Water, sediment and tidewater glaciers: simplistic review and weakly constrained speculations

Richard B Alley
Earth System Science Center and Department of Geosciences
The Pennsylvania State University
University Park PA 16802

Introduction

Ice flow is highly sensitive to basal lubrication, of which we remain somewhat ignorant. Lubrication in turn is linked to the basal water and sedimentary systems, making these important topics for study. I was asked to review the role of subglacial water in behavior of tidewater glaciers. Hence, I start with a few well-known and often-reviewed generalizations that probably accord with most available data, and are probably accurate in most cases. However, I suspect someone could be found who would argue with most or all of these, and accurate, physically based quantitative expressions are lacking for most. I include brief references to representative or key papers. Following this, I sketch two research topics of possible interest that involve the tidewater hydrological and sedimentary systems in some way.

Simple or simplistic review:

⇒ Basal water storage smooths the bed and speeds ice motion (Kamb, 1987).
⇒ Increased basal water pressure increases basal water storage (Iken and others, 1983).
⇒ Steady basal water pressure increases with water supply (Alley, 1996) up to some limit, but steady basal water pressure decreases with water supply above that limit as high-flux, compact-cross-section channels develop in response to thermal dissipation by water flow (Röthlisberger, 1972). The limit is probably different in different situations, tied to ice motion and bed character, and exhibits hysteresis with changing water supply (Walder, 1982).
⇒ A step increase in water supply to a steady system above the thermal-dissipation limit typically causes rapid increase of water pressure and basal water storage, followed by slower decrease in pressure and storage to steady values lower than before the increase (Iken and others, 1983).
⇒ Steady water pressure typically increases with flux without a thermal-dissipation limit for water flowing up a steep slope (Hooke and Pohjola, 1994), because the pressure drop along flow combined with the pressure-dependence of the melting point cause the water flow to absorb heat rather than generating it, thus growing ice that clogs channels.

⇒ Steady water pressure typically increases with flux without a thermal-dissipation limit for water flowing over a very soft bed, because creep of sediment into low-pressure regions balances the tendency toward lower water pressure by melt enlargement of channels (Walder and Fowler, 1994).

⇒ Soft sediments typically increase ice velocity directly through bed deformation (Boulton and Hindmarsh, 1987), and indirectly through burial of bumps that restrain sliding (Anandakrishnan and Alley, 1994) and through creep into low-pressure channels raising water pressure and speeding sliding and bed deformation (Walder and Fowler, 1994), although it may be possible, at least in theory, for soft sediments to slow ice motion by burying exceptionally smooth, polished bedrock under a rougher interface (Weertman, 1964).

⇒ Of the many ways glaciers have to transport sediment, subglacial channels can be especially effective when they are well-developed (Hunter and others, 1996).

Weakly constrained speculation 1: Moraine-shoal upglacier slopes are balanced to achieve sediment continuity.

Consider a large, fast-moving, sediment- and water-rich tidewater glacier such as the Columbia Glacier, southeastern Alaska, USA, with a large sedimentary moraine shoal and the ice front. Assume that basal water channels exist along much of the glacier length, transport sediment efficiently, and so by sediment continuity prevent any deforming subglacial till from becoming as thick as is allowed by the physics of till deformation (Humphrey and others, 1993).

If the moraine shoal is too steep on the upglacier side, basal channels must rise and supercool; hence, they will experience increased pressure and lose water to distributed drainage systems, and will lose transport capacity (Lawson and others, subm.; Strasser and others, 1996). Assuming that all of the basal transport capacity is not lost to gradually rising englacial channels, loss of basal transport capacity will cause deposition at the upglacier end of the moraine shoal, reducing its steep slope.

If the moraine shoal is too gradual on the upglacier side, basal channels will not be perturbed by transit across the shoal, but transport by deformation of basal sediment will increase onto the
moraine shoal in response to the greater sediment thickness downglacier. This will cause erosion of the upglacier end of the moraine shoal, steepening it.

The optimal angle for a static moraine shoal between these limits would be one in which a little of the upglacier fluvial transport capacity is lost, to balance a little increase in bed deformation downglacier, and thus should be one that rises just steeply enough to cause supercooling of channels. A somewhat flatter profile may be appropriate for an advancing shoal, to allow erosion of the upglacier end.

This sketch does not include longitudinal strain of ice and variation of the ice-surface profile and the basal shear stress, and so clearly is far too simplified. Nonetheless, the analysis suggests feedbacks and roles in ice dynamics. For example, as pointed out by Mark Fischer at the meeting, a glacier with a moraine shoal that is overly steep on its upglacier side may experience an excess backstress that tends to stabilize the ice; a glacier with an anomalously gentle shoal will tend to trench into it and so come closer to flotation and possible instability (Nolan and others, 1995).

Weakly constrained speculation 2: Glaciers may grow mountains.

At the meeting, Ross Powell showed data indicating an increase of specific sediment yield (mass per unit area) with increase in basin area for the Glacier Bay region (also in Powell, 1991). Numerous caveats must be attached to such data, related to possibilities of transients in the system, possibilities of different lithologies in different areas, issues related to glaciated vs. non-glaciated regions and their sediment supply, etc. The interpretation thus is not simple.

Suppose we take the simplest possible interpretation of these data, which is that specific sediment yield increases with basin area. A likely interpretation is that confluence of two tributary glaciers creates a glacier tongue that erodes more rapidly than either of the tributary glaciers. Such a model is at least consistent with the common observation of selective linear erosion by glaciers, and the common observation of hanging valleys in deglaciated regions. Such a model is also consistent with some of our knowledge of subglacial erosion, given the general downglacier increase in water flux and water input from the surface to the bed. Variability of subglacial water pressure is especially linked to input of surface water to the glacier bed, and has been linked to fracture of bedrock for plucking. The action of subglacial streams also is likely to increase with water supply.

We thus can interpret the sediment-yield data as suggesting a role for glaciers in increasing local relief. Perhaps more peculiarly, if the sediments are removed efficiently from the glacier toes by
fluvial or marine processes, and if the distance between highland and lowland regions is such that their isostatic response is strongly linked, the isostatic uplift in response to preferential lowland erosion might even raise the upland regions, thus increasing maximum elevations in the area (c.f. Oerlemans, 1984).

I hope that this brief tour through the roles of subglacial water in tidewater dynamics will illustrate the critical importance, the fascinating problems, and the deep ignorance attached to the topic.

References


Tidal forcing of basal seismicity of ice stream C, West Antarctica, observed far inland

S Anandakrishnan and R B Alley
Earth System Science Center and Department of Geosciences
The Pennsylvania State University
University Park PA

Abstract

The seismicity rate beneath the downglacier 85 km of ice stream C, West Antarctica is modulated by the tide. The tide beneath the Ross Ice Shelf modifies the force balance of the ice stream basal environment enough to change the rate of basal microearthquake generation by an order of magnitude. This tidal forcing travels up the ice stream as an attenuating wave at approximately 1.6 m/s and is detectable 85 km from the grounding line. We successfully model this behavior as an elastic ice stream underlain by a viscous substrate of viscosity $\eta$ and thickness $H_b$, and calculate that the substrate has an apparent stiffness $\eta/H_b$ of $O(10^8)$ Pa-s/m. This finding suggests that the conditions of the till layer at the bed of ice stream C are similar to those of ice stream B and that the reason for the recent stagnation of ice stream C is other than loss of till. We further find that the ice stream at the grounding line is more strongly affected by ice-shelf processes than by the basal shear-stress. For more details, please see Anandakrishnan and Alley (1997b).

The flexure of the ice at the grounding line due to the rising and ebbing tide has been detected at the surface of ice shelves (Williams and Robinson, 1980) and at the grounding lines of ice streams (Doake and others, 1987). In all these cases, the detected effects of the tidal forcing on the ice have been restricted to the floating ice and to the region directly inland of the grounding line up to a distance of a few ice-thicknesses. In this study we present evidence that the tidal effects are “seen” far inland (as much as 100 ice-thicknesses). During the austral summer of 1995-96, three arrays of short-period seismometers were deployed 10 km inland from the grounding line, 86 km from the grounding line, and 162 km from the grounding line (Anandakrishnan and Alley, 1997a). At the grounding line station, we detected a significant increase in seismicity at low-tide. We detected a pattern of increased seismicity at the site 86 km inland, but delayed 13 hours relative to low-tide. We hypothesize that the tidal forcing travels up the ice stream as a traveling wave with phase velocity $v = 76 \text{km/13hr} = 1.6 \text{ m/s}$. At the site 162 km inland, we did not detect any significant
change in seismicity over the tidal period and hypothesize that the forcing has decayed below the threshold for activating seismicity.

We model the ice stream system as a beam of thickness $H$ resting on a substrate of thickness $H_b$. The beam is acted upon by a pressure forcing due to the change in hydrostatic pressure associated with tidal fluctuations. On this time scale, ice responds as an elastic material (Sinha, 1978) and we make the beam elastic with Young's modulus $E = 3 \times 10^9$ Pa. We hypothesize that the substrate is a linear viscous material of viscosity $\eta$. For a sinusoidal stress forcing $P$ at the free end of an infinitely long system, the displacement at distance $x$ from the free end is an exponentially decaying sinusoidal travelling wave (Bott and Dean, 1973).

The disturbance moves up the ice stream with phase velocity $v$ that is a function of the forcing frequency the elastic-material parameters, and the viscous-substrate parameters. The amplitude of the disturbance decays to $1/e$ of the forcing amplitude at a distance $X$ from the free end. The phase velocity $v$ and the penetration distance $X$ are related to the material constants of the system by

$$v = 2 \frac{\omega E H}{\sqrt{2(\eta/H_b)}} ,$$

$$X = \sqrt{\frac{2E H}{\omega (\eta/H_b)}} ,$$

where $\omega$ is the forcing frequency, $E$ and $H$ are the elasticity and thickness of the ice, respectively, and $\eta$ and $H_b$ are the viscosity and thickness of the till, respectively.

Given $v$, we can solve for the ratio $\eta/H_b$ (a "basal stiffness"). For the observed $v = 1.6$ m/s, we calculate that $\eta/H_b = 1.5 \times 10^8$ Pa-s/m. For a basal till layer of thickness $H_b = 10$ m, $\eta = 1.5 \times 10^9$ Pa-s and for $H_b = 1$ m, $\eta = 1.5 \times 10^8$ Pa-s. This range of values for till viscosity $\eta$ is not unreasonable (Alley and others, 1987; Iverson and Hooke, 1994). We modeled this system numerically so we could introduce non-linear-viscous tills. The data are best fit by a linear-viscous till, but we cannot rule out a moderate ($p \sim 4$) power-law exponent non-Newtonian till. We find that a high ($p \sim 100$) power-law exponent is unlikely. These hypotheses can be tested by deploying a finer grid of seismometers on the ice stream. The change in wave propagation velocity with distance from the grounding line, and the rate of decay of the propagating wave will distinguish between the various models (as well as resolving the question of whether the propagation time is 13 hours or 13 hours plus some multiple of 24 hours). Note that the ice flow velocity variations (and the associated
basal shear-stress variations) are probably too small to be detected by GPS velocity measurements and the seismic proxy method must be used.

References


Late Holocene advance and retreat of tidewater glaciers in Yakutat Bay and Icy Bay, Gulf of Alaska

David Barclay¹, Julie Gloss¹, Parker Calkin¹ and Gregory Wiles²
¹ Department of Geology, University at Buffalo, Buffalo, NY
² Department of Geology, Macalester College, St Paul, MN

Introduction

The tidewater-calving termini of glaciers in Yakutat Bay and Icy Bay are currently located up to 60 km from the Gulf of Alaska coastline. However, glacial deposits record that these marine embayments and their tributary fjords were completely filled by ice during the late Holocene. Radiocarbon ages, tree-ring dates and historical reports are used to reconstruct the advance and retreat of these tidewater glacier systems during this most recent expansion.

Yakutat Bay

Hubbard Glacier advanced to dam Russell Fiord (Figure 1) around 2700 BP and the terminus then divided to simultaneously flow southward down Disenchantment Bay as well as along the North-West Arm of Russell Fiord (Figure 2a). The advance of the western lobe appears to have been continuous and culminated at about 810 BP (calibrated to calendar years = AD 1245). Advance of the eastern lobe slowed or reached a stillstand at about 2500 BP before reaching to the south end of Russell Fiord after 1730 BP (Figure 2a). Timing of ice advance from Nunatak Fiord is unclear, but the sequence of events during deglaciation suggests that Nunatak Glacier was supplying ice to the eastern lobe of Hubbard Glacier in Russell Fiord by the late Holocene maxima.

Minimum tree-ring dates along the east shore of Yakutat Bay suggest retreat of the western lobe was underway by AD 1327 and was rapid with at least 35 km of recession by AD 1471 (Figure 2b). The Spanish explorer Malaspina entered Disenchantment Bay in AD 1791; a landscape painting made at the time clearly shows features of the inner bay with Hubbard Glacier conspicuously absent from the field of view. This is consistent with journals and maps of Malaspina and Puget (in AD 1794) that indicate Disenchantment Bay was full of icebergs but lack reference to a calving ice cliff.
Figure 1. Location map of Yakutat Bay and Russell Fiord. Shaded areas represent glaciers.

Figure 2. Late Holocene glacier fluctuations in Yakutat Bay and Russell Fiord. Shaded areas represent glaciers, arrows indicate ice-flow directions, solid lines are dated ice-front positions and filled circles are other dated control points.
Considered together these historical sources suggest that in the 1790s the terminus of Hubbard Glacier was north of its current position.

With retreat of the western Hubbard lobe in Yakutat Bay, Nunatak Glacier became the primary source of ice to Russell Fiord. The lobe occupying the South Arm of Russell Fiord remained at its maxima into the late 18th century while a reversal of flow in the North-West Arm of Russell Fiord gave Nunatak Glacier a second terminus that faced the retracted Hubbard Glacier and calved icebergs into tidewater (Figure 2b).

The oldest trees on the terminal moraine at the south end of Russell Fiord germinated around AD 1780 suggesting initiation of ice retreat by this time (Figure 2b). A deep, ice contact proglacial lake expanded northwards with the retreating glacier terminus and drained into the Gulf of Alaska via the Situk River (Figure 1). Coeval retreat of the tidewater terminus in the North-West Arm of Russell Fiord led to the two termini intersecting in about AD 1860. This allowed the lake in the South Arm to drain as recorded by native traditions and by trees which germinated by AD 1866 on the lake treed at the south end of Russell Fiord. The new single calving front of Nunatak Glacier continued its retreat eastwards along Nunatak Fiord and today both East and West Nunatak Glaciers have retreated onto land (Figure 1).

Rates of advance of Hubbard Glacier during the late Holocene based on the radiocarbon data points in Figure 2a vary from about 10 to 46 m/yr. This is in good agreement with observed rates of advance during the past century that range from 16 to 47 m/yr (Trabant and others, 1991). Rates of retreat based on tree-ring and historical data (Figure 2b) vary from about 120 to 230 m/yr for the Yakutat Bay lobe while retreat of the Russell Fiord lobe between AD 1780 and 1860 was at about 400 m/yr. These rates are at the low end of the range of observed retreat rates of Alaskan tidewater glaciers (Meier and Post, 1987).

**Icy Bay**

Icy Bay is located about 100 km northwest of Yakutat Bay on the west side of the Malaspina Glacier (Figure 3). Retreat from an earlier advance left the fjords of inner Icy Bay ice free by about AD 1000 and during the next few centuries Alaskan coastal forest migrated into the deglaciated areas (Porter, 1989; Heusser, 1995).

Tree-ring cross-dates suggest twelve spruce logs at the mouth of Tsaa Fiord were overrun by Guyot and/or Yahtse glaciers between AD 1647 and 1650 (Figure 4a). Descriptions by Russian
Figure 3. Location map of Icy Bay. Shaded areas represent glaciers.

Figure 4. Late Holocene glacier advance in Icy Bay. Shaded areas represent glaciers, solid lines are ice-front positions and filled circles are other dated control points.
explorers in 1788, observations and a painting by Malaspina in 1791, and the journals of Vancouver and Menzies in 1794 all suggest that the Guyot/Yahtse ice margin had advanced to a position close to Kichyatt Point by the 1790s (Figure 4a).

In 1837 the terminus of the Icy Bay glacier system was observed close to the Gulf of Alaska coastline (Belcher, 1843). Tree-ring cross-dates of spruce logs at two sites confirm the complete filling of Icy Bay by ice during the first decades of the 19th century; outer rings of nine in situ stumps along the west shore of the outer bay date between AD 1805 and 1810, while those of 21 in situ stumps near Kageet Point date between AD 1811 and 1819 (Figure 4b).

Early maps record that the Icy Bay glacier had reached its maximum position by 1886 where it remained until 1907. Recession was rapid with over 40 km of retreat by 1978 (summarized in Porter, 1989).

The rate of advance between AD 1650 and 1790 was about 36 m/yr which is consistent with reported advance rates of tidewater glaciers elsewhere in the Gulf of Alaska (Meier and Post, 1987). However, the advance of at least 14 km between 1794 and 1837 is an order of magnitude greater than expected and so may have been a surge or series of surges. Such an interpretation is supported by native accounts of a rapid ice advance in Icy Bay (de Laguna, 1972) and observed surges of Svalbard tidewater glaciers in which termini have advanced as much as 20 km in less than three years (Solheim and Pfirman, 1985).

Conclusions

The late Holocene advance and retreat of tidewater glaciers in Yakutat Bay and Russell Fiord fits the general model of the tidewater glacier cycle (Post, 1975). Rates of movement are within the range of previously reported values (Meier and Post, 1987). In contrast, most of the late Holocene expansion of the Icy Bay glacier system appears to have occurred as a surge or series of surges which advanced the terminus at least 14 km in a period of less than 43 years.

Complete papers detailing the glacial histories of both Yakutat Bay-Russell Fiord and Icy Bay are in preparation.

Acknowledgements

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Le Conte Glacier, Alaska: surface elevation changes and calving retreat

Keith Echelmeyer and Roman Motyka
Geophysical Institute
University of Alaska
Fairbanks, AK 99775

Introduction

Le Conte Glacier is a tidewater glacier in southeast Alaska that has recently undergone a rapid calving retreat. Associated with this retreat are surface elevation changes along the glacier. An understanding of the relationship between these changes in thickness and in the terminus position is important for determining how a tidewater glacier responds to climate change, and how the entire glacier geometry is influenced by changes at the terminus. Here we report preliminary results on the elevation changes of Le Conte Glacier since 1948.

Background

Le Conte Glacier, near Petersburg, Alaska, is the southern-most tidewater glacier in the Northern Hemisphere. It is 35 km long, has an area of 487 km² and a high value of the accumulation area ratio (0.90). The glacier flows through the granodioritic Coast Ranges Batholith. Icebergs are discharged into a narrow, deep fjord; at the present terminus ice thickness is about 310 m and the water depth is 260 m. Surface ice speeds of up to 23 m/d are reached near the terminus.

The glacier retreated about 4 km between 1887 and 1962, it then stabilized until 1994. In late 1994 the glacier began rapid retreat by calving; the present retreat has been about 1300 m (Figure 1). Large calving events have been observed during this retreat and during superimposed seasonal fluctuations in terminus position (Motyka, this volume), with the majority of the ice being discharged by submarine calving.
Figure 1. Aerial photograph (US National Forest Service, July 30, 1985) of the lower Le Conte Glacier showing terminus positions. The crossed circles mark the observation ridge.
Figure 2. Panel a: ground tracks for profiles 1 and 2 on Le Conte Glacier (June, 1996) superposed over the glacier outline; horizontal coordinates are referenced to the NAD27 datum. Panel b: 1996 profile elevation minus the 1948 USGS map elevation as a function of surface elevation, for profile 1 (open circles) and profile 2 (filled circles).

Surface elevation profiles

In June 1996 we made a series of airborne laser profiles (Echelmeyer and others, 1996) along the glacier. These elevation profiles, which are accurate to about 0.3 m, have been compared to the glacier surface as shown on 1948 USGS topographic maps (Figure 2). The variability in the elevation change shown on the upper glacier in this Figure is due to errors in the 1948 map.

The results of the elevation change (Figure 2b) show that there has been significant thinning of the glacier over all of its area. A thinning of about 30 m occurred during the 1948-96 period over the
entire accumulation area (above 920 m); this thinning reached to the highest elevations, 35 km from the terminus. At the present terminus the glacier has thinned by 200 m. There is some uncertainty as to the exact date of the aerial photography used in making the maps, but the thinning has occurred at an average rate of 1 to 4 m/yr along its length.

Discussion

Some aerial photography exists that could be used for better time resolution of elevation changes, especially near the terminus, but these data have not been analyzed yet. The 1996 elevation profiles do indicate several interesting features, however. First, the thinning observed well up from the terminus cannot have been caused by terminal retreat since 1994. The large thinning near the terminus is probably due, in part, to the recent retreat. The terminus was stable from 1962 to 1994, and it only retreated about 700 m from 1948 to 1962. It is thus unlikely that the extensive elevation changes in the upper reaches of the glacier were caused by changes at the terminus over the period 1948 to present.

Based on the timing and magnitude of the observed retreat phases and the thinning, we can speculate on the relations between the two sets of observations. The recent retreat was probably caused by the thinning of the ice. This thinning would lead to a decreased ice flux near the terminus, and a decrease in terminal ice thickness. Such a decrease might then lead to an increase in calving rate, and a subsequent retreat. The cause of the glacier-wide thinning may be climatic, and/or it may be related to the substantial retreat which started in 1887. Further investigation of the patterns of thinning and retreat using existing photographs should help resolve the sequence of events. It is clear, however, that studies of the retreat or advance of tidewater glaciers must take the entire glacier into account, not just the dynamics of the terminal region.

Reference

Synthetic aperture radar (SAR) observations of calving glaciers

Richard R Forster
Byrd Polar Research Center
The Ohio State University
Columbus OH 43210

Introduction

Synthetic aperture radar (SAR) data have the ability to address the two fundamental observational needs of the calving glacier research community (identified in Section 2.2 of this Report), namely 1) documenting and detection of changes at regular intervals and 2) detailed measurements of topography and velocity. The presently orbiting SARs [ERS-1 and ERS-2, JERS-1, and RADARSAT] have a resolution on the order of 25 m and repeat time of about one month (Table 1). Therefore, these sensors are capable of detecting moderate scale changes (> 50 m/month) on a near-global basis, through clouds and darkness, from a time series of simple amplitude images. High resolution velocity (mm/day) and topographic (< 10 m vertical resolution) maps can also be generated using repeat-pass interferometric SAR (InSAR) techniques. However, the present inventory of potential InSAR data of calving glaciers is much more limited than the amplitude data and requires complex post processing.

The theoretical background of InSAR has been discussed in detail (Zebker and Goldstein, 1986; Gabriel and Goldstein, 1988; Gabriel and others, 1989) along with the specific application of the technique to glacier studies (Goldstein and others, 1993; Kwok and Fahnestock 1996; Joughin and

Table 1. Spaceborne synthetic aperture radars

<table>
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<th>Sensor</th>
<th>Wavelength (cm) / Freq. band</th>
<th>Ground Resolution (m)</th>
<th>Repeat Time (day)</th>
<th>Operational Dates</th>
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<td>5.7 / C-band</td>
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<td>3, 35, 176</td>
<td>8/91 - pres.*</td>
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<td>JERS-1</td>
<td>24 / L-band</td>
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<td>24</td>
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<td>LightSAR</td>
<td>TBD / L-band</td>
<td>25</td>
<td>8</td>
<td>proposed</td>
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</table>

* Not routinely acquiring data, operational for specific missions only

71
Figure 1. An interferogram of a portion of the South Patagonian Icefield, Chile, produced from SIR-C acquisitions on October 9 and 10, 1994. The phase contribution due to the different viewing geometries has been removed (flattened) leaving the phase variations to represent surface displacement and topography. The two orbits were almost identical over this area producing a small baseline (28 m) which diminishes the topographic contribution. Therefore, surface displacement dominates and this interferogram can be interpreted as a contour map of ice velocity in the direction of radar illumination (from the left at an incidence angle of 35°). The phase is shown varying from ±π (wrapped) with each cycle of gray tone representing 12.1 cm/day. The phase has be unwrapped and an InSAR DEM used to project the velocity field onto the surface of the ice by Forster and others (in prep.). North is to the bottom left with the ice margin shown as a black line. For a color version of this Figure, see p. 29 of this report.
others, 1996; Rignot and others, 1996a), therefore a simplified description of the process is sufficient for this discussion. Multiple SAR acquisitions over glacierized regions can be combined to produce high resolution ice velocity maps and digital elevation models (DEMs). The technique is based on the fundamentals of interferometry. First, two separate complex SAR images of the same area are acquired from slightly different positions separated in time by one to several days. [Repeat times on the order of several weeks are typically too long because of large ice motion displacement and changes in the scattering properties of the glacier due to melting or precipitation events.] The two images are then coregistered and their phases are subtracted forming a third image, termed an interferogram (Figure 1). The phase of the interferogram at each pixel is a combined effect of the displacement of the ice, the topography, and the different view points of the SAR for each acquisition. The latter two contributions to phase can be removed with ancillary information such as an existing DEM, estimates of the satellite orbits, and ground control points. If no DEM is available, one can be generated from an additional SAR acquisition. Once the displacement contribution to phase has been isolated it can be converted to a map of velocity in the radar line-of-sight by picking a point of zero or known velocity. The result is a map of one component of the ice velocity measured to an accuracy typically better than 1 cm/day for millions of points.

Review of InSAR applications to calving glaciers

Recent results demonstrate the utility and limitations of InSAR data to the study of calving glaciers in polar and temperate conditions. Most of the research has utilized data collected by ERS-1 while in its 3-day repeat orbit phase or data collected by both ERS-1 and ERS-2 during their tandem phase. [Between August 1995 and May 1996 the two ERS satellites where placed in the same orbit with ERS-2 lagging by about 24 hours creating an effective 1-day repeat tandem orbit.] This research has focused on the northern outlet glaciers of Greenland where the lack of surface melting and an undisturbed surface allows the phase of the signal returned to the SAR to remain correlated over periods as long as 6 days (Joughin and others, 1996a). In that study, InSAR derived DEMs where produced with relative accuracies of 2.5 m from ERS-1 and ERS-2 data with repeat times ranging from one to three days. A mini-surge which increased the speed 80 km from the terminus of Ryder Glacier by a factor of three was discovered with a time series of InSAR velocity measurements (Joughin and others, 1996b). The increased velocities were observed in October, 1995, with bracketing slower velocities observed in September and November, thus constraining the
duration of the surge event to less than seven weeks. The effect on the calving flux is not known because the terminus was not contained in the analyzed SAR scenes. Rignot (1996) estimated tidal motion of Petermann Glacier’s floating ice tongue to locate the hinge line from a sequence of three ERS-1 scenes and an existing DEM. The ice discharge was calculated at and below the hinge line using InSAR ice velocities and ancillary ice thickness profiles. The result, when compared to calf-ice production and surface ablation estimates, indicates that basal melting is the dominant process of mass reduction at the ice tongue. Rignot and others (1997) extended this method to measure ice discharge from 13 other glaciers in northern Greenland and found a similar deficiency in the calf-ice production confirming the extensive basal melting of the floating portions of the calving glaciers. For all glaciers combined, the total grounding line flux is less than the estimated balance flux, implying this portion of the Greenland Ice Sheet is thinning. The accuracy of InSAR velocities when using adequate tie-point information was found to be within a few meters per year based on data from Humboldt Glacier (Joughin and others, 1996c).

InSAR analysis for faster moving temperate glaciers experiencing surface melt, during and between the acquisitions, are more problematic, but have been successful using an L-band (24 cm) SAR. The Spaceborne Imaging Radar -C (SIR-C), a dual frequency (C and L-band) Shuttle-based SAR, collected data which included the calving outlet glaciers of the Patagonian Icefields with daily repeat times. The phase was decorrelated at C-band because of the high velocities and strain rates, rotation of seracs, and surface melting on both the North (Rignot and others, 1996a, b) and South Patagonian Icefields (NPI and SPI) (Forster and others, in prep.). However, at the longer wavelength L-band interferograms produced DEMs and velocity maps over most of the glacier surfaces. The advantage of L-band InSAR is due to the measurable displacement limit scaling with the wavelength and the longer wavelength signal also penetrating farther into the glacier surface where the scattering is probably less susceptible to changes from melt and precipitation. Results from the NPI showed that the San Rafael Glacier, Chile, has a central area of concentrated high velocity laterally bounded by slower moving ice (Rignot and others, 1996a). At the calving front velocities accelerate to over 17 m/day with longitudinal strain rates reaching over 1 yr⁻¹. SIR-C data of a portion of the SPI in Chile (Forster and others, in prep.) show that the three outlet glaciers in the radar scene (Europa, Penguin, and HPS-19) act as ice streams flanked by slower moving ice within the icefield (Figure 1). A flow divide between two of the glaciers is mapped by locating a narrow band of near-zero ice velocity. Horizontal ice surface velocity profiles calculated along flowlines show there is a high degree of spatial variability in velocity along the center-lines with velocities reaching over 5.5
m/day. Longitudinal strain rates calculated from these velocities at the locations of the initiation of crevassing agree with theoretical values computed for ice fracture under longitudinal tension.

**Potential for monitoring calving termini with SAR amplitude data**

A more straightforward application of SAR data to the study of calving glaciers uses only the amplitude data, without the phase, in a change-detection and feature-tracking mode. This can be done using pattern recognition techniques to estimate displacement vectors (e.g. Fahnestock and others, 1993). While the resolution of the measurable displacement is degraded from the millimeter scale, available from InSAR, to pixel scale (25 m), the data availability is increased by several orders of magnitude. For many calving glaciers throughout the world there is near-monthly coverage extending over several years. A monitoring program based on these data would allow for intra-seasonal measurements presently not available from visible satellite remote sensing data because of clouds and darkness or aerial photographs which impose additional logistical restrictions. A pilot study of Jakobshavn Glacier using a time series of SAR amplitude and visible remote sensing images observed large seasonal dependencies in calving rates (Sohn and others, this volume).

**Future applications of InSAR**

The previously acquired tandem data represents the best opportunity to continue InSAR observational studies of calving glaciers, because the ERS satellites are no longer in the tandem mode and ERS-1 will only be activated upon special arrangement with the European Space Agency. In addition to the Greenland calving glacier investigations, work is ongoing on the Antarctic ice shelves using the tandem data (Kwok, and others, 1996; MacAyeal and Rignot, 1996). There is extensive tandem data coverage of the calving glaciers of south-east Alaska from October, 1995, to May, 1996, for both ascending and descending orbits. Therefore, monthly velocity maps over the eight month period could begin to address the issue of spatial and temporal intra-seasonal velocity fluctuations (Section 4.3.2 of this Report).

Glacial velocities from a single member of the current fleet of spaceborne SARs may be useful for slower moving glaciers without surface melt or significant disturbance of surface features between acquisitions. These restrictions are the result of the long repeat times and the short C-band wavelength (although JERS-1 has a longer wavelength it also has the longest repeat time; Table 1). In order to produce InSAR products for faster moving glaciers, an L-band SAR dedicated to repeat-
pass interferometry, such as the proposed LightSAR mission, is required. One of the main objectives of the proposed mission (jointly supported by NASA and a commercial consortium) is the generation of InSAR ice velocity and topographic maps of the world’s glaciers and ice sheets. Therefore, the use of InSAR to help answer some of the remaining fundamental questions concerning calving glaciers can currently be increased by more fully utilizing the previously acquired ERS tandem data and in the future may be applied to all substantial calving glaciers.

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The relationship between calving rates, ice velocities and water depths

Terence J Hughes* and James L Fastook**
*Department of Geological Sciences and Institute for Quaternary Studies
**Department of Computer Science and Institute for Quaternary Studies
University of Maine
Orono, ME 04469

Abstract

Our interest in the relationship between fast glacier flow and ice calving dynamics began in 1970, when one of us (Hughes) was part of the First International Deception Island Volcanological Expedition, which was organized to study effects of the 12 August 1970 volcanic eruption. Deception Island (63°S, 60.6°W) is a collapsed volcanic caldera that lies north of the Antarctic Peninsula and that had experienced other eruptions in 1967 and 1969. The 1970 eruption produced several craters. One was 300 - 500 m in diameter and had produced a calving ice wall from 50 - 100 m high at the terminus of a glacier ~1500 m long. Shear bands rose nearly vertically from the bed and curved toward the high calving ice wall on the up-slope side of the glacier (Hughes, 1971). A proposal to study this phenomenon was funded, and field studies were conducted in the 1972-1973 and 1973-1974 Antarctic summers (Hughes, 1973). In the intervening two years, the upglacier ice wall had advanced 200 m into the crater and had lowered from 100 m to ~60 m. Ice flowing into the crater had produced conjugate normal faults. One set constituted ring faults that became crevasses toward the glacier surface, which was covered with volcanic ash. The ring faults were arcuate and extended ~500 m upglacier. The other set was localized within 50 m of the ice wall, and also became crevasses toward the surface. Ice slabs ~3 m wide between crevasses calved into the crater at irregular intervals. Horsts and grabens formed where both sets of faults were active. We studied both sets of faults.

Ice deformation associated with the ring faults was studied by installing two long strain networks along flowlines converging at an angle of 60° toward the crater, by measuring elevation and position changes of individual stakes not included in the strain networks, and by mapping surface elevation changes using aerial photogrammetric measurements on aerial photographs taken before and after the eruption (Hughes and others, 1974; Brecher and others, 1974). Shear along these faults was primarily responsible for lowering the ice surface and the ice wall subsequent to the eruption.
Ice deformation associated with the faults along which ice slabs calved was studied by carving four tunnels into the ice wall at various heights above the bed. These heights were accessible to us because surface ash had flowed into the crater at several sites and produced ash ramps against the ice wall at those sites. We studied strain inside the tunnels by mounting strain networks on the tunnel walls, measuring the offset of ash and dust layers across shear bands in the ice, and by conducting laboratory creep and fabric studies on the ice (Hughes, 1989; Hughes and Nakagawa, 1989). Shear bands associated with the normal faults extended downward, from the bottom of crevasses along which ice slabs calved to the base of the ice wall. Shear across shear bands decreased from up to 3 m near the surface to zero at the bed. This gradient of shear offset was explained by forward bending of the ice wall, like a book can be bent around its binding, with shear in shear bands being like bending shear between pages of the book.

Laboratory fabric studies showed that recrystallization of the ice had produced two single-maximum ice fabrics, one between shear bands that was due to nearly horizontal shear parallel to the bed, and one within the shear bands that was due to nearly vertical bending shear. Laboratory creep studies showed that recrystallization occurred after 30% strain at a shear stress of ~100 kPa, but did not occur at any strain when the shear stress was reduced to 50 kPa. Recrystallization increased the shear strain rate over tenfold, from 1.88 yr\(^{-1}\) to 21.4 yr\(^{-1}\). Creep obeyed the standard flow law for glacial ice

\[ \dot{\varepsilon}_{ZX} = \left( \frac{\sigma_{ZX}}{A} \right)^n \],

(1)

where \( x \) is horizontal, \( z \) is vertical, strain rate \( \dot{\varepsilon}_{ZX} = \dot{\varepsilon}_{XZ} \) by symmetry, \( \sigma_{ZX} = \sigma_{XZ} \approx \tau_0 \) for basal shear stress \( \tau_0 = \rho g H \alpha \), ice density \( \rho \), gravity acceleration \( g \), ice thickness \( H \), and surface slope \( \alpha \), \( n = 2.07, A = 86.3 \text{ kPa yr}^{1/n} \) before recrystallization and \( A = 25.3 \text{ kPa yr}^{1/n} \) after recrystallization (Hughes and Nakagawa, 1989).

Although the calving glacier on Deception Island was virtually temperate, the mean annual air temperature was \(-3 ^\circ C\) and the glacier apparently had a frozen bed. Most calving glaciers have a thawed bed, so sliding at the bed is possible and the bed can deform if it consists of wet sediments or till. In addition, most calving ice walls are grounded in water of variable depth. If the bed of the calving glacier is thawed, the basal water pressure at the ice wall can force basal water between basal ice and the bed and through permeable basal sediments or till for some distance upglacier from the ice wall, as shown in Figure 1. The linear increases of ice and water pressure with depth are shown
exerting horizontal gravitational forces acting on opposite sides of ice columns at the ice wall and at various distances upglacier, depending on water height at the ice wall. Basal water pressure would raise basal water to height $H_w$ in a borehole through ice of thickness $H$. Slope $\alpha$ of the glacier surface decreases as $H_w / H$ increases.

The velocity of a glacier sliding on a wet deforming bed is the sum of an average creep velocity, $u_c$, a basal sliding velocity, $u_b$, and a maximum bed velocity, $u_0$. The average creep velocity for temperate ice is

$$u_c = \frac{2H}{n + 2} \left( \frac{\tau_0}{A} \right)^n.$$  \hspace{1cm} (2)

Experimental and theoretical sliding velocities generally can be represented by the expression

$$u_b = \left[ \frac{\tau_0}{B(1 - P_w / P_b)^2} \right]^m,$$  \hspace{1cm} (3)

where $m$ is a viscoplastic parameter, $B$ is a bed traction parameter, $P_w$ is basal water pressure, and $P_b$ is ice overburden pressure. A number of deformation laws for sediments or till have been proposed, but the limited data available are generally represented by the expression

$$u_0 = 2\dot{\varepsilon}_v \lambda \left[ \frac{\tau_0}{\sigma_0 (1 - P_w / P_b)} \right]^\nu,$$  \hspace{1cm} (4)

where the bed deforms to depth $\lambda$, $\dot{\varepsilon}_v$, is a creep rate that depends on viscoplastic properties of the deforming layer, $\sigma_0$ is its yield stress, and $\nu$ is a viscoplastic parameter such that $1 \leq \nu \leq \infty$.

In Figure 1, horizontal arrows show how water and ice pressure increase with depth, where $P_w = \rho_w g H_w$ and $P_b = \rho g H$ are the respective water and ice pressures at the bed for respective water and ice densities $\rho_w$ and $\rho$, and gravity acceleration $g$. Triangular and rectangular areas enclosing these arrows show the opposing horizontal forces. At an ice wall calving in water of variable depth as water depth $H_w$ increases, water back-pressure increasingly offsets ice forward-pressure, causing basal shear stress $\tau_0$ to decrease with a corresponding decrease in surface slope $\alpha$. Balancing forces and moments at the ice wall leads to the following expression for the calving rate $U_c$ (Hughes, 1992):
Figure 1. Longitudinal gravitational forces at a calving ice wall and upglacier from the ice wall.

Top: horizontal forces acting on an ice column at the ice wall are due to basal ice pressure $P_i$, basal water pressure $P_w$, and basal shear stress $\tau_o$, shown for an ice wall just floating in water so that $\tau_o = 0$ (right), grounded in water so that $\tau_o > 0$ (center), and grounded on land so that $\tau_o >> 0$ (left). Middle: horizontal forces acting on ice columns at the ice wall and upglacier are due to $P_i$, $P_w$ and $\tau_o$, where $P_w = P_i$ for ice just floating at the ice wall (right), $0 < P_w < P_i$ for ice grounded on a thawed wet bed upglacier from the ice wall where $P_w$ is equivalent to water of height $H_w$ (center), and $P_w = 0$ for ice grounded on a thawed or frozen bed further upglacier (left). Bottom: variations of ice height $H$ (solid curve) and water height $H_w$ (dashed curve) above the bed (hatchured line) in boreholes along distance $x$ upglacier from a calving ice wall that is just afloat (left), across a thawed wet bed (center), to a thawed damp bed or even a frozen bed (right).
\[ U_c = \frac{3 \rho g H^4 \theta}{\eta_v X_c} \left( 1 - \frac{P_w}{P_i} \right), \]

where \( \theta \) is the forwarding bending angle of nearly vertical shear bands of the kind studied on Deception Island which cause the ice wall to lean forward, \( X_c \) is the horizontal distance from the ice wall to the transverse crevasses along which the next slab will calve, and

\[ \eta_v = \frac{\partial \sigma_{xz}}{\partial \varepsilon_{xz}} \approx \frac{A^n}{\tau_0} \text{ is the viscoplastic creep viscosity obtained from equation (1) when } \sigma_{xz} \approx \tau_0. \]

Figure 1 shows a longitudinal cross-section along a surface flowline of a calving glacier on a horizontal bed. Horizontal gravitational forces per unit width of the glacier are represented by the areas of triangles and a rectangle containing horizontal arrows on the front and back faces of vertical ice columns, where arrows represent vertical variations of ice and water pressure acting on these faces. When the net horizontal gravitational force is balanced against the longitudinal deviator stress, \( \sigma'_{xx} \), and its longitudinal gradient, \( \frac{\partial \sigma'_{xx}}{\partial x} \), and the traction forces of basal shear stress, \( \tau_0 \), and side shear stress, \( \tau_s \), acting on the respective basal and side faces of the ice columns, and this force balance is combined with the mass balance, the expression for surface slope \( \alpha = \frac{\Delta h}{\Delta x} \) is (Hughes, 1996)

\[ \frac{\Delta h}{\Delta x} = \left(1 - \frac{\rho}{\rho_w}\right) \left(1 - \frac{P_w}{P_i}\right)^2 \left\{ \frac{MH}{Mx + H_0 U_0} - \frac{H^2 R^{n-1}}{Mx + H_0 U_0} \left[ \frac{\rho g H}{4A} \left(1 - \frac{\rho}{\rho_w}\right) \left(\frac{P_w}{P_i}\right)^2\right]^n \right\} + \\
+ H \left(1 - \frac{\rho}{\rho_w}\right) \left(\frac{P_w}{P_i}\right) \frac{\Delta (P_w / P_i)}{\Delta x} + \frac{2 \tau_s}{\rho g W} + \frac{\tau_0}{\rho g H}, \]

where elevation change \( \Delta h \) occurs over horizontal distance \( \Delta x \), \( M \) is surface accumulation rate, \( x \) is distance measured upglacier from the calving ice wall, \( H_0 \) and \( U_0 \) are the respective ice thickness and velocity at \( x = 0 \), \( W \) is the glacier width, \( R \) represents summed ratios of surface strain rates (\( R = 1 \) for one-dimensional flow), and \( \tau_s = 0 \) for the central flowline meeting the calving ice wall.
Figure 2. The ratio of ice velocity $U_S$ to calving rate $U_C$ of a calving glacier at an ice wall grounded in water of depth $H_W$, of height $h$ above the water surface, and moving at various sliding velocities $u_s \approx U_S$. The curves are obtained by combining equations (5) and (6).

Equation (6) applies for variable bed topography when $H = h - h_b$, where $h$ is ice surface height above water level and $h_b$ is bed height above or depth below water level. Setting mass-balance velocity $U_S = M(L - x)/H$ as the sum of $u_c$, $u_s$, and $u_o$ at distance $x$ from the ice wall along a flowline of length $L$ for accumulation rate $M$, equations (2) through (4) can be solved for $\tau_o$ in equation (6) with $L$, $M$, and the quantities in these expressions (Hughes, 1996).
Figure 2 is a plot of the ratio $U_c/U_s$ versus $H_w$ for various values of $h$ and $U_s$. It is obtained by combining equations (5) and (6) for $\theta \approx \Delta h/\Delta x$ over distance $X_c$, for which $H_0U_0 \gg MX_c$, $H_0 \approx H$, and $-U_0 \approx U_s = u_c + u_s + u_o \approx u_s$ because $x$ is positive upglacier, $u_c << u_s$ over $X_c$, and $u_o = 0$ when sediments or till stop deforming at the ice wall. Note that $U_c/U_s$ increases as $U_s$ decreases, such that the calving rate exceeds the ice velocity when ice is slow but lags behind ice velocity when ice is fast, with big changes in $U_s$ causing smaller changes in $U_c$.

The bending shear mechanism observed on Deception Island depends on bed traction, so it should vanish as a calving ice wall becomes afloat. This is seen in equation (5), where $U_c = 0$ when $P_w/P_1 = 1$. As seen in Figure 1, however, the asymmetry of the horizontal gravitational force at the calving front still exists when $\tau_o = 0$ for floating ice, so a bending moment still exists that will cause the calving front to bend downward and the floating glacier behind the calving front to arch upward, as shown in Figure 3. This phenomenon has been analyzed by Reeh (1968) and by Fastook and Schmidt (1982). If a calving front is barely afloat, this downwarping will cause it to maintain contact with the bed until the water becomes deep enough to break contact. Until that happens, the

![Figure 3](image)

**Figure 3.** The bending shear mechanism of crevasse formation before and after a glacier becomes afloat. Left: bending shear at the calving ice wall when the glacier is grounded along its entire length. Right: bending shear after the glacier becomes afloat. Downwarping of the calving ice wall and arching behind the wall allow the ice wall to contact the bed, thereby providing the basal traction needed to allow bending shear. Reeh (1968) analyzed the stress regime when the ice wall is also afloat. Until then, the ice wall can act like a plow that scrapes sediments from bedrock, thereby creating a moraine that dams water under the glacier.
the calving ice wall will scrape across the bed even though the glacier is afloat behind the ice wall. This has two consequences. First, it maintains the bending shear mechanism over distance $X_c$ where bed contact is maintained, so ice slabs can still calve. Second, it will act like a plow that scrapes basal sediments or till from bedrock, pushes this pile forward as the ice wall advances, and creates a terminal moraine that may eventually buttress the ice wall enough to halt its advance. Such moraines are common across fjords that had been occupied by calving tidewater glaciers.

References


Tidewater terminus dynamics in Glacier Bay, Alaska

Lewis E Hunter
U.S. Army Cold Regions Research and Engineering Laboratory
72 Lyme Road
Hanover, NH 03755

Abstract
Asynchronous and complex behavior of glaciers with tidewater termini can often be attributed to the glacier’s response to calving. Any external forcing that can cause the balance between the terminal ice flux and calving flux to shift can influence advance, retreat and stillstand phases. Recent studies in Glacier Bay, Alaska, document periods of terminus stabilization and moraine formation. In this paper, the recent histories of Grand Pacific and Muir glaciers are presented. Grand Pacific Glacier advanced through most of the last half century while Muir Glacier only recently stabilized after 100 years of retreat. Their dynamics appear unrelated to climatic forcing, but instead reflect internal adjustments to calving and glacier dynamics.

Introduction
Observations of tidewater terminating glaciers in Glacier Bay, Alaska (Figure 1), began with Vancouver in 1794 and have been continued by Wright (1887), Reid (1892, 1895), Muir (1902), Cooper (1937), Field (1947, 1964), Goldthwait (1966), Powell (1980, 1984), Hunter (1994) and many others. As a result, the area has the best documented history of tidewater glacier retreat in North America. That history is characterized by diachronous advance and retreat phases (Figure 2). Powell (1991) and Hunter and Powell (1995a) suggest that this diachronous behavior is partly controlled by meteorological variability between the Fairweather Range and Takhinsha Mountains (Figure 1); however, grounding-line sedimentation can rapidly change water depth at the base of a tidewater ice cliff, which imposes a second-order control on calving speed (e.g. Powell, 1991).
Figure 1. Map of Glacier Bay National Park and Preserve showing the locations of (1) upper Muir Inlet and Muir Glacier, and (2) Tarr Inlet with Grand Pacific Glacier.

Sedimentological investigations in Glacier Bay have documented extremely high sedimentation rates (Cowan and Powell, 1991; Powell, 1991; Hunter and others, 1996a) fed by the highest known basin-wide sediment yields (Hallet and others, 1996). Process studies show that morainal banks deposited at the grounding line can aggrade or collapse on the order of several tens of meters in a few weeks to a month (Hunter and Powell, 1995b; Hunter and others, 1996b). Rapid changes in crevasse patterns, terminus profile and ice-flow velocities induced by sediment dynamics (Hunter, 1994) show that terminus stability is closely related to grounding-line water depth (Brown and others, 1982; Powell, 1991).
The growth and decay of morainal banks are controlled by a delicate balance between glaciofluvial, glaciotectonic and gravitational processes (Hunter and others, 1996b). A process hierarchy, based on detailed process studies, shows that it is the balance between glaciofluvial inputs (bedload dumping and suspension settling from overflow plumes) and gravitational processes (slides, slumps and debris flows) that will determine whether a bank will grow or decay (Table 1). Seasonal analyses also show that ice push is a primary mechanism by which banks grow during winter months. Release of debris from glacial transport (basal, subglacial, englacial) by meltout is one to two orders-of-magnitude less than glaciofluvial inputs and is relatively insignificant to the overall sediment budget (Hunter and others, 1996a and b). These sedimentary processes are important because they control the rate of morainal bank buildup and determine whether or not the grounding line will shoal or be maintained during an advance, and thus, dictate a glacier’s ability to build a morainal bank and maintain that bank as the glacier advances into a deep water fjord (e.g. Post, 1975; Mayo, 1988).

**Table 1: Sedimentary process hierarchy**

<table>
<thead>
<tr>
<th>Order</th>
<th>Volume Flow Rate (m$^3$/yr)</th>
<th>Processes</th>
</tr>
</thead>
<tbody>
<tr>
<td>First Order</td>
<td>$(10^6 - 10^7)$</td>
<td>- glaciofluvial dumping</td>
</tr>
<tr>
<td></td>
<td></td>
<td>- mass movements</td>
</tr>
<tr>
<td>Second Order</td>
<td>$(10^5 - 10^6)$</td>
<td>- plume settling and by-pass</td>
</tr>
<tr>
<td></td>
<td></td>
<td>- glaciotectonic deformation</td>
</tr>
<tr>
<td></td>
<td></td>
<td>- soft-bed transport</td>
</tr>
<tr>
<td>Third Order</td>
<td>$(10^4 - 10^5)$</td>
<td>- subglacier freeze-on</td>
</tr>
<tr>
<td></td>
<td></td>
<td>- iceberg rafting</td>
</tr>
<tr>
<td>Fourth Order</td>
<td>$(10^3)$</td>
<td>- dumping by calving</td>
</tr>
<tr>
<td></td>
<td></td>
<td>- ice-cliff meltout</td>
</tr>
</tbody>
</table>

**Terminus behavior**

Recent studies of tidewater termini in Glacier Bay have focused on the glaciers at the heads of Muir and Tarr inlets. Two glaciers in these inlets have opposite histories over the last half century: Grand Pacific Glacier advanced into Tarr Inlet and Muir Glacier retreated in Muir Inlet. In terms of a glacier cycle, the behavior of Grand Pacific Glacier reflects a quasi-stable advance attributable to
disequilibrium at the head of the fjord following catastrophic retreat (e.g. Mann, 1986), while Muir Glacier was nearing the end of its retreat phase.

Grand Pacific Glacier

Maximum tidewater retreat at Grand Pacific Glacier occurred in about 1929, when the terminus was in Canada (Figure 2; Field, 1964). The first decade of advance was characterized by periods of surges followed by rapid retreat. By the early 1940s, this behavior switched to one of continuous advance (~30 m/yr) that was maintained into the early 1990s (Hunter, 1994). The Grand Pacific advance began after Cooper (1937) observed that its grounding line was shoaling. Shoaling continued through the 1960s (Field, 1964) as the glacier advanced. It was noted that fluctuations in glacier dynamics near the terminus changed during the 1970s (Hunter, 1994) as shoaling decreased and grounding-line water depths increased to about 30 m (Brown and others, 1982). Ice flow to the terminus accelerated during that time and was accompanied by thinning of the lower reaches of the glacier (Hunter, 1994). Subsequent observations showed that ice flow velocities decreased as the grounding line again began to shoal, reaching water depths of about 5 m by 1991 (Bograd, 1997).

The jagged seracs and crevasses near the terminus were replaced by a rounded and gently sloping surface by 1996. The height of the ice cliff increased from about 35-40 m in 1996 to 60-70 m by March 1997 as the terminus began to retreat off of the morainal shoal. Lateral sections of the old morainal bank now project into the inlet and are subaerially exposed, having been isolated by this retreat. These observations support the idea that the advance of Grand Pacific Glacier resulted from the glacier trying to adjust its surface slope once rapid calving ceased (e.g. Mann, 1986).

Development of the morainal bank protected the grounding line from deep water (e.g. Post, 1975; Mayo, 1988) and allowed the terminus to advance into Tarr Inlet, which is over 300 m deep (Hunter, 1994).

Muir Glacier

Muir Glacier has generally retreated rapidly (~0.4 km/yr) since pulling off of its moraine at the mouth of Muir Inlet in 1899 (Field, 1947; Goldthwait, 1986; Hunter, 1994). The retreat was punctuated by stillstand and minor readvance phases in 1890-1899, 1926-1929, 1946-1950 and 1971-1974. Each stillstand was followed by periods of accelerated retreat, reaching rates as high as 2.7 km/yr (1976-1977). Retreat slowed after 1979 and a position of stability was achieved around 1986 (Powell, 1991) after 45 km of retreat (Figure 3). The morainal bank became intertidal in the center
Figure 2. Recent retreat and advance history of Grand Pacific Glacier since 1879.
Figure 3. Retreat history of Muir Glacier since 1860.
of the fjord along the glacier during 1988 and aggraded until the grounding line became terrestrial by 1993. Observations between 1994 and 1996 indicate that Muir Glacier continued to thin despite the rapid reduction in calving. Such a response does not fit current models of glacier response to reaching the heads of their fjords (e.g. Mann, 1986). Instead it shows that the surface of the snowfields that feed the terminus through two narrow ice falls dropped to critical levels where ice flow to the terminus is now restricted by bedrock geometry.

Figure 4. Mass balance indices of (A) Grand Pacific and Ferris and (B) Muir Glaciers.
The terminus initially stabilized before morainal bank build-up, showing that bank development may not be the triggering mechanism to stabilization. Mass balance indices (Figure 4) show a rapid change during the mid-1970s, reflecting the loss of slow moving, lateral ice as the terminus rapidly retreated into a narrow segment of the fjord. The change in glacier hypsometry is recorded by a rapid rise in the accumulation area ratio while the equilibrium line altitude was situated in a relatively climatic-insensitive position (on the steep slope of the ice falls). The position of recent stability occurs directly below two ice falls and the confluence of their tributaries. Encroachment of the terminus caused an increase in surface slope, rapid surface lowering (Hunter, 1994), and should have caused an acceleration of ice flow to the terminus. Stabilization, therefore, likely occurred in response to an increasing ice flux to the terminus that eventually balanced the reduced calving flux where the cross-sectional area was reduced. Therefore, the glacier likely stabilized through glaciodynamic adjustments, despite the continuation of rapid calving.

Continued thinning is unexpected and can only be explained by the glacier having become overthinned during the late stages of its rapid retreat. Under normal conditions, slope oversteepening is expected to cause advance after a morainal bank develops as the glacier adjusts its length and surface slope (Mann, 1986; Powell, 1991). However, Muir thinning appears to have reached a critical threshold, imposed by the bedrock configuration under the ice falls that is now restricting ice flow to the terminus.

Conclusions

Recent behavior of Grand Pacific Glacier and Muir Glacier termini appears to be a response to non-climatic forcing. Initiation of advance at Grand Pacific Glacier followed the buildup of a morainal bank. Flow dynamics and advance rates have been linked to the behavior of the morainal bank, supporting the model that tidewater glacier advance into a deep basin is controlled by the development and maintenance of a moraine shoal (Post, 1975; Mayo, 1988).

Stabilization of Muir Glacier during the past decade reflects a type of glaciodynamic response that may be characteristic of tidewater glaciers as they reach the heads of their fjords. Accelerated ice flow and increased surface slopes characterize the lower reaches just upglacier of the calving margin. As the terminus of Muir Glacier retreated around the last bend (Figure 3), increased surface gradients in the terminal reach would have been able to penetrate the two ice falls at the head of the fjord and tap the adjacent snowfields. During the late 1980s, the flux of ice out of these snowfields
would have increased in response to steeper surface gradients near the terminus and appears to have balanced the calving flux, allowing the terminus to stabilize. Rapid fjord infilling during the early 1990s coincided with slowed ice flow through the ice falls (Hunter, 1994). Continued surface lowering during the mid-1990s indicates that ice flow is being restricted through the icefalls above the terminus. Rapid sediment accumulation at the grounding line and transition to a terrestrial terminus by 1993 reflects a dynamics balance between waning flow through the icefalls, as the snowfield thinned, and increased grounding line stability ascribable to lower calving rates at the terminus.

References


Hans Glacier - a tidewater glacier in southern Spitsbergen: summary of some results

Jacek Jania and Marzena Kaczmarska
Department of Geomorphology
Faculty of Earth Sciences
University of Silesia
Sosnowiec, Poland

Introduction

The majority of glaciers in Svalbard terminate in sea water as ice cliffs (Hagen and others, 1993). Grounded tidewater glaciers are the most common type of ice masses there. Therefore, mass loss due to calving must be considered an important component of their mass balance.

Since 1982, studies of Hans Glacier in South Spitsbergen have focused on measurements of calving flux and the factors which influence the calving speed. The results form one of the longest records for Svalbard and the Arctic. The objective of this paper is to summarise the most important results of the research program within the context of data on calving of other Svalbard and Arctic tidewater glaciers.

Hans Glacier was first described during an Austrian expedition in 1872, when a sketch map was made. The first glaciological work was done in 1938 and consisted of large-scale mapping and surface velocity measurements using terrestrial photogrammetry (Pillewizer, 1939). Further investigations were carried out after the Polish Polar Station was established in the vicinity of the glacier in 1957 (Kosiba, 1960; Baranowski, 1977; Figure 1). The station was renovated and modernised in 1978 and has been operating continuously since then. Systematic monitoring of fluctuations of the glacier front and surface velocity by terrestrial photogrammetry started in 1982 (Jania and Kolondra, 1982; Jania, 1988). In 1988, new projects on the glacier were initiated concerning surficial mass balance components, investigations of the glacier drainage system and the internal thermal structure of the glacier. Recently, studies on the energy balance of the ice surface and snow/firn chemistry were initiated.
Figure 1. Location map of Hans Glacier with research posts near its frontal part.

**Hans Glacier**

Hans Glacier is a valley glacier, with a compound basin. Its main accumulation field is connected with adjacent glaciers. The glacier extends from sea level in the Hornsund Fjord to 600 m above sea level (a.s.l.). The glacier covers an area of 57 km$^2$, and is about 16 km long; the maximum ice thickness exceeds 400 m, and the mean surface slope is $1^\circ 40'$. The glacier tongue terminates in a $\sim$1.3 km long active ice cliff. Lateral parts of the glacier rest on land (Figure 1). Temperature measurements in nine holes suggest that the glacier is of the polythermal type. The cold ice layer of 40-100 m thick is underlain by $\sim$60-250 m of temperate ice (Jania and others, 1996). In the accumulation zone, firn and ice are at the pressure melting point temperature throughout the whole thickness. Hans Glacier is considered to be a surge-type glacier.
Climatic characteristics of the area are known from observations at the meteorological station located near the Polish Polar Station (9 m a.s.l.) since July 1978. At sea level, the annual mean air temperature is -4.9 °C (1978-1995) and annual precipitation exceeds 410 mm water equivalent (w.e.). Winter precipitation at higher elevations on the glacier exceeds 1200 mm w.e. The mean air temperature during the summer months is +3.2 °C, and -11.5 °C during the winter. Strong eastern winds in summer and winter are characteristic features of the local climate.

The surface net balance of Hans Glacier was negative during most years between 1989 and 1995 (-0.24 m w.e. on average). The mean winter balance is +0.9 m w.e. and the mean summer balance -1.14 m w.e. for the period. The average ELA is recorded to be about 350 m a.s.l., but its elevation fluctuated between 240 m and 400 m a.s.l. during the period of observation. Due to the small surface slope of the glacier, the AAR varies significantly from year to year (38% is the mean value). Data on the mass balance of the glacier are given in Table 1.

<table>
<thead>
<tr>
<th>Balance year</th>
<th>Winter balance (m w.e.)</th>
<th>Summer balance (m w.e.)</th>
<th>Net balance* (m w.e.)</th>
<th>Net balance** (m w.e.)</th>
<th>ELA (m a.s.l.)</th>
<th>AAR (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1989</td>
<td>0.91</td>
<td>-1.00</td>
<td>-0.09</td>
<td>-0.44</td>
<td>365</td>
<td>39</td>
</tr>
<tr>
<td>1990</td>
<td>0.90</td>
<td>-1.44</td>
<td>-0.54</td>
<td>-0.89</td>
<td>380</td>
<td>27</td>
</tr>
<tr>
<td>1991</td>
<td>1.16</td>
<td>-1.03</td>
<td>0.13</td>
<td>-0.22</td>
<td>280</td>
<td>54</td>
</tr>
<tr>
<td>1992</td>
<td>0.89</td>
<td>-1.16</td>
<td>-0.27</td>
<td>-0.08</td>
<td>380</td>
<td>27</td>
</tr>
<tr>
<td>1993</td>
<td>0.93</td>
<td>-1.61</td>
<td>-0.68</td>
<td>-1.03</td>
<td>400</td>
<td>22</td>
</tr>
<tr>
<td>1994</td>
<td>0.76</td>
<td>-0.56</td>
<td>0.20</td>
<td>-0.15</td>
<td>240</td>
<td>69</td>
</tr>
<tr>
<td>1995</td>
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<td>-1.21</td>
<td>-0.45</td>
<td>-0.80</td>
<td>390</td>
<td>25</td>
</tr>
<tr>
<td>Cumulative</td>
<td>-</td>
<td>-</td>
<td>-1.70</td>
<td>-3.61</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Average</td>
<td>0.90</td>
<td>-1.14</td>
<td>-0.24</td>
<td>-0.52</td>
<td>348</td>
<td>38</td>
</tr>
</tbody>
</table>

Notes: * - net balance of the glacier surface, ** - net balance including calving flux

Some results of studies

Calving flux has been calculated using surveys of the glacier front position changes, velocity of the glacier and cross section of the glacier tongue at the ice cliff. Repeated stereoscopic photos were taken from three permanent photogrammetric bases, located near the glacier terminus (Figure 1). The time parallax (pseudoparallax method) on converted photos was employed (Kolondra, 1984).
Figure 2. Changes of the Hans Glacier front position in the 20th century against the background of the bathymetry of its forebay (thin lines: isobaths in meters, after the bathymetric chart by J. Różański, unpubl.).

The average time interval between series of photographs varied between 3 and 6 weeks except for the period of total darkness during the polar night. The terminus cross section was determined by radio-echo sounding of the sea depth near the ice cliff. Comparison of the results of the low
Figure 3. Dynamics of Hans Glacier in 20th century: surface velocity of the glacier ($V_g$), calving speed ($V_c$), glacier length and rate of change of glacier length, against the background of the sea depth changes ($h_w$) in front of the ice cliff (average data for the whole width of the cliff during the observation period, as well as maximum depth).

frequency echo sounding of the glacier thickness done in 1989 (Glazovsky and others, 1991) was very useful. Changes of the glacier length have been recorded at the centerline and averaged over the whole width of actively calving cliff (1250 m).

Terminus position shows a seasonal pattern of fluctuations, superimposed on a general retreat. In winter, the glacier advances slowly (~50 m) while the summer retreat exceeds 90 m. The mean
annual retreat is about 40 m. General retreat of the glacier has been observed during the 20th century (Figure 2) but with some minor advances (Figure 3), the last of which occurred in 1995.

Glacier velocity has been surveyed by identification of the displacements of artificial and natural targets on the surface of the lower part of the glacier. Results were normalised to two transverse profiles (Figure 1), located close to the front ("F" profile) and about 0.5 km upglacier ("P" profile). The mean velocity at the front profile (~210 m/yr) is more than three times faster than at the "P" profile. The highest velocity occurs during the first part of ablation season. Using data from the photogrammetric surveys, and comparison to measurements of velocity of Kongsvegen Glacier (northwest Spitsbergen) and the water discharge from the adjacent Werenskiold Glacier basin, the annual trend in average glacier velocity was determined (Figure 4). Estimates of the average velocity during the winter are based on the rate of front advance, because calving is either very small or does not occur at all during winter (Jania, 1988).

The depth of the sea was obtained from a navigation chart 1:10 000 (J. Różański, unpubl.) with 10 m contour interval (Figure 2) and sonar sounding along the ice cliff in 1985. The average depth of the sea at the glacier snout varies between 20 and 25 m, while the mean elevation of the ice cliff is 30-40 m.

The annual mean calving speed of Hans Glacier is 250 m/yr. Intense calving begins simultaneously with the summer increase in glacier movement in July. The calving intensity decreases at the end of August, to increase again in autumn (Figure 4). Hans Glacier looses about $22.25 \times 10^6$ m$^3$ of ice annually by calving. This constitutes about 23% of the total mass loss, and is equivalent to a surface ablation rate over the whole glacier of -0.35 m w.e. (in average). As a result, the glacier net balance is negative in every year (Table 1).

Hans Glacier is a grounded tidewater glacier that produces small icebergs or ice brash only. The calving rate changes during the year from ~0 m/d in winter up to ~2.0 m/d in July and the end of August (Figure 4).

Discussion of results

Changes in glacier length, average velocity close to the front, rate of advance and retreat, as well as maximum and average depth of the sea at the ice cliff in the 20th century suggest that calving averaged over longer time periods (a year and more) may be controlled by glacier velocity, instead of
Figure 4. Average and smoothed annual trend in velocity of Hans Glacier at the profile “P” ($V_\text{g}$) and calving speed ($V_\text{c}$) as a function of the seismic activity of the glacier: total daily relative energy of quakes ($E_\text{r}$) and daily total number of glacier quakes (N). The seismic data represent averages for the period 1980 - 1984).

only by the average water depth at the glacier front (Figure 3). The calving speed of Hans Glacier in 1982-1984 was 250 m/yr and the ice velocity 210 m/yr; the average sea depth was 25 m. Measurements carried out on another tidewater glacier in Spitsbergen (Kongsvegen) show a mean calving speed of ~440 m/yr and an ice velocity of 414 m/yr with the average depth at the ice cliff ~47 m (Voigt, 1979). Voigt’s data show that changes in glacier extent in 1962-1964 occurred at the same average depths, while the ice velocity measured in 1962 on the centerline of the glacier was significantly higher than in the summer of 1964. Even with a front advance of ~120 m/yr the estimated calving speed is about 600 m/yr (Jania 1991). Kongsvegen is a surge-type glacier.

The statistical relation between glacier velocity and calving speed is well pronounced when available data from Spitsbergen and Alaskan glaciers are considered (Figure 5). There are also functional relations between glacier movement and the calving speed. Ice velocity controls calving

by stretching and fracturing of the frontal parts of tidewater glaciers. Glacier movement near the ice cliff is mainly due to basal sliding. The glacier terminus velocity is also related to the buoyancy forces (ice thickness/water depth ratio), which reduces friction against the bed. Periodic or episodic
increase in basal sliding causes acceleration of calving. Massive calving events appear during the active phases of surges, as observed in Svalbard in several cases.

The typical mean iceberg production for glaciers in Spitsbergen is 25% of the total mass loss (Jania, 1988; Glazovskyi and others, 1991; Jania and Hagen, 1996). However, Lefauconnier (1994) estimated that Svalbard loses from 7.44 to 9.94 km³ of ice per year, corresponding to about 0.21 - 0.28 m w.e. of surface melting; this represents about 50% of the ablation at the glacier surface and up to 30% of the total mass loss. Despite the uncertainty in the data, the mass loss by calving is very important for the mass budget of the Svalbard glaciers.

Data collected for Arctic glaciers show that calving ranges from several percent of mass loss from total ablation (3 - 4% on De Long Islands) to up to ~40 - 60% in the case of Western Greenland and glaciers in southern Alaska. For other Arctic regions, calving accounts for about 20 to 35% of the total mass loss (Chizhov 1976; Jania 1988; Govorukha 1989; Jania and Hagen 1996). Such differences are probably related to regional climatic conditions. Despite such differences, mass balance measurements carried out in the Arctic point out the negative net balance (including calving flux) for the majority of tidewater glaciers. Calving is an important mode of transposition of ice mass from land to the sea, in addition to the discharge of meltwater, and therefore contributes to the general sea level rise.

Acknowledgements

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References


Documentation of the retreat of Columbia Glacier, Alaska

Robert M Krimmel
U.S. Geological Survey
Ice and Climate Project
Tacoma, Washington

Abstract

The instantaneous surface geometry of the lower reach of Columbia Glacier, Alaska, was recorded with stereo vertical photography 97 times between mid-1976 and early 1997. The shortest time interval between flight dates was 11 days, the longest 330 days. Each flight was made from approximately 7000 meters altitude with a calibrated camera. The resulting stereo models were referenced to geodetic datum using aerotriangulation. These photographs form a temporally irregular basic data set from which changing conditions can be measured. Superposition was used to follow surface features from flight date to the successive flight date. Photogrammetry was used to determine the locations of these traveling features, the terminus position, and surface ice levels.

The long term trend of terminus position (Figure 1) was stability prior to 1980, and accelerating retreat in the early 1980’s, followed by steady retreat of about 0.7 km/year until the mid-1990’s, with reduced retreat rate in the late 1990’s. Within the long-term trend was a seasonal position change in which the terminus tended to be extended in late spring and retracted in late fall (Figure 2). Annually averaged surface ice speed near the terminus was 3-5 m/day prior to the beginning of the retreat, and has since accelerated to about 20 m/day (Figure 3). During any particular flight epoch, speed is greatest at the terminus and decreases up glacier. Speed varies nearly synchronously over the lower reach during the year, with maximum speed in early spring and minimum speed in early fall (Figure 4). Surface altitude has decreased about 20 m/year near the retreating terminus.

Several aspects of the Columbia Glacier regime are undisputed: the glacier is mostly grounded, though some short-lived local flotation may occur; movement is basically by plug flow in the lower reach, with extreme basal shear; and the overall glacier system is losing mass by calving far faster than it is gaining mass by accumulation. The processes driving the system
Figure 1. The changing length of Columbia Glacier. Each dot represents a flight date. The dashed line is a quadratic representation of the smoothed retreat and when removed leaves the seasonal length change shown in Figure 2.

Figure 2. The seasonal length deviation of Columbia Glacier. The numerals indicate the year.
Figure 3. Annual speed of Columbia Glacier in 1978 and 1993.

Figure 4. The seasonal speed deviation of Columbia Glacier. The numerals indicate the year.
are disputed: for instance, what is cause and what is effect between the relations of speed, terminus position, surface altitude, and runoff?, and what is the mechanism causing calving? These questions are relevant to the economic consequences of icebergs in local shipping lanes, and to broader scientific questions relating to the suspected rapid demise of the Laurentide ice sheet, general ice flow and calving processes, and the possible breakup of the West Antarctic ice sheet.

The photographs are in the public domain and reproductions may be purchased from the Geodata Center, Geophysical Institute, University of Alaska, Fairbanks, 99775. Columbia Glacier vertical photographs are a subset of a geographically broader image collection, which includes aerial photographs of western North America glaciers taken by the U.S. Geological Survey, other Federal agencies, universities, contractors, and private concerns beginning in 1960. An index of the photography is on the internet at ‘ftp://rose.gi.alaska.edu/Public_Folder/’ Within the index, each line identifies one photograph with a unique roll-frame number, date, altitude if the photograph is vertical, and glacier name. Of the more than 60,000 photographs in the collection, about 4,000 are of Columbia Glacier.

Photogrammetric results are available in three data reports (Fountain, 1982; Krimmel, 1987; Krimmel, 1992), and a detailed description of the photogrammetric methodology is given by Meier and others (1985).

References


The iceberg discharge process: observations and inferences drawn from the study of Columbia Glacier

Mark F Meier
Institute of Arctic and Alpine Research and Department of Geological Sciences
University of Colorado
Boulder CO 80309-0450

Why Columbia Glacier?

Iceberg discharge is the main source of land ice loss to the ocean in Antarctica as well as a significant contributor to ice loss in Greenland, the Arctic, and the sub-Antarctic islands; most of these glaciers are cold and have floating termini. Why then concentrate on Columbia Glacier, a temperate, grounded tidewater glacier in Alaska, in order to deduce the processes involved in iceberg calving?

Qualitative and episodic observations of calving abound on glaciers in many parts of the world, but we do not yet have a quantitative understanding of the calving process, one capable of making predictions, for instance, on the change of calving discharge to be expected with global warming. Obtaining such an understanding requires a quantitative data set which can be used to formulate and test physical models. Such a data set exists only at Columbia Glacier. As of this writing, we have obtained 103 photogrammetric overflights of the lower reach of this glacier, each of which offers the potential to obtain hundreds of point locations on the glacier which can be followed through time; thus the velocity, strain-rate, and thickness-change fields are very well documented in time and space (Krimmel, this volume). Radio-echo sounding and fathometry have defined the bed geometry. Boreholes to the bed have been used to measure fluctuations of basal water pressure and storage and characteristics of the bed. Surface ablation and mass balance data also exist. In the period of record (1977-1997) the glacier has changed its geometry, flow, and calving rate by large amounts, yielding a unique opportunity to derive and test models with a wide variation of boundary conditions.

By way of contrast, data on flow, geometry, time changes, basal conditions, and calving rates are very sparse for cold, floating glaciers, with some exceptions; the Jakobshavn Glacier in Greenland has been studied intensively for many years. However, none of these data sets matches the richness and completeness of that available at Columbia. In addition, calving from cold, floating glaciers is very episodic, making it difficult to test calving physics on a reasonable time scale. An added
complication with cold glaciers is that thermophysical modeling must be combined with glacier
dynamics and fracture mechanics models, making data requirements more stringent and solutions
more difficult. So few real data exist on the dynamics of cold glacier calving that this subject has
enormous potential for unbridled speculation but less possibility for real quantitative understanding.
It seems logical to try first to understand calving where there are ample data, and then to extend the
results to other more complex situations.

The observational evidence

The following points seem to be established:

1. Tidewater glaciers in Alaska and some other parts of the world exhibit an asynchronous and
   possibly periodic behavior, characterized by slow advance or retreat when the terminus is in
   shallow water, but very rapid calving and retreat accompanied by rapid flow and reservoir
drawdown when the terminus is in deep water (Post, 1975; Mann, 1986; Meier and Post, 1987;
   Clapperton and others, 1989; Warren, 1992; Clague and Evans, 1993; Viens, 1994). Columbia
   Glacier is no exception to this pattern

2. Iceberg calving rates (speeds and discharges) vary seasonally, with a low calving rate in
   February-March and a high rate in late September (Krimmel, pers. comm. 1996; this workshop).

3. Iceberg calving speed shows an approximate linear relation to water depth at the front, when one
   removes the seasonal fluctuation by using annual averages. The fraction of explained variance
   for 52 calving glaciers in Alaska, including both near-steady state and drastically retreating
   glaciers, is 0.81 for the water-depth relation (Brown and others, 1982). For Columbia Glacier
   (1980-93) this fraction is 0.51 (Meier, 1994); using newer and more accurate bathymetry data
   and adding values for 1994-96 changes this value to 0.72. Calving speed can also be related to
   ice thickness at the front.

4. Ice cliff heights, regardless of the state of stability or instability of the terminus or thickness of
   the glacier, have not been observed to be higher than 100 m above the water and are mostly in
   the range of 40-80 m (Brown and others, 1982).

5. Rapid seasonal calving is followed by rapid extension rate toward the calving face; this has been
   shown to exist both in width-averaged time-distance plots of calving retreat and velocity, and in
   two-dimensional plots of strain-rate changes around calving embayments (Meier and others,
   1985).
6. Waves of several dynamic perturbations, evidently triggered at the terminus, can be followed as they propagate and diffuse upglacier. These include the seasonal increases in extension rate (Meier and others, 1985), velocity changes due to tidal changes in the back pressure against the terminus (Krimmel and Vaughn, 1987; Walters and Dunlap, 1987), as well as waves of thinning and velocity increase over multi-year time scales (Krimmel, pers. comm. 1996). We have found no evidence for waves originating upglacier that propagate down to the terminus.

7. The flow of Columbia Glacier, which is predominately due to basal sliding (Kamb and others, 1994), varies with time depending on changes in water pressure and water storage at the bed, with the relative contribution of each to velocity fluctuation depending on the time scale (hourly to seasonal) (Fahnestock, 1991; Meier and others, 1994; Kamb and others, 1994). Evidence exists for a thin and possibly discontinuous layer of deforming till at the bed (Humphrey and others, 1993).

8. The effective pressure at the bed (ice pressure minus water pressure) measured in boreholes in 1987 mostly ranged from 1.0 ± 0.6 bars at a site 12 km above the 1987 terminus, to 0.9 ± 0.9 bars at 5 km above the terminus (Kamb and others, 1994). These numbers correspond to heights above buoyancy of only about 10 m well above the terminus. The effective basal pressure at the retreating terminus, computed from glacier geometry and ice and local water densities with due allowance for crevasse densities, dropped from about 6 bars in 1977 to less than 1 bar in 1992 - 1996.

9. Velocity of the lower reach shows a seasonal variation with a maximum in spring (May) and a minimum in fall (late September), except for effects at the terminus apparently related to seasonal calving retreat (Meier and others, 1985; Krimmel and Vaughn, 1987); this variation appears to occur simultaneously over the reach. In addition, there has been a significant acceleration (up to 5-fold) at the terminus and at fixed points upglacier from 1977 to 1996, and speeds of more than 7 km/yr have been measured in the last few years (Krimmel, 1992; Krimmel, pers. comm. 1996). The pattern of long-term acceleration clearly propagates upglacier. Longitudinal extending strain rates also show significant increases over time. At the same time, the glacier has thinned, and the wave of thinning also propagates upglacier (Krimmel, 1992; Krimmel, pers. comm. 1996).
Some conclusions which can be drawn from these observations

In my opinion, we do not yet have a physical understanding of iceberg calving that will stand up to rigorous scrutiny. A number of suggestions and models have been promoted, but none of these has been verified. Any conclusions, models, or speculations on the iceberg calving process at Columbia Glacier should be based on, and consistent with, the observations listed above. Below I list some conclusions based on the observational data which, I believe, are quantitatively defensible and which might be useful in developing a realistic physical model of iceberg calving.

1. An empirical relation between iceberg calving speed (averaged over a seasonal cycle) and water depth appears to be fairly robust, and therefore a physical model should produce a relation that is not inconsistent with this. The height above buoyancy model, proposed originally by Sikonia (1982), seems useful in explaining aspects of calving, even though it cannot apply when the effective pressure at the bed approaches zero. It does suggest an increase in calving with ice thinning but the cause/effect relationship is undefined.

2. The role of water (pressure as well as amount in temporary storage) in and within the glacier needs to be treated carefully. Basal water pressure is clearly highest in the spring (before the basal hydraulic system has time to respond to increased input), and is generally thought to be the cause of the observed velocity maximum in spring. The role of water in causing crevasse extension is less obvious, but it should be noted that water is almost always available on the lower reach of this glacier. In fact, water standing in crevasses as high as the surface, in local areas, is frequently observed in late winter and spring, and rarely if ever observed it in late summer or fall. Thus if water pressure in crevasses aids calving, one would expect a different seasonal calving pattern. Inferences about the amount of water stored at any instant need to be made carefully, because computing storage change from a realistic input/output rate model is difficult and not well constrained by data (Fahnestock, 1991), and there are strong dependencies on time scales (Kamb and others, 1994).

3. Iceberg calving is largely a problem in fracture mechanics coupled to ice dynamics, and thus analysis of longitudinal extending stresses is of fundamental importance. Ice cliff heights do not become excessive because extending strain rates tend to increase toward the calving front when calving retreat occurs, this produces local thinning, as is clearly indicated by the observations at Columbia Glacier. Higher ice cliffs should cause higher non-hydrostatic stresses (both bending and extensile) which should lead to the higher extending strain rates.
4. There is no evidence that the position of the terminus is controlled by anything other than the balance between the flow of ice to the terminus and the removal of ice by iceberg discharge. These two variables, of course, are determined by the calving boundary condition, the geometry of channel, and the ice flow (mainly basal sliding) process. We see no evidence that changes in these are propagated in from somewhere upglacier, and every indication is that changes at the terminus propagate upglacier. This lends support to the concept that calving is the local driving process, and the ice dynamics changes in response to this driving.

5. Two feedback loops seem to be important. First, increased calving (for whatever reason) leads to increased terminus extension rate (perhaps due to increased unbalanced stresses due to a higher ice cliff), which leads to thinning of the glacier near the terminus, which leads to a decrease in the effective pressure on the bed, which leads to increased basal sliding as explained by established theory (e.g., Bindschadler, 1983), which leads to increased extension, thinning, and sliding, and propagation upglacier. If the calving rate is related to the extension rate, increased calving will strengthen the positive feedback. Second, the potential rate of retreat will be diminished due to the increased basal sliding due to thinning, a negative feedback. The space-time pattern of extension, thinning, and flow for Columbia Glacier appears to be entirely consistent with these feedback processes.

6. The above suggested chain of processes does not invoke any unspecified forcing of changes in terminus position or ice velocity which in turn cause changes in calving by some (also unspecified) secondary process (Van der Veen, 1996). Rather, all steps in the chain described here should be verifiable either through existing glaciological knowledge or further studies.

7. Extension of a calving model to include floating, cold glacier tongues and shelves, which will require coupling with a thermophysics model, seems premature until we understand the basic physics more completely.

8. Finally, it should be noted that the observations and inferences presented here are derived from study of a tidewater glacier in the steady (1977-1980) and in the rapidly-retreating (1981-1996) modes. Additional study needs to be given to the processes involved in the third mode (slow advance) of the tidewater glacier cycle, which apparently involves the motion by erosion and deposition of a morainal bank and subglacial/fjord sediment transport.
References


Deep-water calving at Le Conte Glacier, Southeast Alaska

Roman J Motyka
University of Alaska

Abstract

The mechanics of calving remain poorly understood, largely because direct observations are often difficult. This is particularly true for deep-water glaciers where in terms of volume, submarine calving can be the dominant iceberg-producing process. At Le Conte Glacier in southeast Alaska, deep water and high ice flux combine to produce extremely active calving while the narrow fjord (~1.5 km) and vantage points on adjacent bedrock ridges allow excellent documentation of deep-water calving events. Le Conte Glacier is currently undergoing a calving retreat that began in late 1994 (cf Echelmeyer and Motyka, this volume).

Ice cliffs at Le Conte are generally about 40 to 60 m above water, although seracs as high as 80 m occasionally survive to the calving front. The ice cliffs are commonly vertical or near vertical but at various times they were observed to lean outward. Soundings made after the glacier retreated show that terminus water depths ranged from 250 to 270 m at the time of our calving observations. Ice thickness therefore ranged from 290 to 330 m while the height-above-buoyancy was on the order of 30 m. The terminus region has an average slope of 10°. Surveyed near-terminus ice velocities range from 15 to 23 m/d. The glacier is heavily crevassed to at least the equilibrium line, indicating strong extensional flow. Intersecting crevasse fields produce columns of seracs in the terminus region separated by broad troughs. Daily time-lapse photographs and surveys show that the terminus undergoes a seasonal fluctuation of 300 to 400 m with maximum position attained by mid-May. We observed calving at LeConte Glacier from 300-m and 100-m high bedrock ridges overlooking the terminus and documented nearly 400 events during 25 hours of observations over a 4-day period in mid-May 1996, and nearly 900 events during 65 hours of observations over a 5-day period in mid-May 1997. The vast majority of events were small to moderate size subaerial spalls, collapses, and avalanches. However, of particular interest were periodic episodes of massive calving. These occurred at an average rate of about twice daily and typically involved broad subaerial and submarine sections of the calving front. These events were by far the most voluminous, producing the largest and most numerous icebergs.
Figure 1. Examples of submarine calving at Le Conte Glacier, Alaska. Panel a: Collapse of subaerial ice cliffs preceded emergence of the tabular block of ice along the face; the top of the submarine iceberg reached 40 m above water before collapsing. Panel b: Buoyant forces thrust the oblong submarine iceberg upwards to 80 m above water level after a subaerial serac collapsed (see also p. 21 of this report).
Figure 2. Massive deep-blue submarine icebergs emerged 250 to 300 m from the face following subaerial collapse of a broad section of the glacier terminus. The top of the iceberg in the foreground was driven 35 m above water.

An episode of massive calving usually began with subaerial collapse of seracs and ice cliffs, either by toppling forward or by sliding along fractures or slip planes (Figure 1a). Sections of submarine ice then apparently detached and were propelled upward by buoyancy, their tops emerging as much as 40 to 80 m above water (Figure 1a and 1b). These submarine icebergs were commonly either tabular with long axis parallel to the calving front (Figure 1a) or oblong in shape (Figure 1b), and usually rose vertically parallel to the calving front. Their emergent surfaces were highly irregular indicating complex fracturing. The icebergs were light blue to aqua in color suggesting high bubble content and/or micro-fracturing of the ice surface. Additional sections of submarine ice then emerged and were also propelled upwards with great force. These icebergs were usually very massive with clean fracture surfaces, were clear and deep blue in color, and emerged as much as 200 to 300 m from the subaerial calving face (Figure 2). The time for the entire preceding
sequence to transpire varied considerably. In some cases, all three phases occurred within a span of 2 to 3 minutes. In other cases, delays of several minutes to several hours occurred between phases.

Isolated submarine icebergs, locally known as “shooters”, were occasionally observed to emerge from 50 to as far as 500 m from the subaerial face. Vigorous upwelling usually preceded emergence and continued for several seconds to minutes after the ice blocks surfaced. These blocks appear to have risen vertically as no horizontal component of motion was observed. Although such events have been documented at other calving glaciers their genesis remains a mystery.

The large-scale submarine events are likely related to buoyancy and extensional flow. At the measured water depths and ice cliff heights, the face was at ~ 90% of flotation. Subaerial collapse of a ~50-m-high ice cliff could therefore have helped trigger release of submarine icebergs by rendering the underlying submarine section super-buoyant. Judging from ice color and the sequence in which the icebergs emerged, the light blue to aqua color submarine icebergs appear to have been derived from immediately below water level while the dark blue, clear ice likely come from deeper, basal parts of the calving front. Several of the massive calving events produced deep blue ice bergs of considerable size that emerged at distances of up to 250 m from the terminus. These events plus isolated emergence of submarine icebergs distal from the subaerial face indicate that at times the calving front can support a substantial submarine platform or “toe” of ice. Given the strong extensional flow and buoyant forces at work it is surprising that such a toe can remain intact for any length of time.
Taku Glacier, Alaska: advance and growth of a tidewater glacier

Roman J Motyka
University of Alaska

Abstract

In 1890, Taku Glacier calved icebergs into a 100 m deep tidal basin. Since then the glacier has advanced 7.3 km while all neighboring glaciers have been retreating. Rapid infilling of the tidal basin by glacial and fluvial sediments, and the formation of a shoal moraine progressively reduced and eventually stopped glacier calving, spurring glacier advance. The rate of advance peaked at 160 m/yr during the 1930’s then steadily declined to 15 m/yr after 1960. The total growth of the ablation zone between 1948 and 1989 as determined by photogrammetry was \(-52 \times 10^8\) m\(^3\) water equivalent, with increases in surface elevation ranging from 200 m near the terminus to 40 m at the equilibrium line. In addition, comparison of ice depth soundings made in 1989 and 1990 to early and mid-century pre-advance bathymetric surveys, document compaction and erosion of overridden glaciomarine sediments at rates of 2.0 to 3.4 m/yr along the centerline. In comparison, ice surface elevations in the lower ablation zone have risen 2 to 5 m/yr while ablation averaged 8-10 m/yr. Downward expansion of the glacier by erosion and compaction of soft sediments is estimated at \(-12 \times 10^8\) m\(^3\) water equivalent for the period between 1948 and 1989. The rapid advance during the 1930s may have been facilitated by the compressive dewatering of glaciomarine sediments. The terminus has been stationary since 1988 despite glacier thickening, and erosion may now be retarding advance by entrenching the glacier into its glaciomarine deposits and terminal moraine. The remobilized sediments have gradually filled the head of Taku Inlet. Similar processes likely occurred for glaciers overriding soft sediments in the past, for example expansion of ice sheets into marine basins during the Pleistocene, and are probably happening during the current advance of glaciers into the west arm of Glacier Bay. The data from Taku Glacier indicate care must be taken in estimating glacier mass balance solely from surface elevations in these cases, and that modeling of glacier growth and advance into marine basins needs to take into account rapid advance over soft sediments and their subsequent erosion.
Additional information can be found in the following references:

