Examination of the interplay between glacial processes and exhumation in the Saint Elias Mountains, Alaska

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ABSTRACT

The combination of large, temperate glaciers and rapid crustal convergence in the Saint Elias Mountains (southeastern Alaska, USA, and Yukon Territory and British Colombia, Canada) provides an exceptional opportunity to study the interactions between the tectonic and surface processes that have shaped most active orogens on Earth during much of the Quaternary. This research first provides a review of thermochronometric data sets recording exhumation under two major glacier systems of the Saint Elias Mountains, the Bagley-Bering and the Seward-Malaspina systems. These data sets are integrated over the single glacier systems and used in conjunction with glaciological data to investigate the interactions of glacial erosion and tectonics. Despite their proximity, the glaciological processes and geological settings of these two glacial systems differ significantly. On the east side of the orogen, sediments eroded from bedrock underneath the Malaspina Glacier reflect regions of rapid erosion under the slowly moving Seward Ice Field. Because the Seward Ice Field overlies a localized zone of major faulting and rapid exhumation, the strained and fractured bedrock is primed for erosion. On the west side, the Bering Glacier is the primary outlet for the Bagley Ice Field, which covers half of the crest of the orogen; however, few if any of the sediments at its terminus originate from under the Bagley Ice Field. Sediment transport is likely hindered by subglacial freeze-on processes that reduce the sediment-carrying capacity of subglacial rivers, though glacial surges are likely exceptions that deposit sediment far beyond the active margin of the glacier. Our study concludes that the widely invoked concepts of glacial erosion should be used with caution, as oversimplification can fail to account for important site-specific differences in geologic and glacial conditions.

INTRODUCTION

The interactions among climate, tectonics, and surface processes in active orogens have received much attention through a variety of modeling efforts and field investigations (e.g., Molnar and England, 1990; Beaumont et al., 1992, 2001; Koons, 1995; Zeitler et al., 2001; Wobus et al., 2003; Bookhagen et al., 2005; Hooks et al., 2009). Erosion can significantly influence the geodynamics of active mountain belts (e.g., Brozović et al., 1997; Zeitler et al., 2001). Many studies have focused on fluvial erosion (e.g., Zeitler et al., 2001; Whipple and Meade, 2004), but the importance of glacial erosion has recently gained attention in view of (1) the exceptionally high modern erosion rates documented for many glaciers (e.g., Hallet et al., 1996; Delmas et al., 2009; Koppes and Montgomery, 2009), and (2) the role of glacial erosion in curtailing the height of mountain ranges, a concept known as the “glacial buzzsaw” that highlights the widespread association between the heights of mountains and the snowline (Brozović et al., 1997; Mitchell and Montgomery, 2006; Egholm et al., 2009). Many active orogens were heavily glaciated during much of the Pliocene–Pleistocene (e.g., Himalaya, Andes, European Alps, Southern Alps of New Zealand, northwestern Cordillera of North America), but most of these regions currently contain only small alpine glaciers and remnants of larger ice fields (e.g., Porter, 1989; Meigs and Sauber, 2000; Tomkin and Braun, 2002; Tomkin and Roe, 2007). Consequently, the studies of the coupling between glacial erosion and tectonic processes are largely based on the geomorphology of deglaciated landscapes (Brocklehurst and Whipple, 2002, 2007; Champagnac et al., 2009), and on conceptual (Whipple et al., 1999), analytical (Tomkin and Roe, 2007), and numerical models (Tomkin and Braun, 2002; Tomkin, 2007; Herman and Braun, 2008; Yanites and Ehlers, 2012). Direct field investigations of regions currently being eroded by massive glaciers have only recently gained attention (Ehlers et al., 2006; Herman et al., 2010).

The heavily glaciated Saint Elias Mountains in southeastern Alaska provide a glimpse at how many orogens likely appeared and functioned during much of the Quaternary; the area is an excellent location to study links between glacial erosion and exhumation in a tectonically active mountain belt. We first review existing data of the geology, glaciology, and thermochronometric record from the Saint Elias Mountains. Then, we reexamine the detrital data reported in Enkelmann et al. (2008, 2009) to obtain robust cooling age information over the individual glaciers. This information is then integrated with published glaciological data and new calculations of the hydrological potential for the Bagley-Bering Glacier. The purpose is to gain new insights on the interaction between glacial erosion and rock exhumation for individual glacial systems. We emphasize a number of results that merit further attention. (1) The two major glacier systems covering much of the mountain range share many similarities in terms of their size, ice dynamics, and underlying rock units, but produce significantly different detrital signals of exhumation rates, suggesting a strong influence of the different structural settings within the orogeny on the glacial erosion. (2) From the thermochronometry, the spatial pattern of exhumation rates shows little correlation with the rate of ice motion, and the most intense exhumation in the orogen occurs in a region of relatively slow ice flow.
BACKGROUND

Geology

The tectonic setting of southern Alaska is dominated by the ongoing convergence of the Yakutat terrane with North America (Fig. 1). Recent geophysical studies characterize the Yakutat terrane as buoyant overthickened (15–30 km) oceanic crust (Ferris et al., 2003; Eberhart-Phillips et al., 2006; Worthington et al., 2008, 2012; Christeson et al., 2010) that was transported northward along the dextral Fairweather fault system (Plafker et al., 1977; Plafker, 1987), causing uplift and exhumation in southern Alaska since the Late Eocene–Early Oligocene (Finzel et al., 2011; Benowitz et al., 2011). Today the Yakutat terrane is separated from the Pacific plate by the Transition fault, and from the North American continent by the Aleutian megathrust in the west, the Chugach–Saint Elias thrust fault in the north, and the Fairweather transform fault in the east (Figs. 1 and 2).

Southeastern Alaska comprises a complex mosaic of terranes that have been accreted to North America since the Mesozoic (e.g., Plafker et al., 1994). From north to south these terranes are the Wrangellia, Chugach, Prince William, and Yakutat (Fig. 2). The Border Ranges fault is the suture between the late Paleozoic–Jurassic basement of the Wrangellia terrane in the north and the Chugach terrane in the south (Fig. 2) (Plafker and Roeske, 2007). The Chugach and Prince William terranes represent a Mesozoic–Paleogene subduction accretionary complex, which was metamorphosed and intruded by the Sanak-Baranof plutonic belt during Eocene ridge subduction, forming the Chugach metamorphic complex (Plafker et al., 1977, 1994; Barker et al., 1992; Cowan, 2003; Pavlis et al., 2003; Sisson et al., 2003). Metamorphic grades decrease generally westward, from amphibolite facies gneiss and schist around the eastern Bagley Ice Field and Seward ice field, to greenschist metamorphic rocks in the west (Fig. 2).

The Chugach terrane is separated from the Prince William terrane by the Contact fault, which is thought to have initiated as a subduction thrust during the Paleogene (Plafker et al., 1994). Although the Contact fault cannot be observed directly because it is buried under the thick ice of the Bagley and Seward Ice Fields, it can be inferred with confidence from geodetic measurements, structural analysis of the surrounding region, and geological observations on isolated nunataks (e.g., Bruhn et al.,

Figure 1. Tectonic setting of southern Alaska with major tectonic structures and geographic features, including the Bering and Seward glacier systems (light blue). The Last Glacial Maximum extent is shown by the red line, and the Pleistocene maximum is shown by the blue line; both are from the work of Kaufman and Manley (2004). The Yakutat–North America plate motion vector is from global positioning system measurements by Fletcher and Freymueller (2003), and Pacific plate–North American plate motion is from Kreemer et al. (2003). CSEF—Chugach–Saint Elias fault.
Moreover, it was possibly reactivated during the Yakutat collision, most likely as a dextral compressional transform fault and the western continuation of the Fairweather fault (Fig. 2) (Savage and Lisowski, 1986; Sauber et al., 1997; Bruhn et al., 2004). The Chugach–Saint Elias thrust fault is the suture between the Prince William terrane and the Yakutat terrane in the south. During the Yakutat collision, the Paleogene to Pleistocene sedimentary cover of the Yakutat terrane was accreted to North America in the evolving fold-and-thrust belt (Fig. 2; Plafker et al., 1994). Mainly Eocene to Miocene sediments of the Kultieth and Poul Creek Formations are exposed in the northern and western part of the fold-and-thrust belt, whereas the sediments of the syncollisional Yakataga Formation (younger than 5.6 Ma) are located farther south and offshore (Plafker, 1987; Plafker et al., 1994). In contrast to the Kultieth and Poul Creek Formations that derived from rocks located several hundreds of kilometers farther south, the Yakataga Formation is derived from Chugach terrane rocks and reworked Kultieth and Poul Creek Formations (e.g., Enkelmann et al., 2008, 2010; Perry et al., 2009).
The Seward and Bagley Ice Fields occupy the orogenic spine of the Saint Elias Mountains at ~2000 m elevation (Fig. 2). The Seward Ice Field funnels ice southward through the Seward Throat (~40 km long, 4–6 km wide), a narrow passage through the prominent range that bounds most of the ice field in the south. The ice then spreads out toward the coast into the massive Malaspina piedmont glacier (Figs. 2 and 3). In the area of highest topography near Mount Logan and Mount Saint Elias, the Seward Ice Field connects with the Bagley Ice Field. The Bagley ice flows westward for >100 km before diverging into the Bering and Tana Glaciers. The Bering Glacier (~75 km long, 8–10 km wide) flows south toward the Gulf of Alaska, and the Tana Glacier (~30 km long, 3–4 km wide) drains to the north into the Chitina Valley (Fig. 2).

The syncollisional Yakataga Formation is composed of marine and glaciomarine sediments and records a long history of glaciation starting ca. 5.6 Ma (Plafker, 1987; Lagoe et al., 1989, 1993; Plafker et al., 1994). Glaciers large enough to reach the Gulf of Alaska have generally been sustained since ca. 5 Ma (Péwé, 1975). During the last major glaciation, and likely previous glaciations, the ice was substantially thicker than today and extended over much of the continental shelf (Péwé, 1975; Mann and Hamilton, 1995; Kaufman and Manley, 2004). Glaciers were sufficiently massive to reach the ocean and shed debris-laden icebergs since the latest Miocene–Pliocene, possibly during a period that was warmer than today (Péwé, 1975; Lagoe and Zellers, 1996). Evidence for Last Glacial Maximum (LGM) ice extent to the edge of the continental shelf, 150 km off the present coastline, is given by erosional features on islands in Prince William Sound and submarine features that show glacial erosion or deposition, including a series of prominent shelf-crossing sea valleys (e.g., Mann and Hamilton, 1995; Kaufman and Manley, 2004; Berger et al., 2008a).

It is possible to quantify the amount of glacial erosion by measurements of sediment strata to estimate approximate orogen-wide erosion rates over shorter time scales (10^3–10^5 yr). Based on sedimentation rates of 7.9 mm/yr, Sheaf et al. (2003) estimated minimum erosion rates of 5.1 mm/yr over the past 10,000 yr for the full Saint Elias Mountains. Jaeger et al. (1998) determined that the deposition rates from the previous 100 years vary significantly from those rates averaged over the Holocene; these rates were measured in shallow water off the coast of the Saint Elias Mountains, particularly in fjords and bays into which glaciers had extended during the LGM. However, in deep troughs that had never been overrun by glaciers, the rates on the two times scales were found to be very similar over both the past ~100 yr and the Holocene (Jaeger et al., 1998). Even considering corrections for overestimates of erosion rates due to the enhanced sediment flux carried by increased water discharge as modern glaciers retreat, the coastal Alaskan glaciers have maintained generally high erosion rates (1–10 mm/yr) for more than 10,000 yr (compiled by Koppes and Hallet, 2006; Koppes and Montgomery, 2009). Considerably less is known about erosion rates over longer time scales, 10^3–10^5 yr (Sheaf et al., 2003). These rates may have been slightly lower (2–5 mm/yr), based on thermochronometry data from the region south of the Seward and Bagley ice fields (Berger and Spotila, 2008).

On the scale of specific glaciers, sediment yield and water discharge rates have also been used to determine erosion rates that compare favorably to the regional rates. The erosion rate of the Variegated Glacier near Yakutat Bay was found to average ~3 mm/yr (Humphrey and Raymond, 1994) and Icy Bay glaciers averaged ~9 mm/yr (Koppes and Hallet, 2006). For the Seward-Malaspina system, a sediment yield of 10^6 m^3/yr, if sustained, corresponds to an effective erosion rate of ~11 mm/yr averaged over a basin area of ~5000 km^2 (Molnia et al., 1978; Hallet et al., 1996). For the Bagley-Bering system, contemporary sediment yields correspond to basin-averaged erosion rates of ~2 mm/yr based on rough estimates (Merrand and Hallet, 1996) for non-surge periods. During outburst floods associated with surges, the much higher sediment yields reflect enhanced sediment transport and more vigorous erosion, possibly corresponding to minimum erosion rates of ~10 mm/yr (Merrand and Hallet, 1996).

Figure 3. The Seward-Malaspina glacier system and glacier velocity distribution. Through the Seward Throat, the velocities range from >100 m/yr to ~2000 m/yr (adapted from Headley et al., 2012). The Seward Ice Field shows velocities >25 m/yr (adapted from Ford et al., 2003), reaching a maximum of ~100 m/yr. Mapped faults are shown as black lines, assumed faults (F.) are dashed lines. CSEF—Chugach–Saint Elias fault; UTM—Universal Transverse Mercator.
Both the Seward-Malaspina and the Bagley-Bering-Tana systems surge on a decadal to multi-decadal time scale. The surges likely result from partial or complete blockage of the subglacial (and possibly englacial) fluvial network, and the buildup of basal water that leads to fast sliding (Raymond, 1987). Surges can end abruptly when the subglacial hydrological network opens and connects with the glacier margin, characteristically releasing large volumes of sediment-laden water in outburst floods (Raymond, 1987; Paterson, 1994). This profound reorganization of the subglacial hydrological network during surges has a strong impact on the subglacial transport of sediment (Raymond, 1987). Surging is characteristic of, and has been extensively studied for, several other major glaciers in southeastern Alaska, including the nearby Variegated Glacier in the eastern part of the Saint Elias Mountains (Raymond, 1987; Lingle and Fatland, 1998).

There are other processes that contribute to the erosion and transport of sediment out of mountainous terrain. Periglacial processes, i.e., headwall erosion due to frost weathering, rock fall, and landslides, can also deliver considerable debris to the glacier surface, where it can be carried along with glacier flow and mixed with subglacially eroded sediment at the glacial terminus (e.g., Post, 1972; Lingle and Roche, 2010). Previous work on other glaciers has shown that this supraglacially-sourced sediment can account for a large range of the total sediment flux, from <10% to as much as 25%–60% for small glaciers (Syvitski, 1989; Hunter et al., 1996; Arnould and Meigs, 2005). However, both the Seward-Malaspina and Bagley-Bering-Tana systems cover the vast majority of their drainage areas (~10^3 km^2), with only small portions of the surrounding slopes exposed to subglacial processes, so supraglacial material likely accounts for only a small percentage of the total sediment load. At the bed of the glacier, once loose material is produced through erosion, subglacial rivers are largely responsible for the sediment evacuation and transport through to the rivers that emanate from the termini of many temperate glaciers (e.g., Hunter et al., 1996; Alley et al., 1997; Ruhimaki et al., 2005). For glaciers in Alaska and Canada, 80%–98% of the sediment load is transported by glacial rivers (Stravers and Syvitski, 1991; Hunter, 1994), making glacial-fluvial sediments natural targets for probing the erosional efficiency of glaciers.

**Seward-Malaspina Glacier System**

The Seward Ice Field provides the bulk of the ice forming the piedmont Malaspina Glacier (Fig. 3). Other smaller glaciers, including the Agassiz to the west (Fig. 3), also contribute to the Malaspina lobe but are not considered here. According to work by Washburn (1935), Sharp (1958), and Post (1972), the folded moraines of the Malaspina piedmont reflect a series of surges. Different regions of the lobe are expected to surge at different times, including regions influenced by the Agassiz Glacier, distorting existing moraines (Washburn, 1935; Sharp, 1958). The effects of surges between 1976 and 2006 include surface lowering, heavy crevassing, distortion of moraines on the piedmont lobe, and terminus encroachment into bordering lakes (Muskett et al., 2008).

There are excellent surface velocity measurements available for the Seward-Malaspina system. Synthetic aperture radar interferometry (InSAR) derived surface velocities through the Seward Throat were presented in Headley et al. (2012). Additional surface velocities were measured for the Seward Ice Field (Ford et al., 2003). Surface velocities increase by more than an order of magnitude as the ice from the Seward Ice Field funnels through the Seward Throat, accelerating from ~100 m/yr to >1800 m/yr in the central and lower portion of the throat (Fig. 3). While the surface velocity is well defined for the Seward-Malaspina Glacier, the ice thickness has proven difficult to measure except for through the Malaspina lobe (Conway et al., 2009). These difficulties are due to the extreme constriiction, existence of ample water, and heavy crevassing of the Seward Throat.

**Bagley-Bering-Tana Glacier System**

Much research has focused on the Bering Glacier surge cycle (discussed in detail in Brühn et al., 2010). Evidence of the surges includes drastic changes to the velocity, historical records of the glacier advancing several kilometers into proglacial Vitus Lake, and folded moraine loops on its surface (e.g., Post, 1972; Lingle and Fatland, 1998, 2003). In contrast, the Tana Glacier does not seem to surge or be heavily influenced by the Bering surges (Fatland and Lingle, 2002). The Bering Glacier has surged at least six times in the twentieth century; a well-documented surge occurred in 1993–1995 (Sauber and Molnia, 2004). These surges affect the surface profile and velocity upglacier to within 20 km of the divide from the Seward Ice Field (Lingle and Fatland, 2003). Other effects are evidenced well beyond the glacier’s margin. Sediment flux during the floods that characteristically punctuate the end of surges can exceed normal fluxes by orders of magnitude (Humphrey and Raymond, 1994; Fleisher et al., 2003), and can transport sediment hundreds of kilometers beyond the glacier front (Jaeger et al., 1998). Large outburst floods, with massive sediment delivery, were documented during the 1993–1995 surge of the Bering Glacier (Merrand and Hallet, 1996; Fleisher et al., 2003). The discharge in the Seal River, which drains proglacial Vitus Lake to the Gulf of Alaska, increased by an average of 1100 m^3/s above the base flow of 1550 m^3/s over a 6 day flood period in 1994 (Merrand and Hallet, 1996). Offshore in the Bering Trough, sediment deposits record many discrete packages of rapidly settled sediment; these packages have been interpreted as historical surges that disrupt the normal sedimentation packages (Jaeger et al., 1998; Jaeger and Nittouer, 1999).

While not nearly as extensive as on the Seward-Malaspina, ice motion has been studied through sparse global positioning system (GPS) measurements (Larsen et al., 2007; LeBlanc et al., 2008; Bruhn et al., 2010) and InSAR (Fatland and Lingle, 2002). Summer velocities range from <300 m/yr to >1000 m/yr on both the Bering Glacier and the Bagley Ice Field. However, the longer-term velocity distribution of this glacier system is difficult to determine because of seasonal variations and the surge cycles.

In contrast to the Seward-Malaspina system, the basal topography of the Bering Glacier is well defined, although is lacking for beneath most of the Bagley Ice Field. Figure 4 shows a surface elevation profile measured using airborne laser altimetry along an approximate flowline (Echelmeyer et al., 2002) and the glacier thickness measured using an airborne ice-penetrating radar over most of the Bering Glacier (Conway et al., 2009).

**Thermochronometry in the Saint Elias Mountains**

The main geological events that have been recorded in the Saint Elias orogen using geochronometry and thermochronometry are summarized in Table 1. The tectonic and erosional evolution of the Saint Elias Mountains derived from thermochronometry was reviewed in detail (see Enkelmann et al., 2010). In general, there is a wealth of bedrockapatite U-Th/He and fission-track (FT) ages that refer to cooling through 55–65 °C and 100–110 °C, respectively (e.g., Farley, 2000; Carlson et al., 1999) and reveal rapid exhumation along the southern flanks of the Saint Elias Mountains and along the Fairweather fault (Spotila et al., 2004; Berger et al., 2008b; Berger and Spotila, 2008; Meigs et al., 2008; McAleer et al., 2009; Spotila and Berger, 2010; O’Sullivan et al., 1995). The rapid exhumation coincides with the highest rates of annual precipitation (>5000 mm/yr), whereas older cooling ages on the drier (<1000 mm/yr) north side of the mountain range indicate less rapid exhumation (O’Sullivan and Currie, 1996).
Spotila et al., Meigs et al., 2008; Spotila and Berger, 2010; Enkelmann et al., 2010). At the southern flanks the bedrock cooling ages do not show variations in exhumation along the orogenic strike and were suggested to show the highest rates of erosion (>3 mm/yr) within an orogen-parallel band that coincides with the region between the modern and LGM equilibrium line altitude (ELA) (Berger and Spotila, 2008; Berger et al., 2008a). Zircon U-Th/He and FT cooling ages have also been reported for bedrock and detrital material from the Saint Elias Mountains that refer to cooling through 180 ± 20 °C and 250 ± 40 °C, respectively (Reiners et al., 2004; Brandon et al., 1998). Zircon FT cooling ages from modern proglacial sediments revealed a different pattern of exhumation and erosion, with the most rapid and deep-seated exhumation occurring at the Saint Elias syntaxis region, which is covered by the Seward Ice Field (Enkelmann et al., 2008, 2009, 2010). For these detrital zircon FT studies, sand was collected from the main river channel that emergates from the glacier; 3–4 kg of sand were collected at various spots to obtain a well-mixed sample. Sample locations were usually within 100 m and 1 km from the ice front (Enkelmann et al., 2008, 2009). The measured zircon FT age distributions of individual samples were deconvolved into age populations using the binomial peak-fitting method (Galbraith and Green, 1990; Brandon, 1992, 1996), and the peak age of the age populations and their size with respect to the entire sample have been reported (Enkelmann et al., 2008, 2009, 2010).

**Seward-Malaspina Glacier System**

Proglacial sand samples collected around the Malaspina Glacier revealed the youngest zircon FT cooling ages found in this orogen, with single grain ages as young as 0.5 Ma and prominent component age populations that make up 24%–41% of the entire grains and peak at 2–3 Ma (Table 2; Enkelmann et al., 2009). These young ages suggest a region of localized intense exhumation with cooling rates to 300 °C/Ma. The U/Pb crystallization ages of these young detrital zircons FT grains revealed that cooling is clearly due to exhumation (FT age << U/Pb age) and, more surprisingly, that the source region is composed of Chugach terrane rocks and Yakutat Group rocks (Enkelmann et al., 2009). While the Yakutat Group crops out at the southeastern...
**Table 1. Summary of the Main Geologic Events in Southeast Alaska and the Chronometric Record**

<table>
<thead>
<tr>
<th>Time (Ma)</th>
<th>Event</th>
<th>Chronometric record</th>
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<tr>
<td>0–10</td>
<td>Beginning of subduction collision of thick Yakutat crust and formation of the fold-and-thrust belt</td>
<td>Reset and partially reset bedrock apatite fission-track (FT) and U-Th/He ages from the fold-and-thrust belt, and reset bedrock apatite and zircon FT and U-Th/He cooling ages from the Prince William and Chugach terranes (Spotila et al., 2004; Berger et al., 2008b; Berger and Spotila, 2008; Meigs et al., 2008; Perry et al., 2009; Enkelmann et al., 2010)</td>
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<tr>
<td>10–40</td>
<td>Several phases of exhumation in southern and southeastern Alaska due to Yakutat subduction</td>
<td>Bedrock apatite and zircon FT and U-Th/He cooling ages of the Chugach terrane (O’Sullivan and Currie, 1996; Spotila et al., 2004; Berger et al., 2008b; Berger and Spotila, 2008; Meigs et al., 2008) Bedrock FT, U-Th/He, and Ar/Ar ages in southern Alaska and basin inversion due to flat slab subduction (Benowitz et al., 2011; Finzel et al., 2011) Several Eocene to Miocene zircon FT age populations in glacial sediment originating from the Chugach terrane (Enkelmann et al., 2008, 2010) Similar zircon FT populations found in the syncollisional Yakataga Formation that is sourced from the Chugach terrane (Perry et al., 2009; Enkelmann et al., 2010)</td>
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<tr>
<td>50–60</td>
<td>Metamorphism and magmatism related to the ridge subduction, reactivation of the Border Range fault</td>
<td>Zircon U/Pb crystallization ages of glacial sediment derived from the Chugach terrane (Enkelmann et al., 2008, 2009) Zircon U-Pb, K-feldspar, amphibole, and biotite 40Ar/39Ar ages of various intrusive and metamorphic rocks of the Chugach terrane and Border Range fault zone (Sisson et al., 2003)</td>
</tr>
<tr>
<td>90–125</td>
<td>Last major exhumation and/or cooling phase in the Wrangellia composite terrane</td>
<td>Bedrock zircon FT and U-Th/He ages from the Wrangellia terrane north of the Chitina Valley (O’Sullivan and Currie, 1996; Enkelmann et al., 2010) K-feldspar 40Ar/39Ar multidomain modeling and hornblende, biotite ages of rocks north of the Border Range fault (Sisson et al., 2003; Enkelmann et al., 2010)</td>
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**Bagley-Bering-Tana Glacier System**

Three detrital zircon FT samples from the Bering Glacier were collected from river outlets on the eastern side of the glacier at the lowest elevations possible: 6 m, 60 m, and 70 m (Fig. 2) (Enkelmann et al., 2009). As the Bering Glacier terminates in Vitus Lake, contemporary fluvial sediment samples were not collected further west because the subglacial rivers exit the glacier below lake level (Fig. 2). The age populations of the three individual samples are generally similar in their peak ages (Table 2), suggesting that (1) all the material is representative of sediment currently carried by the main rivers under the Bering Glacier, and (2) the samples are representative of the basin, well mixed, and largely unaffected by local sources within the basin, such as landslide material. The main zircon FT age populations of these three Bering Glacier samples peak between 23 and 70 Ma and are similar to the age population peaks found in the detritus of smaller glacial catchments located east of the Bering Glacier (see Table 2; Enkelmann et al., 2008, 2009); they are also similar to the age population peaks reported from bedrock samples of the Kultiteh and Paul Creek Formations collected east of the Bering Glacier (Meigs et al., 2008; Perry et al., 2009). However, bedrock apatite cooling ages from the same formations are young (0.5–5 Ma; Spotila et al., 2004; Berger and Spotila, 2008; Berger et al., 2008b), but the non-reset ages of the higher-temperature systems indicate that these rocks were exhumed from shallow depths (e.g., Spotila et al., 2004; Meigs et al., 2008; Perry et al., 2009; Enkelmann et al., 2010). Based on apatite and zircon FT studies in the fold-and-thrust belt, Meigs et al. (2008) suggested that only the upper 5 km of the sediments are involved in the cycle of lateral material input and erosional exhumation; this is supported by thermal-kinematic modeling of the Yakutat–North America collision (Enkelmann et al., 2010). North of the Bagley Ice Field bedrock apatite U-Th/He and FT ages are older and range from 8 to 30 Ma (e.g., Spotila et al., 2004; Berger et al., 2008b; Berger and Spotila, 2008; Meigs et al., 2008). Two sand samples collected from the Tana Glacier yielded zircon FT age populations that were similar to one another with small age populations (4%–12%) that peak between 11 and 17 Ma and the main age components that peak at 21–35 Ma (Table 2; Fig. 2) (Enkelmann et al., 2008). The zircon FT age distributions from the Tana Glacier are similar to the result obtained from Granite Creek (CH44; Fig. 2), a catchment located just east of the Tana Glacier and composed entirely of Chugach terrane rocks (see Table 2) (Enkelmann et al., 2008).

**NEW ANALYSIS AND RESULTS**

**Thermochronometry**

The individual detrital zircon FT results presented in Table 2 (see also Enkelmann et al., 2008, 2009) show that samples derived from the same glacier system are similar in their age distributions, suggesting that they represent the same source material. We combine the single-grain FT ages of the samples from the same glacier into a single, aggregate grain age distribution and repeated the peak-fitting procedure (shaded in Table 2; Fig. 2). In this way the age population result of a specific glacier is more...
TABLE 2. ZIRCON FISSION-TRACK AGE POPULATIONS FROM INDIVIDUAL SAMPLES AND NEW PEAK FITTING RESULTS OF THE COMBINED AGES FROM TANA, BERING, AND MALASPINA GLACIER SEDIMENTS COMPARED WITH THOSE OF THE CHUGACH TERRANE AND A COMBINED YAKUTAT TERRANE SAMPLE

<table>
<thead>
<tr>
<th>Sample Location</th>
<th>N</th>
<th>P0</th>
<th>P1</th>
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<tr>
<td>Tana Glacier</td>
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<tr>
<td>CH44*</td>
<td>104</td>
<td>8.2 ± 2.3 (5)</td>
<td>21.5 ± 2 (48)</td>
<td>34.5 ± 3 (47)</td>
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<tr>
<td>CH46*</td>
<td>103</td>
<td>13.5 ± 2.2 (9)</td>
<td>21.7 ± 10 (85)</td>
<td>35.6 ± 6 (6)</td>
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<tr>
<td>CH47*</td>
<td>105</td>
<td>10.7 ± 2.2 (4)</td>
<td>17.4 ± 1.8 (12)</td>
<td>27.1 ± 2 (49)</td>
<td>35.2 ± 3 (35)</td>
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<tr>
<td>Tana combined (CH46, CH47)</td>
<td>208</td>
<td>12 ± 3 (5)</td>
<td>18.3 ± 3 (24)</td>
<td>25.3 ± 4 (48)</td>
<td>34.8 ± 5 (23)</td>
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<tr>
<td>Bering Glacier</td>
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<tr>
<td>Y8†</td>
<td>102</td>
<td>14.2 ± 3 (4)</td>
<td>30.2 ± 2 (62)</td>
<td>55 ± 4 (34)</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Y9†</td>
<td>103</td>
<td>23.3 ± 2 (15)</td>
<td>45 ± 3 (50)</td>
<td>71 ± 5 (28)</td>
<td>114 ± 13 (7)</td>
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<tr>
<td>Y10†</td>
<td>104</td>
<td>24.7 ± 2.7 (10)</td>
<td>34.4 ± 3 (32)</td>
<td>52 ± 3 (47)</td>
<td>77 ± 8 (11)</td>
<td></td>
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<tr>
<td>Bering combined (Y8, Y9, Y10)</td>
<td>309</td>
<td>11.8 ± 3.5 (2)</td>
<td>28 ± 3.5 (26)</td>
<td>44 ± 8 (44)</td>
<td>67 ± 11 (23)</td>
<td>111 ± 30 (3)</td>
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<tr>
<td>Malaspina Glacier</td>
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<tr>
<td>Y11†</td>
<td>103</td>
<td>14.2 ± 1 (22)</td>
<td>24.1 ± 1 (33)</td>
<td>64 ± 3 (43)</td>
<td>56 ± 2 (40)</td>
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<tr>
<td>Y12†</td>
<td>100</td>
<td>8.7 ± 1.5 (3)</td>
<td>34.4 ± 1.5 (37)</td>
<td>45 ± 2 (44)</td>
<td>68 ± 4 (31)</td>
<td>107 ± 10 (6)</td>
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<tr>
<td>Malaspina combined (Y3, Y7, Y11, Y12)</td>
<td>411</td>
<td>10.7 ± 3 (3)</td>
<td>34.4 ± 4 (34)</td>
<td>46 ± 4 (35)</td>
<td>67 ± 6 (21)</td>
<td>116 ± 11 (6)</td>
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<td>Consolidated Yakutat cover samples</td>
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<tr>
<td>Kultieth Formation*</td>
<td>250</td>
<td>28 ± 27 (27)</td>
<td>39 ± 55 (28)</td>
<td>78 (8)</td>
<td></td>
<td></td>
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<tr>
<td>Poul Creek Formation*</td>
<td>200</td>
<td>29 ± 37 (19)</td>
<td>41 ± 52 (52)</td>
<td>63 (11)</td>
<td></td>
<td></td>
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<td></td>
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<tr>
<td>IB1†</td>
<td>333</td>
<td>26.4 ± 4 (34)</td>
<td>42 ± 5.9 (49)</td>
<td>68 ± 12 (17)</td>
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<tr>
<td>IB3†</td>
<td>20</td>
<td>41 ± 5.7 (71)</td>
<td>67 ± 15 (29)</td>
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Note: Zircon fission-track ages are from Enkelmann et al. (2008, 2009). Sample locations are shown in text Figure 2. Zircon grain age distributions were deconvolved into component age populations using binomial peak fitting (Brandon, 1992, 1996). Gray shaded bands indicate the results of the peak fitting of grouped samples from specific glacier systems or bedrock areas. P—peak of modeled age populations (in Ma ± 2σ). Parentheses indicate the size of the age population (in percent). N—number of single grain ages.

* Ages from Enkelmann et al. (2008).
† Ages from Enkelmann et al. (2009).
§ Ages from consolidated samples of the Kultieth and Poul Creek Formations from Perry et al. (2009; no error bars given).
** Ages from consolidated samples of the Kultieth and Poul Creek Formations from Meigs et al. (2008).
Glacial processes and exhumation in the Saint Elias Range

from the Prince William terrane (Enkelmann et al., 2008). Rather, the Tana sediments show the typical Eocene–Miocene cooling signal of the Chugach terrane that is also found in the Kultiieth Formation and Paul Creek Formation rocks that were sourced by the Chugach rocks and are now located in the fold-and thrust belt (Perry et al., 2009) (samples Y3,7,11,12; Fig. 2).

Glaciology

To integrate the glaciological and detrital thermochronometric data, we investigate the transport capacity of the subglacial hydrological system of the Bagley-Bering-Tana system. Based on estimates of precipitation and on measurements of proglacial discharge compared to the size of the basin, meltwater from the entire Bagley-Bering system drains primarily to the Gulf of Alaska (Merrand and Hallet, 1996; Josberger et al., 2006). Discharge from the Tana Glacier is considered negligible, though it is admittedly not well studied. The lack of sediment derived from under the Bagley Ice Field in the discharge from the Bering Glacier leads us to examine the sediment transport capacity of the subglacial hydrological system during quiescent (i.e., non-surge) phases of the glacier. We use the known thickness profile for the glacier to determine whether water can flow along the entire length of the glacier, and whether supercooling and refreezing of water might impede water flow and sediment transport under the glacier.

The direction of subglacial water flow is determined by the direction of the hydraulic potential gradient (\( \nabla \Phi \)) at the glacier bed (Röthlisberger, 1972; Shreve, 1972): \( \nabla \Phi = g \rho \frac{V_s}{z_s} - \frac{V_z}{z_b} \), where \( \rho \) is the density of ice, \( g \) is gravitational acceleration, \( z_s \) is the measured elevation of the glacier surface, and \( z_b \) is the bed elevation (Shreve, 1972). The expression shows that the glacier surface gradient tends to dictate the direction of subglacial water flow. At the glacier bed, an impermeable bed with reverse gradient (dipping upglacier) does not necessarily preclude water flow, as it does subaerially. However, in cases of relatively steep, opposing bed surfaces, at least 10 times steeper than the ice surface, subglacial water flow and sediment transport would be blocked.

Using the definition of hydraulic potential and the known ice thickness profile (Fig. 4B), we calculate that the subglacial hydraulic potential is sufficient to drive subglacial water flow downglacier, despite overdeepenings under the Bagley Ice Field (Fig. 4B). As the glacier surface reflects the dynamically induced stresses within the moving glacier over an uneven bed, we can reasonably approximate the basal pressure as the weight of overlying ice per unit area, which can be determined simply from the surface and bed topography (e.g., Gudmundsson, 2003). With the current glacier surface elevation and slope, ice under the Bagley Ice Field would have to be much thicker than the thickest ice that has currently been measured. Whereas radar measurements have only revealed maximum ice thicknesses of just slightly over 1 km (Gim et al., 2008), ice at 40 km down the glacier’s length (Fig. 4B) would need to be 3000 m thick for a reverse bed gradient between 40 and 60 km to prevent subglacial water flow toward the Bering terminus. This analysis suggests that subglacial water should flow from the Bagley Ice Field to the Bering Glacier despite overdeepenings, consistent with estimates of the size of the basin that is required, based on estimates of regional ice melt rates, to account for the measured proglacial discharge (Merrand and Hallet, 1996; Josberger et al., 2006).

A downglacier decrease in subglacial hydraulic potential is a necessary but not sufficient condition for effective sediment transport; other requirements arise from thermodynamic constraints on the carrying capacity of the hydrological network. Because the ice is at the pressure melting point, water driven from higher to lower pressure areas can be supercooled due to the melting point dependence on pressure. With further depressurization, this supercooled water can freeze at the base of the glacier. This process and its consequences, with respect to erosion and sediment transport, have been studied extensively (Röthlisberger, 1972; Hooke, 1991; Alley et al., 2003). In general, basal freezing is expected if the bed ascends downglacier more steeply than 20%–70% of the glacier surface slope; this percentage depends on the air content of the water (Alley et al., 2003). Such steeply ascending bed slopes occur on the downstream end of overdeepening in the Bagley-Bering system: the bed slopes commonly reach >20% and occasionally exceed 70% of the glacier surface slope (Fig. 4B). Direct evidence for subglacial supercooling under the Bering Glacier has been reported; it includes frazil ice growth on sediment traps suspended in Vitus Lake, due presumably to freezing of supercooled water emanating from the glacier terminus in the lake (Fleisher et al., 1998).

DISCUSSION

Transport Capacity and Surges of the Bagley-Bering Glacier

The effects of subglacial supercooling are likely to influence sediment transport under the Bering Glacier. Supercooling and subsequent freezing could limit the amount of sediment transferred through the subglacial hydrological system. Subglacial freezing would tend to choke or close subglacial conduits, thereby constricting the flow and decreasing the bedload sediment-carrying capacity of the conduits, as this capacity increases nonlinearly with discharge (e.g., Alley et al., 1997). This decrease, together with the relatively low hydraulic potential gradient on the upglacier side of transverse bedrock ridges, would be conducive to sediment deposition (Alley et al., 1998). In the case that conduits are completely blocked, only sediment that is incorporated in the ice accreted at the base or entrained higher in the glacier would move beyond the region of ice freeze-on. Over time, material that is stored under the glacier because of loss of sediment-carrying capacity would be poorly represented in the detrital samples collected from rivers at the glacier margin.

During surges the subglacial hydrological system transforms considerably. Surges are likely initiated when the system is not able to convey the meltwater at the glacier bed, causing the basal water pressure to generally rise and accelerate sliding. Ultimately, the system fails, marking the end of the surge, and releases major sediment-laden floods (Humphrey and Raymond, 1994; Jaeger and Nittouer, 1999; Fleisher et al., 2003). In front of the Bering Glacier, the massive discharge of the Seal River increased by ~70% during the 1994 flood and the suspended sediment flux increased 8 or 9 fold (Merrand and Hallet, 1996); this highlights the sensitive dependence of sediment flux on discharge (e.g., Alley et al., 1997). Further support for sediment evacuation from the Bagley-Bering system largely during glacier surges is evident in the Bering Trough, where fine-grained, high-porosity mud deposits, representative of fast sedimentation, abruptly punctuate coarser and bioturbated sand and mud layers (Jaeger et al., 1998; Jaeger and Nittouer, 1999). We hypothesize that much of the sediment eroded below the Bagley Ice Field is evacuated during surges. The youngest zircon FT age population found in the Bering Glacier and Tana Glacier sample that peaks ca. 12 Ma, but comprises only <5% of the entire sample, may highlight the cooling signal from underneath the Bagley Ice Field, which is not transported efficiently during quiescent periods. This hypothesis could be tested through a study of sediments deposited by such floods proglacially and offshore.

Rapid Erosion Underneath the Seward-Malaspina Glacier System

The large age population peak ca. 2 Ma found in the Malaspina sediment is exceptional within the orogen and indicates very fast erosion in
the catchment, particularly north of the Seward Throat under the sluggish Seward Ice Field, where ice surface velocities do not exceed 100 m/yr (Ford et al., 2003) (Fig. 3). Material from around the accumulation area of the glacier found at the terminus demonstrates that, unlike for the Bering Glacier, material is successfully evacuated by the subglacial drainage system despite the existence of a number of overdeepenings under the Seward Throat (Sharp, 1958; Muskett, 2007; Headley et al., 2012).

The Seward Throat (Fig. 3) seems an ideal location to consider the impact of sliding velocity on erosion rate, as the surface velocity increases by more than an order of magnitude as ice funnels from the Seward Ice Field through the narrow gap in the range (Fig. 3). The erosion rate is generally expected to scale with basal sliding speed (Hallet, 1979, 1996; Humphrey and Raymond, 1994). Numerical analysis of glacier dynamics has shown that sliding accounts for the majority of the ice motion through this fast-moving reach (Headley et al., 2012). Erosion in the Seward Throat might be expected for many reasons. The exceptional convergence of ice funnelling through this region accounts for extremely rapid flow, fast sliding (exceeding 800 m/yr in places), and by inference, fast erosion (Headley et al., 2012). The Seward Throat also traverses the Malaspina fault, which is the northeastern continuation of the Pamplona fault zone; the active deformational front of the Yakutat–North American subduction boundary (Figs. 1 and 2) (Plafker et al., 1994; Bruhn et al., 2004; Worthington et al., 2008), where many active faults converge (Chapman et al., 2008). GPS measurements show the highest rates of crustal convergence across this fault zone (Elliott et al., 2008). However, we lack supporting evidence for rapid exhumation from the detrital thermochronometry. High rates of exhumation are not apparent from the detrital thermochronometry because the zircons of the underlying rocks (belonging to the Yakutat terrane) are not reset. Sediment strata have not been heated high enough to erase entirely the cooling signal from their source area, before they were exhumed again.

In contrast, the young zircon grains originating from Chugach terrane rocks under the Seward Ice Field indicate an unexpected region of rapid exhumation in an area with low surface velocities and where rapid exhumation is not expressed by the few bedrock ages that have been reported from this area (O’Sullivan and Currie, 1996; Enkelmann et al., 2010). Particularly, the 10 Ma zircon U-Th/He age from the Seward Ice Field nunatak suggests large gradients in exhumation above and beneath the glacier.

A major finding reported in recent studies in this region is that a distinct orogen-parallel band of rapid exhumation and long-term, focused erosion, inferred from bedrock apatite U-Th/He ages, corresponds in a rough sense to the region between the modern and LGM ELA (Berger and Spotila, 2008; Berger et al., 2008b). This was consistent with the common assumption that the glacial erosion rates tend to be highest near the regional ELA. The ELA is where the long-term accumulation and ablation of ice are balanced, and therefore where the ice flux is maximized (e.g., Andrews, 1972; Brozović et al., 1997; Egholm et al., 2009). However, we suggest that a direct connection between the region of maximum erosion rate and the ELA is unlikely due to a large tectonic influence.

Defining the location of the ELA in the Saint Elias Mountains is not a trivial task. First, the current ELA is located high up in the mountain range, with estimated altitudes ranging from ~1000 m to 1100 m for the Seward-Malaspina (Meier et al., 1971; Pévé, 1975) and the Bagley-Bering systems (Pévé, 1975; Tangborn, 2002; Berger and Spotila, 2008). Throughout much of the Quaternary, however, the ELA was ~300–600 m lower (Pévé, 1975), when glaciers were also much more extensive. The ELA likely intersected the ice surface tens of kilometers above the current shoreline due to both the shallow surface gradients and the substantial thickness of these massive glaciers, although the ice thickness history is not established. In this region, the location of the ELA likely swept through a broad region during the Quaternary as the glaciers expanded over 100 km onto the continental shelf during recurring periods of cooler climate and lower sea level (Mann and Hamilton, 1995) (Fig. 1).

A reinterpretation of the apatite U-Th/He ages in the fold-and-thrust belt suggests that a simple correspondence between young cooling ages and the region of modern and past ELAs is perhaps too simple: the zone of rapid exhumation was found to roughly parallel the active northeast-striking thrust fault systems (Enkelmann et al., 2010) (Fig. 2A). In light of this recent work, we stress that the highest erosion rates, inferred from the highest cooling rates, and deepest exhumation in the Saint Elias orogen occur at the tectonic corner of the Yakutat indenter instead of a focused band related to the ELA position (Fig. 2). Generally, a zone of rapid erosion would tend to persist and account for the deep exhumation only if it is offset by rock uplift, otherwise the erosion by the glacier would carve deep basins that would eventually cease to be eroded (e.g., Alley et al., 2003). Moreover, a region where faulting continuously strains, weakens, and fractures the rock may be an optimal setting for efficient glacial erosion (Laitakari et al., 1985; Dühnforth et al., 2010). The effect of crustal strain on making the bedrock more erodible can focus erosion, crustal deformation, and rock uplift into narrow regions, contributing to the interplay between tectonics, erosion, and topography (e.g., Zeitler et al., 2001; Koons et al., 2011).

**Integration of Thermochronometry and Glaciology**

On an orogen scale, the Saint Elias Mountains have been showcased as an archetype of climatic influence on tectonics (Spotila et al., 2004; Berger et al., 2008b; Whipple, 2009). There is no doubt that climate affects glaciers, and that glaciers are efficient at eroding and evacuating sediments from the range. However compelling, using simplified stand-ins such as the ELA or the sliding velocity for the many complicated processes related to glacial erosion appears to not always capture the full rock exhumation history of this region, especially when considering the changing tectonic and climatic histories. This study helps build a basis for further and more detailed analysis of the spatial and temporal patterns of glacial erosion in conjunction with sedimentation and erosion on the continental shelf over the Quaternary. By combining existing thermochronologic data into bulk samples representing single glaciers, we have a snapshot of from where and how glacially eroded bedrock has been transported. By integrating these results with glaciological evidence for patterns of erosion rate and sediment transport, we can make interpretations of how larger scale patterns of exhumation and glaciers interact.

**CONCLUSIONS**

Our integration of glaciological and thermochronometric data over the Saint Elias Mountains is a powerful tool to reveal new insights on how glacial erosion shapes orogens over geological time scales. This study emphasizes that, even for glaciers that occupy the same climatic and orogenic setting, the patterns of erosion rates and sediment evacuation are influenced both by glacier-specific factors (basin architecture, glacial dynamics, subglacial hydrology) and local tectonics. Therefore, a detailed consideration of both glaciology and tectonics, on scales ranging from the thickness of glaciers to the width of the orogen, is necessary as a basis for understanding the spatial patterns of glacial erosion and rock exhumation.

The Seward Ice Field is a locus of rapid rock exhumation even though ice velocities are modest. This suggests that the underlying bedrock is particularly erodible not because the lithologies
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are inherently easy to erode, but because the bedrock is more pervasively fractured due to major faulting in the region. Erosion is also expected to be rapid where immense volumes of ice rapidly funnel through the nearby Seward Thrust. This, however, cannot be considered in more detail due to the nonreset zircon cooling ages.

Although the Bering and Tana Glaciers are the main outlets of the Bagley Ice Field, the sediment transport capacities of their subglacial hydrological systems generally appear to be insufficient to evacuate rock debris produced by erosion and stored under the Bagley Ice Field. Sediment transport may be hindered by loss of carrying capacity due to supercooling and freeze-on processes, except during surges. The possibility that much of the sediment eroded below the Bagley Ice Field is evacuated during surges could be tested through a study of sediments deposited by such floods proglacially and offshore.

ACKNOWLEDGMENTS

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REFERENCES CITED


