GEOPHYSICAL RESEARCH LETTERS, VOL. 33, L24504, doi:10.1029/2006GL028467, 2006

Rapid erosion of soft sediments by tidewater glacier advance: Taku Glacier, Alaska, USA

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Received 13 October 2006; accepted 14 November 2006; published 22 December 2006.

[1] Taku Glacier in southeast Alaska has advanced 7.5 km over the last 115 years, overriding its own glaciomarine and outwash sediments. We have documented rapid erosion of these sediments by comparing radio echo soundings (RES) along five transects (2003–2005) to earlier RES surveys (1989 and 1994) and to early bathymetric surveys of the proglacial fjord. Erosion rates, $\dot{E}$, reached 3.9 ± 0.8 m a^{-1} (1989–2003) at a distance, L, of 5.4 km from the 2003 terminus, where ice thickness, H, is 610 m. $\dot{E}$ averaged 2.0 ± 0.1 m a^{-1} (1940–2005) at L = 3 km (H = 350 m), and 1.5 ± 0.2 m a^{-1} (1952–2005) at L = 1.5 km (H = 250 m). Detailed mapping over a 4 km^2 area of the terminus revealed a deeply incised channel in line with a major outlet stream. Glaciofluvial processes must play the dominant role in the subglacial erosion and removal of these unlithified sediments. **Citation:** Motyka, R. J., M. Truffer, E. M. Kuriger, and A. K. Bucki (2006), Rapid erosion of soft sediments by tidewater glacier advance: Taku Glacier, Alaska, USA, *Geophys. Res. Lett.*, *33*, L24504, doi:10.1029/2006GL028467.

1. Introduction

[2] During expansions, glaciers can override unlithified sediments, which are often composed of glaciomarine sediments, moraines, and outwash deposited during prior retreats. Erosion of these soft sediments during glacier advance results in their redistribution in proglacial environments, producing significant landscape changes. These remobilized sediments accumulate on fjord and ocean floors, thus submarine strata can preserve records of glacier advance [*Jaeger et al.*, 2001; *Gulick et al.*, 2004]. Such erosion has not been well quantified, therefore it has not been addressed in models of continental ice sheet expansion [e.g., *Hildes et al.*, 2004] and of advancing tidewater glaciers [e.g., *Nick and Oerlemans*, 2006], even though. erosion and redistribution of stored sediment may have been an important control on the dynamics of advancing glaciers [*Meier and Post*, 1987; *Post and Motyka*, 1995; *Larson and Mooers*, 2005].

[3] Glacier advance over thick sections of till also has implications for the mechanics of basal sliding [e.g., *Truffer et al.*, 2001; *Hooke*, 2005], development of subglacial drainage systems [*Walder and Fowler*, 1994; *Ng*, 2000a, 2000b], and entrenchment of termini in over-deepened basins [*Nolan et al.*, 1995; *Björnsson*, 1996]. Erosion of overridden sediments may eventually result in bedrock being reached with concomitant changes in the style of basal sliding and subglacial drainage. Erosion of soft sediments may also continue under stable or retreating conditions until sediment supply has been exhausted.

[4] Data on glacier erosion of unlithified subglacial sediments is lacking in part because the Earth is currently in a regime of general glacier retreat, thus opportunities for such studies are rare. One exception is Taku Glacier, which has advanced 7.5 km since 1890 (Figure 1). We have taken advantage of this to investigate erosion rates using repeat ice-penetrating radio echo soundings (RES) and by comparing these results to a series of bathymetric surveys that were periodically conducted in the proglacial fjord during the last 115 years.

2. Taku Glacier

[5] Taku Glacier (700 km^2, 60 km long), is a temperate, maritime glacier in the granitic Coast Range mountains that lie on border between southeast Alaska and British Columbia; it is the largest glacier draining the 4000 km^2 Juneau Icefield. After more than a century of calving retreat that began in 1750 [*Motyka and Begét*, 1996], Taku Glacier began steadily advancing into the north end of Taku Inlet [*Motyka and Post*, 1995; *Motyka and Echelmeyer*, 2003] (Figure 1). The present advance of Taku Glacier is the result of dynamical changes that many tidewater glaciers undergo following calving retreats [*Meier and Post*, 1987] and a large accumulation to total area ratio of about 0.90 during the last century [*Post and Motyka*, 1995]. The maritime climate produces both large amounts of winter snowfall and high rates of summer ablation, typically 10–12 m a^{-1} near the terminus [*Pelto and Miller*, 1990], thus Taku Glacier has a high rate of mass exchange with velocities at its equilibrium line altitude ($\sim$1000 m) of about 1 m d^{-1}. The glacier ended in tidewater during the early 20th century, with active calving in the 100+ m-deep fjord [*Post and Motyka*, 1995]. The glacier has advanced since then, gradually filling in the fjord with glaciomarine sediment, outwash deposits, and ice. Additional fluvial contributions came from Taku River [*Motyka and Post*, 1995]. In a previous study [*Nolan et al.*, 1995], we showed that the glacier is actively excavating and entrenching itself into these overridden sediments. Here we present new data that greatly expand on our earlier results and document rapid erosion throughout a broad sector of the terminus lobe; we then explore implications of our findings.

3. Methods

[6] We used a 2 to 5 MHz monopulse radar transmitter deployed on foot to measure ice thickness. Spacing between measurements was approximately 30 m along

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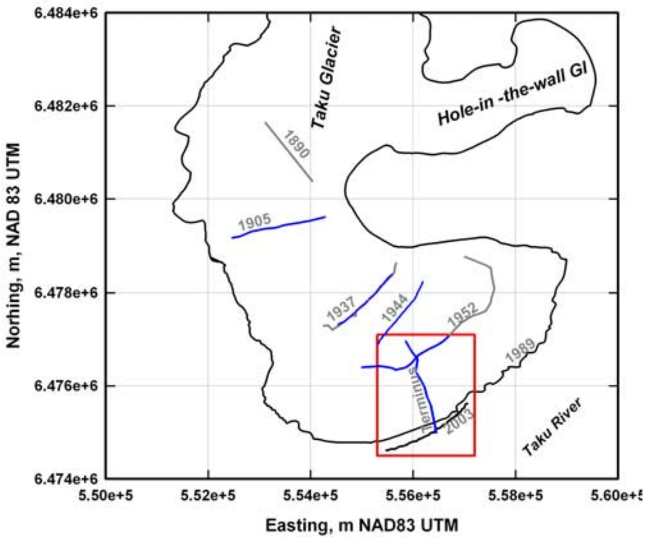

Figure 1. Lower Taku Glacier showing locations of RES transects (blue) and outline of terminus bed map in Figure 4 (red box). Historic terminus positions also shown.

the majority of our transects, but in some cases it was 100 m or more. Crevasse fields often impeded travel and dictated positioning of antennas. An oscilloscope was coupled to the receiving antenna and travel times were measured from the oscilloscope trace; waveforms were stored for additional post-processing and analysis. The accuracy of the RES is primarily limited by the uncertainty in travel time, which was read to ±0.1 μs (= ±8 m) or better. Another source of uncertainty arises from the wave speed in temperate ice: 168 ± 2 m μs^{-1}. When comparing RES, it is safe to treat this error as systematic (which cancels), because the wave speed depends on water content of the ice, which should be similar for all soundings. For comparisons to bathymetry, RES wave speed uncertainty translates to ±7 m for our greatest ice thickness, $H = 610$ m, and ±3.5 m for $H = 300$ m. Bathymetric surveys conducted by the U.S. Coast and Geodetic Survey in 1890, 1937 and 1952 and obtained from the National Ocean Service (NOS) provided baseline data for fjord depths directly in front of the advancing glacier with uncertainties of ±2 m, ±1 m, and ±1 m respectively. Antenna positions in 2003–2005 were determined by dual frequency differential GPS relative to a fixed base station in front of the terminus and have estimated horizontal and vertical accuracies of ±0.3 m and ±0.5 m respectively. Earlier RES locations (1989 and 1994) have estimated positional uncertainties of about ±1 m. Standard propagation of error analysis was used to estimate 1σ values of calculated results.

4. Results

[7] We conducted soundings along four cross-glacier transects, two of which coincide with terminus positions during bathymetric surveys in 1937 and 1952 (Figure 1). The northernmost transect, P-1905, lies at the approximate position of the 1905 terminus, about $L = 5.4$ km, where L is the distance from the 2003 terminus along the glacier centerline. 1905 fjord depths were linearly interpolated by estimating the amount of sediment that filled this region between the 1890 NOS survey and 1905, using sedimentation rates computed by comparing NOS surveys. The deepest soundings made along P-1905 showed that the glacier bed was as much as 270 m below sea level (bsl) in 2003 (Figure 2a). Crevasse fields limited spacing between soundings so a detailed assessment of bed geometry is not possible. The returns nevertheless document that up to 190 m of sediments have been eroded since 1905. Comparison with a 1989 sounding near the centerline shows that erosion there averaged 3.9 ± 0.8 m a^{-1} between 1989 and 2003 compared to 1.3 ± 0.1 m a^{-1} for 1905–1989.

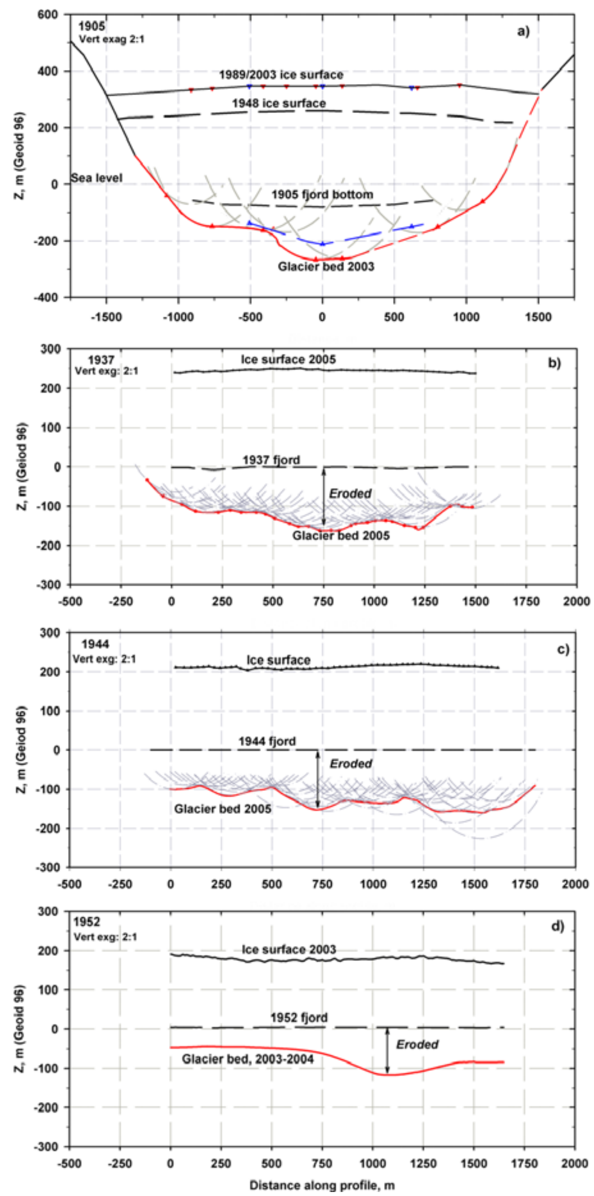

Figure 2. Ice surface, fjord bathymetry, and RES bed profiles: (a) location "1905"; (b) location "1937"; (c) location "1944"; (d) location 1952. Black dashed lines are fjord bathymetry at terminus in respective location years. Red lines show estimated bed locations from 2003–2005 RES. Blue dashed line in Figure 2a represents estimated bed location from 1989 RES.

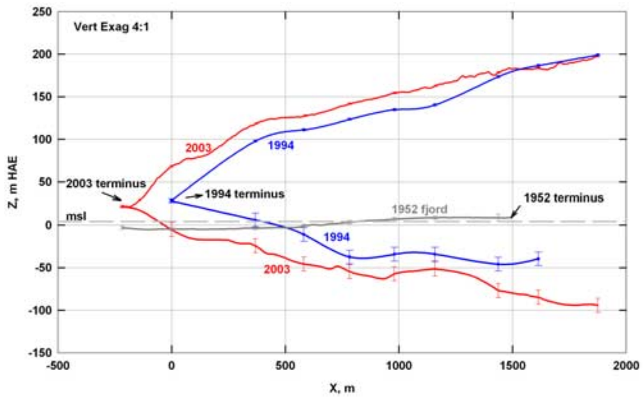

Figure 3. Longitudinal transect from terminus. Blue (red) lines indicate glacier bed and ice surface in 1994 (2003) with 1σ error bars. Fjord bathymetry for 1952 (gray line) also shown.

Although the surface of the glacier gained elevation between 1948 and 1989, it has remained nearly constant since then (Figure 2a). In contrast, the glacier sole entrenched itself into sediments by about 55 m at the centerline between 1989 and 2003.

[8] The next three profiles (Figures 2b–2d) correspond to terminus positions in 1937, 1944, and 1952 (L = 3.1, 2.3, and 1.4 km respectively). Table 1 summarizes the averages of erosion (E_{ave}), erosion rate ($\dot{E}_{ave}$), and ice thickness (H_{ave}) along each profile between respective time intervals. Within uncertainty limits $\dot{E}_{ave}$ and H_{ave} are nearly identical for profiles P-1937 and P-1944 (2 m a^{-1}, 300 m); while P-1952 shows lower values for both $\dot{E}_{ave}$ and H_{ave} (1.5 m a^{-1}, 250 m). Erosion has scoured sediments to depths of up to 160 m bsl along P-1937 and P-1944, and to 120 m bsl along P-1952. The final transect is a 2 km long longitudinal line starting from the 2003 terminus and perpendicular to it (Figure 3). Elevations of the glacier surface were photogrammetrically derived from 2003 air photos while the 2003–2004 bed geometry was determined from a dense array of soundings conducted in this region (see below). Also plotted in Figure 3 are the 1952 fjord bottom and the results of a 1994 RES survey along this same transect. $\dot{E}_{ave}$ increased from 1.1 ± 0.2 m a^{-1} (1952–1994) to 3.1 ± 1.2 m a^{-1} (1994–2003) while ice thickness increased an average of ~50 m between 1994 and 2003. Over half of this thickening reflects entrenchment into subglacial sediments; the remainder, a rise in surface elevation.

[9] Our final result is a ~4 km^2 map of the glacier bed at the terminus produced from numerous RES soundings (average spacing 50 m) obtained during 2003–2004 (Figure 4). This map documents a ~250 m wide subglacial channel incised into sediments to depths of 140 m bsl

Table 1. Average Erosion Rates and Ice Thicknesses for Cross-Glacier Transects[a]

Profile	L, km	H_{ave}, m	E_{ave}, m	$\dot{E}_{ave}$, m a^{-1}
1937	3.1	370 ± 9.0	123 ± 9.1	1.8 ± 0.13
1944	2.3	341 ± 8.9	128 ± 9.4	2.1 ± 0.15
1952	1.4	252 ± 8.6	78 ± 8.6	1.5 ± 0.17

[a]L is the distance from terminus along centerline; H_{ave}, E_{ave}, and $\dot{E}_{ave}$ are the average ice thickness, total erosion, and erosion rate along profile, respectively.

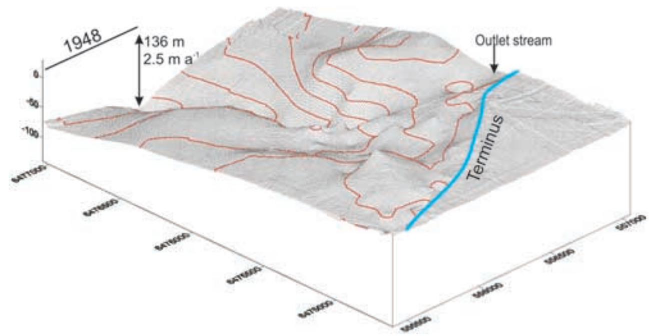

Figure 4. Wireframe map of glacier bed at terminus section bounded by red box in Figure 1 as determined from dense array of RES soundings in 2003 and 2004. Black (blue) line shows 1948 (2003) terminus position. Outlet stream also indicated.

since 1948, the date when the glacier first began overriding this region. The surrounding subglacial terrain is being eroded at much slower rates. The channel is in line with the mouth of a major outlet stream.

5. Glaciofluvial Erosion and Sediment Transport

[10] Glaciofluvial erosion is the only mechanism capable of removing large amounts of sediment from beneath a glacier. Sediment transport through a basal freeze-on layer is relatively small [e.g., *Lawson et al.*, 1998; *Alley et al.*, 2003], as demonstrated by a crude order of magnitude estimate, loosely based on observations: Assume that a 2 m thick basal ice layer contains 50% sediments and moves 100 m a^{-1}. The terminus lobe is roughly 6 km long. This would result in 6 × 10^5 m^{3}a^{-1} of sediment transport, or averaged over the area of active erosion (~20 km^2) this amounts to about 0.03 m a^{-1} of erosion rate, far less than what we observed. Similarly, sediment transport by subglacial till deformation cannot account for a majority of the observed erosion [e.g., *Boulton*, 1996; *Truffer et al.*, 2000].

[11] Taku Glacier's setting in a marine environment insures high mass turn-over rates, and large amounts of water are readily available. Sediment erosion is a non-linear function of water flux and it depends on sediment concentration in the water, turbulent flow along conduit walls, and supply of sediment to the conduit. A quantitative assessment of fluvial sediment transport is difficult, however, because it requires knowledge of the distribution of water flux. There are virtually no observations of the hydrology of glaciers overlying till, and only few models (*Walder and Fowler* [1994, hereinafter referred to as *WF*] and *Ng* [2000a, 2000b], summarized by *Hooke* [2005, p. 223–230]).

[12] According to *WF*, two stable regimes of water flow exist: (i) localized flow through large channels cut into the ice (similar to the classic R-channels [*Röthlisberger*, 1972]), and (ii) distributed flow through shallow and broad canals cut into till. The effective pressure N (ice overburden P_i minus water pressure P_w) determines which regime is stable. At high effective pressure, (i) is stable. This seems to be the case near the terminus where water flow is concentrated and exits the glacier along relatively large outlet streams flowing in a well defined subglacial valley (Figure 4). This tends to localize erosion. Further upstream,

water flow seems to become more distributed, presumably along a network of canals (ii). Typically, these canals are too small to be resolvable in RES, but our results do indicate the existence of several shallow depressions, that could be indications of distributed flow. Sediment erosion is then also more distributed. In *WF*'s model, erosion along canals is a decreasing function of N. This is due to the rheological weakening of the till at low N, which proves more important than the reduced driving force at low N.

[13] As the glacier thickens, N changes as well. A thickening of the glacier leads to an increase in P_i. If thickening results in entrenchment, one should also expect an increase in water pressure. This is due to the fact that water has to be driven up to approximately sea level, which requires a certain pressure gradient P_w/L, where L is the distance to the terminus. If there are no major changes in water flux and in the nature of the drainage system, this pressure gradient has to be maintained, and entrenchment by ΔH must result in an increase in water pressure by $\Delta H \rho_w g$. This changes the effective pressure: $N_{new} = P_{i,new} - P_{w,new} = P_i + \Delta H \rho_i g - P_w - \Delta H \rho_w g = N - \Delta H(\rho_w - \rho_i) g < N$.

[14] An additional effect is that L increases as the glacier advances and the emergent base level at the terminus rises as proglacial sediment accumulate. These effects also require an increase in P_w and therefore a decrease in N to maintain a pressure gradient. It is therefore reasonable to assume that erosion rates should increase with increasing ice thickness, which matches our observations. What is more, this process sets up a positive entrenchment feedback loop, which insures accelerated erosion and entrenchment as long as the glacier remains in positive balance.

6. Summary and Further Implications of Findings

[15] 1) Our results show that soft sediment erosion of 1–4 m a^{-1} can occur for valley glaciers in maritime climates, and that the erosion rate increases as ice thickens. We have shown that these increases can be explained by decreasing N as a glacier thickens through entrenchment, which leads to a positive feedback loop. Such erosion is likely to have taken place during expansion of the Cordilleran and Continental ice sheets. In addition, soft bed sliding has been implicated in the rapid advance of the Puget lobe of the Cordilleran ice sheet during the late Pleistocene [*Brown et al.*, 1987] and probably also contributed to the rapid advance of Taku Glacier in the early 20th century [*Motyka and Post*, 1995]. Soft sediments and their erosion therefore play a significant role in a glacier's dynamics and need to be taken into account in modeling glacier and ice sheet advances [*Larson and Mooers*, 2005]. Once unlithified sediment is totally eroded, bedrock will be reached, substantially reducing proglacial sediment delivery and changing the dynamics of basal motion.

[16] 2) Down-fjord redistribution of subglacially excavated unlithified sediments can have a profound effect on the evolution of the local and regional landscape. Taku Inlet, which was a 100 m deep fjord in 1890 is now shoal at a distance of 20 km from the 1890 terminus [*Motyka and Post*, 1995; *Post and Motyka*, 1995]. Extrapolating to larger scales, evidence of sediment fluxes from rapid excavation and redistribution during periods of major glacier advance must be present in the oceanic sediment records on continental shelves. Such records can help delineate advance and retreat cycles of glaciers in coastal waters [*Jaeger et al.*, 2001].

[17] 3) High erosion rates have positive and negative feedbacks on the advance of a tidewater glacier. Proglacial redeposition of subglacially eroded sediments leads to shoal conditions which reduce or eliminate calving at the glacier terminus thereby enhancing further advance [*Hunter et al.*, 1996; *Post and Motyka*, 1995; *Motyka and Truffer*, 2006]. On the other hand the combined effects of raising the terminal moraine and reducing the rate of ice-surface rise through entrenchment into subglacial sediments reduce the ability of the ice to over-ride the moraine, thus retarding advance [*Kuriger et al.*, 2006].

[18] 4) The deep troughs that tidewater glaciers excavate to well below sea level during their advance will lead to such glaciers' rapid demise once a retreat is initiated. Retreat from the shoal moraine places the terminus into deep water, exposing it to calving. Deep water accelerates calving and retreat, which then becomes almost entirely independent of climate conditions [*Meier and Post*, 1987; *Post and Motyka*, 1995].

[19] 5) High erosion rates can result in errors when assessing glacier change solely from surface elevation measurements. Such measurements cannot account for subglacial erosion and therefore could have considerable error.

[20] 6) Bedrock erosion rates cannot always be directly inferred from measurements of proglacial sediment fluxes [e.g., *Hallet et al.*, 1996]. A significant fraction of the flux may be remobilized sediment rather than sediment derived from primary bedrock erosion. Even stable or retreating tidewater glaciers may still be eroding significant quantities of unlithified sediments.

[21] **Acknowledgments.** Support for this work was provided by U.S. National Science Foundation Grant OPP-0221307. Additional support was provided by the Geophysical Institute, University of Alaska. We wish to thank J. Amundson, E. Boyce, and M. Hekkers for valuable field assistance. The manuscript benefited from comments by J. Walder and an anonymous reviewer.

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