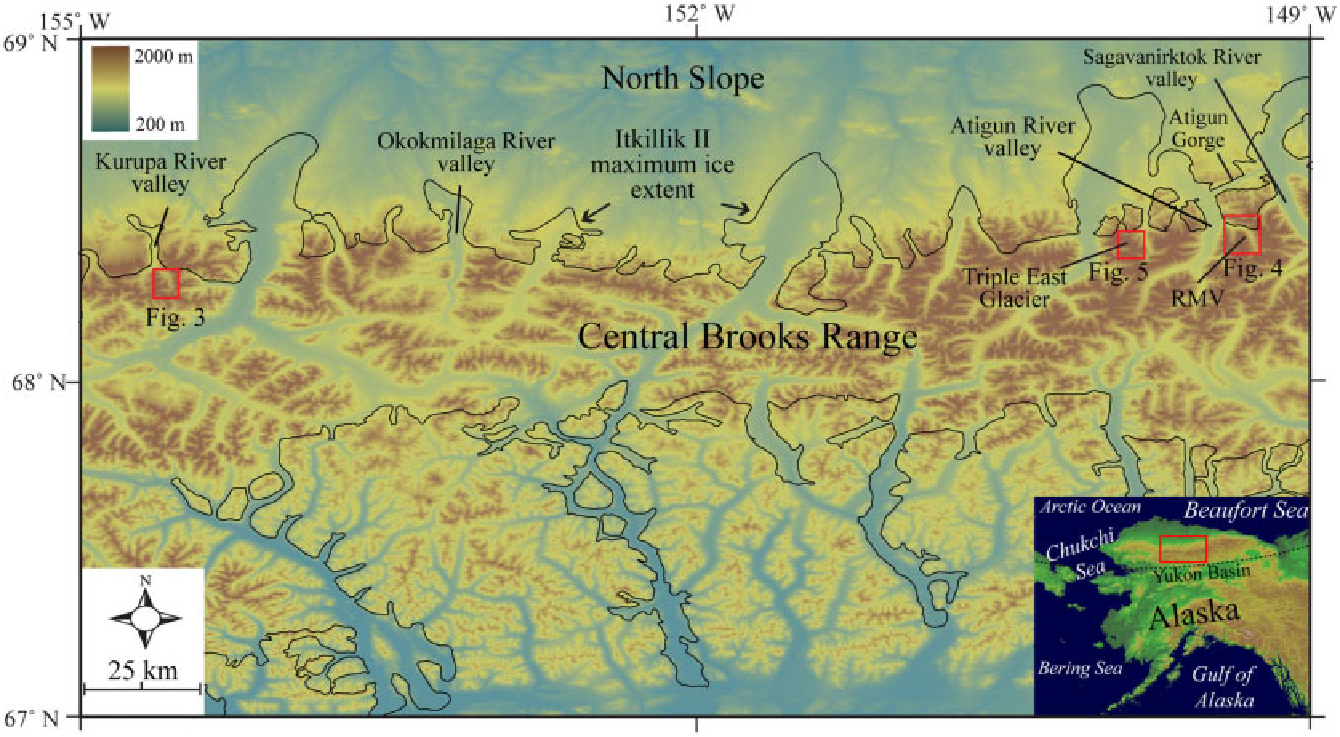

Figure 1. Digital elevation model of north-central Brooks Range. Inset: location of Brooks Range within eastern Beringia. Locations mentioned in text are shown; RMV, Roche Moutonnée valley. Ice limits from Hamitton (1986). This figure is available in colour online at wileyonlinelibrary.com.

cirque glaciers) occupying the highest peaks (Molina, 2007). Nearly all of the modern glaciers are sheltered behind north-facing cirques (Ellis, 1982). Bedrock in the north-central Brooks Range is dominated by highly deformed Devonian sedimentary and meta-sedimentary strata (Brosgé *et al.*, 1979). Most relevant for ^{10}Be dating are the resistant, quartz-rich members of the Kanayut Conglomerate (Upper Devonian). Although Pleistocene drift limits are well mapped in the Brooks Range, and the chronology of several Itkillik II maxima phases (late Wisconsin glaciation) is generally known (Hamilton, 2003; Balascio *et al.*, 2005; Briner and Kaufman, 2008), the timing of deglaciation in the Brooks Range is less well known (Hamilton, 1986).

We collected samples for ^{10}Be dating to constrain the timing of deglaciation of the headwaters in the Kurupa River and Atigun River valleys. We chose these study locations because both valleys are accessible, contain the target lithology for ^{10}Be dating, and were previously mapped and dated with lichenometry (Ellis and Calkin, 1984; Lamb, 1984). The Kurupa River valley is situated in the west-central Brooks Range (Fig. 1). Here, Itkillik II terminal moraines lie ~10 km north of the range front and ~25 km downvalley of cirque headwalls (Hamilton, 1982). Although the glacial chronology has not been previously studied in the Kurupa River valley, radiocarbon ages ranging from 12.6 to 15.4 cal ka BP from the Okokmilaga River valley, ~60 km east (Fig. 1), provide minimum limiting ages on timing of deglaciation from Itkillik II ice limits (Hamilton, 1982).

Approximately 200 km to the east of Kurupa River valley is the Atigun River valley (Fig. 1), which served as a major trough for ice drainage during the last glaciation (Hamilton, 1986). We investigated the deglaciation of Roche Moutonnée valley (informal name, Fig. 1), which is a tributary to the Atigun River valley ~15 km south of the northern range front.

Terminal moraines deposited by glaciers sourced in the Atigun River valley during the early phase of the Itkillik II glaciation are located ~30 km from the cirque headwalls of Roche Moutonnée valley (Fig. 1). During the final phase of the Itkillik II glaciation in the Atigun River valley, ice spilled eastward through Atigun Gorge and into the Sagavanirktok River valley (Fig. 1), forming a piedmont lobe and partially damming the Sagavanirktok River (Hamilton, 2003). Radiocarbon ages dating an aggradation event associated with an ice advance into the Sagavanirktok River valley are 15.2 ± 1.4 and 15.5 ± 1.2 cal ka BP (Fig. 1); an additional radiocarbon age of 13.8 ± 0.7 cal ka BP is interpreted as a minimum limiting age on deglaciation (Fig. 1; Hamilton, 2003).

Our Holocene glacial chronology arises from two cirques located ~200 km apart, both of which contain pre-LIA moraines based on previous mapping (Ellis and Calkin, 1984; Lamb, 1984). In the Kurupa River valley, we collected samples from boulder surfaces from the outermost pre-LIA moraine fronting Fireweed West glacier [all glacier names are informal from Ellis and Calkin (1984) and Lamb (1984)]; previous mapping and lichenometric data suggested that a suite of moraines stabilized at 750 ± 150, 1650 ± 330 and 3700 ± 740 cal a BP (Lamb, 1984). We also dated moraine boulders from the Triple East glacier, which hosts one of the oldest lichenometry-inferred Holocene moraines in the north-central Brooks Range (Ellis and Calkin, 1984; Fig. 1). Ellis and Calkin (1984) assigned lichen ages of 400 ± 80, 1900 ± 380, 3200 ± 640, 3700 ± 740 and 5000 ± 1000 cal a BP to the moraines fronting Triple East glacier.

Methods

Sites were selected for ^{10}Be dating by considering potential snow and debris cover and post-glacial erosion. We sampled bedrock and boulder surfaces marked by glacial polish and striations, indicating negligible post-glacial erosion. Samples from bedrock were collected from at least 0.5 m above surrounding landscape and from elevated sites within the valley bottom, increasing the likelihood of the site to be windswept of snow and free of previous cover by glacial sediment (Fig. 2). We sampled sites along the valley center where the duration of glacier occupation and efficiency of subglacial erosion are most likely to have reduced the possibility of isotopic inheritance. In some cases we sampled

Figure 2. Examples of boulder and bedrock sample sites. (A) Bedrock knob (sample 11RMV-08) in the middle of the valley and ~40 m above the valley bottom. (B) Perched erratic (11RMV-07) on glacially eroded bedrock (11RMV-08). (C) Typical Holocene moraine boulder (10FWW-02). (D) Bedrock sample site (10KRV-09) with late Holocene moraine in the background. This figure is available in colour online at wileyonlinelibrary.com.

erratic boulders that rested directly on exposed bedrock (Fig. 2). All moraine boulders that we sampled had tops ~1 m above the surrounding landscape and had no evidence of post-depositional movement (Fig. 2). The Fireweed West glacier moraines are probably ice cored, and thus some downslope movement is likely to have occurred. However, we avoided portions of the moraine with boulders that exhibited obvious signs of rotation, aided with lichenometry. Thus, despite some post-depositional down-valley motion of the moraines, many boulders on the moraine crests are probably in the same orientation today as they were when they were deposited originally. All samples were from horizontal to near-horizontal surfaces, and we avoided boulder edges and corners. Geographic coordinates and elevation for each sample were determined with a handheld GPS device, and checked on a 1:63 360-scale USGS topographic map. Topographic shielding was measured in the field at each sample site with a clinometer and compass.

Samples were collected with a hammer and chisel from ice-sculpted meta-sandstone and conglomerate bedrock surfaces ($n = 8$), erratics perched on bedrock ($n = 2$) and moraine boulders ($n = 7$). All samples were processed at the University at Buffalo Cosmogenic Isotope Laboratory. Samples were crushed and sieved to isolate the 250- to 425-μm size fraction. Samples were pretreated in dilute HCl and HNO_3-HF acid baths. Quartz was isolated with a series of heated HNO_3-HF acid baths, and in some cases large grains with visible inclusions were removed by hand. ^{9}Be carrier (~0.37–0.80 mg; 405 p.p.m.) was added to each sample prior to dissolution in concentrated HF. Beryllium was extracted using ion-exchange chromatography, selective precipitation with NH_4OH and final oxidation to BeO.

^{10}Be/^{9}Be accelerator mass spectrometry measurements were completed at the Lawrence Livermore National Laboratory Center for Mass Spectrometry and normalized to standard 07KNSTD3110 (Nishiizumi *et al.*, 2007). Ratios for process blanks ($n = 2$) averaged $4.39 \pm 2.06 \times 10^{-15}$. ^{10}Be ages were calculated assuming no snow shielding and no erosion using the CRONUS-Earth online exposure age calculator (http://hess.ess.washington.edu/math; Version 2.2; Balco *et al.*, 2008).

We used the north-eastern North America ^{10}Be production rate of 3.93 ± 0.19 atoms g^{-1} a^{-1} (Balco *et al.*, 2009) and the constant-production scaling scheme of Lal (1991) and Stone (2000). There are no ^{10}Be production-rate calibration sites in Alaska; thus, we choose the north-eastern North America ^{10}Be production rate because recent ^{10}Be production rates from elsewhere in the Arctic yield a similar value (Fenton *et al.*, 2011; Briner *et al.*, 2012). The north-eastern North American ^{10}Be production rate yields ages ~12% older than the globally averaged rate from Balco *et al.* (2008) when using the 'St' scaling scheme. The use of alternative scaling schemes results in ages that are up to ~4.9% older or ~1.5% younger than the ages that we present.

Results

Late Pleistocene deglaciation

We ^{10}Be-dated five samples from four sites in the eastern portion of upper Kurupa River valley (Fig. 1). Four ^{10}Be ages from valley-bottom bedrock samples and one ^{10}Be age from an erratic boulder range from 15.9 ± 0.5 to 13.8 ± 0.3 ka (Table 1, Fig. 3), and span 1.5–6.5 km from the cirque headwall. The ages increase with distance from cirque headwalls. Samples from the erratic boulder and bedrock upon which it lies yield statistically identical ^{10}Be ages of 13.9 ± 0.3 ka (erratic) and 13.8 ± 0.3 ka (bedrock). Five ^{10}Be ages from four sites within Roche Moutonnée valley range in age from 50.4 ± 1.2 to 14.7 ± 0.5 ka (Table 1, Fig. 4) spanning 3–10 km from the cirque headwall. Although the ages in Roche Moutonnée valley are considerably older than in Upper Kurupa valley, they too decrease in age toward the headwalls. Samples from an erratic boulder and bedrock upon which it is perched yield ^{10}Be ages of 14.7 ± 0.5 and 17.1 ± 0.3 ka, respectively.

Mid Holocene moraines

In the upper Kurupa River valley, we adopted the Holocene moraine mapping from Lamb (1984), who showed an end-moraine complex fronting Fireweed West glacier consisting of

Table 1. ^{10}Be ages, sample data and location information.

Sample	Sample type	Latitude (°N) (DD)	Longitude (°W) (DD)	Elevation (m a.s.l.)	Sample height (m)	Thickness (cm)	Shielding correction	Quartz (g)	Be carrier added (g)	^{10}Be/^{9}Be	^{10}Be/^{9}Be uncertainty	^{10}Be (atoms g^{-1})	^{10}Be uncertainty (atoms g^{-1})	^{10}Be age (ka)
Kurupa River Valley														
10FWW-01	Moraine boulder	68.19317	154.51979	1499	1.5	1.5	0.979	30.0549	0.3687	1.421E-13	4.150E-15	4.69E+04	1.38E+03	2.8 ± 0.1
10FWW-02	Moraine boulder	68.19309	154.51994	1508	1.0	1.0	0.979	35.0691	0.3704	1.589E-13	2.373E-15	4.52E+04	6.86E+02	2.6 ± 0.1
10FWW-03	Moraine boulder	68.19267	154.51959	1509	1.0	1.0	0.979	35.5775	0.3718	1.335E-13	3.392E-15	3.75E+04	9.65E+02	2.2 ± 0.1
10KRV-03	Bedrock	68.22772	154.50264	1267	0.5	2.0	0.993	29.3285	0.8008	2.904E-13	8.403E-15	2.15E+05	6.22E+03	15.9 ± 0.4
10KRV-05	Bedrock	68.21104	154.49136	1296	2.0	2.0	0.983	35.0942	0.8029	3.368E-13	6.296E-15	2.09E+05	3.90E+03	15.2 ± 0.3
10KRV-07	Erratic Boulder	68.20375	154.48836	1318	1.5	2.0	0.972	35.2659	0.3711	6.702E-13	1.241E-14	1.91E+05	3.54E+03	13.8 ± 0.3
10KRV-08	Bedrock	68.20383	154.48865	1314	1.5	3.0	0.972	35.0602	0.3722	6.596E-13	1.218E-14	1.90E+05	3.50E+03	13.9 ± 0.3
10KRV-09	Bedrock	68.19965	154.52574	1476	1.0	3.0	0.981	35.0313	0.3724	7.570E-13	1.408E-14	2.18E+05	4.05E+03	13.8 ± 0.3
Atigun River Valley region														
11TE-01	Moraine boulder	68.33395	149.76705	1846	1.0	2.0	0.993	40.0375	0.8035	1.616E-13	4.755E-15	8.78E+04	2.58E+03	4.1 ± 0.1
11TE-02	Moraine boulder	68.33423	149.76735	1849	0.75	2.5	0.993	30.0701	0.7962	1.545E-13	6.141E-15	1.11E+05	4.41E+03	5.1 ± 0.2
11TE-03	Moraine boulder	68.33510	149.76785	1833	1.5	2.5	0.993	30.0480	0.8031	1.361E-13	2.526E-15	9.85E+04	1.83E+03	4.6 ± 0.1
11TE-04	Moraine boulder	68.33532	149.76832	1826	1.7	1.5	0.993	35.1430	0.7982	9.146E-14	2.618E-15	5.63E+04	1.61E+03	2.6 ± 0.1
11RMV-02	Bedrock	68.30927	149.14107	1246	1.5	4.0	0.983	35.0755	0.8014	3.238E-13	5.990E-15	2.00E+05	3.71E+03	15.6 ± 0.3
11RMV-07	Erratic Boulder	68.33232	149.15842	1242	1.0	1.0	0.990	32.9613	0.7965	2.975E-13	9.993E-15	1.95E+05	6.54E+03	14.7 ± 0.5
11RMV-08	Bedrock	68.33235	149.15790	1235	1.0	1.5	0.990	35.1294	0.7958	3.644E-13	6.764E-15	2.24E+05	4.15E+03	17.1 ± 0.3
11RMV-13	Bedrock	68.36522	149.25767	1019	2.0	1.5	0.995	32.2053	0.8042	3.630E-13	6.732E-15	2.46E+05	4.55E+03	22.6 ± 0.4
11RMV-15	Bedrock	68.38003	149.31113	876	0.5	1.0	0.996	35.0198	0.8037	7.759E-13	1.759E-14	4.82E+05	1.09E+04	50.4 ± 1.2

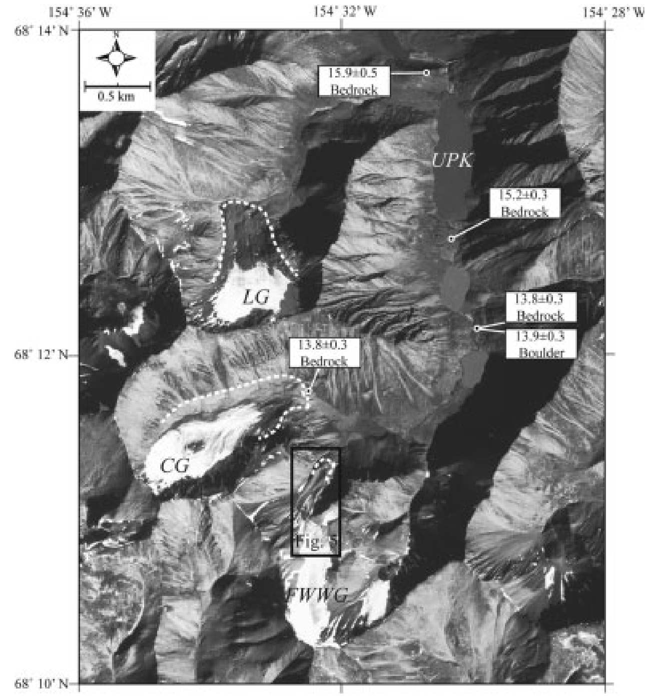

Figure 3. Aerial photograph of the headwaters of the Kurupa River valley. The Itkillik II ice limit was ~25 km north downvalley. Numbers are ^{10}Be ages reported in cal ka BP. LIA moraines are outlined by white dashed lines. USGS aerial photograph acquired in 1982. LG, Lupine glacier; CG, Cotton Grass glacier; FWWG, Fireweed West glacier; UPK, Upper Kurupa Lake.

multiple ridges, the outermost of which has lichen diameters consistent with pre-LIA deposition (Supporting Information, Figs S1 and S2). Our ^{10}Be analyses on three moraine boulders from the outermost moraine fronting this glacier yield ages of 2.2 ± 0.1, 2.6 ± 0.1 and 2.8 ± 0.1 ka (Table 1, Fig. 5). Boulder

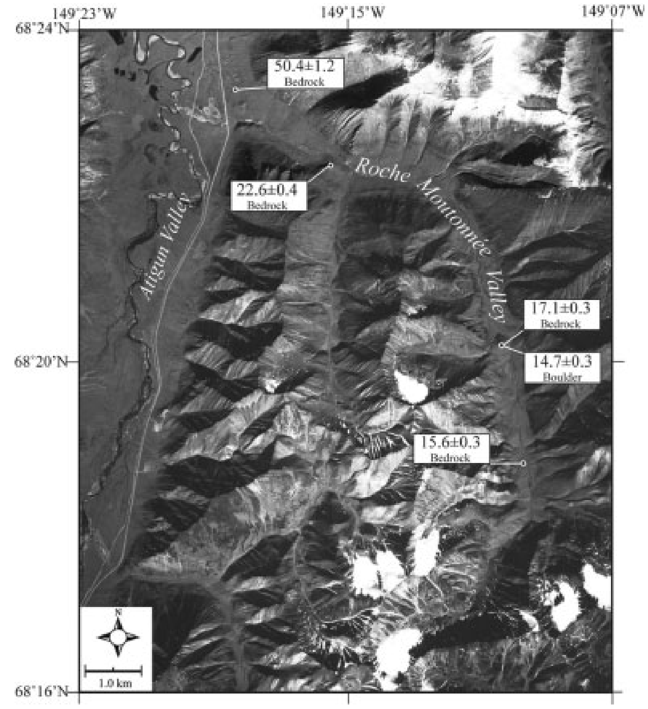

Figure 4. Aerial photograph of the Atigun River valley and headwaters of the Roche Moutonnée valley. The Itkillik II ice limit was 15 km downvalley (north). ^{10}Be ages are reported in cal ka BP. USGS aerial photograph acquired in 1978.

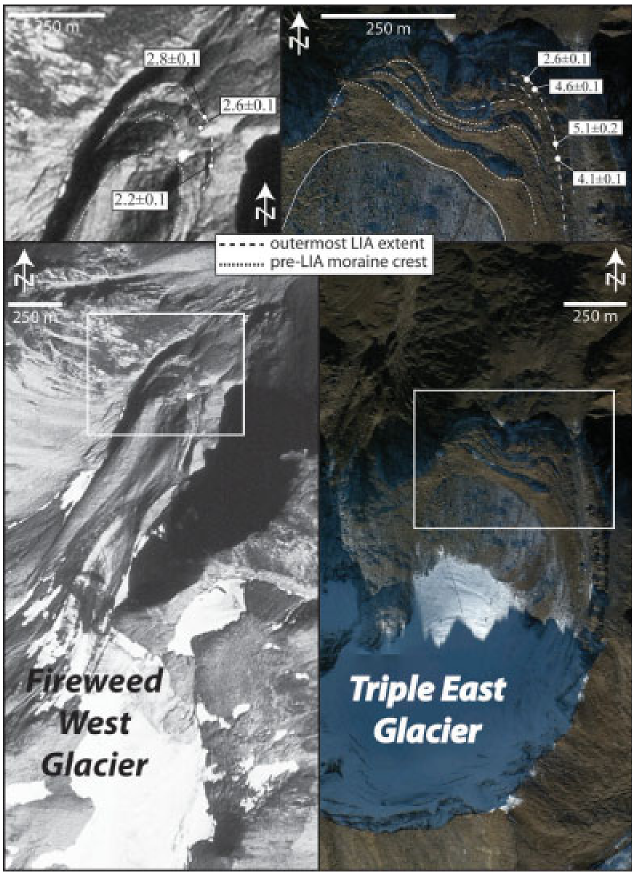

Figure 5. Vertical views of Fireweed West and Triple East glaciers showing Neoglacial moraine crests and locations of samples for ^{10}Be ages (reported in cal ka BP). USGS aerial photographs acquired in 1982 (Fireweed West), and Triple East Glacier is shown in a Google Earth image acquired on 19 September 2010. This figure is available in colour online at wileyonlinelibrary.com.

dimensions range from $2 \times 1.5 \times 1.5$ m for the smallest boulder to $2 \times 2 \times 1.5$ m for the largest boulder.

At Triple East Glacier, we adopted the mapping from Ellis and Calkin (1984), who showed five moraine crests on the north-eastern portion of the latero-end moraine complex (supporting Fig. S3). We sampled four boulders from the outermost of these moraine crests, which yield ages of 2.6 ± 0.1, 4.1 ± 0.1, 4.6 ± 0.1 and 5.1 ± 0.2 ka (Table 1, Fig. 5). Boulder dimensions range from $1 \times 1 \times 1$ m for the smallest boulder to $2.5 \times 2 \times 1.7$ m for the largest boulder.

Interpretation and discussion

Deglaciation

In the upper Kurupa River valley, the ^{10}Be ages suggest that deglaciation of the range front occurred prior to ~16 ka. Glaciers retreated into the upper catchment by 15.9 ka, and ice retreated to within Neoglacial limits, or perhaps disappeared entirely, around 13.8 ka. The youngest age on bedrock of 13.8 ± 0.3 ka, from just beyond the outermost Holocene moraine, implies that the cirques were ice-free or mostly ice-free at this time (Fig. 2). We believe these ages are not influenced by inheritance because they decrease upvalley and the paired erratic and bedrock samples yield identical ages. The timing of deglaciation of the Upper Kurupa River valley is consistent with the minimum radiocarbon age from Okokmilaga valley of 15.3 ± 1.8 cal ka BP (Hamilton, 1986). Glacier retreat appears to have slowed between 15.2 and 13.8 cal ka BP because of the close proximity of these ages, although no geomorphic evidence (e.g. moraine) suggests a standstill or re-advance in this vicinity.

In the Roche Moutonnée valley, several of the valley-bottom bedrock ages are apparently too old to represent the timing of deglaciation because they are inconsistent with previously published radiocarbon ages (Hamilton, 1986, 2003). We attribute the older-than-expected ^{10}Be ages in Roche Moutonnée valley to inheritance. We interpret the ages of samples 11RMV-8 (17.1 ± 0.3 ka), 11RMV-13 (22.6 ± 0.4 ka) and 11RMV-15 (50.4 ± 1.2 ka) to be influenced by inheritance (Table 1); although we have more confidence in the ages of 11RMV-2 (15.6 ± 0.3 ka) and 11RMV-7 (14.7 ± 0.5 ka) because they are similar to independent chronologies in the Brooks Range of when deglaciation occurred, we cannot rule out completely that they lack inheritance. Minimum limiting radiocarbon ages from Hamilton (1986) suggest ice was perhaps 10 km downvalley in Atigun River valley at ~15.5 ka cal BP. If the age of 14.7 ± 0.5 ka from an erratic is not influenced by inheritance, it implies rapid deglaciation at ~15 ka.

Inherited cosmogenic isotopes could be the result of partially cold-based glaciers with low rates of subglacial erosion that led to incomplete resetting of the bedrock surface. In addition, the Kanayut Conglomerate, including the beds that we sampled in the Roche Moutonnée valley (Fig. 2), is known to be particularly erosion resistant (Brosgé *et al.*, 1979), whereas the siliclastic and arkosic sandstones that we sampled from the Kurupa River valley are more easily eroded. Differences in erodability may explain the differences in inheritance between the two valleys.

Regional comparison

Previously published glacier chronologies in the Brooks Range indicate an early maximum extent and early retreat from late Pleistocene positions compared with elsewhere in Alaska (Briner and Kaufman, 2008). However, radiocarbon ages of ~15 cal ka BP are only minimum ages on late Pleistocene moraine abandonment at the range front (Hamilton, 1982, 1986). Previous ^{10}Be dating in the north-eastern Brooks Range

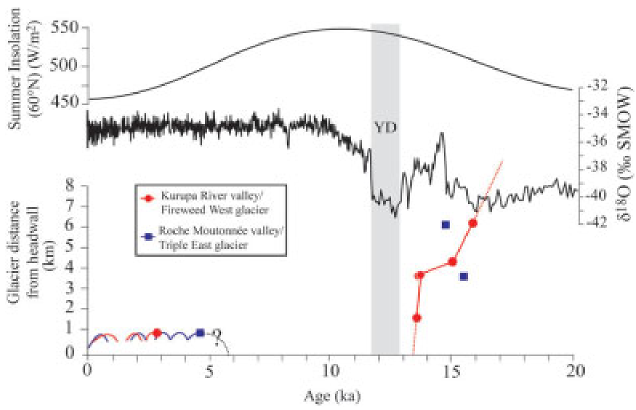

Figure 6. Glacier time–distance diagram for the two study valleys. The ^{10}Be ages from the Kurupa River valley suggest that glaciers had retreated into the cirques prior to the Younger Dryas (YD). Glacier expansion was underway during the middle Holocene, based on data from Triple East (TE) glacier. Other late Holocene glacier fluctuations are based on moraine mapping and lichenometry by Ellis and Calkin (1984) and Lamb (1984). The Greenland isotope record from Stuiver and Grootes (2000) and summer insolation from Berger and Loutre (1991) are shown for comparison. This figure is available in colour online at wileyonlinelibrary.com.

revealed glaciers there reached their maximum at ~30–27 ka, and had retreated to within ~12 km of their cirque headwalls prior to ~21 ka (ages adjusted according to the production rate used here) (Balascio *et al.*, 2005; Briner *et al.*, 2005). Our ^{10}Be ages from the Kurupa River valley indicate that ice had retreated to within ~5 km of their cirque headwalls prior to ~16 ka and had retreated into their cirques by ~14 ka (Fig. 6). In southern Alaska, some glaciers were still expanded far downvalley ~16 ka (Briner and Kaufman, 2008). Evidence that glaciers were well downvalley at ~17–16 ka in many south and central Alaska valleys, while at the same time glaciers were within a few kilometers of their cirques in the Brooks Range, reflects the spatial heterogeneity in Alaskan climate.

Our ^{10}Be ages provide the opportunity to investigate potential Lateglacial advances as observed elsewhere in north hemisphere, such as during the YD. The ^{10}Be age of 13.8 ± 0.3 ka from the upper Kurupa River valley immediately outboard of the Neoglacial moraines of Cottongrass Glacier (Figs 2 and 3) suggests that any advance during the YD was less extensive than advances during Neoglaciation. Previous studies in the Brooks Range failed to identify glacier advances during the YD (e.g. Hamilton, 2003; Balascio *et al.*, 2005). The lack of evidence for a glacier advance during the YD in the Brooks Range is compatible with findings from Kokorowski *et al.* (2008), who showed similar-to-present or even higher-than-present temperatures across central and northern Alaska during the YD. Cooling during the YD is more strongly expressed in southern Alaska, but was heterogeneous (Kaufman *et al.*, 2010). Elsewhere in the Arctic, including Baffin Island and Svalbard, glaciers were smaller then than during the LIA (Mangerud and Landvik, 2007; Briner *et al.*, 2009). On the other hand, isotope records from permafrost on the North Slope indicate cooling during the YD (Meyer *et al.*, 2010), but this is interpreted to reflect lower winter temperatures that might not have led to significant glacier growth. On the Arctic Foothills, lake levels fell and streams incised their floodplains during the YD (Mann *et al.*, 2002), further indication of lowered effective moisture at that time.

Neoglaciation

At both Fireweed West and Triple East glaciers, ^{10}Be ages on the outermost moraine include one age noticeably younger than the others (Table 1). Averaging the closely spaced ages and excluding the youngest age from each dataset yields ages of 2.7 ± 0.2 ka ($n = 2$) at Fireweed West glacier and 4.6 ± 0.5 ka ($n = 3$) at Triple East glacier (Fig. 6). Alternatively, we could use the oldest age as a best estimate of the moraine emplacement age, assuming that there was no inheritance and that moraine degradation was the main cause for the age spread (Heyman *et al.*, 2010; Applegate *et al.*, 2012). However, given that these moraines were deposited close to cirque headwalls and contain abundant angular blocks that were probably transported supraglacially, we hesitate to rule out inheritance. Alternatively, the moraines are composite features deposited during multiple periods when glaciers expanded onto their former end moraines, and each ^{10}Be age represents the timing of deposition of that particular boulder on the moraine crest. Although the tightly nested ridges of the end moraine complex suggest that, as a whole, it is a composite feature, we hypothesize that each ridge crest was built during a single maximum. Regardless, despite the difficulty in interpreting the ^{10}Be ages, we favor the interpretation that the average of the oldest ages is the best approximation for the age of moraine stabilization.

We compare our ^{10}Be chronology with the previously published moraine ages determined from lichenometry (Fig. 7).

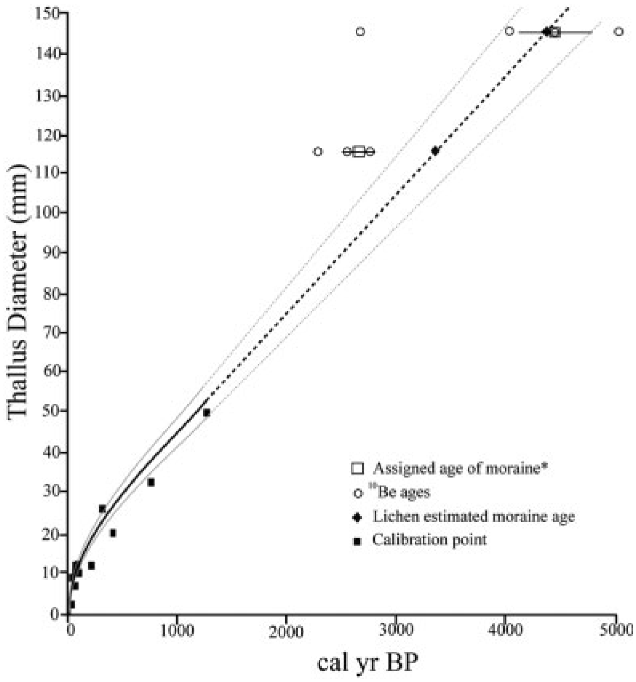

Figure 7. Growth curve for *Rizocarpon geographicum* (*sensu lato*) and ^{10}Be ages associated with lichenometrically dated moraines (modified from Calkin and Ellis, 1980; Calkin and Solomina, 2003). Gray lines bracketing the growth curve are $\pm$ 20% uncertainty. *Assigned moraine age is the average of two ages, excluding the youngest age. Error bars show 2-sigma range.

Lichenometry has an estimated ~20% age uncertainty, and calibration points are limited to the last millennium (Calkin and Ellis, 1980; Solomina and Calkin, 2003). Using the lichen growth curve from Ellis and Calkin (1984), which was converted to calendar years by Solomina and Calkin (2003), our ^{10}Be ages on Holocene moraines overlap with the 20% uncertainty of the lichenometric age assignments. For example, Lamb (1984) reported lichen diameters that yield an age of 3.7 ± 0.7 ka cal BP on the Fireweed West moraine, and our average ^{10}Be age is 2.7 ± 0.2 ka. Our average ^{10}Be age of the Triple East moraine of 4.6 ± 0.5 ka also agrees with the previously estimated age of 5.0 ± 1.0 ka cal BP (Ellis and Calkin, 1984). Based on these ^{10}Be ages, it seems that the lichen growth curve developed by Ellis and Calkin (1984) can be reliably projected beyond the control points using a linear function as they had suggested, at least to the middle Holocene. The exponential from of the growth curve that was considered by Solomina and Calkin (2003) yields ages that are much too old relative to the ^{10}Be ages. The second-order polynomial form that was suggested by Sikorski *et al.* (2009) yields ages that are intermediate between the linear and exponential fits.

These ^{10}Be ages demonstrate that glaciers in the Brooks Range experienced advances in the middle Holocene that were as extensive as during the LIA. In part, the pre-LIA moraine preservation may be due to the inability of smaller subsequent glaciers to overrun previously emplaced moraines (Porter, 2007). Furthermore, those moraine complexes that are ice cored might have post-depositional, net downslope movement, and thus escape subsequent disturbance; however, many moraine sequences throughout the Brooks Range also lack ice cores. Regardless, glaciers during the middle Holocene were at least similar in size to those during the LIA. This is at odds with insolation as the main driver of glacier growth, which seems to explain the fact that glaciers were most extensive during the LIA in many areas throughout the northern hemisphere (Barclay *et al.*, 2009; Briner *et al.*, 2009; Menounos *et al.*, 2009). On the other hand, that some northern hemisphere glaciers exceeded or became very close to their late Holocene maxima prior to the LIA is not exceptional, and has been documented at a number of locations (e.g. Briner *et al.*, 2009; Maurer *et al.*, 2012; Schimmelpfennig *et al.*, 2012).

The relatively minor expansion of Brooks Range glaciers during the LIA compared with expansions during the Neoglacial could be due to a progressive decrease in moisture availability through the late Holocene as Arctic sea ice expanded (England *et al.*, 2008; Antoniades *et al.*, 2011; Funder *et al.*, 2011). Sikorski *et al.* (2009) concluded that limited lowering of equilibrium line altitudes in the Brooks Range compared with elsewhere in Alaska during the LIA was due to drier-than-present conditions, consistent with the notion of moisture starvation due to expanded Arctic sea ice. This decrease in moisture may have been most pronounced in the Arctic; although middle Holocene advances are recognized elsewhere (Barclay *et al.*, 2009; Menounos *et al.*, 2009), to our knowledge the Brooks Range is the only location in the northern hemisphere yet identified where multiple glaciers seem to have expanded beyond their LIA positions as early as the middle Holocene.

Conclusion

Our ^{10}Be ages provide new evidence for the timing of deglaciation and Neoglaciation in the north-central Brooks Range. The ^{10}Be inventory in bedrock surfaces in Roche Moutonnée valley appears to be significantly influenced by isotopic inheritance, which we attribute to the erosion-resistant beds within the Kanayut Conglomerate, possibly to weakly erosive polythermal glaciers. A ^{10}Be age immediately outboard of late Holocene moraines in the upper Kurupa River valley (Fig. 2), supported by additional ^{10}Be ages slightly downvalley, suggests that glaciers in the Brooks Range were either absent or smaller during the YD than during the Neoglacial. The lack of glacial advance associated with the YD agrees with other records in the Arctic, despite climate records showing pronounced YD cooling, including permafrost records that indicate significant winter cooling along the North Slope (Meyer *et al.*, 2010). It has been suggested that cooling during the YD in the Arctic was marked by strong seasonality with cold winters and increased sea-ice cover (Denton *et al.*, 2005) as well as aridity (Alley *et al.*, 1993), but with relatively mild summer conditions. In addition, glaciers in the Brooks Range are probably precipitation-sensitive (Sikorski *et al.*, 2009), and thus it is possible that strong aridity contributed to the lack of any extensive glacier expansion during the YD.

The ^{10}Be ages for a middle Holocene moraine dating to $\sim$4.6 ka provide the first numerical ages for the maximum Neoglacial expansion following early Holocene warmth (e.g. Kaufman *et al.*, 2004). Our ^{10}Be ages agree with lichenometric ages of middle Holocene moraines within the 20% uncertainty of lichenometry based on the original linearly extrapolated growth curve of Ellis and Calkin (1984). This suggests that the lichenometrically dated moraine sequence, which features dozens of pre-LIA moraines across the Brooks Range, is generally robust. The ^{10}Be ages confirm that glaciers in the Brooks Range reached similar extents during several glacier expansions during the Neoglacial. The effect of orbitally driven decreasing summer insolation in the Brooks Range may have been surpassed by decreasing moisture availability.

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Abbreviations. LIA, Little Ice Age; YD, Younger Dryas.

Supporting information

Additional supporting information can be found in the online version of this article:

Supporting Figure S1. Aerial photograph (same as in Figure 5) and lichen map from Lamb (1984) of Fireweed West Glacier. Bottom panel shows oblique photograph taken from airplane flight in July 2010 showing Fireweed West Glacier moraines.

Supporting Figure S2. Surficial geomorphic map from Lamb (1984) of the upper Kurupa River valley. See Figure 3 for aerial photograph of same area.

Supporting Figure S3. Aerial photograph and lichen map from Ellis and Calkin (1984) of the Triple East Glacier.

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