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Glacier changes in southeast Alaska and northwest British Columbia and contribution to sea level rise

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Received 1 June 2006; revised 21 August 2006; accepted 21 September 2006; published 24 February 2007.

[1] The digital elevation model (DEM) from the 2000 Shuttle Radar Topography Mission (SRTM) was differenced from a composite DEM based on air photos dating from 1948 to 1987 to determine glacier volume changes in southeast Alaska and adjoining Canada. SRTM accuracy was assessed at ±5 m through comparison with airborne laser altimetry and control locations measured with GPS. Glacier surface elevations lowered over 95% of the 14,580 km^2 glacier-covered area analyzed, with some glaciers thinning as much as 640 m. A combination of factors have contributed to this wastage, including calving retreats of tidewater and lacustrine glaciers and climate change. Many glaciers in this region are particularly sensitive to climate change, as they have large areas at low elevations. However, several tidewater glaciers that had historically undergone calving retreats are now expanding and appear to be in the advancing stage of the tidewater glacier cycle. The net average rate of ice loss is estimated at 16.7 ± 4.4 km^3/yr, equivalent to a global sea level rise contribution of 0.04 ± 0.01 mm/yr.

Citation: Larsen, C. F., R. J. Motyka, A. A. Arendt, K. A. Echelmeyer, and P. E. Geissler (2007), Glacier changes in southeast Alaska and northwest British Columbia and contribution to sea level rise, *J. Geophys. Res.*, *112*, F01007, doi:10.1029/2006JF000586.

1. Introduction

[2] Recent studies have documented that the majority of alpine glaciers world-wide have been losing mass during the past century and have been contributing to global sea level rise [e.g., *Dyurgerov and Meier*, 1997; *Arendt et al.*, 2002; *Rignot et al.*, 2003; *Paul et al.*, 2004; *Dyurgerov and McCabe*, 2006]. Although temperate alpine glaciers represent a small fraction of the world's ice mass they are contributing significantly to sea level rise through mass wastage. *Arendt et al.* [2002] showed from analysis of small-aircraft laser altimeter data that ice masses in Alaska and neighboring Canada are thinning so rapidly that they made a larger contribution to global sea level rise than the Greenland Icesheet during the later half of the 20th century. Nowhere in Alaska have these effects been more dramatic than along the southern coastal mountains, where high annual accumulation rates (up to 4 m/yr water equivalent (weq)) and severe annual ablation (up to −14 m/yr weq) [*Pelto and Miller*, 1990; *Eisen et al.*, 2001; *Motyka et al.*, 2002] result in extremely high rates of ice mass exchange.

[3] Most glaciers along the Gulf of Alaska have been retreating since achieving their Little Ice Age (LIA) maximums sometime between 1750 and 1900 AD, in some cases quite rapidly [*Goodwin*, 1988; *Mann and Ugolini*, 1985; *Motyka and Beget*, 1996; *Calkin et al.*, 2001; *Larsen et al.*, 2005]. In this paper we examine glacier changes that have occurred in southeast Alaska and adjoining Canada during the last half of the 20th century by comparing a digital elevation model (DEM) derived from the C band NASA Shuttle Radar Topography Mission (SRTM) flown 11–22 February 2000 to a DEM from the U.S. Geological Survey National Elevation Dataset (NED) combined with the Terrain Resource Information Management Program (TRIM) DEM from Natural Resources Canada.

[4] Our study area includes 14,580 km^2 of glaciers, extending from 55°N, just south of the Stikine Ice Field (the southernmost major ice field that spans the border between southeast Alaska and northwest British Columbia), to 60°N, the approximate limit of SRTM coverage (Figure 1). With a few exceptions, the NED is based on maps that were produced from 1948 aerial photography (Figure 2), and thus the comparison for southeast Alaska generally documents 52 yrs of ice elevation change. Coverage in Canada uses the British Columbia TRIM DEM derived from 1982 and 1987 photography. The difference in the dates of the original aerial photography is problematic, as glacial wastage in this region has been shown to be generally accelerating over the latter half of the 20th century [*Arendt et al.*, 2002]. However, our DEM comparison offers a significant advance in areal coverage over previous glacial change studies here, and we show that the variation of surface elevation changes between various individual glaciers is far greater than the factor of 2 increase in the rate of glacial wastage during the latter part of the 20th century [*Arendt et al.*, 2002]. Our intent in this study is to focus on the spatial rather than temporal variations in regional glacier change and the contribution of these changes to sea level rise, and to

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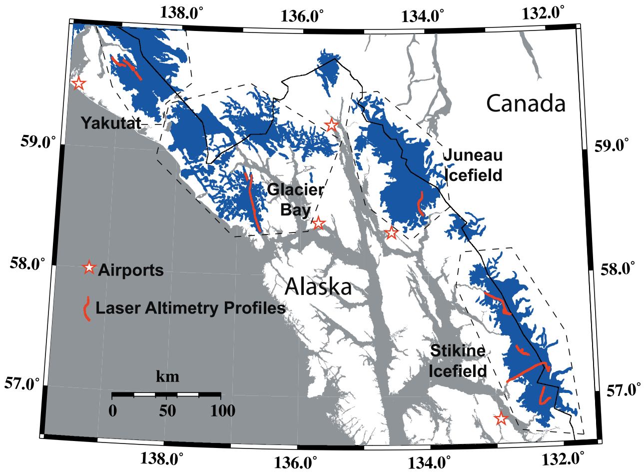

Figure 1. Location map. Glaciers are shown in blue, covering a total of 14,580 km^2. The four main glacier regions are shown by the dashed outlines (Yakutat, Glacier Bay, Juneau and Stikine Ice Fields). Airports and altimetry profiles used as ground control points for the SRTM DEM are shown in red.

explore what might be causing these dramatic and varied changes.

[5] Glaciers in our study area are temperate alpine and predominately maritime that receive abundant precipitation: 7 m/yr to 1.5 m/yr weq with a strong inland gradient (http://www.wrcc.dri.edu/pcpn/ak.gif). Many glaciers in southeast Alaska are now or have recently been tidewater glaciers that calve icebergs and discharge meltwater directly into the sea. Numerous others calve into large proglacial lakes. The main goal of this study is to characterize the ice elevation changes exhibited over this large array of glaciers. Although laser altimetry elevation data are more accurate (±0.3 m), logistical costs prevent complete coverage of all glaciers within a given region, requiring extrapolation of elevation changes to the entire region. Detailed annual mass balance measurements exist for only three glaciers in our region and are either of short duration, of uneven quality or both [*Pelto and Miller*, 1990; *Eisen et al.*, 2001; *Motyka et al.*, 2002]. The spatial coverage afforded by differencing two DEMs allows us to map a surprising variety of glacial changes within our study area. In addition, the method of comparing DEMs has advantages over altimetry and conventional mass balance for estimates of volume contribution to sea level rise because it eliminates the problem of extrapolating from a few known to a large number of unknown glaciers.

2. Methods

[6] We obtained the SRTM and NED DEM for our study region from the USGS web server (http://seamless.usgs.gov), and the Canadian TRIM DEM from Geobase (http://www.geobase.ca). The NED DEM for the majority of our region is based on 1948 photography, and the Canadian DEM is based on 1982 photography south of 59 N, and 1987 photography north of 59 N. Thus our comparison generally documents elevation change between 1948 and 2000 in southeast Alaska, and between 1982/1987 and 2000 in Canada (Figure 2).

[7] However, there are a few noteworthy exceptions (Figure 2). The NED photo base for most of the northern half of the Stikine Ice Field (which includes the tidewater glaciers Dawes, South Sawyer, and Sawyer) (Figure 3) is from August 1961. A portion of the Yakutat Glacier (Figure 3) has photos from 1972. Hypsometry for several glaciers in northeast Glacier Bay, including Muir, Burroughs, and Carroll glaciers (Figure 3), was revised based on air photos from August 1972. Similarly, part of the lower Taku Glacier (Figure 3) hypsometry was adjusted based on its terminus position in 1971 air photos. These revised maps for northeast Glacier Bay and the Taku Glacier are the basis for those portions of the NED DEM. We were able to obtain the original contour maps based on 1948 photographs for both the Taku glacier and northeast Glacier Bay. We digitized glacier contours on these 1948 maps and then modified the NED DEM using these data, so as to better normalize the time span between DEMs in these areas. Similarly dated historic maps are not available for those portions of the Stikine Ice Field and the Yakutat Glacier highlighted in Figure 2, and we were unable to normalize the time span between DEMs there.

[8] The SRTM DEM is available both in 30 m and 90 m spacing. We used the 90 m spacing because we found that both resolutions provided similar results for our large regional coverage. To ascertain vertical precision and any vertical frame bias, we compared surface elevations over 5 airfields (Figure 1) with elevations determined using precision GPS. Airfields were chosen because (1) they are large flat areas, so a large number of DEM grid cells can be compared with the GPS elevations and (2) they are non-vegetated, which is important because SRTM data often gives elevations to the top of the forest canopy, which can be 30–50 m high in southeast Alaska. This comparison indicated no vertical frame bias, and standard deviation of the elevation difference on the airfields is 5 m (i.e., the mean difference between GPS and SRTM on the airfields is 0 ± 5 m).

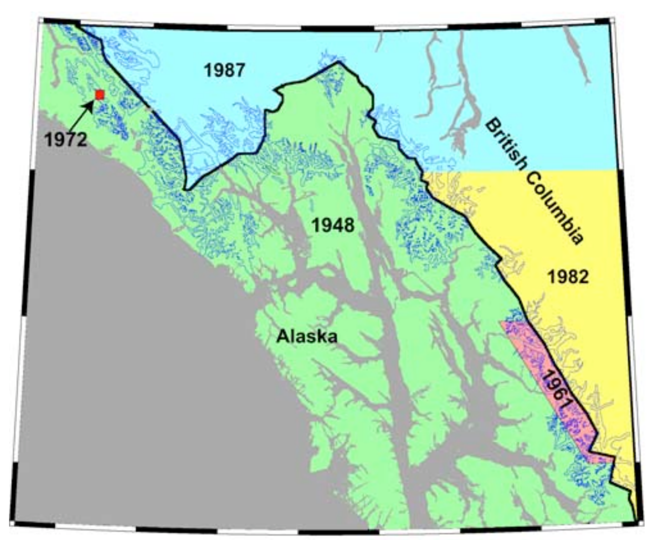

Figure 2. Dates of air photos used to construct the topo maps that formed the earlier DEM.

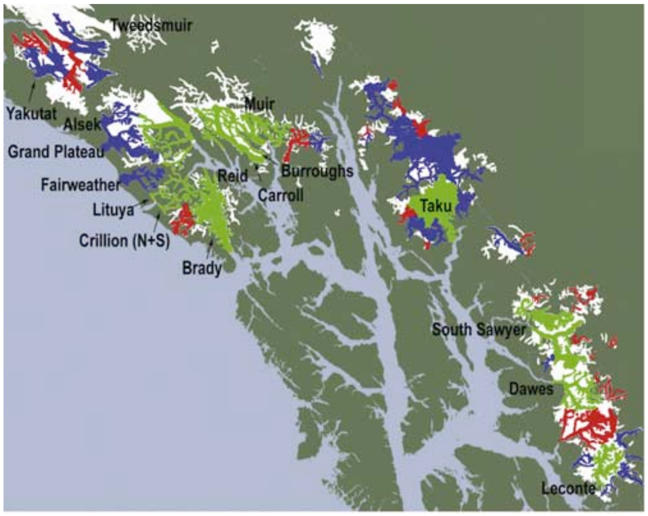

Figure 3. Locations of specific glaciers named in text. The color scheme is the same as in Figures 8 and 9, with green for tidewater glaciers, blue for lake calving glaciers, and red for land terminating. Glacier cover not included in the glacier class analysis discussed in the results section yet displayed in Figures 8 and 9 is shown in white.

[9] The NED DEM for Alaska has a grid spacing of 45 m and was derived by digitizing the original contour maps. These maps have inaccuracies related to a variety of factors [see *Arendt et al.*, 2002, supporting material (SM)]. The nominal random error in elevation is 15 m or one half of the contour interval. However, these uncertainties can be greater at higher elevations and especially in glacier accumulation zones where featureless snow cover can make stereo perception and photogrammetric mapping difficult, and 30 to 45 m errors are possible [*Adalgeirsdöttir et al.*, 1998; *Arendt et al.*, 2002]. We follow these prior assessments of elevation error in Alaskan USGS maps with the assumption that no additional error is introduced in digitizing these maps.

[10] Before differencing, the NED was transformed from the Alaska NAD27 horizontal datum to WGS84, the horizontal datum used in the SRTM. The vertical datums also differ: the SRTM vertical datum is EGM96, while in Alaska the NED is NGVD29. This is problematic, as no standard exists to transform NGVD29 heights in Alaska to any other datum. Indeed, the National Geodetic Survey (NGS) describes the NGVD29 as neither mean sea level, the geoid, nor any other equipotential surface. To estimate the offset between these vertical datums, we found NGS benchmark descriptions with elevations published in both datums, and used a constant, average difference found at tidal benchmarks in Juneau, Haines and Skagway. To test the accuracy of this estimate, we again used the 5 airfields (Figure 1), this time as locations where the elevations of a large number of pixels from both DEMs were directly compared. Both assessments find the NGVD29 to be lower than the EGM96 by 2.3 ± 0.6 m (1σ).

[11] The SRTM was interpolated to the same 45 m grid as the transformed NED. Both the NED and Canadian DEMs were then subpixel registered to the SRTM using an image warping program (the USGS Astrogeology Integrated Software for Imagers and Spectrometers, ISIS). This last step is critically important as small offsets between the DEMs can lead to large errors in an elevation difference map [e.g., *Berthier et al.*, 2004]. The results of differencing the DEMs were then masked with outlines of the glacial cover. The Global Land Ice Measurements from Space project (http://www.glims.org) has digitized a number of glacier outlines in our study area using Landsat images [*Beedle et al.*, 2006]. We used *Beedle et al.*'s [2006] work as a base to redigitize and extend outlines to meet our need for a complete glacier outline database across our study area. We then adjusted these outlines using USGS and Canadian contour maps to more accurately reflect initial conditions in terminus regions where pronounced changes in ice-covered area have occurred at most glaciers. However, on the few advancing glaciers, the outlines were generated solely from new imagery to reflect the present ice-covered area. In general, the masking of the elevation difference map includes ice-covered area if ice was present at the time of either DEM.

3. Accuracy and Error Analysis

[12] To estimate accuracy of SRTM data over glaciers, we compared SRTM data to light aircraft laser altimetry data obtained in late August of 2000 at 12 glaciers distributed throughout the study area (Figure 1). Vertical precision of the altimetry technique is $\sim$0.3 m with points spaced every $\sim$1.2 m along the centerline of each glacier [*Echelmeyer et al.*, 1996; *Adalgeirsdöttir et al.*, 1998; *Arendt et al.*, 2002]. When comparing these data, the SRTM DEM was extrapolated between grid cells with a bicubic interpolation. Approximately 5×10^4 laser altimeter derived elevations were compared to the SRTM DEM (Figure 4). A linear trend was fitted to the differenced data, and the standard deviation about this trend indicates an overall accuracy of the SRTM data over glaciers to be slightly better than ±5 m (Figure 4b).

[13] There are several reasons for elevation differences to be correlated with elevation in the SRTM to laser comparison. The slope (2.6 m per 1000 m elevation) and offset (-2.5 m at zero elevation) of the trend with elevation is what one might expect from the seasonal difference between the two measurements (late August 2000 for the altimetry measurements versus February 2000 for SRTM), associated with ablation, accumulation, densification, and ice flow. The contour maps that form the basis for the earlier DEMs for southeast Alaska and adjoining Canada were derived from aerial photography flown during late summer, so adjusting the SRTM to a late summer surface is desirable in order to examine nonseasonally affected changes in surface elevation. In addition to seasonal differences, the elevation difference trend is also partially caused by radar penetration depth and its dependence on elevation [*Rignot et al.*, 2001]. We take the linear trend shown in Figure 4 to account for both seasonal and radar penetration effects and use it to adjust the SRTM DEM as a function of elevation prior to differencing with the earlier DEM.

[14] *Berthier et al.* [2006] have found SRTM elevation biases over non-glacier-covered terrain, but with a stronger dependence on elevation than we find. These biases may be introduced in the processing of the SAR data used for the

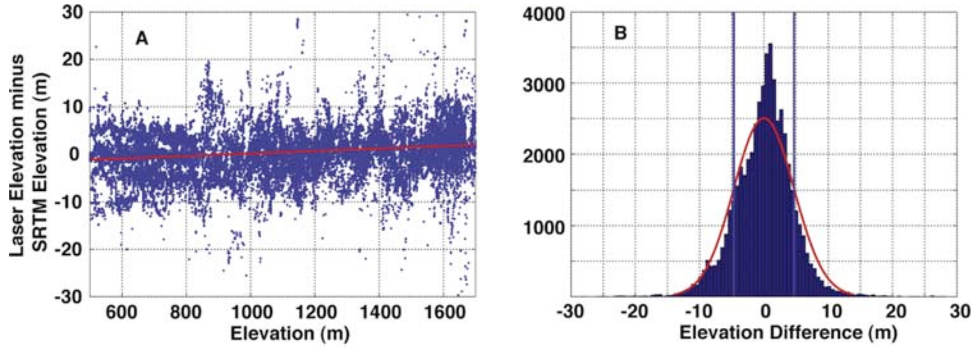

Figure 4. (a) Laser minus SRTM elevation difference versus elevation. (b) Distribution of laser minus SRTM elevation difference. The red curve is a Gaussian fit over the distribution, and the blue vertical lines bracket the standard deviation (SD = ±5 m).

SRTM DEM and may increase more rapidly at higher elevations [*Berthier et al.*, 2006; B. Rabus, personal communication, 2005]. The range of elevations over which we compare SRTM and laser altimetry elevations is somewhat lower than the elevations used in the analysis of *Berthier et al.* [2006]. Furthermore, our only comparison over non-glacier-covered terrain of SRTM elevations to GPS derived elevations is limited to near sea level at the 5 airfields shown in Figure 1.

[15] We use 5 m as a measure of uncertainty of the SRTM elevation data. For NED and Canadian DEM data, we assume an uncertainty of ±15 m for elevations below the equilibrium line altitude (ELA) (which averages about 1000 m for our region) and ±30 m above the ELA. Altimetry comparisons have suggested that individual sections of the original contour maps may have systematic errors of up to −13 m in the Chugach Range and +45 m in the Brooks Range due to poor ground control or photogrammetric errors, with a normal range on the order of ±2 m [*Arendt et al.*, 2002, SM; *Arendt et al.*, 2006, SM]. The uncertainty in estimating the offset in vertical datums is ±0.6 m. The first two errors are random in nature while the second two are systematic, but overall the errors associated with contour maps dominate. Standard propagation of the random errors leads to a combined per pixel 1σ uncertainty in elevation change from our DEM comparisons of ±16 m at lower elevations and ±30 m at higher elevations, with an additional 2.6 m systematic error possible. For rate of elevation change, uncertainty values at low and high elevations are ±0.3 m/yr and ±0.6/yr for 1948–2000 (most of southeast Alaska), ±0.4 m/yr and ±0.8/yr for 1961–2000 (parts of the Stikine Ice Field in Alaska), ±0.9 m/yr and ±1.6 m/yr for 1982–2000 (BC, south of 59 N), and ±1.2 m/yr and ±2.3 m/yr for 1987–2000 (BC, north of 59 N).

[16] To test the uncertainties estimated above, we differenced the two DEMs over non-ice-covered areas. The mean of the off-ice difference map is zero, suggesting that our transformation between vertical datums prior to differencing the DEMs was correct. However, the median is slightly positive and, indeed, the distribution of non-ice elevation differences is strongly non-Gaussian (Figure 5). This is problematic, as the use of standard deviation (SD) as an estimate of the spread of the data is now not valid. We instead use the interquartile range (IQR) as an estimate of the errors associated with the DEM differencing. We find an IQR of ±16.5 m over more than 1×10^8 pixels comparisons (representing over 200,000 km^2 of non-ice-covered area). This error approximation is just slightly greater than the formal errors estimated above. When expressed in terms of rate of elevation change errors, this IQR corresponds to ±0.3 m/yr. Some minor dependence in the magnitude of this error estimate was found with elevation and slope. It should be noted, however, that most of the higher elevations in our difference map are ice covered, and so an independent test of the errors associated with accumulation areas (as discussed above) is not feasible.

[17] An additional source of error is the uncertainty in mapping of ice covered area. Sensitivity tests in which several different glacier outline databases were used in volume calculations in the western Chugach Mountains show that volume change estimates there varied 10%

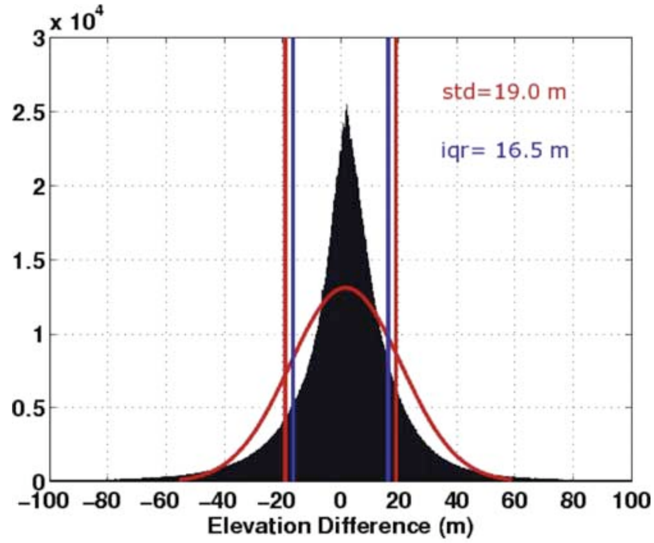

Figure 5. Distribution of DEM difference map values over nonglacier covered areas as used to estimate glacier elevation change errors. The red curve is a Gaussian over the distribution, showing poor quality of fit. The standard deviation (“std,” bracketed by red lines) is 19.0 m, but a more robust estimate of the spread is shown by the interquartile range (“iqr,” bracketed by blue lines), which is 16.5 m for this distribution.

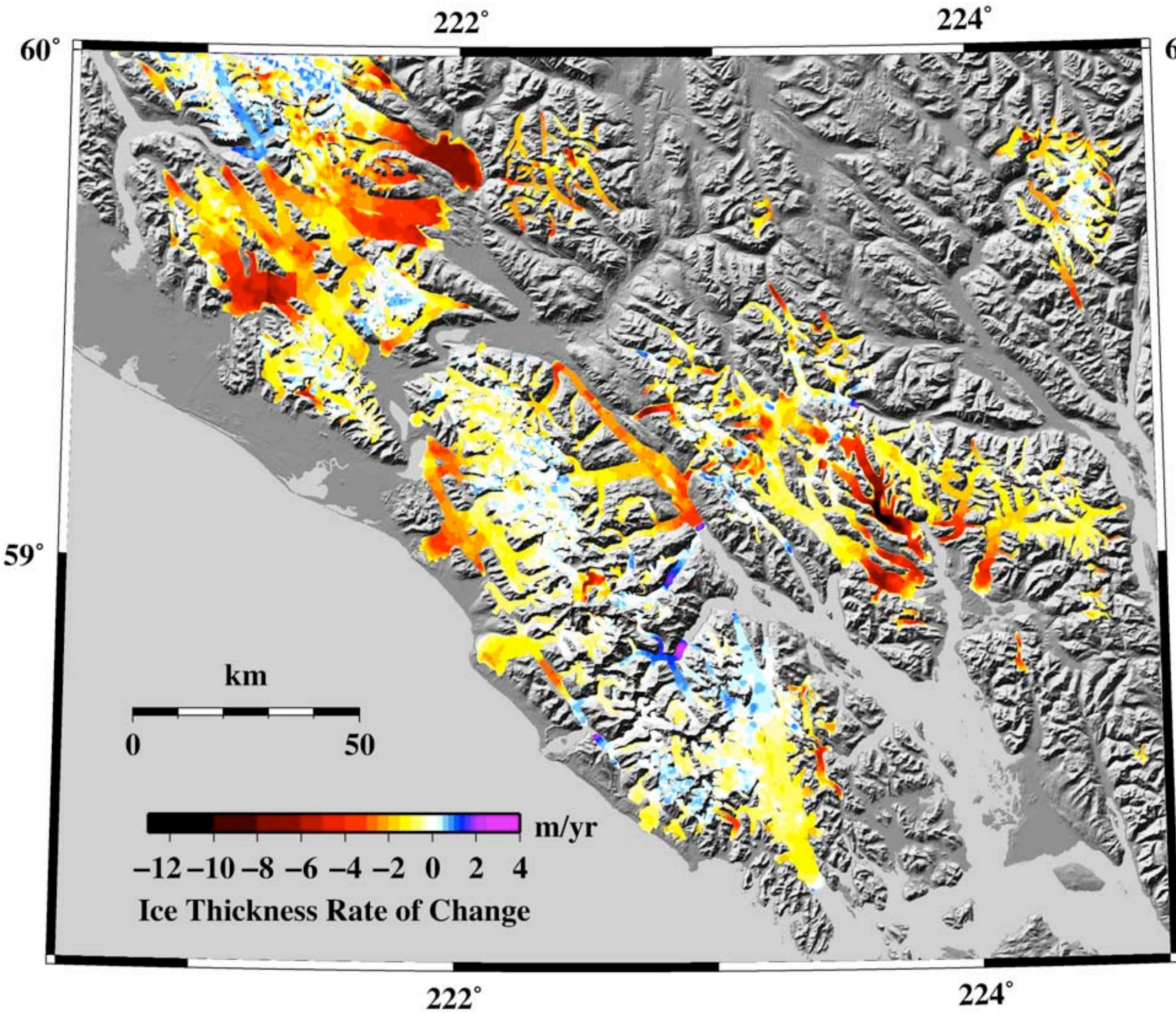

Figure 6. Glacier surface elevation changes, Yakutat region and Glacier Bay region.

between the various outlines used [*Arendt et al.*, 2006]. For this study, we have only one glacier outline database available and so are not able to determine the errors introduced from glacier area uncertainties.

4. Results: Elevation and Volume Changes

[18] With a few notable exceptions, the majority of glaciers in southeast Alaska and adjoining Canada showed strong thinning and retreat during this period. For regions where the photo base is from 1948, we have measured up to a maximum of 640 m total elevation change (on the lower reaches of the Muir Glacier). The greatest changes occurred at lower elevations but large changes are also apparent at higher elevations. Interestingly, 5% of the study area experienced some amount of thickening, particularly Taku Glacier. The net ice loss over the region between the photo-based DEM and the SRTM is 870 ± 140 km^3, corresponding to a total contribution to sea level rise of 2.4 ± 0.4 mm.

[19] Because of the different time frames of the photo-based DEMs, we present the elevation changes as rates to allow better comparison across the entire region. Figures 6 and 7 illustrate the dramatic glacier changes that have occurred throughout southeast Alaska and adjoining Canada during the last half of the 20th century. When the map of volume changes is divided by the map of time spans between DEMs (Figure 2), the average rate of volume loss is 16.7 ± 4.4 km^3/yr.

[20] By combining data from two periods (1948–2000 and 1982/1987–2000), this calculation of an average volume rate assumes that there has been no acceleration in rates of ice loss over the later half of the 20th century, which is not the case, as *Arendt et al.* [2002] have shown. However, the wide range of individual glacier responses and behaviors shown in Figures 6 and 7 suggests that the task of accurately monitoring and characterizing this entire glacier ensemble both spatially and temporally requires considerably more data than either this study or the current scope of the laser altimetry program have available. If anything, our results for the volume loss rate of 16.7 ± 4.4 km^3/yr for this region may underestimate current ice loss rates, or those of the last decade of the 20th century.

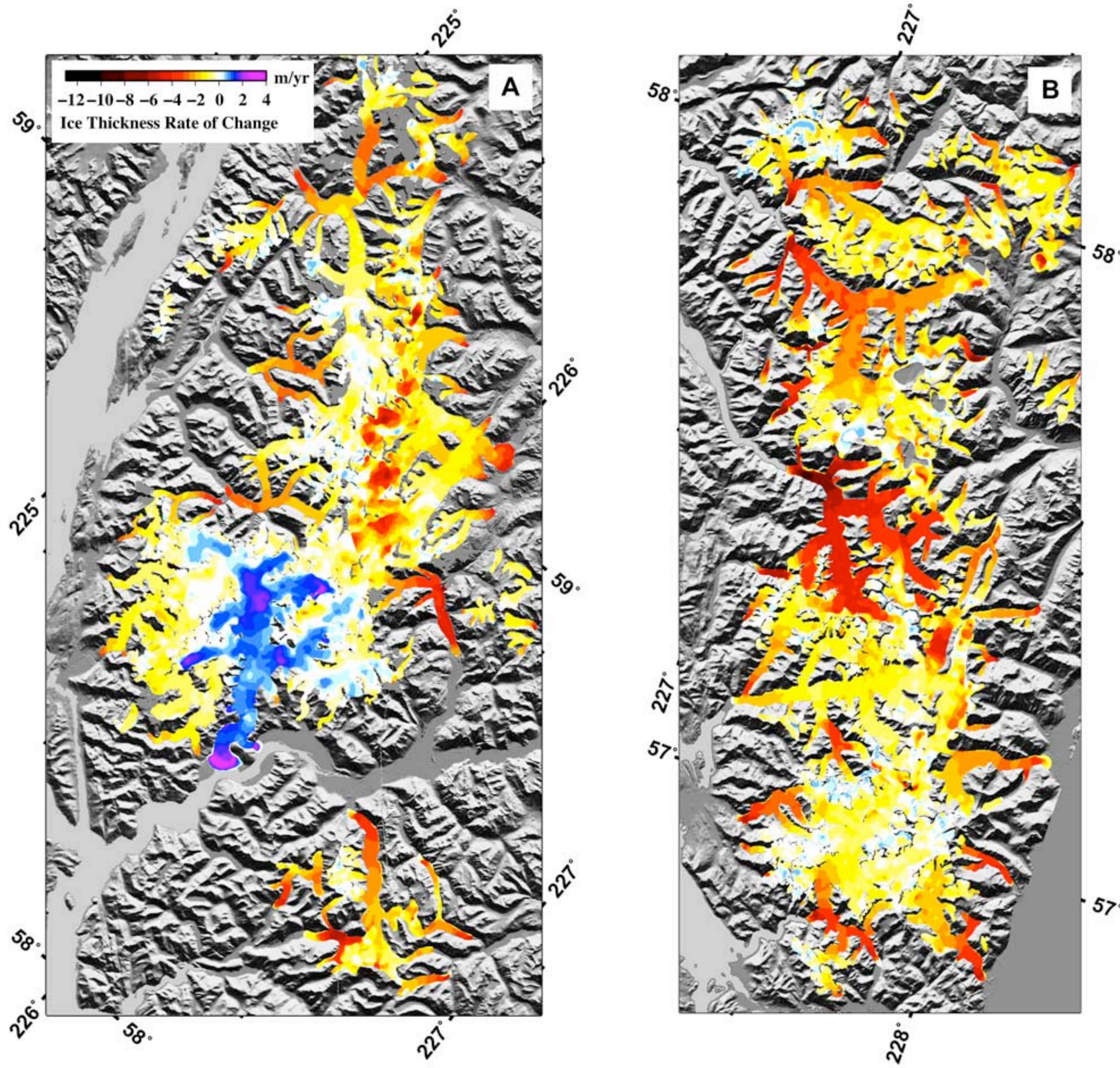

Figure 7. Glacier surface elevation changes, (a) Juneau Ice Field and (b) Stikine Ice Field.

[21] Most of the glacierized terrain in southeast Alaska and adjoining Canada falls into four distinct regions: Yakutat, Glacier Bay, Juneau Ice Field, and Stikine Ice Field (Figure 1). Glaciers in our study area account for about 17% of the total glacier area in Alaska and adjoining Canada. This glacier-covered area is three times the area of Swiss Alps glaciers and equal to the combined area of the Northern and Southern Patagonia Ice Fields. The volume changes documented here show that southeast Alaska and adjoining Canada contributed an average of 0.04 ± 0.01 mm/yr to sea level rise during the later half of the 20th century, assuming all the volume lost is ice with a density of 0.9 kg/m^3. The rate of sea level rise contribution we find from glaciers in southeast Alaska and adjoining Canada is effectively equal to that from all Patagonia glaciers during the period 1968/1975–2000 (0.042 mm/yr) [*Rignot et al.*, 2003].

[22] To obtain a rough estimate of the relative contributions to ice loss from different classes of glaciers we analyzed ice volume changes on 74 individual glaciers: 32 land terminating glaciers (covering 2054 km^2), 20 tidewater glaciers (covering 4033 km^2), and 22 lake calving glaciers (covering 2870 km^2). The total area analyzed by glacier class thus comprises about 62% of the region's total glacier covered area (Figure 3). Figure 8 shows the wide variations in area-averaged thinning rates on various individual glaciers from across our study region. The results of this analysis show that over two thirds of the losses are coming from calving glaciers and that the losses from lake glaciers are slightly greater than those from tidewater

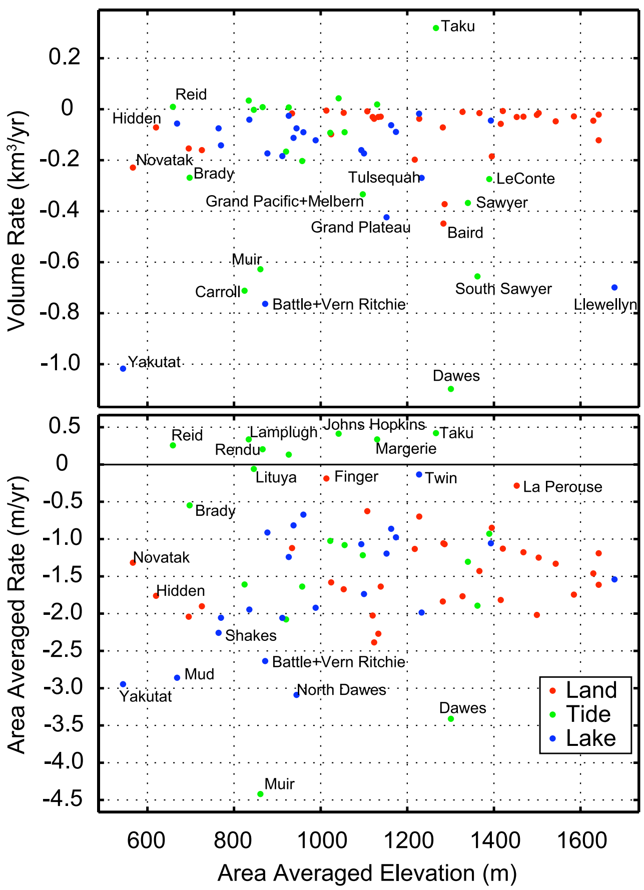

Figure 8. (top) Volume change rates versus area-averaged elevation. (bottom) Area-averaged thinning rates versus area-averaged elevation.

glaciers (Figure 9). Perhaps the most surprising result of this analysis is that lake calving glaciers are thinning faster per area than the tidewater calving glaciers of southeast Alaska, and, in particular, one lake calving glacier (Yakutat) has very nearly the largest rate of volume loss of any glacier in the region.

5. Discussion

5.1. Climate Change

[23] Climate change is commonly invoked as a factor causing the negative trend in global glacier mass balances. The surface mass balance of glaciers is largely determined by the magnitude of summer air temperatures, which represents the variability in solar radiation and sensible heat available for melting, and winter snowfall, which determines the net surface accumulation. Previous studies have shown that annual air temperatures in Alaska have increased during the past 50 years, with winter increases approximately double those occurring during the summer [*Stafford et al.*, 2000]. *Rasmussen and Conway* [2003] suggest the summer temperature increases were sufficient to explain the widespread glacier mass loss in Alaska and northwestern Canada, but little work has been done to investigate the potential effects of winter warming on the distribution and type of winter precipitation in this region. Precipitation is an important component of the mass balance in maritime regions such as southeast Alaska, where accumulation rates are as high as 4 m weq/yr [*Eisen et al.*, 2001; *Motyka et al.*, 2002; *Pelto and Miller*, 1990], compared to 1–2 m weq for Alaska interior continental glaciers [*March*, 2003; *Mayo et al.*, 2004].

[24] We obtained climate data from the National Climate Data Center for 1948 to 2000 for weather stations located in our study area and calculated daily means of temperature and precipitation for each season: annual, winter (November to April), and summer (May to October). The long-term changes in temperature and precipitation were calculated with a linear regression, and are shown in Table 1. We assume that the climate trends do not vary with elevation and represent climate trends at the glaciers, while recognizing that at times the local mountain conditions can vary substantially from those occurring at low elevations. Mean annual air temperature measured at Juneau, Sitka, and Yakutat increased from 1948 to 2000. Total precipitation (rain and snow) increased slightly at Juneau and Sitka but at Yakutat, total precipitation increased by about 1.5 m/yr. A slight increase in winter precipitation was observed at both Juneau and Sitka and a much stronger increase in Yakutat. However, the 1.8°C increase in average winter temperature would likely have driven snow lines higher in altitude.

[25] Because the vast majority of glacier covered area in our study area is below 2200 m in elevation, the effect of higher temperatures probably dominates over increased precipitation, causing negative mass balance. For those glaciers with a substantial part of their accumulation areas at high elevation, increased precipitation would translate into greater snowfall and these glaciers could benefit and thicken. For example, some high areas of the Fairweather

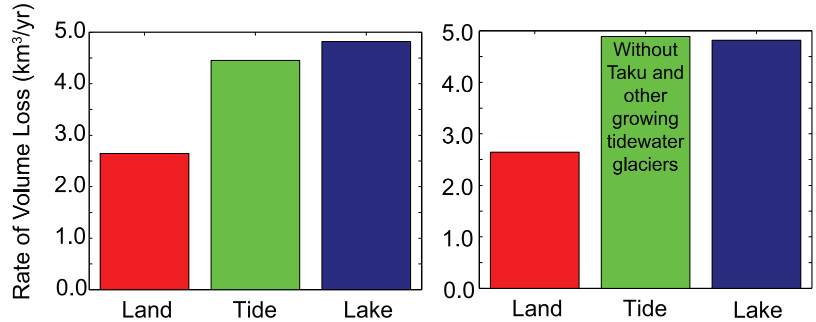

Figure 9. (a) Total contributions by glacier type of the individual glaciers analyzed. (b) Same as Figure 9a, but with the advancing tidewater glaciers removed from the tidewater sum.

Table 1. Temperature and Precipitation Change, 1948–2000: Annual, Winter, and Summer[a]

	Temperature, °C			Precipitation, mm		
Location	Summer	Winter	Annual	Summer	Winter	Annual
Juneau	1.0	1.6	1.5	81	141	223
Sitka	0.5	1.8	1.1	44	95	163
Yakutat	0.8	1.9	1.3	811	646	1528

[a]Winter is November–April, and summer is May–October.

Range indicate surface elevation increases (Figure 6). However, the results for these areas are especially subject to the greater errors associated with photogrammetric mapping of accumulation zones, and may not be significant.

5.2. Glacier Hypsometry and ELA

[26] The above comparison between glacier and climate changes assumes a static glacier surface geometry. In reality, the response of the volume of a glacier to climate is complicated because the changing surface configuration acts as a feedback which affects the response. Through feedback effects, glaciers usually attain a new equilibrium volume through terminus retreat, reducing the amount of area at low elevations where balances are most negative. An opposite, destabilizing effect occurs as the glacier surface elevation decreases. However this effect is usually secondary to those occurring due to terminus changes. The timescales over which these dynamic adjustments occur are on the order of decades, so that the volume changes of a glacier represent the integrated effects of climate during and before the period of measurement [*Elsberg et al.*, 2001]. Therefore we note that our discussion in the previous section is a first-order approximation of the effects of climate on glacier changes.

[27] Glacier surface geometry also plays a role in determining the sensitivity of glaciers to climatic changes. The equilibrium line altitude (ELA) separates the accumulation and ablation zones on a glacier, and its fluctuations relative to the glacier hypsometry determine the glacier sensitivity to climate change. For example, a glacier with broad, flat areas near the long-term ELA would be more sensitive to climatic fluctuations than one with little area near the ELA. We found the regional average ELA was between 900 to 1100 m, based on an analysis of the change in inflection of contours on the topographic maps [*Leonard and Fountain*, 2003], and an examination of available mass balance data. Assuming a summer temperature increase of 0.8°C, the ELA would increase by 100 m, if precipitation and radiation balance remained constant [*Hooke*, 1998]. Except for glaciers flowing from the Fairweather Range (which rise to 4600 m), glaciers in our study region drain mainly from relatively low mountain ranges (<2200 m) and have accumulation areas that predominately lie below 1800 m in elevation. The Brady and Yakutat Glaciers are particularly sensitive to changes in climate because their accumulation areas are near or even below the long-term ELA (Figures 3 and 6). The ice field that contains the Yakutat Glacier, which is losing ice at rates of up to 8 m/yr, mostly lies below 1000 m and has essentially lost its accumulation area. This ice field will likely disappear completely under current conditions. Thinning rates on the Brady Ice Field, which lies mostly below 1200 m, are also losing mass at a rate which is above the regional average (2 to 3 m/yr).

5.3. Glacier Dynamics

5.3.1. Tidewater Glaciers

[28] Some of the largest ice losses in southeast Alaska are occurring at tidewater glaciers that are known to have undergone calving retreats during all or part of the time periods covered in this study, e.g., LeConte, South Sawyer, Dawes, and Muir Glaciers (Figures 3, 6, and 7). These glaciers have experienced up to 640 m of thinning in their terminal reaches (Muir) and 100 m or more at higher elevations. Tidewater glaciers become unstable when the terminus retreats from its protective shoal into a deep basin and rapid calving ensues [*Post*, 1975]. Although the retreat may be triggered by climate (or other factors), these calving retreats become independent of climate as described in the "tidewater glacier cycle" [*Meier and Post*, 1987; *Post and Motyka*, 1995]. Once begun, the retreat phase of this cycle is subject to increasing positive feedbacks as surface slopes and flow velocities substantially increase at the terminus and throughout the length of such glaciers, causing significant drawdown of the parent ice field and further increasing calving flux [*Pfeffer et al.*, 2000; *O'Neel et al.*, 2001]. This phase of the tidewater cycle has been observed to lead to terminus retreat in excess of 1 km/yr in southeast Alaska. Surface mass balance data on tidewater glacier is sparse, but interpolation of available data indicates that for retreating tidewater glaciers calving is by far the dominant mode of ice loss (>90%) [*Brown et al.*, 1982; *O'Neel et al.*, 2003]. Because of the magnitude of ice loss relative to noncalving glaciers is so large, retreating tidewater glaciers tend to be the dominate contributor to sea level rise in coastal Alaska [cf. *Arendt et al.*, 2006]. The importance of this assertion is that a significant proportion of current ice loss in southeast Alaska is due to the dynamics of tidewater glaciers and not directly forced by climate change, aside from the initiation of the calving retreats.

[29] In contrast to widespread glacier wastage and retreat in southeast Alaska and adjoining Canada, seven glaciers in the region are growing. Taku Glacier, which drains from the Juneau Ice Field (Figure 7), is the most notable, having advanced over 7 km since 1890. Surface elevations have increased over 200 m in its terminus area and over 100 m at higher elevations since 1948 [*Motyka and Echelmeyer*, 2003]. Causes of the Taku Glacier advance have been addressed in several articles [*Motyka and Beget*, 1996; *Post and Motyka*, 1995; *Nolan et al.*, 1996] and are primarily related to the advance phase of the tidewater glacier cycle [*Post and Motyka*, 1995]. Briefly, when a tidewater glacier suffers a large calving retreat most of its ablation area is lost. This leads to positive imbalance, and the glacier responds with advance and growth [*Post*, 1975]. The terminus of Taku Glacier has now emerged above tidewater and it no longer calves. Other glaciers that were in advance phase of the tidewater glacier cycle in between the dates of the two DEMs analyzed here include Johns Hopkins, Reid, and Lampaugh in Glacier Bay, and North Crillon in Lituya Bay, which all continue to calve, and Lituya Glacier in Lituya Bay and Art Lewis Glacier in Nunatak Fiord, which are now grounded above sea level (Figure 6). Even for stable and advancing tidewater glaciers, calving can often be the

dominant mode of annual ice loss [*Brown et al.*, 1982; *Trabant et al.*, 1991].

5.3.2. Lake-Terminating Glaciers

[30] Many lacustrine glacier systems have also experienced significant ice losses in southeast Alaska, e.g., among the largest are Yakutat, Grand Plateau, and Alsek glaciers, whose lower reaches have thinned by 200 to 350 m (Figures 3 and 6). These three glaciers all calve icebergs into large lakes that formed as the glaciers retreated from their LIA terminal moraines into overdeepened basins. In fact, many proglacial lakes of all sizes have formed throughout the region as a result of retreat, often during the period covered by this study, and the development of these lakes thus introduces a calving component into the balance equation of these glaciers [*Viens*, 2001]. Unfortunately, except for Mendenhall Glacier, data on calving speeds and surface mass balance on lake calving glaciers in our study area are virtually nonexistent. Thus it is difficult to assess the relative importance of calving dynamics in overall ice loss for lake-terminating glaciers. Although studies have shown that calving rates for lake-terminating glaciers tend to be much lower than for their tidewater calving cousins for equivalent depths (see *van der Veen* [2002] for a review), calving losses can apparently still play a significant roll in glacier mass balance for some deep water lake terminating glaciers in Patagonia [*Venteris*, 1999; *Warren and Aniya*, 1999]. At the other extreme, calving losses at Mendenhall Glacier, a small valley glacier near Juneau, account for only 4% of the total ice losses although calving has been an important agent of glacier retreat [*Motyka et al.*, 2002; *Boyce et al.*, 2007]. In contrast to tidewater glaciers, only one lake-terminating glacier is known to be currently advancing in southeast Alaska. South Crillion Glacier, located just south of Lituya Bay in the Fairweather range, is advancing into Crillion Lake. However, South Crillion's advance is probably driven by the larger-scale tidewater dynamics of North Crillion Glacier, to which South Crillion is joined.

5.3.3. Surging Glaciers

[31] Glacier surges constitute another flow instability that can contribute to glacier changes (see *Raymond* [1987] and *Harrison and Post* [2003] for reviews) and we attribute the strong thinning observed in the terminus region of Tweedsmuir Glacier (Figures 3 and 6) to the aftermath of a surge in 1973.

5.3.4. Stranded Glaciers

[32] A subsidiary effect of calving glacier retreat is that tributary glaciers can become stranded by the retreat of the main trunk glacier and drawdown of the parent ice field. Glacier Bay, which is now surrounded by numerous discrete glaciers and small isolated ice fields, contained a huge continuous ice field with ice up to 1.5 km thick that covered more than 6000 km^2 as recently as 250 years ago [*Larsen et al.*, 2005]. The 120 km retreat and collapse of the parent LIA ice field between 1750 and 1929 AD stranded many tributary glaciers. Some were entirely isolated from any source of accumulation and are now simply wasting away (e.g., Burroughs Glacier). Other glaciers have had their accumulation areas substantially reduced as the ice fields feeding the LIA tidewater glaciers disappeared and are now susceptible to the effects described earlier as a consequence of hypsometry and rising ELA. The strong ice losses exhibited at Yakutat, Novatak and Nunatak Glaciers may in part be attributable to the demise of Nunatak Glacier during the 19th century [*Barclay et al.*, 2001]. Post-LIA retreats in Tracy Arm, Endicott Arm, and LeConte Bay also contributed to isolation and wastage of several tributary glaciers.

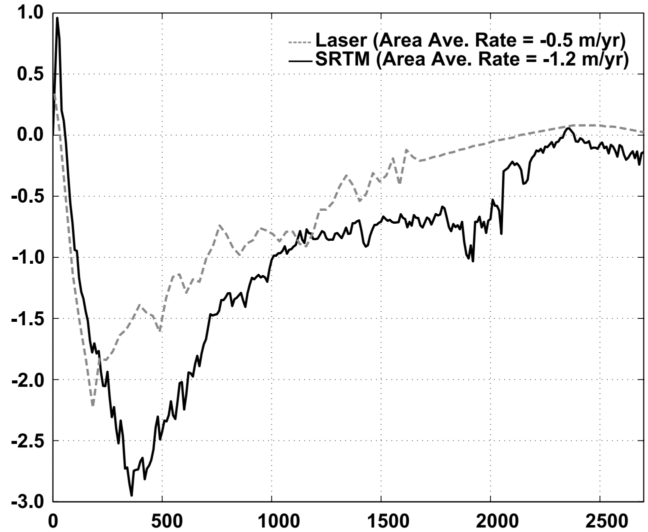

Figure 10. Comparison of DEM differencing results and laser profiling. The solid black line shows the rate of elevation change averaged in 10 m elevation bins across all of the glacier covered area analyzed here. The dashed gray laser profile curve is what was used to extrapolate from the few profiled glaciers in southeast Alaska to the remaining unprofiled glacier area [*Arendt et al.*, 2002].

5.4. Comparison of Losses to Laser Altimetry Estimates

[33] *Arendt et al.* [2002] compared small aircraft laser altimetry profiles acquired over 70 glaciers in Alaska and neighboring Canada to elevations from USGS and Canadian topographic maps. Results of the analysis were extrapolated to nonprofiled glaciers to estimate losses from all ice-covered areas in NW North America. Twelve glaciers were profiled in southeast Alaska at the time of the study. We compared our results to the results from this earlier extrapolation and found that area-averaged thinning rates for southeast Alaska in the earlier study may have been underestimated by more than a factor of two (Figure 10). We attribute this discrepancy to two factors: (1) the small number of profiled glaciers and (2) tidewater and lake calving glaciers were underrepresented in southeast Alaska at the time of 2002 study. Although the time period covered by our analysis does include both the "early" (~1950 to ~1995) and "recent" (~1995–2001) periods discussed by *Arendt et al.* [2002], and therefore should include some effect of the accelerated wastage of the recent period, this effect is not enough to account for a factor of two difference in thinning rates averaged over the whole period.

[34] *Arendt et al.* [2002] specifically removed tidewater glacier data from the composite profiles used in their extrapolation of measured rates of surface elevation change. The reasoning behind this choice is clear: one would not want to extrapolate thinning rates from rapidly disintegrating tidewater glaciers to the remaining ice coverage in a

region. However, by extrapolating thinning rates from non-tidewater glaciers to a region which does indeed have a significant percentage of tidewater (and lacustriane) glacier area, *Arendt et al.* [2002] implicitly assumed that these unmeasured calving glaciers were not making an above average contribution, individually, to the regional volume loss. In the absence of data from the calving glaciers in a region, this method of extrapolation is the only practical approach. That it appears to have underestimated the volume loss in our study area by more than a factor of two strongly emphasizes the need for greatly expanded laser altimetry over as many glaciers as possible in order to obtain an accurate assessment of glacier change. Repeated profiling of glaciers strongly affected by glacier dynamics that may be contributing disproportionately to ice loss, particularly calving glaciers, is critically important to determine present volume rates and contributions to sea level rise [e.g., *Arendt et al.*, 2006].

6. Conclusions

[35] The majority of glaciers in southeast Alaska and adjoining Canada are thinning, many of them very rapidly. The rate of volume loss is 16.7 km^3/yr, similar to that of Patagonia from 1968/1975 to 2000 [*Rignot et al.*, 2003], with both southeast Alaska and Patagonia having similar area of ice cover. We attribute this wastage to a combination of factors including climate change, calving glacier dynamics, and glacier hypsometry relative to rising ELA. The generally low elevation and geometry of glaciers and ice fields in southeast Alaska and adjoining Canada make them particularly susceptible to any climate change causing an ELA rise. Over two thirds of the ice losses are occurring at glaciers that we have identified as either tidewater or lake calving glaciers. The large losses at retreating tidewater glaciers are clearly the result of glacier dynamics. Once initiated, these calving losses are largely independent of climate change and can be an order of magnitude greater than ice losses driven solely by climate change. Generally, once climate renders a tidewater calving glacier unstable, ice losses increase dramatically. The rapid retreats of large trunk glaciers can strand tributary glaciers, in turn rendering them much more vulnerable to changing climate. The relative importance of calving at lake-terminating glaciers is more equivocal because we lack the data necessary to make this assessment. However, as examples here and also in Patagonia show, calving dynamics is likely to play an important role for lake calving glaciers that terminate in deep water.

[36] The dramatic glacier changes we have documented in southeast Alaska and adjoining Canada could serve as an analogue for the changes that Greenland may undergo as the outlet glaciers there undergo catastrophic retreat. However, unlike Greenland or Patagonia [*Rignot et al.*, 2003], several glaciers in southeast Alaska are also growing, some quite robustly. Except for a few areas high in the Fairweather Range, all of the growing glaciers in southeast Alaska are doing so in response to earlier dynamic forcings, not recent climatic conditions, and are now in the advancing stage of the tidewater glacier cycle.

[37] Glaciers in southeast Alaska and adjoining Canada contributed 0.04 ± 0.01 mm/yr to global sea level rise during the latter part of the 20th century. The massive ice wastage is helping drive the region's rapid post-LIA glacier rebound, with regional rates of isostatic uplift (up to 32 mm/yr) that are the highest presently documented [*Larsen et al.*, 2005]. Our results indicate that glacier thinning in southeast Alaska and northwest British Columbia is about double that previously reported based on sparsely distributed laser altimetry profiles [*Arendt et al.*, 2002]. This difference emphasizes the need to expand ongoing laser altimetry measurements to as many glaciers as possible in future programs in order to monitor accurately the ongoing contribution to sea level rise from these glaciers.

[38] **Acknowledgments.** This manuscript benefited greatly from reviews by Mark Meier, Robert Anderson, and an anonymous reviewer. I am especially grateful to Etienne Berthier for an informal review that led to significant corrections. Support for this work was provided by NASA grants NAG5-13760 (NRA-01-OES-05) and NNG-04GH64G (NRA-03-OES-03). Matthew Beedle, Bruce Raup, and Chris Helms, in association with the GLIMS project (Global Land Ice Measurements from Space, http://www.glims.org), helped greatly with technical details and Landsat image selection used to build the glacier outline database employed herein. Open source software from Frank Warmerdam (http://fwtools.maptools.org) was used extensively for GIS-type tasks, and in particular the GDAL utilities were used for raster manipulation and reference frame transformation (http://www.gdal.org). The Generic Mapping Tools software [*Wessel and Smith*, 1998; http://gmt.soest.hawaii.edu/) was used extensively for the volume calculations and for figures.

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