Thinning of the south dome of Barnes Ice Cap, Arctic Canada, over the past two decades

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ABSTRACT

Between 1970 and 1984, ground surveys were carried out along a flowline extending from the top of the south dome of Barnes Ice Cap to the margin. Over this time span, the ice cap thinned an average of 1.7 m, or 0.12 m yr^{-1}. By comparing the 1984 survey with elevations derived from satellite imagery in 2006, we find that it has now thinned an additional 16.8 ± 7.7 m, or an average of ~0.76 ± 0.35 m yr^{-1}. Laser altimeter profiles show that between 2004 and 2006, the thinning rate was 1.0 ± 0.14 m yr^{-1}. A correlation between mass balance and mean summer temperature at nearby weather stations, developed over the period of the ground surveys, permits independent estimates of the thinning rate. These estimates are in excellent agreement with those based on satellite imagery. The acceleration in thinning is consistent with meteorological records documenting an increase in the number of positive degree-days (atmospheric warming) in the region.

Keywords: Barnes Ice Cap, global warming, glacier retreat, satellite remote sensing.

INTRODUCTION

It is widely recognized that climate warming is affecting glaciers all over the world. Long records that can be used to document these changes on polar ice caps and ice sheets are not as common as those from temperate mountain glaciers, yet the impacts of climate change are predicted to be largest in the polar regions (ACIA, 2005), and future sea level will be modulated in large part by ice masses there.

More than twice as much glacial ice is in the Canadian high Arctic, above ~74°N, than in the low Arctic (Oerlemans et al., 2005). Perhaps for this reason, the best available records of ice cap thinning in this region have been from the high Arctic. Abdalati et al. (2004), for example, measured thinning rates between 1995 and 2000 along ~2200 km of flight lines in the high Arctic, but only ~700 km in the low Arctic.

The thinning rate is less in the high Arctic. Between 1962 and 2000, negative net mass balances on Devon Ice Cap (75°N) and on White Glacier on Axel Heiberg Island (79°N) (Johansson, 2002) were ~1/3 those we find, herein, on the south dome of Barnes Ice Cap (69°N) on Baffin Island. Abdalati et al. (2004) note a similar contrast in their more extensive data set from the late 1990s. This contrast reflects a northward decrease in the number of positive degree-days, averaged over the period 1980–1999, from 378 at Dewar Lakes (68.6°N) to 195 at Alert on northern Ellesmere Island (82.5°N).

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Christensen et al. (2007) project that by the end of the twenty-first century the mean annual temperature in the Arctic will have increased ~5 °C, relative to the 1980–1999 mean, with the warming predominately during the autumn and winter. If this projected warming occurs, a large fraction of the world's glaciers and small ice caps are likely to have vanished by the end of the twenty-first century. This will contribute ~0.2 m to the global average sea level rise (Meehl et al., 2007).

With the warming scenario proposed by Christensen et al., we estimate that the increase in the number of positive degree-days will be ~327 at Dewar Lakes but only ~222 at Alert. In addition, changes in precipitation patterns by the end of the twenty-first century, with an increase in winter precipitation that increases northward from <5% in the low Arctic to 30%–40% in the high Arctic (Christensen et al., 2007), may moderate the effects of increasing temperature on mass balance in the high Arctic. Thus, the sensitivity of glaciers to temperature change is greater in the low Arctic, and their contribution to early twenty-first century sea level rise is projected to be nearly the same as that of the significantly larger ice masses in the high Arctic (Oerlemans et al., 2005, their Fig. 2A).

Heretofore, the published mass balance record for the south dome of Barnes Ice Cap covered the period from 1962 to 1983 with minor gaps (Hooke et al., 1987). In this paper we extend this record forward to 2006, revealing a dramatic increase in the rate of thinning. We also show that this thinning is a response to rising temperatures, contrary to assertions by Abdalati et al. (2004).

Barnes Ice Cap covers ~5900 km^2 on Baffin Island. In 1970 Holdsworth (1975) set 43 stakes in a series of overlapping strain nets along an ~10 km flowline extending from the top of the ice cap's south dome to its northern margin. By 1984 only 24 of these stakes were recoverable (Fig. 1B). Surveys, carried out nearly annually from fixed points in the glacier forefield to some or all of these stakes (Holdsworth, 1975; Hooke et al., 1987), revealed that by 1984 the ice cap had thinned an average of ~1.7 m along this flowline, or ~0.12 m yr^{-1}. Nearly annual mass balance measurements were also made during this period.

During the period of the ground survey, the equilibrium line varied from ~600 m to >900 m, and the winter balance was typically ~0.5 m (water equivalent, w.e.). During positive-balance years, net accumulation was composed of both firn and superimposed ice in roughly equal quantities. Balance gradients were typically ~0.006 m yr^{-1} w.e. in the ablation area and ~0.001 m yr^{-1} w.e. in the accumulation area (Hooke et al., 1987). The maximum velocity along the flowline occurred between 1 and 3 km from the margin and was ~7 m yr^{-1}. The 10 m ice temperature was ~–10 °C, and basal temperatures were ~–10 °C near the margin and –4 °C beneath the divide (Hooke et al., 1987).

With the use of satellite remote sensing data, we have now obtained surface elevations for 2006 along the same flowline, and for 2004 and 2006 along a nearby line. In addition, we have

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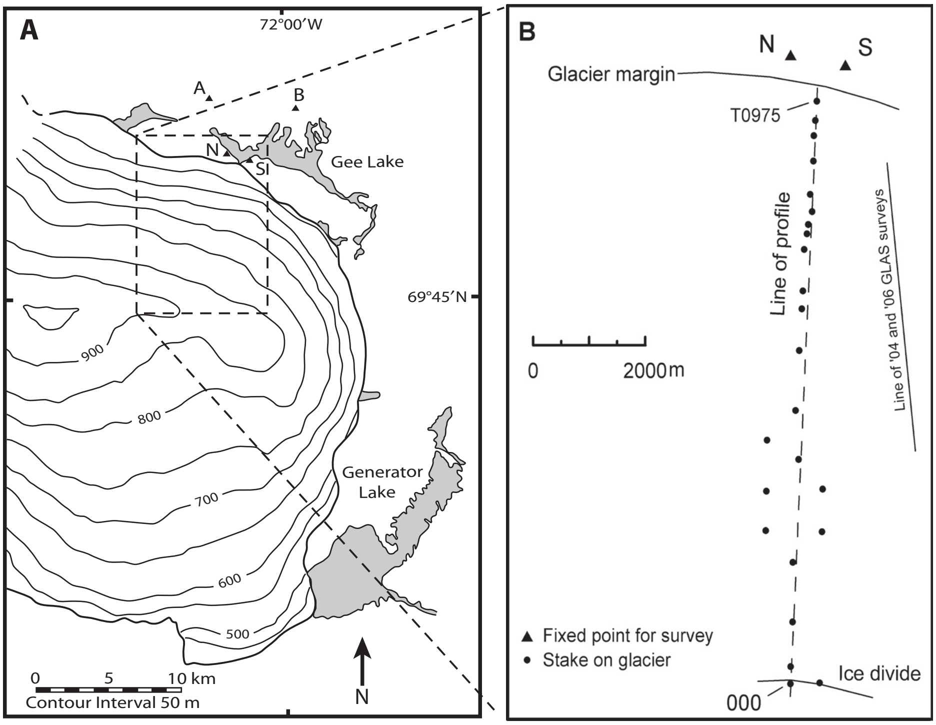

Figure 1. A: Map of south dome of Barnes Ice Cap showing location of the flowline and of fixed points used for ground-based surveys. B: Locations of stakes on the flowline and of lines of the 2004 and 2006 GLAS surveys. Stakes at end points of profile are labeled, and coordinates are given in Table 1.

independently estimated the change in elevation since 1984 using temperature measurements from nearby weather stations and a correlation between mean summer temperature and net mass balance obtained from the earlier studies.

Herein, we describe the procedures used to obtain the updated surface elevations and discuss the results.

COORDINATE SYSTEMS

The fixed points used for the ground surveys are A, B, N, and S in Figure 1. Geodetic and Universal Transverse Mercator (UTM) coordinates for points A and B were obtained from G. Holdsworth (1979, written commun.) and are based on an earlier unpublished survey by J.P. Henderson in 1966. Geodetic and UTM coordinates for points N and S are based on triangulation from points A and B (see Holdsworth, 1975, and Hooke et al., 1987, for details). These four fixed points are all referenced to the Clarke 1866 ellipsoid. Point B is in UTM zone 19N, while points A, N, and S, and the stakes on the glacier, are in zone 18N.

The digital elevation model (DEM) used in this study was extracted from stereo satellite images collected by the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) sensor and is referenced to the WGS-84 ellipsoid in the UTM projection (Abrams et al., 2001). When an ASTER image straddles a UTM zone boundary, the whole 60 × 60 km image is "forced" into a single zone. In the case at hand, the DEM is in UTM zone 18N.

Elevation data from the Geoscience Laser Altimeter System (GLAS), onboard the Ice, Cloud, and land Elevation Satellite (ICESat), were used to check the elevations of the ASTER DEM and were also used to create additional elevation profiles along the ice cap. The GLAS data are referenced to the TOPEX/Poseidon-JASON ellipsoid and are provided in geodetic coordinates (Schutz, 2002). Horizontal differences between WGS-84 and TOPEX/Poseidon ellipsoids are on the order of a few centimeters, while elevation differences (~50 cm at the equator) increase with latitude (http://www.nsidc.colorado.edu/data/icesat/faq.html).

To develop a consistent reference frame, compatible with the satellite data, we translated points A, B, N, and S to the WGS-84 ellipsoid and then reprojected them into UTM 18N coordinates. Point B was warped into UTM zone 18N so that distances measured from B across the UTM zone boundary were accurate. GLAS data were likewise converted to the WGS-84 ellipsoid. Table 1 lists the resulting geodetic and UTM coordinates for points A, B, N, and S.

The 24 stakes remaining in 1984 were not in a straight line (Fig. 1). Therefore, we fit a straight line through them and chose the 18 stakes closest to the line (rms deviation 43 m) to create an elevation profile from stake T0975 to the divide (Fig. 2). This profile was also referenced to the WGS-84 ellipsoid.

ASTER AND GLAS GEOLOCATION AND ELEVATION UNCERTAINTIES

ASTER DEMs are constructed by using images from the visible near-infrared (VNIR) bands 3N and 3B (Fujisada et al., 2005). The viewing angle of VNIR 3N is nominally nadir, while that of 3B looks aft ~27° in the along-track direction. The spatial resolution of each scene is 15 m, while the resulting DEM has a spatial resolution of 30 m. There is a small geolocational uncertainty in the ASTER DEM. DEMs generated with the same software used in this analysis have reported geolocation uncertainties <50 m, and elevation uncertainties of ~15 m (Fujisada et al., 2005; Stearns and Hamilton, 2007).

The footprint of the laser spot produced by GLAS is ~70 m, and the separation between successive spots is 172 m (Schutz et al., 2005). Carabajal and Harding (2005) report a horizontal geolocation error 2.4 ± 7.3 m for Release 22 data. GLAS data used in this study are from Release 28. Incremental changes in the processing of the signal waveform, as noted in the release notes of the intervening releases, have reduced the geolocation errors reported by Carabajal and Harding (2005). The accuracy of GLAS elevation data depends upon a host of orbital and atmospheric parameters (Brenner et al., 2003), but the magnitude of elevation error depends, in large part, on the incidence angle of the laser with Earth's surface. Carabajal and Harding (2005) report the vertical error to be 0.04 ± 0.13 m per degree of incidence angle. Shuman et al. (2006) estimate a relative elevation accuracy for all of Antarctica, based on statistical analysis of repeat and crossover tracks, of ±0.14 m with a precision of slightly more than 0.02 m. In any case, whatever the absolute elevation uncertainties of GLAS

TABLE 1. COORDINATES OF FIXED POINTS AND OF END POINTS OF 1984 PROFILE IN DECIMAL DEGREES AND IN UTM METERS, ZONE 18N, REFERENCED TO THE WGS-84 ELLIPSOID

Point	Latitude	Longitude	Northing	Easting
A	69.870459167N	72.141636111W	7,753,998.207	609,749.553
B	69.863673611N	71.978194444W	7,753,544.711	616,058.192
N	69.834309116N	72.084511995W	7,750,074.308	612,134.491
S	69.829703799N	72.065992929W	7,749,601.407	613,003.756
T0975	69.826021329N	72.081429167W	7,749,157.050	612,296.689
000	69.746312908N	72.184536667W	7,740,090.185	608,741.775

data, they are insignificant compared with those of an ASTER DEM.

Given these spatial resolution, geolocation, and elevation uncertainties, little is to be gained by using the UTM coordinates of individual stakes on the glacier to derive elevations from an ASTER DEM. Thus, we focus on changes in the profile along the entire flowline.

CHANGES IN PROFILE

To determine the change in elevation since 1984, we extracted elevations along the flowline from an 8 August 2006 ASTER DEM (Fig. 2). The undulations above 700 m in this profile are an artifact introduced by the commercial software used to process the DEM. To extract a useful profile, we thus fit a fourth-order polynomial to these data and also to the 1984 ice surface elevations at the stake locations. The difference between these profiles, calculated at 654 points along the flowline and averaged, is 16.8 ± 7.7 m,[1] giving a thinning rate of 0.76 ± 0.35 m yr^{-1} averaged over the 22 yr. The transverse strain rate, averaged along the flowline, was compressive in the late 1970s, resulting in an estimated thickening of ~0.17 m between 1970 and 1984 (Hooke et al., 1987). This was more than offset by the thinning due to the negative mass balance. Although the strain became slightly less compressive between 1970 and 1984, it is unlikely that a continuation of this trend could explain a significant fraction of the recent thinning. We therefore attribute it to climate change, a possibility that we now explore in greater depth.

As an independent check on these results, we calculated the change in elevation using meteorological data from nearby weather stations. With the use of 17 yr of data between the 1960s and 1980s, Hooke et al. (1987, their Fig. 5) established a relation between mean summer temperature (defined as June–September) recorded at Dewar Lakes and net balance measured on the ice cap. To use this relation for our analysis, we required temperature data from Dewar Lakes for 1984–2006. Unfortunately, data for the years 1993 through 1999 and 2004 are missing from the Dewar Lakes record. However, there is a good correlation (r = 0.854) of degree-day data between Dewar Lakes and Clyde River, a station on the east coast, 226 km from Dewar Lakes (Jacobs, 1989). The Clyde River data are complete through 2005. We thus used a linear regression between mean summer temperature at Dewar Lakes and that at Clyde River for the period 1959–2005 to estimate the missing Dewar Lakes data. A cubic equation with an rms deviation from the measured values of 0.11 was then fit to earlier data of Hooke et al. (1987), and the net balance (in meters of water equivalent per year, m yr^{-1} w.e.) was calculated for the years 1984–2005. Summing these annual values, converting to ice equivalent, and incorporating the rms deviation above yields a total decrease in elevation of 17.6 ± 0.5 m, or 0.80 ± 0.02 m yr^{-1}. In the original work of Hooke et al. (1987), however, there was a discrepancy between thinning based on surveyed profiles and that based on mass balance measurements, suggesting that the latter underestimated the mass balance rate by ~0.06 m yr^{-1}. Hooke et al. (1987) attributed this to failure to account adequately for internal accumulation due to the refreezing of meltwater. Taking this into consideration results in a thinning rate over the past 22 yr of 0.74 ± 0.02 m yr^{-1}. Either result is in excellent agreement with the change based on the ASTER DEM.

This agreement differs from results reported by Abdalati et al. (2004) based on airborne laser surveys in the springs of 1995 and 2000. They found thinning rates on the south and southwest sides of the south dome that were comparable to ours, but were unable to find a consistent relation between thinning rate and temperature. This was apparently due to a problem with the data set used by Abdalati et al. (2004). They reported mean summer temperatures for 1995–2000 at Clyde River that were 0.09 K cooler than the 52 yr mean, but the data file presently available (http://climate.weatheroffice.ec.gc.ca/climate_normals/stnselect_e.html) shows temperatures for this period that were 0.8 K warmer than the long-term average. Furthermore, Abdalati et al. (2004) used the mean June–August temperature, whereas Hooke et al. (1987, p. 1557) found a better correlation with field measurements when summer was defined as June–September.

As a further check on these results, we examined GLAS data sets from 10 February 2004 and 24 February 2006. The relevant satellite tracks lie ~1700 m east of the 1984 stake line (Fig. 1), are more or less parallel to each other, and are ~60 m apart. They have 29 overlapping elevation measurements covering the elevation range from ~540 to ~750 masl. The mean decrease in elevation of these 29 points was 2.0 ± 0.14 m for the two-year period, or 1.0 ± 0.07 m yr^{-1}. For comparison, the mean thinning based on the summer temperatures for this two-year period is –0.95 m yr^{-1}, again in excellent agreement.

As was the case between 1970 and 1984, the thinning between 1984 and 2006, and between 2004 and 2006, was greatest at lower elevations (Fig. 3). This is not unexpected, as a uniform increase in mean temperature will result in more positive degree-days at lower elevations than at

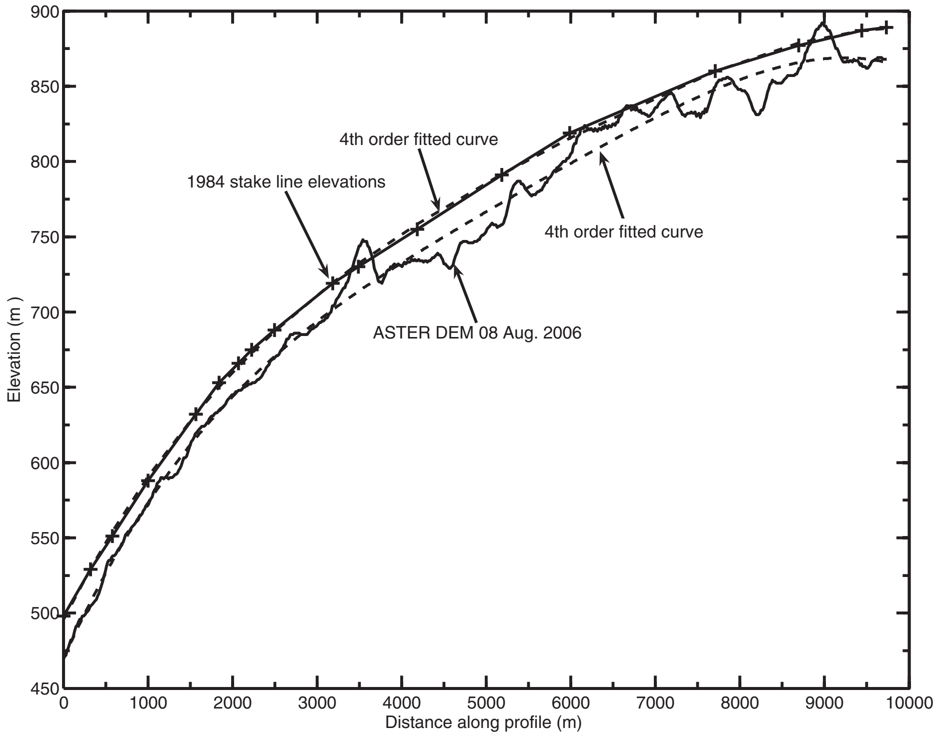

Figure 2. Elevation profiles along the flowline based on the 1984 ground-based survey and the 2006 ASTER DEM.

[1]The uncertainty is the rms deviation between the DEM and its fourth-order polynomial approximation (9.8 m) minus rms deviation between the 1984 profile and its fourth-order polynomial (2.1 m). With uncertainties of this magnitude, the 1984 surveyed profile (<±0.05) is effectively exact. Thus, in subtracting the 2.1 m we are assuming that the actual 2006 profile differs from the polynomial in the same sense and by the same amount, an assumption that is supported by the observation (Fig. 2) that the deviations from the polynomials at comparable elevations are similar in sense.

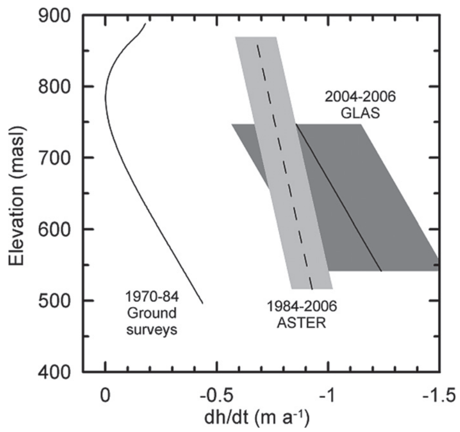

Figure 3. Variation in thinning rate, dh/dt, with elevation. The 1970–1984 data are based on ground surveys and are thus most accurate, meriting the nonlinear fit shown. The uncertainty is judged to be approximately the width of the line. The dashed line for the ASTER data is a linear fit (r = 0.60) to the difference between the two fourth-order profile curves in Figure 2. The uncertainty shown by the shaded area is based on the rms difference between the linear fit and the actual difference between the curves. The solid line for the GLAS data is also a linear fit (r = 0.35), and the shaded area is based on the rms difference, ~0.30 m, between this fit and the actual data. Based on the uncertainty in a single GLAS measurement reported by Shuman et al. (2006), ±0.14 m, we expect an uncertainty in the change over 2 yr to be only ~±0.1 m. We attribute the larger rms deviation to the fact that the 2004 and 2006 GLAS flight lines are ~60 m apart, which is only slightly smaller than the footprint of the GLAS laser.

higher ones. For example, with a mean annual temperature at the margin of 0 °C, a sinusoidal annual temperature cycle with a range of 12 K, and a lapse rate of -0.006 K m^{-1}, a 1 K increase in mean annual temperature results in 192 more positive degree-days per year at the margin but only 70 at the top of the south dome. Also playing a role in this process is the decrease in albedo at lower elevations as snow and firn are replaced by darker ice during the melt season. Counteracting it is ice flow, which tends to raise the ablation area and depress the accumulation area. This general pattern of greatest thinning near the margin is also observed on Penny Ice Cap on Baffin Island, as well as on the ice caps or ice fields of the high Arctic (Braun et al., 2004; Abdalati et al., 2004). These observations point to the importance of increased summer temperatures in driving ice cap wastage.

CONCLUSIONS

The south dome of Barnes Ice Cap has been thinning at an average rate of ~0.76 m yr^{-1} for the past 22 yr. This is seven times the thinning rate between 1970 and 1984. In 2004 the rate appears to have accelerated still further, to ~1.0 m yr^{-1}, although we obviously do not know whether this high rate will be sustained. The thinning is consistent with that expected from mean summer temperatures and a correlation between mean summer temperature and mass balance established with data from 1963 to 1983. As expected, the thinning rate was highest near the margin. If the projections for global warming over the next century become a reality, the future is bleak for these ice masses. The consequent change in the pattern of summer runoff will likely have ecological implications.

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