

Accelerating thinning of Kenai Peninsula glaciers, Alaska

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[1] Temperate mountain glaciers are thinning at high rates and significantly contributing to sea level rise. Due to these glaciers' remote locations, remote sensing and digital elevation models (DEMs) are an effective way to calculate their thinning rates and contribution to sea level rise. Comparisons of 1950s United States Geological Survey (USGS) and 2000 Shuttle Radar Topographic Mission (SRTM) DEMs with DEMs produced from Lidar profiles of glacier center-lines indicate thinning rates from the mid-1990s to 1999 ($-0.72 \pm 0.13 \text{ m y}^{-1}$) accelerated by a factor of 1.5 as compared with 1950 to mid-1990s ($-0.47 \pm 0.01 \text{ m y}^{-1}$) for glaciers on the Harding Icefield. Overall, comparison of USGS and SRTM DEMs indicate the Harding Icefield and Grewingk-Yalik Glacier Complex, Alaska, are thinning $-0.61 \pm 0.12 \text{ m y}^{-1}$ from 1950 to 1999. **Citation:** VanLooy, J., R. Forster, and A. Ford (2006), Accelerating thinning of Kenai Peninsula glaciers, Alaska, *Geophys. Res. Lett.*, 33, L21307, doi:10.1029/2006GL028060.

1. Introduction

[2] Mountain glaciers around the world are currently thinning at a high rate and have been estimated to be contributing $0.3\text{--}0.5 \text{ mm y}^{-1}$ of melt water to sea level rise [Meier *et al.*, 2003]. Comparatively, the Greenland and Antarctic Ice Sheets are together estimated to contribute 0.4 mm y^{-1} over this same time period prior to 2005 [Rignot and Thomas, 2002; Krabill *et al.*, 2004]. While mountain glaciers are contributing about as much melt water to sea level rise as Greenland and Antarctica, the total area of mountain glaciers is merely 4% of the Earth's total glacial area [Dyurgerov and Meier, 1997].

[3] While changes in mountain glaciers have been an important topic of study over the last several decades, much of the research must be conducted in the field. This can often be expensive and limit the researchers to only a few glaciers due to their remoteness. As a result, only ~40 out of >160,000 mountain glaciers have a continuous mass balance record over the last 20 years [Dyurgerov and Meier, 1997]. More recently studies have begun to explore the potential to observe glacial changes through active remote sensing methods. This has involved airborne profiles using laser altimetry to measure elevation changes [Arendt *et al.*, 2002; Abdalati *et al.*, 2004], and observing glacier surface elevations using the NASA/JPL spaceborne platform, Shuttle Radar Topography Mission (SRTM) [Muskett *et al.*, 2003; Rignot *et al.*, 2003]. By using remote sensing, it is

possible to observe large and inaccessible glaciated regions thereby gaining a better understanding of mountain glacier changes and contribution to sea level rise.

[4] The Kenai Peninsula is located in south-central Alaska (Figure 1). The glacial area being analyzed on the peninsula is ~2480 km² with ~1872 km² in the NE portion (Harding Icefield) and ~462 km² in the SW (Grewingk-Yalik Glacier Complex). Many of the smaller glaciers closely surrounding but not connected to either of the two icefields were also included in the analysis, which cover an area of ~146 km². The Kenai Peninsula glaciers are temperate and there are no known surging glaciers.

[5] A previous study compared USGS topographic map contours from 1950–52 with airborne lidar elevations along glacier center-lines from 1994/96 for the Harding Icefield. Results show an average surface lowering of –21 m and a volume loss of –34 km³ for the ~43 year period [Adalgeirsdottir *et al.*, 1998]. More recently a study of terminus retreat on the Harding Icefield and Grewingk-Yalik Glacier Complex found that most of the glaciers have receded significantly since 1973 [Hall *et al.*, 2005].

[6] This study revisits the Harding Icefield using the lidar data and similar techniques as Adalgeirsdottir *et al.* [1998], along with Shuttle Radar Topographic Mission DEMs to determine more recent changes of 13 individual glaciers and measure changes in glacier thinning rates. While this involves using extrapolation methods to estimate area averaged surface elevation and volume changes, an overall comparison of contiguous gridded surfaces (i.e. USGS and SRTM DEMs) was also conducted to measure changes directly for all of the glaciers in the Kenai Peninsula study area from 1950 to 1999.

2. DEM Processing

2.1. USGS and SRTM DEMs

[7] The USGS DEM of the Kenai Peninsula was created from topographic maps, which in turn were created from airphotos that were flown in 1950–52. Therefore, the DEM is only as accurate as the topographic maps. The DEM has a 2×3 arc second (latitude $\times$ longitude) posting ($30 \times 60 \text{ m}$ pixel size) at this latitude. The horizontal locations are calibrated to the 1927 North American Datum (NAD27) and the vertical elevations are calibrated to the 1929 National Geodetic Vertical Datum (NGVD29). Nominal absolute vertical accuracy of the USGS DEM is $\pm 15 \text{ m}$ [U.S. Geological Survey, 2004].

[8] In February of 2000, the SRTM obtained continuous coverage of elevation data from approximately 60° North to 57° South Latitude produced from single pass C-band (5.6 cm wavelength) interferometric SAR data [Farr and Kobrick, 2001]. The final product resulted in DEMs adjusted to World Geodetic System 1984/Earth Gravitational Model 1996 (WGS84/EGM96) Geoid elevations and

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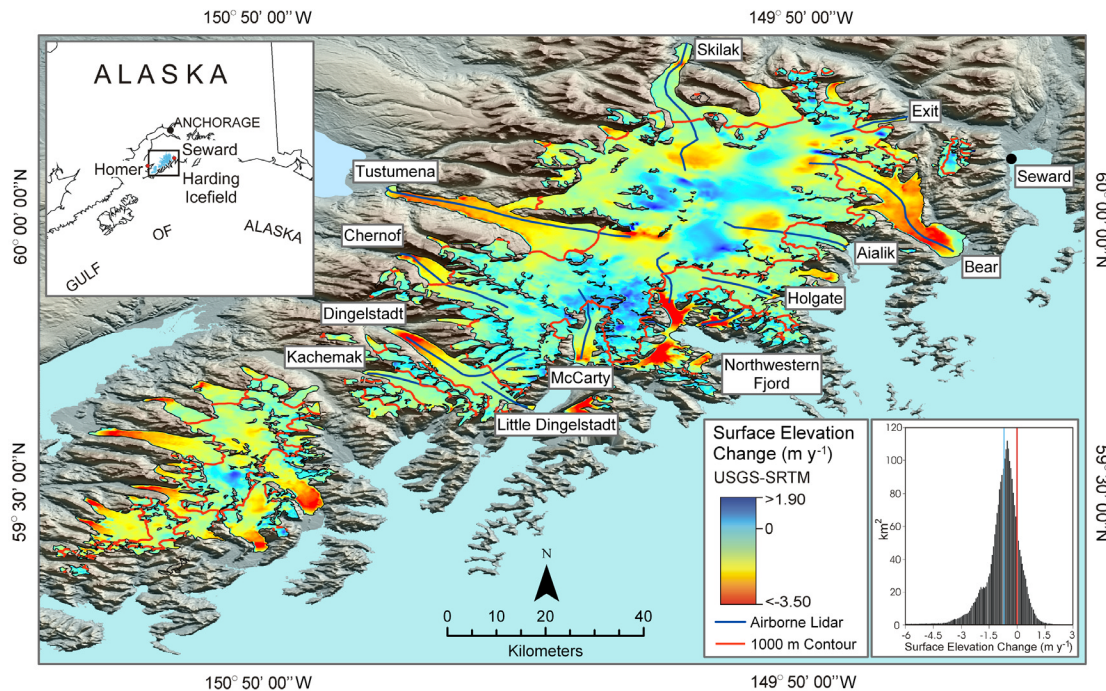


Figure 1. Surface elevation change rates for the Harding Icefield between 1950 and 1999. The histogram shows the distribution of surface elevation changes. Blue line of the histogram is mean surface elevation change (-0.61 m y^{-1}); red line separates thickening (+) and thinning (−) rates.

WGS84 horizontal datum, with a posting of 1×2 arc seconds ($15 \times 30 \text{ m}$ pixels) for the Kenai Peninsula.

2.2. Datum Transformations

[9] SRTM DEM elevations were first adjusted from WGS84/EGM96 Geoid heights to WGS84 Ellipsoid heights, and then to North American Vertical Datum 1988 (NAVD88) elevations (National Geodetic Survey, Computation of Geoid99 Geoid Height, 2005 (last accessed December 2005), http://www.ngs.noaa.gov/cgi-bin/GEOID_STUFF/geoid99_prompt1.prl). Since there are no known transformation models between NAVD88 and NGVD29 elevations for Alaska, ten benchmark elevations were sampled throughout the Kenai Peninsula to determine an estimate between the datums (National Geodetic Survey, NGS Datasheet Page, 2006 (last accessed August 2006), <http://www.ngs.noaa.gov/cgi-bin/datasheet.prl>). This resulted in finding an $\sim 2 \text{ m}$ offset (cm residuals) with NGVD29 being lower; therefore, 2 m were subtracted from the SRTM NAVD88 elevations. The SRTM DEM was also adjusted to the NAD27 horizontal datum, and resampled to the same pixel size ($30 \times 60 \text{ m}$) for consistency with the USGS DEM.

2.3. DEM Offsets and Error Analysis

[10] A preliminary subtraction was conducted between the USGS and SRTM DEMs, analyzing flat unvegetated terrain to determine the absolute vertical offset. Pixels with slopes $< 5^\circ$ and unvegetated (as determined by analyzing an intersection of USGS digital maps with 2002 Landsat ETM+ imagery) within 20 km of the icefields were used in the analysis (5819 pixels). The mean difference for these areas was $4.0 \pm 7 \text{ m}$ (1σ) with the SRTM DEM lower than

the USGS DEM, and no evident spatial pattern in the residuals. The mean difference was then added to the SRTM DEM for all further analysis. Another concern with SRTM DEMs is a significant decrease in elevation accuracy in steep terrain. This was noticeable during the offset assessment with the USGS DEM, however the vast majority of the icefields have low slopes and was not a concern. To avoid any problems with steep terrain, mountainous areas including nunataks on the icefields were masked.

[11] Along with random errors in the USGS DEMs, systematic errors created during the photogrammetric interpretation of the air photos can be significant. Contour floating due to lack of image contrast is possible in bright, flat, snow covered terrain, and is a problem in the accumulation areas of icefields. Also, others have noted problems in previous Harding Icefield USGS DEMs where elevations were significantly different from the original topographic contour maps [Adalgeirsdottir *et al.*, 1998]. However, these problems appear to be corrected as the more recent USGS DEM used in this study shows consistent elevations as compared to the original USGS topographic maps.

[12] To accurately measure glacial surface elevation changes, it is important to compare elevation data from the same seasons (or as close as possible) due to differences in elevation caused by snow accumulation and melting throughout the year. The USGS DEMs were created from topographic maps which were created from air photos obtained near the end of the ablation season. In this case, the snow line would be near its highest elevation allowing for bare ice to be exposed. At this point the end-of-summer snow could be considered firm with a denser structure than snow. Therefore, it would be important to try to determine the elevation of bare ice (ablation area) and firm (accumu-

lation area) of the end-of-summer 1999 season to compare with the 1950s DEM.

[13] Snow cover during the mid-winter (February) acquisition of the SRTM DEM has the potential to mask the true ice/firn surface elevation from the end of the 1999 summer season. However, previous C-band topographic mapping has been observed to penetrate through cold dry snow by as much as 9 ± 2 m [Rignot *et al.*, 2001], and the theoretical penetration depth is >10 m [Ulaby *et al.*, 1986]. Snow accumulation estimates during SRTM acquisition on the Malaspina Glacier, located ~ 500 km east of the Kenai Peninsula but within a similar climatic regime, show accumulation amounts of ~ 7 m at 1700 m [Muskett *et al.*, 2003], which are the highest elevations on the Kenai Peninsula icefields. Therefore, it is likely that the C-band radar signal from SRTM penetrated the dry snow cover and reached the ice or firn surface below. If the SRTM surface does include a small contribution from the winter snow pack then our subsequent calculations of ice/firn surface lowering are underestimated. With this stated, there are still significant concerns with the accuracy of SRTM in high-elevation accumulation areas. SRTM elevations are likely to be lower by as much as 10 m at these higher elevations, however the bare ice area elevations closely resemble true end-of-summer 1999 elevations, as was noted in a study of SRTM accuracy of ice surfaces of glaciers in the Alps [Berthier *et al.*, 2006].

[14] Random error estimates for the thinning rates were determined using a similar method as Rignot *et al.* [2003]. The $\pm 1\sigma$ of the USGS and SRTM DEM offsets (7 m) are assumed to be random. Since each icefield USGS contour level is an independent sample point, the error in the elevation change is determined by dividing the 7 m offset by the square root of the number of contour intervals on the icefield (53) from USGS maps. The resulting error bound is ± 1.0 m. Systematic errors such as those due to contour floating, were estimated for the overall (USGS-SRTM) difference at elevations above 1200 m, where late summer snow cover is prevalent. To estimate this error, 15 m (absolute USGS vertical accuracy) were added to the USGS DEM elevations above 1200 m, and a new volume change was calculated. The difference between the original and error estimated volume changes was considered the systematic error. The overall error was calculated by taking the quadrature sum of the random and systematic errors (noted in section 3.2).

2.4. Glacial Area Determination

[15] The outline for the entire glacial study area was determined from 1950s USGS topographic maps. This works well for the overall volume change analysis, however for the calculation of the surface elevation rate changes, time dependent glacial extents were needed due to changes in the termini. First, the outlines provided by Echelmeyer *et al.* [2002] were used mostly to provide a glacial boundary (for the 13 glaciers) in the accumulation area, as these data contained termini that are to the 1950s extent. These glacial boundaries have been updated since the results of Adalgeirsdottir *et al.* [1998] were published (A. Arendt, personal communication, 2006). Therefore, due to substantial differences in the process of determining several glacier boundaries, the results from this study vary from the

original. Next, new glacier boundaries for all areas were created for the more recent time period (mid-1990s) from Landsat TM (1986) and ETM+ (2002) imagery where the average outline of the glaciers between the two years was determined. This provided an approximate mid-1990s boundary, and due to the assumption of little change in area between mid-1990s and 1999, a third glacial area for 1999 was not created.

3. Ice Surface Elevation Changes

3.1. Rates of Change for Selected Harding Icefield Glaciers: 1950s to Mid-1990s and Mid-1990s to 1999

[16] To investigate rates of surface elevation change, airborne Lidar profiles that were flown over 13 Harding Icefield glaciers in 1994/96 (Figure 1) were compared with USGS (1950s) and SRTM (1999) DEMs. This produced two different surface change rates; the first from 1950–52 to 1994/96, and a more recent rate from 1994/96 to 1999. The 1994/96 Lidar data obtained from the National Snow and Ice Data Center [Echelmeyer *et al.*, 2002] were collected in WGS84 heights above the ellipsoid with a vertical accuracy of 0.30–0.35 m, and were adjusted to NAD27 horizontal datum and NAVD88 vertical datum, as well as compensating late spring snow depths which were on the order of 1–5 m mostly at higher elevations [Adalgeirsdottir *et al.*, 1998]. The Lidar data was also adjusted from NAVD88 to NGVD29 by subtracting 2 m from the elevations.

[17] The methods for calculating area averaged surface elevation changes resemble those given by Adalgeirsdottir *et al.* [1998]. This was done by dividing the glaciers into sections bounded by the contour lines 50 feet (15.2 m) up and down glacier from an individual Lidar point, (Lidar points used in the original study were located along a profile at every 100 foot (30.5 m) USGS contour crossing). Each contour bounded section was then assigned the elevation of the Lidar point centered in that section. DEMs of the ablation areas for the 13 glaciers were created to represent each time period (i.e., USGS 1950s, Lidar mid-1990s, and SRTM 1999). The surface lowering rates were determined by dividing the volume change of each glacier by the temporally averaged glacial area for each time period; (the volume change for the mid-1990s to 1999 was divided by the glacier area of mid-1990s due to little change in the glacial area). Elevation change was only computed below 1000 m or 1200 m due to the unreliable nature of SRTM at higher elevations [Berthier *et al.*, 2006]. Above this threshold, all elevation changes were assumed to be zero due to little change for each glacier, as shown by Adalgeirsdottir *et al.* [1998].

[18] Surface elevation changes in the ablation area were compared for the 13 glaciers on the Harding Icefield (Table 1). While the Lidar/DEM comparisons do not produce a complete understanding of the surface elevation changes for the entire icefield, they do show important trends of changes since the mid-1990s. The average elevation change between the 1950s USGS DEMs and the 1994/96 Lidar produced DEMs for all 13 glaciers was -0.47 ± 0.01 m y^{-1} (Table 1). For the same 13 glaciers a comparison of the Lidar produced DEMs and SRTM DEMs results in an average elevation change of -0.72 ± 0.13 m y^{-1} , between 1994/96 and 1999. It's important to note that 12 of the 13 glaciers

Table 1. Area Averaged Glacier Surface Elevation Change Rates Below 1000 m, and Glacier Areas^a

Glacier	ΔZ , m y ⁻¹		A , km ²	
	1950–1994/96	1994/96–1999	1950s	Mid-1990s
Exit	-0.10 ± 0.01	-0.25 ± 0.12	40.5	39.5
Bear	-0.80 ± 0.01	-1.01 ± 0.10	227.4	217.9
Aialik	$+0.002 \pm 0.01$	-0.51 ± 0.10	87.1	87.0
Holgate	-0.23 ± 0.01	-1.37 ± 0.12	67.4	67.3
Northeastern	-2.33 ± 0.01	-2.26 ± 0.20	14.2	11.5
Northwestern ^b	-0.94 ± 0.01	-1.03 ± 0.21	113.2	100.4
McCarty	-0.15 ± 0.01	-0.35 ± 0.15	124.6	119.4
Little Dingelstadt	-0.20 ± 0.02	-0.27 ± 0.09	34.8	34.5
Skilak	-0.13 ± 0.01	-0.16 ± 0.12	202.3	196.1
Tustumena	-0.43 ± 0.01	-0.52 ± 0.09	300.5	298.9
Chernof ^b	-0.67 ± 0.02	-1.11 ± 0.19	63.5	56.5
Dingelstadt ^b	-0.89 ± 0.01	-1.89 ± 0.18	72.7	67.2
Kachemak	-0.39 ± 0.01	-0.84 ± 0.13	23.8	22.8
Total Average	-0.47 ± 0.01	-0.72 ± 0.13		

^aError was calculated using the method from section 2.3. The $\pm 1 \sigma$ of the offsets were 3.6 m for 1950–1994/96, and 3.3 m for 1994/96–1999. ΔZ is change in surface elevation in m y⁻¹; A is glacier area in km²; (+) indicates thickening, (–) indicates thinning.

^bCalculation determined for area below 1200 m.

increased their thinning rates, with the average thinning rate increasing by 1.5 times in the more recent time period. Six glaciers more than doubled their thinning rates (Table 1).

3.2. Mean Lowering for All Kenai Peninsula Glaciers: 1950s to 1999

[19] In addition to the lowering rate of the ablation area for the 13 glaciers, the volume change of the entire glacial study area of the Kenai Peninsula was estimated. The USGS and SRTM DEM subtraction indicated that from 1950–52 to 1999 there was an area averaged surface elevation lowering of -30.0 ± 6.0 m (-0.61 ± 0.12 m y⁻¹) (Figure 1). The glacial volume change from 1950–52 to 1999 was -72.1 ± 15.0 km³ (-1.47 ± 0.30 km³ y⁻¹). While difficulties occur in calculating melt water equivalence due to differences in density of snow, firn, and ice, the majority of the volume change (70 percent) of the Kenai Peninsula glaciers occurred in the ablation zone below 1000 m (Figure 1). Therefore, estimated water equivalence was calculated using a combined ice density ratio of 0.81 g/cm³, (0.9 g/cm³ for bare ice below 1000 m and 0.6 g/cm³ for firn above 1000 m) resulting in -58.4 ± 12.1 km³ of potential melt water (-1.19 ± 0.24 km³ y⁻¹), which equates to 0.0033 ± 0.0006 mm y⁻¹ of sea level rise. These rates were calculated by dividing by the dominant time period (49 years) since the majority of the photos for the USGS DEM were taken in 1950 and the last ablation season for the SRTM DEM was fall 1999.

[20] The tidewater glaciers of the Northwestern Fjord appear to have experienced the greatest amount of thinning (Figure 1), likely due to rapid retreat caused by tidewater effects. During tidewater glacier retreat, calving rates rapidly increase due to the association of deep water at the termini, and back pressure is reduced up-glacier allowing glacier thinning and stretching, leading to increased calving and retreat [Meier and Post, 1987]. Therefore, the rapid thinning of the tidewater glaciers is more likely due to tidewater effects than other influences, such as climate. While thinning dominates the changes on the icefields, apparent thickening has occurred at higher elevations (greater

than ~ 1300 m) (Figure 1). However, it is possible that the magnitude of this change is in question due to problems associated with SRTM data at the higher accumulation areas (discussed in section 2.3).

3.3. Comparison With Other Mountain Glaciers

[21] The average thinning rate of Malaspina Glacier (~ 5000 km²) between 1972/73 and 2000 was -1.7 ± 0.19 m y⁻¹ [Muskett et al., 2003], whereas for the Patagonian Icefields of South America ($\sim 17,200$ km²) from 1968/75 to 2000 the rate was -1.0 ± 0.1 m y⁻¹ [Rignot et al., 2003]. Other mountain glaciers of the world have also been thinning at rapid rates since 1990, such as Vernagt (-0.87 m y⁻¹) in the Alps and Place (-1.4 m y⁻¹) in SW British Columbia [Dyrgerov and Meier, 2004]. While these glaciers' thinning rates are comparable to the Kenai Peninsula glaciers for similar time periods, compared to the global average mountain glacier thinning rate (-0.13 m y⁻¹) from 1960 to mid-1990s [Dyrgerov and Meier, 2004] the Kenai Peninsula glaciers are rapidly thinning.

4. Meteorological Data Analysis

[22] To determine the potential causes of the Kenai Peninsula glacier thinning, analysis of annual mean temperature, and total snowfall (not shown) were conducted for data from the two closest meteorological stations, Seward and Homer, (Figures 1 and 2), (Desert Research Institute, Alaska Climate Summaries, 2005 (last accessed June 2006), <http://www.wrcc.dri.edu/summary/climsmak.html>) from 1950 to 2000. Periods of negative air temperature anomalies occurred mostly before the mid-1970s for both stations. The anomalies have been consistently positive since the mid-1970s for Homer, but at Seward mean annual temperatures began to more closely match the 50 year mean from the late 1980s through 1996 (Figure 2). This change from mostly

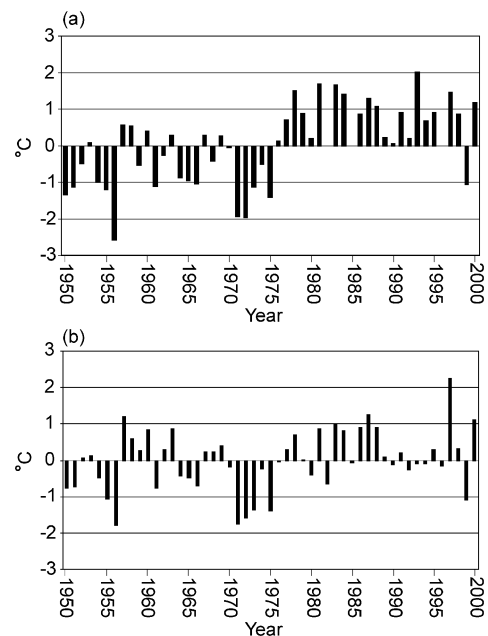


Figure 2. Annual mean temperature anomalies from the 1950–2000 mean for (a) Homer, and (b) Seward. For locations see inset in Figure 1.

negative to mostly positive anomalies is consistent with a shift in the interdecadal Pacific signal in 1976/77 related to increases in Pacific Ocean sea surface temperatures and atmospheric circulation in the Northern Hemisphere [Frauenfeld *et al.*, 2005]. Seasonal temperature anomalies were also computed to determine if the warming trend could be responsible for increased glacier thinning. These anomalies indicate much of the annual warming occurred during the winter when temperatures continued to be below 0°C, in particular at higher elevations. This suggests the causes of the glacial thinning are more complex than simple temperature increases.

[23] Both stations are missing snowfall data, especially throughout the 1990s. Despite this, there is a trend of decreasing snowfall amounts since the mid-1970s for the Seward station, while at Homer amounts have generally been without a trend. Overall, precipitation changes for this region are not statistically significant [Stafford *et al.*, 2000], and therefore difficult to analyze in relationship with the glacial changes discussed.

[24] The slower thinning rates for 1950 to 1994/96 emphasize the influence of the cooler annual temperatures between 1950 and 1975 (Figure 2). It is likely that most of the thinning that occurred from 1950 to 1994/96 began after 1975, and increased during the 1990s as noted with other glaciers in the region [Rasmussen and Conway, 2004], as well as the 13 Harding Icefield glaciers. Despite the relationship of increased temperatures and glacial thinning, it is difficult to fully conclude the reasons for the thinning of Kenai Peninsula glaciers due to lack of sufficient meteorological data, especially at higher elevations.

5. Summary

[25] Despite temperate mountain glaciers of the world having much smaller ice masses than Greenland and Antarctica their rapid wasting is a significant concern for imminent sea level rise. The average of 13 glaciers on the Harding Icefield increased their thinning rates by 1.5 times since the mid-1990s ($-0.72 \pm 0.13 \text{ m y}^{-1}$) as compared to the time period 1950 to 1994/96 ($-0.47 \pm 0.01 \text{ m y}^{-1}$). The volume and surface elevation changes of the Kenai Peninsula glaciers (1950 to 1999) was $-72.1 \pm 15.0 \text{ km}^3$ of glacial mass with a thinning rate of $-0.61 \pm 0.12 \text{ m y}^{-1}$. This is equivalent to $-1.19 \pm 0.24 \text{ km}^3 \text{ y}^{-1}$ of melt water, which is as much as one percent when compared to the total estimate of mountain glacier contribution to sea level rise since the 1950s [Meier *et al.*, 2003]. This is a large amount of melt water considering the relatively small size of these glaciers, which highlights the pressing need for more up-to-date and accurate monitoring of the Kenai Peninsula glaciers and other “small” mountain glaciers of the world to better understand their response to climate change and potential influence on sea level rise.

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