Latest Pleistocene Advance and Collapse of the Matanuska – Knik glacier system, Anchorage Lowland, Southern Alaska

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Summary: *Here we document evidence for a major ice advance in Alaska and perhaps the mid-latitudes of the Northern Hemisphere, which falls between the LGM and major late glacial climate oscillations.*

Abstract

At the end of the last ice age, glacier systems worldwide underwent dramatic retreat. Here, we document the advance and retreat of a glacier system with adjacent marine- and land-based components during the latter part of the Termination. As part of our study, we utilize three lines of evidence: lithologic provenance, geomorphic mapping, and radiocarbon ages derived from lake cores to reconstruct glacier extent and timing of advance and retreat within our study area centered at N 61.50°, W 149.50°, just north of the city of Anchorage, Alaska. Two glaciers, sourced in the Talkeetna and Chugach Mountains, flowed down the Matanuska and Knik Valleys, forming a coalesced lobe that advanced onto the Anchorage Lowlands and terminated at Elmendorf Moraine. Our reanalysis of existing radiocarbon chronology from an overridden glaciomarine deposit places an advance at 16.8 ka BP. We use the presence of lithologies unique to the Matanuska catchment in glacial drift to delineate the paleoflow lines and to estimate the suture line of the two glacier systems. The eastern side of the lobe, attributed to ice flow from the Knik Valley, was in contact with elevated marine waters within the Knik Arm fjord, and thus retreat was likely dominated by calving. Geomorphic evidence suggests the western side of the

lobe, attributed to ice flow from Matanuska Valley, retreated due to stagnation. We constrain retreat of the combined Matanuska and Knik lobe with thirteen new radiocarbon ages, in addition to previously published radiocarbon ages, and with geomorphic evidence suggesting the retreat occurred in two phases. Retreat from the Elmendorf Moraine began between 16.8 and 16.4 ka BP. A second, faster retreat phase occurred later and was completed by 13.7 ka BP. With the 140 km of total retreat occurring over ~3,000 years or less. This pattern of glacial advance and retreats agrees well with the deglacial histories from the southern sectors of the Cordilleran Ice Sheet, as well as many other alpine glacier systems in the western U.S. and northern Alaska. This agreement in the behavior of glacier systems may indicate that climate oscillated over western North America early in deglaciation before it was recorded in other proxies such as ice cores. Furthermore, the period in which we note mountain glacier collapse in northwestern North America is synchronous with the worldwide glacial termination raising questions about intrahemispheric linkages.

Keywords: Anchorage Lowland, deglaciation, Elmendorf Moraine, Matanuska Glacier, Knik Glacier, Termination

1.0 Introduction

The structure and cause of the last glacial termination have long been an area of interest, as it provides a view into the most recent transition from glacial to interglacial climate. Recently, much attention has focused on the final stages of the Termination (Putnam et al., 2013; Svendsen et al., 2015; Pendleton et al. 2015). However, the structure of climate change between the peak of full-glacial conditions and the subsequent highly variable, oscillating, late glacial (18-11.6 cal ka BP) climatic conditions is less clear. Denton et al. (2006) called the period between 17.5 to 14.5 cal ka BP the “Mystery Interval” because several enigmatic changes occurred in the climate system, including the H-1 iceberg discharge (Bond et al., 1992; Hemming, 2004), the possibility of hyper-cold winters in Greenland (Denton et al., 2005), the expansion of closed lakes in the Great Basin (Benson, 1993; Garcia and Stokes, 2006), and the major recession of mountain glaciers in both the Southern (Denton et al., 1999; Schaefer et al., 2006; Strelin et al., 2011; Putnam et al., 2013) and Northern Hemispheres (Schlüchter, 1988; Young et al., 2011; Stroven et al., 2014). Brocker and Putnam (2012) have divided the Mystery Interval into an older “Big Dry” (17.5 to 16.1 cal ka BP) and a younger “Big Wet” (16.1 to 14.5 cal ka BP) periods reflecting global hydrology changes. Here we refine the pattern of glacier advance and retreat during the “Mystery Interval” in southern Alaska and compare this recession to other areas in western North America. To place the existing chronology into context, we delineated the past flow lines of two confluent glacial lobes using drift clast lithology, assessed the distribution of glacial and deglacial landforms, and expanded the existing chronology with radiocarbon ages from terrestrial organics retrieved from basal-lake sediments.

1.1 Regional Glacial History

In Alaska, one record of climate change is the signature of mountain glacier fluctuations. At present, the best dated records are from the Brooks Range in northern Alaska (Briner et al.,

2005; Badding et al., 2013; Pendleton et al., 2015), Alaska Range in central Alaska (Porter et al., 1983; Briner et al., 2005; Dortch et al., 2010), and the Ahklun Mountains in southwestern Alaska (Kaufman et al., 2003; Briner and Kaufman, 2008; Kaufman et al., 2012). Advances associated with the Last Glacial Maximum (LGM; 22-18 cal ka BP) are constrained to ~21 cal ka BP across the state (Briner et al., 2005; Dortch et al., 2006; Pendleton et al., 2015; Kaufman et al., 2003). Moraines indicating glacial advances, or stabilizations, inboard of the LGM extent are rarer in the glacial record. Pendleton et al. (2015) collected samples from moraine boulders within the LGM extent in the Brooks Range which place a local advance at $\sim 17.2 \pm 10$ cal ka BP, and Dortch collected samples from boulders on deposits assigned to the Carlo Stage in the Alaska Range which yielded exposure ages of ~ 19 -17 cal ka BP. Further abroad, the nearby northern sector of the Cordilleran Ice Sheet achieved a maximum position in the Mackenzie Mountains during the LGM (McConnell Glaciation) at ~ 18 cal ka BP, while nearby alpine glaciers reached a maximum between 22-17 ka (Stroven et al., 2010; Stoven et al., 2014).

1.2 Anchorage Lowlands Glacial History

Situated in south-central Alaska, the Anchorage lowland lies on the northern margin of the Cook Inlet (Fig. 1). Continental climate dominates the interior areas while a maritime climate characterizes southern coastal Alaska (Stafford et al., 2000). The Anchorage lowlands are bounded to the south by the Chugach Mountains and the north by the Talkeetna Mountains. Modern glaciers found in the Matanuska and Knik valleys in the northern part of the Anchorage lowlands are fed by accumulation areas in the Chugach Mountains, which receive moisture from Prince William Sound and the Gulf of Alaska to the south (Fig. 1).

During the Naptowne Glaciation (Reger et al., 1995), a glacier system, hereafter referred to as the Matanuska-Knik Glacier, sourced in the Chugach and Talkeetna Mountains advanced across the Anchorage Lowlands forming the Elmendorf Moraine (Fig. 1). The Matanuska Lobe flowed 170 km from the southern Talkeetna Mountains and the northern Chugach Mountains, while the Knik Lobe flowed 120 km from the southern Chugach Mountains. The confluent Matanuska-Knik Glacier flowed across the Anchorage Lowlands to a common ~ 100 km long terminus marked by the Elmendorf Moraine. Stratigraphically below the Elmendorf Moraine and associated drift lies the Bootlegger Cove Formation (BCF), containing a fossiliferous horizon, as well as marine clay, silt, and sand layers.

The assemblage of macrofossils from the BCF includes gastropods (*Buccinum cf.*, *Buccinum physematum*, *Cryptonatica clausa*, *Cryptonatica sp.*), pelecypods (*Clinocardium ciliation*, *Clinocardium sp.*, *Hiatella arctica*, *Macoma cf.*, *Macoma calcaria*, *Mya truncata*), and barnacles (*Balanus sp.*). While late glacial estimates of past relative sea level (RSL) are scarce from the Gulf of Alaska, a few control points exist in the Cook Inlet area. Minimum constraints during the start of local deglaciation place RSL at least 10 meters higher than present from 19.1-18.7 cal ka BP (Mann and Hamilton, 1995; Reger and Pinney, 1995; Shugar et al., 2014). Reger et al. (1995) estimate a minimum of 86 meters of isostatic depression in this area at ~ 17 cal ka BP, with relative sea level being 36 m higher than present prior to 16.3 cal ka BP. Finally, a

radiocarbon-dated peat found at 24 meters above present sea level places RSL below that level between 16.8-14.6 cal ka BP (Rubin and Alexander, 1958).

Radiocarbon ages derived from BCF shells in front of and below the Elmendorf Moraine range from 17.6 to 15.7 cal ka BP, providing a maximum limit on the timing of moraine emplacement, and thus a constraint on the timing of the advance (Schmoll et al., 1972; Reger et al., 1995; Table 1; Fig. 2). Although nine ages provide constraints on the BCF, only one shell was collected directly beneath the Elmendorf drift, and it yielded an age of 17.5-14.3 cal ka BP (W-2389; Schmoll et al., 1972). Technically, this radiocarbon age should be the closest maximum bracket on the deposition of the Elmendorf Moraine, though this age is slightly younger, yet overlapping with the range of ages from the entire assemblage of nine samples (Schmoll et al., 1972; Reger et al., 1995). The implications of using either the single age or the range of ages, as a maximum constraint will be discussed further in the context of our new chronology in the discussion section.

The existing deglacial chronology tracks retreat of the Matanuska-Knik Glacier over 140 km during the late glacial (Table 1; Fig. 2). Organic matter from a kettle, Lorraine Lake, on the Elmendorf Moraine yields an age of 15.4-15.0 cal ka BP (Kathan et al. 2004), constraining the deposition of the Elmendorf Moraine to within the “Mystery Interval” (Denton et al., 2006). Up valley from the Elmendorf Moraine, at the Hay Flats site, organic-rich silt recovered from a borehole yielded an age of 15.3-11.3 cal BP (GX-15241; Combellick, 1999). Further up valley, at Hundred-Mile Lake, ~5 km north of the current Matanuska Glacier terminus, Yu et al. (2008) dated a terrestrial shell 25 cm above the base of a lake core that yielded an AMS age of 13.3-13.0 cal ka BP (ETH-29990). By employing an age model utilizing terrestrial macrofossil and shell pairs at higher stratigraphic levels, Yu et al. (2008) estimated an age of 14.5 cal ka BP for deglaciation at Hundred-Mile Lake. The furthest up valley constraint on ice margin recession comes from the Matanuska Glacier Bog, located 3 km from present-day Matanuska Glacier. Exposed sediment sections in a lateral moraine contain organic sediments underlying till. Williams (1986) collected peat samples that yielded ages of 15.9-15.4 cal ka BP (BETA-11174) and 14.7-13.7 cal ka BP (USGS-2175). Tom Ager of the USGS re-dated this section using accelerator mass spectrometry analysis and found ages of 13.7-13.5 cal ka BP (WW-3623) and 13.4-13.3 cal ka BP (WW-3618). We note that Ager dated sedge seeds from the lowest peat exposed at the time of sampling, while Williams’ (1986) ages are derived from peat samples, which may contain contamination from coal deposits in the Matanuska Glacier catchment. Due to ongoing active erosion at the site, it is uncertain whether Williams and Ager sampled the same horizon. Thus, the offset in ages may be due to the fact that samples were collected from different horizons. The Hundred-Mile Lake and Matanuska Glacier bog sites collectively indicate that the Matanuska Lobe retreated to less than 5 km from the present position by at least 13.7 cal ka BP, and possibly before the end of the “Mystery Interval”.

2.0 Methods

2.1 Provenance

We use drift provenance mapping to distinguish ice-flow paths from the Matanuska and Knik catchments. The two glacier systems derive clasts from two accreted terranes with different lithologies (Winkler, 1992), thus drift from the Matanuska Valley should include lithologies from both Chugach and Peninsular terranes (Fig. 1), whereas the Knik Valley is situated only in the Chugach terrane. Local bedrock exposures of these lithologies were mapped by Winkler, (1992) and verified during our field work (outcrops are plotted in Fig. 2 as gray stippled polygons). We identified and recorded every rock type from random samples of 100-pebble minimum collections at 131 provenance stations in the study area (Fig. 4, Table S2) for a total of 16,920 pebbles counted in this study. We acknowledge that this region has undergone multiple glaciations, and the deposits sampled reflect the most recent advance. Thus, there is a possibility that the reworking of older glaciofluvial deposits may inject some inherited material into our dataset. Our investigations follow a drift provenance approach successfully implemented by Evenson and Clinch (1987), Kerr and Wilson (2000), Lilliesköld (1990), Shilts (1993, 1996), McClenaghan et al. (1997), and Kjarsgaard et al. (2004). The results of our survey were plotted in GIS software to visualize the spatial distribution of clast lithologies in the context of the geomorphic features described in the next section.

2.2 Geomorphology

Our geomorphic investigations focus on large-scale landforms related to the deglacial retreat history. The majority of the geomorphic features used in this study were already mapped in previous studies, such as Reger and Updike (1983), Yehle et al., (1990), Schmoll et al., (1999), and maps produced by the Alaska Division of Geological & Geophysical Surveys (DGGs; [http:// www.dggs.dnr.state.ak.us/pubs](http://www.dggs.dnr.state.ak.us/pubs)) and the United States Geological Survey (USGS; <http://www.usgs.gov/pubprod/publications>). These previous studies identified most of the eskers, moraines, streamlined bedforms, kettle lakes, and ice-stagnation features presented in Figure 3. The preponderance of our fieldwork verified, refined, and sampled these features through field observations guided by 1-m digital orthoimagery from the Alaska Geospatial Data Clearinghouse, 1:25,000 USGS topographic contour maps, and 1:25,000 geologic maps where available.

2.3 Chronology

Sediment cores were collected from post-glacial basins, using a modified, hydraulically assisted, Livingstone–Wright piston corer. Organic matter from basal contacts was sampled from split cores. Terrestrial macrofossils, such as wood, seeds, and leaves, were selected to avoid hard water effects or the potential contamination of coal debris. Fossils were submitted to ETH and Beta Analytic for accelerator mass spectrometer (AMS) radiocarbon analysis. Radiocarbon ages were calibrated using CALIB Rev. 7.1, with the IntCal13 datasets for terrestrial macrofossils (Reimer et al., 2013). Previously reported ages on marine shells were recalibrated with the MARINE13 dataset (Reimer et al., 2013) with the default marine correction factor of 405 yr

(Bard et al. 1990). Ages are reported in calibrated years as the two-sigma age range (Table 1). In the discussion, we examine the implication of local marine corrections (ΔR).

For the Elmendorf Moraine, we employ existing and new radiocarbon ages that lie in both maximum and minimum limiting stratigraphic positions to derive a probability estimate for the timing of advance. This approach employs the stratigraphic order of the samples to determine the most probable age of the moraine. Our method differs from that of Sharon (2001) in that we compute the calibrated probability density functions (PDF) of the radiocarbon dates, instead of the un-calibrated PDFs. For a review of methods for incorporating stratigraphic information into chronology problems see Buck et al. (2003). Kelly et al. (2015) describes the specific formulation we apply. The moraine age estimates yielded by this method is dependent on the subset of the available ages are used for the maximum and minimum brackets. The effects of sample selection are analyzed in the discussion. For the discussion, the reported ages from other sites have been listed, recalibrated, and subjected to the same analysis as outlined here allowing for comparisons among sites. It is acknowledged that for some of the older published ages, especially those from the BCF, there may be inconsistencies in the correction for isotope fractionation. Since for the time of interest here that effect is on the order of decades, we argue that our consideration of the larger uncertainty of the marine correction factor in the discussion will accommodate any issues derived from the isotope fractionation.

3.0 Results

3.1 Provenance

Our survey of clast lithology identified forty-two unique rock types (Table S1), but 91% of the clasts fall into one of six general rock groups: granites, leucocratic-mafic dike lithologies, greywacke, phyllite, mudstone, and *mélange* lithologies (Fig. S1). Sample sites within the Knik Valley yielded only Chugach Terrane rocks, such as greywacke, phyllite, and *mélange* lithologies. In contrast, sample sites in the Matanuska Valley yielded a combination of Chugach and Peninsular Terrane rocks, with the specific Peninsular lithologies puddingstone (a distinctive pebbly mudstone) and siderite making these samples distinctive as Matanuska Valley tracers. Of our 131 provenance stations, 58% contained up to 8 tracer clasts with a mean of 3 tracer clasts, all of which are located north of Knik Arm (Fig. 2). In contrast, none of our 23 provenance stations south of Knik Arm contained these rock types.

In addition to the Matanuska and Knik Valleys, we also examined the lithologic provenance of the BCF. These marine deposits were emplaced during the retreat of a calving ice margin. We collected 200 dropstones from the upper part of the BCF where it is exposed along the Knik Arm under the Elmendorf Moraine (Fig. 4). No puddingstone or siderite rocks were identified in the BCF. Instead, the BCF dropstones were predominantly granite, greywacke, phyllite and banded metavolcanic rocks, lithologies common to the Chugach Terrane that outcrops along the Knik Valley.

3.2 Geomorphology

Geomorphic mappings from previous studies are assembled and discussed below. Glacial-geomorphic features relevant to the deglacial pattern are concentrated in the Anchorage Lowland to the north and west of the Knik Arm. Glacial bedforms include drumlins, ice-molded bedrock, flutes, and streamlined basal terrain, with ice-stagnation features such as large esker complexes, kames, and kettled terrain. Preservation of landforms in the Knik Valley is poor due to Holocene outburst floods and fluvial erosion (Post and Mayo, 1971), as well as mass wasting. Streamlined landforms, ice stagnation features, and lateral moraines are common in the Anchorage Lowland and to the northeast into the lower Matanuska Valley (Schmoll and Dobrovolsky, 1972).

3.2.1 Eskers and Stagnation Deposits:

Eskers and stagnation deposits such as kames, hummocky till, and kettled drift cover roughly 970 km², just over 50%, of the area of the lowland north of the Knik Arm. This assemblage of depositional landforms is dominated by more than 360 individual northeast- to west-southwest trending esker ridges (Fig. 3; see also Reger and Updike, 1983). The eskers are concentrated into two major fan-shaped areas or swarms: one is a few kilometers inboard of the Elmendorf Moraine and a second is approximately 30 km up the flowline to the northeast (Fig. 3). Broadly, each of these areas includes many large eskers with interspersed kettles and hummocky drift features. Individual esker ridges range from 1 km to just over 50 km long with heights of up to 25 m. The western (closer to Elmendorf Moraine) esker area is approximately 250 km² in area and extends toward a pitted outwash plain to the southwest of the Elmendorf Moraine. The eastern (farthest from the Elmendorf Moraine) esker area extends to the lower reaches of the Matanuska Valley. A smaller number of eskers are also found in the Matanuska Valley forming narrower ridges on the valley floor. Additional ice stagnation features are observed along the post-glacial Palmer Terrace (37 m), which is accentuated with kettles.

In contrast, stagnation deposits are only found within a few kilometers of the region of the Elmendorf Moraine proximal to current Knik Arm. Farther up the flowline along the Knik Arm, streamlined bedforms such as drumlinized and fluted ground moraine are the most common depositional features (Fig. 3d). Although this terrain includes some minor intervening kames and kame terraces (Yehle et al., 1990) and rare small eskers 1 km long and <10 meters high, the character of the landscape is markedly different than the region to the north and west of Knik Arm.

3.2.2 Till Ridges:

The area between the two esker clusters to the north of the Knik Arm is primarily till covered and is characterized by streamlined till bedforms and till ridges oriented transverse to flow (Fig. 3). Previous studies and our field observations document more than 20 flow-transverse ridges covering an area just over 100 km². The ridges have a maximum relief of 34 m, with continuous crests up to 7 km long. Test pits and natural sediment exposures demonstrate that they are primarily composed of till, with sub-rounded striated cobbles and boulders, and are

locally superimposed by smaller gravel ridges of 1 to 2 meters in height transverse to ice flow. The origin of these features remains unclear, although speculation has included ribbed moraines (Reger and Updike, 1983), flood features (Wiedmer et al., 2008), as well as remnants of recessional end moraines or basal-crevasse fillings (Reger pers. comm., 2015).

3.2.3 Moraines:

The Elmendorf Moraine forms a prominent ice marginal deposit that spans 104 km along the western and southwestern margin of the Matanuska-Knik glacier. Over the region north of the Knik Arm, the moraine has an average relief of 80-90 m, with a width varying between 2 and 9 km. Along the segment south of the Knik Arm, the Elmendorf Moraine is smaller, with a topographic relief of ~40 m and a width of ~2 km. Our field observations document that the Elmendorf Moraine (Fig. 3) is composed of till overlain in places by ice-stagnation deposits. In contrast to the abundant streamlined landforms in the lowland, our field investigation failed to reveal any well-defined recessional moraines in the lowland or the valleys. Lateral moraines are present along the Talkeetna and Chugach mountain fronts adjacent to the Anchorage lowland (Reger and Updike, 1983; Schmoll et al., 1999).

3.2.4 Fjord:

The Knik Arm is an estuarine fjord occupying the distal (southern) half of the former Matanuska-Knik Glacier footprint. A seismic profile and offshore drilling in Knik Arm (Fig. 3, profile X-X') reveals a U-shaped trough cut into till to a depth of 60 m below modern sea level, filled by up to 43 m of glaciolacustrine and Holocene marine sediments (Shannon and Wilson, 2005). The coastal bluffs along the Knik Arm expose the glaciomarine BCF, which is overlain by the drift of the Elmendorf Moraine (Schmoll et al. 1972, 1999).

3.3 Chronology:

Broadly, our new chronology provides new insight into the timing and style of retreat of the Matanuska-Knik Lobe over ~140 km between ~17-14 ka BP (Fig. 2; Table 1). The oldest post-glacial material we collected is from Otter Lake near the southern edge of the Elmendorf Moraine and yields an age of 16.9-16.4 cal ka BP (Beta-237552), providing the closest minimum constraint on moraine deposition. Younger ages in the area are derived from wood fragments collected 25 cm above the basal contact at Redshirt Lake (14.6-14.1 cal ka BP, UC-29777,) and 5 cm above the basal contact at Long Lake (14.5-14.1 cal ka BP, UC-29778).

The oldest age north of the Knik Arm is 15.4-14.7 cal ka BP, collected at Reed Lake (ETH-32186). A wood sample from Wolverine Lake yields an age of 14.2-13.8 cal ka BP (Beta-237553). All other ages north of the Knik arm in the lowland area are younger ranging from 10.4-10.3 cal ka BP at Canoe Lake (Beta-237554) to 13.6-13.4 cal ka BP in Loon Lake (Beta-237559). The deglaciation pattern south of the Knik Arm is less well established, as lake basins are far fewer. The oldest age south of the Knik Arm is from Beach Lake, ~15 km behind the Elmendorf Moraine, and it is 16.2-15.8 cal ka BP (Beta-237556).

4.0 Discussion and Interpretation

Here we interpret and discuss the two glacier lobes, reconstruct the timing of the Elmendorf readvance and then estimate the timing of inland glacier retreat. Then, we consider the readvance and retreat in the context of other glacial systems in Alaska and western North America.

4.1 Matanuska and Knik Glacier Ice Flow Pathways:

Our provenance survey identifies distinctive Peninsular lithologies such as puddingstone and siderite solely at provenance stations north of Knik Arm, suggesting that the ice in this region was sourced from the Matanuska catchment. In addition, we collected clasts from the BCF marine deposits yielded no puddingstone or siderite rocks and matched the pebble lithology of sites south of Knik Arm suggesting that the BCF is associated only with ice flow sourced in the Knik catchment. Thus, we interpret the northern edge of the Knik Arm to be the approximate location of the suture line between the two ice lobes sourced in the Matanuska and Knik Valleys (Fig. 4).

4.2 The timing of advance of the Matanuska-Knik Glacier System

The timing of Elmendorf Moraine deposition is bracketed by radiocarbon ages from shells in glaciolacustrine sediments of the BCF, and by radiocarbon ages derived from lake cores on and inboard of the moraine. However, as a result of various factors, the ages from stratigraphically above and below the moraine overlap. While this overlap likely indicates the deposition of the moraine occurred during a brief interval of time, it makes identifying the precise timing of readvance challenging.

The maximum ages of the Elmendorf Moraine are derived from marine shells. Although there are nine ages from the BCF, only one dated shell lies physically under the Elmendorf Moraine. Therefore, for the maximum bracket, we can either take the total range of radiocarbon dates from the BCF (18.5-16.6 to 16.0-15.3 cal ka BP, Table 1) associated with a macrofossil rich horizon within the marine sediments (Schmoll et al., 1972), or the single shell date of 17.5-14.3 cal BP (W-2389) from directly below the moraine inferred to correlate to the same macrofossil rich horizon (Schmoll et al., 1972, Reger et al. 1995). The former approach assumes the organic horizon has the same age throughout the formation, while the latter relies on an age proximal to the moraine. In the former, any difference in ages is reflected in the variability in the individual samples; the latter provides a tighter stratigraphic context but is based on only one analysis and the error associated with it.

For the minimum bracket, the simplest approach is to employ the oldest radiocarbon age recovered from postglacial sediments. Long Lake, Redshirt Lake, Otter Lake, Lorraine Lake, and Cairn Point (Fig. 2; Table 1) all lie within 5 km of the Elmendorf Moraine. Of these nearby sites, Otter Lake is the oldest as it yielded tree twigs less than 5 cm from the non-glacial/ glacial

contact that provides an age of 16.9-16.4 cal ka BP. The next oldest being Beach Lake at 16.2-15.8 cal ka BP.

The choice of either maximum age, from the single shell sample or range of all shells, overlaps with the minimum age from Otter Lake. Possible explanations include: 1) the Otter Lake organics were deposited before the Elmendorf Moraine was created; 2) the marine-correction factor (taken here as 405 yr; Bard et al., 1990) for the shell ages is not correct, and that a larger local ΔR must be applied; or 3) that the deposition of the marine unit, the deposition of the moraine, and the establishment of organic material at Otter Lake occurred within the resolution of our radiocarbon chronology. We believe the first possibility to be implausible, as it violates geomorphic principles, leaving both 2 and 3 as possible scenarios.

The minimum - maximum approach employed here depends on the shapes of the two input PDFs. Since these are determined by the individual samples included, we present multiple cases to show the effects of these choices on the age estimate before proposing our preferred interpretation. First, we examine the different results if we use either of two cases for the maximum, all nine BCF ages or sample W-2389 only. As the radiocarbon ages are based on marine material, either scenario depends on the values of marine corrections for the estuarine setting in the Anchorage area in the late glacial. We note here that marine correction factors for this site are not well established in part because of dynamic ocean circulation during this time. We have identified three possible marine correction estimates: 1) a marine correction of 0 yr with the argument that the environment was dominated by meltwater, and meltwater does not have significant old dissolved carbonate; 2) take the standard correction of 405 yr as it allows comparison; or 3) consult the marine reservoir correction database (<http://intcal.qub.ac.uk/marine/>, accessed June 5th, 2015) and use the three closest entries to Anchorage to derive an additional correction of 415 ± 60 yr (McNeely et al., 2006). Thus to estimate the age of the readvance that formed the Elmendorf Moraine, we calculate six different cases (two variations in the included samples and three possible marine corrections), cover the probable range of possibilities (Table 2).

Our estimation algorithm computes the most probable age between the inputted minimum and maximum PDF's. Thus we report the minimum (at 95% level), the best, and maximum (at the 95% level; Table 2). When all BCF ages are used the central estimates range from 16.7 to 16.2 cal ka BP. In contrast, when using only sample W-2389 (Schmoll et al., 1972), the central estimate is slightly older and ranges from 16.9 to 16.7 cal ka BP. Our preference is to reduce any uncertainty from stratigraphic issues, thus we prefer the single maximum bracket. Moreover, we adopt the standard marine correction to be consistent with convention. For further discussion then, we provisionally adopt an age for formation of the Knik segment of the Elmendorf Moraine as 16.8 cal ka BP with a 95% confidence level that the true age lies between 17.7 and 16.5 cal ka BP (case 2b, Table 2). As we can see no geomorphic break, cross-cutting relationships, discordant outwash fans, and because there is no chronologic evidence to the contrary, we take the age of the Knik segment where shells were collected for radiocarbon analysis to be the same

as the Matanuska segment of the Elmendorf Moraine, and thus we apply our age determination to the whole moraine.

4.3 Regional context of the Elmendorf advance

The Elmendorf Moraine appears to have formed at the same time as moraines in several other areas. In Northern Alaska, a ^{10}Be chronology from the Brooks Range documents moraine deposition at ~17 cal ka BP, the “late Itkillik II re-advance” (Pendleton et al., 2015). Indeed, this record documents a similar pattern of deglaciation, with readvances at our site and in the Brooks Range that is coincident with inferred cooling across Beringia from ~18-17 cal ka BP (Clark et al., 2012). Further abroad, on the southwestern margin of the Cordilleran Ice Sheet, the Puget Lobe advanced at 19.4 cal ka BP reaching its maximum extent at 17.3 cal ka BP (Porter and Swanson, 1998; ages recalibrated, see supplemental material). The Des Moines Lobe of the Laurentide Ice Sheet, 2,300 km east of the Puget Lobe, reached its southernmost limit after 16.8 cal ka BP (Lowell et al. 1999), with basal ages above the till indicating retreat began prior to 16.3 cal ka BP (Lepper et al. 2007). This collection of ages documents a pattern of glacier expansion superimposed on a general pattern of retreat during the early stages of the most recent termination in the mid-high latitudes of the North America.

Specifically, the spatial distribution of radiocarbon ages in the Puget Lobe indicates that glacier expansion was underway for at least 2,000 years prior to the maximum being reached under positive mass accumulation conditions. The maximum ice positions, represented by moraines, are often associated with periods of mass balance equilibrium. However, an alternative possibility is that age of the maximum extent represents the transition in mass balance conditions from positive to negative (Lowell et al. 1999). We point out that the maximum extents in the Alaska Lowland, Puget Sound, and near Des Moines, Iowa were maintained only for brief intervals of time. Thus we suggest these moraines reflect the transition from positive to negative mass balance, rather than a period of positive mass balance. Therefore, we interpret the coincident moraine emplacement at ~17 ka to indicate a mass balance shift from positive to negative in Alaskan alpine glaciers, as well as sectors of the Cordilleran and Laurentide Ice Sheets.

Collectively these glacier expansions, occurring from the Brooks Range to the north-central US and reflecting independent alpine glaciers to ice sheets, took place during the first half of the Mystery Interval. Broecker and Putnam (2012) call this time the Big Dry reflecting falling pluvial lake levels in the Great Basin of the western US. While the climatic mechanism responsible for the Big Dry remains enigmatic, we note a correlation to a shift in the mass balance of glacier systems in the mid to high latitudes of North America. Thus the climatic mechanism responsible for the Big Dry must also account for a climatic shift significant enough to reverse glacier expansion ~17 cal ka BP.

4.4 Geomorphology of the collapse of the Matanuska-Knik Glacier Lobe

The collapse of the Matanuska-Knik Glacier lobes is consistent with a transition to warming at 16.8 cal ka BP. After the Elmendorf Moraine had formed, the contributing lobes underwent a rapid collapse as reflected in the geomorphology. We believe the different style of deglaciation of the two lobes can be attributed to terrestrial retreat in the Matanuska lobe versus the marine-calving in the Knik lobe. Although the retreat of the Matanuska and Knik lobes appear to have been partly driven by different processes, the present chronology cannot resolve any difference in timing.

We place our chronologic results in the context of the geomorphology of the Matanuska Valley, making connections between glacial-geomorphic features in the Anchorage Lowlands and the style of recession implicated by the chronology. Two large regions of eskers exist: one swarm is found terminating at the Elmendorf Moraine, with a second emanating from the mouth of the Matanuska Valley (Fig. 6). Preservation of both large and small eskers over the span of each esker swarm is diagnostic of ice disintegration rather than systematic ice retreat (Lauriol and Gray, 1987; Aylsworth and Shilts 1989). However, esker swarms with high concentrations of eskers are unusual and may indicate massive down wasting (Brennand and Sharpe, 1993). Furthermore, the individual eskers are distinct, indicating that the glacier cover remained in place long enough to allow the two large swarms to form. The presence of esker swarms, kettles, and lack of organized ice marginal features leads us to conclude that the Matanuska lobe collapsed in place. In contrast, we interpret the distribution of ground moraine and drumlinized landforms along the Knik flowline to indicate that the Knik ice lobe continued actively flowing during deglaciation, rather than stagnating in place.

There are multiple possible scenarios for esker creation beneath the Matanuska Lobe, though all require a source of water that is routed to the glacier's base. One possibility is the subglacial drainage of a glacially dammed lake in the Copper River Valley (Williams and Galloway, 1986; Wiedmer et al. 2008). This scenario depends on the lake waters reaching Tahneta Pass east of the present Matanuska Glacier at an elevation of 907 m (Williams, and Galloway, 1986). Additionally, ice in both the Nelchina and Tazlina Valleys must be retracted as the pass lies west of both of these valleys. Weidmer et al. (2011) argue this is unlikely, as he identifies the Matanuska Valley eskers and till ridges as dunes indicative of a mega-flood. We feel that until more evidence emerges from this area on the timing and nature of a possible lake drainage event, the formation of these two separate swarms by a single mega-flood is problematic. Rather, we suggest these esker swarms represent two events of meltwater originating from meltwater production within the Matanuska basin. This basin is large enough to create the necessary meltwater, and the geometry is such that meltwater would funnel directly into the Anchorage lowland. For a more detailed treatment of the possible flood or lack thereof, we suggest reading the comment and reply on the topic (Reger et al., 2011; Weidmer et al., 2011).

4.5 Timing of glacier collapse of the Matanuska-Knik Glacier

The distribution of our chronologic control points identifies two intervals of retreat. For the oldest, we take the age of the Elmendorf Moraine to be the oldest control point at 16.8 cal ka BP, and use Reed Lake (15.4-14.7 cal ka BP; ETH-32186; 49 km up ice) and Wolverine Lake (14.2-13.8 cal ka BP; Beta -237553; 72 km up ice) as the closest controls on glacier retreat up valley (Table 1, Fig. 5). Wolverine Lake falls up valley of both esker swarms, thus providing a minimum age on the deposition of the up-ice esker swarm. Given the distinct swarms of eskers, it would appear there were two generations of meltwater flow after 16.8 and before 14.0 cal ka BP during periods of ameliorating climate.

For the younger interval, radiocarbon ages from Reed Lake and the Matanuska Bog near the present glacier margin provide bracketing constraints. The ages on the up valley position are derived from both radiometric and AMS dating techniques, and range from 15.9-15.4 to 13.4-13.32 cal ka BP (Table 1; Williams, 1986; Ager, unpublished). The older ages, if correct, would imply near instantaneous retreat across the 80 kilometers between Reed Lake to the Matanuska bog.

While we cannot demonstrate that the older radiometric ages are incorrect, to avoid potential issues with contamination from local coal deposits, and to maintain instrumental consistency with the ages established down valley, we only utilized AMS ages. For this reason, the two unpublished AMS ages of Tom Ager of 13.8-13.6 cal ka BP (WW-3623) and 13.4-13.3 cal ka BP (WW-3618) are considered the best brackets for the northeastern most section of Matanuska retreat (Table 1). This analysis yields an average retreat rate during the initial phase of glacier recession of 65 m/a and an average rate of 57 m/a during the later retreat phase, rates of retreat comparable to other terrestrial glacier systems (Kelley et al., 2015). Given the stratigraphic context, these are minimum estimates the pace of a retreat that lasted at least 3.1 cal ka.

The retreat of ice across the Knik footprint is less well-constrained than along the Matanuska flow line due to the paucity of ages. A basal age from Otter Lake, 5 km up flowline, 16.9-16.4 cal ka BP, overlaps with our estimate for the age of the Elmendorf Moraine, suggesting the retreat off the moraine occurred rapidly. At Beach Lake, 22 km up flow, the basal age is at 16.2-15.8 cal ka BP. These ages are older than from lakes a similar distance from the Elmendorf Moraine along the Matanuska flowline. This contrast allows for three possible scenarios: 1) retreat initiated sooner at the head of the Knik lobe than at the Matanuska lobe; 2) retreat across the Knik lobe progressed faster than across the Matanuska lobe; 3) the retreat was the same in both glacier systems, but minimum limiting chronology are closer to the actual timing of deglaciation in the Knik lobe. We feel that the geomorphic evidence in the two valleys makes scenario 1 unlikely (see the previous discussion of Elmendorf Moraine). Scenario 3 is possible. However, we favor scenario 2, as the drumlinized and fluted ground moraine present in the Knik lobe suggest marked frontal retreat occurred, in contrast to the stagnation deposits indicating slower retreat through the Matanuska Lobe. We believe this faster retreat was facilitated by calving in the deeper waters of the Knik Arm.

4.6 Context for the Matanuska-Knik Glacier Lobe retreat

The marine record from the Gulf of Alaska supports our timing for the recession of the Matanuska-Knik Glacier. Planktonic foraminifera records document an increase in freshwater input starting at 16.9-16.5 cal ka BP and continuing till 13.9-13.7 cal ka BP, implying and increase in regional glacier melt (Davies et al., 2011). Furthermore, the same cores display a transition from ice-proximal sediments to more hemipelagic sedimentation at 15.2-14.4 cal ka BP, possibly reflecting the regional retreat of marine based glaciers, such as the Knik lobe, on to land, thus decreasing its sediment contribution to the Gulf of Alaska (Davies et al., 2011).

The synchronicity of retreat between glaciers in the Anchorage Lowlands, Brooks Range, and Puget Lobe (Fig. 7) is marked, with all three sites undergoing retreat after synchronous moraine deposition. The timing of retreat of western North American glacier systems corresponds with warming in the southern mid-latitudes, where climate reached near-interglacial conditions by 16.8 ka BP (Moreno et al., 2015), mountain glaciers on the South Island of New Zealand began to recede at 17.7 cal ka BP (Putnam et al., 2013), and glaciers in southernmost South America collapsed at ~16.8 cal ka BP (Hall et al., 2013). While retreat of both the Laurentide and Cordilleran Ice Sheets following the LGM appear concordant with steadily rising temperatures recorded in Antarctica and the Southern Hemisphere, a comparable warming during this time is not recorded in the Greenland cores (Fig. 8; Brook et al., 2005). The $\delta^{18}\text{O}$ data suggest a slight warming may have been recorded in Greenland around 24 cal ka BP (Alley et al. 2002), but the onset of the dramatic Bølling warming did not occur until much later, around 14.7 cal ka BP (Severinghaus and Brook 1999). The onset of any local expression of the Bölling warming is clearly much later than the onset of the collapse of the Matanuska-Knik glacier, mountain glaciers in the Brooks Range, the Puget, and Des Moines Lobes, and other glacier systems worldwide. Our glacial reconstruction from the high latitudes, placed in context with other glacial records, show large-scale ice collapse in the Anchorage Lowland is concordant with middle-latitude glaciers worldwide. This growing dataset fits well with the concept of a globally synchronous glacial termination (e.g. Denton et al., 2010) that started at 17 cal ka BP.

5.0 Conclusions

Ice advance into the Anchorage Lowland is bracketed by radiocarbon ages on marine shells from stratigraphically below the advance till (Reger et al. 1995; Schmoll et al. 1999), and on material recovered in lake cores from stratigraphically above the moraine. The marine shells, which form the oldest bound on moraine emplacement, yield a peak summed probability distribution ages of 16.8 cal ka BP. Retreat in the Anchorage Lowland, constrained by radiocarbon ages from lake cores, was underway prior to 16.4 cal ka BP. Clast provenance data delineates a paleo-suture line between the coalesced Matanuska and Knik lobes. Ice on the southern sector along the Knik route was in contact with the marine waters of the Knik Arm, where ice calved into at least 47 m of water. This calving margin had retreated at least 20-25 km by 15.8 cal ka BP.

Along the Matanuska flowline, our geomorphologic analysis argues for a two-phased retreat. Specifically, the leading edges of two large esker swarms are interpreted to mark two ice terminal positions. Initial retreat from the moraine at 16.8 cal ka BP, crossing the first down ice esker swarm by 14.9 cal ka BP. Ice disintegration was well underway in the second esker swarm near the Matanuska Valley before 14.9 cal ka BP, and the ice margin had retreated out of the Anchorage Lowlands and into the Matanuska Valley by 14.0 cal ka BP. The data show that Matanuska ice retreated from the entrance of the Matanuska Valley ~70 km to the modern terminus by 13.7 cal ka BP. This second step of retreat at 14.9 - 13.7 ka BP is consistent with records of global ice collapse during the Bølling (Ridge et al., 2012; Hughes et al., 2015; Kelly et al., 2016), yet the climate event responsible for the initial retreat off the Elmendorf Moraine is more enigmatic. Our findings demonstrate that late glacial events in the Anchorage Lowland are not an isolated Alaskan phenomenon; rather, they contribute to a growing dataset, which cumulatively supports the notion that major glacial terminations are globally driven events.

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Figure Captions

Figure 1. Shaded relief INSAR image of the study area, with an inset map showing the position (white box) of our study area within Alaska. Transparent white overlay demarks area contributing to the last major advance of the Matanuska-Knik Lobe of the Cordilleran Ice Sheet (12,360 km²). Dashed line traces the Border Ranges fault separating the Chugach Terrane (south) from the Peninsular Terrane (north). Present glacier extent in white. Imagery from U.S. Geological Survey EROS Alaska Field Office.

Figure 2. Location of samples collected for radiocarbon dating. White boxes contain key calibrated radiocarbon ages, with all sample sites from this study noted, numbers are keyed to Table 1 for all ages. The inset map shows the location of marine shell samples used for the maximum age constraint, with letters denoting the collection location of specific samples: GX-20127 (A); AA-2227(B); GX-20128 (C); GX19989 (D); W-2367 (E); AA-2226 (F); W-2389 (G); W-2369 (H); W-2151 (I). Inset map LiDAR, from the Alaska Division of Geological & Geophysical Surveys.

Figure 3. Shaded relief INSAR image of the study area with key geomorphic features labeled. Heavy dashed line denotes the maximum extent of the Elmendorf advance, with light dashed lines outlining two esker swarms. White boxes correspond to LIDAR insets of till ridges (A), lateral moraines (B), esker complex, containing some kettles and hummocky drift deposits (C), and drumlins and flutes (D). Also, note the second prominent drumlin field south of Knik Arm. Black line (X to X') denotes the position of the seismic profile across Knik Arm shown in (E).

Figure 4. Dispersal of lithologic tracers found at sample stations (black circles Peninsular Terrane pebbles found, open circles indicate stations with no Peninsular clasts). The short dashed lines indicate the location of the Matanuska-Knik Lobe catchment and the heavy gray line delineates the provenance boundary interpreted as the approximate boundary between the Matanuska catchment to the north and west and Knik catchment to the south and east. Locations of Peninsular Terrane outcrops denoted as stippled gray polygons after mapping by Winkler (1992).

Figure 5. Interpreted pattern of glacier advance and retreat. Ages used to constrain our interpretation are shown as probability distributions. The retreat can be no younger but could be older than shown. Age of the Elmendorf Moraine derived as discussed in the text.

Figure 6. Conceptualized model of ice retreat in the Knik and Matanuska Lobes. The two-step retreat of the lowland is indicated by downstream limits of two esker swarms where the ice terminus is interpreted to have stabilized during the retreat. Of particular interest are those drainage channels (purple) flowing into the north side of the Knik Arm from the Matanuska Lobe. These channels are deeply incised into the glacial drift in front of the leading edge of the second esker swarm. The inset map depicts a simplified chronology of retreat within the two lobes.

Figure 7. Comparison probability distribution functions on the timing of major glacial advances in Alaska and the Cordilleran. From left to right: the Matanuska-Knik glaciers (this study), the Puget Lobe (Porter and Swanson, 1998), alpine glacier in the Brooks Range (Pendleton et al.,

2015). For each system, the left panel plots the raw ages used whereas the right panel plots the probability curves. Matanuska-Knik and Puget Lobe systems there are radiocarbon ages from minimum (red) and maximum (blue) stratigraphic positions. For the Brooks Range, the chronology is based on ^{10}Be dating. The summed probabilities for the minimum and maximum sets are shown in respective colors. The black maximum data only includes W-2389 because of its location below the Elmendorf Moraine. However, an alternative model including all ages from the Bootlegger Cover Formation is shown as the blue dashed probability line. In that same panel, the horizontal dashed lines represent the extreme limits of the marine correction sensitivity test reported in Table 2. Details for samples used from the Puget Lobe and Brooks Range datasets can be found in Table S3.

Figure 8. Comparison of the Matanuska, Des Moines, and Puget Lobes retreat patterns with ice core records. The retreat in all lobes begins before any documented warming in the NGRIP ice core and well before the Bölling warming ~14.7 ka (Severinghaus and Brook 1999). The final retreat of the ice margin in the Matanuska Valley to near its present location is accomplished during the Bölling time. Calving controls some of the retreat in the Puget Lowland, nevertheless the overall patterns of these diverse areas are more similar to the pattern of temperature changes recorded in Antarctica than the pattern recorded in Greenland (Brook et al., 2005). Retreat patterns adopted from the Puget lowland from Porter and Swanson (1998) and the Des Moines Lobe from Lepper et al. (2007).

Table 1: Mantanuska-Knik Chronology

Site ID, Locality	Lab ID	Latitude ^	Longitude ^	Cm above contact	Material	Age ± 1σ 14C yrs BP	Calibrated Range (cal yr BP)	Median cal.	Reference
								solution ± 2σ	
1, Bootlegger	W-2151	61.201°	-149.998°	n/a	marine shell	13,690 ±400	17190-14600	15,940±1290	Schmoll et al., 1972
2, Bootlegger	W-2369	61.201°	-149.998°	n/a	marine shell	14,900 ±350	18470-16650	17,620±910	Schmoll et al., 1972
3, Bootlegger	W-2367	61.243°	-149.979°	n/a	marine shell	14,300 ±350	17780-15870	16,830±960	Schmoll et al., 1972
4, Bootlegger	W-2389	61.280°	-149.921°	n/a	marine shell	13,750 ±500	17450-14340	16,010±1560	Schmoll et al., 1972
5, Bootlegger	AA-2226	61.257°	-149.941°	n/a	marine shell	14,308 ±140	17290-16370	16,840±460	Reger et al., 1995
6, Bootlegger	GX-19989	61.240°	-149.986°	n/a	marine shell	14,100 ±90	16870-16240	16,530±320	Reger et al., 1995
7, Bootlegger	GX-20128	61.239°	-149.990°	n/a	marine shell	14,078 ±214	17150-15920	16,520±630	Reger et al., 1995
8, Bootlegger	AA-2227	61.242°	-149.998°	n/a	marine shell	13,470 ±120	16020-15270	15,650±370	Reger et al., 1995
9, Bootlegger	GX-20127	61.249°	-150.041°	n/a	marine shell	13,994 ±90	16700-16090	16,380±310	Reger et al., 1995
10, Cairn Point	W-2589	61.257°	-149.876°	n/a	basal bog age	12,350 ±350	15580-13540	14,500±1044	Yehle et al., 1990
11, Lorraine Lake	n/a	61.293°	-149.952°	23 cm > basal	seed	12,755±45	15360-15040	15,200±160	Kathan et al., 2004
12, Otter Lake	Beta -237552	61.288°	-149.736°	5 cm > basal 1 cm > bottom of core	twig fragments	13,790 ±50	16930-16430	16,680±250	This study^
13, Walden Lake**	Beta -237558	61.349°	-149.631°		seeds	9,940 ±40	11410-11240	11,340±180	This study^
14, Knik Lake	Beta -237557	61.460°	-149.735°	At basal	seed/stem frag	11,610 ±40	13560-13340	13,440±110	This study^
15, Beach Lake	Beta -237556	61.406°	-149.556°	At basal	wood/ stem frag	13,330 ±50	16230-15840	16,040±200	This study^
16, Mirror Lake	Beta -237560	61.427°	-149.415°	35 cm > basal	wood/ stem frag	10,120 ±60	12010-11590	11,740±300	This study^
17, Hay Flats	GX-15241	61.520°	-149.254°	n/a	organic	11,400 ±720	15320-11270	13,330±2020	Combellick,1990
18, Redshirt Lake	UC-29777	61.604°	-150.199°	25 cm > basal	wood	12,345 ±35	14640-14120	14,320±260	This study^
18, Redshirt Lake	ETH-3218 5	61.604°	-150.199°	25 cm > basal	wood	12,180 ±110	14560-13750	14,080±410	This study^
19, Long Lake	UC-29778	61.711°	-150.086°	5.5 cm > basal;	wood	12,305 ±35	14510-14070	14,230±220	This study^
20, Loon Lake	Beta -237559	61.601°	-149.750°	12 cm > basal At bottom of core	wood	11,680 ± 40	13580-13440	13,510±70	This study
21, Reed Lake	ETH-32186	61.663°	-149.317°		wood	12,680 ±90	15390-14650	15,070±370	This study
22, Little Su	GX-5019	61.710°	-149.230°	n/a at bottom of core	organic	9,155 ±215	10870-9680	10,340±700	Reger & Updike, 1983
23, Canoe Lake **	Beta -237554	61.558°	-149.191°		wood/ stem frag	9,200 ±40	10440-10250	10,360±120	This study^
24, Wolverine Lake	Beta -237553	61.664°	-148.968°	16 cm > basal	wood	12,150 ±50	14170-13830	14,030±170	This study^
25, Fish Lake	Beta -237555	61.783°	-148.568°	At basal	Trigg frags	10,350 ±40	12390-12030	12,210±180	This study^
26, Hundred Mile Lake	ETH-29990	61.811°	-147.836°	25 cm > basal	Shell *	11,320 ±85	13340-13040	12,790±180	Yu et al., 2008
27, Matanuska Glacier bog	USGS-2175	61.796°	-147.804°	n/a	peat	13,100 ±60	15970-15430	15,720±270	Williams, 1986
27, Matanuska Glacier bog	Beta-11174	61.796°	-147.804°	n/a	peat	12,210 ±120	14690-13760	14,140±460	Williams, 1986
27, Matanuska Glacier bog	WW-3623	61.796°	-147.804°	10 cm > basal	seed	11,810 ±40	13750-13550	13,650±130	Ager, unpublished
27, Matanuska Glacier bog	WW-3618	61.796°	-147.804°	10 cm > basal	seed	11,495 ±40	13440-13260	13,350±90	Ager, unpublished

Note: * hardwater effect accounted for by analysis of nearby terrestrial matter; ** Indicates the basal contact at this site was not encountered in drilling. ^Italic text indicates coordinates were estimated from publication.

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