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Twentieth century thinning of Mendenhall Glacier, Alaska, and its relationship to climate, lake calving, and glacier run-off

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Abstract

Mendenhall Glacier is a dynamic maritime glacier in southeast Alaska that is undergoing substantial recession and thinning. The terminus has retreated 3 km during the 20th century and the lower part of the glacier has thinned 200 m or more since 1909. Glacier-wide volume loss between 1948 and 2000 is estimated at 5.5 km^3. Wastage has been the strongest in the glacier's lower reaches, but the glacier has also thinned at higher elevations. The shrinkage of Mendenhall Glacier appears to be due primarily to surface melting and secondarily to lake calving. The change in the average rate of thinning on the lower glacier, <1 m a^{-1} between 1948 and 1982 and >2 m a^{-1} since 1982, agrees qualitatively with observed warming trends in the region. Mean annual temperatures in Juneau decreased slightly from 1947 to 1976; they then began to increase, leading to an overall warming of ~1.6 °C since 1943. Lake calving losses have periodically been a small but significant fraction of glacier ablation. The portion of the terminus that ends in the lake is becoming increasingly vulnerable to calving because of a deep pro-glacial lake basin. If current climatic trends persist, the glacier will continue to shrink and the terminus will recede onto land at a position about 500 m inland within one to two decades. The glacier and the meltwaters that flow from it are integral components of the Mendenhall Valley hydrologic system. Approximately 13% of the recent average annual discharge of the Mendenhall River is attributable to glacier shrinkage. Glacier melt contributes 50% of the total river discharge in summer.

Keywords: glacier retreat; climate change; glacier calving; glacier run-off; Alaska

1. Introduction

Temperate glaciers have been the focus of an increasing number of studies because of their sensitivity to climate (Hodge et al., 1998; Hooker and Fitzharris, 1999) and because of their predicted role in sea-level rise (e.g., Meier, 1993; IPCC, 1995). However, the number of glaciers world-wide that have long-term mass balance records is quite small (Dyurgerov and Meier, 1997). This limits the accuracy of sea-level and climate-change assessments based on the recent glacial record. Despite the fact that Alaska has the largest number and volume of non-polar glaciers in the world, only three glaciers in Alaska have mass-balance time series spanning 30 years or more: Wolverine Glacier, a maritime glacier on the Kenai Peninsula, Gulkana Glacier, located in the

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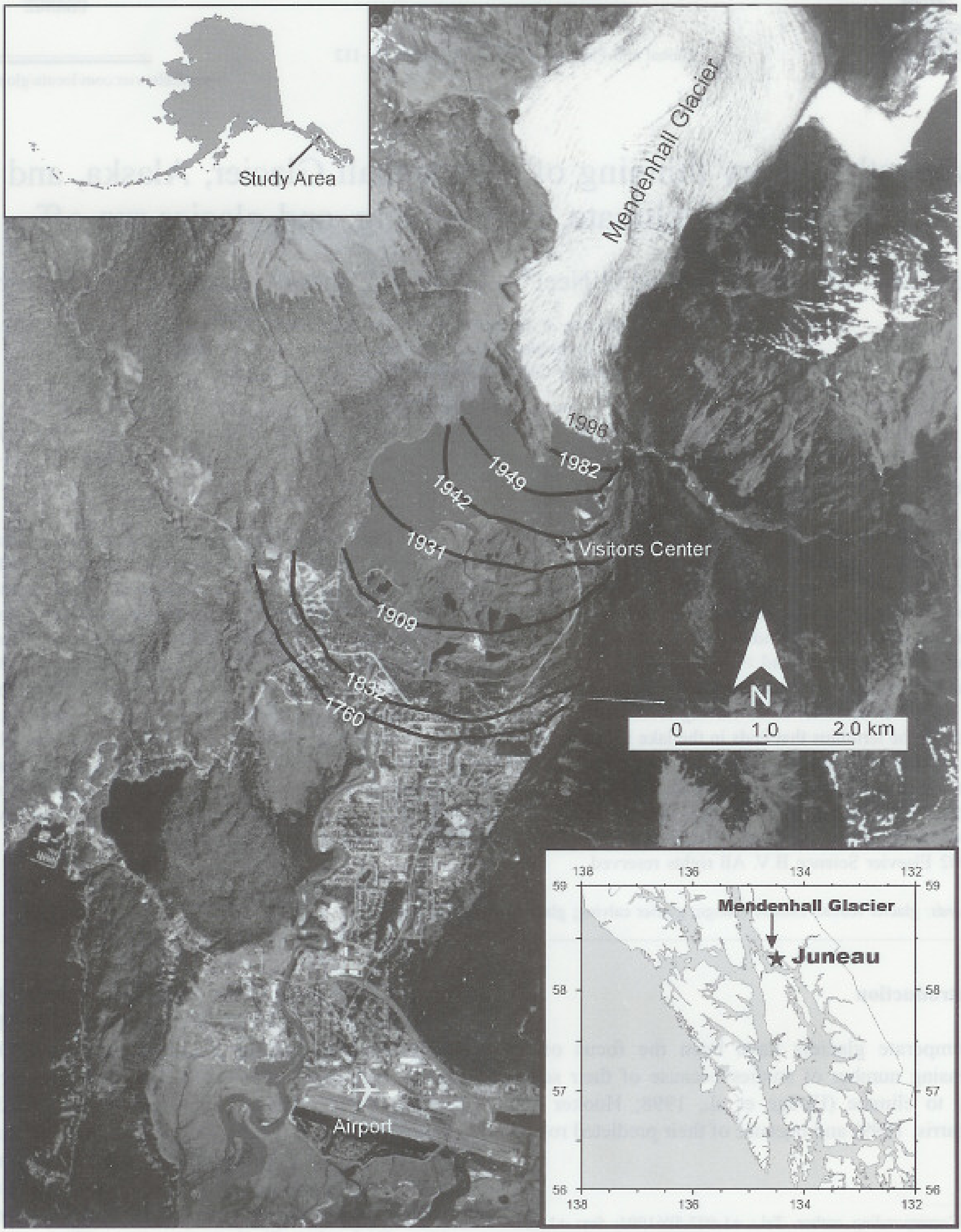
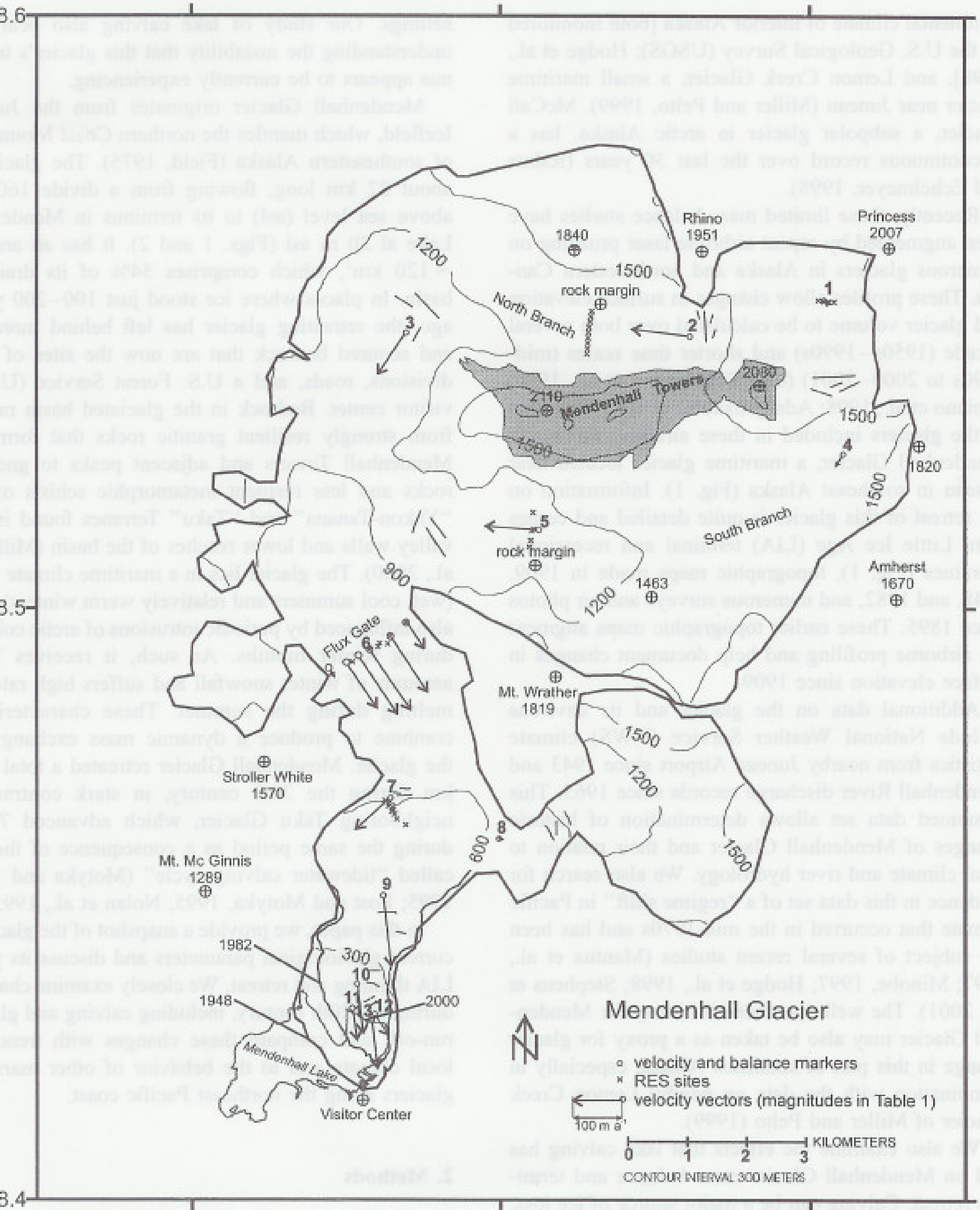


Fig. 1. Location map and aerial photo of Mendenhall Glacier. Several post-LIA terminus positions are also shown. (Photo: NASA, 7-25-1996).



Mendenhall Glacier. Contours and base map are from USGS maps and DEM based on 1948 photography. Terminus boundaries for 1948, and 2000 are shown. Mass balance and velocity markers are indicated by circles, RES sites by crosses. Dark lines locate major icefalls. See for enlargement of terminus area and RES sites there.

continental climate of interior Alaska [both monitored by the U.S. Geological Survey (USGS); Hodge et al., 1998], and Lemon Creek Glacier, a small maritime glacier near Juneau (Miller and Pelto, 1999). McCall Glacier, a subpolar glacier in arctic Alaska, has a discontinuous record over the last 30 years (Rabus and Echelmeyer, 1998).

Recently, these limited mass balance studies have been augmented by repeat airborne laser profiling on numerous glaciers in Alaska and northwestern Canada. These profiles allow changes in surface elevation and glacier volume to be calculated over both several decade (1950s–1990s) and shorter time scales (mid-1990s to 2000–2001) (e.g., Echelmeyer et al., 1996; Sapiano et al., 1998; Adalgeirsdottir et al., 1998). One of the glaciers included in these airborne surveys is Mendenhall Glacier, a maritime glacier located near Juneau in southeast Alaska (Fig. 1). Information on the retreat of this glacier is quite detailed and comes from Little Ice Age (LIA) terminal and recessional moraines (Fig. 1), topographic maps made in 1909, 1948, and 1982, and numerous surveys and air photos since 1895. These earlier topographic maps augment the airborne profiling and help document changes in surface elevation since 1909.

Additional data on the glacier and its environs include National Weather Service (NWS) climate statistics from nearby Juneau Airport since 1943 and Mendenhall River discharge records since 1965. This combined data set allows determination of historic changes of Mendenhall Glacier and their relation to local climate and river hydrology. We also search for evidence in this data set of a "regime shift" in Pacific climate that occurred in the mid-1970s and has been the subject of several recent studies (Mantua et al., 1997; Minobe, 1997; Hodge et al., 1998; Stephens et al., 2001). The well-documented history for Mendenhall Glacier may also be taken as a proxy for glacier change in this part of southeast Alaska, especially in combination with the data on nearby Lemon Creek Glacier of Miller and Pelto (1999).

We also examine the effects that lake calving has had on Mendenhall Glacier mass balance and terminus retreat. Calving can be a major source of ice loss, for example, Columbia Glacier, but there are very little data in the literature documenting the contribution of calving losses to overall glacier mass balance and its effect on glacier retreat, especially in lacustrine settings. Our study of lake calving also bears on understanding the instability that this glacier's terminus appears to be currently experiencing.

Mendenhall Glacier originates from the Juneau Icefield, which mantles the northern Coast Mountains of southeastern Alaska (Field, 1975). The glacier is about 22 km long, flowing from a divide 1600 m above sea level (asl) to its terminus in Mendenhall Lake at 20 m asl (Figs. 1 and 2). It has an area of ~120 km^2, which comprises 54% of its drainage basin. In places where ice stood just 100–200 years ago, the retreating glacier has left behind moraines and scoured bedrock that are now the sites of subdivisions, roads, and a U.S. Forest Service (USFS) visitor center. Bedrock in the glaciated basin ranges from strongly resilient granitic rocks that form the Mendenhall Towers and adjacent peaks to gneissic rocks and less resilient metamorphic schists of the "Yukon-Tanana" and "Taku" Terranes found in the valley walls and lower reaches of the basin (Miller et al., 2000). The glacier lies in a maritime climate zone (wet, cool summers and relatively warm winters); it is also influenced by periodic intrusions of arctic cold air during winter months. As such, it receives large amounts of winter snowfall and suffers high rates of melting during the summer. These characteristics combine to produce a dynamic mass exchange on the glacier. Mendenhall Glacier retreated a total of 3 km during the 20th century, in stark contrast to neighboring Taku Glacier, which advanced 7 km during the same period as a consequence of the so-called "tidewater calving cycle" (Motyka and Post, 1995; Post and Motyka, 1995; Nolan et al., 1995).

In this paper, we provide a snapshot of the glacier's current glaciological parameters and discuss its post-LIA thinning and retreat. We closely examine changes during the 20th century, including calving and glacier run-off, and compare these changes with trends in local climate and to the behavior of other maritime glaciers along the northeast Pacific coast.

2. Methods

2.1. Glacier mass balance

We measured the mass balance of Mendenhall Glacier over the period August 1997–September

2000 following the glaciological method. Annual mass balances are reported in water equivalent (w.e.); calving losses at the terminus are included in the annual ablation. Thirteen to 14 mass balance markers were drilled into the glacier surface at elevation intervals of approximately 300 m. Because of the large melt rates at lower elevations (>10 m a^{-1} w.e.), the lowermost holes were redrilled midway through the summer. Ablation wires were used in these lower reaches to reduce problems with excessive pole heights and to combat heat conduction to surrounding ice. Summer ablation at higher elevations was monitored using coupled lengths of metal conduit.

The extreme amount of snowfall at high elevations on Mendenhall Glacier (>10 m w.e. in places) makes determining annual accumulation a difficult problem. We dug snow pits near the Taku–Mendenhall divide (Fig. 1) in late summer of each year (1997–2000), but the previous summer's surface was reached only in 1998. Snow densities were measured in the pits. In addition, we used exposures in the walls of crevasses and snow probes to determine snow depths. Equilibrium line altitudes (ELA) were measured each autumn using differentially corrected GPS (DGPS) to an accuracy of a few meters.

2.2. Ice velocity

Surface ice velocity was determined by measuring the positions of the mass balance markers and at the flux gate markers with DGPS, typically during the spring and fall each year. These DGPS surveys have a positional accuracy of ±0.3 m.

2.3. Radio echo sounding

A standard radio echo sounder (RES, Watts and Wright, 1981) was used to determine ice thickness and bed elevations at several points on the glacier, including across two transverse sections. This 5-MHz ice radar is capable of sounding approximately 600 m in temperate ice with an accuracy of about 10 m. RES surface positions were determined using DGPS.

2.4. Lake bathymetry

Lake bathymetry, required for discussions of sedimentation and calving flux, was obtained using an acoustic depth sounder (transducer: 600 W, 200 kHz and either 15° or 5° beamwidths) from a small inflatable boat. The accuracy of the soundings is estimated to be about 2%, and GPS sounder positions were accurate to about 1 m. We compiled about 11,500 soundings and combined these with approximately 1100 collected by J. Fleisher (pers. comm., 2000), to produce a detailed bathymetric chart of the lake.

2.5. Terminus positions and glacier thinning

Post-LIA terminus positions on recessional moraines were obtained from Lawrence (1950), Miller (1975), and Lacher (1999), as well as from aerial photos. The earliest historic record of terminus position comes from an International Boundary Commission map made in 1895. A contour map of the lower glacier and surrounding region at a scale of 1:62,250 was produced by Knopf (1912) in 1909 and provides the earliest topographic record of the glacier, but only up to an elevation of 800 m. We estimate the accuracy of the glacier topography on this map to be about ±30 m. The terminus position was resurveyed several times between 1931 and 1949 by personnel from the U.S. Coast and Geodetic Survey and by the USFS (source: USFS Juneau Ranger District archives).

USGS 1:63,260 scale maps, Juneau B2 and C2, cover the entire glacier and were made from 1948 aerial photography. We used the USGS DEM derived from these maps for constructing profiles and calculating volume changes. Elevation accuracy of the maps, and of the derived DEM, is estimated to be ±15 m below the 1948 snow line (about 1000 m asl); the accuracy at higher elevations is not as good because of the featureless snow cover. A second, larger scale USGS map based on 1982 photography (1:25,000, Juneau B2 NW) has an accuracy of about ±10 m, but that map extends only up to the 800 m level on the glacier. Post-1982 terminus positions were obtained from aerial photos and from direct measurement. Since 1997, terminus position has been seasonally determined using DGPS.

Two methods were employed to examine glacier thinning. In the first, centerline profiles of the glacier surface were constructed from the 1909 and 1982 maps and from the 1948 map/DEM, and then differenced. The second method uses an airborne laser altimetry system developed by Echelmeyer et al.

(1996). Elevations are determined every 1.5 m along the profiles to an accuracy of 0.1–0.3 m. Airborne profiles of the glacier surface were obtained in August 1995, 1999 and 2000. The aircraft's navigation system allows the pilot to fly the original flight paths in succeeding years with reasonable repeatability, and elevation changes can be determined at several thousand crossing points along the various flight paths. We calculated changes in elevation between the 1948 map and the 1995 profiles, the 1995 and 1999 profiles, and those in 1999 and 2000. Because a 1982 DEM is not available, elevation differences between 1948 and 1982 centerline profiles were subtracted from the 1948 to 1995 changes to determine lower glacier thinning between 1982 and 1995.

For the years that we have complete elevation change coverage at all elevations, we computed volume changes and glacier-wide average changes in thickness following methods described in Echelmeyer et al. (1996). Long-term average mass balances (w.e.) can be obtained from these volume changes by assuming that Sorge's Law (Paterson, 1994) holds at all elevations. This assumption may introduce small errors in the calculated average mass balances if old firn is removed by melting near the equilibrium line, exposing bare ice.

3. Results

3.1. Current mass balance

The results of summer and annual 1998 and 2000 mass balance measurements for Mendenhall Glacier are shown in Fig. 3. "Summer balance" refers to that determined from late-May to late-August, while the balance year was taken to be from 1 September to 31 August. The 2000 ELA was estimated to be ~ 820 m (±50 m) based on ground observations of the firn line at the end of August and from airborne profiling; this is considerably lower than the 1990s' average ELA, which we estimate from airborne observations to be about 1100 m. The ELA in 1999 and 1998 was ~1100 and ~1250 m, respectively. The 2000 balance year saw exceptionally high accumulation at higher elevations. The end-of-melt season snow pack, as measured in crevasses near the head of the glacier in early September 2000, ranged from 7.5 to 9.5 m in thickness. Using the measured snow density of 600 kg m^{-3} gives an average of 5.1 m w.e. In comparison, annual accumulation in 1998 at site 1 was only 1.6 m w.e. (Fig. 3). Only an estimate of the annual accumulation near site 1 is available for the 1999 balance year. At the end of summer 1999, we were unable to

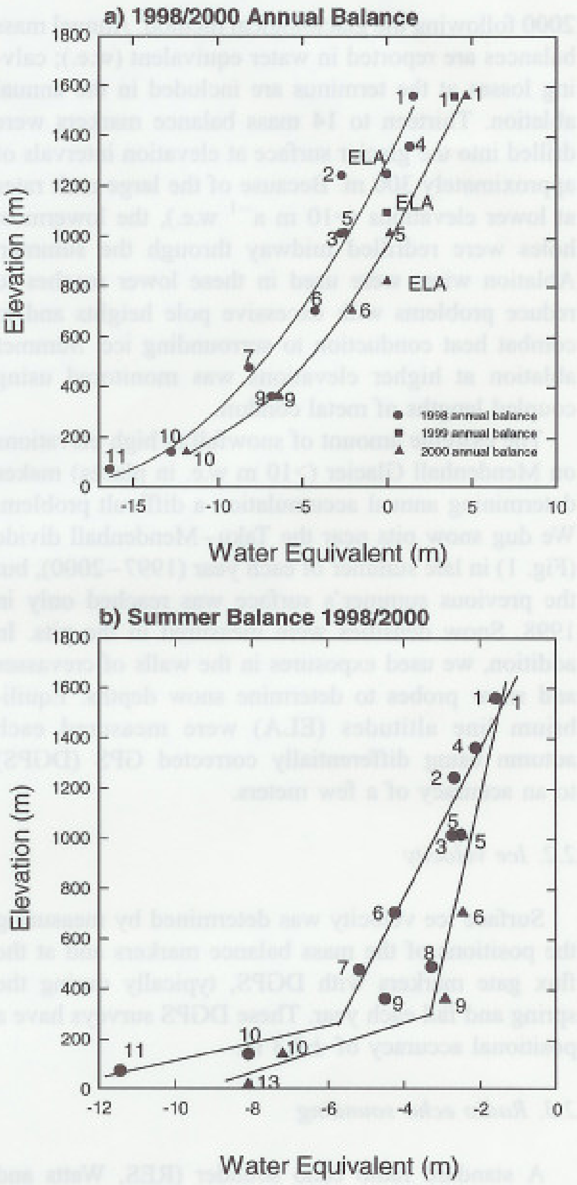


Fig. 3. Mass balance versus elevation for Mendenhall Glacier, 1998 and 2000 balance years: (a) annual balance, (b) summer balance. 1999 ELA and 1999 annual accumulation at site 1 are also shown.

detect a summer surface with a 1.3-m snow probe from the base of a 5.3-m-deep pit, so an annual accumulation of ~4.0 m w.e. is a minimum estimate. We did not measure annual ablation at any markers in 1999.

In contrast to the accumulation at high elevations, ablation at the lowest elevations was similar for both 2000 and 1998 (cf. marker 10, where ablation was ~−12 w.e. m for both years). However, ablation at all higher elevations was significantly lower in 2000 compared to 1998 (Fig. 3). These differences are reflected in the glacier-wide annual mass balance, which was +1.1 m w.e. for 2000 and −1.5 m w.e. for 1998. Note that these annual balances also include a small fraction of ice loss due to calving, as discussed below. Given the errors in these calving measurements, plus the limited number of balance markers and snowpits, we estimate the error in the glacier-wide balances to be about ±0.3 m.

3.2. Bathymetry

Mendenhall Lake did not exist before 1930; it was formed as the glacier retreated from 1931 to the present. The current area of the lake is ~3.4 km^2 and its volume is ~0.05 km^3. Results of our recent bathymetric survey are shown in Fig. 4. The southern portion of the lake is relatively shallow, generally 15

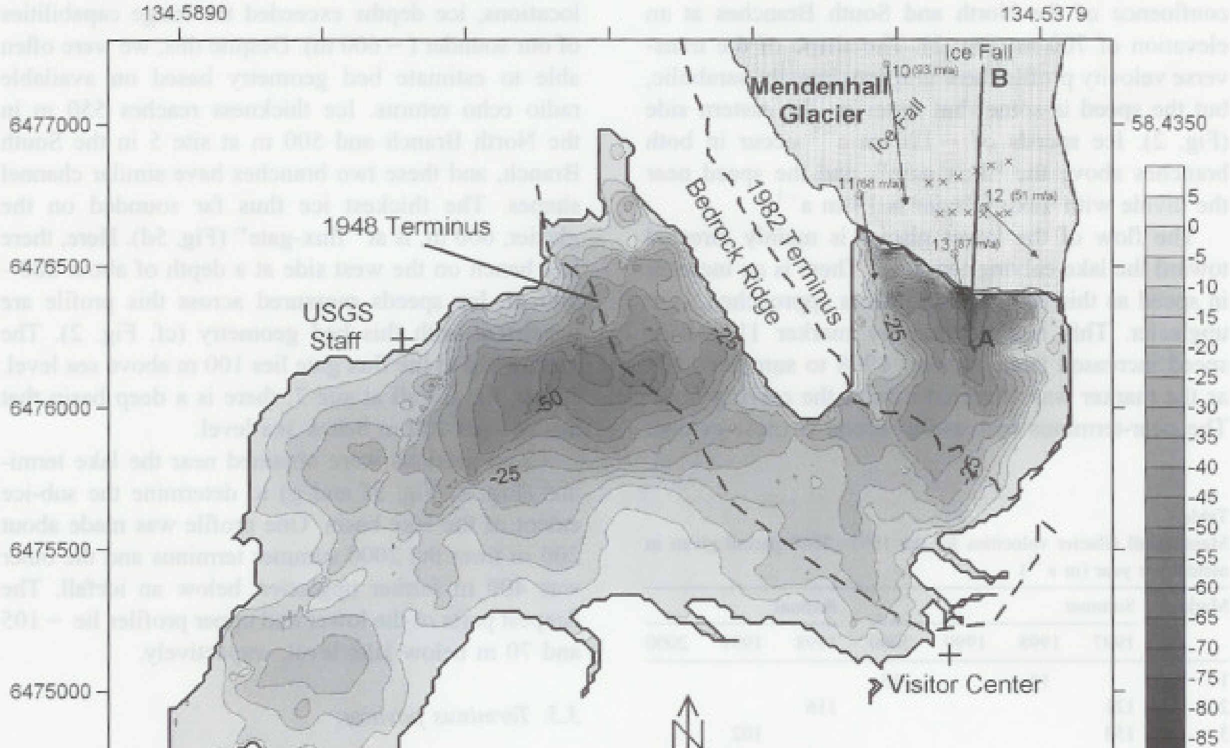


Fig. 4. Mendenhall Lake bathymetry as measured in summer 2000. UTM and geographic coordinates are NAD27. X's mark radio echo sounding positions. Velocity vectors are also shown. Early-fall terminus positions in 1948 and 1982 are shown as dashed lines. The change in position between 1982 and 2000 documents differential retreat of land and lake terminating sectors. Line AB marks the transect shown in Fig. 9.

m or less. The lake deepens to the north and has two prominent basins: one lies south of the prominent bedrock ridge and the other lies directly in front of the present terminus. Both basins are ~70 m deep with respect to average lake level, which places them well below sea level.

3.3. Ice motion

Summer and annual ice speeds are listed in Table 1 (see Figs. 2 and 4 for marker locations). Summer speeds are faster than annual speeds; this reflects increased basal motion, probably from increased meltwater to the base. The maximum speed of 160 m a^{-1} was observed at marker 6 in the "flux gate" below the confluence of the North and South Branches at an elevation of 700 m (Fig. 2). The shape of the transverse velocity profile there is approximately parabolic, but the speed is somewhat faster on the eastern side (Fig. 2). Ice speeds of ~120 m a^{-1} occur in both branches above the "flux gate", and the speed near the divide with Taku Glacier is 14 m a^{-1}.

The flow of the lower glacier is mainly directed toward the lake calving terminus. There is an increase in speed as this calving terminus is approached from upglacier. This is illustrated by marker 11, whose speed increased from summer 1998 to summer 1999 as the marker was advected toward the calving front. The near-terminus increase in speed is most evident from markers 12 to 13 (Table 1; Fig. 4), where ice speed nearly doubles over a distance of about 200 m. Marker 13 was only about 50 m from the terminus by August 2000. This strong extensional flow causes crevassing and ice thinning, making it more susceptible to calving.

Ice in "Suicide Basin" is nearly stagnant, with a speed of 2 m a^{-1} (marker 8 in Fig. 2). Moraine patterns suggest that this tributary was previously more active.

Table 1
Mendenhall Glacier velocities for the 1997–2000 period given in meters per year (m a^{-1})

Marker	Summer				Annual		
	1997	1998	1999	2000	1998	1999	2000
1		14		13			
2	121				116		
3	153					102	
4	40	36			34		
5	106	118		117	105	111	107
6	172	167		169		145	160
7		121				96	
8		2					
9	135	131					138
10		116	111	104		93	99
11		64	81			68	
12			67	59			51
13				83			87[a]

[a] Two-year average (1998–2000).

3.4. Ice thickness

Ice thickness was measured at several sites (Figs. 2 and 4), with results shown in Fig. 5. At some locations, ice depths exceeded the range capabilities of our sounder (~600 m). Despite this, we were often able to estimate bed geometry based on available radio echo returns. Ice thickness reaches 550 m in the North Branch and 500 m at site 5 in the South Branch, and these two branches have similar channel shapes. The thickest ice thus far sounded on the glacier, 600 m, is at "flux-gate" (Fig. 5d). Here, there is a bench on the west side at a depth of about 200–300 m. Ice speeds measured across this profile are consistent with this bed geometry (cf. Fig. 2). The glacier bed at the flux gate lies 100 m above sea level. Below the icefall at site 7, there is a deep basin that lies at least 100 m below sea level.

Cross sections were obtained near the lake terminus (Fig. 4; Fig. 5f and g) to determine the sub-ice extent of the lake basin. One profile was made about 200 m from the 2000 summer terminus and the other was 400 m farther upglacier, below an icefall. The deepest parts of the lower and upper profiles lie ~105 and 70 m below lake level, respectively.

3.5. Terminus position

The LIA maximum (ca. 1760) and several additional terminus positions are shown in Figs. 1 and 4. Terminus position along the centerline is plotted in Fig. 6 as a function of time since the LIA, using all available information. The length of the glacier has decreased by about 16% since the LIA, and most of this retreat occurred during the 20th century. Retreat rates peaked during mid-20th century and at the end of the century. In recent years, episodes of calving

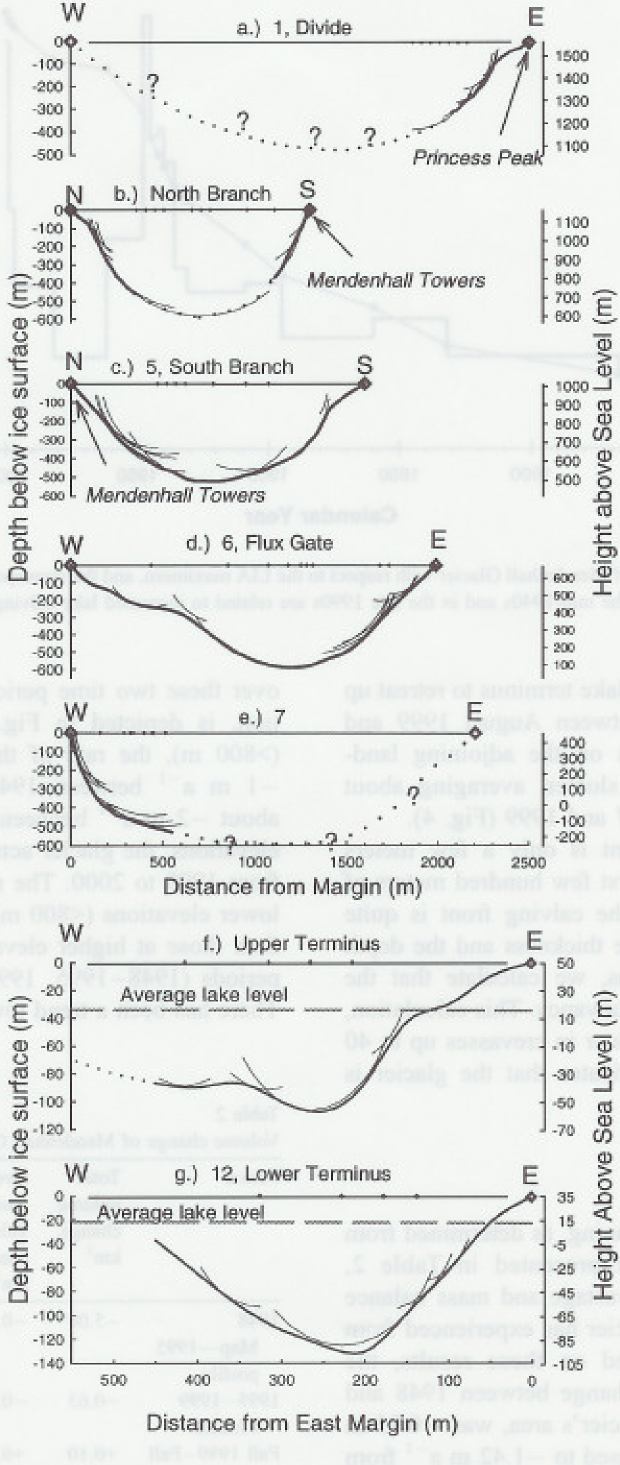


Fig. 5. RES cross sections. See Figs. 2 and 4 for locations. Arcs indicate boundaries of RES returns. Thickest ice (600 m) occurs at "flux gate".

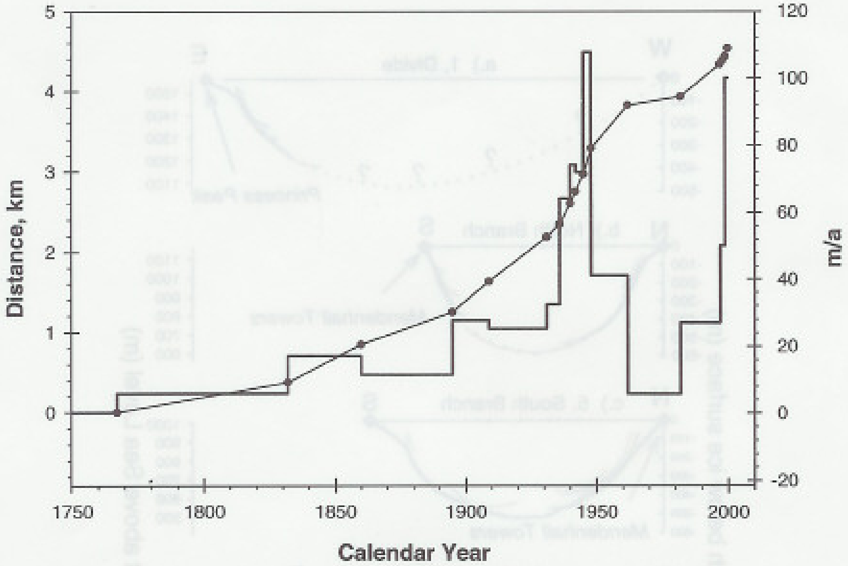


Fig. 6. Centerline terminus retreat of Mendenhall Glacier with respect to the LIA maximum, and the corresponding rate of retreat since the LIA. Peaks in retreat rate that occur in the mid-1940s and in the late 1990s are related to increased lake calving.

have caused portions of the lake terminus to retreat up to 100 m a^{-1}, such as between August 1999 and August 2000. Retreat rates on the adjoining land-based terminus are much slower, averaging about 10–15 m a^{-1} between 1997 and 1999 (Fig. 4).

The present calving front is only a few meters above lake level, and the first few hundred meters of the glacier surface above the calving front is quite flat. Given the measured ice thickness and the depth of the lake at the terminus, we calculate that the terminus is within 90% of buoyancy. This calculation, plus observations of lake water in crevasses up to 40 m back from the face, indicates that the glacier is primed for calving.

3.6. Glacier thinning

The results of glacier thinning, as determined from the airborne profiling and presented in Table 2, document the remarkable wastage and mass balance deficit that Mendenhall Glacier has experienced from 1948 to the present. Based on these results, the average rate of thickness change between 1948 and 1995, averaged over the glacier's area, was −0.95 m a^{-1}, and this thinning increased to −1.42 m a^{-1} from 1995 to 1999. The average rate of thickness change over these two time periods, as a function of elevation, is depicted in Fig. 7a. At higher elevations (>800 m), the rate of thickness change was about −1 m a^{-1} between 1948 and 1995, compared to about −2 m a^{-1} between 1995 and 1999. At higher elevations, the glacier actually increased in thickness from 1999 to 2000. The rates of thickness change at lower elevations (<800 m) were much more negative than those at higher elevations during all three time periods (1948–1995, 1995–1999, and 1999–2000). There has been a trend toward more rapid thinning in

Table 2
Volume change of Mendenhall Glacier, airborne laser profiles

Period	Total volume change, km^3	Average annual volume change, $km^3\ a^{-1}$	Area averaged ΔZ, m a^{-1}	Area averaged ΔZ, m a^{-1} w.e.
1948 Map—1995 profile	−5.00	−0.11	−0.95±0.11	−0.86
1995–1999 Profiles	−0.63	−0.16	−1.42±0.03	−1.29
Fall 1999–Fall 2000	+0.10	+0.10	+0.93±0.12	+0.85

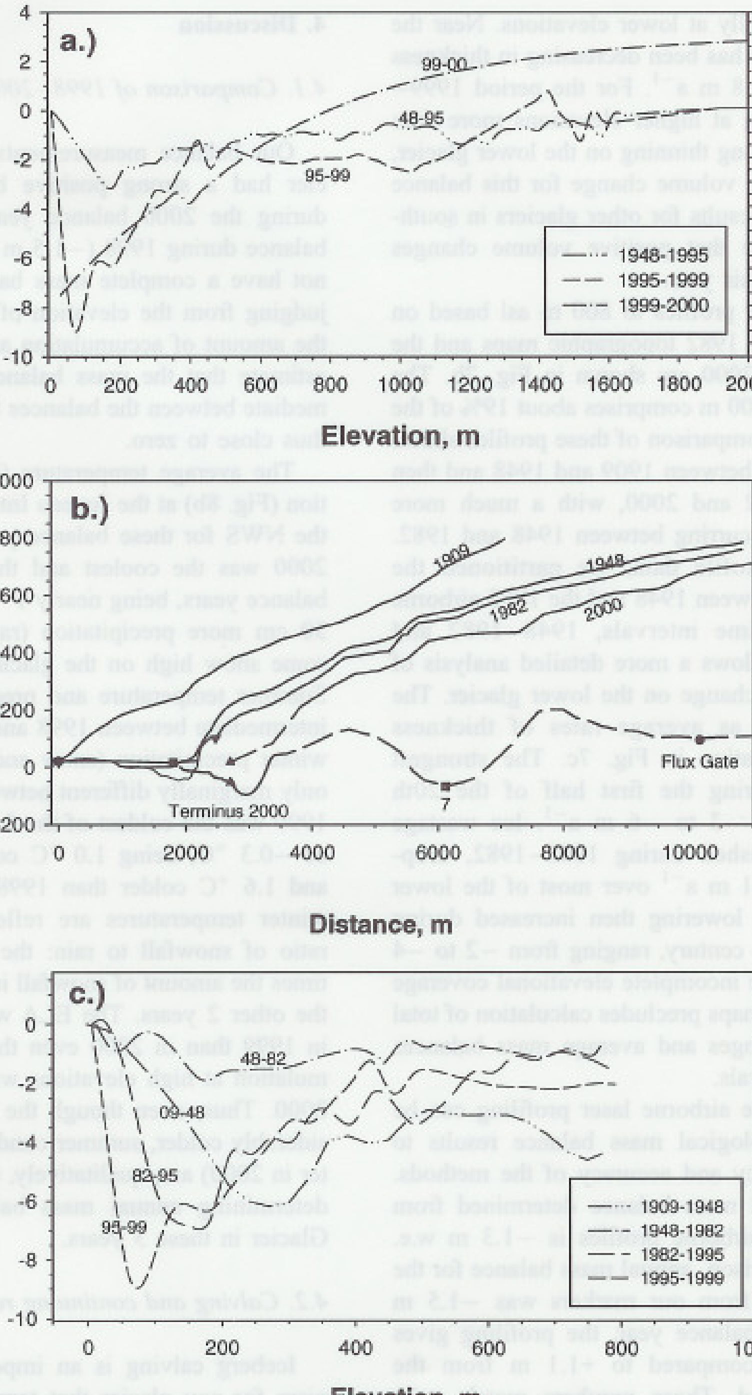


Fig. 7. Changes in glacier surface elevation as a function of elevation. (a) Rate of change over the entire glacier as determined by airborne altimetry (1995, 1999, 2000) and the 1948 topographic map. (b) Centerline surface profiles of lower glacier up to 800 m asl from topographic maps (1909, 1948, 1982) and from airborne profiling. The glacier bed, from RES, is also shown. (c) Rate of change in thickness (surface elevation) on the lower glacier up to 800 m asl.

4. Discussion

4.1. Comparison of 1998–2000 mass balances

Our surface mass balance measurements indicate that the glacier had a strong positive balance (+1.1 m w.e.) during the 2000 balance year and strong negative balance during 1998 (−1.5 m w.e.). Although we do not have a complete mass balance record for 1999, judging from the elevation of the ELA in 1999 and the amount of accumulation at higher elevations, we estimate that the mass balance for 1999 was intermediate between the balances for 1998 and 2000, and thus close to zero.

The average temperature (Fig. 8a) and precipitation (Fig. 8b) at the Juneau International Airport from the NWS for these balance years show that summer 2000 was the coolest and the wettest of the three balance years, being nearly 1 °C cooler and receiving 25 cm more precipitation (rain in Juneau, possibly some snow high on the glacier) than summer 1998. Summer temperature and precipitation for 1999 are intermediate between 1998 and 2000. In comparison, winter precipitation (snow and rain at the airport) is only marginally different between the 3 years. Winter 1999 was the coldest of the 3 years, with an average of −0.3 °C, being 1.0 °C colder than winter 2000 and 1.6 °C colder than 1998. These differences in winter temperatures are reflected in near-sea-level ratio of snowfall to rain: the airport received three times the amount of snowfall in 1999 than in either of the other 2 years. The ELA was significantly higher in 1998 than in 2000, even though the annual accumulation at high elevations was similar in 1999 and 2000. Thus, even though the 1999 winter was considerably colder, summer conditions (cooler and wetter in 2000) are, qualitatively, the dominant factors in determining annual mass balances of Mendenhall Glacier in these 3 years.

4.2. Calving and continuing retreat

Iceberg calving is an important ablation mechanism for any glacier that terminates in an ocean or lake, and it is also an important mechanism of glacier [illegible]

recent years, especially at lower elevations. Near the terminus, the glacier has been decreasing in thickness at a rate of about −8 m a^{-1}. For the period 1999–2000, the thickening at higher elevations more than compensated the strong thinning on the lower glacier, leading to a positive volume change for this balance year. Our profiling results for other glaciers in southeast Alaska indicate that negative volume changes were the norm for [illegible]

Centerline glacier profiles to 800 m asl based on the 1909, 1948, and 1982 topographic maps and the airborne profile of 2000 are shown in Fig. 7b. The glacier area below 800 m comprises about 19% of the total glacier area. Comparison of these profiles shows substantial thinning between 1909 and 1948 and then again between 1982 and 2000, with a much more moderate change occurring between 1948 and 1982. Using the 1982 profile allows us to partition the elevation change between 1948 and the 1995 airborne profile into two time intervals, 1948–1982 and 1982–1995. This allows a more detailed analysis of trends in thickness change on the lower glacier. The results are plotted as average rates of thickness change versus elevation in Fig. 7c. The strongest losses occurred during the first half of the 20th century, at rates of −2 to −6 m a^{-1}, but rates substantially diminished during 1948–1982, dropping to less than −1 m a^{-1} over most of the lower glacier; the surface lowering then increased during the latter part of the century, ranging from −2 to −4 m a^{-1}. Note that the incomplete elevational coverage on the topographic maps precludes calculation of total glacier volume changes and average mass balances for these time intervals.

The results of the airborne laser profiling can be compared to glaciological mass balance results to check for consistency and accuracy of the methods. The average annual mass balance determined from the 1995 to 1999 airborne profiles is −1.3 m w.e. (Table 2). In comparison, the glaciological mass balance for the 1998 balance year from our analysis was −1.5 m w.e. For the 2000 balance year, the profiling gives +0.9 m (w.e.) as compared to +1.1 m from the glaciological method. These numbers match quite well, given the uncertainties inherent to the glacio- [illegible] recent years, especially at lower elevations. Near the terminus, the glacier has been decreasing in thickness at a rate of about -8 m a^{-1}. For the period 1999–2000, the thickening at higher elevations more than compensated the strong thinning on the lower glacier, leading to a positive volume change for this balance year. Our profiling results for other glaciers in southeast Alaska indicate that positive volume changes were the norm for this year.

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4. Discussion

4.1. Comparison of 1998–2000 mass balances

Our balance measurements indicate that the glacier had a strong positive balance (+1.1 m w.e.) during the 2000 balance year and strong negative balance during 1998 (-1.5 m w.e.). Although we do not have a complete mass balance record for 1999, judging from the elevation of the ELA in 1999 and the amount of accumulation at higher elevations, we estimate that the mass balance for 1999 was intermediate between the balances for 1998 and 2000, and thus close to zero.

The average temperature (Fig. 8a) and precipitation (Fig. 8b) at the Juneau International Airport from the NWS for these balance years show that summer 2000 was the coolest and the wettest of the three balance years, being nearly 1 °C cooler and receiving 30 cm more precipitation (rain in Juneau, possibly some snow high on the glacier) than summer 1998. Summer temperature and precipitation for 1999 are intermediate between 1998 and 2000. In comparison, winter precipitation (snow and rain at the airport) is only marginally different between the 3 years. Winter 1999 was the coldest of the 3 years, with an average of -0.3 °C, being 1.0 °C colder than winter 2000 and 1.6 °C colder than 1998. These differences in winter temperatures are reflected in near-sea-level ratio of snowfall to rain: the airport received three times the amount of snowfall in 1999 then in either of the other 2 years. The ELA was significantly higher in 1999 than in 2000 even though the annual accumulation at high elevations was similar in 1999 and 2000. Thus, even though the 1999 winter was considerably colder, summer conditions (cooler and wetter in 2000) are, qualitatively, the dominant factors in determining annual mass balances of Mendenhall Glacier in these 3 years.

4.2. Calving and continuing retreat

Iceberg calving is an important ablation mechanism for any glacier that terminates in an ocean or lake, and it is also an important mechanism of glacier retreat. There are important unanswered questions regarding calving dynamics, in particular the differences between tidewater calving and freshwater lake

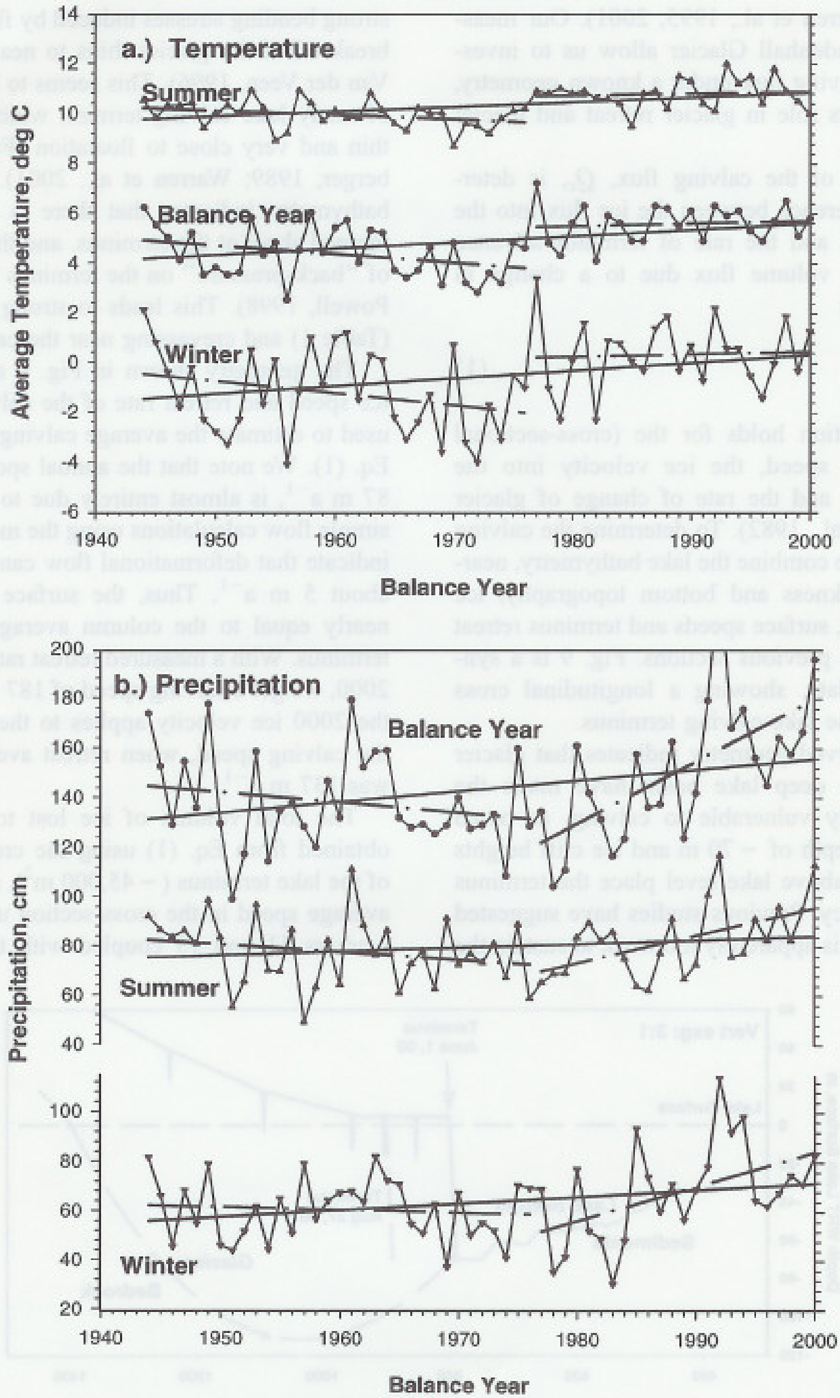

Fig. 8. Average seasonal and balance year temperatures (a) and precipitation (b) at the Juneau International Airport weather station. Solid lines are linear regressions fitted to the entire data set; they show a general increase in both temperature and precipitation over the period of record. Dash–dot lines highlight separate linear regressions fitted to data from 1943 to 1976 and from 1977 to 2000. These regressions show a general decline in both temperature and precipitation from 1943 to 1976 and a reversal of this trend after 1977.

calving (e.g., Warren et al., 1995, 2001). Our measurements on Mendenhall Glacier allow us to investigate the lake calving flux under a known geometry, and to quantify its role in glacier retreat and glacier mass balance.

Quantification of the calving flux, Q_c, is determined as the difference between the ice flux into the calving front, Q_i, and the rate of terminus advance expressed as the volume flux due to a change in glacier length, Q_t:

$$Q_c = Q_i - Q_t \tag{1}$$

A similar relation holds for the (cross-sectional average) calving speed, the ice velocity into the calving terminus, and the rate of change of glacier length (Brown et al., 1982). To determine the calving speed and flux, we combine the lake bathymetry, near-terminus ice thickness and bottom topography, ice surface elevations, surface speeds and terminus retreat rates discussed in previous sections. Fig. 9 is a synthesis of these data, showing a longitudinal cross section through the lake calving terminus.

First, the observed geometry indicates that glacier thinning and the deep lake basin have made the glacier particularly vulnerable to calving, as noted earlier. A water depth of ~70 m and ice cliff heights of only 5–10 m above lake level place the terminus very near buoyancy. Previous studies have suggested that temperate ice is apparently too weak to sustain the strong bending stresses induced by floatation, and will break off if the glacier thins to near buoyancy (e.g., Van der Veen, 1996). This seems to be especially true of many lake calving termini, which are often quite thin and very close to floatation (Funk and Rothlesberger, 1989; Warren et al., 2001). In addition, our bathymetry indicates that there is little or no submerged shoal at the terminus, and thus there is a lack of "back-pressure" on the terminus lobe (Fischer and Powell, 1998). This leads to strong extensional flow (Table 1) and crevassing near the calving terminus.

The geometry shown in Fig. 9 and the measured ice speed and retreat rate of the calving front can be used to estimate the average calving speed, following Eq. (1). We note that the annual speed of marker 13, 87 m a^{-1}, is almost entirely due to basal motion, as simple flow calculations using the measured geometry indicate that deformational flow can only account for about 5 m a^{-1}. Thus, the surface velocity is very nearly equal to the column average velocity at the terminus. With a measured retreat rate of 100 m a^{-1} in 2000, we get a calving speed of 187 m a^{-1}. Assuming the 2000 ice velocity applies to the two prior years, the calving speed, when retreat averaged 50 m a^{-1}, was 137 m a^{-1}.

The total volume of ice lost to calving can be obtained from Eq. (1) using the cross-sectional area of the lake terminus (~45,000 m^2), an estimate of the average speed in the cross section using the speed at markers 11 and 13 coupled with the known cross-

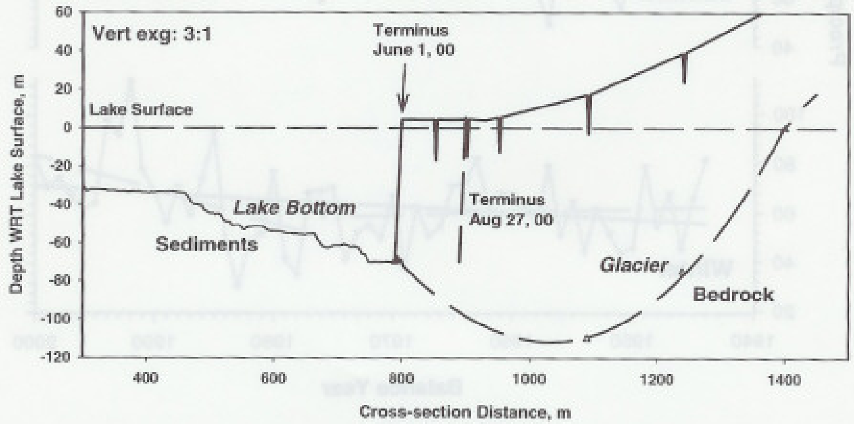

Fig. 9. Cross section through calving terminus based on lake bathymetry, RES, surveyed glacier surface, and surveyed terminus position. Location of cross section is shown in Fig. 4. Continued lake-calving retreat will lead to increased instability because the subglacial basin deepens behind the present terminus.

sectional geometry (Nye, 1965), and the measured variations in retreat rate across the lake terminus. This gives an annual calving flux, Q_c, of $-6.4\ (\pm 1.0) \times 10^6$ $m^3\ a^{-1}$ w.e. for balance year 2000 and about $-4.0 \times 10^6\ m^3\ a^{-1}$ w.e. for 1998 and 1999. For comparison, total ablation losses due to surface melting in 2000 equaled $-112 \times 10^6\ m^3\ a^{-1}$ w.e. Thus, loss by calving was about 6% of ice lost by surface melting. Similarly, the annual volume change between 1948 and 1999 was $-104 \times 10^6\ m^3\ a^{-1}$ w.e., so the calving flux was again about 6% of the long-term average glacier-wide volume loss. These comparisons show that lake calving was relatively minor compared to surface melt in terms of total volume loss. However, it was an important mechanism of Mendenhall Glacier terminus retreat throughout the period of study.

When a glacier retreats into deeper water, be it ocean or freshwater, the rate of retreat often increases (Brown et al., 1982; Funk and Rothlesberger, 1989; Warren et al., 2001). Our ice thickness measurements indicate that the basin is about 30 m deeper 200 m behind the August 2000 terminus (Fig. 9). This significant overdeepening will probably increase the potential instability of the glacier as it calves back. There is no indication in the bathymetry of a growing sublacustrian moraine that would increase "back-pressure" on the calving terminus, and because of the resilient bedrock in the Mendenhall basin, it is unlikely that sedimentation in this overdeepening will occur at a rate sufficient to counter this instability (Hunter et al., 1996).

The glacier bed remains below average lake level about 500 m upstream of the present terminus (Fig. 9). Given that the current rate of retreat is 50–100 m a^{-1}, and the fact that the basin will get deeper as the glacier recedes, we can expect the terminus to continue to retreat another 500 m over the next one to two decades, and then become a terrestrial terminus. During the early part of its present retreat phase (about 1940–1980, Figs. 1 and 4), the entire terminus was calving into the lake. Then the western part of the terminus retreated up and over a bedrock sill, where it has terminated on land ever since. Surface melting of about 15 m a^{-1} w.e. causes continued but slow retreat of this land terminating section of the terminus. The largest proportion of the ice flux on the lower glacier is now directed toward the lake calving section of the terminus (Fig. 2). However, this increased ice flux is substantially less than the present calving flux. Thus, from Eq. (1), the rate of retreat of the calving part of the terminus is significantly greater than that at the land terminating part, as shown in Fig. 4. This increased retreat rate will likely lead to a change in the geometry of the terminus, with an indentation following the bedrock channel inland of the present lake terminus.

4.3. Ice thickness changes

At about 500 m elevation (site 7), the glacier thinned 100 m between 1909 and 1948, and thinned another 100 m between 1948 and 2000, with an overall thinning of nearly 35%. At the flux-gate (700 m asl), the 1948–2000 thinning was 80 m, or 12% of the 1948 thickness (Fig. 7b). Near the equilibrium line (~1150 m), the ice has thinned by 50–60 m between 1948 and 2000, again approximately 10% of the thickness. These are substantial changes in ice thickness, and they will affect both the speed of the glacier [by as much as four times the percentage change in thickness (Echelmeyer, 1983)] and the position of the terminus. Of course, we must expect a time lag between the thinning and the retreat of the terminus. An estimate of the response time for length changes of Mendenhall Glacier is about 45 years, following the ideas of Harrison et al. (2001) and assuming an average thickness of about 400 m, an ablation rate on the lower glacier of -12 m a^{-1} and a balance rate gradient on the lower glacier (~0.008 m $m^{-1}\ a^{-1}$) obtained from the curve in Fig. 3.

4.4. Ice flux as a check on volume loss and mass balance

Our measurements of ice thickness and surface speeds at the "flux-gate" allow us to calculate the ice volume flux through the cross section of the glacier at this location. Because this flux must equal the volume loss over the glacier below this section, we can use the calculated flux as an independent estimate of the combined volume loss due to thinning and the surface-integrated mass balance of the lower glacier plus the calving loss. That is, the volume flux through the "flux-gate" cross section, Q_f, is given by the sum of three terms:

$$Q_f = Q_{sb} + Q_c + \Delta V \tag{2}$$

The ice flux, Q_f, can be determined by summing over the product of average column velocities [the surface velocity, from Fig. 2, scaled by a shape factor (Nye, 1965)] and the area of each column in the cross section, using the measured cross-sectional shape in Fig. 5d. The integrated surface balance below the cross section, Q_{sb}, is evaluated by summing the product of specific mass balance for each elevation increment, b_j, from Fig. 3, and the surface area within that elevation increment, S_j over N increments from the terminus up to the flux-gate:

$$Q_{sb} = \sum_{j=1}^{N} b_j S_j \qquad (3)$$

Q_c are losses by calving into the lake, as averaged over the past several years. ΔV is the change in glacier volume downstream of the cross section (due both to thinning and any changes in area, such as those that arise during terminus retreat). If the glacier were in equilibrium with the present climate, there would be no change in thickness or terminus position, and ΔV would be zero. Solving Eq. (2) for the volume change gives a measure of the state of the glacier. The flux through the gate is currently 0.82×10^8 m^3 a^{-1} w.e. We have only 2 years of detailed mass balance records for the lower glacier, and we used the average of these to compute the surface balance below the gate. This will tend to underestimate average surface ablation because 2000 had the coolest, wettest summer in the past several years. The resulting (minimum) integrated surface balance, Eq. (3), is 1.06×10^8 m^3 a^{-1} w.e. The calving losses have averaged about 0.05×10^8 m^3 a^{-1} w.e. over the past several years (Section 4.2). From Eq. (1), we obtain a minimum estimate of the volume loss: -0.29×10^8 m^3 a^{-1} w.e. In comparison, the computed annual volume change below the flux gate using the 1995 and 1999 airborne profiles is -0.38×10^8 m^3 a^{-1} w.e. These numbers are in reasonable agreement, given the uncertainties in the various measurements and the use of the average 1998–2000 specific balances.

4.5. Long-term trends

Despite the strong positive balance in 2000, the long-term trend at Mendenhall Glacier has been decidedly negative (Fig. 7; Table 2). Except for a standstill or brief re-advance during the 1830s, Mendenhall Glacier has been in steady recession since the late 18th century (Fig. 6). Recession accelerated during the 20th century. Over 60% of the 20th century retreat transpired between 1909 and 1948 (Fig. 6), a trend echoed in the strong thinning of the lower glacier during this period (Fig. 7). Weather records for Juneau before 1943 are unreliable and therefore we cannot ascertain how much of this thinning and subsequent shrinkage is attributable to temperature or precipitation changes. However, we note that the average rate of retreat began increasing between 1931 and 1948, after Mendenhall Lake began forming. During the latter part of that period, the lake terminus was nearly 2 km wide (twice the current width) and the proglacial basin lake was at least 70 m deep in places (Fig. 4). Thinning at the terminus would have rendered the glacier increasingly unstable as the ice thickness approached buoyancy. The retreat rate between 1945 and 1948 was the highest on record before 1997–2000 (Fig. 6).

Glacier buoyancy has been suggested as a major factor in determining the rate of calving (Warren et al., 2001). The conditions in Mendenhall Lake and the terminal region of the glacier suggest that increased calving was at least partially responsible for the higher-than-average rate of retreat during the mid-1940s. Increased calving is commonly accompanied by increases in ice flux to the calving front that causes a draw-down of ice and an increased rate of glacier thinning (Hughes, 1986).

The trends at Mendenhall Glacier during the latter half of the 20th century can be compared to weather data recorded at Juneau since August 1943 (Figs. 8a and b), although such a comparison must necessarily be qualitative because of the limited balance record. A linear regression fitted to the Juneau data shows that the average annual temperature has risen by ~1.6 °C over the period of record, while "winter" temperature has risen by ~1.9 °C and "summer" by ~1 °C. Annual precipitation has increased by about 15 cm, with winter precipitation increasing slightly more than summer. We have not investigated changes in cloudiness or incoming radiation, but the 1 °C rise in average summer temperature undoubtedly increased ablation rates over the glacier. It may also be that the rising winter temperature in this maritime climate has caused an increase in the amount of precipitation falling on the glacier as rain, as has been found on South Cascade Glacier in Washington state by McCabe and Fountain (1998). Both of these effects would lead to more negative glacier mass balances, and thus we conclude that rising temperature is the dominant climatic factor affecting Mendenhall Glacier mass balance during the second half of the 20th century.

We must invoke a caveat at this stage and remember that this conclusion is based on the weather records at the Juneau Airport, 10 km from the terminus, and not at the glacier. Although temperatures on the glacier are likely to follow a similar pattern of warming, there are strong variations in precipitation in the Juneau area caused by orographic effects. Precipitation at the glacier may differ substantially from that at the airport, particularly at higher elevations. However, we have no pre-existing data on the corresponding variables measured on the glacier, and climatological modeling of these effects is beyond the scope of this study.

In addition to the long-term trends in temperature and precipitation, there are shorter term variations. For example, there appear to be shifts in both temperature and precipitation around 1976–1977. If a linear regression is fitted to the temperature record from the 1940s to 1976, it indicates an overall cooling trend (Fig. 8a). After that, the temperature began increasing. Precipitation also shows a shift from a decreasing trend from the 1940s to about 1976, to an increasing one from about 1977 onward (Fig. 8b, note regression lines over individual time periods). The changing trends in precipitation are counter to the observed changes in mass balance (slower rates of thinning with decreasing precipitation and faster thinning with increasing precipitation), giving further support to the dominant role of temperature in determining Mendenhall Glacier's mass balance. Increased winter air temperatures can significantly reduce snow precipitation at higher elevations (McCabe and Fountain, 1998). Mendenhall Glacier is particularly susceptible because its accumulation zone lies at a relatively low elevation.

These changing trends, and their timing, agree qualitatively with the "regime shift" around 1977 that have been noted in other climatological records (Mantua et al., 1997; Minobe, 1997; Stephens et al., 2001). The "regime shift" includes a warming of eastern equatorial Pacific Ocean waters by as much as 1.5 °C and a cooling of North Pacific waters of 1 °C, in conjunction with an atmospheric shift in sea level pressure around the mid-1970s (Stephens et al., 2001). Hodge et al. (1998) have discussed evidence for a 1977 regime shift in the mass balance records of Wolverine Glacier (located on the Kenai Peninsula, Alaska) and South Cascade Glacier in Washington State, two maritime glaciers that are located 3000 km apart. Mass balance for Wolverine Glacier shows strongly negative mass balance from 1965 (start of measurements) to 1977. In contrast, South Cascade Glacier was generally just slightly negative or positive between 1959 (start of measurements) to 1977. Both Wolverine and South Cascade showed a change in mass balance in the decade following 1977, interestingly in reverse phase, with Wolverine showing an increase in balance while South Cascade decreased. However, both glaciers suffered decreasing mass balance after 1989.

Lower Mendenhall Glacier experienced a dramatic slowdown in the rate of thinning during the period 1948–1982. This slowdown is unlikely to be due to changes in lake calving, given the small fraction that present-day calving contributes to total mass balance. It is more probable that the declining temperatures experienced before 1977 caused a reduction in ablation, and possibly even led to some positive annual balances. The increased rate of thinning after 1982, and the associated negative mass balances measured over the entire glacier, agrees qualitatively with the post-1977 local temperature increase.

Although our balance record involves mainly long-term averages, the measured and inferred changes in Mendenhall Glacier mass balance over the last 50 years are qualitatively in phase with those of South Cascade Glacier, some 1600 km to the south. These mass balance trends parallel the continuation of the mid-1970s "regime shift" of warmer temperatures in eastern equatorial Pacific Ocean waters and cooler waters in the North Pacific though 1998 (Stephens et al., 2001). Mass balance records at nearby Lemon Creek Glacier from 1953 to 1997 show similar trends as Mendenhall Glacier (Miller and Pelto, 1999), but the advancing Taku Glacier (Motyka and Beget, 1996), which is also nearby, appears to have had a positive balance almost continuously from 1958 to 1986 (Pelto and Miller, 1990).

4.6. Effects on valley hydrology

Our airborne altimetry gives an annual glacier volume change from fall 1995 to fall 1999 of −0.16 km^3 a^{-1}. Converting to water equivalent, this averages to ~146 million m^3 of water being released each year from the glacier as a result of excess melting of the glacier. Based on the discharge record from the USGS Mendenhall River gaging station (http://www.waterdata.usgs.gov/) from 1996 to 1999, this excess ice melt contributes about 13% of the annual discharge from Mendenhall Lake. Water that was locked up as glacier ice during and following the LIA is now being released, contributing to river discharge and, ultimately, contributing to global sea-level rise.

Most glacier melting at Mendenhall occurs during June, July, and August (Fig. 3), and this melt contributes significantly to the summer discharge of Mendenhall River. Average river discharge was nearly identical for both 1998 and 2000 summers at about 82 m^3 s^{-1}. The average run-off from glacier melt computed from our summer balance (Fig. 3b) was 30 m^3 s^{-1} in 2000 and 41 m^3 s^{-1} in 1998, or about 37% and 50% of the average summer discharge in Mendenhall River in these 2 years, respectively. Higher summer precipitation in 2000 apparently offset the lower melt rate during that year. Average summer discharge in Mendenhall River was remarkably constant between 1966 and 2000, ranging from 65 to 90 m^3 s^{-1}, despite significant variations in total summer precipitation (18–45 cm). Apparently, glacier melt increases during summers of low precipitation, due to a larger number of positive degree days, thus making up for the reduced precipitation in the total river discharge. From this analysis, we conclude that glacier melt is important for maintaining stream flow during prolonged periods of reduced precipitation in southeast Alaska, as has been found elsewhere (e.g., Fountain and Tangborn, 1985).

5. Conclusions

Mendenhall Glacier experienced dramatic thinning during the 20th century, up to 200 m at lower elevations since 1909, and up to 50 m at higher elevations since 1948. The terminus retreated 3 km during this period, creating proglacial Mendenhall Lake. Despite a strong positive balance for the 2000 balance year, the terminus region continues to thin dramatically, at the rate of 8 m a^{-1} since 1995. The glacier has lost 5.5 km^3 of ice since 1948. These dramatic changes appear to be caused primarily by climatic changes and secondarily by lake calving. Qualitative comparison of glacier thinning to Juneau climate statistics indicates that shrinkage of Mendenhall Glacier is mainly due to local warming: the average annual temperature at Juneau has increased ~1.6 °C since 1943. Based on recent mass balance measurements, glacier response appears to depend strongly on summer conditions. Juneau temperatures generally decreased between 1947 and 1976 and have been steadily rising since. These trends are reflected in the rate of thinning on the lower glacier, which averaged ~1 m a^{-1} between 1948 and 1982, and twice that since 1982. South Cascade Glacier responded in a similar fashion, but Wolverine Glacier, at nearly the same latitude, did not (Hodge et al., 1998). These changes correlate qualitatively with a regime shift in Pacific Ocean temperatures (Minobe, 1997; Stephens et al., 2001).

Calving losses have periodically been a small but significant fraction (6%) of ice loss from the glacier, whereas the rate of terminus retreat has at times been greatly affected by calving. These observations emphasize that care must be taken when interpreting terminus retreat in lacustrine environments as representative of glacier health. Ice thickness soundings show that the basin in front of the present terminus deepens behind the calving front, and that this sublacustrine basin extends upglacier for at least another 500 m. The lake terminus is now particularly vulnerable to calving because of the low ice cliffs and deepening basin. If current climatic trends persist, we can expect continued, rapid mass loss from the glacier. Calving rate may also increase, and in the absence of a simultaneous increase of ice velocity, this will lead to recession of the lake terminus onto land, perhaps within one or two decades.

The glacier and the meltwaters that flow from it are integral components of the Mendenhall Valley ecosystem. The glacier contributes significantly to Mendenhall River discharge, up to 50% or more during summer months. Excess glacier ice melt from glacier thinning accounts for ~13% of the current annual discharge of the Mendenhall River.

Based on qualitative observations of other glaciers in northern southeast Alaska (e.g., Field, 1975), our long-term mass balance records for Mendenhall Glacier and that of Miller and Pelto (1999) for Lemon Creek Glacier are representative of regional glacier trends, supporting arguments that non-polar glaciers are contributing significantly to the world-wide rise in sea level (Dyurgerov and Meier, 1997).

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