Rapid and early deglaciation in the central Brooks Range, Arctic Alaska

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ABSTRACT

Alpine-style glaciation was rare in the Arctic during the last glaciation because ice sheets occupied most of the glaciated high latitudes. Due to the tight coupling of alpine-glacier fluctuations with climate, the geomorphic evidence of such fluctuations in the Brooks Range, Alaska (USA), presents a unique opportunity to study past climate changes in this portion of the Arctic. We use cosmogenic ^{10}Be exposure dating to directly date Last Glacial Maximum (LGM) terminal moraines and deglaciation in the central Brooks Range. ^{10}Be ages from moraine boulders indicate that the LGM culminated at ca. 21 ka and was followed by substantial retreat upvalley prior to a second moraine-building episode culminating at ca. 17 ka. Subsequent rapid deglaciation occurred between ca. 16 ka and 15 ka, when glaciers receded to within their Neoglacial limits. Initial deglaciation after the LGM was likely caused by ice sheet–induced atmospheric circulation changes and increasing insolation. Brooks Range glaciers largely disappeared during Heinrich Stadial 1, prior to significant warming in the North Atlantic region during the Bølling-Allerød, but coincident with global CO_2 rise. Glacier fluctuations during the late-glacial period, if any, were restricted to within their Neoglacial extents. This new chronology suggests that ice sheet–modulated atmospheric circulation and global CO_2 dominate glacial climate forcings in Arctic Alaska.

INTRODUCTION

The Brooks Range, Arctic Alaska (USA), contains a rich record of glaciation dating back to the Pliocene (Hamilton, 1982, 1986; Briner and Kaufman, 2008). Unlike most of the North American Arctic, which was covered by ice sheets during the Last Glacial Maximum (LGM; ca. 26.5–19 ka), the Brooks Range supported a network of alpine glaciers. Tight coupling between alpine glaciers and climate makes glacial landforms in the Brooks Range important paleoclimate indicators for the as yet poorly constrained late Quaternary climate history in the North American Arctic.

Northern Hemisphere deglaciation following the LGM was driven by increased summer insolation at northern high latitudes and increased global atmospheric CO_2 concentration (Clark et al., 2009; Shakun et al., 2012). However, variability in the timing of deglaciation across the Northern Hemisphere due to regional atmospheric and oceanic forcings is well documented (Clark et al., 2012). Absent from these studies are alpine glacier chronologies from the Arctic. Notwithstanding foundational glacial-geologic mapping and radiocarbon ages on outwash sequences (Hamilton, 1986), Pleistocene glacial deposits in the central Brooks Range have yet to be sufficiently dated, leaving the timing and underlying climate forcings of the LGM and subsequent deglaciation unclear.

We use cosmogenic ^{10}Be to date LGM terminal moraines, post-LGM end moraines, and deglacial features in several valleys of the central Brooks Range—features that are difficult to date with radiocarbon. A previous study that applied ^{10}Be dating to late Pleistocene moraines in the northeastern Brooks Range had mixed results because the moraines were heavily affected by permafrost processes and had few large boulders (Balascio et al., 2005a). Badding et al. (2013) provided the first ^{10}Be ages of LGM deglaciation, but their chronology focused primarily on Holocene features. Our study provides new ages on the local LGM and subsequent glacier fluctuations during deglaciation, producing the most complete glacial history following LGM termination in this region of the Arctic. This new chronology also highlights possible climate-forcing mechanisms, including changes in atmospheric circulation and CO_2 concentrations.

BACKGROUND

The Brooks Range stretches ~1000 km from the Chukchi Sea in the west to the Alaska-Yukon border in the east (Fig. 1). Radiocarbon ages from outwash sediments in the central Brooks Range provide a framework chronology of late Pleistocene glaciation that has remained largely unchanged for over 30 yr (Detterman et al., 1958; Hamilton, 1982). During the late Wisconsin (30–10 ka; locally termed the Itkillik II glaciation), glaciers reached beyond the Brooks Range front in most valleys during two phases, which were dated by Hamilton (1986) to ca. 29–27 cal. (calibrated) kyr B.P. and 24–20 cal. kyr B.P. Glaciers were relatively restricted, with equilibrium-line altitudes lowering by only 300–600 m, much less than elsewhere in the Americas (Balascio et al., 2005b).

Prominent moraines upvalley of Itkillik II main-phase moraines were mapped as the late Itkillik II re-advance across the Brooks Range (including the Ernie Pass area; Fig. 1) and delimit a pause or re-advance during deglaciation (Hamilton, 1986, 2003; Hamilton and Labay, 2011). Using bracketing radiocarbon ages of 15.2 +0.40/–0.24 cal. kyr B.P. and 13.2 +0.17/–0.14 cal. kyr B.P. from alluvium onlapping a late Itkillik II re-advance moraine (Fig. 1) in the Atigun Gorge region, Hamilton (2003) argued that active ice must have been present in order to block the river and cause deposition. However, ^{10}Be ages on deglacial bedrock surfaces 30–40 km upvalley of the moraine are 15.6 ± 0.3 ka and 14.7 ± 0.3 ka (Badding et al., 2013), suggesting that rapid deglaciation was occurring at the same time as the proposed late Itkillik II re-advance. These conflicting chronologies leave the age of late Itkillik II re-advance moraines uncertain, and the deglacial climate history unclear.

METHODS

Four valleys in the central Brooks Range with prominent Itkillik II landforms and/or exposed deglacial surfaces of quartz-rich metasedimentary or granitic lithologies yielded 25 ^{10}Be ages from moraine boulders (n = 14), erratic boulders perched on bedrock (n = 10), and ice-sculpted bedrock (n = 1; Figs. DR1–DR4 in the GSA Data Repository[1]). Moraine boulders in stable positions were sampled from Itkillik II lateral deposits just adjacent to the Alapah River valley and from a late Itkillik II re-advance moraine in the Ernie Pass area (Fig. 1; Hamilton and Labay,

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[1]GSA Data Repository item 2015147, Figures DR1–DR4 (sample photographs, sampling protocol, preparation, and production rate selection), Table DR1 (^{10}Be CRN data), and Table DR2 (statistical treatment of moraine boulders), is available online at www.geosociety.org/pubs/ft2015.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

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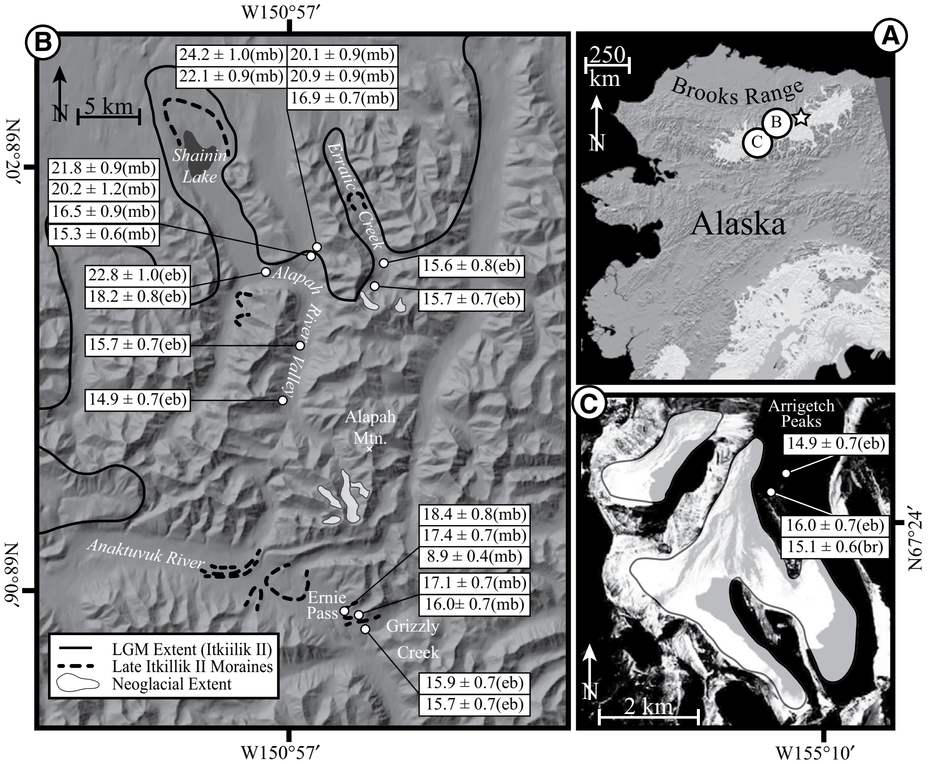

Figure 1. A: Alaska (USA) showing ice extent during Last Glacial Maximum (LGM, white shading; Kaufman et al., 2011) and showing locations of B, C, and Atigun Gorge (star). B: Shaded relief map of north-central Alapah Mountain area showing maximum extent of glaciers during Itkillik II glaciation, late Itkillik II moraines, and Neoglacial ice extent. C: South-central Arrigetch Peaks area with shaded Neoglacial extent (aerial photo base). ^{10}Be ages (in ka) are from moraine boulders (mb), erratic boulders (eb), and bedrock (br).

2011). Erratic boulders upvalley of Itkillik II terminal moraines perched on bedrock above the valley floor were selected to minimize post-depositional modification and snow shielding. The bedrock surface was selected based on the presence of glacial scour and no evidence of erosion or sediment cover.

Samples were processed at the University at Buffalo (New York, USA) Cosmogenic Isotope Laboratory following standard procedures (see the Data Repository for details on sample preparation). ^{10}Be ages were calculated using the CRONUS-Earth exposure-age calculator (Balco et al., 2008) assuming no snow shielding and no erosion, and using the Arctic ^{10}Be production rate (Young et al., 2013) with the constant-production scaling scheme of Lal (1991) and Stone (2000) (see the Data Repository). The lack of snow or erosion corrections means the calculated ages may be slightly too young when compared to other dating methods. Ages are reported with 1σ external uncertainties.

^{10}Be CHRONOLOGY

Nine boulders sampled from Itkillik II terminal deposits in the Alapah River valley yield ^{10}Be ages that range from 24.2 ± 1.0 ka to 15.3 ± 0.6 ka (Fig. 1B; Table DR1 in the Data Repository). Outliers (the three youngest and one oldest ages) were identified by mean square of weighted deviates and χ^2 statistics (Table DR2). We interpret the boulders with the three youngest ages (16.9 ± 0.7, 16.5 ± 0.9, and 15.3 ± 0.6 ka) as having been exposed sometime after emplacement due to moraine degradation, which is common in Alaskan moraines (Briner et al., 2005; Badding et al., 2013). We suspect that the boulder with the aberrantly old age was influenced by prior exposure (cf. Heyman et al., 2011). Given the spread of ages, the average of the five remaining ^{10}Be ages (21.0 ± 0.8 ka) is most appropriate for the culmination of the main Itkillik II phase.

Five boulders from the late Itkillik II re-advance moraine in the Ernie Pass area yield ^{10}Be ages of 18.4 ± 0.8 ka to 8.9 ± 0.4 ka (Fig. 1B; Table DR1). The youngest age is an outlier (Table DR2), likely a product of boulder exhumation. The average age of the remaining four boulders is 17.2 ± 1.0 ka. Two erratic boulders located ~0.3 km upvalley of the late Itkillik II moraine in the Ernie Pass area yield ^{10}Be ages of 15.9 ± 0.7 ka and 15.7 ± 0.7 ka, indicating that glacier ice had retreated from the moraine by this time.

In the Alapah River valley, two erratic boulders located ~14.5 km upvalley from the limit of Itkillik II ice yield ^{10}Be ages of 22.8 ± 1.0 ka and 18.2 ± 0.8 ka. These boulders either are affected by inheritance or represent an earlier period of deglaciation between the multiple late Itkillik II moraine-building phases mapped in the area (Fig. 1). Two additional erratic boulders located 21.0 km and 24.5 km upvalley of the Itkillik II terminal moraine yield ^{10}Be ages of 15.7 ± 0.7 ka and 14.9 ± 0.7 ka, respectively (Fig. 1B). In the Erratic Creek valley, erratic boulders located ~3.5 and 2.0 km downvalley from the Neoglacial limit yield ^{10}Be ages of 15.7 ± 0.7 ka and 15.6 ± 0.8 ka, respectively (Fig. 1B). In the Arrigetch Peaks, samples from bedrock and an erratic boulder <100 m beyond the maximum Neoglacial moraine yield ^{10}Be ages of 15.1 ± 0.6 ka and 16.0 ± 0.7 ka, respectively. A second erratic boulder 0.5 km downvalley of the Neoglacial limit yields a ^{10}Be age of 14.9 ± 0.7 ka (Fig. 1C; Table DR1).

We compiled all of the accepted ^{10}Be ages onto a single composite time-distance diagram by normalizing the elevation range between Itkillik II moraines and the cirque floor in each valley (Fig. 2). Taken together, the ^{10}Be ages from the central Brooks Range provide a chronology from Itkillik II terminal moraines to deglaciation into the cirques. The ages show that the terminal LGM moraine (Itkillik II) formed by 21 ± 0.8 ka. Glaciers then retreated upvalley before re-advancing or stabilizing to build late Itkillik II re-advance moraines at 17.2 ± 1.0 ka. Erratic boulders immediately upvalley of late Itkillik II re-advance moraines at Ernie Pass indicate that deglaciation was underway again by ca. 15.8 ± 0.8 ka (n = 2), and ages on additional erratic boulders from the Alapah River and Erratic Creek valleys show that ice had receded farther upvalley between 15.7 ± 0.7 and 14.9 ± 0.7 ka. This timing of deglaciation of central Brooks Range valleys is supported by erratic boulders and bedrock from the Arrigetch Peaks area that place the ice to within <0.5 km of Neoglacial limits by 15.3 ± 0.6 ka (n = 3).

DISCUSSION

We find no moraine evidence for an Itkillik II advance at ca. 30–25 ka (Hamilton, 1986; Briner et al., 2005), but it is possible that an Itkillik II advance overran previously deposited Itkillik II moraines. Our oldest exposure age of 24.2 ± 1.0 ka could be related to an earlier Itkillik II advance, although there is no evidence to further constrain this. The culmination of the maximum Itkillik II glaciation in the central Brooks Range at ca. 21 ka is in agreement with ages on LGM terminal moraines from elsewhere in Alaska (e.g., Kaufman et al., 2011). The age also agrees with that of a morphostratigraphically similar moraine in the northeast Brooks Range dated to ca. 21 ± 3.0 ka (n = 4; recalculated using the same ^{10}Be production rate used here; Balascio et al., 2005a).

The late Itkillik II re-advance moraine (ca. 17 ka) in our study area was built earlier than the

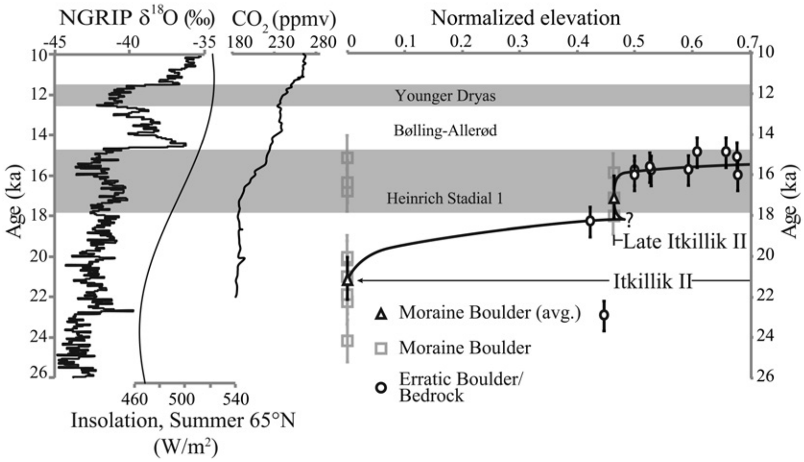

Figure 2. Time-distance diagram of glacier fluctuations spanning the last glacial termination, central Brooks Range (Alaska, USA). Location of ^{10}Be ages from multiple valleys are normalized relative to elevation of Itkillik II moraine (0) and cirque floor (1). Greenland δ^{18}O from Andersen et al. (2004), insolation from Berger and Loutre (1991), atmospheric CO_2 from Monnin et al. (2001). NGRIP—North Greenland Ice Core Project; avg.—average.

previously published radiocarbon chronology suggests (Hamilton, 2003), and in some Brooks Range valleys, its position upvalley from Itkillik II terminal moraines requires a period of substantial deglaciation between ca. 21 ka and ca. 17 ka. This period of deglaciation coincides with regional warming across northern and central Alaska (Kurek et al., 2009), and with glacial retreat in the Alaska Range and Ahklun Mountains (Briner et al., 2005; Howley and Licciardi, 2008). However, the timing of glacial retreat in the Brooks Range differs from that of western conterminous U.S. glaciers and Northern Hemisphere ice sheets, most of which began retreating from their terminal LGM positions between ca. 19 ka and 16 ka (Schaefer et al., 2006; Clark et al., 2009; Young et al., 2011). Additionally, this initial retreat in the Brooks Range occurred prior to the sharp rise in atmospheric CO_2 at ca. 17.5 ka (Monnin et al., 2001).

This early deglaciation might be related to the impact of the expanded Laurentide Ice Sheet (LIS) on atmospheric circulation. Simulations by global climate models show that the LIS caused significant warming and drying in the Alaska-Yukon region during the LGM (Roe and Lindzen, 2001; Otto-Bliesner et al., 2006), which agrees with temperature reconstructions based on chironomids (Kurek et al., 2009) and pollen (Bartlein et al., 2011). Increasing Northern Hemisphere insolation beginning at ca. 21 ka (Fig. 2; Berger and Loutre, 1991) likely enhanced the warm and dry conditions, promoting deglaciation from ca. 21 ka to 17 ka. The decrease in the size of the LIS at ca. 19 ka (Dyke, 2004) might have resulted in a reorganization of atmospheric circulation, reintroducing cold Arctic air over northern Alaska, decreasing summer temperatures, and culminating the late Itkillik II re-advance at ca. 17 ka. The timing of this re-advance also compares well with inferred cooling across Beringia from ca. 18 ka to 17 ka (Clark et al., 2012).

Following deposition of the late Itkillik II re-advance moraines, glaciers in the central Brooks Range apparently retreated rapidly, exposing all middle and upper valley reaches between ca. 16 and 15 ka and reaching their Neoglacial limits by ca. 15 ka. This second phase of deglaciation seems to have occurred simultaneously across the central Brooks Range and is supported by additional ^{10}Be ages from Brooks Range valleys to the east and west (Badding et al., 2013). Global atmospheric CO_2 levels rose dramatically at 17.5 ka (Monnin et al., 2001), likely resulting in increased temperatures in northwestern Alaska (Kurek et al., 2009) and perhaps outcompeting atmospheric effects driven by the LIS.

Although deglaciation following deposition of late Itkillik II re-advance moraines coincides with ages of moraine abandonment and deglaciation at ca. 17–14 ka in the western conterminous U.S. (Young et al., 2011), Brooks Range glaciers reached their cirques earlier. Glaciers in the western U.S. show evidence for continuing deglaciation until ca. 13 ka (Young et al., 2011), at which time Brooks Range glaciers were upvalley of their Neoglacial limits or possibly even absent. In the North Atlantic sector of the Arctic, decreasing sea-surface and winter temperatures occurred from 17.8 ka to 15.0 ka, and there was limited deglaciation during Heinrich Stadial 1 (Barker et al., 2009), which is at odds with Brooks Range deglaciation. By the time substantial warming began in the North Atlantic region during the Bølling-Allerød (ca. 14.7 ka), glaciers in the central Brooks Range had already retreated to within their Neoglacial limits. Our record suggests that Brooks Range glaciers show tighter coupling to CO_2 forcing than at lower latitudes. This chronology and correlation across records suggests that during the LGM, glacier fluctuations in the Alaskan Arctic were largely modulated by LIS dynamics until atmospheric CO_2 took over as the main driver of deglaciation following the LGM.

Any glacier re-advance during the late-glacial period (ca. 14–10 ka) was within Neoglacial limits. Previous studies in the Brooks Range have also found no evidence of a Younger Dryas (YD) glacial expansion (Hamilton, 1986; Balascio et al., 2005a; Badding et al., 2013), which agrees with climate records from central and northern Alaska indicating near-present or higher-than-present summer temperatures and reduced effective moisture during the late-glacial period (Kokorowski et al., 2008; Kurek et al., 2009). Glacial records from Baffin Island and Svalbard show that glaciers were more extensive during Neoglaciation than during the YD, and also point to aridity and seasonality as controlling factors (Mangerud and Landvik, 2007; Young et al., 2012).

CONCLUSIONS

The direct dating of Itkillik II moraines and deglacial features in the central Brooks Range bridges a gap in the glacial chronology from the LGM to Neoglaciation, and provides the most complete record of late Pleistocene alpine glaciation in the Arctic during the last glacial termination. Substantial glacier retreat between ca. 21 ka and ca. 17 ka precedes global CO_2 increase, but is coincident with a rise in orbitally driven summer insolation and is consistent with the impact of the LIS on upstream atmospheric circulation. A contracted LIS at ca. 19 ka may have induced cooling in the Brooks Range, leading to the late Itkillik II re-advance. Following this cool period, Brooks Range glaciers retreated rapidly from ca. 16 ka to 15 ka. This chronology suggests that increasing atmospheric CO_2 at ca. 17.5 ka became the dominant climate forcing in this region, driving swift glacier retreat. The lack of evidence for a glacier advance during the late-glacial period here and at some other Arctic sites contrasts with lower-latitude settings where late-glacial moraines are more widespread, and suggests strong latitudinal variations of climate forcings within the Northern Hemisphere at this time. This new chronology of alpine glaciation in the Arctic during and following the LGM suggests that ice sheet–influenced atmospheric patterns and greenhouse gases are the competing drivers of climate and glacier activity in this region.

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