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Late Pleistocene and Holocene glaciation of the Fish Lake valley, northeastern Alaska Range, Alaska

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ABSTRACT: We reconstructed a chronology of glaciation spanning from the Late Pleistocene through the late Holocene for Fish Lake valley in the north-eastern Alaska Range using ^{10}Be surface exposure dating and lichenometry. After it attained its maximum late Wisconsin extent, the Fish Lake valley glacier began to retreat ca. 16.5 ka, and then experienced a readvance or standstill at 11.6 ± 0.3 ka. Evidence of the earliest Holocene glacial activity in the valley is a moraine immediately in front of Little Ice Age (LIA) moraines and is dated to 3.3–3.0 ka. A subsequent advance culminated at ca. AD 610–900 and several LIA moraine crests date to AD 1290, 1640, 1860 and 1910. Our results indicate that ^{10}Be dating from high-elevation sites can be used to help constrain late Holocene glacial histories in Alaska, even when other dating techniques are unavailable. Close agreement between ^{10}Be and lichenometric ages reveal that ^{10}Be ages on late Holocene moraines may be as accurate as other dating methods. Copyright © 2009 John Wiley & Sons, Ltd.

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KEYWORDS: Holocene glaciation; cosmogenic exposure dating; lichenometry; Alaska Range.

Introduction

Alpine glaciers serve as a valuable proxy to infer past climatic conditions because they are sensitive to regional changes in temperature and precipitation (Lowell, 2000). Furthermore, detailed records of past glaciation provide the context for ongoing glacier response to climate change and aid in the assessment of climate model results. For example, a recent global circulation model (GCM; e.g. Otto-Bliesner *et al.*, 2006) depicts Alaska as experiencing warmer-than-present temperatures during the Last Glacial Maximum (LGM). The existing record of alpine glaciation in Alaska reveals considerable spatial variability in the timing of Late Pleistocene and Holocene glacier expansion (Briner and Kaufman, 2008). To further tighten the spatial variability of mountain glaciation across Alaska, additional studies with absolute chronologies of Pleistocene and Holocene glaciation are required.

The Alaska Range contains a long and detailed record of Late Pleistocene and Holocene glaciation (Calkin, 1988; Briner *et al.*, 2005; Briner and Kaufman, 2008). Because of relatively arid conditions on the northern, rain-shadow side of the range, glacier expansion during the LGM was relatively limited and moraine sequences spanning the Late Pleistocene through the Little Ice Age (LIA, ca. AD 1250–1850; Bradley, 2000) are often preserved within single valleys. The LIA advance generally represents the most extensive Holocene glaciation in Alaska (e.g. Calkin *et al.*, 2001). However, in some cases, older Neoglacial moraines are preserved (e.g. Denton and Karlén, 1977; Ellis and Calkin, 1984) and provide a rare opportunity to date pre-LIA glacier maxima. Precise dating of these moraines is critical to understanding temporal and spatial differences of glacial fluctuations between separate mountain ranges, and between regional and global climate patterns, which are commonly asynchronous (Gillespie and Molnar, 1995).

In this study we use cosmogenic ^{10}Be exposure dating of moraine boulders and lichenometry to reconstruct the Late Pleistocene to late Holocene glacial history of a small valley located on the northern flank of the Alaska Range (Fig. 1). Suitable boulders for ^{10}Be dating of moraines in Alaska are generally sparse because most mountain valleys across the state tap into friable metasedimentary rocks and because exposed rock in Alaska is strongly affected by weathering. This study benefits from its location within a glacially eroded granitic pluton, which yields large, durable boulders that are well preserved on moraine surfaces. Moreover, the study area lies only 75 km east of the Delta River valley, which serves as the regional reference locality for the late Wisconsin glaciation (locally the Donnelly glaciation; Péwé, 1953). The current geochronology for this glaciation is limited (e.g. Péwé, 1965,

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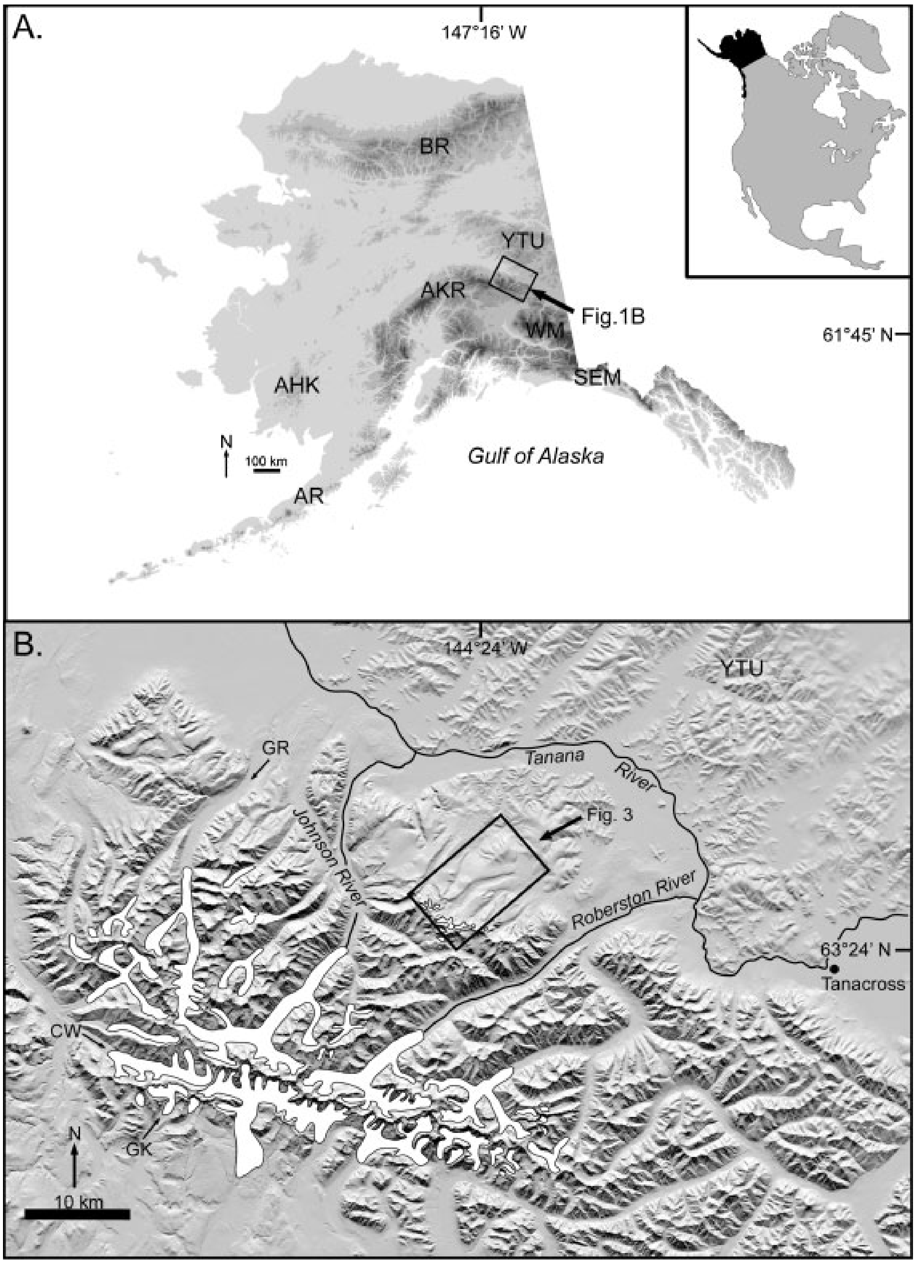

Figure 1 (A) Location map of Alaska showing the Brooks Range (BR), Yukon–Tanana Upland (YTU), Ahklun Mountains (AHK), Alaska Range (AKR), Wrangell Mountains (WM), St Elias Mountains (SEM) and the Aleutian Range (AR). (B) East-central Alaska Range with locations of study area, Gulkana Glacier (GK), Canwell (CW), Gerstle Valley (GR) and Tanacross. Main valley glaciers in east-central Alaska and in the Fish Lake valley massif are shown in white

1975), and this study provides additional chronological control that can be correlated reliably with the nearby reference locality.

Previous ^{10}Be dating in Alaska and elsewhere has focused almost exclusively on Late Pleistocene moraines (e.g. Briner *et al.*, 2001, 2005; Balascio *et al.*, 2005a; Briner and Kaufman, 2008), and only a few attempts have been made to obtain the analytical precision necessary to apply the technique to Holocene materials (e.g. Douglass *et al.*, 2005; Seong *et al.*, 2007). Our study area at high latitude and high elevation is suitable for obtaining cosmogenic exposure ages because effects of the geomagnetic field on production rate scaling are negligible at high latitudes and study area elevations ensure ^{10}Be production rates ~2.6–3.9 times greater than at sea level (Stone, 2000; Gosse and Phillips, 2001). This study also benefits from a well-developed lichen growth curve for the region initially developed by Begét (1994) and later modified by Solomina and Calkin (2003). We use the modified curve from Solomina and Calkin (2003) to assess the validity of ^{10}Be ages by comparing them with new lichenometric ages from the same surfaces.

Physical setting

The Alaska Range is Alaska's highest mountain range and spans over 700 km from east to west (Fig. 1). Elevations rise to over 6100 m above sea level (a.s.l.) at the Mt Denali summit, and the range decreases in elevation towards the Aleutian Range to the south-west and towards the Wrangell Mountains to the east. The Alaska Range contains approximately 13 725 km^2 of ice, including thousands of glaciers ranging from isolated cirque glaciers <1 km^2 to large valley glaciers exceeding 500 km^2 (Molina, 2007).

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The study region in east-central Alaska is affected by maritime/transitional and continental climate regimes (Fig. 1). The maritime/transitional climate exists south of the range and is characterised by relatively mild temperatures (3.8°C mean annual temperature) and high precipitation (up to 2400 mm a^{-1}), while a continental climate distinguished by marked summer–winter temperature swings and low precipitation (~320 mm a^{-1}), dominates north of the range (Alaska Climate Research Center, 2008). These climate regimes create a sharp south–north temperature and precipitation gradient across the range (Clapperton, 2001). Orographic effects prevent the penetration of warm, moist air from the Gulf of Alaska to the Alaska and Wrangell–St Elias Ranges, and the region is also far removed from cold, arctic conditions that prevail in the Brooks Range. Tanacross (Fig. 1), located on the north side of the Alaska Range and ~50 km east of the study area, has a mean winter (JFM) temperature of −23.8°C, a mean summer temperature (JJA) of 13.1°C, and 280.7 mm annual precipitation, with most occurring during the summer and early fall months (1949–2003; Western Regional Climate Center, www.wrcc.dri.edu, 2007).

The Fish Lake valley (63° 31′ N, 144° 29′ W, informal name) extends north from a massif that lies just north of the main topographic divide of the eastern Alaska Range (Fig. 1). The Johnson and Robertson rivers border the massif to the west and east, respectively, and drain northward into the Tanana River. Bedrock in this area and the adjacent Yukon–Tanana Upland is composed of a variety of Precambrian metamorphic rocks (mainly schists and quartzites) with intrusions of Cretaceous granite (Richter *et al.*, 1975). Cirques in the study massif are occupied by small (<2 km^2) debris-covered glaciers and a few clean-ice glaciers that generally lie within 2 km of their headwalls. In the Fish Lake valley, Pleistocene moraines extend approximately 15 km downvalley and are marked by hummocky morphology and abundant large boulders with extensive lichen cover.

Two separate and well-exhibited Holocene moraine sequences are preserved within the Fish Lake valley headwaters. One lies within the main valley, where end moraines range in elevation from ~1500 to 1465 m a.s.l. The second lies in a small tributary valley located ~2 km south-east of the main valley, where moraines range in elevation from ~1780 to 1600 m a.s.l. Moraines in both the main and tributary valleys are ice-cored and generally unvegetated, but moraines in the tributary valley are more sharp-crested, voluminous and bouldery. Presently, relatively debris-free glacier snouts have retreated and detached from their ice-cored moraines.

Methods

Field mapping of moraine crests, measuring lichen diameters and sampling boulders for ^{10}Be dating were completed in 2006 and 2007 (Fig. 2). Moraines in the main Fish Lake valley (Figs 3 and 4) and in the tributary valley (Fig. 4) were first mapped on aerial photos, and then field checked using US Geological Survey (USGS) 1:63 360-scale topographic maps. Because some segments of the ice-cored Holocene moraines showed geomorphic evidence of post-depositional downslope migration, we targeted sections of the moraines with the

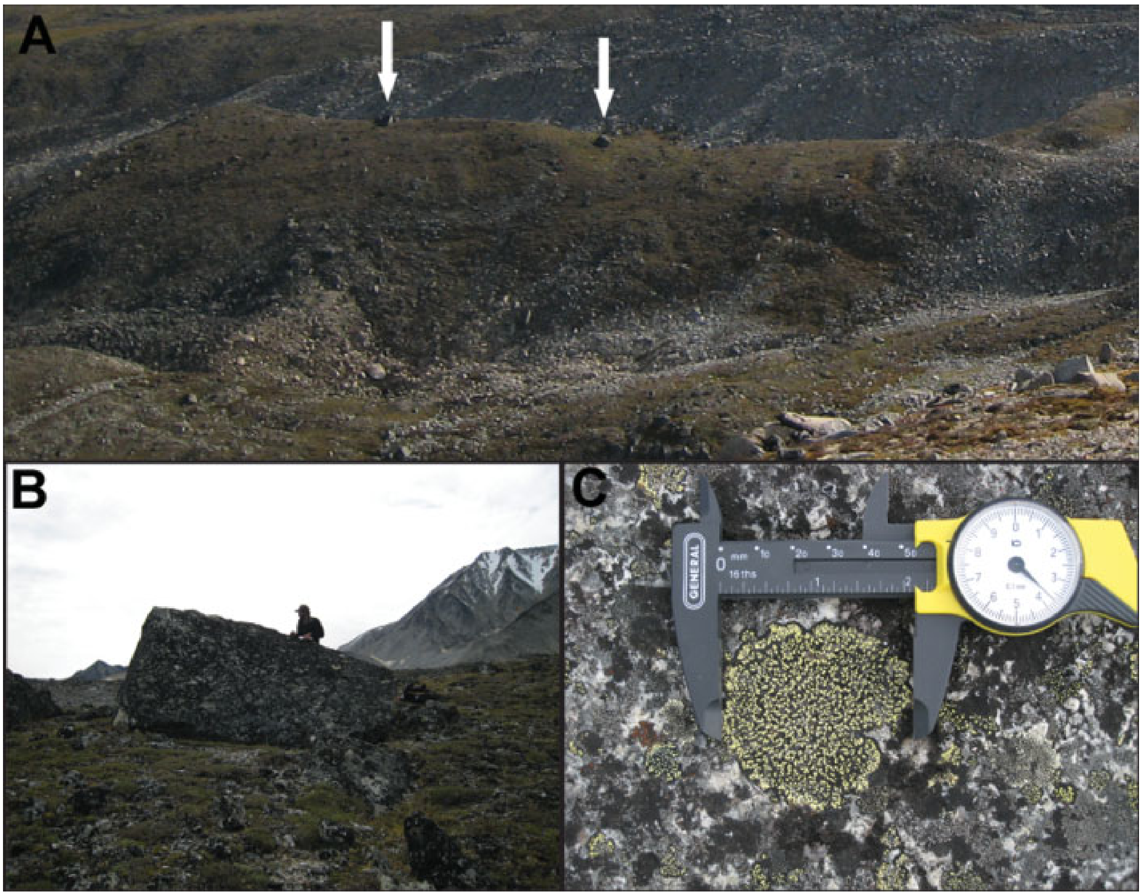

Figure 2 (A) Prominent moraine ridge in the tributary valley. Arrows point to two boulders (US07-04; US07-05) sampled for ^{10}Be dating. (B) Boulder US07-04 with a person standing behind for scale. (C) Typical *Rhizocarpon geographicum s.l.* thallus measured in the field

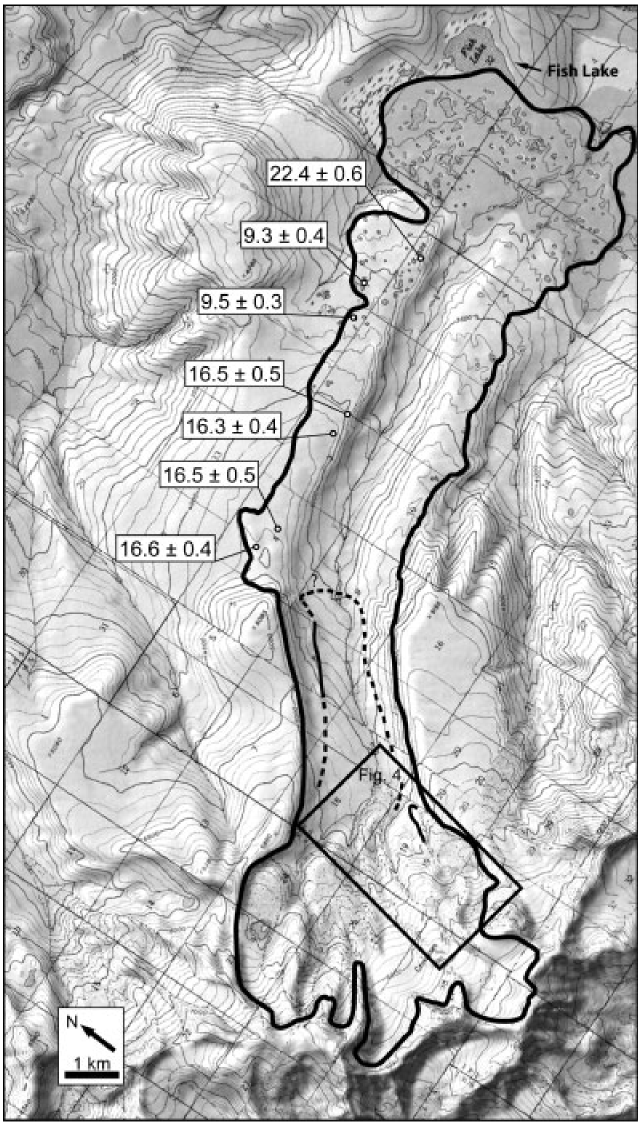

Figure 3 Fish Lake valley with LGM ice limit (solid line), boulder locations and ^{10}Be ages (ka). Also shown are upvalley moraines attributed to a Late Pleistocene readvance or standstill (dashed line). USGS Mt Hayes (C-2) 1: 63 360 topographic map with 100 ft contours

largest lichens as these were likely to have been the most stable compared to neighbouring sections of moraine with smaller lichen diameters. Boulders sampled for ^{10}Be dating were only collected from these most stable moraine sections.

^{10}Be surface exposure dating

Previous Alaskan studies (e.g. Briner *et al.*, 2001, 2002, 2005; Balascio *et al.*, 2005a; Matmon *et al.*, 2006) have demonstrated that careful sampling is critical to the successful application of ^{10}Be dating. As moraines degrade over time, boulders are either exhumed or transported downslope; both processes lead to exposure ages that are younger than the age of deposition (Putkonen and Swanson, 2003). Exhumation is of concern for Late Pleistocene moraines, especially where the moraines are hummocky, indicating that buried ice may have melted out over several thousands of years following deposition. In contrast, late Holocene moraines in the study area are clast-supported and have had little time to degrade, thereby reducing the risk of sampling an exhumed boulder. On the other hand, late Holocene moraines in the study area show evidence for ice cores that cause surface instability, although these moraine segments were avoided.

Twenty-three quartz-rich granodiorite boulders were sampled for ^{10}Be dating (Table 1). The study area is well above the tree line and is generally windswept of snow; however, we only sampled boulders taller than 1 m to minimise the shielding effects of snow cover. Boulders resting in surface depressions where snow might accumulate were avoided, as were boulders showing obvious signs of movement. The uppermost horizontal surface was sampled on all boulders, and an effort was made to sample in the middle of the boulder to reduce potential edge effects (Masarik and Wieler, 2003). Topographic shielding was recorded by measuring the angle of inclination from the boulder surface to the surrounding skyline. The elevation of the sample sites was measured using a handheld GPS receiver and 1:63 360 topographic maps with 100 ft contour intervals. Elevation uncertainty is estimated to be <10 m.

All samples for ^{10}Be dating were prepared at the University of Buffalo Cosmogenic Isotope Laboratory following modified procedures from Kohl and Nishiizumi (1992). To increase the precision of ^{10}Be measurements of young (i.e. Holocene) samples, we dissolved approximately 50–60 g of quartz per sample. For older, Late Pleistocene samples approximately 30–40 g of quartz was dissolved. ^{10}Be contents were measured at the Center for AMS at Lawrence Livermore National Laboratory and the Purdue Rare Isotope Measurement Laboratory. Reported ^{10}Be/^{9}Be ratios were normalised to ^{10}Be ICN standards (^{10}Be $t_{1/2} = 1.36 \times 10^6$ a, 07KNSTD3110) prepared by K. Nishiizumi (Nishiizumi *et al.*, 2007). Ratios of ^{10}Be/^{9}Be for dissolution process blanks averaged $2.12 \pm 0.03 \times 10^{-14}$ ($n = 6$).

We used the CRONUS age calculator (http://hess.ess.washington.edu/math/) to calculate surface exposure ages with and without erosion using the ^{10}Be production rate specified for sea level at high latitude (SLHL) of 4.96 atoms g^{-1} a^{-1} (Table 1; Balco *et al.*, 2008). This production rate is from the new averaging procedure of calibration samples (Balco *et al.*, 2008), resulting in an SLHL value that is ~3% lower than the production rate previously used elsewhere in Alaska (i.e. 5.1 atoms g^{-1} a^{-1}). Site-specific production rates are scaled for latitude, elevation, topographic shielding and sample thickness (Balco *et al.*, 2008). We employ the constant production model of Lal (1991) and Stone (2000) because the influence of the Earth's magnetic field on ^{10}Be production rate is negligible at the study area's relatively high latitude (~63° N; Gosse and Phillips, 2001). Different scaling schemes may slightly alter the ^{10}Be ages (by <8%), but the relative chronology of our exposure ages remains the same. ^{10}Be ages are reported with analytical uncertainty at one standard deviation that ranges between 3% and 7%. Additional error comes from uncertainties in ^{10}Be production rate (±6%; Gosse and Stone, 2001) and elevational scaling procedures (± 5%; Gosse and Phillips, 2001). We estimate a total systematic uncertainty of 5–10%.

Lichenometry

We used an existing growth curve for lichen species *Rhizocarpon geographicum* for the central Alaska Range for lichenometric ages of late Holocene moraine stabilisation (Begét, 1994; Solomina and Calkin, 2003). In their review of lichen growth curves in Alaska, Solomina and Calkin (2003) identified and discarded an anomalous control point dated at

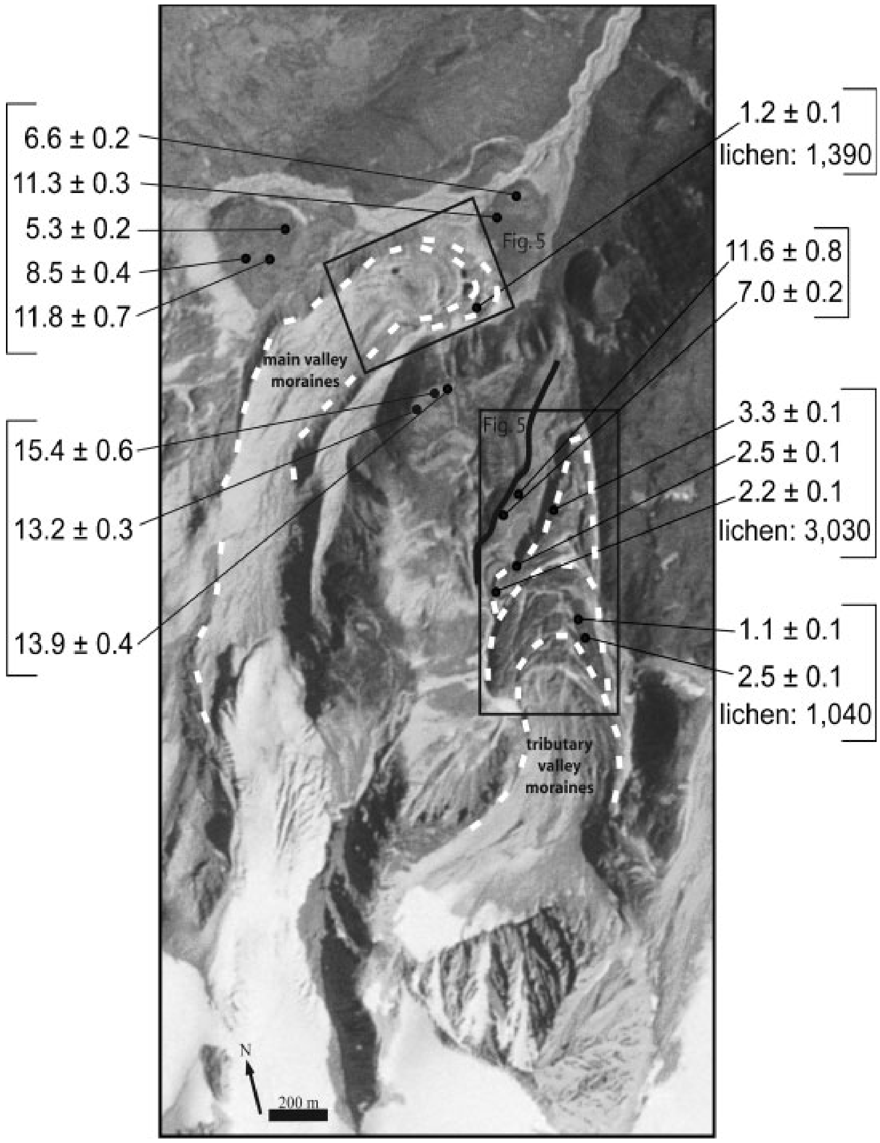

Figure 4 Cirque region of Fish Lake valley (1980 aerial photograph) with mapped moraines in the main and tributary valleys (location shown in Fig. 3). Boulder locations and ^{10}Be ages (ka) are marked by black dots. Lichen ages (years of exposure from AD 2007) are shown with corresponding ^{10}Be ages. Black line marks a latest Pleistocene ice limit and white dashed lines mark Holocene limits. The latest Pleistocene limit in the tributary valley is considered coeval with the latest Pleistocene readvance or standstill limit shown in Fig. 3. Only the outer two Holocene moraines in the main valley are shown here (see Fig. 5 for full moraine sequence)

800^{14}C a BP used by Begét (1994) in his growth curve, and we adopt this suggestion to calculate stabilisation ages. Two equations were used to calculate stabilisation ages. A logarithmic curve was used for thallus diameters (D) $\leq$ 50 mm (great growth portion of the calibration curve, age $= 41.861 \times 10^{(0.011679 \times D)}$. For lichen diameters > 50 mm (age $= -1430 + 32.074 \times D$), a linear equation was used.

Identification of the green-black lichens to the species level was not possible in the field; lichens in this study are referred to as *Rhizocarpon geographicum sensu lato* (*s.l.*; Locke *et al.*, 1979). Lichen thalli were measured on seven moraine crests in Fish Lake valley and three moraine crests in the tributary valley (Fig. 5). Up to three stations were established on each moraine where lichens were suitable for measurement; however, three of the 10 moraines were too small and unstable for more than one measurement station. To ensure consistency, only lichen thalli with distinct margins and circular to slightly elliptical geometries were measured. Lichen diameters were measured with dial callipers and recorded to the nearest millimetre. Following Calkin and Ellis (1980), lichens exceeding 20% in size of the next largest lichen were discarded and considered anomalous. We use the single largest lichen on a moraine crest to calculate the age of stabilisation (Solomina and Calkin, 2003).

Results

^{10}Be ages

Twenty-three boulders sampled from Late Pleistocene and Holocene laterofrontal and ground moraines in Fish Lake valley range in age from 22.4 ± 0.6 to 1.1 ± 0.1 ka, assuming no correction for boulder surface erosion (ε_0; Table 1). ^{10}Be ages were also calculated using a minimum erosion rate (ε_{min}) of 1.0 mm ka^{-1} and a maximum erosion rate (ε_{max}) of 4.8 mm ka^{-1} (Table 1). Minimum erosion was derived from the arid, eastern Canadian Arctic, whereas maximum erosion was calculated from erratics in the Yukon–Tanana Upland (i.e. Bierman *et al.*, 1999; Briner *et al.*, 2005). We provide maximum and minimum

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Table 1 Sample information in the Fish Lake valley

Sample	Latitude (N)	Longitude (W)	Elevation (m a.s.l.)	Sample height (m)	Thickness (cm)	Shielding correction	Be carrier added (g)	^{10}Be (10^5 atoms g^{-1})	^{10}Be age (ka) ε_0	^{10}Be age (ka) ε_{min}	^{10}Be age (ka) ε_{max}
Late Pleistocene terminal moraine (main valley; Fig. 3)											
FL06-01	63° 32.939′	144° 21.434′	1064	1.2	3	1.000	0.3571	2.835 ± 0.071	22.4 ± 0.6	22.8 ± 0.6	24.6 ± 0.7
FL06-02	63° 32.237′	144° 22.136′	1058	1.5	2	1.000	0.3554	1.181 ± 0.051	9.3 ± 0.4	9.3 ± 0.4	9.6 ± 0.4
FL06-03	63° 32.713′	144° 24.977′	1234	1.0	2	1.000	0.3431	2.431 ± 0.070	16.5 ± 0.5	16.7 ± 0.5	17.6 ± 0.6
FL06-04	63° 32.734′	144° 25.616′	1262	1.5	2	1.000	0.3569	2.468 ± 0.064	16.3 ± 0.4	16.5 ± 0.4	17.5 ± 0.5
FL06-05	63° 32.666′	144° 27.303′	1333	1.0	2.5	1.000	0.3497	2.636 ± 0.074	16.5 ± 0.5	16.7 ± 0.5	17.7 ± 0.5
FL06-06	63° 32.825′	144° 28.141′	1347	1.0	2	1.000	0.3575	2.698 ± 0.070	16.6 ± 0.4	16.9 ± 0.4	17.8 ± 0.5
FL06-12	63° 33.100′	144° 23.485′	1129	1.0	2	1.000	0.3575	1.293 ± 0.038	9.5 ± 0.3	9.6 ± 0.3	9.9 ± 0.3
Deglacial erratics (Fig. 4)											
US07-09	63° 30.278′	144° 31.575′	1595	1.0	2	0.995	0.3505	2.967 ± 0.115	15.4 ± 0.6	15.6 ± 0.6	16.4 ± 0.7
US07-10	63° 30.283′	144° 31.570′	1593	1.0	3	0.995	0.3734[a]	2.721 ± 0.076	13.9 ± 0.4	14.1 ± 0.4	14.8 ± 0.4
US07-11	63° 30.266′	144° 31.604′	1598	1.2	3	0.995	0.3714[a]	2.588 ± 0.067	13.2 ± 0.3	13.3 ± 0.4	13.9 ± 0.4
Latest Pleistocene readvance in main valley (Figs 3 and Fig. 4)											
FL07-01	63° 30.615′	144° 30.956′	1449	1.5	3	0.993	0.3737[a]	1.147 ± 0.038	6.6 ± 0.2	6.6 ± 0.2	6.8 ± 0.2
FL07-02	63° 30.602′	144° 31.057′	1458	1.0	3	0.994	0.3722[a]	1.990 ± 0.054	11.3 ± 0.3	11.5 ± 0.3	11.9 ± 0.3
FL07-06	63° 30.666′	144° 32.123′	1544	2.0	2.5	0.997	0.3741[a]	1.000 ± 0.036	5.3 ± 0.2	5.3 ± 0.2	5.4 ± 0.2
FL07-07	63° 30.630′	144° 32.355′	1576	2.0	2.5	0.998	0.3520	1.567 ± 0.078	8.5 ± 0.4	8.6 ± 0.4	8.8 ± 0.4
FL07-08	63° 30.624′	144° 32.235′	1546	1.7	2.5	0.997	0.3562	2.167 ± 0.129	11.8 ± 0.7	12.0 ± 0.7	12.4 ± 0.8
Latest Pleistocene re-advance in tributary valley (Fig. 4)											
US07-04	63° 30.018′	144° 31.360′	1689	2.0	3	0.995	0.3562	2.448 ± 0.169	11.6 ± 0.8	11.7 ± 0.8	12.1 ± 0.9
US07-05	63° 29.995′	144° 31.404′	1687	1.8	2	0.994	0.3739[a]	1.494 ± 0.038	7.0 ± 0.2	7.0 ± 0.2	7.2 ± 0.2
Holocene moraines in main valley (Fig. 4)											
FL07-04	63° 30.439′	144° 31.226′	1494	1.2	2.5	0.993	0.3537	0.211 ± 0.012	1.2 ± 0.1	1.2 ± 0.1	1.2 ± 0.1
Holocene moraines in tributary valley (Fig. 4)											
US07-01	63° 29.703′	144° 31.108′	1768	1.5	2	0.991	0.3717[a]	0.553 ± 0.029	2.5 ± 0.1	2.5 ± 0.1	2.5 ± 0.1
US07-02	63° 29.708′	144° 31.115′	1776	1.5	2	0.991	0.3739[a]	0.257 ± 0.027	1.1 ± 0.1	1.1 ± 0.1	1.1 ± 0.1
US07-06	63° 29.892′	144° 31.335′	1724	1.2	2.5	0.993	0.3739[a]	0.542 ± 0.029	2.5 ± 0.1	2.5 ± 0.1	2.5 ± 0.1
US07-07	63° 29.940′	144° 31.210′	1721	1.0	3	0.994	0.3721[a]	0.725 ± 0.031	3.3 ± 0.1	3.4 ± 0.1	3.4 ± 0.1
US07-08	63° 29.892′	144° 31.581′	1710	1.2	3	0.989	0.3719[a]	0.463 ± 0.028	2.2 ± 0.1	2.2 ± 0.1	2.2 ± 0.1

[a] Denotes use of a low-level Be carrier (405 $\mu g\, g^{-1}$). All other samples use SPEX brand Be carrier (1000 $\mu g\, g^{-1}$).
All samples are standardized to 07KNSTD3110.
$\varepsilon_{min} = 1.0\, mm\, ka^{-1}$; $\varepsilon_{max} = 4.8\, mm\, ka^{-1}$.

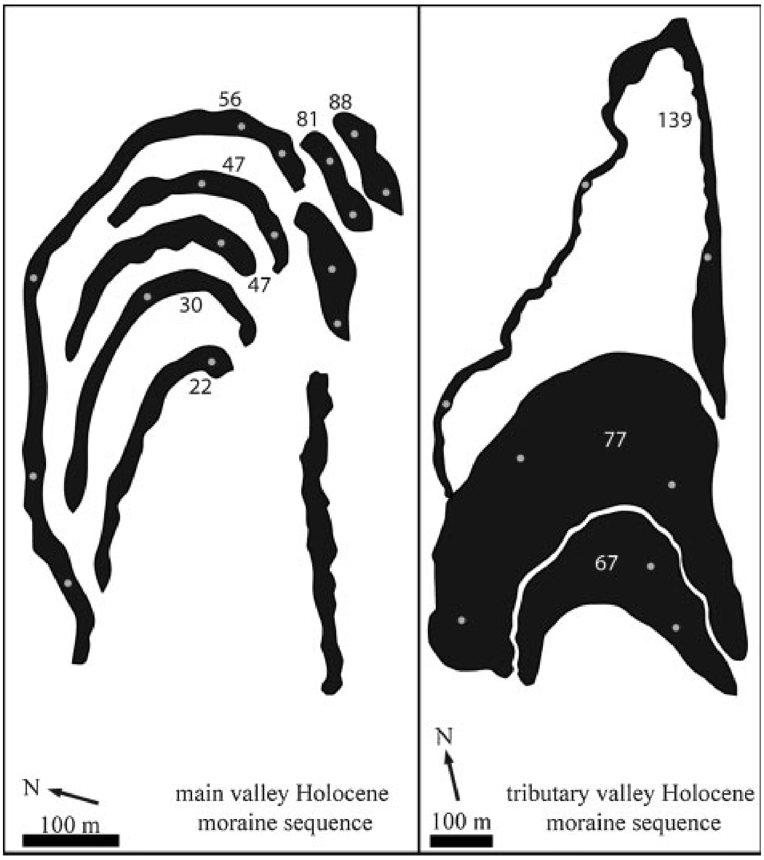

Figure 5 Diagram of moraine crests in the main and tributary valleys (location shown in Fig. 4). Dots mark lichen measurement stations and numbers represent the maximum lichen diameter (mm) measured on each moraine crest. Note the change in scale and direction of the north arrow

erosion rates to illustrate the range of potential uncertainty with absolute ^{10}Be ages; however, the following discussion is based on ε_0 ages only. Finally, as Fish Lake valley is relatively arid and samples were collected from windswept locations, error attributed to snow shielding should be minimal and hence no correction was made.

The oldest ^{10}Be ages come from a left-lateral moraine within terminal late Wisconsin drift (Fig. 3). Boulders in the late Wisconsin moraine range from 22.4 ± 0.6 to 9.3 ± 0.4 ka ($n = 7$; Table 1), with a mean of 16.5 ± 0.1 ka ($n = 4$) after excluding two boulders that were most likely exhumed (FL06-02; FL06-12) as indicated by their much younger exposure ages (9.3 ± 0.4 and 9.5 ± 0.3 ka). Additionally, an outlier (FL06-01) yielding an age of 22.4 ± 0.6 ka is not included in our calculations for reasons discussed below.

A second group of boulders ($n = 3$) resting on ground moraine ~2 km from the cirque headwall, ~200 m above the main valley floor and adjacent to tributary valley moraines (Fig. 4), has ^{10}Be ages of 15.4 ± 0.6, 13.9 ± 0.4, and 13.2 ± 0.3 ka, and a mean age of 14.2 ± 1.2 ka. The two younger erratics (US07-10; US07-11) display some evidence of surface spallation, suggesting the depositional age of these erratics may be closer to that of the oldest boulder (US07-09, 15.4 ± 0.6 ka).

A left-lateral moraine extending ~2.5 km downvalley of late Holocene ice limits (Fig. 3) provides evidence for a latest Pleistocene readvance or standstill in Fish Lake valley. No boulders suitable for ^{10}Be dating were found on its crest, but five boulders resting on ground moraine downvalley of late Holocene moraines, and within the ice limit, were sampled in order to date the retreat from this margin (Fig. 4). Ages from this unit in the main valley are 11.8 ± 0.7, 11.3 ± 0.3, 8.5 ± 0.4, 6.6 ± 0.2 and 5.3 ± 0.2 ka. A moraine in the tributary valley appears to be coeval with the latest Pleistocene moraine in the main valley based on extent of tundra and lichen cover, morphostratigraphic position and weathering features; namely, the boulders lacked striations but exhibited some grain-to-grain surface relief. Two boulders sampled from this end moraine (US07-04; US07-05) have ages of 11.6 ± 0.8 and 7.0 ± 0.2 ka (Fig. 4). Combined, these seven boulders range in age from 11.8 ± 0.7 to 5.3 ± 0.2 ka (Table 1).

In the main valley, one boulder was sampled from the outermost Holocene moraine and has a ^{10}Be age of 1.2 ± 0.1 ka (Fig. 4; Table 1). Boulders from two Holocene moraines in the tributary valley were also sampled. Three boulders on the older of the two moraines yield ^{10}Be ages of 2.2 ± 0.1, 2.5 ± 0.1 and 3.3 ± 0.1 ka (Fig. 4). Only two boulders were sampled from the younger moraine, and are 2.5 ± 0.1 and 1.1 ± 0.1 ka (Fig. 4).

Lichenometry

Maximum lichen diameters on the late Holocene moraines in the main valley are 88 ($n = 104$), 81 ($n = 133$), 56 ($n = 88$), 47 ($n = 137$), 47 ($n = 86$), 30 ($n = 96$) and 22 ($n = 86$) mm on progressively younger moraines crests (Fig. 5; Table 2). In the tributary valley, maximum lichen diameters on three progressively younger moraines are 139 ($n = 84$), 77 ($n = 65$) and 67 mm ($n = 120$). Applying maximum lichen diameters to the growth curve (Begét, 1994; Solomina and Calkin, 2003) indicates that moraines in the main valley stabilised at ca. AD 610, 840, 1640, 1860, 1910 and 1930. Moraines in the tributary valley stabilised at ca. 1020 BC (ca. 3 ka), and AD 970 and 1290. Ages of stabilisation were rounded to the nearest decade and are summarised in Table 2 and Fig. 6. The uncertainty associated with lichenometric ages is typically cited at about ±20% (e.g. Calkin and Ellis, 1980).

Table 2 Lichenometry in Fish Lake valley

Location	Largest lichen (mm)	Mean of largest 5 (mm)	Standard deviation (5 largest; mm)	Total thalli measured (n)	Years of exposure (largest lichen)	Year of stabilization (AD)
Main valley	88	78	7	104	1393	610
	81	75	7	133	1168	840
	57	54	2	88	366	1640
	47	44	2	137	148	1860
	47	42	3	86	148	1860
	30	29	1	96	94	1910
	22	18	2	86	76	1930
Tributary valley	139	106	23	84	3028	BCE 1020
	77	73	3	65	1040	970
	67	61	4	120	719	1290

Lichenometric ages (years of exposure from 2007) were calculated using the single largest lichen applied to the growth curve from Solomina and Calkin (2003). For lichen diameters $\leq$50 mm a logarithmic curve is used (great growth portion of the calibration curve) where age $= 41.861 \times 10^{(0.011679 \times D)}$. A linear equation representing the long-term growth section of the calibration curve was used for lichen diameters >50 mm, where age $= -1430 + 32.074 \times D$. $D =$ lichen diameter (mm).

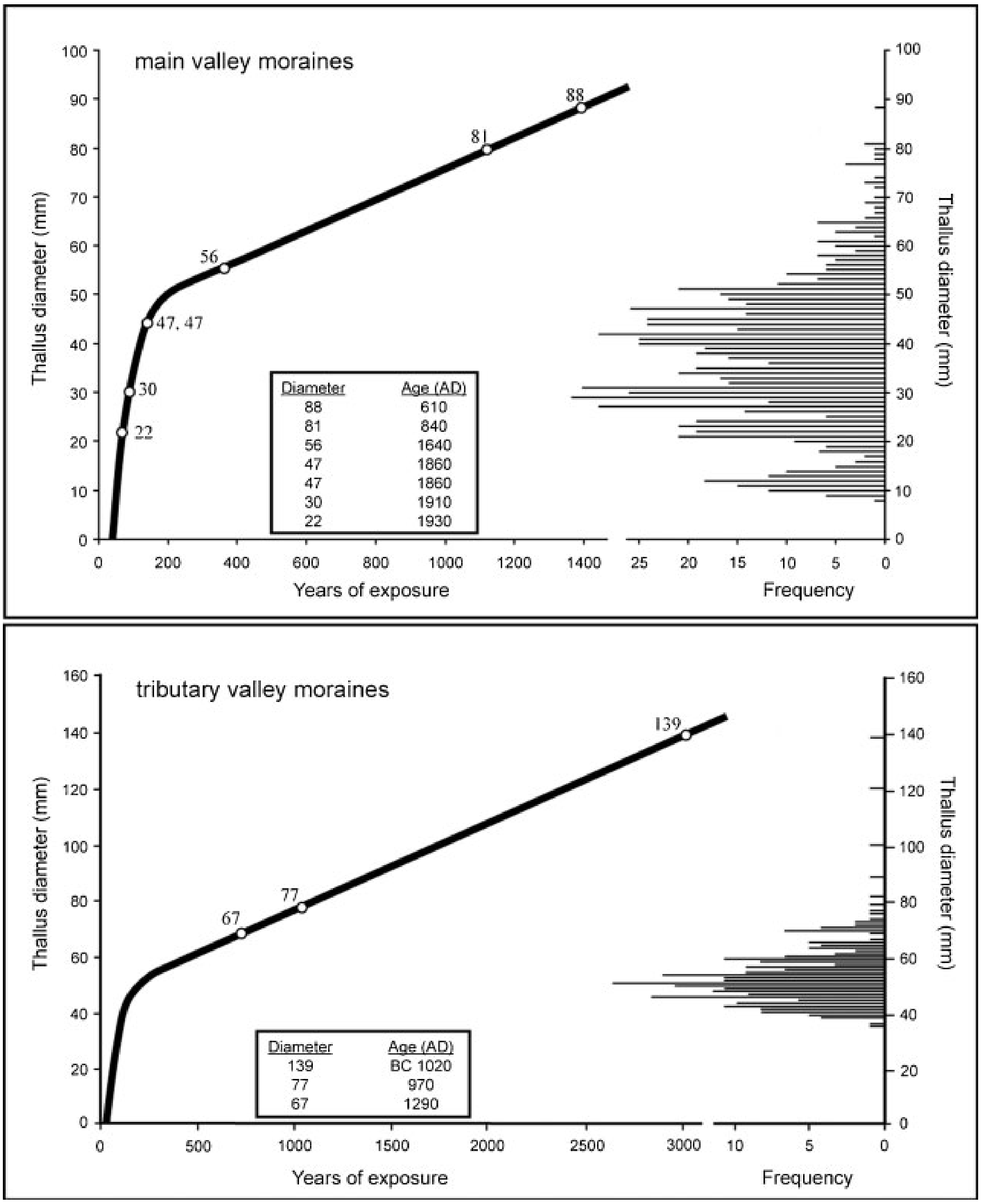

Figure 6 Lichenometry results in Fish Lake valley. All measurements are shown on the histograms and the maximum diameter (Table 2) is used to calculate moraine stabilisation age after excluding thalli >20% larger than the next largest thallus on each moraine. Growth curve is from Solomina and Calkin (2003)

Discussion

We follow the interpretations made elsewhere in Alaska suggesting that the oldest ^{10}Be ages from boulders resting on laterofrontal moraines most closely date the timing of deglaciation and subsequent moraine stabilisation (Briner *et al.*, 2005). However, within a set of ^{10}Be ages multiple inferences can be made, less so when independent geochronological control such as lichenometry is available, as is the case for late Holocene moraines in the Fish Lake valley. Furthermore, ages based on lichenometry should be treated cautiously and considered minima because ice-cored moraines are generally unstable during ice melt-out or as the moraine is mobilised into a rock glacier.

Late Pleistocene

Boulders on the late Wisconsin lateral moraine in Fish Lake valley have a mean age of 16.5 ± 0.1 ka after excluding the two young outliers (9.3 ± 0.4 and 9.5 ± 0.3) that were most likely exhumed post deposition. Boulder FL06-01, yielding an age of 22.4 ± 0.6 ka, was excluded from our calculations and we suggest three possibilities for this outlier: (1) it contains isotopic inheritance, which yields an age older than its deposition, which occurred ca. 16.5 ka; (2) 22.4 ± 0.6 ka is the age of deglaciation and boulders in the 16.5 ka cluster were later exhumed; or (3) the ^{10}Be age is an accurate representation of its exposure history and can be attributed to a separate, more extensive ice margin. We favour the third scenario because this boulder rests approximately 3 km downvalley of the boulders used in our age calculation (Fig. 3) and it is unlikely that ca. 6 ka passed between deglaciation and surface stabilisation, assuming deglaciation occurred at ca. 22 ka. An age of ca. 22 ka is also consistent with another glacial chronology in Alaska (e.g. Briner *et al.*, 2005) that indicates glaciers began retreating from their late Wisconsin maximum positions ca. 23–21 ka. However, we cannot rule out other scenarios.

Three erratics with a mean of 14.2 ± 1.2 ka (oldest boulder $= 15.4 \pm 0.6$ ka) mark the timing of subsequent deglaciation from the late Wisconsin maximum ice extent. Ice from both the main and tributary valley cirques likely coalesced during full-glacial conditions, and then became disconnected as the glacier retreated towards its headwalls (Fig. 3).

Three boulders resting on ground moraine and one from an end moraine were used to date the retreat from the next oldest ice limit in Fish Lake valley. We excluded four boulders from our calculation for several reasons. The wide spread of ages for these boulders (5.3 ± 0.2 to 8.6 ± 0.4 ka; Table 1; Fig. 7)

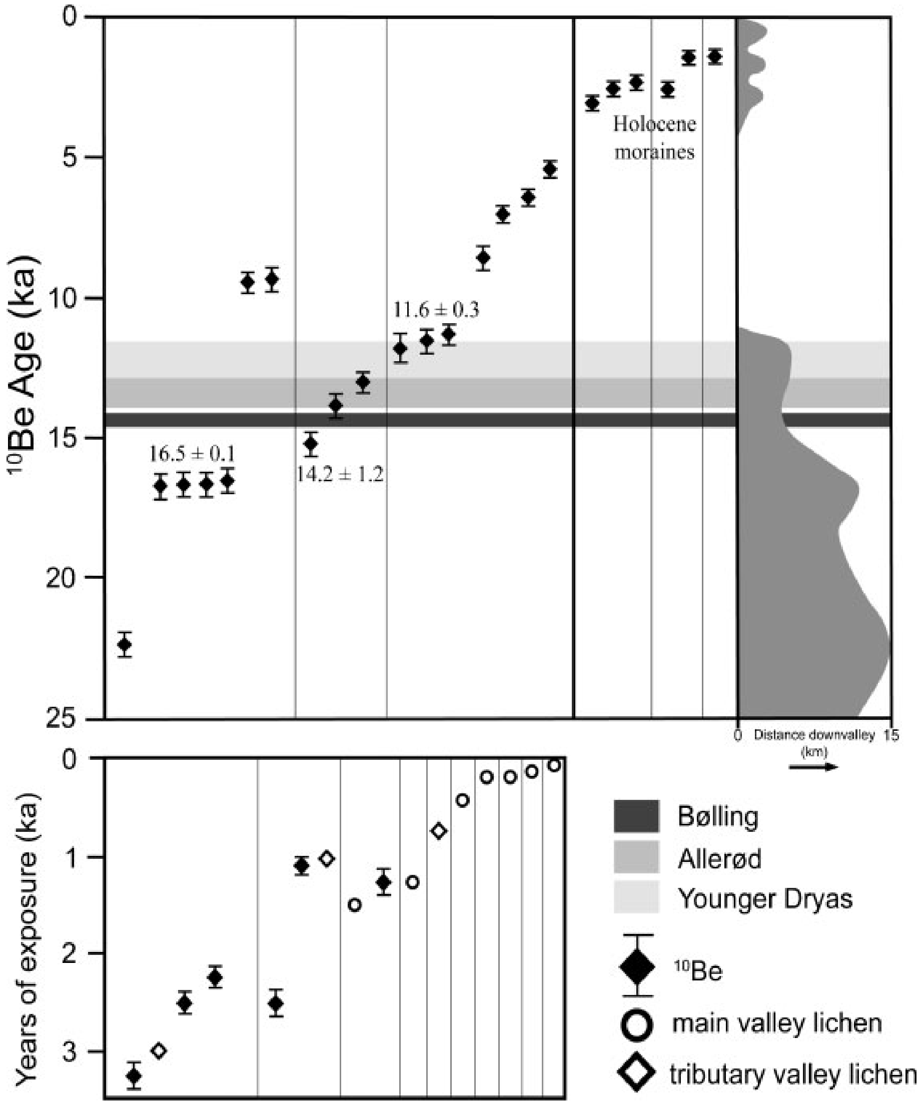

Figure 7 Summary diagram of ^{10}Be and lichen ages. Vertical lines demarcate mapped units in the study valley. Error bars on ^{10}Be ages are the 1σ AMS uncertainty (Table 1). Upper box shows all ^{10}Be ages and a time–distance diagram of glacier extent downvalley relative to the 2007 margin. Lower box shows only Holocene ^{10}Be ages combined with lichenometric ages from the same moraine (note change in axis). ^{10}Be ages and lichenometric ages in both boxes have been ordered by decreasing age within each map unit on the *x*-axis

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indicates that exhumation is responsible for the skewed age distribution. Isotopic inheritance, which would give an age older than the deposition, would not appear as a tight cluster of ages centred at 11.6 ± 0.3 ka ($n = 3$) calculated from the remaining boulders. Additionally, the younger ages are suspect because there is no evidence to suggest that glaciers were advancing or depositing moraines well beyond their LIA extent during the early Holocene, a period of time thought to be warmer than today (Kaufman *et al.*, 2004). Thus, we hypothesise that a readvance or standstill during overall retreat in Fish Lake valley culminated ca. 11.6 ka $\pm$ 0.3 ka.

Late Holocene

The late Holocene moraines preserved in the main valley suggest at least five periods of glacier advance and stabilisation. The outermost moraine crest has a ^{10}Be age of 1.2 ± 0.1 ka (ca. AD 700–900) and lichenometric age of ca. AD 610, while the next inner moraine has a lichenometric age of 840 AD. Taking into account the $\pm$20% uncertainty of lichenometric ages, the methods produce ages that overlap. The two outermost moraines were most likely deposited sometime just prior to or during AD 610–840, which is in close agreement with the single ^{10}Be age. Lichenometric ages of ca. AD 1640, 1860 (on two moraines), 1910 and 1930 indicate the stabilisation ages for successive ice margins.

Moraines in the tributary valley reveal a different history of glaciation during the late Holocene. Used independently, both the ^{10}Be and lichen ages suggest the oldest moraine was deposited sometime between ca. 3.3 ± 0.1 and 2.2 ± 0.1 ka (Fig. 4), but this age assignment can be refined by using the two methods in conjunction. Briner *et al.* (2005) suggested that the oldest age within a group of ^{10}Be ages most closely constrains the timing of moraine abandonment. Using this approach, 3.3 ± 0.1 ka is the most likely time of moraine abandonment, which agrees with the lichen age of ca. 3 ka (Fig. 4 and Table 2). However, because of an unusually large range in lichenometric sizes across the moraine crest ($1\sigma = 23$ mm; Table 2), our confidence in the lichenometric age is limited. Nonetheless, we assign moraine abandonment age of ca. 3.3–3.0 ka for the earliest Holocene advance in the study area.

The next oldest moraine in the tributary valley is constrained by ^{10}Be ages of 2.5 ± 0.1 and 1.1 ± 0.1 ka, and a lichenometric age of AD 970. We hypothesise that inheritance accounts for the anonymously old ^{10}Be age (2.5 ± 0.1 ka; Fig. 4), and that the agreement between the younger ^{10}Be age (1.1 ± 0.1 ka) and the lichenometric age (Table 2 and Figs 4 and 7) indicates that this moraine stabilised ca. 1.1 ka (ca. AD 800–1000). Alternatively, the oldest lichen was not found and this moraine stabilised before ca. 1.0 ka; however, 65 lichen thalli were measured and older thalli were found elsewhere. Thus we favour an age of ca. 1.1 ka for this moraine. The youngest moraine in the tributary valley is dated only by lichenometry to ca. AD 1290.

Ice-cored moraines in both valleys display evidence of limited downslope migration characteristic of rock glaciers or rock-glacerised moraines. Some moraine sections show evidence for a mobilised ice core, but distinct primary crests can be identified along most moraines in both the main and tributary valleys, indicating relative surface stability. Downslope migration of ice-cored moraines may explain the different glacial records between the two valleys. For example, moraines marking past ice extent would be preserved as they moved downslope and beyond the extent of the next younger ice limit. Younger glacier advances would simply butt up against, but not overrun, pre-existing moraines during brief readvances. Moraines in the main valley rest on a relatively gentle slope compared to the steeper tributary valley, where downslope migration could be more pronounced. This process may explain the preservation of the oldest Holocene moraine in the tributary valley and not the main valley.

Late Pleistocene glacial history of the north-eastern Alaska Range

Here we combine ^{10}Be ages from Late Pleistocene and late Holocene surfaces with lichenometry to reconstruct a history of glaciation in the Fish Lake valley since the LGM. These chronological inferences are in good agreement with other records elsewhere in Alaska.

Limited evidence in the Fish Lake valley points to initial ice retreat from its LGM maximum position ca. 22.4 ± 0.6 ka. Although this is based on just one ^{10}Be age, it is consistent with other regional cosmogenic exposure ages on LGM moraines in central Alaska. Briner *et al.* (2005) showed that ice retreated from its late Wisconsin maximum extent at ca. 23–21 ka in the Yukon–Tanana Upland. In the western Alaska Range, ice began retreating from maximum positions between ca. 21 and 20 ka in both the Swift River valley and in the Denali National Park region (Ten Brink and Waythomas, 1985; Briner *et al.*, 2005; Briner and Kaufman, 2008). The local Donnelly glaciation is only constrained by two radiocarbon ages. A minimum age of $23\,610 \pm 1110$ cal. a BP (CALIB v. 5.0.2; http://calib.qub.ac.uk/calib/; Stuiver *et al.*, 1998) is from organic matter at the top of outwash gravel near Delta Junction and a maximum age of $30\,080 \pm 1058$ cal. a BP (CalPal; http://www.calpal.de/) is from wood fragments beneath till in Gerstle Valley (Fig. 1), located $\sim$40 km west of Fish Lake valley (Hamilton, 1982; Porter *et al.*, 1983). These bracketing ages are consistent with our ^{10}Be age, which is interpreted as a minimum age for deglaciation from the LGM limit.

A tight cluster of ^{10}Be ages suggests that ice remained near its LGM limit as late as 16.5 ka. Whether this margin can be attributed to a readvance or standstill during overall retreat is unclear, but several moraines in the Alaska Range indicate a similar ice margin history. In the east-central Alaska Range, Matmon *et al.* (2006) dated end moraines attributed to a readvance at ca. 17–16 ka, and erratics near the Nenana River provided an age of ca. 17 ka (Dortch, 2006). Deglaciation of the Nenana River valley at ca. 15.5 ka was also inferred from erratics on high massifs (Dortch, 2006; Briner and Kaufman, 2008). Furthermore, these data suggest that the final LGM termination in the Fish Lake valley and central Alaska may have been broadly synchronous with LGM termination at mid latitudes occurring about 17.1–17.4 ka (e.g. Schaefer *et al.*, 2006).

Deglaciation of the lower Fish Lake valley by ca. 15–14 ka was followed by a readvance or standstill culminating at 11.6 ± 0.3 ka. The timing of ice retreat following this readvance or standstill may be coincident with the termination of the North Atlantic Younger Dryas (YD) event (12.9–11.6 ka; Alley, 2000; Fig. 7). A recent compilation of several lake and peat records throughout Beringia reveals a spatially complex climatic pattern during the YD chronozone (Kokorowski *et al.*, 2008). The strongest evidence for cooling exists in southern Alaska, and is most likely associated with lowered sea-surface temperatures in the North Pacific. Evidence for cooling is limited in central Alaska, but a few proxy records suggest a climate shift during the YD in the region (Kokorowski *et al.*, 2008). A lake-level reconstruction from nearby Birch Lake indicates that a dry period occurred during the first half of the YD, followed by an increase in moisture (Abbott *et al.*, 2000). This suggests that, at least during the latter half of the YD, increased moisture availability could have contributed to a glacier readvance or standstill. Moraines in two other places in Alaska provide additional evidence for a glacial advance during the YD. In the Ahklun Mountains a readvance terminated sometime between 11.7 and 11.0 ka based on ^{10}Be and ^{26}Al ages (Briner *et al.*, 2002), and moraines dated by ^{10}Be in the central Alaska Range were likely formed between 11.7 and 11.0 ka (Matmon *et al.*, 2006).

Late Holocene glacial history of the central Alaska Range

The onset of Neoglaciation occurred at several sites across Alaska between ca. 3.6 and 3.0 ka (e.g. Calkin, 1988; Calkin *et al.*, 2001; Levy *et al.*, 2004; McKay and Kaufman, 2009), consistent with our findings in Fish Lake valley. Additionally, in the central Alaska Range, advances of the Black Rapids, Canwell, and Gulkana glaciers were documented between ca. 3.6 and 3.4 ka (Calkin, 1988; Begét, 1994). Moraines in the main Fish Lake valley stabilising between AD 610 and 840 may be correlative with widespread advance of southern coastal Alaskan glaciers centred between AD 400 and 700 (Reyes *et al.*, 2006; Wiles *et al.*, 2008). Moreover, a geochemical proxy temperature reconstruction at Farewell Lake located in the western Alaska Range suggests that a pronounced cooling occurred around AD 600 (Hu *et al.*, 2001). A new proxy temperature record from Hallet Lake in the Chugach Range also shows pronounced cooling around AD 600 (McKay *et al.*, 2008). A moraine in the tributary valley is centred at ca. AD 900, taking into account both ^{10}Be and lichenometric results, and is likely coeval with the outermost Holocene moraines in the main valley. Although this is not a period of widespread glacier advance in Alaska, moraines dating to around AD 900–950 were found in the Brooks Range (Evison *et al.*, 1996; Calkin, 1988), and Denton and Karlén (1977) suggested a period of ice advance in the St Elias Mountains between ca. AD 720 and 900. Alternatively, we cannot rule out the possibility that this moraine was deposited before 900 AD and is coeval with the two oldest Holocene moraines in the main valley (ca. AD 610–840), considering the age uncertainties associated with both ^{10}Be and lichenometric dating.

The innermost moraine in the tributary valley was dated by lichenometry to AD 1290. It most likely stabilised following a period of LIA cooling centred on AD 1250 that has been reported from several regions of Alaska (e.g. Wiles *et al.*, 2004). Lichenometric moraine ages of ca. AD 1640 and 1860 in the main valley are both in agreement with well-established periods of glacial expansion during the LIA in the Wrangell Mountains. Based on overrun logs in glacial forefields and lichenometrically dated moraines, glaciers in the Wrangell Mountains were advancing during the mid to late 1600s and into the 1800s (Wiles *et al.*, 2002). Additionally, moraine stabilisation at ca. AD 1910 is consistent with early 20th-century moraine-building events also recorded in the Wrangell Mountains. The youngest inner moraine in the main valley (ca. AD1930) marks the ice position before the historical period of glacier retreat.

Equilibrium-line altitudes

We estimated palaeoglacier equilibrium-line altitudes (ELA) for Late Pleistocene and Holocene ice limits using well-defined end and lateral moraines in the main and tributary valleys. Ice margins were reconstructed from moraines mapped on 1980 aerial photographs, then field checked. ELA estimates were based on the accumulation area ratio (AAR) method. AAR values have been measured annually from 1966 to 2006 at the Gulkana Glacier, located ~50 km southwest of Fish Lake valley (Fig. 1), and we use the average AAR of 0.61 over this time period (United States Geological Survey, 2007, http://ak.water.usgs.gov/glaciology/gulkana/balance/index.html). However, the AAR method assumes a steady-state glacier mass balance and the Gulkana Glacier has experienced an overall negative mass balance over the observed interval. Therefore, we consider a 0.61 AAR and corresponding ΔELAs to be minimum values. Using the 2007 glacier margins, the modern ELAs in the main and tributary valleys are 2190 and 2040 m a.s.l. The aerial photos show that glacier termini have been relatively stable since 1980. Using the most distal Holocene moraine, we calculated only one Holocene ELA in each valley because the moraines are closely nested. The ELA during the LGM, the latest Pleistocene limit, and the Holocene was approximately 975, 420, and 240 m lower than the modern ELA in the main valley, respectively. In the tributary valley, the latest Pleistocene and Holocene advances were ~350 and 270 m lower than modern. These ELA reconstructions in Fish Lake valley suggest that, during the latest Pleistocene and late Holocene advances, ELAs were ~42% and 25% of full LGM conditions. An ELA depression of ~975 m in Fish Lake valley is almost double typical ELA depressions (~500 m) in Alaska during the LGM (e.g. Manley *et al.*, 2003; Balascio *et al.*, 2005b). It is possible that significant debris cover on the Fish Lake valley glacier contributed to a large ELA depression during the LGM. However, our ELA depressions are from a single valley, and previous studies noted significant valley-to-valley variation in ELA depressions and 500 m typically represents an average value over a broad region.

Summary and conclusions

Initial LGM moraine abandonment in Fish Lake valley occurred at ca. 22.4 ka, and ice remained near its terminal position until ca. 16.5 ka. These results, combined with several other central-Alaskan records, imply that termination of the LGM in central Alaska is generally synchronous with terminations at mid latitudes (e.g. Schaefer *et al.*, 2006). The new geochronological results from Fish Lake valley add to the evidence that glaciers in Alaska reached their maximum extent around the time of the global LGM. The global circulation model output implies that glaciers should have been restricted during the LGM (e.g. Otto-Bliesner *et al.*, 2006; CCSM3), yet ELA lowering in Fish Lake valley was similar to elsewhere around the globe (e.g. Broecker and Denton, 1990), which seems to contradict the GCM simulation. On the other hand, ELA lowering in Fish Lake valley stands out relative to elsewhere in Alaska. More work is needed to determine how ELA estimates from Fish Lake valley relate to a general W–E pattern of ELA lowering in Alaska associated with south-westerly moisture transport (e.g. Kutzbach *et al.*, 1998; Manley *et al.*, 2003; Balascio *et al.*, 2005b).

Deglaciation of the Fish Lake valley ca. 15–14 ka was followed by a Late Pleistocene readvance or standstill terminating by 11.6 ka, coeval with the end of the North Atlantic YD event. This adds to the limited evidence in Alaska suggesting that valley glaciers advanced in response to YD cooling. More tightly constrained and widespread chronologies are needed to verify advances of Alaskan valley glaciers during the YD. Neoglaciation was most likely initiated in the Fish Lake valley 3.3–3.0 ka, and at least two glacial advances occurred between AD 610 and 900. The moraine dated to AD 610 is consistent with temperature reconstructions from the western Alaska Range (e.g. Hu *et al.*, 2001) and the Chugach Mountains (e.g. McKay *et al.*, 2008), suggesting a pronounced cooling centred at AD 600. LIA moraines dating to ca. AD 1290, 1640, 1860 and 1910 are consistent with the ages of LIA moraines elsewhere in Alaska. Moreover, the preservation of pre-LIA moraines directly in front of LIA moraines shows that the extent of ice advance during both of these time periods was similar, contrasting the notion that LIA advances typically represent maximum Holocene positions as has been shown for other places in Alaska.

The relative agreement between lichenometric ages and ^{10}Be ages supports the reliability of the *Rhizocarpon geographicum* growth curve for the Alaska Range. The results imply that using these two dating methods in conjunction provides more accurate ages of moraines than if the methods are used independently. This may be particularly true for deposits dating to older portions of the lichen growth curve, which are often poorly constrained. Finally, this study illustrates the future potential for using ^{10}Be to date Holocene glacier fluctuations in high-elevation areas of Alaska where organics needed for radiocarbon dating are unavailable.

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