Surface roughness over the northern half of the Greenland Ice Sheet from airborne laser altimetry

C. J. van der Veen,[1] Y. Ahn,[2] B. M. Csatho,[3] E. Mosley-Thompson,[2,4] and W. B. Krabill[5]

Received 14 May 2008; revised 23 October 2008; accepted 14 November 2008; published 8 January 2009.

[1] Surface roughness, defined as the standard deviation of small-scale elevation fluctuations from the linear trend over 0.5 km, can be estimated from high-resolution airborne laser altimetry. Here we present results for the northern half of the Greenland Ice Sheet using laser data collected in May 1995. Roughness is smallest in the central region straddling the ice divide, increases in amplitude toward the coast, and appears to be correlated with slope of the ice surface. For most of the study region surface roughness is 8 cm or less (<2.5 cm water equivalent). In smaller regions associated with fast flow, larger values are found. Comparison of the size of small-scale topographic disturbances with the spatial noise estimated from five closely spaced ice cores drilled in northwest Greenland shows good agreement. Similar correspondence was found earlier using nine ice cores from the Summit region. These results indicate that the airborne laser altimeter provides an efficient platform for characterizing the statistical nature of the snow surface over large areas of the polar ice sheets.

Citation: van der Veen, C. J., Y. Ahn, B. M. Csatho, E. Mosley-Thompson, and W. B. Krabill (2009), Surface roughness over the northern half of the Greenland Ice Sheet from airborne laser altimetry, *J. Geophys. Res.*, *114*, F01001, doi:10.1029/2008JF001067.

1. Introduction

[2] The surface morphology of ice sheets is characterized by variations on scales ranging from a few centimeters to several kilometers. Larger-scale undulations with a typical wavelength of 3–4 times the ice thickness result from ice flow over basal topography [e.g., *Budd and Carter*, 1971; *Whillans and Johnsen*, 1983] or may be regularly patterned dune-like features, or "megadunes," occupying the low accumulation area of the East Antarctic Plateau [*Swithinbank*, 1988; *Fahnestock et al.*, 2000]. These topographic features are stationary and predictable. In contrast, small-scale surface irregularities associated with sastrugi and other wind-sculpted features, form over periods of hours to days and are inherently transient and unpredictable in a deterministic sense.

[3] Knowledge of the statistical properties of small-scale spatial irregularities, otherwise referred to as surface roughness, is important for a number of applications. First, interpretation of annually resolved climate records derived from ice cores is complicated by noise contained in these records. Inferred annual accumulation rate shows great variability as a result of interannual climate variations superimposed on small-scale transient irregularities in the snow surface and preserved in the core record. To separate both contributions, either multiple closely spaced core records [*van der Veen and Bolzan*, 1999], or some independent estimate of the spatial noise contribution are needed [*van der Veen et al.*, 1998]. Second, knowledge of surface microrelief is a prerequisite for the analysis of satellite data. The return signal of radar and laser altimeters is influenced by the microrelief of the reflecting surface [e.g., *Gardner*, 1992; *Davis and Moore*, 1993; *Yi and Bentley*, 1994; *Rees*, 2001; *Zwally et al.*, 2002]. In the visible wavelength domain surface roughness affects the albedo and the bidirectional reflectance distribution function (BRDF) of snow and ice surfaces. *Warren et al.* [1998] measured the effect of orientation of sastrugi relative to the solar zenith angle on the BRDF and found that bidirectional reflectance measurements best represent the albedo when viewed at nadir or within 20° of nadir. *Leroux and Fily* [1998] developed a snow-reflectance model assuming sastrugi are regularly spaced rectangular protrusions with the same orientation. *Leroux et al.* [1998] found generally good qualitative agreement between ground-based measurements of bidirectional reflectance of snow and model predictions, although quantitative agreement was lacking. From these studies it is clear that characterization of small-scale ice sheet topography is of importance for many diverse remote-sensing applications. Third, as suggested by *Herzfeld et al.* [1999], spatial patterns in surface roughness may contain information about dynamical processes acting on glaciers, and their interaction with climatic processes. Finally, surface roughness affects the exchange coefficients in the standard formulations for the turbulent fluxes of momen-

[1]Department of Geography and Center for Remote Sensing of Ice Sheets, University of Kansas, Lawrence, Kansas, USA.
[2]Byrd Polar Research Center, Ohio State University, Columbus, Ohio, USA.
[3]Department of Geology, University at Buffalo, State University of New York, Buffalo, New York, USA.
[4]Department of Geography, Ohio State University, Columbus, Ohio, USA.
[5]Goddard Space Flight Center, Wallops Flight Facility, Wallops Island, Virginia, USA.

0148-0227/09/2008JF001067

tum, sensible, and latent heat, thereby impacting the energy balance of the upper snow surface [*Oerlemans*, 2001, sect. 3.8].

[4] The literature is somewhat confusing when it comes to defining surface roughness. *Oerlemans* [2001, section 3.8] and *Davis and Zwally* [1993] use the root-mean-square surface roughness, whereas *Yi et al.* [2005] define surface roughness as the standard deviation of the differences between measured elevations and a fitted line. Where the average surface is horizontal, both definitions provide the same result. Other quantities have been proposed to characterize the surface roughness, including the correlation length, the root-mean-square of slope deviations, and the semivariogram [*Rees and Arnold*, 2006; *Herzfeld et al.*, 1999, 2000]. For the purpose of investigating variations in surface roughness and potential effects on ice core records and interpretation of radar altimeter data, prior studies considered the standard deviation of surface perturbations from a linear trend. This approach, adopted here, is consistent with theoretical models for radar and laser backscattering, in which surface roughness is defined as height deviations from a sloping planar surface [*Gardner*, 1992; *Davis and Moore*, 1993; *Yi and Bentley*, 1994].

[5] There are few measurements available that relate to small-scale topography on ice sheets, primarily because traditional methods used to map topography at high spatial resolution are labor intensive and necessarily restricted to small areas. *Long* [1961] determined relative elevations in a 100 by 100 m study area near Byrd Station, West Antarctica, from 250 dowels placed in the firn, ~6 m apart. Near Dome C, East Antarctica, *Palais* [1980] measured relative elevations at six different times in a "bamboo forest" consisting of poles placed 3 m apart in a 10 by 10 grid. The measurement spacing in these studies is too great to provide estimates of surface roughness on spatial scales of centimeters and up. *Herzfeld et al.* [1999, 2000] designed and employed a multichannel instrument (the "Glacier Roughness Sensor") to measure surface roughness on Jakobshavn Isbræ, west Greenland, at a resolution of 20 cm across track and at 1 cm vertical accuracy. Surveys were carried out in areas of 175 m length (along-track direction) and 75 m width (across-track direction), deemed optimal to typify characteristic morphological features, yet small enough to be surveyed by a person-hauled instrument. *Jezek* [2007] employed a 1-m-long comb gauge as well as smaller handheld comb gauges to obtain quantitative estimates of surface roughness in southwest Greenland. *Rees and Arnold* [2006] used an airborne lidar system together with ground-based surveys to investigate the roughness of the snow-free surface of Midre Lovénbreen, Svalbard, on spatial scales ranging from 1 mm to 300 m.

[6] The major drawback of traditional methods used to map surface roughness is their labor intensity and the comparatively high costs of logistics required to survey a small region. With the advent of accurate remote-sensing techniques, it is now possible to study the morphology of larger areas of the polar ice sheets more expediently. *Partington et al.* [1989] derived estimates of surface roughness for the Wilkes Land plateau in East Antarctica by fitting models for satellite radar altimeter returns to measured averaged wave forms. A similar study was conducted by *Davis and Zwally* [1993] who compared inferred surface roughness over the Greenland and Antarctic ice sheets. As noted in that study, the accuracy of surface roughness estimates depends on how well the contributions from surface scattering and volume scattering to the radar return signal can be separated. *Yi and Bentley* [1994] developed a model on the basis of a variable combination of surface and volume scattering to derive surface roughness from radar altimetry over the inland parts of the East Antarctic Ice Sheet. More recently, *Yi et al.* [2005] used 8 day repeat orbit data from the Ice, Cloud, and land Elevation Satellite (ICESat) to produce a map of surface roughness for the entire Greenland Ice Sheet. Simultaneous multiangular measurements, obtained by the Multiangle Imaging SpectroRadiometer (MISR) aboard NASA's Terra satellite, were used by *Nolin et al.* [2002] and *Nolin and Payne* [2007] to derive proxies of ice sheet and sea ice surface roughness from backward and forward-scattered radiances.

[7] In the studies referred to above, surface roughness is defined as the standard deviation of height deviations from a local or regional trend (usually assumed to be linear). Values derived in various studies differ substantially. For example, *Davis and Zwally* [1993] report roughness values ranging between 15 and 48 cm along an East Antarctic transect, whereas *Partington et al.* [1989] found a range from 70 to 160 cm. Part of these differences may result from different averaging scales considered in the various studies. However, at the present, it is not clear how to evaluate the relation between estimated values from satellite sensors and the true variation of surface roughness [*Davis and Zwally*, 1993].

[8] In a previous study, *van der Veen et al.* [1998] used high-resolution airborne laser altimetry to determine surface roughness in the Summit region, central Greenland. They found a standard deviation of the surface roughness of 1.6 cm water equivalent (weq). Because annual layers in ice cores are bounded by two buried surface topographies, the corresponding standard deviation of annual layer thickness is 2.3 cm weq. This estimate is in agreement with an independent assessment of the nine shallow ice cores available from that region, which indicates a spatial variability of 2.5 cm weq [*van der Veen and Bolzan*, 1999]. The present study extends the work of *van der Veen et al.* [1998] by analyzing additional laser-altimeter data over the northern half of the Greenland Ice Sheet. A further comparison between altimeter-derived roughness and spatial variability estimated from five ice cores in the Humboldt region of northwest Greenland is presented to demonstrate the applicability of our methods to the study of roughness over polar ice sheets.

2. Data Acquisition, Analysis, and Validation

2.1. Airborne Laser

[9] Surface elevation measurements of most representative regions of the Greenland Ice Sheet were obtained as part of the NASA Program for Arctic Climate Assessment (PARCA), using airborne laser altimetry several times since the early 1990s. The conical scanning Airborne Topographic Mapper (ATM) was used for most of the missions, but during May 1995, one of the two available laser systems was deployed in a profiling mode. These data (which include the two smaller segments considered by *van der Veen et al.* [1998]) are used here to estimate surface

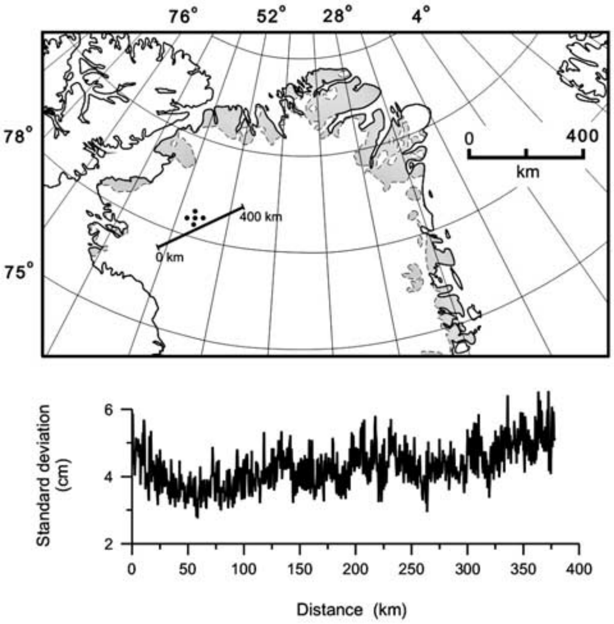

Figure 1. (top) Location map showing the five Humboldt core sites and the nearest ATM flightline. (bottom) Surface roughness along the ATM flightline.

roughness over the northern part of the Greenland Ice Sheet. At a nominal operating altitude of 400 m, the illuminated footprint is near circular with a diameter of ~1 m. With a pulse rate of 2000 Hz and a nominal aircraft speed of 125 m/sec, the resultant footprint spacing is ~6 cm.

[10] After removing obvious outliers from the ranging data, consecutive averages of five adjacent points were taken to reconstruct the elevation of the snow surface. Other averaging schemes to minimize random noise contained in the data produced essentially the same results. Surface roughness was derived following the procedure described by *van der Veen et al.* [1998]. That is, all flightline segments are divided into 0.5 km sections over which the large-scale slope is considered constant, allowing the average surface to be determined by least squares fitting a straight line to the data in the 0.5 km section. Differences between the actual elevations and the best fit regression line are considered to represent small-scale perturbations. The standard deviation of these perturbations is taken as a measure of the roughness of the surface.

2.2. Humboldt Ice Cores

[11] In support of the PARCA project, five cores were drilled in 1995 at a remote site identified as Humboldt after the major outlet glacier in the region. The central core (78.527°N; 56.83°W; 1995 m above sea level) was 146.5 m long and extended back to 1153 A.D. Four shorter (~21 m long) cores, at a distance of 25 km from the central core in each of the four radial directions, were drilled to explore the spatial variability of the preserved records (Figure 1 (top)). Each core was carefully dated using the well preserved seasonal variations in the concentration of insoluble dust (measured at Byrd Polar Research Center) and calcium (measured at the University of Arizona). *Mosley-Thompson et al.* [2001, Figure 3] illustrate the seasonal variations in dust content that provided excellent timescales. The thickness of each year's accumulated mass is determined from successive dust concentration minima characteristic of winter precipitation, using density (obtained from measured core dimensions and weight) to convert layer thickness to water equivalent. The seasonal variations in δ^{18}O were not preserved beyond a few years because of the low accumulation rate (~14 to 15 cm/a weq, where a is years) in this region. The annual dating of cores was also supported by the identification of the well known 1952 and 1963 peaks in Beta radioactivity from atmospheric thermonuclear testing. The five cores contain a 66-year period of overlap from 1928 to 1994 (the most recent complete year of accumulation). Annual accumulation rates for each of the cores are shown in Figure 2.

2.3. Validation

[12] *Van der Veen et al.* [1998] found good agreement between laser-derived surface roughness and the estimated spatial variability preserved in nine ice cores covering an 160 × 160 km area in the Summit region [*van der Veen and*

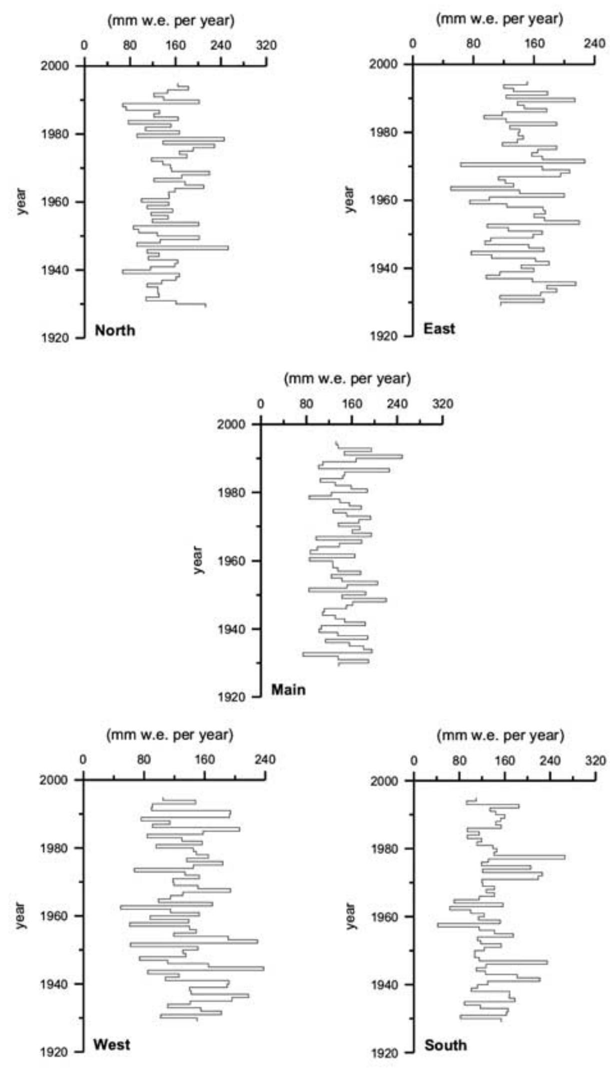

Figure 2. Annual accumulation measured in the five Humboldt ice cores.

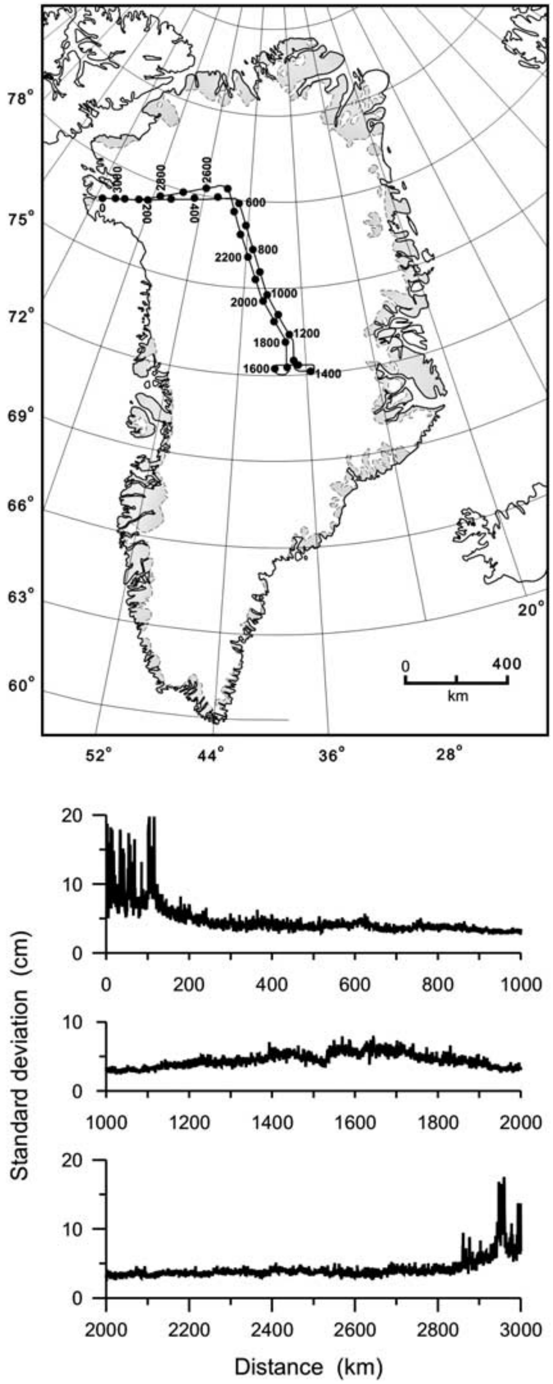

Figure 3. Location map and calculated surface roughness along the 18 May 1995 flightline.

Bolzan, 1999]. A further validation of the current approach is possible using the five Humboldt cores.

[13] The procedure for separating interannual variability and spatial noise from multiple neighboring cores is described in detail by *van der Veen and Bolzan* [1999]. In short, the assumption is made that each record of annual accumulation consists of a long-term average accumulation signal (unique to each core), interannual fluctuations in snowfall associated with synoptic-scale variability (assumed to be uniform over the region covered by the cores, and thus common to all five cores), spatial noise associated with microrelief of the snow surface which is unique to each core, plus random measurement errors. The average signal may include slowly varying climate trends and effects from large-scale surface undulations, and is obtained by averaging annual accumulations for each core over the length of the core record. Next, interannual fluctuations are estimated by subtracting this average from each core record, and averaging the resulting five records of annual deviations. Finally, subtracting these interannual fluctuations from annual deviations of each core record leaves the spatial and measurement noise preserved in each record. It should be noted that, because both spatial noise and measurement errors are random in space and time, these two contributions cannot be separated in a deterministic sense. However, if the standard deviation associated with uncertainties in the measurement can be estimated independently, the standard deviation associated with spatial noise can be inferred from the core records [*van der Veen and Bolzan*, 1999].

[14] The greatest source of uncertainty in annual accumulation rates derives from ambiguity in the timing of winter dust concentration minima. Following *van der Veen and Bolzan* [1999], the standard deviation of the timing of each annual stratigraphic horizon is taken to be one month. With an average accumulation rate in this region of 15 cm/a weq, the associated measurement error is 2.0 cm weq. From the core records, the standard deviation associated with regional interannual fluctuations in accumulation is 1.7 cm weq. Following the procedure outlined above, this yields a standard deviation for the spatial noise of 2.5 cm weq.

[15] The location of the flightline segment nearest to the Humboldt cores is shown in Figure 1 (top); surface roughness along this line is shown in Figure 1 (bottom). For this segment, the standard deviation of the actual elevation deviations is 4.8 cm. To convert this value to water equivalent, a near-surface density of 330 kg/m^3 is adopted on the basis of surface density determined at the Humboldt core sites. This gives 1.7 cm weq as the estimate for spatial noise. In core records, each annual layer is bounded by two undulating surfaces, and the standard deviation associated with spatial noise is 2.4 cm weq, which is essentially the same as obtained from the five core records.

[16] To summarize, analysis of the Humboldt cores, as well as the earlier analysis of the more centrally located Summit cores [*van der Veen and Bolzan*, 1999], show good agreement between estimates of surface roughness from core records and from airborne laser altimetry. This suggests that the procedure developed by *van der Veen et al.* [1998] and adopted here, to evaluate surface roughness from laser-altimeter data provides a good estimate of this quantity over the entire Greenland Ice Sheet, with the possible exception of near coastal regions (see discussion below).

3. Surface Roughness Over Northern Greenland

3.1. Flightline Results

[17] Laser altimetry missions were flown on 4 days in May 1995 (18, 22, 23, and 24 May). Calculated surface roughness (expressed as standard deviation of actual surface topography from a linear trend; note that these values have not been converted to water equivalent) is shown in Figures 3–6, each corresponding to one of the 4 days laser

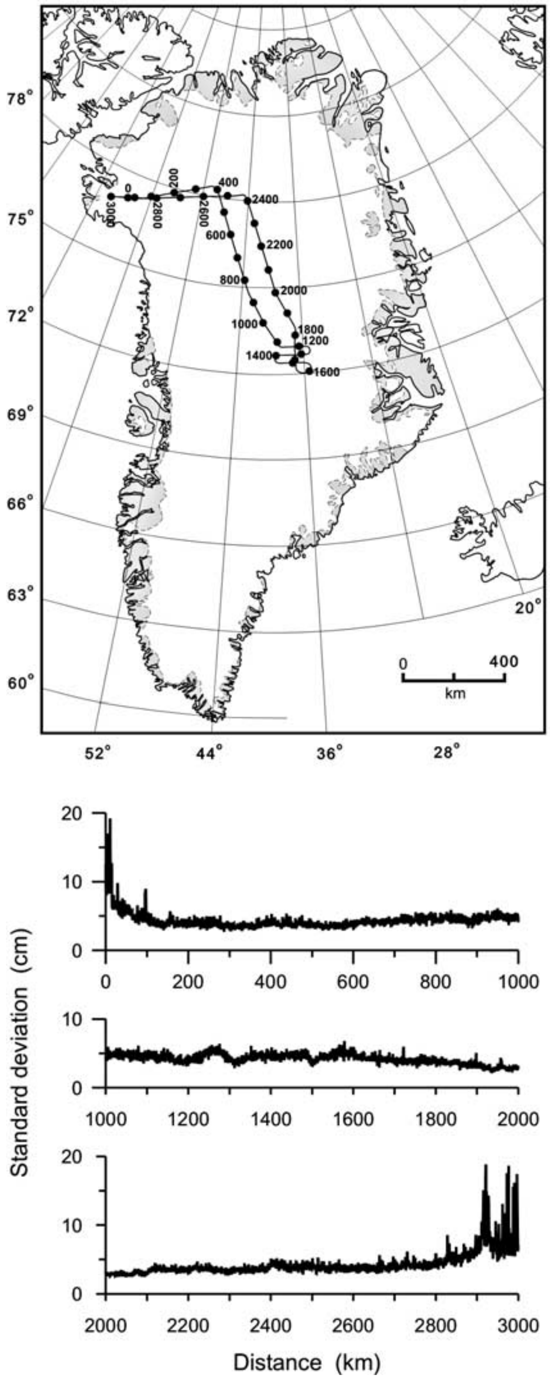

Figure 4. Location map and calculated surface roughness along the 22 May 1995 flightline.

altimetry was conducted. Surveys consisted of a closed loop and the distance along this loop (indicated on the maps) is used as a spatial coordinate, except for the 24 May survey, which is split into two separate flightlines, omitting noisy data over the heavily crevassed lower reaches of Jakobshavn Isbræ.

[18] Surface roughness is greatest at the beginning and end of the flightlines, corresponding to regions of lower ice surface elevation where ice speeds are generally larger and the surface may be dissected by crevasses. Note that the

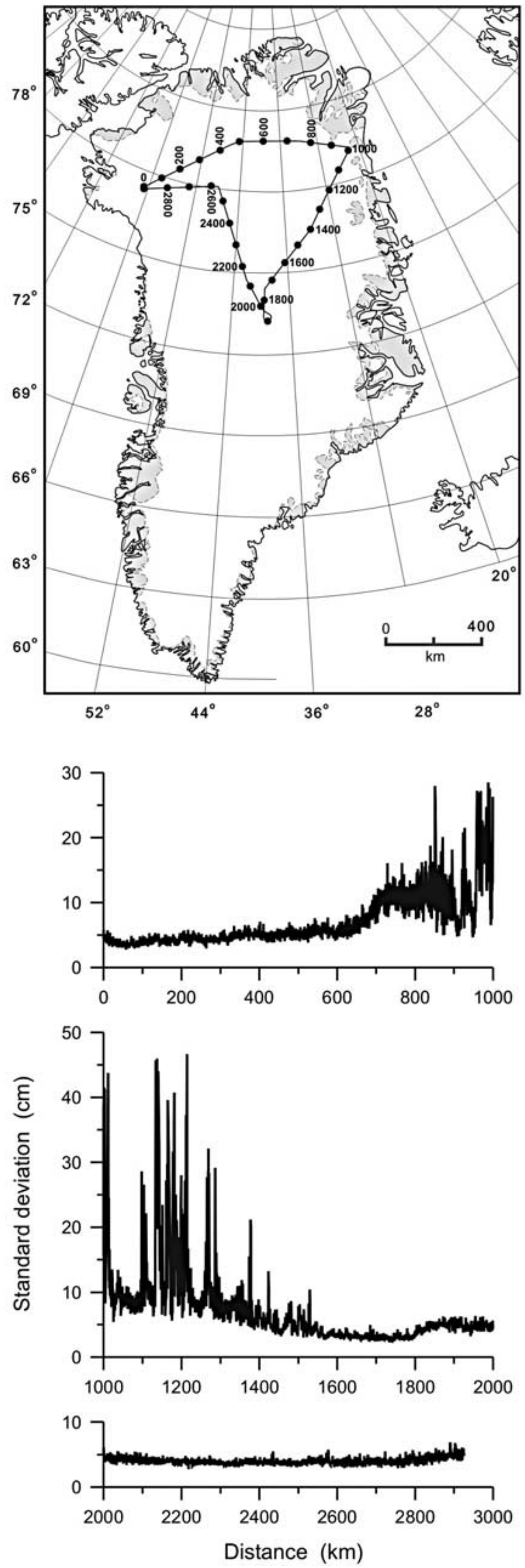

Figure 5. Location map and calculated surface roughness along the 23 May 1995 flightline.

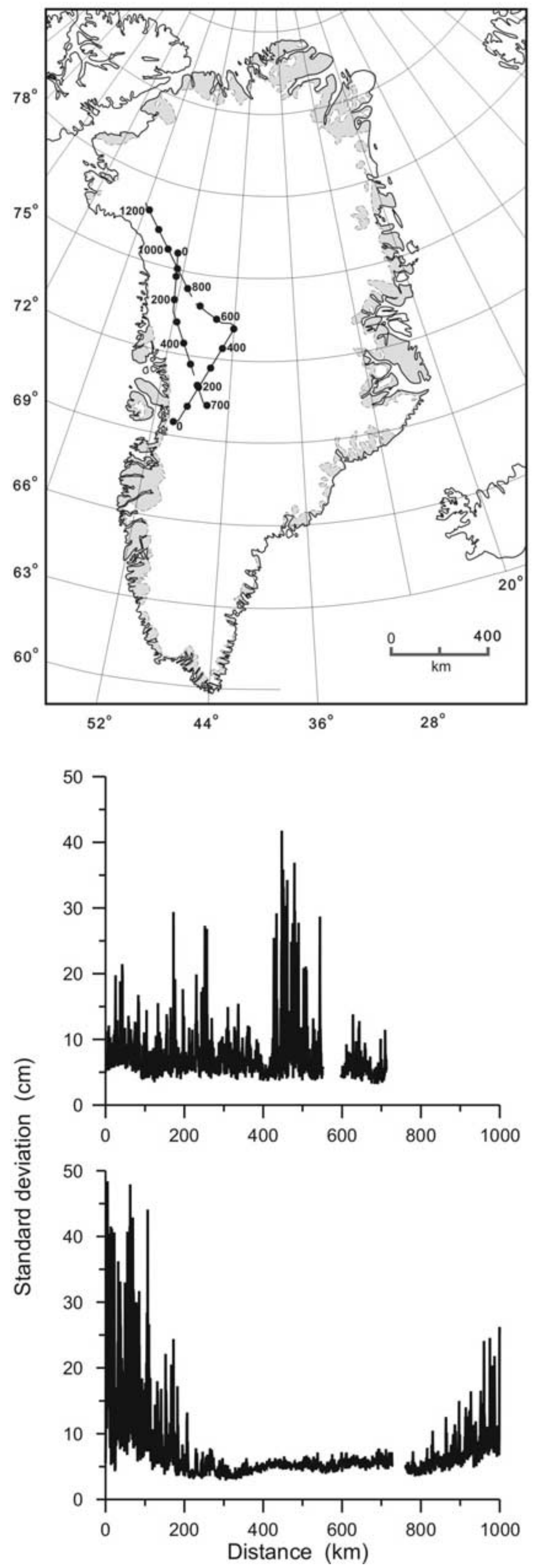

Figure 6. Location map and calculated surface roughness along the 24 May 1995 flightline. Data gaps are in regions where laser energy did not reach the surface because of low cloud or fog.

24 May survey was primarily restricted to lower elevations along the west coast of Greenland, where ice surface slopes are generally larger than in the interior. Consequently, these profiles (Figure 6) exhibit a much greater variability than the other profiles. For most of the interior, however, the standard deviation or surface roughness is rather uniform. An exception occurs in northeast Greenland (Figure 5, from ~800 to ~1200 km) where the surface is exceptionally rough. This locally rough surface is associated with the Northeast Ice Stream.

[19] *Van der Veen et al.* [1998] considered two 100-km-long flightline segments in the Summit region. Their north-south transect was approximately aligned with the local ice crest and showed a slightly smaller roughness than the perpendicular east–west trending segment. This difference may be more coincidental than signifying a true directional effect on surface roughness. Inspection of the roughness profiles shown in Figures 3–6 shows similar subtle differences along flightlines oriented in the same direction. The lack of systematic variations along surveys where the flight path changes directions suggests there is no obvious directional dependence of surface roughness.

3.2. Spatial Patterns

[20] Combining results from all four flightlines, a contour map showing the spatial pattern of roughness over the ice sheet can be produced. This map is shown in Figure 7 and

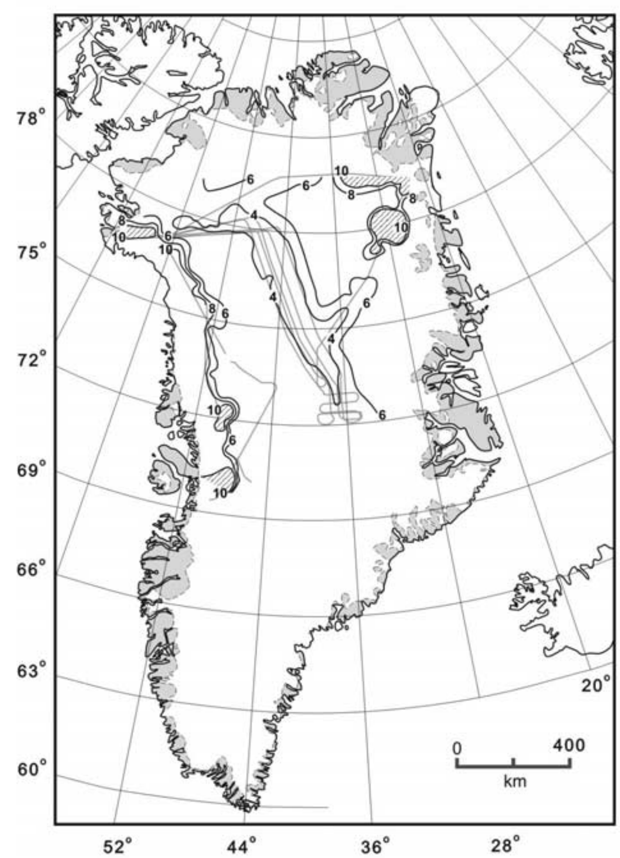

Figure 7. Contour map of surface roughness (in cm) over the northern half of the Greenland Ice Sheet. Flightlines are shown in light gray.

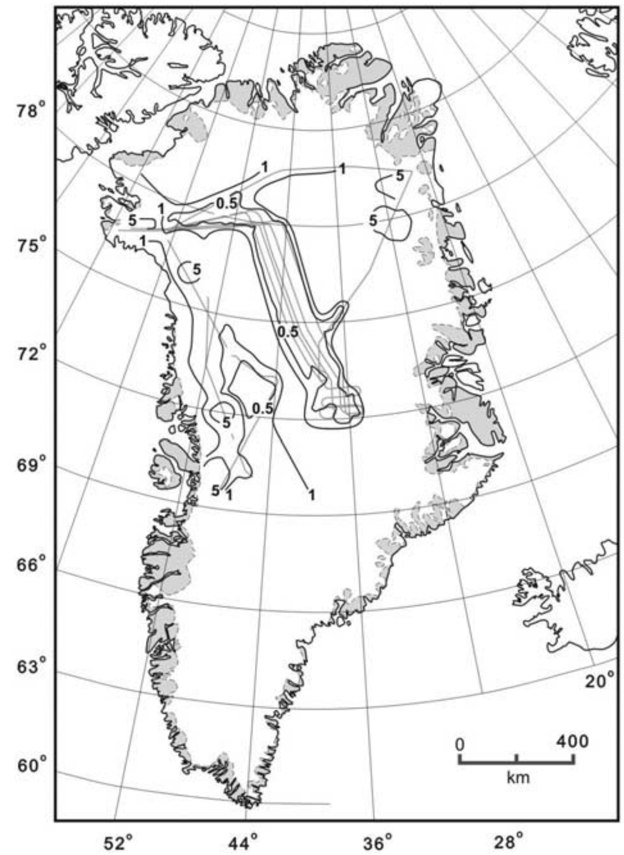

Figure 8. Contour map of the standard deviation of surface roughness (in cm) over the northern half of the Greenland Ice Sheet. Flightlines are shown in light gray.

reiterates the comparative homogeneity of surface roughness. Roughness is least pronounced in the central region straddling the ice divide, and increases in amplitude toward the coast. For most of northern Greenland, surface roughness is 8 cm or less (<2.5 cm weq). Larger values are found in smaller regions associated with fast flow of Jakobshavn Isbræ and the Northeast Ice Stream. The region of large variability centered on 72^{o}N may reflect the onset of fast flow on Rinks Isbræ or could be associated with the local maximum in surface accumulation [*Bales et al.*, 2001]. In this region, relatively easy access of humid air masses passing through the Disko Bugt area onto the ice sheet may enhance accumulation [*Reeh*, 1989] and increase surface relief.

[21] To evaluate the variability in surface roughness in more detail, the standard deviation of roughness is considered. This standard deviation is calculated for overlapping 25-km-long segments of the profiles shown in Figures 3–6. A contour map of the results is presented in Figure 8. As was expected from inspection of the profiles, variability in surface roughness is smallest near the ice divide (standard deviation less than 0.5 cm) whereas the greatest variability occurs on Jakobshavn Isbræ and the Northeast Ice Stream.

[22] The apparent colinearity between contours shown in Figures 7 and 8 and the orientation of flightlines is not an artifact of the contouring method used. Contouring was done using ordinary kriging to interpolate the data to a regular grid. Other methods, such as minimum curvature, yielded essentially the same maps. Flightlines considered in this study mostly followed elevation contours which explains why the contours reflect the paths of flightlines.

4. Discussion

[23] Comparison of the contour map of roughness obtained here (Figure 7) with the map produced by *Yi et al.* [2005] shows that the spatial pattern is similar, following closely their map of surface slope. That is, smallest roughness values are found in the interior where slopes are small, with a gradual increase toward the steeper coastal regions. Actual roughness values derived by *Yi et al.* [2005] are much greater (up to several meters) than those shown in Figure 7. The main reason for this difference appears to be that *Yi et al.* [2005] used a 10 km window to define the local linear trend. Their roughness values range from 0.1 to 0.2 m in central Greenland to more than 10 m at the northern edge. Because of the much longer segments through which a straight line is fitted, surface undulations with wavelengths of several km contribute to these roughness estimates. Using a smaller averaging window (0.5 km in this study) reduces this contribution and provides a better estimate for small-scale spatial noise.

[24] Several studies have suggested that the roughness is better described by a spectrum of scales [e.g., *Burrough*, 1981, 1986; *Gilbert*, 1989]. *Arnold and Rees* [2003] suggest that surface roughness on a glacier exhibits self-similar or fractal behavior at a variety of spatial scales, up to tens of meters. *Rees and Arnold* [2006] investigated the roughness of the snow-free Midre Lovénbreen, Svalbard, on scales ranging from 1 mm to 300 m, and found that the roughness can be described by scale-free or fractal models at spatial scales less than $\sim$0.1 m, or greater than a few meters. Our choice for 0.5 km segments to estimate surface roughness reflects a compromise between the need to sample a sufficiently large number of sastrugi in each section, and the need to minimize effects from less transient topography. Calculated roughness values change slightly if the length of the window is increased or decreased by several hundred meters. The agreement with roughness estimates from ice cores provides some a posteriori justification for the choice of 0.5 km segments.

[25] In comparing the laser altimeter data, which apply to a 5-day period in May 1995, with the annually resolved core records, the implicit assumption is made that surface roughness is time invariant. Measurements conducted at Summit indicate that the amplitudes of sastrugi are several times larger in February and March than in June to February, possibly as a result of sustained strong winds and cold temperatures during the winter [*Albert and Hawley*, 2002]. On the other hand, that study found no significant seasonal variation in snow surface roughness at wavelengths less than 30 m. The more regional study of *Nolin and Payne* [2007] considered the lower regions of the Jakobshavn Isbræ drainage basin. For most of the study area they found surface roughness increasing over the course of the summer. At the highest elevations considered, however, seasonal variations were found to be minimal. The present study excludes the low elevation periphery of the ice sheet, and the assumption of temporal stationarity may be valid. This

issue warrants further investigation where seasonal roughness estimates can be obtained.

5. Conclusions

[26] When used in profiling mode, the airborne NASA ATM provides an efficient platform for characterizing the statistical nature of the snow surface over large areas of the polar ice sheets. Surface roughness, defined here as the standard deviation of height differences from a linear trend over 0.5 km, derived from laser altimeter measurements agrees favorably with estimates derived from multiple closely spaced ice core records. While a quantitative correlation was not attempted, comparison between the surface roughness map derived in this study and the map of ice surface slope from *Yi et al.* [2005] reveals similar spatial distribution. Results for the northern half of the Greenland Ice Sheet show small roughness values in the low slope interior regions straddling the ice divide, and increasing values toward the steeper coastal regions. Localized regions of particularly high roughness appear to be associated with regions of fast flow.

[27] **Acknowledgments.** This research was supported in part by the National Science Foundation under grant 0424589 (KU), by NASA grant NAG5–5032 (OSU), and NASA's ICESat program (UB and KU). Funding for laser data collection was from the PARCA Program under the NASA Cryospheric Sciences Program. We would like to thank Gordon Hamilton, Robert Hawley, and two anonymous reviewers for valuable comments and feedback that helped improve the paper significantly.

References

Albert, M. R., and R. L. Hawley (2002), Seasonal changes in snow surface roughness characteristics at Summit, Greenland: Implications for snow and firn ventilation, *Ann. Glaciol.*, *35*, 510–514, doi:10.3189/172756402781816591.

Arnold, N. S., and W. G. Rees (2003), Self-similarity in glacier surface characteristics, *J. Glaciol.*, *49*, 547–554, doi:10.3189/172756503781830368.

Bales, R. C., J. R. McConnell, E. Mosley-Thompson, and B. Csatho (2001), Accumulation over the Greenland Ice Sheet from historical and recent records, *J. Geophys. Res.*, *106*, 33,813–33,825, doi:10.1029/2001JD900153.

Budd, W. F., and D. B. Carter (1971), An analysis of the relation between the surface and bedrock profiles of ice caps, *J. Glaciol.*, *10*, 197–209.

Burrough, P. A. (1981), Fractal dimensions of landscapes and other environmental data, *Nature*, *294*, 240–242, doi:10.1038/294240a0.

Burrough, P. A. (1986), *Principles of Geographic Information Systems for Land Resources Assessment*, *Monogr. Soils Resour. Surv.*, vol. 12, Oxford Univ. Press, Oxford.

Davis, C. H., and R. K. Moore (1993), A combined surface- and volume-scattering model for ice-sheet radar altimetry, *J. Glaciol.*, *39*, 675–686.

Davis, C. H., and H. J. Zwally (1993), Geographic and seasonal variations in the surface properties of the ice sheets by satellite-radar altimetry, *J. Glaciol.*, *39*, 687–697.

Fahnestock, M. A., T. A. Scambos, C. A. Shuman, R. J. Arthern, D. P. Winebrenner, and R. Kwok (2000), Snow megadune fields on thee Antarctic plateau: Extreme atmosphere-ice interaction, *Geophys. Res. Lett.*, *27*, 3719–3722, doi:10.1029/1999GL011248.

Gardner, C. S. (1992), Ranging performance of satellite laser altimeters, *IEEE Trans. Geosci. Remote Sens.*, *30*, 1061–1072, doi:10.1109/36.175341.

Gilbert, L. E. (1989), Are topographic data sets fractal?, *Pure Appl. Geophys.*, *131*, 241–254, doi:10.1007/BF00874489.

Herzfeld, U. C., H. Mayer, W. Feller, and M. Mimler (1999), Glacier roughness surveys of Jakobshavn Isbræ drainage basin, west Greenland, and morphological characterization, *Z. Gletscherkd. Glazialgeol.*, *35*, 117–146.

Herzfeld, U. C., H. Mayer, W. Feller, and M. Mimler (2000), Geostatistical analysis of glacier-roughness data, *Ann. Glaciol.*, *30*, 235–242, doi:10.3189/172756400781820769.

Jezek, K. C. (2007), Surface roughness measurements on the western Greenland Ice Sheet, *Tech. Rep. 07-01*, 20 pp., Byrd Polar Res. Cent., Columbus, Ohio.

Leroux, C., and M. Fily (1998), Modeling the effect of sastrugi on snow reflectance, *J. Geophys. Res.*, *103*, 25,779–25,788, doi:10.1029/98JE00558.

Leroux, C., J.-L. Deuzé, P. Goloub, C. Sergent, and M. Fily (1998), Ground measurements of the polarized bidirectional reflectance of snow in the near-infrared spectral domain: Comparison with model results, *J. Geophys. Res.*, *103*, 19,721–19,731, doi:10.1029/98JD01146.

Long, W. E. (1961), Glaciology, Byrd Station and Marie Byrd Land traverse, 1958–1959, *Rep. 2*, Ohio State Res. Found., Columbus, Ohio.

Mosley-Thompson, E., J. R. McConnell, R. C. Bales, Z. Li, P.-N. Lin, K. Steffen, L. G. Thompson, R. Edwards, and D. Bathke (2001), Local to regional-scale variability of Greenland accumulation from PARCA cores, *J. Geophys. Res.*, *106*, 33,839–33,851, doi:10.1029/2001JD900067.

Nolin, A. W., and M. C. Payne (2007), Classification of glacier zones in western Greenland using albedo and surface roughness from the Multi-angle Imaging Spectroradiometer (MISR), *Remote Sens. Environ.*, *107*, 264–275, doi:10.1016/j.rse.2006.11.004.

Nolin, A. W., F. M. Fetterer, and T. A. Scambos (2002), Surface roughness characterizations of sea ice and ice sheets: Case studies with MISR data, *IEEE Trans. Geosci. Remote Sens.*, *40*, 1605–1615, doi:10.1109/TGRS.2002.801581.

Oerlemans, J. (2001), *Glaciers and Climate Change*, 148 pp., A. A. Balkema, Lisse, Netherlands.

Palais, J. M. (1980), Snow stratigraphic investigations at Dome C, east Antarctica: A study of depositional and diagenetic processes, M.S. thesis, 146 pp., Ohio State Univ., Columbus, Ohio.

Partington, K. C., J. K. Ridley, C. G. Rapley, and H. J. Zwally (1989), Observations of the surface properties of the ice sheets by satellite radar altimetry, *J. Glaciol.*, *35*, 267–275.

Reeh, N. (1989), Dynamic and climatic history of the Greenland Ice Sheet, in *Quaternary Geology of Canada and Greenland*, edited by R. J. Fulton, pp. 793–822, Geological Survey of Canada, Ottawa.

Rees, W. G. (2001), *Physical Principles of Remote Sensing*, 2nd ed., 343 pp., Cambridge Univ. Press, Cambridge, U. K.

Rees, W. G., and N. S. Arnold (2006), Scale-dependent roughness of a glacier surface: Implications for radar backscatter and aerodynamic roughness modeling, *J. Glaciol.*, *52*, 214–222, doi:10.3189/172756506781828665.

Swithinbank, C. (1988), Satellite image atlas of the world, Antarctica, *U.S. Geol. Surv. Prof. Pap. 1386-B*, U.S. Geol. Surv., Washington, D. C.

van der Veen, C. J., and J. F. Bolzan (1999), Interannual variability in net accumulation on the Greenland Ice Sheet: Observations and implications for mass balance measurements, *J. Geophys. Res.*, *104*, 2009–2014, doi:10.1029/1998JD200082.

van der Veen, C. J., W. B. Krabill, B. M. Csatho, and J. F. Bolzan (1998), Surface roughness on the Greenland Ice Sheet from airborne laser altimetry, *Geophys. Res. Lett.*, *25*, 3887–3890, doi:10.1029/1998GL900041.

Warren, S. G., R. E. Brandt, and P. O'Rawe Hinton (1998), Effect of surface roughness on bidirectional reflectance of Antarctic snow, *J. Geophys. Res.*, *103*, 25,789–25,807, doi:10.1029/98JE01898.

Whillans, I. M., and S. J. Johnsen (1983), Longitudinal variations in glacial flow: Theory and test using data from the Byrd Station Strain Network, Antarctica, *J. Glaciol.*, *29*, 78–97.

Yi, D., and C. R. Bentley (1994), Analysis of satellite radar-altimeter return wave forms over the east Antarctic ice sheet, *Ann. Glaciol.*, *20*, 137–142.

Yi, D., H. J. Zwally, and X. Sun (2005), ICESat measurements of Greenland Ice Sheet surface slope and roughness, *Ann. Glaciol.*, *42*, 83–89, doi:10.3189/172756405781812691.

Zwally, H. J., et al. (2002), ICESats laser measurements of polar ice, atmosphere, ocean, and land, *J. Geodyn.*, *34*, 405–445, doi:10.1016/S0264-3707(02)00042-X.

Y. Ahn and E. Mosley-Thompson, Byrd Polar Research Center, Ohio State University, Scott Hall Room 108, 1090 Carmack Road, Columbus, OH 43210-1002, USA.

B. M. Csatho, Department of Geology, University at Buffalo, 411 Cooke Hall, Buffalo, NY 14260, USA.

W. B. Krabill, Goddard Space Flight Center, Wallops Flight Facility, Wallops Island, VA 23337, USA.

C. J. van der Veen, Department of Geography, University of Kansas, Lawrence, KS 66045, USA. (cjvdv@ku.edu)