Cosmogenic exposure dating of late Pleistocene moraine stabilization in Alaska

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ABSTRACT

Seventy-three new $^{10}Be/^{26}Al$ ages from 57 moraine boulders and 2 tors, together with 43 previously published cosmogenic exposure ages from 41 moraine boulders, allow us to critique the use of cosmogenic exposure (CE) dating of moraine boulders in Alaska. Boulder exhumation during moraine degradation likely gives rise to the largest uncertainty in constraining the timing of initial moraine stabilization following ice retreat. Isotopic inheritance appears to be most important for moraines deposited close to their cirque headwalls. Boulder-surface (bedrock) erosion rate can be roughly constrained and leads to a range in moraine stabilization ages. Snow-cover history is difficult to constrain, but its effect is thought to be minor for the tall boulders sampled.

Despite these complications, the CE ages provide important new information regarding the timing of the last and penultimate glaciations in Alaska. Three penultimate moraines yielded CE ages that overlap with marine isotope stage (MIS) 4/early MIS 3 (45–65 ka) rather than MIS 6 (ca. 140 ka). Based on a combination of our new CE chronologies and existing ^{14}C ages from six study areas, glaciers retreated from their local late Wisconsin maxima: ca. 24–27 ka, Kokrines Hills (west-interior Alaska); ca. 24–26 ka, northeastern Brooks Range (NE Alaska); ca. 21–23 ka, Yukon Tanana Upland (east-interior Alaska); ca. 22 ka, Ahklun Mountains (SW Alaska); ca. 20 ka, western Alaska Range (central Alaska); ca. 16–18 ka, Chuilnuk Mountains (SW Alaska). Overall, glacier retreat was concurrent with the peak of the last global glacial maximum, probably in response to limited moisture availability.

Keywords: cosmogenic exposure dating, moraine, Alaska, last glacial maximum, penultimate glaciation.

INTRODUCTION

Alaska is unique within the Arctic because, unlike most northern regions that were covered by vast ice sheets during the late Pleistocene, Alaska encompasses numerous mountain ranges, each with an independent record of glaciation (Fig. 1). Mountain glaciers are key indicators of Pleistocene climate, and the glacial record in Alaska provides a rich source of paleoclimate information for the North Pacific region. Alaska is also one of the most accessible regions of the Arctic to study late Pleistocene alpine glaciation, yet uncertainty lingers regarding the spatial and temporal pattern of ice extent during the late Wisconsin (roughly 30–10 ka; ka is used as thousand of years before present) and penultimate glaciations.

The strikingly limited extent of late Wisconsin glaciers in Alaska relative to previous advances has long been recognized (Péwé et al., 1965; Hopkins, 1982; Hamilton et al., 1986a; Kaufman and Manley, 2004). Most mountain ranges preserve clear morainal evidence for the extent of pre–late Wisconsin (penultimate) glaciers. These deposits are rarely datable by ^{14}C, either because organic matter is too sparse or because the moraines are beyond the range of ^{14}C dating. Although the timing of late Wisconsin glacier advances has been constrained in a few regions, the ages of glacier fluctuations remain poorly known in most areas across Alaska. For example, it is still debated whether the penultimate advance occurred prior to the last global interglaciation (e.g., marine isotope stage (MIS) 5e; e.g., Begét, 2001), or during an early stade of the last glaciation (early Wisconsin sensu lato; late MIS 5 or MIS 4; Hamilton, 1994).

The advent of cosmogenic radionuclide exposure dating (Gosse and Phillips, 2001; Cockburn and Summerfield, 2004) has profoundly improved the ability to date glacial deposits and has now been applied nearly worldwide (e.g., Phillips et al., 1997; Gualtieri et al., 2000; Shanahan and Zreda, 2000; Barrows et al., 2001; Briner et al., 2001; Owen et al., 2001; Miller et al., 2002; Landvik et al., 2003; Stone et al., 2003). Cosmogenic exposure (CE) dating in glaciated settings has focused on moraine boulders and has provided direct ages for dozens of moraines deposited by continental ice sheets

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and alpine glaciers, in many places providing the first available quantitative chronology. While the technique has proven successful, previous studies have brought to light potential complications, especially when CE dating is used on degrading landforms such as moraines (e.g., Hallet and Putkonen, 1994; Zreda et al., 1994; Putkonen and Swanson, 2003).

We used CE dating to establish the timing of moraine deposition and subsequent stabilization for 23 moraines across Alaska. Here, we highlight the complications that we have faced while trying to establish glacial chronologies over the past seven years. Our results from the Ahklun Mountains (Briner et al., 2001, 2002), our first study site, helped guide our sampling strategy and age interpretations for late Pleistocene moraines elsewhere in Alaska. We discuss the outcomes that contribute to our understanding of the late Pleistocene glacial history of Alaska, as well as many challenges that have hampered our ability to constrain moraine ages.

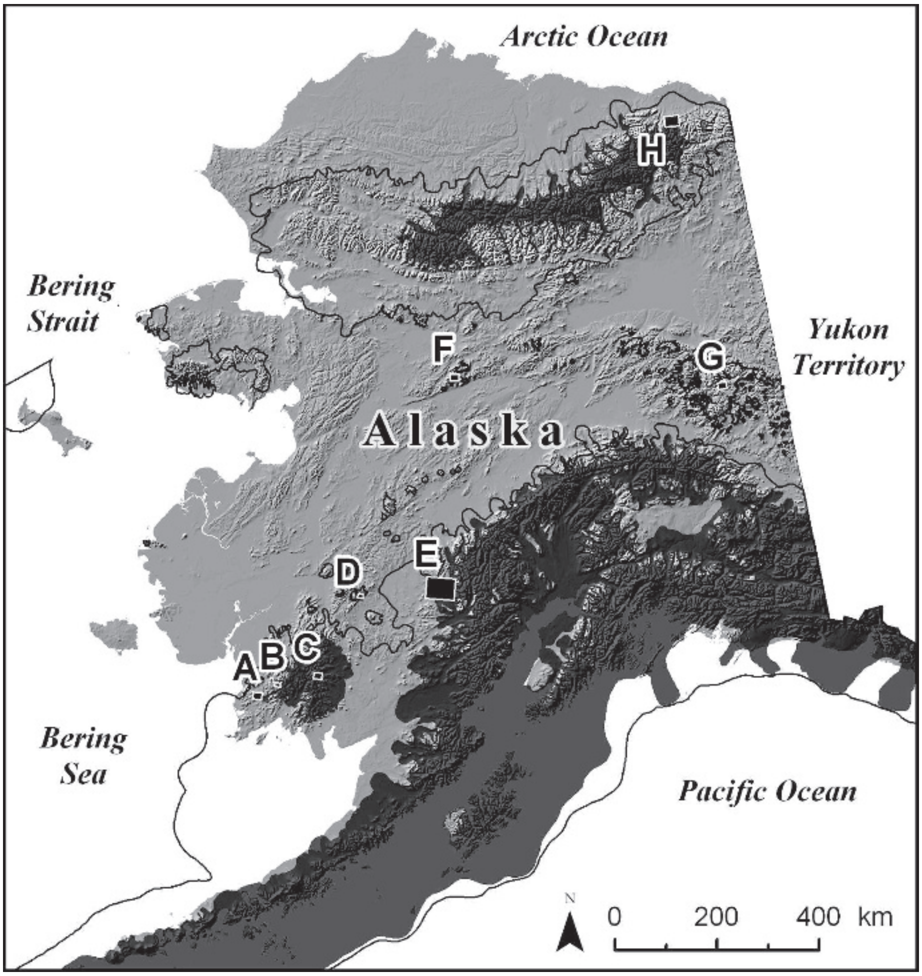

Figure 1. Map of Alaska showing study locations in the Ahklun Mountains (A–C), Chuilnuk Mountains (D), Lime Hills region (E), Kokrines Hills (F), Yukon-Tanana Upland (G), and northeastern Brooks Range (H). Dark shading depicts late Wisconsin ice extent; bold line outlines maximum extent of Pleistocene drift (from Manley and Kaufman, 2002).

COSMOGENIC EXPOSURE DATING CONSIDERATIONS

The uniform build-up of cosmogenic nuclides in earth surfaces assumes stable landscape conditions. Moraines do not fully satisfy this condition, especially during the early stages following abandonment by a glacier margin. Rather, they degrade through time by processes that ultimately remove material from the moraine crest and deposit it at the slope base (e.g., Bursik, 1991; Hallet and Putkonen, 1994). Based on moraine morphometry studies, it appears that Alaskan moraines degrade no faster than moraines in alpine environments at lower latitudes (e.g., Kaufman and Calkin, 1988; Peck et al., 1990; Manley et al., 2001; Balascio et al., 2005; versus Burke and Birkland, 1979; Colman and Pierce, 1986; Bursik, 1991). Nonetheless, we attempted to minimize the complications associated with moraine degradation by careful boulder sample selection. To reduce the chances of sampling exhumed boulders, we targeted large boulders, generally standing at least 1 m above the moraine surface. We avoided boulders on unstable slopes and searched for boulders along kilometers of moraine crest (using helicopter-based surveys in some cases). We also considered the stability of each boulder by favoring boulders with wide bases, which are unlikely to have rolled as they were exhumed or transported short distances below the moraine crest. We avoided boulders involved in obvious mass wasting, including those at the toes of solifluction lobes.

Many glacial valleys that we visited lacked boulders that satisfied our sampling criteria. Moraines lacking at least three large boulders could not be dated; on average, we dated four boulders on each moraine. We attribute the relative scarcity of suitable boulders to the limited distribution of durable lithologies and proximity to outcropping sources. The overall scarcity is probably a function of lithology and transport distance rather than significantly accelerated rock-weathering rates; for example, the bedrock erosion rate as determined at one site in this study is similar to rates measured elsewhere (e.g., Small et al., 1997). Excessive postdepositional erosion of boulder surfaces can lead to CE ages that differ from the depositional age of a moraine. We chose boulders that were no more weathered than neighboring boulders.

Finally, shielding by snow cover can lead to a CE age that is younger than the moraine age. The tall boulders from moraine crests that we tended to sample are also least prone to snow-shielding effects. Because most of our sampling sites are from moraine ridges above treeline, they are exposed to wind, which helps to further reduce snow shielding. Details on CE age methodology can be found in the Data Repository.[1]

LATE PLEISTOCENE GLACIAL HISTORY OF ALASKA: PREVIOUS WORK, NEW COSMOGENIC EXPOSURE AGES, AND INTERPRETATIONS

The glacial history of Alaska has been recently reviewed by Kaufman and Manley

[1]GSA Data Repository item 2005109, cosmogenic exposure dating methods, Table DR1: cosmogenic exposure ages (arranged in decreasing age per moraine), and Table DR2: cosmogenic isotope, derived maximum erosion rates, is available on the Web at http://www.geosociety.org/pubs/ft2005.htm. Requests may also be sent to editing@geosociety.org.

(2004), which built upon past reviews (e.g., Péwé et al., 1965; Péwé, 1975; Hamilton, 1986a, 1994). While Alaska was extensively glaciated during late Pleistocene ice advances, large portions of the state escaped glacial cover (cf. Manley and Kaufman, 2002). The Alaska Range, from the Aleutian Islands to the southeastern panhandle, was covered by amalgamated ice that formed the western extension of the Cordilleran Ice Sheet. Other major centers of glaciation in Alaska were the Brooks Range in the north and the Ahklun Mountains in the southwest, each of which contained ice caps separate from the Cordilleran Ice Sheet. Outside of these major centers of ice, smaller mountain ranges and massifs scattered across the state supported hundreds of smaller valley glaciers. Here, we focus on the CE ages and their contribution to understanding the timing of late Wisconsin and penultimate advances in Alaska. Most mountain ranges exhibit evidence for multiple glacier fluctuations during the late Wisconsin, and we have analyzed samples from downvalley moraine sequences to determine whether their ages were distinguishable using CE dating. The presentation is organized according to regions of the state where we have worked, beginning with an overview of previous work, then followed by our results and interpretations.

Ahklun Mountains

The Ahklun Mountains, southwestern Alaska (Fig. 1), have been studied extensively in the last decade, and a detailed mid- and late-Quaternary glacial history has been produced by a combination of surficial mapping, stratigraphic, and lake-core studies, coupled with a suite of geochronological methods (Kaufman et al., 1996; Briner and Kaufman, 2000; Briner et al., 2001; Manley et al., 2001; Kaufman et al., 2001a, 2001b; Briner et al., 2002; Kaufman et al., 2003; Levy et al., 2004; Axford and Kaufman, 2004). During multiple Pleistocene glaciations, the Ahklun Mountains hosted an ice cap over its east-central spine that expanded radially outward, extending farther to the south and west than to the north and east; isolated alpine glaciers occupied the highest valleys beyond the ice cap margin (Fig. 2). Early and late Wisconsin drift was deposited during the locally termed Arolik Lake and Klak Creek glaciations, respectively (Briner and Kaufman, 2000). In most valleys, the drift of both glaciations is composed of several (2–5) moraine belts formed by outlet glaciers of the central ice cap (Manley et al., 2001). Drift of the Klak Creek glaciation consists of one or two end moraines (Briner and Kaufman, 2000).

Previous Age Control

Age constraints on the late Pleistocene glacial history of the Ahklun Mountains, similar to all other glaciated ranges in Alaska, are few and far between. The age of Arolik Lake (early Wisconsin) drift is bracketed by two dates from the southern Ahklun Mountains. Kaufman et al. (2001b) report a thermoluminescence age of 70 ± 10 ka on lava-baked sediments that underlie Arolik Lake drift and provide a maximum-limiting age on the glaciation. Manley et al. (2001) report a minimum limiting age of >39.9 ^{14}C ka on organic material that overlies Arolik Lake drift. The age of the Klak Creek (late Wisconsin) drift is provided by several ^{14}C determinations that bracket one or more of the hummocky drift belts that comprise the Klak Creek drift. Manley et al. (2001) report a close minimum-limiting age of 19,950 ± 250 yr B.P. for the next-to-oldest hummocky drift belt of the Klak Creek glaciation (all finite ^{14}C ages are reported as calibrated ages using CALIB 4.4.1). The age of the outermost hummocky drift belt of the Klak Creek glaciation is best constrained by overflow sediments of a glacier-dammed lake that were recently recovered in the extant Arolik Lake. Radiocarbon ages bracket the maximum extent of ice during the Klak Creek glaciation between 24 ka and 22 ka (Kaufman et al., 2003).

In summary, the Ahklun Mountain ice cap expanded during the Arolik Lake glaciation sometime between 70 ± 10 ka and 40 ka. Glaciers retreated and a prominent interstadial unit was deposited around the Bristol Bay lowland (Lea et al., 1991). Glaciers then readvanced and reached their maximum extent of the Klak Creek glaciation between 24 ka and 22 ka.

Cosmogenic Exposure Ages

The largest concentration of CE ages from Alaska comes from the Ahklun Mountains. These include 32 ^{36}Cl ages on Arolik Lake and Klak Creek drift (4 and 28, respectively; Briner et al., 2001), and seven ^{10}Be and four ^{26}Al ages from a latest Pleistocene readvance moraine (Briner et al., 2002). This relatively large data set includes many CE ages undoubtedly influenced by inheritance (yielding an apparent age older than deposition) and exhumation (yielding an apparent age younger than deposition). On some moraines, multiple boulders (3–6 per moraine) yielded a tight cluster of CE ages, while on others, the age series are widely scattered. Our confidence in the chronological integrity of the CE data is high for moraines with well-clustered ages and lower for moraines with widely scattered ages.

Of the seven moraines dated by Briner et al. (2001), all in the southwestern sector of the range (Fig. 2), two are from ice-cap outlet glacier moraines. One of the moraines was deposited during the Arolik Lake glaciation; four granodiorite boulders sampled from atop a thin, stable ground moraine sheet deposited in flat saddle between two valleys have ^{36}Cl ages that range from 57.5 ± 1.6 to 63.9 ± 2.3 ka (Fig. 3A). Unadjusted for boulder-surface erosion, the ages average 60.3 ± 3.2 ka; using reasonable estimates for boulder erosion and snow cover, the average age ranges between 50.2 ± 2.0 and 65.1 ± 4.1 ka (Briner et al., 2001). Because the sampled boulders are ≥2 m high, snow-cover estimates are probably maximums, and we favor the younger part of this age range. Nevertheless, the age range falls within the bracketing ages of correlative drift in the southern Ahklun Mountains (Fig. 3A) and suggests that the penultimate drift in the Ahklun Mountains was deposited during the early Wisconsin rather than prior to the last interglaciation (MIS 6).

Five metabasaltic boulders from a hummocky moraine deposited by an ice-cap outlet glacier (Gusty Lakes moraine) during the Klak Creek glaciation yielded ^{36}Cl ages that range between 20.7 ± 0.4 and 9.4 ± 0.6 ka. Because the Gusty Lakes moraine is close and easily correlated to the moraine with tight ^{14}C constraints (between 24 and 22 ka from Arolik Lake and >20 ka from 20 km upvalley; see above), we can independently establish its age and evaluate the ^{36}Cl ages. This pattern of ages provides insights into the CE age distribution for a moraine (Fig. 3B), while keeping in mind that ^{14}C ages provide limiting maximum and minimum ages, whereas CE ages date moraine stabilization, which occurs following ice retreat but continues as the moraine degrades through time (Hallet and Putkonen, 1994; Putkonen and Swanson, 2003). Reasonable boulder-surface erosion rates do not affect these particular CE ages (Briner et al., 2001), and the impact of snow cover on these relatively tall boulders is likely small. The decreasing age spread with increasing age (Fig. 3B) suggests that these boulders were unlikely to have been influenced by inheritance. Instead, the age spread suggests that one or more of these boulders were exhumed or rolled following initial moraine stabilization. Thus, once ice retreated from the Gusty Lakes moraine ca. 22 ka, the moraine largely stabilized during the following two millennia, as two of the boulders yield ages of 20.7 and 20.2 ka and a third is 19.3 ka. The other two boulders (17.6 and 9.4 ka) were apparently exhumed following initial moraine stabilization.

The CE age distribution on the independently dated Gusty Lakes moraine provides a case study for interpreting CE age distributions on other moraines that typically lack independent

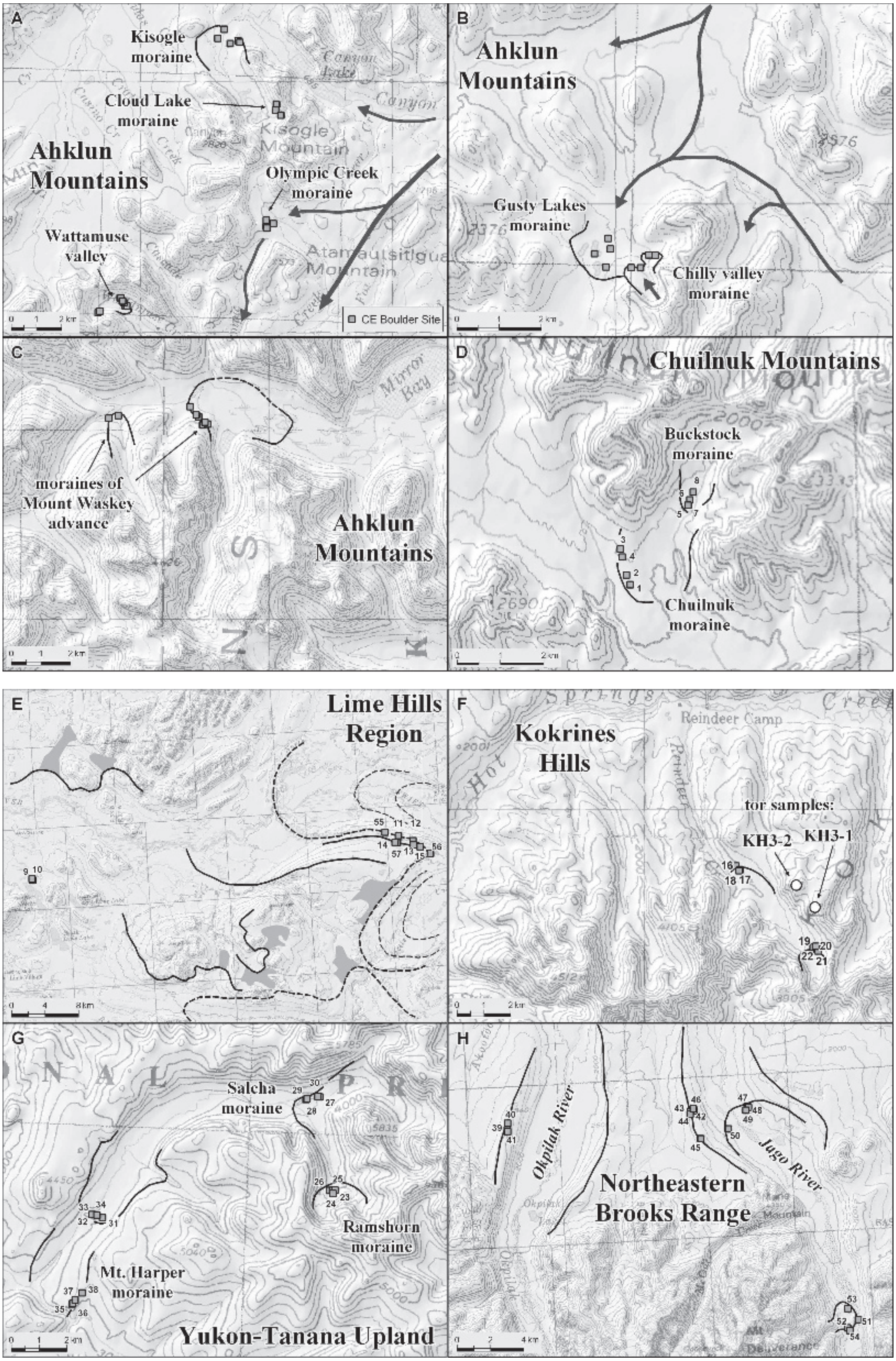

Figure 2. The eight study regions, showing representative glacial geology and CE age locations (squares). Numbers refer to CE age locations listed in Table DR1 (see footnote 1). Bold lines with arrows in panels A and B show ice cap outlet glacier flow paths. Black lines in all panels are moraines. In panel E, dashed lines are late Wisconsin moraines and solid lines are penultimate moraines based on moraine morphometry; gray areas are ice-dammed paleolakes. Study locations shown in Figure 1.

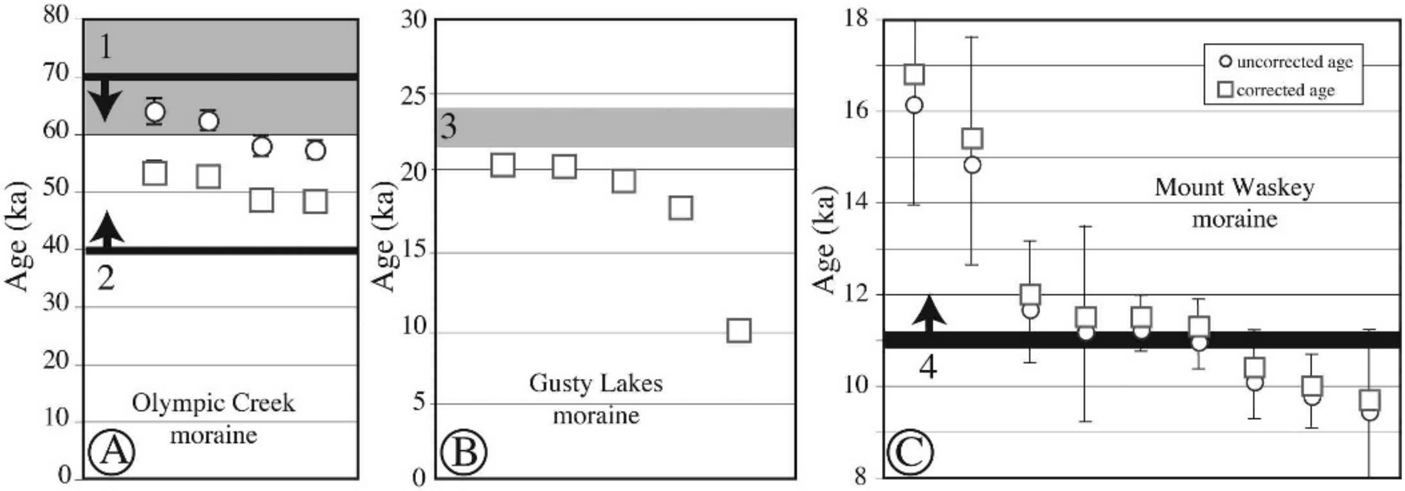

Figure 3. CE ages from moraines in the Ahklun Mountains with independent age constraints. (A) The Olympic Creek moraine is younger than 70 ± 10 ka (denoted by line 1) and older than ca. 40 ka (line 2). Four ^{36}Cl ages become younger when corrected for boulder-surface (bedrock) erosion, and suggest an Ahklun Mountain ice cap advance around the MIS 4/3 boundary. (B) Ice retreated from the Gusty Lakes moraine ca. 22 ka, the time when ice-dammed lake sedimentation (shaded zone 3) ceased at a correlative moraine (note change in age axis). The five ^{36}Cl ages from the Gusty Lakes moraine are not sensitive to boulder-surface erosion, and show that CE ages best date moraine stabilization. (C) The Mount Waskey moraine was deposited before ca. 11 ka (line 4); ^{10}Be/^{26}Al ages younger than the timing of Mount Waskey moraine deposition indicate moraine degradation.

age control. Without an independent age check, there would be several approaches to interpreting the CE ages. The first distinction is whether the CE ages constrain the timing of moraine formation (when the glacier is still present), moraine abandonment (deglaciation), or moraine stabilization (shortly following deglaciation). The Gusty Lakes moraine ages seem to most closely constrain the timing of moraine stabilization. Second are the different approaches to calculating the moraine age, namely by calculating the arithmetic mean, weighted mean, or simply to favor the oldest age or ages as the most probable moraine age. The CE data from the Gusty Lakes moraine yield an average age of 17.4 ± 4.7 ka, 19.5 ± 1.4 ka after excluding the obvious 9.4 ka outlier, and a weighted mean of 19.6 ± 1.4 ka. With this data set, the best approach is probably to favor the oldest age, in this case 20.7 ± 0.4 ka, which is ~1000 yr younger than the timing of moraine abandonment as determined by ^{14}C ages. Thus, the ^{36}Cl ages from the Gusty Lakes moraine support weighing the oldest age or ages within an age cluster (this becomes more complicated in cases that lack an age cluster; see below) to obtain the age of initial moraine stabilization, which occurs following deglaciation.

Another example from the Ahklun Mountains supports the conclusions drawn from the Gusty Lakes moraine CE ages. Briner et al. (2002) obtained CE ages on nine granodiorite moraine boulders (five ^{10}Be ages, two ^{26}Al ages, and two ^{10}Be/^{26}Al average ages) from moraines deposited during the late-glacial, Mount Waskey advance. Waskey Lake, which is impounded by a moraine deposited during the Mount Waskey advance, began accumulating sediment following 11.0 ± 0.2 ka (Levy et al., 2004); thus the basal lake sediments provide a minimum age of ca. 11 ka for ice retreat from the Mount Waskey advance. The CE age distribution includes two ages ca. 15 ka, and seven that range between 11.7 ± 1.2 and 9.4 ± 1.5 ka (Fig. 3C). The two oldest boulders are interpreted as containing inherited cosmogenic isotopes. Of the seven ages that compose the main age cluster, three are younger than the minimum age of the advance, based on the ca. 11 ka ^{14}C age (Fig. 3C). Thus, similar to the Gusty Lakes moraine, at least three of the Mount Waskey moraine boulders were apparently exhumed or rolled following initial moraine stabilization. Similarly, weighting the oldest ages rather than taking the average of all of the analyses seems to provide the best correspondence with the independent age.

Alpine glacier moraines of the local LGM (Last Glacial Maximum; Klak Creek glaciation) were dated in three valleys in the western Ahklun Mountains (Briner et al., 2001) and provide further insights into how to interpret moraine boulder CE ages. In one valley, ^{36}Cl ages were obtained on five metabasaltic moraine boulders from the maximum Klak Creek moraine, as well as five from a cirque moraine upvalley. Both moraine ages cluster between ca. 15 and 19 ka; boulders on the downvalley moraine include a ca. 3 ka outlier, and the upvalley moraine includes a ca. 24 ka outlier. The downvalley and upvalley moraines have average ^{36}Cl ages of 17.5 ± 0.9 and 16.7 ± 1.4 ka, respectively, once the outliers are excluded. The oldest CE ages from the downvalley and upvalley moraines are 18.6 ± 1.8 and 18.1 ± 1.4 ka, respectively. Thus, it appears that either the alpine glacier that occupied this valley did not retreat until ca. 3 ka following the Ahklun Mountain ice cap retreat (Briner et al., 2001) or that moraine stabilization lagged behind the stabilization of the Gusty Lakes moraine. Alternatively, the boulders to first stabilize on the moraine were not sampled. The ca. 24 ka age on one of the cirque moraine boulders is probably influenced by inheritance, likely arising from prior exposure on the cirque headwall.

In a second valley, ^{36}Cl ages were obtained on four metabasaltic boulders from an alpine glacier end moraine. The ages are scattered, ranging from 26.7 ± 1.0 to 6.2 ± 0.2 ka. The moraine directly overlies the Gusty Lakes moraine, thus, stabilized after ca. 21 ka, suggesting that some of its boulders contain inherited radionuclides while others were exhumed long after deglaciation.

In a third valley, granodiorite boulders were sampled from a terminal and recessional moraine. Compared with other CE ages from Ahklun Mountain moraines, the ages are unexpected. Boulders on the terminal moraine have ages that range between 57.7 ± 4.4 and 20.4 ± 0.9 ka (n = 6), with no strong cluster, but a weak one at ca. 30 ka. The three boulders analyzed from the recessional moraine have ^{36}Cl ages of 47.0 ± 1.3 ka, 26.4 ± 1.6 ka, and 25.9 ± 0.8 ka. Thus, the ages from this valley are older and more scattered than the CE ages from elsewhere on similar-aged deposits, suggesting persistent inheritance issues, although the cause remains uncertain.

Chuilnuk Mountains

We sampled a moraine sequence for CE dating in the Chuilnuk Mountains (Fig. 1), ~100 km NNE of the Ahklun Mountains. This small mountain range was studied in detail by Waythomas (1990), who delimited glacial drift of four relative ages that probably span from the mid-Quaternary to the Holocene. The Chuilnuk Mountains and the neighboring Kiokluk Mountains are each composed of a deeply dissected, high-elevation plutonic core from which glacially eroded valleys are oriented radially outward. The ranges were apparently overwhelmed during the maximum extent of Quaternary ice cover, but ice was restricted to the valleys, terminating near the range front, during the penultimate (Chuilnuk) and last (Buckstock) glaciations (Waythomas, 1990).

Previous Age Control

Prior to our study, no absolute age control was available in the Chuilnuk Mountains. Detailed relative-weathering studies by Waythomas (1990) provide the only chronological assessment of the drift units. The majority of relative-weathering criteria (surface morphometry and soil development characteristics) distinguish the Chuilnuk from the Buckstock drift. Waythomas suggested that the penultimate drift is early Wisconsin rather than MIS 6 in age because no significant break in the extent of relative weathering separated the deposits. Waythomas (1990, p. 166) reported that "in the Chuilnuk and Kiokluk Mountains, relative-weathering characteristics between Chuilnuk and Buckstock age moraines suggest that the deposits are not significantly different in age." Nonetheless, because early Wisconsin eolian deposits appear to be absent in the adjacent Holitna Lowland, Waythomas (1990) ultimately correlated the Chuilnuk drift to MIS 6 and the Buckstock drift to MIS 2.

Cosmogenic Exposure Ages

The four most stable of five sampled granodiorite boulders from each of a Chuilnuk and Buckstock moraine were chosen for ^{10}Be analysis (Fig. 2). Four ^{10}Be ages from the Chuilnuk moraine range between 9.3 ± 1.8 and 15.8 ± 1.8 ka, and four ^{10}Be ages from the Buckstock moraine range from 14.3 ± 1.2 to 17.7 ± 1.5 ka (Table DR1 [see footnote 1]; Fig. 4). Correcting these ages to account for snow cover and boulder-surface erosion is difficult. We ignore any possible effect of snow cover. For minimum boulder-surface erosion, we adopt a value of ~1 mm k.y.$^{-1}$ reported from the eastern Canadian Arctic by Bierman et al. (1999). This value was calculated from the ^{10}Be concentration in gneissic bedrock assumed to be at steady-state with regard to surface erosion and ^{10}Be production and is probably no lower than in Alaska where climate is warmer and wetter than the

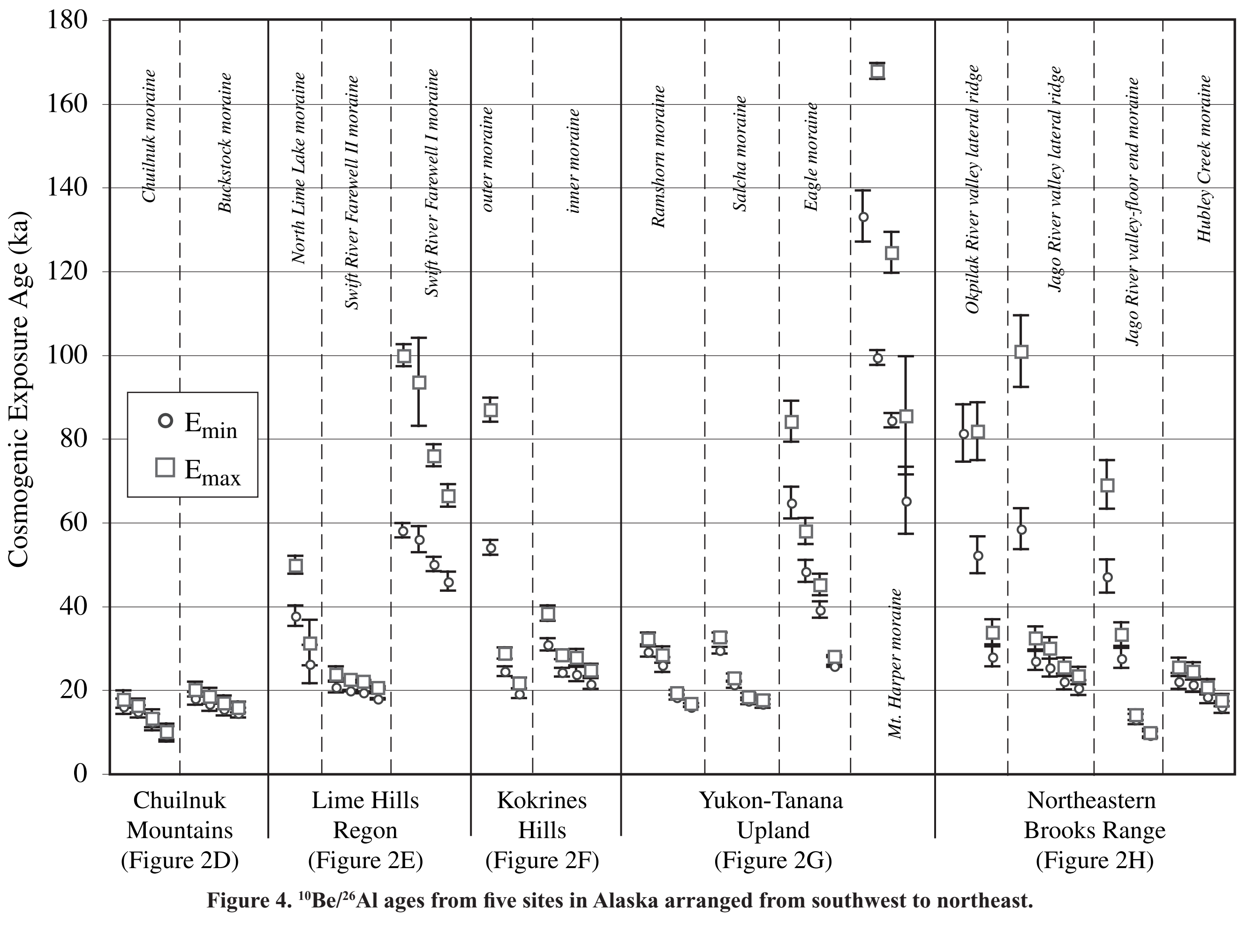

Figure 4. $^{10}Be/^{26}Al$ ages from five sites in Alaska arranged from southwest to northeast.

eastern Canadian Arctic. We used cosmogenic nuclides to obtain a maximum bedrock erosion rate for interior Alaska (from the Kokrines Hills; see below) by analyzing samples from upland tors assumed to be in steady-state with respect to erosion. The tors are composed of quartz monzonite with feldspar phenocrysts up to 2 cm long; ^{10}Be concentrations indicate a maximum erosion rate of <8.2 ± 0.8 mm k.y.$^{-1}$ (n = 2; Table DR2 [see footnote 1]). This value is similar to results from the Wind River Range in the Rocky Mountains where upland summit bedrock erosion rate is ~5–10 mm k.y.$^{-1}$ (Small et al., 1997). Moraine boulder-surface erosion rate is estimated at ~0–1 mm k.y.$^{-1}$ from the same mountain range (Phillips et al., 1997), indicating that glacially excavated bedrock initially might weather slowly relative to long-term bedrock erosion. Thus, our value of 8 mm k.y.$^{-1}$ is most likely a far maximum. For the CE ages from the Chuilnuk Mountains, as well as for the remainder of the sites discussed below, we report CE ages with both minimum (ε_{min}: 1 mm k.y.$^{-1}$) and maximum (ε_{max}: 8 mm k.y.$^{-1}$) estimates of boulder-surface erosion rates.

Incorporating the uncertainty associated with the range in possible boulder-erosion rates, the average of four ^{10}Be ages from the Chuilnuk moraine ranges between 13.1 ± 2.9 (ε_{min}) and 14.3 ± 3.4 ka (ε_{max}), and the ages from the Buckstock moraine ranges between 16.1 ± 1.5 (ε_{min}) and 17.8 ± 1.9 ka (ε_{max}) (Table DR1 [see footnote 1]; Fig. 4). The age clusters contain no obvious sample with inheritance but do contain obvious exhumed boulders. Using the "oldest-age method" (as for the Ahklun Mountains results discussed above) to provide a constraint for moraine stabilization suggests a ca. 16 (ε_{min}) to ca. 18 ka age (ε_{max}) for the Chuilnuk moraine, and a ca. 18 (ε_{min}) to ca. 20 ka (ε_{max}) age for the Buckstock moraine (Table DR1 [see footnote 1]). This approach leads to a greater geochronological discrepancy, however, because the Chuilnuk moraine lies downvalley from, and is therefore stratigraphically older than, the Buckstock moraine. The larger age spread among CE ages from the Chuilnuk moraine, along with the age inversion, suggests that either the oldest boulders on the Chuilnuk moraine were not sampled or that the moraine is degrading faster than the Buckstock moraine for some unknown reason. In any case, none of the CE ages is indicative of a pre–late Wisconsin age, but rather support a late Wisconsin age for both moraines.

Lime Hills Region

The Lime Hills region of the foothills of the western Alaska Range (Fig. 2) are marked by a rich display of end and lateral moraines that emanate from several river valleys in the area, typical of the northern and western Alaska Range fronts (Hamilton and Thorson, 1983). The glacial geology of the area has been previously mapped at a regional scale only (e.g., Coulter et al., 1965). The Farewell area to the immediate north was mapped by Kline and Bundtzen (1986; building on earlier work of Fernald, 1960), who identified a fivefold glacial sequence that spans from the late Tertiary to the Holocene. For this study, dozens of ice margins were delimited by mapping moraines and ice-dammed lakes on air photographs along two major river valleys (the Stony and Swift River valleys) and several of their tributaries (Fig. 2). We identified the position of a significant morphostratigraphic break, inferred to separate late Wisconsin from pre–late Wisconsin moraines along the Swift River valley. Large moraine boulders were sampled for CE dating from age classes of both moraines (Fig. 2).

Previous Age Control

Previous chronological control on late Pleistocene glacial advances in the Lime Hills area does not exist. The moraine sequence is similar to that in numerous other valleys across the northern Alaska Range; they consist of at least two major drift units, each deposited in multiple phases (e.g., Ten Brink and Waythomas, 1985; Kline and Bundtzen, 1986; Thorson, 1986). This similarity has provided the basis for correlations, and the chronology from elsewhere has been applied to the moraines in the Swift and Stony River valleys. In the Farewell area, for example, deposits of Farewell I (early Wisconsin) and Farewell II (late Wisconsin) glaciations are separated by a ^{14}C age of ca. 34 ka (Kline and Bundtzen, 1986).

Farther northeast, for example in Denali National Park and in the Nenana River valley, glacier deposits have received modest attention, and a fourfold sequence of late Wisconsin moraines is reasonably well dated. Porter et al. (1983) provide the most detailed review to date, although not the most recent (Hamilton, 1994), on the existing ^{14}C dates for the timing of late Wisconsin glacier advances. Several maximum-limiting ^{14}C ages constrain the initial late Wisconsin advance to sometime after ca. 27 ka (Porter et al., 1983). In Denali National Park, the maximum late Wisconsin (McKinley Park [MP] I) moraine is constrained between 21.4 ± 0.7 and 20.6 ± 0.5 ka (Ten Brink and Waythomas, 1985; Werner et al., 1993). Three younger phases are constrained between 20.6 ± 0.5 and 19.9 ± 0.3 ka (MP II; Werner et al., 1993; Child, 1995), 15.1 ± 0.7 and 12.3 ± 0.5 ka (MP III; Child, 1995; Ten Brink and Waythomas, 1985), and 12.3 ± 0.5 and 11.0 ± 0.2 ka (MP IV; Ten Brink and Waythomas, 1985).

Cosmogenic Exposure Ages

The boulders on the late Pleistocene moraines in the Lime Hills area, specifically on the lateral moraines bordering the Swift River and Little Underhill Creek valleys, are the largest and probably the most stable boulders we sampled in Alaska. We sampled ≥3-m-high granite boulders from a sharp-crested terminal moraine deposited during the Farewell II glaciation (we apply the Farewell nomenclature to the moraines in the Lime Hills area). We also sampled >1.5-m-high granite boulders from a much more subdued moraine ridge ~1 km beyond the Farewell II terminal moraine that we assigned to the Farewell I glaciation. The surfaces of boulders on Farewell II moraines are smooth with little grain relief; in contrast, the Farewell I boulders exhibit significant surface weathering, including weathering pits up to 10 cm deep. Finally, we sampled two other ground-moraine boulders near North Lime Lake, ~45 km downvalley from the Swift River lateral moraine site, which were deposited by ice spilling north from the Stony River valley (Fig. 2).

Four CE ages for Farewell I lateral moraine boulders range from 55.4 ± 1.4 to 44.2 ± 1.9 ka and average 52.5 ± 5.6 (ε_{min}) to 83.9 ± 15.5 ka (ε_{max}; Table DR1 [see footnote 1]; Fig. 4). The moraine stabilization age according to the "oldest-age method" ranges between ca. 58 (ε_{min}) and ca. 100 ka (ε_{max}). The relatively tight age cluster indicates that none of the boulders contains significant isotopic inheritance. Rather, the age spread is more likely due to postdepositional exhumation or differential boulder surface weathering among the sampled boulders. Nonetheless, the ages lie within the MIS 5–3 age range, as opposed to MIS 6, and suggest that the Farewell I glaciation occurred during the early Wisconsin. The ages also overlap the age range for the early Wisconsin (Arolik Lake) glaciation in the Ahklun Mountains (ca. 65 to ca. 50 ka).

Four large, stable boulders on the Farewell II terminal moraine were chosen for ^{10}Be and ^{26}Al analysis and yield individual ^{10}Be/^{26}Al average ages that range between 21.3 ± 0.9 and 18.2 ± 0.7 ka (Table DR1 [see footnote 1]; Fig. 4). Accounting for erosion, the ages average between 19.6 ± 0.9 (ε_{min}) and 22.3 ± 1.2 ka (ε_{max}). The moraine stabilization age according to the "oldest-age method" ranges between ca. 21 (ε_{min}) and ca. 24 ka (ε_{max}). However, unlike the Farewell I boulders, which contain 10-cm-deep weathering pits, Farewell II boulder surfaces are relatively smooth and lack irregularities greater than roughly 1 cm of grain relief. The lack of obvious boulder surface erosion suggests that their erosion-corrected age lies toward the younger value. The sampled Farewell II moraine in the Lime Hills region can be compared to the late Wisconsin terminal moraine in Denali National Park (MP I), which was previously dated between ca. 21.4 and ca. 20.6 ka. The age of the Swift River lateral moraine overlaps with the MP I moraine age, supporting the CE ages based on ε_{min} and a minimal correction for snow cover.

A third landform in the Lime Hills area, hummocky ground moraine near North Lime Lake, was sampled for CE dating. In contrast to the sites just discussed, the moraine at North Lime Lake is heavily forested, and the two sampled boulders, both 2–3 m high, were covered by several centimeters of moss in places. One boulder yielded an uncorrected ^{10}Be age of 25.6 ± 4.5 ka, and the other yielded an uncorrected ^{10}Be/^{26}Al average age of 36.5 ± 2.3 ka (Table DR1 [see footnote 1]; Fig. 4). Taking into account boulder-surface erosion, the ages span ca. 26.1 ± 4.6 and 37.6 ± 2.5 ka (ε_{min}) to 31.2 ± 5.4 and 49.8 ± 2.2 ka (ε_{max}). In this forested setting, low-lying moraines are less likely to be wind swept; thus, snow cover may have been significant. Because we analyzed only two boulders, and their ages are >10 k.y. apart, assigning this moraine to a Farewell I versus Farewell II age is precluded.

Kokrines Hills

The Kokrines Hills comprise the rolling uplands north of the Yukon River in west-interior Alaska (Fig. 1). Roughly a dozen small glaciers occupied north-facing cirques and valleys in the highest sectors of the Kokrines Hills during the late Pleistocene. Prior to the late Pleistocene, during the maximum phase of Quaternary ice cover, the hills were occupied by larger centers of ice (Kaufman and Manley, 2004). Late Pleistocene glaciers in the region have been compiled at the regional scale only (e.g., Coulter et al., 1965); no detailed glacial mapping has taken place in the Kokrines Hills to our knowledge.

Previous Age Control

There is no age control for the glacial deposits in the Kokrines Hills. However, the twofold moraine sequence easily recognized in air photographs of many valleys is suggestive of an early and a late Wisconsin moraine, similar to that observed in better-studied ranges surrounding the Yukon-Koyukuk region and elsewhere around the state (Hamilton, 1994). Our field-based impression of the relative degree of moraine weathering in comparison with other moraines that we have studied from around the state also suggests that the outer and inner valley moraines are early and late Wisconsin, respectively.

Cosmogenic Exposure Ages

We visited a north-facing valley in the south-central part of the Kokrines Hills and sampled four of the most suitable boulders from each of two moraines located 1.5 and 5.0 km downvalley from the cirque headwall (Fig. 2). Large feldspar phenocrysts in the quartz monzonite boulders seem to give rise to relatively high rates of boulder-surface erosion. This is evidenced by resistant veins that stand ~2 cm in relief and weathering pits 2–3 cm deep on boulders of both moraines. We estimated that the rate of bedrock erosion at this site (see above) is <8.2 ± 0.8 mm k.y.$^{-1}$, based on ^{10}Be concentrations in tors from the unglaciated upland interfluves.

Three 1–2-m-high boulders on the outer moraine have ε_{min} ^{10}Be ages of 54.0 ± 1.8, 24.4 ± 1.1, and 19.1 ± 1.2 ka, and ε_{max} ages of 86.8 ± 2.8, 28.7 ± 1.3, and 21.6 ± 1.3 ka (Table DR1 [see footnote 1]; Fig. 4; the fourth sample contained insufficient quartz). Given these ages, it is difficult to place this moraine unequivocally into the early or late Wisconsin. Either the oldest boulder is closer to the moraine age and the younger two were more recently exhumed, or the oldest boulder contains inherited isotopes and the younger two ages are closer to the timing of initial moraine stabilization.

Four ≥2-m-high boulders from the inner moraine yielded ε_{min} ^{10}Be ages of 30.8 ± 1.5, 24.2 ± 1.0, 23.7 ± 1.7, and 21.4 ± 1.3 ka, and ε_{max} ages of 38.2 ± 1.8, 28.4 ± 1.2, 27.8 ± 2.0, and 24.7 ± 1.5 ka (Table DR1 [see footnote 1]; Fig. 4). The relatively tight cluster of the younger three ages suggests that the oldest age of this series is influenced by isotopic inheritance. Averaging the younger three ages indicates that the inner moraine stabilized ca. 23 (ε_{min}) to 27 ka (ε_{max}) or at ca. 24–28 ka using the "oldest-age method." Considering again the ages of boulders from the outer moraine, it too included one boulder with an age in the 24–28 ka range, which might support the late Wisconsin age interpretation.

Yukon-Tanana Upland

The Yukon-Tanana upland (YTU; Fig. 1) encompasses a hilly landscape in east-interior Alaska between the Yukon and Tanana river valleys. During peak glacial intervals, the YTU contained dozens of centers of ice, ranging between independent massifs with a few glaciers to larger centers of ice with dozens of valley glaciers (Weber, 1986). With its radially oriented glaciated valleys and widely spaced massifs, the YTU preserves a complicated moraine record that records an extensive glacial history. The glacial features in the area have been mapped in detail (e.g., Weber and Hamilton, 1984; Weber, 1986). In some valleys, moraine sequences represent four periods of ice expansion; most late Wisconsin moraines were formed by cirque and small valley glaciers (Weber, 1986). From old to young, glacial episodes in the YTU are termed the Charlie River, Mount Harper, Eagle, Salcha, and Ramshorn, of which the youngest four have a moraine record (Weber, 1986).

Previous Age Control

There is little age control for the moraines in the YTU. Their subdivision into glacial episodes is based on relative-weathering criteria and correlations with neighboring glaciated regions. The four youngest moraines were assigned by Weber (1986) to middle Pleistocene (Mount Harper), early Wisconsin (Eagle), late Wisconsin (Salcha), and late Holocene (Ramshorn) glaciations. The Salcha and Eagle moraines have been correlated with the last and penultimate moraines, respectively, in neighboring areas. For example, the Eagle and Salcha glacial episodes have long been correlated with the Delta and Donally glaciations, respectively, of the Alaska Range to the south, and with the Reid and McConnell glaciations, respectively, in the Yukon Territory (Weber, 1986; Thorson, 1986; Hamilton, 1994).

Until recently, the Eagle, Delta, Healy (in the Nenana River valley), and Reid glaciations were all placed into the early Wisconsin (e.g., Porter et al., 1983). Berger et al. (1996) suggested that the Reid glaciation was older than the Sheep Creek tephra and for the first time firmly dated the tephra to ca. 190 ka, an age later supported by Westgate et al. (2001). And Begét and Keskinen (2003) improved the tephro-stratigraphic context of deposits of the Delta glaciation and showed that the glaciation is coeval with MIS 6 because it predates the Old Crow tephra and postdates the Sheep Creek tephra (Begét, 2001; Begét and Keskinen, 2003). Thus, the penultimate moraines in the Yukon and the Delta River drainage of the northern Alaska Range are MIS 8 and MIS 6, respectively. This calls into question the age assignment of the Eagle moraines in the YTU.

Cosmogenic Exposure Ages

We visited the type area in the YTU, the Ramshorn Creek valley, and sampled a fourfold moraine sequence that represents the Mount Harper, Eagle, Salcha, and Ramshorn glacial episodes (Fig. 2). We chose four of the most suitable granitic boulders from each moraine for CE dating; the samples were analyzed using a mix of ^{10}Be ages, ^{26}Al ages, and ^{10}Be/^{26}Al average ages (Table DR1 [see footnote 1]). The longest CE chronology in Alaska is from the YTU, representing three glacial episodes, with CE ages ranging from ca. 16 to ca. 120 ka.

The oldest CE ages in Alaska to the authors' knowledge come from the Mt. Harper moraine, ~20 km downvalley from the cirque headwall (Fig. 2). Four uncorrected CE ages range between 61.8 ± 2.2 and 119.2 ± 3.9 ka (Table DR1 [see footnote 1]; Fig. 4; the latter age, a minimum, can also be interpreted in terms of a maximum bedrock erosion rate of 4.8 mm k.y.$^{-1}$ and provides a local ε_{max} for the YTU boulders, Table DR2 [see footnote 1]). The ages from this moraine are more scattered than from any other single landform dated in Alaska. They average 95.5 ± 28.7 (ε_{min}) and >125.9 ± 41.2 ka ($\varepsilon_{max\text{-}YTU}$; excluding the oldest age, which is used to calculate $\varepsilon_{max\text{-}YTU}$). The "oldest-age method" suggests that the Mt. Harper glacial episode occurred during MIS 6, although an even older age cannot be precluded.

The age of the Eagle moraine, located ~15 km downvalley from the cirque (Fig. 2), is presently controversial. Four CE ages uncorrected for erosion range between 25.1 ± 1.0 and 61.4 ± 3.9 ka, with no apparent age cluster (Table DR1 [see footnote 1]; Fig. 4). Average boulder ages range between 44.4 ± 16.4 (ε_{min}) and 53.8 ± 23.6 ka ($\varepsilon_{max\text{-}YTU}$). The moraine is clearly pre–late Wisconsin in age, as revealed by the antiquity of the boulder ages combined with their scatter, typical of older moraines. The "oldest-age method" may provide the best estimate for the age of the Eagle moraine because there is no obvious old outlier that ranges between ca. 65 (ε_{min}) and ca. 84 ka ($\varepsilon_{max\text{-}YTU}$), and suggests that the Eagle glacial episode is early Wisconsin (cf. Weber, 1986).

The Salcha moraine lies ~6 km downvalley from its cirque headwall (Fig. 2) and was assigned a late Wisconsin age by Weber (1986). The four dated boulders from the Salcha moraine yield uncorrected CE ages ranging between 16.4 ± 1.1 and 28.7 ± 1.3 ka (Table DR1 [see footnote 1]; Fig. 4). Accounting for boulder-surface erosion, the CE ages average 21.2 ± 5.9 (ε_{min}) and 22.8 ± 6.9 ka ($\varepsilon_{max\text{-}YTU}$). The oldest Salcha moraine boulder is statistically older than the mean of the others and thus appears to contain inherited isotopes. The "oldest-age method" for the Salcha moraine (excluding the old outlier) results in ages between ca. 21 (ε_{min}) and ca. 23 ka ($\varepsilon_{max\text{-}YTU}$). The CE ages support Weber's (1986) assignment of the Salcha to the late Wisconsin.

The Ramshorn moraine is only 1.5 km downvalley from the cirque headwall (Fig. 2) and was previously assigned a late Holocene age (Weber, 1986). Four boulders yield uncorrected CE ages that range from ca. 16 to 28.4 ± 1.1 ka (Table DR1 [see footnote 1]; Fig. 4). The Ramshorn moraine boulders average 22.3 ± 6.3 (ε_{min}) and 24.3 ± 7.3 ka ($\varepsilon_{max\text{-}YTU}$). The "oldest-age method" is difficult to apply to the Ramshorn CE ages because of their bimodality (ca. 26–29 and 16–18 ka; ε_{min}), which may be due to the proximity of the moraine to the cirque headwall (see discussion). Because this moraine is morphostratigraphically younger than the Salcha moraine, the older mode of ages almost certainly arises from inheritance. The "oldest-age method" for the Ramshorn moraine (excluding the boulders with suspected inheritance) suggests the age lies between ca. 18 (ε_{min}) to ca. 19 ka ($\varepsilon_{max\text{-}YTU}$), clearly within the late Wisconsin.

Northeastern Brooks Range

The Brooks Range (Fig. 1), trending east-west across northern Alaska, is the largest center for Quaternary glaciation in Alaska outside of the Cordilleran Ice Sheet. Glaciers expanded to the north and south from the central crest and were mostly composed of long, complex, and interconnected valley glaciers. The glacial geology of the Brooks Range has received considerable research attention over most of its western and central valleys (e.g., Detterman et al., 1958; Hamilton and Porter, 1975; Hamilton, 1982; Hamilton, 2001). However, little research attention has been paid to the Pleistocene glacial geology of the northeastern sector of the range, until the recent work of Balasio et al. (2005).

Previous Age Control

Until the present study, no Pleistocene glacial chronology existed for the northeastern Brooks Range. In the central Brooks Range, late Pleistocene moraines were deposited during the Itkillik glaciation (Hamilton, 1994). Glaciers expanded up to 40 km north of the northern range front during the Itkillik I phase and up to 25 km north of the range front during the Itkillik II phase (Hamilton, 1982). Drift of the Itkillik I phase is considered to be early Wisconsin in age based on relative-weathering criteria, infinite ^{14}C ages, and its presence above the 140 ka Old Crow Tephra (Hamilton, 1986b, 2003). The Itkillik II phase is firmly placed into the late Wisconsin based on numerous ^{14}C ages that are stratigraphically tied to Itkillik II outwash downvalley of associated moraines, which constrain advances and retreats of central Brooks Range glaciers (Hamilton, 1082). Glaciers reached their maximum extent of the Itkillik II phase after ca. 27 ka and first retreated ca. 25.5 ka (Hamilton, 1982 [Radiocarbon ages were converted to calendar ka using a graphical interpretation of Fig. 1A *in* Kitagawa and van der Plicht (1998)]). A subsequent advance, almost as extensive as the first, occurred after ca. 23 ka; glaciers then retreated between ca. 22 and ca. 18.5 ka (Hamilton, 1982). Itkillik II glaciers at last receded following a final readvance during the late glacial (Hamilton, 1982).

Cosmogenic Exposure Ages

We investigated moraines in the Jago and Okpilak River valleys in the northeastern Brooks Range (Fig. 2) and sampled granitic boulders from Itkillik-age moraines for CE dating. New glacial mapping, relative-weathering data, and 16 ^{10}Be ages are reported in Balascio et al. (2005). Here we review the ages in the context of our chronologies from elsewhere in Alaska and compare them to Hamilton's (1982) ^{14}C chronology.

Three moraines were sampled near the range front along the Jago River valley and one moraine at the range front along the Okpilak River valley (Fig. 2). The northern range front here forms an abrupt transition from high, glaciated peaks to low-relief Arctic coastal plain with a narrow (<5 km) zone of foothills. High (>200 m above river level), massive lateral ridges emanate from the range front on either side of the Jago and Okpilak river valleys and presumably are composite moraines formed over numerous glacial cycles. We sampled boulders from the tops of the left lateral ridge in both the Jago and Okpilak areas. Three boulders atop the Okpilak River valley moraine have uncorrected ^{10}Be ages of 27.3 ± 0.9, 50.0 ± 1.3, and 76.0 ± 1.9 ka (Table DR1 [see footnote 1]; Fig. 4). Five boulders atop the Jago River valley lateral moraine have uncorrected ^{10}Be ages that cluster between 20.1 ± 0.7 and 26.3 ± 0.7 ka and include an older age of 55.7 ± 1.4 ka.

The cluster of ages from the Jago lateral moraine averages 23.7 ± 3.0 (ε_{min}) and 27.8 ± 4.1 ka (ε_{max}; ε_{max} is locally substantiated with the oldest sample from the Okpilak ridge, which provides a maximum bedrock erosion rate of 7.8 mm k.y.$^{-1}$). The "oldest-age method" suggests a moraine age between ca. 27 (ε_{min}) and ca. 32 ka (ε_{max}). Because the climate of this part of Alaska is more similar to Arctic Canada (source of ε_{min}) than other sites discussed thus far, we favor ε_{min} as the more appropriate boulder-surface erosion rate. The ca. 56 ka age on the Jago lateral moraine appears to be a clear outlier that we attribute to inheritance. The age of the Okpilak ridge is uncertain; correlation with the Jago ridge supports the suggestion that the two older boulders from the Okpilak ridge contain inherited isotopes. Alternatively, we sampled different age deposits in the different valleys, considering that the large ridges were clearly formed through successive glacial cycles. Alternatively, both ridges might be pre–late Wisconsin in age, and the young age cluster on the Jago ridge records accelerated moraine degradation and consequent boulder exhumation during the late Wisconsin. On the other hand, the stabilization age indicated by the ca. 24 ka (ε_{min}) average age or the ca. 27 ka (ε_{min}) age by the "oldest-age method" is consistent with Hamilton's (1982) age constraints for deglaciation from the Itkillik II terminal moraine at ca. 25 ka.

Four ^{10}Be ages were also obtained from a range-front end moraine within the valley bottom (stratigraphically younger), below the left-lateral ridge site along the Jago River valley (Fig. 4). Uncorrected ^{10}Be ages exhibit unexpected scatter, ranging between 9.2 ± 0.4 and 42.3 ± 1.1 ka (Table DR1 [see footnote 1]). If indeed Itkillik II ice reached the lateral ridge tops at the range front along the Jago River valley, then this moraine is also of Itkillik II age, and the scatter reveals both inheritance and exhumation issues.

The fourth moraine sampled in the northeastern Brooks Range is also in the Jago River valley, ~8 km upvalley from the range front. The moraine was deposited by ice that readvanced onto the Jago River valley floor from the Hubley Creek tributary valley, following the evacuation of ice from the trunk valley (i.e., a readvance). Four uncorrected ^{10}Be ages range from 15.6 ± 0.5 to 21.6 ± 0.4 ka (Table DR1 [see footnote 1]; Fig. 4). The CE ages average 19.4 ± 2.8 (ε_{min}) and 22.0 ± 3.6 ka (ε_{max}), and the "oldest-age method" suggests a moraine stabilization age of ca. 22 (ε_{min}) to ca. 26 ka (ε_{max}). Taking the ε_{min} values, deglaciation from the Hubley moraine (ca. 19 ka [average] to ca. 22 ka [oldest-age method]) overlaps with deglaciation from the second phase of Itkillik II glaciation in the central Brooks Range (ca. 18.5 to ca. 21 ka; Hamilton, 1982).

DISCUSSION

In total, we have obtained 114 CE ages on 98 Alaskan moraine boulders from 23 moraines. Given the size of this data set and the variety of physical settings that have been sampled, we have gained new insight into the use of CE dating on Alaskan moraine boulders. Our investigations include examples of both failures and successes at establishing the ages of moraines using CE dating. In addition, the few cases where moraine ages are constrained independently provide clues as to the most reliable interpretation of CE age distributions.

Issues in CE Dating of Moraine Stabilization

The major potential complications in CE dating of moraine boulders include moraine degradation and boulder exhumation, inheritance, boulder-surface erosion, and snow cover. All the CE ages from the independently dated Gusty Lakes moraine in the Ahklun Mountains are younger than the timing of moraine abandonment, suggesting that in this case, CE ages closely date moraine stabilization and that individual CE ages represent subsequent boulder exhumation. This has been found and predicted previously (e.g., Zreda et al., 1994; Hallet and Putkonen, 1994; Phillips et al., 1996; Phillips et al., 1997; Putkonen and Swanson, 2003) and has led us to prefer the "oldest-age method" for determining the timing of ice retreat from a moraine position, which is especially suitable where there is no evidence of inheritance. In the case of the Gusty Lakes example, the "oldest-age method" suggests an age that is still ~1000 yr younger than the independently established timing of moraine abandonment, which could indicate that the oldest boulders were not sampled. The use of the "oldest-age method" is further corroborated by the CE age distribution on the Mount Waskey moraine.

In some cases, however, individual CE ages are too few or too scattered to confidently constrain the timing of moraine stabilization. The best example is the end moraine on the Jago River valley floor at the range front in the northeastern Brooks Range, where the CE ages are spaced roughly evenly between ca. 9 and ca. 45 ka. Another example is the drift near North Lime Lake in the Lime Hills region, where two samples are ca. 25 and ca. 38 ka.

Inheritance can also compromise a cluster of CE ages on moraine boulders. Identifying boulders with inheritance is not always straightforward, although its effect can often be identified as an old outlier. For terminal moraines, this is probably a reasonable criterion, as bedrock sources for boulders with prior exposure would likely have been exposed for relatively long periods of time (>10 k.y.). However, in the case of recessional or readvance moraines, shorter intervals of prior exposure might make a CE age influenced by inheritance less obvious. For example, the Mount Waskey moraine in the Ahklun Mountains probably stabilized between 11.0 and 12.4 ka, but two boulders yielded older ages of ca. 15 ka, which is also glacial-geologically meaningful as the time of regional deglaciation (e.g., Briner et al., 2001; Manley et al., 2001). In a second example, the suite of boulders from the youngest (Ramshorn) moraine in the Ramshorn Creek valley in the YTU did not contain statistical outliers but had an average age older than the late Wisconsin terminal moraine slightly downvalley. The almost certain effect of inheritance in boulders on the Ramshorn moraine and other moraines outside of the YTU suggests the possibility that inheritance is related to the proximity of the moraine to the cirque headwall. Debris that travels a shorter distance appears to have a higher chance of containing inherited radionuclides. This relationship was also found at the Cloud Lake cirque moraine in the Ahklun Mountains, where we sampled a boulder with inheritance, whereas boulders from the moraine farther downvalley did not contain inheritance. Similarly, we sampled a boulder with inheritance from the Chilly Valley moraine, ~2 km from its cirque headwall, but not from the nearby Gusty Lakes moraine, which was formed by ice that traveled tens of km. Finally, boulders on the Kokrines Hills moraines appear to have inheritance, and they were deposited by small valley glaciers only a few kilometers downvalley from their headwall.

Boulder-surface (bedrock) erosion and snow cover are additional sources of uncertainty. We attempted to account for boulder-surface erosion by estimating local bedrock erosion rates using tors in interior Alaska and the oldest moraine boulders. While constraints on bedrock erosion can be made, the actual rate for each individual moraine chronology is a function of many variables including boulder lithology and local climate. If the late Wisconsin terminal moraine in the Lime Hills region is the same age as the independently dated moraine in Denali Park, then the influence of snow cover and boulder erosion are minor. Similarly, if the CE age clusters accurately represent moraine age in the northeastern Brooks Range and these moraines correspond to the ^{14}C chronology from the central Brooks Range, then there is little room for erosion to be greater than ε_{min} or for snow cover to be significant.

The Age of the Penultimate Glaciation

While the CE-age data set provides insights into potential complications, in many cases it also provides reliable chronological information. The data show that the penultimate drift corresponds with the early Wisconsin (cf. Hamilton, 1994) rather than MIS 6 or earlier. In every location where we dated penultimate moraines, the CE ages fall between ca. 40 and ca. 75 ka (Fig. 5). The complicating factors discussed above cannot account for the >60 k.y. age difference required to place boulders from penultimate moraines within MIS 6. Robust evidence has recently emerged that the penultimate glaciation is MIS 6 age in the northernmost Alaska Range (Begét and Keskinen, 2003) and MIS 8 in the Yukon Territory (Westgate et al., 2001; Huscroft et al., 2004). This contrasts with our results that indicate early Wisconsin ages for penultimate

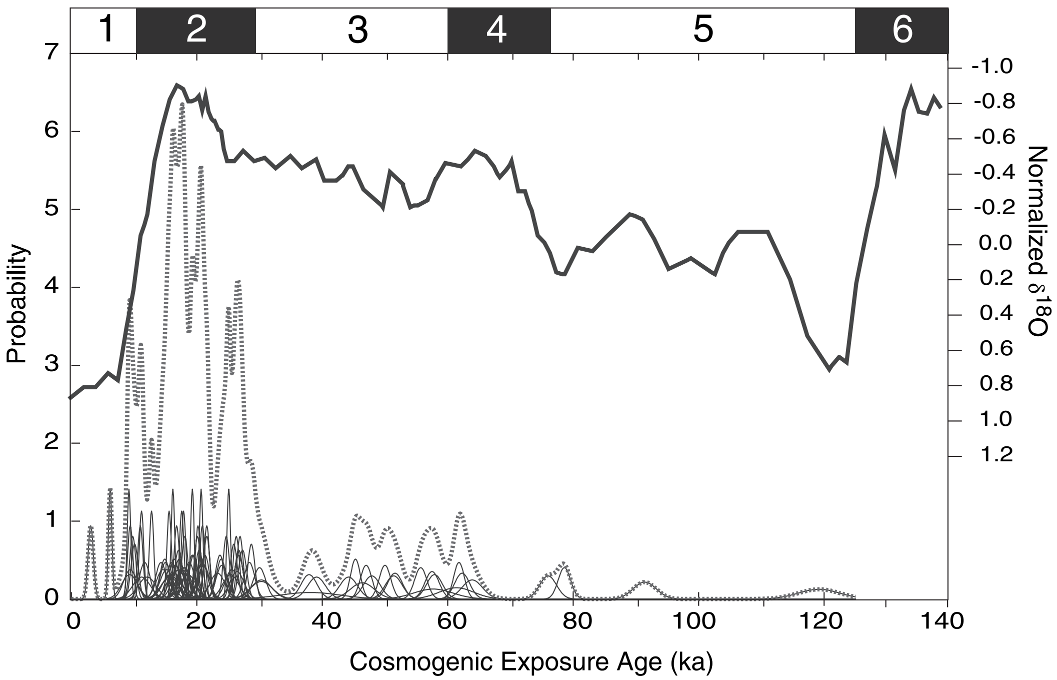

Figure 5. Probability distributions for the 95 moraine boulder cosmogenic ages from Alaska discussed in this paper compared to global ice volume (and marine isotope stages) from Martinson et al. (1987; bold line at top). Summed probability shown as the dotted line above the 95 individual probability distributions. The distribution shows an early Wisconsin age cluster (65–40 ka), and a trimodal late Wisconsin age cluster that represents three discrete intervals of moraine stabilization across the state.

moraines as nearby as the YTU. This apparent discrepancy should be pursued further.

Glaciers expanded farther during the early Wisconsin than during the late Wisconsin across Alaska, indicating that climate conditions were more favorable for glaciation during late MIS 4/early MIS 3 than during MIS 2. Global ice volume was higher during MIS 2 than MIS 4/3 (Martinson et al., 1987), implying that ice-volume fluctuations in Alaska were asynchronous with global-ice volume. This likely arose from a combination of temperature depression with moisture availability near the MIS 4/3 boundary that led to relatively extensive glaciers. On the other hand, peak global-ice volume during the late Wisconsin resulted in a sea level low enough to move the moisture source for Alaskan glaciers ~500 km away (e.g., Brigham-Grette, 2001).

Late Wisconsin Glaciation

The CE ages constrain the timing of late Wisconsin deglaciation in six regions that span Alaska from southwest to northeast. A compilation of all uncorrected CE ages shows three strong modes in the distribution of ages (Fig. 5). The modes imply that deglaciation from terminal moraines occurred in two discrete intervals, at ca. 24–26 and ca. 17–21 ka. The third, youngest mode is dominated by ages from the late glacial moraine in the Ahklun Mountains (Briner et al., 2002).

The primary mode of boulder ages at ca. 17–21 ka includes samples from the Ahklun Mountains, the Lime Hills region, and the YTU. The CE ages from the younger portion of this mode come from the Chuilnuk Mountains. The CE ages from the Brooks Range and Kokrines Hills comprise the older, secondary mode at ca. 24–26 ka, where glaciers first receded during this earlier time (cf. Hamilton, 1982). Three CE ages in this mode are from boulders with inheritance from the YTU. The time when glaciers reached their terminal late Wisconsin positions, where it has been established by radiocarbon dating, also varies from place to place, ranging from ca. 27 ka in the Brooks Range (Hamilton, 1982) to ca. 21 ka in Denali Park (Ten Brink and Waythomas, 1985) to ca. 24 ka in the Ahklun Mountains (Kaufman and Manley, 2004).

Glaciers across the state advanced to and retreated from their terminal late Wisconsin positions at different times. The difference in ages appears to exceed the uncertainties in radiocarbon and CE dating, as is supported where the CE ages can be compared with independently dated moraines. Considering the complexity and vastness of Alaskan physiography, the variety of its glacial systems (ranging from small cirque glaciers to major ice cap outlets), and its position in relation to prominent centers of atmospheric circulation features, we expect some heterogeneity to the spatio-temporal pattern of climate change and glacier response to changing climate. One commonality among all deglacial records is the recession during or slightly before the peak of the last global glacial maximum. This is most likely related to limited moisture availability at that time as lowered sea level exposed vast areas of shallow continental shelf and sea ice capped the adjacent seas.

Comparison with Other CE Studies in Beringia

Despite the emergence of CE dating as a popular tool for reconstructing glacier chronologies worldwide, few other studies have taken place in Beringia. Gualtieri et al. (2000) report 16 ^{36}Cl ages, and Brigham-Grette et al. (2003) report 12 ^{36}Cl ages from two mountain ranges in northeastern Siberia. Similar to the results from many of the moraines we studied, the Siberian data are scattered, best providing age constraints at stadial-level time scales. For example, four boulders from the Sartan (late Wisconsin) drift in the Pekulney Mountains yielded two ca. 16 ka ages, with an old outlier at ca. 24 ka and a young outlier at ca. 9 ka (Brigham-Grette et al., 2003), and 14 ages from Sartan drift in the Koryak Mountains range between ca. 10 and ca. 22 ka (Gualtieri et al., 2000). Four boulder and four glacially eroded bedrock samples associated with the Zyryan (pre–late Wisconsin) limit in the Pekulney Mountains provide a rare opportunity to compare these two sample types in Beringia. At one site, ages on sculpted bedrock range between ca. 56 and ca. 69 ka, whereas Zyryan moraine boulder ages range between ca. 37 and ca. 42 ka (Brigham-Grette et al., 2003). Assuming that any prior accumulation of cosmogenic isotopes in the bedrock samples was removed and that erosion rates are similar for these samples, the younger ages of the boulders may indicate that they stabilized sometime after ice retreat from the Zyryan moraine. Thus, the CE ages from Alaskan moraines, combined with the northeastern Siberian CE ages, support the interpretation that CE ages date moraine stabilization, which occurs sometime following deglaciation.

Implications

These results provide guidelines for interpreting a suite of CE ages on moraine boulders. The timing of a glaciation is often determined as the mean of a suite of moraine boulder CE ages. Implicit in this approach is the assumption that moraines form as soon as glaciers reach their maximum extent and that the moraine boulders stabilize before glaciers retreat. While these assumptions may be valid in some settings, we have found that glaciers can persist at their terminal margin for several thousand years (e.g., the Gusty Lakes moraine in the Ahklun Mountains), and that moraines stabilize following ice retreat. Our results suggest that the distribution of CE ages from multiple boulders on a single moraine represent the process of moraine stabilization (cf. Hallet and Putkonen, 1994; Phillips et al., 1996; Putkonen and Swanson, 2003), and that they are best interpreted as minimum-limiting ages on the timing of deglaciation, rather than the timing of glacier *advance* per se.

Interpreting CE ages as minimum ages of glacier retreat has been adopted in other settings, especially for penultimate moraines. For example, dozens of CE ages on Bull Lake deposits at and near its type locality in the Wind River Range, Wyoming, which are presently interpreted as MIS 6 in age (Sharp et al., 2003; Pierce, 2004), have yielded CE ages that range from ca. 60 to ca. 150 ka, with the majority between ca. 60 and ca. 100 ka (Phillips et al., 1997). The unexpectedly young CE ages were not attributed to boulder-surface erosion or snow cover but rather to moraine degradation (Phillips et al., 1997). The next oldest moraine segment (Sacagawea Ridge) was dated as well and consistently yielded the oldest ages in the data set (Phillips et al., 1997). Overall, the Wind River Range data, along with our data, particularly from the YTU, suggest that while scattered, CE ages on pre–late Wisconsin moraines generally provide enough age information to distinguish moraine ages between MIS 4 and MIS 6 or older.

CONCLUSIONS

Our CE-age data set from 23 Alaskan moraines provides an opportunity to explore factors that complicate the use of CE dating on Pleistocene moraines. The data set can be explored for internal consistencies, and it can be compared in a few locations with independently dated moraines. Interpreting the age of moraine deposition from a suite of CE ages on moraine boulders is often difficult. The ages seem to most closely reflect the timing of moraine stabilization, which is a minimum constraint on ice recession. CE ages generally cluster more tightly on late Wisconsin moraines than on penultimate moraines, indicating that moraine degradation processes significantly scatter CE ages by boulder exhumation. However, even on some late Wisconsin moraines, the scatter of CE ages is significant. More samples are needed from penultimate moraines to achieve a comparable level of certainty as for late Wisconsin moraines. In addition, moraines proximal to cirque headwalls appear to contain more boulders with inherited radionuclides than moraines farther downvalley. Finally, depending on how individual CE ages from a single moraine are distributed, the "oldest-age method" may provide the closest constraint on moraine stabilization.

The CE ages from all three penultimate moraines that we sampled are early Wisconsin in age (~MIS 4), revealing strong paleoclimatic forcing for glacier growth prior to the late Wisconsin but after the last interglacial period. Older ages for the penultimate glaciation, as reported for regions neighboring the YTU, suggest that glacier response was spatially variable. Late Wisconsin glaciers appear to have advanced to and retreated from their late Wisconsin terminal moraines at slightly different times across the state. Local maxima may have occurred earliest in northern Alaska and latest in southwestern Alaska. Glaciers were by and large receding from their late Wisconsin terminal moraines around the last global glacial maximum, probably responding to lowered sea level and limited moisture availability.

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