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Journal:	*Journal of Quaternary Science*
Manuscript ID:	draft
Wiley - Manuscript type:	Review Article
Date Submitted by the Author:	n/a
Complete List of Authors:	Briner, Jason; University at Buffalo, Geology Kaufman, Darrell; Northern Arizona University, Geology
Keywords:	Alaska, glaciation, late Pleistocene, chronology, alpine glacier

Late Pleistocene mountain glaciation in Alaska: Key chronologies

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ABSTRACT: Moraine sequences of mountain glaciers can be used to infer spatial and temporal patterns of climate change across the globe. Alaska is the most accessible high-latitude location in the Northern Hemisphere and contains a rich record of alpine glaciation. Here, we highlight the key chronologies from three mountain ranges in Alaska that reveal the timing and spatial extent of late Pleistocene glaciation. The most extensive glacier advance of the last glaciation occurred prior to the last global glacial maximum. Cosmogenic exposure ages from moraine boulders in three sites spanning 800 km indicate that this penultimate advance most likely occurred during marine isotope stage (MIS) 4 or early MIS 3. During MIS 2, more limited glacier expansion generated multiple moraines spanning from prior to the global last glacial maximum through the late glacial period. Glaciers retreated from their terminal positions ~27 to 25 ka in arctic Alaska and ~22 to19 ka in southern Alaska. Moraines in at least two ranges date to 12 to 11 ka, indicating a glacial advance during the Younger Dryas period. Reconstructed equilibrium-line altitudes of both penultimate and MIS 2 glaciers were lowered only 300-600 m, much less than elsewhere in the Americas. Alaska is documented to have been more arid during MIS 2, perhaps due in large part to the exposure of the Bering-Chukchi platform during eustatic sea-level lowering. The restricted ice extent is also consistent with the output of climate models that simulate a lack of significant summer cooling.

KEYWORDS: Alaska, glaciation, late Pleistocene, chronology, mountain glacier

Introduction

Alaska is often characterized as a land of extremes, and the same applies to its glacial geology. The state presently hosts the largest valley glaciers in North America, yet during the Pleistocene, it encompassed the largest unglaciated expanse on the continent. Presently (c. 1970), glaciers cover about 75,000 km^2 of the state and are distributed among 14 centers of glacerization (Molnia, 2007). During the last global glacial maximum, the area of glacier cover expanded by ten times, to 727,800 km^2 (Kaufman and Manley, 2004), and encompassed several lower-elevation massifs that are presently unglaciated. The vast majority of this expansion involved glaciers that surround the Gulf of Alaska. This amalgamation of coalescent ice caps and piedmont lobes formed the northwestern extension of the Cordilleran Ice Sheet (Hamilton and Thorson, 1983). Like their modern counterparts, these glaciers benefited from a proximal source of moisture, a persistent atmospheric circulation pattern that drove moist air inland, and adiabatic cooling associated with the extraordinary mountainous terrain. In contrast, the interior part of the state was never extensively glaciated. The Cordilleran ice formed an effective barrier to moisture derived from the Gulf of Alaska. And, prevailing southwesterly winds dried as sea ice expanded and global sea level lowered, exposing the Bering-Chukchi platform. The only significant centers of glacier growth beyond the Cordilleran Ice Sheet were the Brooks Range in arctic Alaska and the Ahklun Mountains in the southwest.

Because most of Alaska was never glaciated, mountain glaciers expanded onto piedmonts where they left moraines dating to multiple glaciations. The ages of these moraines are known from a few places where they have been correlated with radiometric ages on organic matter or volcanic products interbedded with outwash (Hamilton, 1994). With the advent of cosmogenic exposure dating, direct ages on moraine stabilization have recently been obtained from several mountain ranges in Alaska (Briner et al., 2005). The growing database of tephra marker beds has further refined the ages of glacier deposits (Begét and Keskinen, 2003).

In this paper, we summarize the key late Pleistocene mountain glacier chronologies currently available in Alaska. This is the first detailed review of mountain glacier chronology in Alaska since Hamilton's (1994). It benefits from a recent compilation of late Wisconsin state-wide glacier extents (Kaufman and Manley, 2004) and a recent summary of Quaternary alpine glaciation in Alaska (Kaufman et al., 2004). The most complete and robust chronologies are from the Brooks Range (northern Alaska), the Alaska Range (central Alaska), and the Ahklun Mountains (southwestern Alaska). Late Wisconsin moraines are well dated in other parts of Alaska, for example on the Alaska Peninsula (Mann and Peteet, 1994; Stilwell and Kaufman, 1996) and the Kenai Peninsula (Reger and Pinney, 1996). Here, we focus on the sequences that include moraines deposited during both the late Wisconsin and the penultimate glaciations so the relative extent of glaciers through the late Pleistocene can be assessed.

The ages of late Pleistocene glacial features are primarily based on either cosmogenic exposure dating (mostly using ^{10}Be) or ^{14}C. Cosmogenic exposure ages from surface boulders on moraines date the glacier retreat and subsequent stabilization of the landform. Briner et al. (2005) discuss alternative interpretations of clusters of cosmogenic exposure ages from moraine boulders in Alaska and concluded that the oldest ages in a cluster generally yielded the best agreement with independent age information where available. Because this method relies heavily on just the single oldest age (excluding obvious outliers with inheritance; e.g., those that are $>2\sigma$ from the average of the others), Briner et al. (2005) reported moraine ages as the range between the oldest age and the average age (excluding outliers). All cosmogenic exposure ages reported here are also presented in this way. The uncertainty listed following the average age represents the 1σ variability among boulders. Additional uncertainties result from shielding effects related to snow cover and rock-surface erosion rates. All cosmogenic exposure ages reported here are unmodified from their original publications, and in all cases are based on the isotope production rates. Although there are differences in other calculations, such as altitude scaling, shielding and erosion effects, these should be relatively minor (<10% of the age). In contrast to exposure ages, ^{14}C ages generally bracket the timing of glacier fluctuations and must be interpreted in context of the morphostratigraphic position of the sample. All ^{14}C ages have been calibrated to calendar years using CALIB (v5) (Stuiver and Reimer, 1993) and are reported in cal ka BP (hereafter "ka"). Most ages are rounded or should be considered approximate at the millennium scale, even where this is not stated explicitly.

Brooks Range

The Brooks Range (Fig. 1) forms the northernmost drainage divide in northwest North America. It spans ~1000 km east to west across northern Alaska from the Alaska-Yukon border to the Chukchi Sea. Summit elevations increase eastward, exceeding 2700 m asl in the northeast. The range encompasses hundreds of small, sub-polar valley glaciers sheltered behind the highest north-facing cirque headwalls (Calkin and Ellis, 1980). The Brooks Range is the largest center of 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 extensive suite of moraines in the Itkillik River area, central Brooks Range, serves as the reference locality for late Pleistocene glaciations of the Brooks Range (Fig. 2; Hamilton, 1986a). Moraines are subdivided into the Itkillik I (older) and Itkillik II (younger) advances (Hamilton and Porter, 1975; Fig. 3). 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). Recent detailed mapping in the Itkillik River area resulted in further subdivision of the glacial deposits (Hamilton, 2003). The Itkillik I glaciation was subdivided into two phases based on differences in postglacial modification of moraines. We refer to the moraines deposited during the Itkillik I glaciation as the "penultimate" moraines. The Itkillik II (late Wisconsin) glaciation was also subdivided into two primary phases, including a maximum advance and a later readvance. Each of these phases of the Itkillik II glaciation is represented by two distinct moraines in the Itkillik River area (Hamilton, 2003).

Two phases of the Itkillik I advance recognized in the central Brooks Range are older than non-finite ^{14}C ages of 53 ka, and are believed to be younger than the last interglacial maximum (marine isotope stage (MIS) 5e; Hamilton, 1994). In the Noatak basin of the western Brooks Range, two separate advances are younger than the 140 ka Old Crow tephra and older than 36-34 ka (Hamilton, 2001). There are no published luminescence or cosmogenic exposure ages on Itkillik I (penultimate) drift in the Brooks Range.

The subsequent Itkillik II glaciation in the Brooks Range (Fig. 2) is bracketed in both the central (Hamilton, 1982) and western Brooks Range (Hamilton, 2001) between 30 and 13 ka. Numerous ^{14}C ages have been reported from Itkillik II outwash in the Koyakuk River area on the south side of the range (Hamilton, 1982). The outwash has been correlated with moraines upvalley and thereby has been used to infer the timing and position of glacier fluctuations in the central Brooks Range. The maximum Itkillik II glaciation occurred between about 27 and 25 ka, and was followed by an advance almost as extensive as the first after 23 ka. Alluviation of outwash streams seems to have ceased 15 ka (Hamilton, 1982). In the north-central Brooks Range, where a detailed sequence of Itkillik I and II moraines has been mapped in the Itkillik River area (Fig. 2A), a readvance at the northern range front led to rapid alluviation of a landside-dammed valley from 15.1 to 13.3 ka (Hamilton, 2003). The broad troughs between the range front and the cirques contain a suite of end moraines, but they have yet to be dated.

The outer two ridges of a prominent nested-moraine sequence in the Jago River valley, northeastern Brooks Range (Fig. 2B) have been correlated with the Itkillik II glaciation, and have been dated with ^{10}Be on moraine boulders (Balascio et al., 2005a). The Itkillik II terminal moraine in the Jago River valley, which projects 12 km to the north of the range front, stabilized between 27 and 23.7 ± 3.0 ka. A prominent end moraine 8 km upvalley from the range front, which was deposited at the mouth of a tributary valley that contains the Hubly Glacier, stabilized between 22 and 19.4 ± 2.8 ka (Balascio et al., 2005a).

ELAs have been reconstructed for smaller, topographically constrained Itkillik II glaciers across the Brooks Range (Balascio et al., 2005b). ELAs rise from west to east at 1.4 m km^{-1}, and are highest in the northeastern sector of the range, where the highest summits presently support the largest glaciers in the range. The Itkillik II equilibrium line altitude (ELA) surface is generally parallel to the modern, and is about 250 m lower on average (Balascio et al., 2005b). ELAs for Itkillik I glaciers are difficult to reconstruct because most glacier ice was interconnected and divides demarking their source areas are poorly defined. During the Itkillik I glaciation, ice was tens of kilometers more extensive than during the Itkillik II. Considering the low-gradient of the valleys, however, ELAs were likely only a few tens of meters lower during the Ikillik I glaciation than Itkillik II.

Alaska Range

The Alaska Range (Fig. 4) was occupied by western extension of the Cordilleran Ice Sheet during the late Pleistocene. In some portions of the range, the ice comprised a series of interconnected ice fields. Along the west and northern flanks of the Alaska Range, ice formed smaller, independent valley glacier systems. Moraine sequences in valleys across the northern Alaska Range typically consist of at least two major drift units (early and late Wisconsin), each deposited during multiple phases (e.g., Ten Brink and Waythomas, 1985; Kline and Bundtzen, 1986; Thorson, 1986). Several valleys within the Alaska Range have a long history of glacial-geologic research and a local nomenclature of glacial deposits (Hamilton, 1994).

The age of the penultimate drift in the Alaska Range is best constrained in three localities. A moraine sequence deposited along the Delta River valley beyond the northern Alaska Range front (Fig. 4) constitutes the reference locality of the Donnelly (late Wisconsin) and Delta (penultimate) glaciations (Péwé, 1953; Fig. 3). An outwash terrace that grades to the Delta moraine is overlain by the Old Crow Tephra (140 ka), suggesting that it pre-dates the late Pleistocene (Begét and Keskinen, 2003). A more detailed moraine sequence in the Nenana River valley, north-central Alaska Range (Fig. 4; Wahrhaftig, 1958; Thorson, 1986) was the focus of a recent exposure-dating study. Dortch (2006) obtained nine ^{10}Be ages on boulders from landforms created during the Healy glaciation, which is thought to be the equivalent to the moraine deposited in the Delta River valley during the Delta glaciation (Fig. 3; Hamilton, 1994).

The Healy landforms, excluding one young outlier, range between 60 and 55.7 ± 3.7 ka (Dortch, 2006). At a third locality, in the Swift River valley of the western Alaska Range (Fig. 4), Briner et al. (2005) mapped a sequence of moraines and correlated them with the Farewell I (penultimate) and Farewell II (late Wisconsin) moraines in the nearby Farewell region (Fig. 3; Kline and Buntzen, 1986). The Farewell I equivalent moraine, dated by four ^{10}Be ages, stabilized between 58 and 52.5 ± 5.6 ka (Briner et al., 2005).

The most robust ^{14}C chronologies for late Wisconsin moraines in the Alaska Range come from Denali National Park, the Nenana River valley (Fig. 4), and a few additional valleys. In the McKinley (Denali National Park) and Nenana River valleys, a four-fold sequence of late Wisconsin moraines is well dated, and Porter et al. (1983) provide the most detailed review of the timing of late Wisconsin glacier fluctuations. Several maximum-limiting ^{14}C ages constrain the initial late Wisconsin advance to sometime after 27 ka (Hamilton, 1982; Porter et al., 1983). In Denali National Park, the late Wisconsin (McKinley Park (MP) I) terminal moraine was deposited 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).

Recent ^{10}Be exposure dating (Dortch, 2006) provides additional ages on the late Wisconsin moraines, including landforms of Riley Creek age in the lower Nenana River valley (Fig. 4) (equivalent to MP deposits in Denali National Park and Donnelly deposits in the Delta River valley; Fig. 3). Landforms of the Riley 1 (oldest) and Riley 2 glaciations produced a wide distribution of ^{10}Be ages, ranging between 61 and 8 ka. Deposits of the Carlo glaciation (youngest) produced a tighter cluster of ages between 19 and 17.2 ± 1.3 ka. Dortch (2006) also dated late Wisconsin landforms in the upper portion of the Nenana River drainage basin. Thirteen erratics from the Reindeer Hills, a massif that protrudes from the upper Nenana River lowland, average 16.6 ± 2.0 ka. A group of young erratics from the highest elevations of the massif cluster around 15.5 ± 0.8 ka (n = 5), which may record the timing of deglaciation of the summit by local glaciers. If so, then the lower valley walls of the massif were deglaciated between 19 and 17.3 ± 2.3 ka (Dortch, 2006).

In the Swift River valley of the western Alaska Range (Fig. 4), four ^{10}Be ages from the largest (2-6 m high) and most stable moraine boulders that we have seen in Alaska constrain the age of the late Wisconsin (Farewell II equivalent) terminal moraine to between 21 and 19.6 ± 0.9 ka (Briner et al., 2005). In the central Alaska Range, moraines offset by prominent faults in five valleys were recently dated by ^{10}Be to determine slip rates (Matmon et al., 2006). The moraines are located well upvalley from late Wisconsin terminal moraines, and their ages can be divided into an older age group of 17-16 ka (2 moraines) and a younger group of 13-12 ka (3 moraines). All moraines were dated by at least three ^{10}Be ages, and two of the younger moraines were particularly well dated. Both are within 2 km of extant glacier snouts; one is 11.7 to 11.0 ± 0.5 ka (7 samples), and the other 14.2 to 12.2 ± 1.3 ka (11 samples).

To summarize the late Pleistocene glacial chronology in the Alaska Range, ^{10}Be ages from two sites indicate that moraines of the penultimate glaciation stabilized between 60 and 55. The ^{14}C and ^{10}Be ages suggest that the late Wisconsin terminal moraines were deposited 21 to 20 ka, followed by retreat to an ice margin between 19 and 17 ka. Later readvances seem to have occurred between 17 and 16 ka, and 14 and 12 ka. Finally, the latest Pleistocene advance is dated by ^{14}C in McKinley Park and by ^{10}Be in the eastern Alaska Range to between 12 and 11 ka.

Ahklun Mountains

The Ahklun Mountains, a 150 by 200 km range in southwestern Alaska (Fig. 1), were covered by the largest ice mass in western Alaska. The range has been the focus of Quaternary research in the last decade, and a detailed mid- and late-Quaternary glacial history has emerged through surficial mapping, and 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; Axford and Kaufman, 2004; Levy et al., 2004). During the late Pleistocene, the Ahklun Mountains hosted an ice cap over its east-central spine that expanded radially, extending farther to the south and west than to the north and east (Fig. 5); isolated alpine glaciers occupied the highest valleys beyond the ice cap margin. In most valleys, late Pleistocene drift is composed of several moraine belts formed by outlet glaciers of the central ice cap (Manley et al., 2001).

The penultimate drift (deposited during the locally-termed Arolik Lake glaciation; Fig. 3) is dated in several locations across the range. In the southern Ahklun Mountains, Kaufman et al. (2001a) report a thermoluminescence (TL) age of 70 ± 10 ka on lava-baked sediment that underlies penultimate drift and provides a maximum-limiting age on the glaciation. Manley et al. (2001) report a minimum ^{14}C age of 39.9 ka on organic material that overlies Arolik Lake drift. In the western Ahklun Mountains, Briner et al. (2001) used four ^{36}Cl exposure ages on erratic boulders deposited in the Goodnews River valley to constrain the age of the Arolik Lake glaciation to between 56 and 53.8 ± 2.6 ka. Thus, the ^{36}Cl ages on boulders deposited during the Arolik Lake glaciation fit well between the TL maximum age of 70 ± 10 ka and the ^{14}C minimum age of 40 ka. These ages are in general agreement with amino acid and luminescence ages from glacial-estuarine sediments of the penultimate glaciation in the Bristol Bay lowland (Fig. 4), which ranged between 90 and 55 ka (Kaufman et al., 1996). Collectively, these ages indicate a major glaciation in the Ahklun Mountains roughly coincident with MIS 4; in the Bristol Bay low lands, however, we cannot exclude the possibility that the advance culminated late during MIS 5.

The age of the late Wisconsin drift (deposited during the locally-termed Klak Creek glaciation; Fig. 3) is known from several ^{14}C determinations from hummocky moraine belts and associated deposits in the western Ahklun Mountains. In the southwestern Ahklun Mountains, the late Wisconsin glaciation is well dated by ^{14}C ages that bracket the sediment from a glacier-dammed lake that overflowed into Arolik Lake. The arrival to, and the retreat from, the maximum position reached by the Goodnews River valley outlet glacier are tightly constrained in lake sediment cores to between 24 and 22 ka (Kaufman et al., 2003). Four ^{36}Cl ages from boulders on the terminal moraine in a nearby valley range between 21 and 19.6 ± 1.5 ka (Briner et al., 2001). Manley et al. (2001) report a minimum ^{14}C age of 19.9 ± 0.3 ka for next-to-oldest hummocky drift belt deposited during the late Wisconsin. Thus, following the deposition of the terminal moraine between 24 and 22 ka, ice in the Ahklun Mountains deposited a second moraine just before 20 ka.

Following several minor fluctuations and extensive ice stagnation, late Wisconsin glaciation in the Ahklun Mountains concluded with a late-glacial readvance represented by several small, single-crested vegetated moraines a few kilometers downvalley of extant glaciers in some, but not all, of the highest valleys in the range. In the Mt. Waskey valley (Fig. 5), a sediment core that penetrated to glacial-lacustrine mud in Waskey Lake has a basal ^{14}C age of 11.0 ± 0.2 ka (Levy et al., 2004). The lake is impounded by the Mt. Waskey moraine. Briner et al. (2002) obtained exposure ages on nine granodiorite boulders (five ^{10}Be ages, two ^{26}Al ages, and two $^{10}Be/^{26}Al$ average ages) from this and from morphostratigraphically similar moraines in a neighboring valley. Excluding two old outliers, the moraines stabilized between 11.7 and 10.6 ± 0.8 ka. Because the basal age from Waskey Lake suggests that the moraines are older than 11 ka, the best estimate for their stabilization age is between 11.7 and 11 ka.

Late Wisconsin ELAs have been estimated from reconstructed cirque and valley glaciers surrounding, and independent of, the Ahklun Mountains ice cap (Manley et al., 1997). These ELAs range from 600-800 m asl in the north, to 280-480 in the southwest, and average 540 ± 140 m asl, roughly 200-400 m lower than the ELAs of modern glaciers in the highest portion of the Ahklun Mountains. The gradient of the ELAs sloped 1.7 to 2.5 m km^{-1} toward the southwest during the late Wisconsin (Manley et al., 1997). In the western Ahklun Mountains, several early Wisconsin valley glaciers have reconstructed ELAs that are 50-90 m lower than late Wisconsin ELAs using the same techniques (Briner and Kaufman, 2000).

Discussion

Temporal and Spatial Patterns of Late Pleistocene Glaciation in Alaska

The application of new geochronological methods in Alaska has greatly improved the understanding of the timing of mountain glacier fluctuations during the late Pleistocene. This is especially true for the penultimate glacier advance, which for decades was suspected to post-date the last interglaciation (Hamilton, 1986b, 1994; Hamilton et al., 2001). The penultimate advance culminated between 60 and 50 ka, based on cosmogenic exposure ages of moraine boulders in three valleys from sites up to 800 km apart (Fig. 6). Thus, we conclude with some certainty that the largest advance of mountain glaciers during the late Pleistocene occurred prior to the global LGM, and likely during MIS 4 or early during MIS 3. Although not yet dated in the Brooks Range, penultimate moraines there are likely of similar age, because they post date the Old Crow tephra (Hamilton, 2001). Penultimate drift in some locations might pre-date the late Pleistocene, such as in the Delta River valley (Begét and Keskinen, 2003). In other valleys of the north Alaska Range, however, penultimate drift is late Pleistocene age (Dortch, 2006), in agreement with ages from elsewhere in the state, suggesting that the relative extent of glacier advances in the Delta River valley was anomalous. A pulse of loess deposition in the Tanana River valley (Begét, 2001) that appears to coincide with MIS 4 supports the notion of a regionally significant early Wisconsin glacier advance in the north Alaska Range (Fig. 6).

New chronologies have also improved the ages of mountain glacier fluctuations during the late Wisconsin. Although still sparse, the chronologies across Alaska show some pattern in timing of the maximum extent of mountain glaciers during MIS 2. Many of these chronologies are based on cosmogenic exposure ages of moraine boulders, which likely date the timing of moraine stabilization upon glacier retreat (Briner et al., 2005). In northern Alaska, glaciers retreated from their late Wisconsin terminal moraines by 25 ka, compared to 22 to 20 ka in central and southern portions of the state. The age of advance phase of late Wisconsin glacier expansion is constrained in very few places: in Denali National Park in the Alaska Range, glaciers neared their late Wisconsin limit around 22 ka, and around 24 ka at Arolik Lake in the

Ahklun Mountains. Thus, the retreat of Brooks Range glaciers seems to have occurred several thousand years before the advance of glaciers in central and southern Alaska during MIS 2.

Drift deposited during MIS 2 has been dated throughout Alaska in areas other than the three mountain ranges discussed here. Although dozens of limiting radiocarbon ages loosely constrain moraines to MIS 2 (e.g., Hamilton, 1994; Mann and Hamilton, 1995), only a few additional localities have tight age control. On the upper Alaska Peninsula, radiocarbon ages from river bluffs constrain MIS 2 advances to between ~30 and 14.8 ka (Stilwell and Kaufman, 1996). On nearby Kodiak Island, the maximum MIS 2 advance occurred between ~26 and 17.8 ka (Mann and Peteet, 1994). Outlet glaciers that filled Cook Inlet, south-central Alaska, retreated from their MIS 2 maximum positions by ~19.4 ka (Reger and Pinney, 1996).

Following the maximum phase of the late Wisconsin, glaciers across the state constructed end moraines during subsequent periods of stabilization or re-advance. Although most glaciated valleys across Alaska contain multiple moraines, few have been dated, hampering state-wide comparisons; however, glaciers in many valleys built sizeable moraines near terminal moraines shortly following their initial retreat. In the Ahklun Mountains, for example, prominent end moraines were deposited about 20 ka, and in the Alaska Range, end moraines post-dating the terminal moraine formed around 19 ka. In both cases, glaciers stabilized near their former limits for one or two thousand years following the maximum phase.

Of particular interest is the evidence for a glacier re-advance in Alaska concurrent with the North Atlantic Younger Dryas event. In the Ahklun Mountains, a late-glacial advance culminated 11.7 to 11 ka in some of the highest tributary valleys (Briner et al., 2002). In the northern Alaska Range, the MP-IV advance is dated by ^{14}C to between 12.3 and 11 ka (Ten Brink and Waythomas, 1985), the same age as one of the moraines along the northern range front dated by ^{10}Be to between 11.7 and 11 ka (Motmon et al., 2006). A ^{14}C age on sediment overlapping a moraine in the Kenai Mountains, south-central Alaska, might correlate with the Younger Dryas (Reger et al., 1995), and other proxy climate records from Alaska clearly attest to a climatic reversal during the Younger Dryas (e.g., Hu et al., 2006). Nonetheless, widespread evidence for a glacier re-advance during the Younger Dryas has yet to be revealed across Alaska. The youngest late-glacial re-advance in the Brooks Range, for example, occurred prior to the Younger Dryas, between 15 and 13 ka (Hamilton, 2003). Thus, glaciers across the state register re-advances during the last glacial-interglacial transition, but only in a few places can they be considered a candidate for a glacier advance during the Younger Dryas.

Correlations with Adjacent Regions

Late Pleistocene mountain glacier chronologies are emerging worldwide, including in regions adjacent to Alaska, known collective as Berignia. In northeastern Siberia, Gualtieri et al. (2000) report 16 ^{36}Cl ages, and Brigham-Grette et al. (2003) report 12 ^{36}Cl ages from two mountain ranges (Pekulney and Koryak Mountains) where the glacial morphostratigraphy is similar to Alaska. The best-dated early Wisconsin glacial feature in northeastern Russia is glacially eroded bedrock with ^{36}Cl ages ranging between 69 and 56 ka. Although ages on late Wisconsin drift are scattered, they indicate that terminal and younger end moraines were deposited between 24 and 16 ka. An outwash terrace graded to an end moraine behind the terminal late Wisconsin moraine, thought to be close in age to the terminal moraine, is dated by a cluster of three ^{14}C ages from organics within the outwash that average 18.7 ± 0.5 ka.

In the western Yukon Territory, Canada, ^{10}Be ages have recently been obtained from penultimate drift deposited by a lobe of the Cordilleran Ice Sheet that emanated from the St. Elias and Coast mountains. Four ages on 1.5- to 3.7-m-high erratics range between 54 and 53.3 ± 1.3 ka, providing the first evidence that the penultimate drift in western Yukon (= Gladstone glaciation) dates to MIS 4 or early during MIS 3 (Ward et al., 2007). In contrast, penultimate drift derived from the Selwyn lobe of the Cordilleran Ice Sheet in central Yukon (= Reid glaciation) is younger than the Sheep Creek tephra (Westgate et al., 2001) and older than radiometrically dated basalt (Huscroft et al., 2004), and is correlated with MIS 8 age.

Given the few well-dated records of the penultimate glaciation in Beringia, it is difficult to characterize temporal patterns across the broader region. Although the penultimate drift dated from sites spanning 800 km across Beringia appears to coincide with MIS 4 or early MIS 3, the extent to which glacier maxima were attained synchronously from place to place is not known. A similar conclusion was reached based on the frequency and source of ice-rafted detritus (IRD) in the North Pacific: Athough the mass accumulation rate of IRD was high if not higher during MIS 4 than MIS 2 (Hewitt et al., 1997), significant variations in source and timing of IRD suggest regional controls on iceberg input (St. John and Krissek, 1999). The maximum MIS 2 advance seems to have occurred earliest in arctic Alaska (27-25 ka) and later (24-20 ka) in regions more strongly influenced by the Pacific Ocean.

Paleoclimate Controls

The most recent state-wide compilation of snowline estimates for the late Wisconsin was based on cirque-floor altitudes (Péwé, 1975). These show a spatial pattern similar to the modern snowline, namely a southwest moisture source and prominent orographic effects on the windward and leeward side of major mountain ranges. More recent studies indicate that glacier ELAs were generally 300-600 m lower across Alaska during the LGM (Hamilton and Porter, 1975; Kaufman and Hopkins, 1986; Mann and Peteet, 1994; Stillwell and Kaufman, 1996; Manley et al., 1997; Briner and Kaufman, 2000; Balascio et al., 2005b). This relatively minor ELA lowering contrasts with a more typical mid-latitude value of 1000 m (Broecker and Denton, 1990) and has long been attributed to arid conditions related to increased continentality resulting from the emergence of the Bering-Chukchi platform during eustatic sea-level lowering (e.g., Hopkins et al., 1982). Moisture sources may have been further restricted as sea-ice cover expanded over the Aleutian basin in southern Bering Sea to the southwest (Sancetta et al., 1984) and the Beaufort Sea in the north (Phillips and Grantz, 1997). Further southwest, in the northwestern Pacific, however, more recent multi-proxy evidence indicates that sea-surface temperature was not significantly lower at 20 ka compared with the Holocene (Sarnthein et al., 2006). Similarly, in the Gulf of Alaska, dinoflagellate cyst assemblages indicate little change in temperature and sea-ice cover (de Vernal et al., 2005). On land, cold and dry conditions during the late Wisconsin are inferred from pollen records, which reveal a sparsely vegetated landscape dominated by herbaceous tundra across Alaska (e.g., Anderson et al., 2004). Hydrologic-balance models informed by lake-level evidence indicate considerable reduction in effective moisture (Barber and Finney, 2000). In contrast, paleobotanical and fossil insect data from central and northern Bering land bridge indicate mesic conditions during the late Wisconsin (Elias et al., 1997), and a relatively mild LGM temperature depression (Elias, 2001). Pollen and lake-status indicate that, although generally cold and arid, central Beringia may have been slightly more mesic than interior Alaska (Ager, 2003).

The paleoenvironmental evidence for cold conditions in Alaska contrasts with results of paleoclimate modeling for the LGM. General circulation models (GCMs) consistently show enhanced southwesterly flow of warm air into Alaska (e.g., Kutzbach et al., 1998). Recent simulations using Community Climate System Model version 3 (CCSM3) clearly depict significantly warmer-than-modern (pre-industrial) annual temperature across Alaska during the LGM, although the simulated warming diminishes with the height of Laurentide Ice Sheet (Otto-Bliesner et al., 2006). Seasonally resolved output from CCSM3 (B. Otto-Bliesner, pers comn, 2007) shows that the warming occurs during both winter and summer months. The model also shows decreased precipitation across Alaska, except for the Gulf of Alaska. The models are consistent with the paleo-glacier evidence that clearly attests to limited ice extent in Alaska compared with most northern high-latitude regions. We suggest that glacier expansion in Alaska was limited not only by decreased precipitation, which is well known from the paleoenvironmental record, but also by a lack of significant summer cooling during the LGM.

The growing evidence for maximum late Pleistocene glaciation during MIS 4 or early during MIS 3 in Alaska summarized here contrasts with the global marine oxygen-isotope record, which features maximum ice volume late during the last glacial cycle. Many mountain glaciers at lower latitudes in North America attained their maximum extent during MIS 2 (Gillespie and Molnar, 1995; Pierce, 2004). Previous studies have emphasized evidence for “out-of-phase” glaciations in Beringia (e.g., Brigham-Grette, 2001; Kaufman et al., 2001a). Glaciers in northeastern Siberia and western Alaska expanded onto the continental shelf several times during the Pleistocene. They deposited glacial-marine sediment hundreds of kilometers inboard the shelf edge, implying that eustatic sea level was high during the maximum phase, and supporting the hypothesis that large glacier expansions in Alaska require a proximal source of moisture. Sea level probably fell below the shelf break to expose the Bering-Chukchi platform following substage 5a, and the transition between MIS 5a and 4, around 75 ka (based on orbitally tuned global marine oxygen isotopes; Martinson et al., 1987), has been suggested as a candidate for extensive glacier growth in Beringia (Brigham-Grette, 2001). Eustatic sea level rose again during MIS 3. Dated coral reefs in the Pacific and other evidence reviewed by Cabioch and Ayliffe (2001) indicate a transgression to within 30 to 60 m of present, seemingly high enough to inundate a large portion of the continental shelf in central Beringia. This proximal moisture source would have enhanced moisture availability during this interval. During MIS 2, in contrast, moisture availability decreased as sea level fell from the shelf break. In addition, GCM simulations show that, as the Laurentide Ice Sheet grew, the Aleutian low-pressure system strengthened (Otto-Bliesner et al., 2006). The instrumental data demonstrate that a stronger, eastward-shifted low steers storms away from western Alaska and into the Gulf of Alaska (Rodionov et al., 2005). Increased winter storminess would have nourished the Cordilleran Ice Sheet over the coastal ranges. The higher ice enhanced the orographic barrier and narrowed passages for low-level moisture transport, further depleting moisture in interior Alaska during the LGM.

Summary and Conclusion

This paper focused on the most robust late Pleistocene mountain glacial chronologies currently available in Alaska. New cosmogenic exposure ages combined with ^{14}C, luminensence, and tephra-based ages have improved the geochronological control on the glacial history of Alaska. Although previously suspected to be early Wisconsin in age (Hamilton, 1994), new numerical ages place the penultimate glaciation into MIS 4 or early MIS 3. During the late Wisconsin, glaciers appeared to have deposited terminal moraines earlier (27-25 ka) in arctic Alaska than in southern Alaska (24-20 ka). Glaciers remained close to their maximum extent for thousands of years following the local glacial maximum. Although their ages are generally not well constrained, the numerous end moraines upvalley of terminal moraines document the response of glaciers to climate change through the late-glacial period. Finally, glacier advances in a few valleys may be correlative with the Younger Dryas event.

Among the most notable features of late Pleistocene glaciation in Alaska are: 1) more extensive glaciation during MIS 4/3 than during MIS 2; 2) relatively restricted glacier extent, requiring only modest (300-600 m) ELA lowering compared to the mid-latitudes; and 3) an earlier MIS 2 maximum extent in the arctic- versus Pacific-dominated portions of the state. These features likely relate to temporal and spatial patterns of moisture availability, with more moisture available during MIS 4/3 than during MIS 2. In addition, relatively mild summers may have combined with arid conditions during MIS 2 to limit glacier expansion. Similar to temporal patterns elsewhere, such as in the Andes Mountains where the maximum MIS 2 glaciation coincided with the global LGM in the south (Kaplan et al., 2004) but pre-dated it in the north (Smith et al., 2005), the timing of peak MIS 2 glaciation in Alaska differed by several thousand years. Glaciers in Alaska probably retreated from their terminal MIS 2 limit prior to ~19-17 ka, the interval of common mid-latitude glacier retreat in both hemispheres recently recognized by Schaefer et al. (2006).

We have focused on the few areas where the ages of mountain-glacier moraine sequences are reasonably well known. For these, the prominent penultimate advance has been dated to within the last glaciation, and the timing of the maximum phase of the MIS 2 glaciation is secure. In many areas of the state, however, the glacial geology has been studied at the reconnaissance level only, and numerical age control is lacking. In the Brooks Range in arctic Alaska, for example, the penultimate drift is undated. Although recent efforts have revealed a systematic temporal pattern to the deposition of MIS 2 terminal moraines across the state, age control on the numerous moraines younger than the terminal moraine is sparse, including during the late-glacial period. As new information on the ages and extent of glacier fluctuations continues to be generated, Alaska's alpine glacier record combined with glacier-climate models will lead to improved and quantitative understanding of the paleoclimate controls on glaciation.

Acknowledgments

Our glacial-geologic research in Alaska was supported by NSF grants OPP-9977972 and OPP-9977974 to DSK. This compilation benefited from stimulating discussions with many people, including Yarrow Axford, Nick Balascio, Jason Dortch, Thomas Hamilton, William Manley, Al Werner and the entire INQUA Mountain Glacier working group.

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Figure Captions

Figure 1 Alaska showing the extent of glacier ice during the late Wisconsin (from Kaufman and Manley, 2004; available online by Manley and Kaufman, 2002) and areas discussed in this paper where moraines sequences spanning the late Pleistocene have been well dated. Inset shows extent of coalescent ice sheets over North America during the last glacial maximum from Dyke et al. (2002).

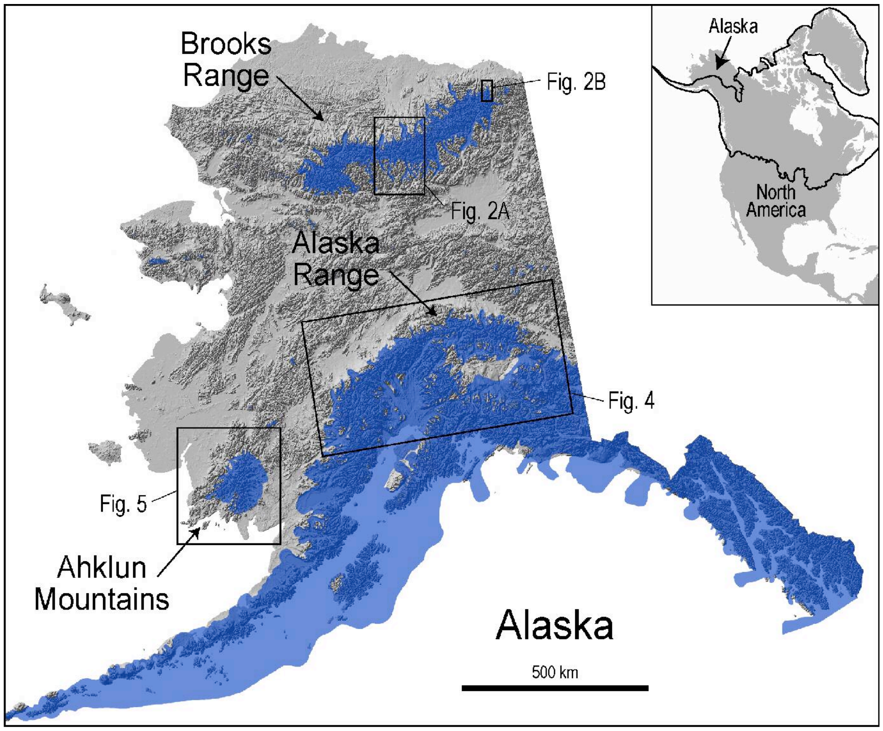
Figure 2 (A) Central and (B) northeastern Brooks Range showing the extent of glaciers during the penultimate and late Wisconsin glaciations with locations of key ages. Map areas are shown in Fig. 1.

Figure 3 Correlation chart showing approximate ages and local nomenclature for glacial intervals in areas discussed in the text. The dating method that the age constraints are based on is listed.

Figure 4 North Alaska Range showing the extent of glaciers during the penultimate and late Wisconsin glaciations with locations of key ages. Map area shown in Fig. 1; explanation of map symbols in Fig. 2.

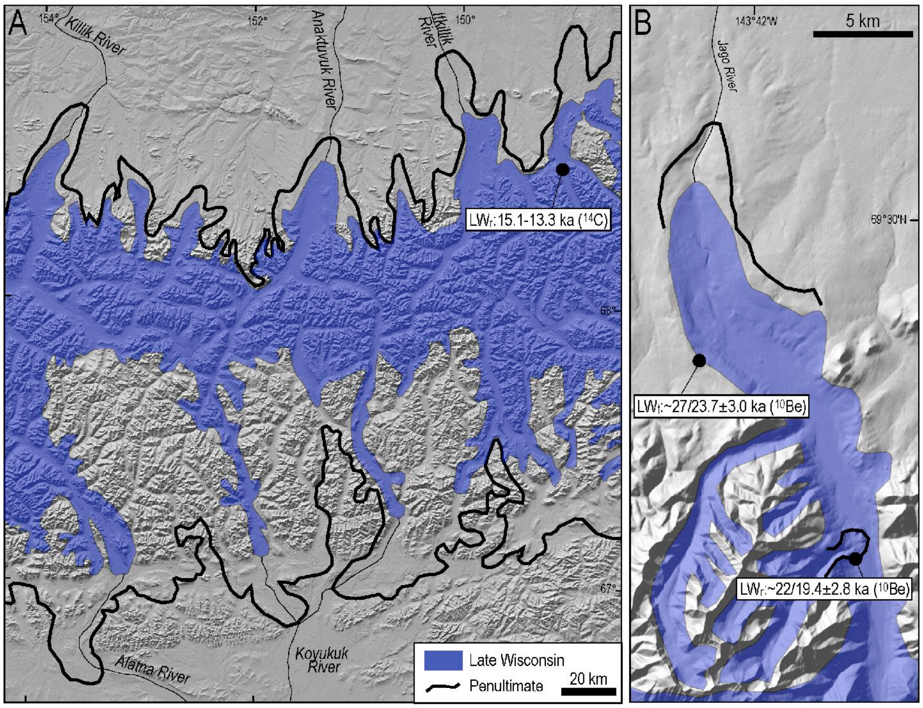
Figure 5 Ahklun Mountains showing the extent of glaciers during the penultimate and late Wisconsin glaciations with locations of key ages. Map area shown in Fig. 1; explanation of map symbols in Fig. 2.

Figure 6 Time-distance diagrams for glaciers in the three areas discussed in the text. The magnetic susceptibility (MS) profile for Fairbanks loess (Begét, 2001) and the position of soils (S) and the Old Crow tephra (OCT) is shown for comparison. The global marine oxygen-isotope record (Martinson et al., 1987) and marine isotope stages (MIS) shown for reference. Solid lines = securely dated glacier extent; dashed lines = approximate glacier extent.



Briner and Kaufman, Figure 1

190x244mm (300 x 300 DPI)



Key to labels in Figures 2, 3 and 4

LW=Late Wisconsin, EW=Early Wisconsin, P=Penultimate, subscripts: t=terminal moraine, r=recessional or readvance moraine
(^{14}C)=radiocarbon age, (^{10}Be/^{26}Al)=cosmogenic exposure age, (TL)=thermoluminescence age, (OCT)=Old Crow Tephra

Briner and Kaufman, Figure 2

190x249mm (300 x 300 DPI)

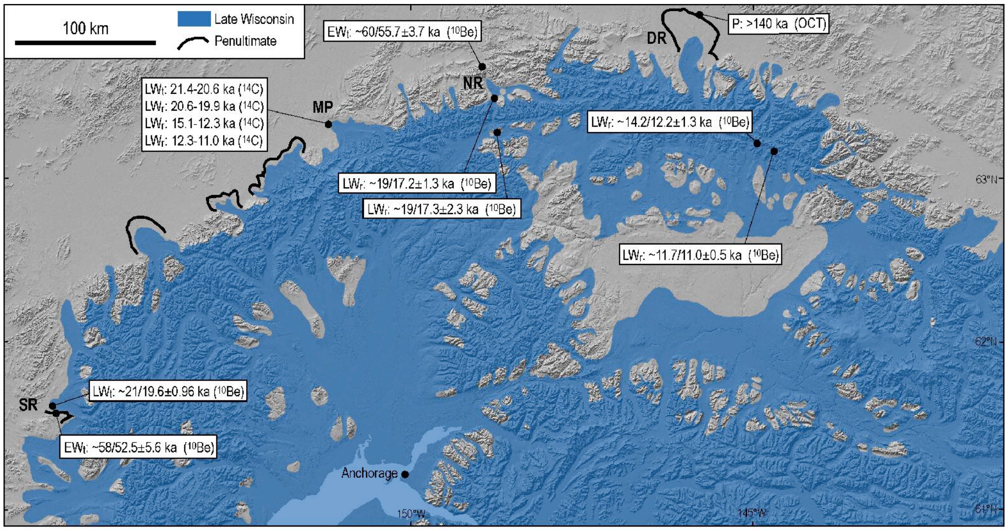
		Brooks Range[1] (fig. 2)	Alaska Range (fig. 4)				Ahklun Mountains[6] (fig. 5)
			Western[2]	Denali N.P.[3]	Nenana River[4]	Delta River[5]	
Late Pleistocene	Late Wisconsin	Itkillik II (r) **15** to 13 ka (^{14}C) Itkillik II (r) 22-19 ka (^{10}Be) Itkillik II (t) 27-24 ka (^{14}C, ^{10}Be)	Farewell II (t) 21-20 ka (^{10}Be)	McKinley Park IV (r) **12.3** to 11.0 ka (^{14}C) McKinley Park III (r) **15.1** to 12.3 ka (^{14}C) McKinley Park II (r) **20.6** to 19.9 ka (^{14}C) McKinley Park I (t) **21.4** to 20.6 ka (^{14}C)	Carlo (r) 19-17 ka (^{10}Be) Riley Creek II (r) Riley Creek I (t)	Donnelly	Mt. Waskey 11.7-11.0 ka (^{10}Be, ^{26}Al, ^{14}C) Klak Creek (r) **22** to 20 ka (^{14}C, ^{36}Cl) Klak Creek (t) **24** to 22 ka (^{14}C)
Late Pleistocene	Early Wisconsin	Itkillik I	Farewell I (t) 58-53 ka (^{10}Be)	?	Healy 60-56 ka (^{10}Be)	?	Arolik Lake **70** to 56-54 ka (TL, ^{36}Cl)
	MIS 6	Sagavanirktok	Selatna	?	Lignite Creek	Delta >140 ka (tephra)	?

Key: (t)=terminal moraine, (r)=readvance or recessional moraine, **bold**=maximum age, plain font=minimum age.

References for chronolgies: 1: Hamilton, 1982, 2003; Balascio et al., 2005a; Briner et al., 2005. 2: Briner et al., 2005. 3: Ten Brink and Waythomas, 1985; Werner et al., 1993; Child, 1995. 4: Dortch, 2006. 5: Begét and Keskinen, 2003. 6: Briner et al., 2001, 2002; Kaufman et al., 2003b).

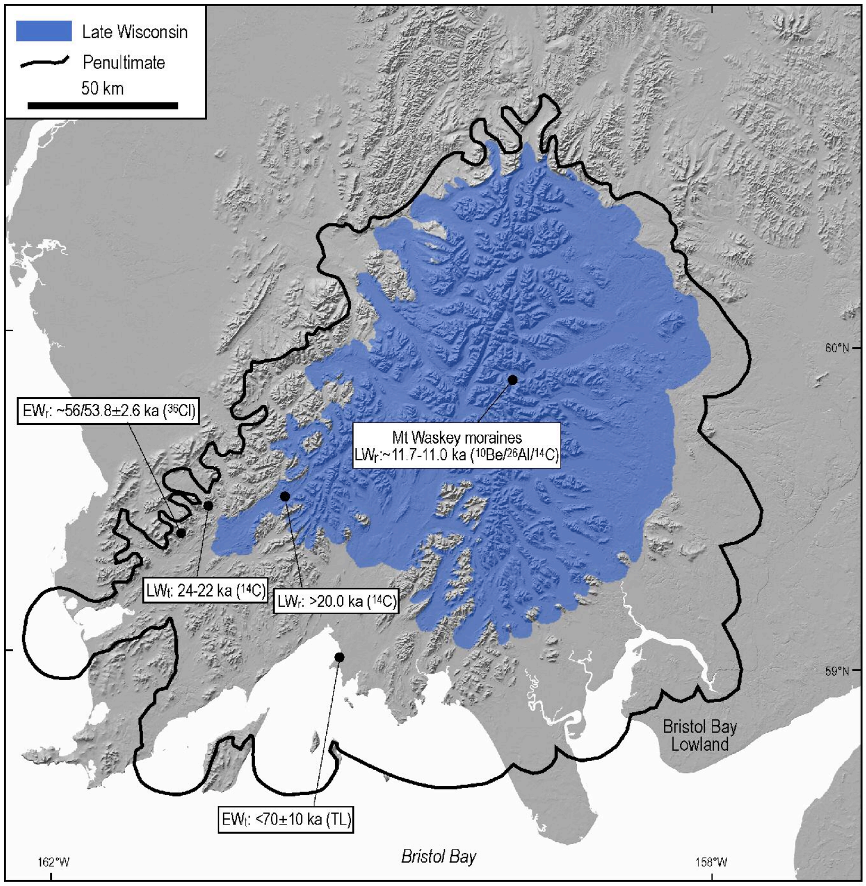
Briner and Kaufman, Figure 3

154x239mm (300 x 300 DPI)



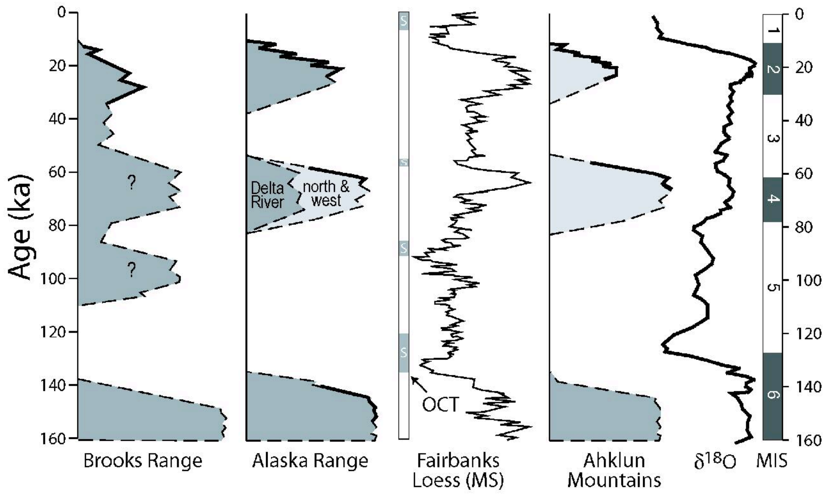
Briner and Kaufman, Figure 4

249x184mm (300 x 300 DPI)



Briner and Kaufman, Figure 5

178x248mm (300 x 300 DPI)



Briner and Kaufman, Figure 6

169x242mm (300 x 300 DPI)