Elevation changes of ice caps in the Canadian Arctic Archipelago

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[1] Precise repeat airborne laser surveys were conducted over the major ice caps in the Canadian Arctic Archipelago in the spring of 1995 and 2000 in order to measure elevation changes in the region. Our measurements reveal thinning at lower elevations (below 1600 m) on most of the ice caps and glaciers but either very little change or thickening at higher elevations in the ice cap accumulation zones. Recent increases in precipitation in the area can account for the slight thickening where it was observed but not for the thinning at lower elevations. For the northern ice caps on the Queen Elizabeth Islands, thinning was generally <0.5 m yr^{-1}, which is consistent with what would be expected from the warm temperature anomalies in the region for the 5 year period between surveys, and appears to be a continuation of a trend that began in the mid-1980s. Farther south, however, on the Barnes and Penny ice caps on Baffin Island, this thinning was much more pronounced at over 1 m yr^{-1} in the lower elevations. Here temperature anomalies were very small, and the thinning at low elevations far exceeds any associated enhanced ablation. The observations on Barnes, and perhaps Penny, are consistent with the idea that the observed thinning is part of a much longer term deglaciation, as has been previously suggested for Barnes ice cap. On the basis of the regional relationships between elevation and elevation change in our data, the 1995–2000 mass balance for the archipelago is estimated to be -25 km^3 yr^{-1} of ice, which corresponds to a sea level increase of 0.064 mm yr^{-1}. This places it among the more significant sources of eustatic sea level rise, though not as substantial as the Greenland ice sheet, Alaskan glaciers, or the Patagonian ice fields. *INDEX TERMS:* 1640 Global Change: Remote sensing; 1699 Global Change: General or miscellaneous; 1827 Hydrology: Glaciology (1863); 1863 Hydrology: Snow and ice (1827); *KEYWORDS:* ice caps, Arctic, mass balance

1. Introduction

[2] Glaciers and ice caps outside of Greenland and Antarctica contain enough water to raise sea level by an estimated 0.5 m [*Church et al.*, 2001]. Although their long-term potential contribution to sea level is far less than that of Greenland or Antarctica, their smaller size and more temperate characteristics may make their near-term contributions, estimated at up to 20 cm in the next 100 years [*Church et al.*, 2001], more substantial. Among these ice masses, the Canadian ice caps are of particular importance primarily because they are situated in the Arctic, where the regional-scale positive albedo feedbacks and interactions between ocean, atmosphere, and sea ice are expected to amplify the effects of climate change [*Cubasch et al.*, 2001] and also because, with an area of 152,000 km^2 [*Ommanney*, 2002], they are among the largest of the Arctic glaciers and ice caps. In light of recent observations of the Greenland ice sheet indicating significant thinning in the warmer regions below 2000 m [*Krabill et al.*, 2000] and studies showing significant wastage of Alaskan glaciers [*Arendt et al.*, 2002], understanding the mass balance of other Arctic ice masses is a matter of increasing interest.

[3] In an effort to assess the current mass balance of the major Canadian ice caps, we conducted precise airborne laser surveys in spring of 1995 using NASA's Airborne Topographic Mapper (ATM) on board the NASA P-3 aircraft and repeated these measurements again in spring of 2000 on a commercial Twin Otter platform. The locations of these survey lines are shown in Figure 1. The survey trajectories were designed to cross the broadest and longest portions of the major ice caps, and where possible, they were planned to cover some of the more significant outlet glaciers.

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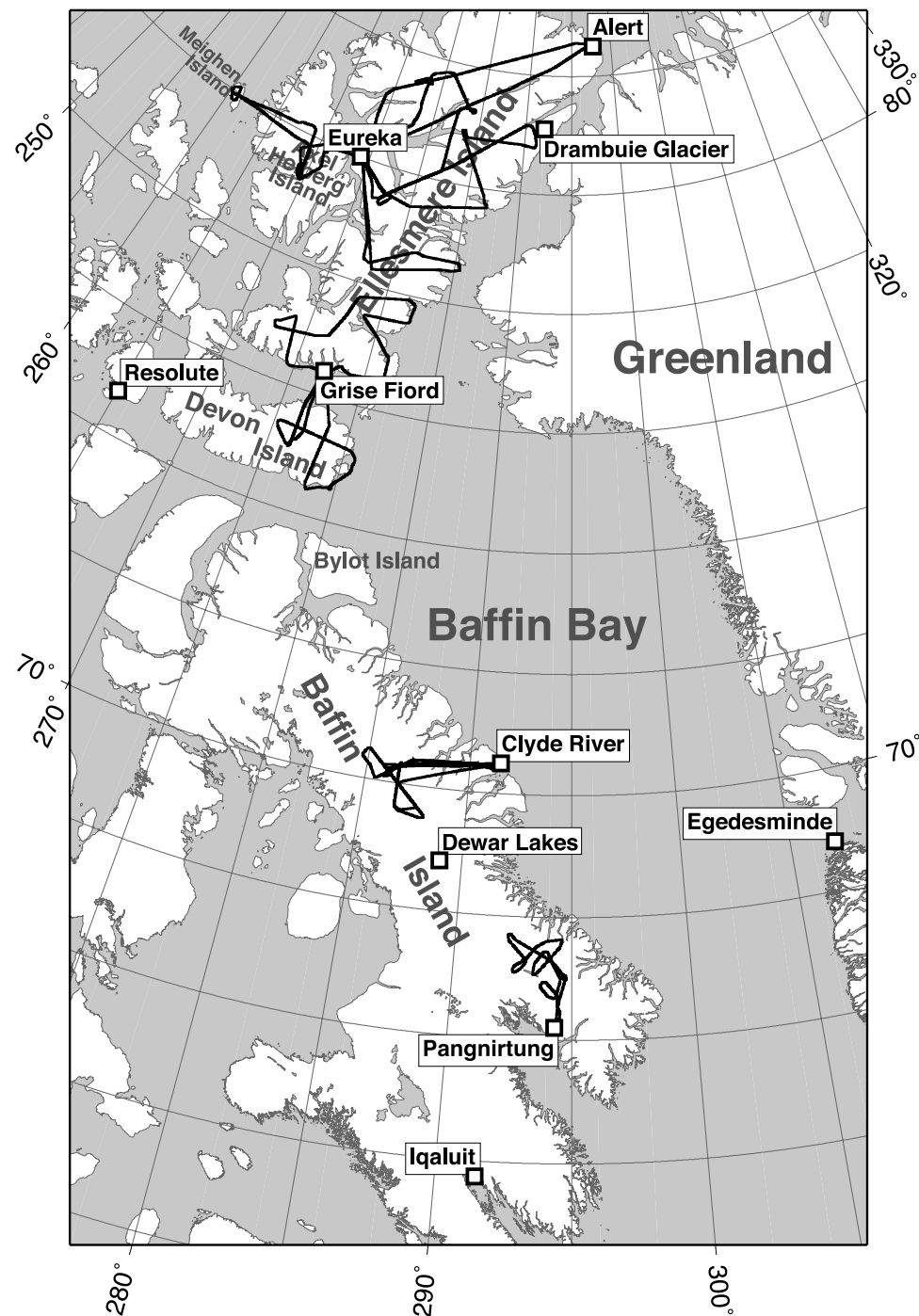
Figure 1. Location map of the 2000 flight lines (repeat surveys of the 1995 lines) in the Canadian Arctic Islands. Flights were conducted out of Pangnirtung, Clyde River, Grise Fiord, and Eureka. Weather station data used in this analysis were from Eureka, Alert, Resolute, Clyde River, Iqaluit, Egedesminde, and Dewar Lakes.

[4] The ATM combines precise ranging capability with Global Positioning System (GPS) techniques to retrieve surface elevation to a root-mean-square error of 10 cm or better [*Krabill et al.*, 1995, 2002]. As such, it provides a valuable tool for measuring changes in the ice cap surface elevations by means of repeat surveys when there is adequate time separation between measurements. The Canadian ice cap campaigns, like those in Greenland, were

conducted over a 5 year time interval in an effort to minimize the effects of year-to-year variability.

[5] The surveys are designed to provide a spatially broad assessment of ice cap thickening and thinning rates over the 5 year time interval. These complement the in situ measurements of accumulation and mass balance that have been ongoing as part of the Canadian glaciology program [*Koerner*, 2002]. These in situ observations, while local in

nature, span a period of more than 4 decades in some cases and provide a historical context for some specific locations.

2. Method

[6] The surveys were carried out in the spring of 1995 and 2000 (late May to early June), just prior to the start of the ablation season, in an attempt to minimize seasonal effects. The 1995 surveys were conducted with the first-generation ATM (ATM-1) flown on a P3-B Orion aircraft. A 250 μJ laser pulse was directed toward the ground by a nutating mirror angled at 10°. The mirror scan rate was 10 Hz, and the laser pulse repetition frequency (PRF) was 1 kHz. This combination of parameters results in an elliptical swath of measurements below the aircraft of roughly 140 m width that is generally centered beneath the aircraft. At a cruising altitude of ~450 m above ground level and a speed of nearly 300 kts (150 m s^{-1}), the spacing between laser shots is roughly 5 m on average, and the size of the spot illuminated by the laser is on the order of half a meter. The 2000 surveys were conducted on a Twin Otter aircraft with a smaller, lower-power version of the ATM (ATM-2). ATM-2, with a laser pulse energy of 125 μJ, operated at a scan rate of 20 Hz, with a 15° scan angle and a PRF of 3 kHz. The Twin Otter aircraft was also flown at the same altitude but at a speed of roughly 130 kts (65 m s^{-1}). The greater scan angle, higher PRF, and slower aircraft speed resulted in a wider swath of close to 250 m and a denser sampling, with shots spaced ~2 m on average.

[7] Using advanced GPS navigation techniques [*Wright and Swift*, 1996], the original survey trajectories were reflown, generally to within a few tens of meters across-track. Such precise repeatability on the second flight was achieved by feeding real-time GPS data to a computer display in the aircraft cockpit, which simultaneously displayed the survey swath from 5 years earlier. A software package developed at NASA's Wallops Flight Facility took input from the GPS data stream during the 2000 flights, calculated the ground locations of the laser scans, and overlaid the laser scan pattern as it was being collected onto a map containing the 1995 scan locations. The pilots would then steer the plane in such a way as to maximize overlap of the current swath with the previous swath. This served to maximize repeat coverage, and it also permitted the pilot to immediately correct any deviations from the previous trajectory.

[8] Change estimates were made by comparing each individual laser pulse surface return from the 2000 campaign with the returns from the 1995 season located within a 1 m horizontal search radius. With the 5 year time separation, elevation changes in excess of 2 cm yr^{-1} are considered significant [*Krabill et al.*, 1999]. In some instances, low clouds prevented the retrieval of surface elevations during parts of the flights. This was especially true for the eastern part of Devon ice cap, which was cloudy during the 1995 surveys, and for the western part of Meighen ice cap, which was cloudy during the 2000 surveys. For the most part, however, more than 90% of the surveys yielded useful data for comparison.

3. Results

[9] The annual elevation changes (*dh/dt*) for the period 1995–2000 are shown in Figures 2 and 3. Thinning is more

evident on the warmer southern ice caps on Baffin Island, Penny, and Barnes than on the more northerly ones. The general character of the elevation changes is such that thinning is evident at the lower elevations near the ice cap edges, while in the higher, more central regions, diminished thinning rates, or thickening, is observed.

[10] There are exceptions, however, the most obvious of which are the regions in the Prince of Wales ice field that are below 950 m. Here there is wide scatter in the *dh/dt* versus elevation plots (Figure 3), and no real relationship with elevation is evident. Above 950 m, however, thinning tends to be greatest at the lower elevations, ranging from 25 cm yr^{-1} in the 950–1000 m elevation band and diminishing to roughly zero above 1500 m. Similarly, Northern Ellesmere Island shows some areas of low-elevation thickening on the northeastern side near Alert. These areas are the exceptions, however.

[11] While qualitatively, most of the individual ice caps or groups of ice caps in a specific region exhibit a clear inverse relationship between elevation and thinning or thickening rate, a strongly quantitative relationship cannot easily be extracted. This is clearly evident in the data from the flight segments over the Barnes ice cap (Figure 4), which shows the most pronounced thinning. The Barnes ice cap is fairly smooth with well-defined margins, is almost axially symmetric, and is landlocked around most of its perimeter. The surrounding land minimizes complexities associated with marine boundaries, such as we see in Greenland. Still, even this apparently "simple" ice cap shows spatial variability in its behavior, particularly on its southwest side [*Andrews et al.*, 2002], with evidence of past surges [*Holdsworth*, 1977] or "local creep slumps," in which creep and basal sliding are enhanced by the accumulation of a bottom layer of meltwater in the warmer central parts of the ice cap [*Shoemaker*, 1981]. There is also some level of iceberg calving in the small (5- to 10-km-wide) surrounding lakes.

[12] Data from the different Barnes flight segments reveal variable relationships between *dh/dt* and elevation, which can be seen in Figure 4b as three distinct linear trends that converge at higher elevations, and each of these can be mapped to a segment of flight line in Figure 4a. The topmost linear feature (A) in Figure 4b (which corresponds to segment A in Figure 4a) shows almost no correlation between *dh/dt* and elevation. The northwestern section of the flight along the major ice cap axis (segment B) shows the strongest relationship between *dh/dt* and elevation of approximately 25 cm yr^{-1} per 100 m of elevation change. The scattering of points bound by these two extremes represents the elevation dependence of most of the remaining segments.

[13] These varying dependencies are not just unique to Barnes but are evident in many of the regions surveyed, as indicated by the nature of the scatter at low elevations in Figure 3. One reason is the varied topography found in some of the more mountainous regions, which strongly influences atmospheric circulation characteristics and creates highly localized effects. This is especially the case for accumulation, which has a strong orographic dependence, but ablation rates can also be affected as the latent heat flux is a direct function of not just temperatures but surface winds and humidity. A clear relationship may exist on one
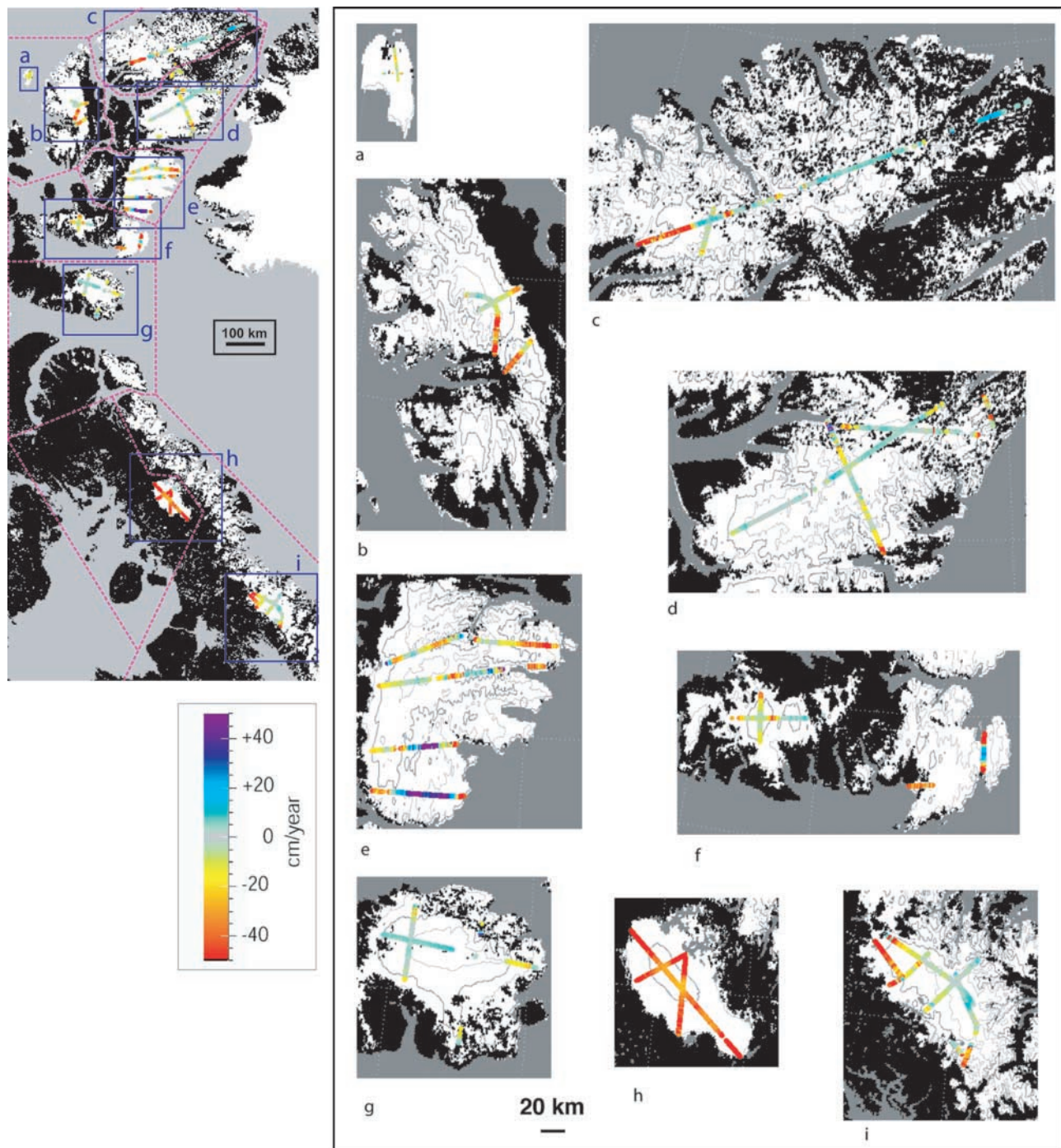
Figure 2. Rates of elevation change (*dh/dt*) along flight lines. The map on which the changes are overlaid shows ice (white with gray contour lines), land (black), and water (gray) and was derived from the U.S. Geological Survey Global Land Cover Characteristics Data Base Version 2.0 (http://edcdaac.usgs.gov/glcc/globe_int.asp). Also shown in dotted magenta lines on the leftmost figure are the regions referred to later in the text for extrapolation of measured *dh/dt* for overall mass balance calculations. Individual ice caps are shown in the expanded figures on the right for (a) Meighen Island, (b) Axel Heiberg, (c) Northern Ellesmere Island, (d) Agassiz ice cap, (e) Prince of Wales ice field, (f) Southern Ellesmere Island, (g) Devon Island, (h) Barnes ice cap, and (i) Penny ice cap.

side of an ice cap, with a clear but different relationship on another side. As a result, *dh/dt* can vary widely for a given elevation, even on an individual ice cap. Also contributing to the scatter in Figure 3 is the fact that many of the flight lines in a region are over several small ice caps and glaciers rather than one large one. Each is subject to its own unique local climate and dynamic conditions.

[14] In the absence of a consistent relationship between *dh/dt* and elevation, even for a relatively simple ice cap such as Barnes, it is difficult to extrapolate these results
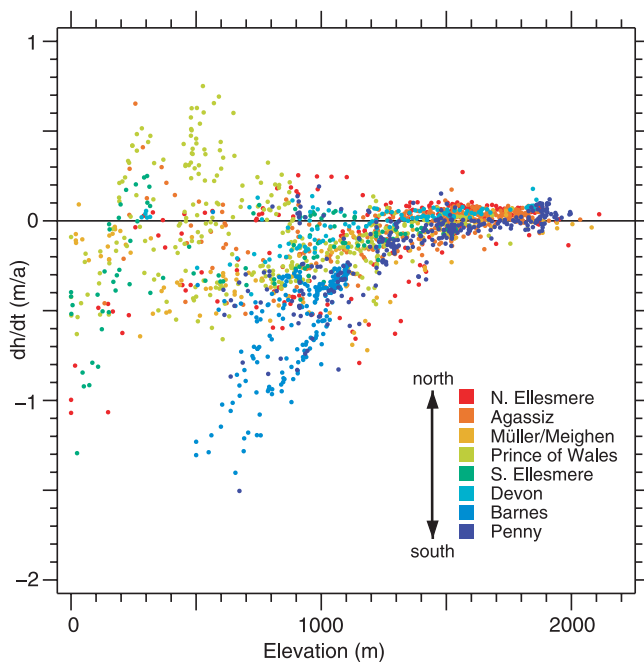
Figure 3. Elevation changes as a function of surface elevation for regions a–h in Figure 2. With some exceptions (most notably, Prince of Wales), thinning is evident in most areas and is most pronounced at the lower elevations. At high elevations, there is consistently negligible thinning, or, in some cases, thickening, for all of the ice caps surveyed. There is also some indication of a zonal dependence, with the more southerly ice caps showing greater thinning rates than the more northerly ones.

to a highly accurate mass balance estimate for the ice caps as a whole. They do, however, offer strong indications of the overall state of ice cap balance. As was observed in Greenland [*Krabill et al.*, 2000], the lower elevations generally show significant thinning by >1 m yr^{-1} in some cases, while the higher elevations are either thinning less or thickening slightly.

[15] These results are consistent with field data on Devon and Agassiz ice caps, which show a zero or slightly positive balance anomaly at high elevations but a negative anomaly at lower elevations (unpublished update of *Koerner* [1996]). The ATM data over Meighen ice cap, an ~90 km^2 stagnant ice cap that is only about 120 m thick, show a widespread thinning of between 0.1 and 0.2 m yr^{-1}, which agrees with the estimated 15 cm yr^{-1} surface lowering from the in situ observations for the same period.

4. Discussion

### 4.1. Climatological Interpretation

[16] In order to interpret the observed elevation changes, we must consider the impacts of recent climate conditions on ice cap accumulation and ablation. Unfortunately, measurements are sparse in the area, but a 52 year (1948–2000) precipitation record from several stations in Nunavut is available in the Historical Canadian Climate Data Set. Monthly data were compiled from Clyde River, Eureka, Iqaluit, and Alert (shown in Figure 1), and their deviations

from the climatological mean are reported in the data set (R. Whitewood, personal communication, 2001). Using these data, *Zhang et al.* [2000] reported an increasing trend in annual precipitation in the latter half of the 20th century along with an increase in the fraction of precipitation that falls as snow. Together, these would suggest an increasing trend in accumulation on the ice caps. Of more direct relevance to our study, however, is the nature of the anomaly during the period of our survey, 1995–2000.

[17] Analysis of the 52 year record from these four stations indicates that for the region as a whole (averaging these four stations), accumulation was ~15% higher during the 1995–2000 time period than for the 1951–1980 climatological mean to which these data are historically compared. This is true whether summer precipitation, which may fall as rain rather than snow, is included in the calculations or not. Moreover, the trend of increasing fraction of precipitation that falls as snow [*Zhang et al.*, 2000] suggests an accumulation increase. With an average precipitation rate of ~220 mm yr^{-1} in the region (R. Whitewood, personal communication, 2001), this 15% precipitation anomaly is 33 mm yr^{-1}. If we assume a snow density of 333 kg m^{-3}, this translates to roughly 10 cm yr^{-1} in surface elevation changes.

[18] We recognize that these precipitation anomalies are only very crudely representative, in a relative sense, of conditions on the ice caps. The rough terrain lends itself to spatially variable and highly localized climate conditions, which make point measurements (fraught with their own uncertainties) difficult to extrapolate. However, for relative assessment of year-to-year variability, they do provide useful insight. Moreover, the anomalies are clearly positive, consistent with the computed trends and opposite in sign to what would be needed to explain the observed thinning at the lower elevations. We therefore conclude that the thinning observed on most of the ice caps is not a result of reduced accumulation. At the higher elevations (>1600 m), where ablation rates are small, there is slight thickening at an average of about 5 cm yr^{-1}. This is qualitatively consistent with the accumulation anomaly but smaller in magnitude, suggesting either an overestimate in the accumulation anomaly, or a reduction of the thickening, by the mechanisms responsible for thinning elsewhere on the ice caps.

[19] A similarly long temperature record is available from the Goddard Institute for Space Studies (GISS) at http://www.giss.nasa.gov/data/update/gistemp/station_data. These records, compiled by the National Climate Data Center, reveal some interesting temperature characteristics in the region based on data from Iqaluit, Clyde River, and Eureka. Analysis of the 52 year record of coastal station temperature data at locations in Figure 1 (Table 1) shows generally warm summer (June, July, and August) temperature anomalies in 1995–2000 when compared to that of the entire record. Clyde River was an exception, however, showing no appreciable deviation from the mean. These anomalies were greater for annual temperatures, and in this case, Clyde River showed a negative (cool) anomaly. These characteristics are consistent with the warming trends reported for 1948–1998 [*Zhang et al.*, 2000] and 1971–2000 [*Vorosmarty et al.*, 2001]. The latter was based on an update of *Chapman and Walsh* [1993]. While the summer anomaly is of primary interest because it most directly
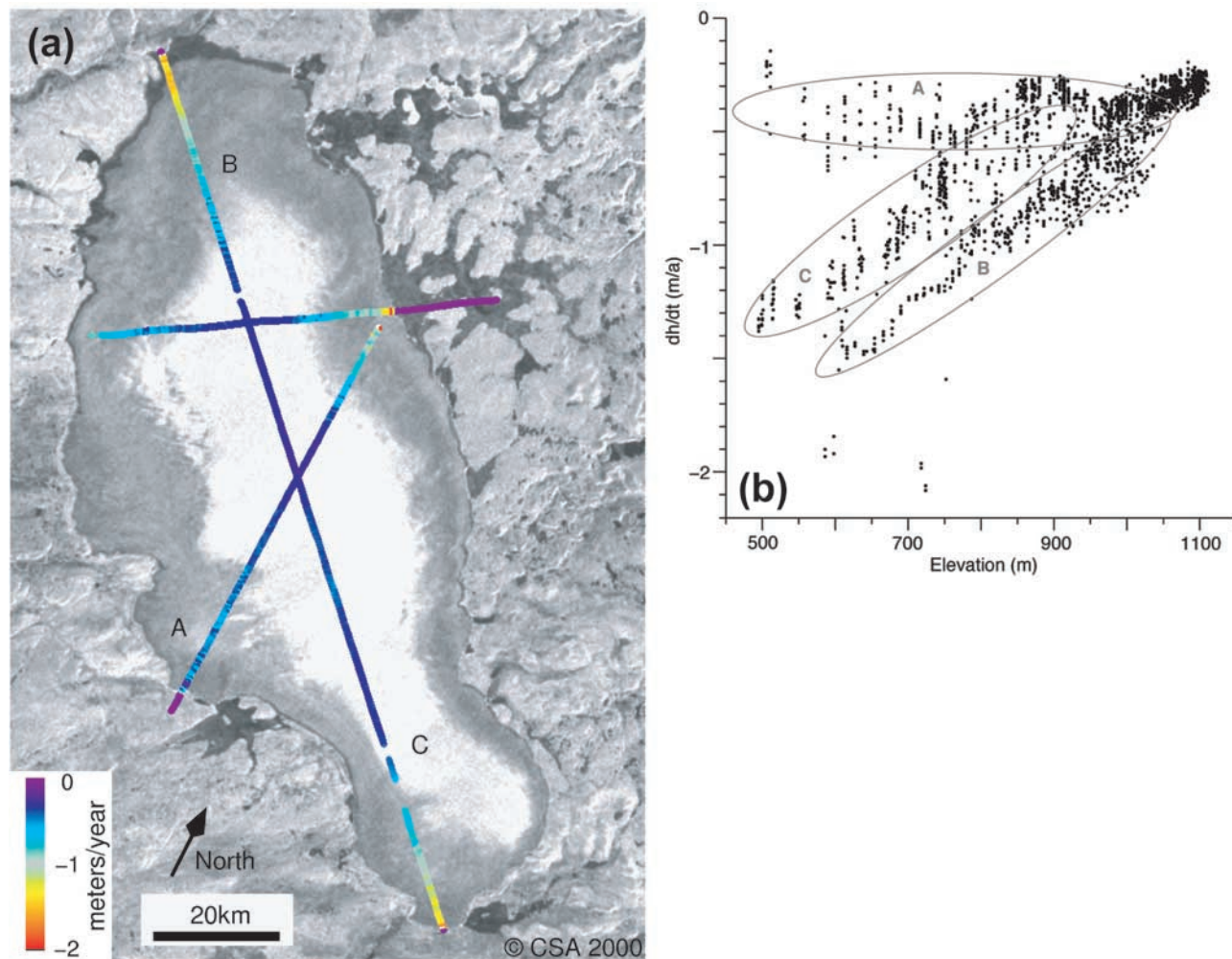
Figure 4. (a) Elevation change rate (dh/dt) for the Barnes ice cap along the three survey flight lines overlaid on a RADARSAT image from October 2000. The light and dark regions of the ice cap roughly represent the accumulation and ablation zones, respectively, with more substantial thinning clearly apparent in the ablation zones. (b) Scatterplots of those changes as a function of surface elevation. Segments of the various flight lines (A−C) are marked in Figure 4a, and their corresponding sections in the scatterplot are indicated with ovals in Figure 4b.

affects mass loss by ablation, the temperatures during the rest of the year are important because they affect densification and determine the energy needed to raise the snow and ice to the melting point. The temperature characteristics for the various stations are given in Table 1.

[20] Assuming that the summer melt season lasts ~3 months (supported by the coastal station data, which consistently show temperatures greater than zero for June, July, and August but less for the other months), the summer temperature anomaly can be converted to a positive-degree-day anomaly (PDD′) by multiplying it by the number of days in the summer melt season. Subsequently, the elevation change associated with the anomaly ($\Delta h'$) can be approximated by adapting the positive-degree-day method for ablation calculation [*Reeh*, 1991] as follows:

$$\Delta h' = k(\text{PDD}')\frac{\rho_w}{\rho_s}, \qquad (1)$$

where k is the PDD factor proportionality constant, ρ_w is the density of water, and ρ_s is the density of the snow or ice at the surface. *Braithwaite* [1981, 1995] reports a PDD factor for ice in Arctic Canada as being 0.0063 meters of water equivalent (m we)/PDD, and he provides no estimates for snow. However, his measurements were made at elevations of 210 m on White Glacier (on Axel Heiberg) and at 300 m on Sverdrup Glacier (on Devon Island). These glaciers are generally steeper than the main ice caps, and PDD factors often vary from glacier to glacier as a result of their slopes, aspects, and local climates. For the Greenland ice sheet the PDD factor is generally assumed to be 0.008 m we/PDD for ice and 0.003 m we/PDD for snow [*Braithwaite*, 1995]. In this study we use the Greenland values as they should represent conditions more consistent with the ice caps surveyed. We recognize that k is highly variable and is a major source of uncertainty, but for the purposes of roughly estimating the contribution of the 1995–2000 warm temperatures to the rate of elevation change (dh/dt) it is reasonable. More precise estimates are the subject of ongoing work. Assuming a snow density of 333 kg m^{-3} and an ice density of 917 kg m^{-3}, the $\Delta h'$ in equation (1)

Table 1. Temperature Characteristics From Various Coastal Stations in the Canadian Arctic and One in Western Greenland (Egedesminde)[a]

Location (Figure 1)	Summer (June, July, August)			Annual		
	52 Year Mean	1995–2000 Mean	1995–2000 Anomaly	52 Year Mean	1995–2000 Mean	1995–2000 Anomaly
Iqaluit	6.0	6.3 (6.3)	+0.29	−9.4	−9.0	+0.43
Clyde River	3.1	3.0	−0.088	−12.3	−12.5	−0.16
Resolute	2.0	2.1 (2.7)	+0.064	−16.4	−15.5	+0.94
Eureka	3.6	4.4 (4.7)	+0.70	−19.6	−18.4	+1.2
Egedesminde	4.8	4.9 (5.7)	+0.16	−5.0	−4.7	+0.27

[a]All values are in °C. Values from 1995 were not available for Clyde River, but the 1995 anomaly at the other Canadian stations (given in parentheses) ranged from zero at Iqaluit (the nearest station) to 0.6°C at Resolute (the farthest station), suggesting that even in the absence of 1995 data, the Clyde River anomalies are reasonable.

becomes 0.009 m of elevation change/PDD′ for both ice and snow (where m in this case refers to meters of elevation change). This is the same approach used by *Krabill et al.* [2000] and *Abdalati et al.* [2001].

[21] The resulting elevation change rates corrected for the 10 cm accumulation increment are given in Table 2. The values reported are for the lowest elevations. At the high elevations, in the accumulation zones, the contribution of the summer temperature anomaly approaches zero, leaving only the accumulation effects.

[22] Figures 2 and 3 indicate that for the ice caps on Ellesmere, Axel Heiberg, and Meighen Islands (the first six ice caps listed in Table 2), observed thinning is generally <50 cm yr^{-1}, with a few exceptions. The summer anomaly at nearby Eureka is 0.7°C. Over a 3 month melt season (following equation (1)), this translates to a surface lowering of nearly 58 cm, or 48 cm after accounting for accumulation effects. The ablation losses are likely to be slightly greater when we consider that a warmer melt season is likely to be a longer melt season. Thus we can conclude that the elevation changes on these ice caps are consistent with what would be expected based on temperature anomalies in the region.

[23] The temperature anomaly in the vicinity of Devon Island is estimated to be 0.23°C, leading to an estimated surface lowering of 20 cm yr^{-1}, or 10 cm yr^{-1} after adjusting for increased accumulation. This, too, is on the order of what was measured, so we attribute the slight or zero thinning observed along the Devon flight lines to regional climate anomalies during the survey period. If these anomalies were to continue, we should expect to see the ice caps continue to thin accordingly.

[24] The closest station to the Barnes ice cap, which thinned by more than a meter per year in some places, is Clyde River. Clyde River data showed a negative temperature anomaly, so we would expect to see a slight thickening in this area. The discrepancy with the observed thinning of >1 m yr^{-1} at lower elevations is quite large. For the region around Penny ice cap the estimated summer anomaly is 0.12°C, and the resulting ablation anomaly is 10 cm yr^{-1}. This is effectively canceled out by increased accumulation.

[25] It is important to note, however, that Clyde River temperature data were missing for 1995, so the anomaly is only based on data from the four summers of 1996–1999. Across Baffin Bay, however, in Egedesminde, a full record indicates a summer anomaly of only 0.15°C, so it is reasonable to assume no appreciable temperature-related component to the thinning on Barnes. Furthermore, at the nearest Canadian station in our record, Iqaluit, the 1995 summer temperature (6.3°C) was identical to the average of the other 4 years in the 1995–1999 period, so the absence of 1995 in the summer temperature anomaly calculation is not considered to be significant.

[26] Additionally, since this is the area of greatest change and smallest apparent temperature anomaly, we need to consider how representative the Clyde River data from the coast are of the conditions at the interior of Baffin Island, where the only temperature data available are from Dewar Lakes (Figure 1) for the period 1958–1989. We compared the 31 year record to that of Clyde River for the same period. The summer temperature correlation between the two stations was $R = 0.9$, and their summer mean values over the period were identical at 2.9°C. The only significant difference between the two stations was that the standard

Table 2. Elevation Change Estimates Based on Summer Temperature Anomalies From the Nearest Coastal Weather Stations

Location (Figure 2)	Coastal Stations (Figure 1)	Temperature Anomaly, °C	Associated dh/dt,[a] cm yr^{-1}
Northern Ellesmere ice caps	Eureka	0.7	−48
Prince of Wales ice caps	Eureka	0.7	−48
Agassiz	Eureka	0.7	−48
Southern Ellesmere ice caps	Eureka	0.7	−48
Mueller ice cap	Eureka	0.7	−48
Meighen ice cap	Eureka	0.7	−48
Devon ice cap	Eureka, Resolute, Clyde River	0.23	−9
Barnes ice cap	Clyde River	−0.09	+17
Penny ice cap	Clyde River, Iqaluit, Egedesminde	0.12	0

[a]After adjusting for the 10 cm yr^{-1} regional accumulation anomaly.

deviation at Dewar Lakes was 60% greater than that of Clyde River (1.6°C versus 1.0°C) as a result of their relative proximity to Baffin Bay and its moderating effects. These results indicate that the Clyde River temperature data serve as an appropriate proxy for conditions at the interior of Baffin Island and further suggest that the survey period was not anomalously warm in the Barnes and Penny regions.

[27] Thus we conclude that unlike the changes observed in the Queen Elizabeth Islands (QEI), the thinning of the ice caps on Baffin Island is not a result of anomalous melting during the late 1990s. It is possible that these changes are related to dynamic effects, as has been the case in many parts of Greenland [*Krabill et al.*, 2000; *Abdalati et al.*, 2001], but the extensive thinning in areas far removed from outlet glaciers suggests otherwise. It is more likely that the observed thinning is part of the ongoing adjustment to warming after the Little Ice Age, as has been suggested for Barnes ice cap by *Jacobs et al.* [1993, 1997]. In this case, recent climate conditions are only a small part of a larger background phenomenon.

4.2. Role of Dynamics

[28] Quantifying the effects of ice cap and glacier dynamics on this overall balance is particularly difficult because, unlike temperature, no simple parameterizations exist that allow estimates from the current observation record. While velocities can be derived from satellite imagery, the extent to which they are above or below balance velocities requires either a knowledge of the accumulation and ablation values for flux calculations at a level that does not exist on these larger scales or, at a minimum, a substantial temporal record of satellite and in situ velocity measurements, which we do not have. For the purposes of this analysis we consider only the effects of temperature and accumulation anomalies. Such considerations seem to sufficiently account for the elevation changes that we observe on the ice caps on the Queen Elizabeth Islands. Moreover, unlike climatological controls, dynamic behavior is basin-specific. Consequently, these effects would likely appear as inconsistencies in the *dh/dt* versus elevation relationships on individual ice caps or glaciers that are in close proximity to one another. While we do see some variability, it is not of such magnitude that it would contribute substantially to the overall analysis.

[29] A separate dynamics issue that needs to be considered is the impact of surges on our analysis. *Copland et al.* [2003] did a comprehensive assessment of the prevalence of surge-type glaciers in the Canadian High Arctic (the Queen Elizabeth Islands). While they found evidence of significant surge-type behavior or likelihood in 51 glaciers in the region, 15 were observed to be actively surging during 1999, which is consistent with the time of our measurements, and of these, four were part of our survey. However, surging glaciers typically have distinct signatures that we see in the elevation change data. The order-of-magnitude acceleration associated with actively surging glaciers should produce thinning in the surge area that would be much larger than the meter per year that we observed. If the surveys were made during quiescence, resulting *dh/dt* profiles would tend to show thickening upslope of the region of surge, as driving stresses build in advance of the next surge, and thinning downslope, as the downstream ice

continues to move toward the glacier terminus without being dynamically balanced by the accumulating upstream ice. Such elevation profiles were evident in several of our Greenland outlet glacier surveys [*Abdalati et al.*, 2001], but they are only clearly evident in one of the Ellesmere Island flight lines (at 77.5°N, 282°E). There are some subtle suggestions in the other flight lines, but the magnitude of the changes, which are typically <0.5 m yr^{-1}, suggests that the impact on the overall assessment is not substantial.

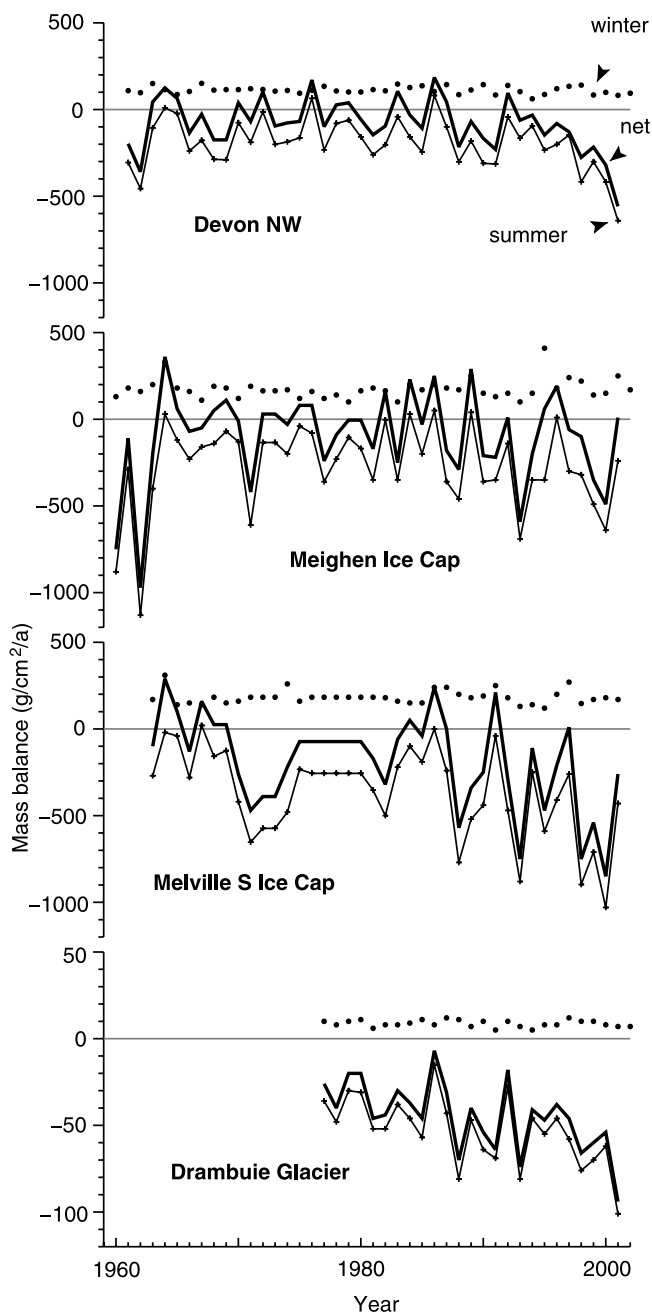## 4.3. Recent Historical Mass Balance Context

[30] Long records of in situ data for the ice caps and glaciers are sparse; however, there have been 4 decades of consistent mass balance measurements in the Canadian Arctic on Devon, Meighen, and Melville ice caps and 25 years of mass balance data on Drambuie Glacier on Agassiz ice cap. Unfortunately, the Devon laser surveys were not as comprehensive as the others due to bad weather in 1995, so data are limited, and the Melville ice cap, located nearly 900 km west of Devon Island, is far from the ATM survey lines. However, these measurements, along with those on the Meighen ice cap, help provide a broader picture of how the observed changes in the region fit into an historical context.

[31] These in situ data (Figure 5) show that while net balance in the areas measured has generally been negative for much of the time dating back to when records began in the 1960s, the 1995–1999 period was generally more negative than in most of the previous period. This is likely due in part to the strong warm temperature anomaly in the summer of 1998 [*Zhang et al.*, 2000]. However, the same period does form part of a trend toward increasingly negative balances, which began in the mid-1980s. The apparent acceleration of mass loss during the survey period has also been observed in Alaskan glaciers in recent years [*Arendt et al.*, 2002]. It is important to note that the net balance is clearly a strong function of the melt and does not appear to be correlated with the accumulation characteristics on each of the ice caps. The strong link to melt further supports the idea that the recent temperature anomalies, and not accumulation variations, drove much of the observed thickness changes. A detailed look at the elevation dependence of *dh/dt* on Devon ice cap (Figure 6) shows a more negative 1995–1999 balance anomaly at lower elevations than at high, and in fact, at the highest locations on the ice cap, the anomaly is positive, suggesting less thinning, or a slight thickening, as was observed with the ATM. The Devon in situ measurements are the most comprehensive, but this same elevation dependence is characteristic of the other field sites in the region.

4.4. Estimated Contributions to Sea Level Rise

[32] To date, there has been no reliable observational estimate of the contributions to sea level rise from the ice caps in the Canadian Arctic Archipelago. In situ measurements have historically been very local in nature; the capability of satellite radar altimetry is severely limited on these smaller ice masses because of their steep slopes and relatively rough topography, and Shuttle Imaging Radar altimetry, which has been used successfully in on the Patagonia ice fields [*Rignot et al.*, 2003], did not provide coverage at these high latitudes. Consequently, the airborne
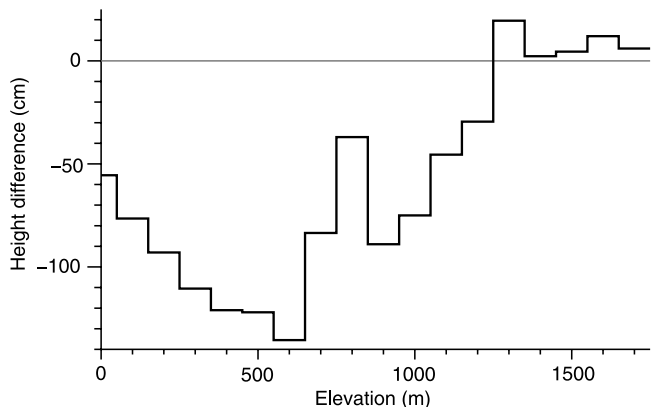
Figure 5. In situ-derived mass balance estimates for several Canadian ice caps and glaciers. Separate curves are shown for summer, winter, and net (algebraic sum of summer and winter).

[33] To make these calculations, we used three data sets: (1) elevation change from our measurements; (2) ice cover extent data provided in digital form by Natural Resources, Canada; and (3) elevation from the U.S. Geological Survey GTOPO30 global digital elevation model (DEM) (http://edcdaac.usgs.gov/gtopo30/gtopo30.asp). The elevation change rates calculated from our data were resampled to a 625×625 m rectangular grid by averaging all the values within each grid cell. For each ice cap/ice field the dh/dt values were binned according to elevation in 50 m intervals. Using the median dh/dt values within each bin, we determined the relation between dh/dt and elevation (alternatively, we could have used the mean, but the median is less sensitive to the influence of outliers and yields a more conservative estimate by roughly 6%). We then calculated the total mass balance for each surveyed ice cap by integrating (over the ice-covered grid cells) the product of the elevation-derived dh/dt and grid cell area. The mass balance estimates for the ice caps that we surveyed are given in the Table 3 (third column). During the 1995–2000 time period the total volume change rate for these ice caps, which represent nearly two-thirds of the total ice-covered area in the Canadian Arctic Archipelago, was 15.5 km^3 yr^{-1}.

[34] To extend this estimate to the entire archipelago, we divided the area into different regions (shown in the left side of Figure 2 with magenta lines) and repeated the above procedure. For the most part, the regions were defined by the proximity of unsurveyed ice masses to those that we did survey. The only exception was in the ice cover on Northern Baffin Island, which we included in the Penny region despite its proximity to Barnes. Most of these glaciers and ice caps were at lower elevations than the ice on Barnes, and the surrounding terrain was more similar to Penny than the smoother characteristics of Barnes. Consequently, Penny provided a better basis for the extrapolation. Moreover, because the mass loss on Barnes is the most rapid of all the areas we surveyed, extending its results to the other regions runs the risk of excessively negative mass balance estimates for these smaller ice masses. On Bylot Island the climate and topographic characteristics are more similar to Devon ice cap, so it was included in the Devon region.

laser altimetry measurements presented here provide the first reasonable basis for estimating total mass balance of these ice masses and the associated contributions to sea level change. Because of the unique behavior of each ice mass and the fact that not all ice caps and glaciers in the area were surveyed, the calculation has a large degree of uncertainty. Despite these limitations, however, these observations provide the best means of making this important assessment for the period surveyed, so the calculation is worth making.

Figure 6. Difference between 1995–2000 mass balance and that from 1961 to 1994 on Devon NW, plotted as ice thickness change. Differences are shown as a function of ice cap elevation.

Table 3. Estimates of Volume Change Rates (ΔV) of Each Surveyed Ice Area as Well as the Estimated ΔV for the Entire Region[a]

Location	Ice Cap Area, km^2	Ice Cap ΔV,[b] km^3 yr^{-1}	Regional Ice Area, km^2	Regional ΔV,[b] km^3 yr^{-1}
North Ellesmere	27,666	−3.52	28,212	−3.59
Agassiz	17,523	−1.23	21,937	−1.98
Mueller/Meighen	8,267	−1.84	11,800	−2.87
Prince of Wales	19,325	−1.70	19,475	−1.68
South Ellesmere	9,165	−1.80	10,156	−1.91
Devon	12,794	−0.81	22,156	−1.83
Barnes	5,671	−3.43	5,956	−3.74
Penny	7,335	−1.13	28,506	−7.56
Total	107,746	−15.46	148,198	−25.16

[a]As described in Figure 2.

[b]ΔV for ice areas that were not surveyed was calculated using the *dh/dt* versus elevation relationships of the nearest surveyed ice cap, as described in the text.

Using the DEM and *dh/dt* relationships, and integrating as before, we calculated the balance for each region (Table 3). Our results show a combined total mass balance for the 1995–2000 time period of −25 km^3 yr^{-1}.

[35] It is reasonable to assume that these volume changes are generally losses of ice (rather than firn) since the thinning observed was primarily in the ablation zones, while thickening or little change was observed at higher elevations in the accumulation zones. Assuming that this is the case, the contribution of the Canadian ice caps to sea level rise during the survey period is estimated to be 0.064 mm yr^{-1}.

[36] Clearly, there is a large uncertainty in this estimate that arises from our limited coverage and the extrapolation of behavior from one glacier region to another. Dynamic effects vary from place to place, and the impact on our results is very difficult to quantify. However, we took steps throughout our calculations to be conservative. When compared to the recent estimates that have been made for some of the larger ice masses of the world (Greenland at 51 km^3 yr^{-1} in the mid-1990s [*Krabill et al.*, 2000], Alaska at 57 km^3 yr^{-1} from the mid-1950s to the mid-1990s and at 105 km^3 yr^{-1} in the late 1990s [*Arendt et al.*, 2002] (after adjusting their reported water equivalent values for the density of ice), and the Patagonian ice fields at 16.7 km^3 yr^{-1} over the last 3 decades of the 20th century and at 41.9 km^3 yr^{-1} from 1995 to 2000 [*Rignot et al.*, 2003]), the mass loss from the Canadian ice caps is appreciable.

[37] It is clear that the anomalously warm period between surveys in the Canadian High Arctic resulted in higher mass loss values than in previous years. This is suggested in the climatological record and is evident in the limited glaciological record. The future balance conditions will depend how sustained these warm anomalies continue to be. However, it is important to recognize that roughly half of the mass loss is from the ice caps on Baffin Island, which experienced essentially no anomaly at all. Thus we conclude that even if temperatures in the vicinity are not especially warm, the sea level contributions of the Canadian ice caps will continue to be significant. If the region warms, they will be even greater.

5. Conclusions

[38] A set of airborne laser altimetry surveys over the major ice caps on the Canadian Archipelago in 1995 and 2000 revealed thinning at lower elevations on most of the ice caps and glaciers but very little change or thickening at higher elevations in the ice cap accumulation zones. This thinning was much more pronounced on the southern ice caps on Baffin Island (Penny and Barnes) than on the more northerly ice caps on the Queen Elizabeth Islands. The elevation dependence agrees with the in situ observations on Meighen, Agassiz, Devon, ice caps in Nunavut, and the Melville ice cap in the Northwest Territories.

[39] The high-elevation thickening is consistent with recent accumulation anomalies at surrounding coastal weather stations and is also consistent with an increasing trend in accumulation reported for the region. Recent warm temperatures during the 1995–2000 time period appear sufficient to explain the observed low-elevation thinning of roughly 0.5 m yr^{-1} observed in most ice caps in the Queen Elizabeth Islands. On Baffin Island, however, where thinning was most pronounced, temperatures during the period were not particularly warm, and in some cases they were slightly cool. Thus we believe that the observed changes of about 1 m yr^{-1} are not directly attributable to climate conditions of the late 1990s but are most likely part of the ongoing response to a longer-term deglaciation in recent centuries [*Jacobs et al.*, 1993, 1997].

[40] At low elevations, *dh/dt* is widely scattered, even for individual ice caps, but at high elevations the changes converge such that for all of the ice caps, most of the scatter is bound by about a 20 cm yr^{-1} range. Collectively, the ice caps showed thinning at elevations below about 1600 m elevation and thickening above 1600 m. There were exceptions, however, such as the Prince of Wales ice field, which thickened even at low elevations, and the Barnes ice cap, which is thinning substantially at all measured locations. A broad extrapolation of these changes is limited by the fact that they are sometimes very local in nature, varying from one side of an ice cap to the other and from glacier to glacier.

[41] In situ measurements on several ice caps on the Queen Elizabeth Islands show that mass balance at the field sites between 1995 and 2000 was more negative than observed over the preceding 2 or 3 decades and that the changes were strongly driven by melt. The relationship provides further evidence that the survey period was anomalously warm and that these warm temperatures were responsible for some of the thinning at the lower elevations. However, these longer in situ data records also suggest that the thinning of the late 1990s appears to be a continuation of a phenomenon that began in the mid-1980s.

[42] We estimate the 1995–2000 mass balance of the ice caps in the Canadian Arctic Archipelago to be −25 km^3 yr^{-1}, which corresponds to a sea level rise contribution of 0.064 mm yr^{-1}. This estimate, along with others, supports the idea that Earth's shrinking ice cover is contributing to sea level rise at a more substantial rate than was previously believed.

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