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Winter accumulation in the percolation zone of Greenland measured by airborne radar altimeter

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[1] We here determine the surface elevation and the winter snow accumulation rate along a profile in the percolation zone of the Greenland Ice Sheet from data collected with ESA's Airborne SAR/Interferometric Radar Altimeter System (ASIRAS) in spring 2004. The altimeter data show that in addition to a backscatter peak at the air-snow interface a dominant second peak occurs. This second peak appears due to the strong scattering properties of the last summer surface layer. A robust re-tracking algorithm was developed that enables the tracking of both interfaces. Utilizing this algorithm, the winter snow thickness is estimated to 1.50 ± 0.13 m. This compares favorably with field measurements (1.43 ± 0.04 m). The snow depth estimates in combination with snow-density measurements of 420 kg m^{-3} give a mean winter mass accumulation rate of 63 cm water equivalent (w.e.) and a spatial variation of ±6 cm w.e. Furthermore a strong correlation is found between surface gradient and accumulation rate, with higher accumulation rate in flatter areas. The approach adopted here has significant potential for remote measurements of winter snow accumulation rate across ice sheets at larger spatial scales. **Citation:** Helm, V., W. Rack, R. Cullen, P. Nienow, D. Mair, V. Parry, and D. J. Wingham (2007), Winter accumulation in the percolation zone of Greenland measured by airborne radar altimeter, *Geophys. Res. Lett.*, *34*, L06501, doi:10.1029/2006GL029185.

1. Introduction

[2] The stability of the Greenland Ice Sheet in the future is of fundamental importance for society, any major changes in mass balance will clearly impact on global sea level and may impact on the strength of the ocean thermohaline circulation and Arctic climate feedbacks [*Jungclaus et al.*, 2006]. A critical component and a major source of error in the mass balance budget is net snow accumulation rate. Several studies have been applied to obtain accumulation rate in Greenland via: (1) direct measurements of firn cores, snow pits and precipitation measurements and; (2) indirect methods using remote sensing data [e.g., *Davis*, 1995; *Bolzan and Jezek*, 2000; *Drinkwater et al.*, 2001; *Munk et al.*, 2003; *Kanagaratnam et al.*, 2004; *Nghiem et al.*, 2005]. In using the first method, *Bales et al.* [2001] generated a gridded mean annual accumulation map in 20 km spatial resolution. He reported accumulation rate uncertainties in certain areas in the order of 20%. *Drinkwater et al.* [2001] and *Munk et al.* [2003] used scatterometer data from different sensors (ESCAT, NSCAT, QuickSCAT, ERS-1) in combination with various scattering models to obtain estimates of accumulation rate. However, their generated maps of accumulation rate in 25 km spatial resolution are restricted to the dry snow zone. Another method was introduced by *Nghiem et al.* [2005] for the percolation zone of Greenland. They revealed a direct correlation of the backscatter of QuickSCAT data to accumulation rate and produced maps for the 2001/2002 and 2002/2003 freezing seasons in 25 km spatial resolution. All approaches together show the possibility to obtain a rather good estimate of the mean accumulation rate for different time scales (years to decades) and at larger spatial scales (>20 km). However, estimates in accumulation rate in high accuracy and at smaller spatial scales ($10^1 – 10^4$ m) and the influence of the topography to their spatial distribution are not well known. In addition, the revealed accumulation rates are heavily loaded with uncertainties, due to the small scale variation of snow stratigraphy and topography [*Van der Veen et al.*, 2001]. Here we present measurements obtained with the Airborne SAR Interferometric Radar Altimeter System (ASIRAS). ASIRAS operates at a Ku band (13.65 GHz) carrier frequency and bandwidth of 1 GHz [*Mavrocordatos et al.*, 2004] and was used for the first time during spring 2004 [*Cullen et al.*, 2006]. This high bandwidth results in high vertical resolution, allowing mapping of single layers of the upper snow pack, including the surface. Recently, *Hawley et al.* [2006] demonstrated that for the dry snow zone of Greenland the ASIRAS measurements revealed distinct stratified layers (due to inter-annual density variations) down to a depth of 10 m below the surface. We extend those studies to the percolation zone with measurements along the EGIG-line near T05 (69°51′N 47°15′W, 1940 a.s.l., Figure 1), where previous firn core studies revealed a temporal variability in accumulation rate [*Fischer et al.*, 1995]. In the percolation zone, covering more than one third of the ice sheet, melt-freeze cycles and other complex processes of snow metamorphosis cause a distinct layering of the snow pack [*Benson*, 1962]. Strong subsurface reflections where obtained by *Zabel et al.* [1995] using ground penetrating radar and snow pit studies along a 100 m transect at Dye-2 (66.5°N, 46.3°W). Similar, more recent studies

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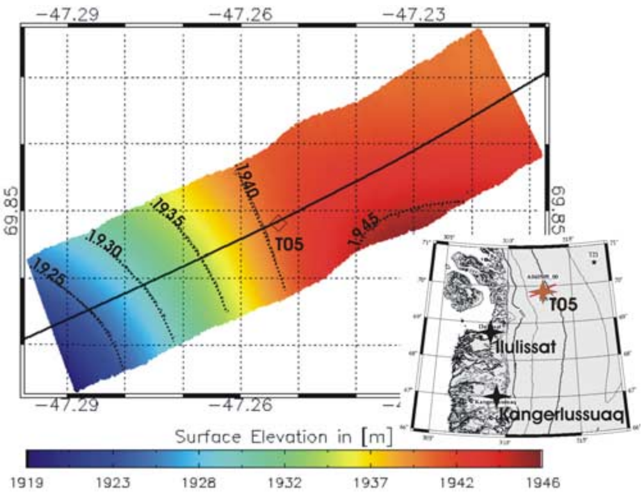

Figure 1. Laser scanner DEM of the region around T05 (diamond) with the subtrack of a 2.7 km long section of a nadir processed ASIRAS profile (black line). (inset) Map of studied area at the western flank of Greenland.

where obtained by *Scott et al.* [2006] around T05. Together, they satisfactorily showed that the uppermost strong reflector corresponded to an ice layer, which was buried with winter snow and originated from the previous summer melt. In our study we investigate the influence of the uppermost melt horizon on the ASIRAS echo and develop a new technique to measure winter snow-accumulation rate over extended areas of the percolation zone by using ASIRAS data and an adapted re-tracking technique.

2. Acquisition and Processing of Airborne and Ground-Based Measurements

[3] Airborne measurements were carried out by Alfred Wegener Institute for Polar and Marine Research (AWI) and ESA using the AWI Do-228 aircraft (Polar4) along a profile in western Greenland during the spring of 2004. The Polar4 was equipped with ASIRAS, an Airborne Laser Scanner (ALS), a single beam laser, an inertial navigation system and two GPS receivers. A ground station at Ilulissat was used for differential GPS post-processing. The ALS was operated at 80 Hz with a scan angle of 45° and the nadir looking laser was used to check the ALS quality. Measurements were conducted at height levels between 1100 and 1200 m above the ground surface, resulting in an ALS swath width of about 800 m and an ALS footprint separation of 1 m along- and 7 m across-track. The ASIRAS footprint is 15 m across track. The high along-track resolution of 4.5 m is achieved by SAR processing [*Raney*, 1998; *Wingham et al.*, 2006]. The accuracy of the ALS range measurements were tested in conjunction with a single beam laser along a runway, and were determined to be within the instruments stated specification of ±3 cm. ASIRAS was calibrated using runway overflights and the ALS elevation model. The accuracy of ASIRAS after calibration was within ±2 cm of the ALS elevation model.

[4] During the ASIRAS overflights, simultaneous ground measurements were made at site T05 on the Greenland EGIG line (Figure 1). These measurements provide information on surface elevation and snowpack stratigraphy. Snow pits and shallow firn cores were used to characterize the density structure in the top 5 m of the snowpack and firn. Nine snow pits were dug to the previous year's summer surface, one at T05 and then at 1, 10, 100 and 1000 m intervals parallel and perpendicular to the EGIG line. These were to determine small to medium scale length changes in the density of the upper snowpack. In addition, an upward looking corner reflector was installed with its tip located 2.24 m above the snow surface to compare real surface elevation and surface elevation as measured by ASIRAS.

3. Signal Analysis — Re-Tracking of Surface and Internal Layers

[5] Preliminary analysis of ASIRAS data shows that in the dry snow zone the major part (approx. 80%) of the total backscattered energy originates from the upper 30 cm of the snowpack, whereas in spring in the percolation zone approx. 85% of the total backscattered energy is returned from below 30 cm. Furthermore, the received total power of a typical percolation zone echo is about 10 dB higher than the total power of a typical dry snow-zone echo. The reason for this disparity in backscatter is a dominant second peak in the waveform from the percolation zone. Figure 2a shows a series of normalized power echoes along a 2.7 km long profile near T05 measured in May 2004. Distorted echoes due to aircraft roll >1° were not used for this analysis. Qualitatively we can distinguish between the surface signal and a strong volume signal which appears about 1.5 m below the surface and which can be traced along the whole section. Figure 2b shows a typical power echo close to T05 where in

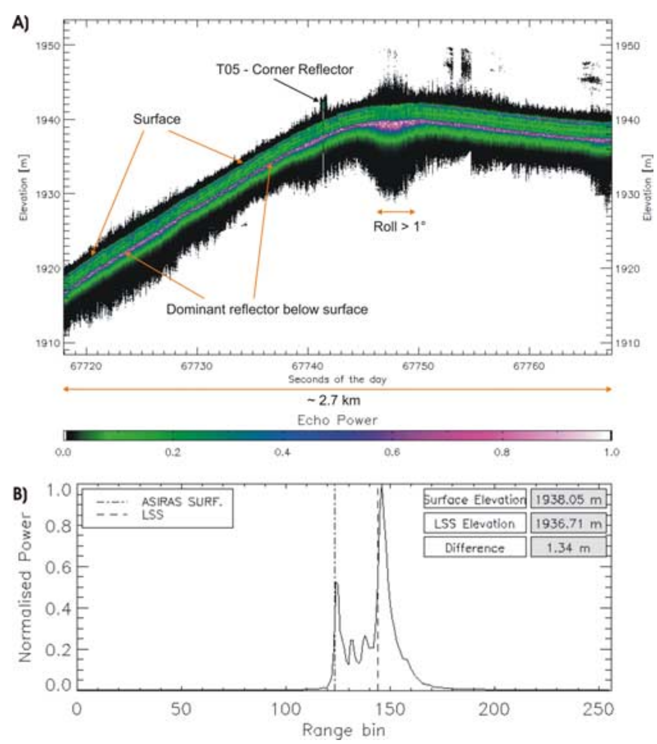

Figure 2. (a) Series of normalized ASIRAS power echoes near the test site T05. (b) A typical power echo in the percolation zone and re-tracking points of the surface and the last summer surface (LSS).

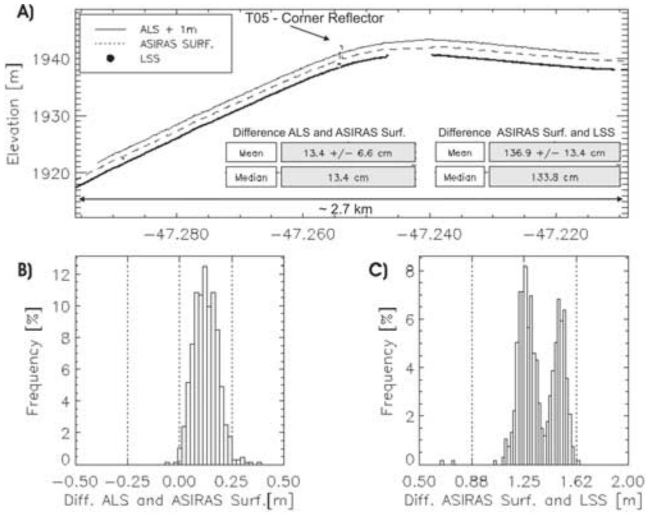

Figure 3. Percolation zone profile re-tracked with the threshold spline retracker algorithm. (a) ASIRAS surface and LSS compared to the ALS surface. (b) Histogram of surface elevation difference between ALS and ASIRAS surface. (c) Histogram of elevation differences between ASIRAS surface elevation and LSS.

addition to the surface signal (around range bin 122), a dominant volume signal (around range bin 143) appears.

[6] In order to track both the first part of the surface and the additional larger second peak, a modified re-tracker was used, termed here the Threshold-Spline Re-tracker Algorithm (TSRA). Its functionality is similar to the one of *Ferraro and Swift* [1995], where a threshold re-tracking point of 50% of the maximum peak amplitude was chosen as the surface elevation estimate. However, here it is applied to both peaks. An example for the performance of the re-tracker is shown for a typical waveform in Figure 2b. The left vertical line corresponds to the surface whereas the right vertical line shows the position of the re-tracked subsurface layer. Surface elevation statistics of the TSRA along the 2.7 km long profile show a mean difference between ALS and ASIRAS surface elevation of 0.13 ± 0.07 m (Figure 3a). This difference, also seen in the comparison of the measured and re-tracked corner reflector height, may be caused by the simplicity of the re-tracker which does not take into account the interaction of the radar altimeter pulse with the rough snow surface and the penetration of the radar pulse.

[7] The velocity of the electro-magnetic wave in snow was derived by the mixing formula of *Looyenga* [1965] using an average snow density of 420 kg m^{-3}, obtained from the snow pit measurements. This, together with the two way travel time, yields the median elevation difference between the ASIRAS surface and the lower dominant radar horizon of 1.37 ± 0.13 m (Figure 3c). Along the profile the power response of the subsurface horizon can be observed to be twice as high as the surface response, indicating a very strong density contrast between internal layers.

4. Winter Snow Accumulation Rate

[8] In the snow pits around T05, a clear melt horizon was found at 1.43 ± 0.04 m depth below the largely homogeneous winter snow pack. Similar pit characteristics and high radar backscatter were obtained by *Jezek et al.* [1994], when performing snowpit and surface-based microwave radar measurements at nearby Crawford Point (69.85°N and 47.12°W). Below the winter snowpack, a heterogeneous zone of metamorphosed snow and ice lenses is seen in shallow firn cores from T05. This change in density structure is responsible for the strong volume scattering within the ASIRAS echoes, which clearly exceeds the near surface return (Figure 2b).

[9] Small variations in the echo are visible between the surface and the last summer surface (LSS), which can be related to small density variations due to the presence of wind crusts (Figures 2a and 2b). However, these echoes have a lower amplitude than the much stronger reflection from the LSS and we therefore conclude, that the latter can be used to directly infer the amount of the winter accumulation rate within the test area. Along the test profile, a winter accumulation rate (over approx. eight months) of 63 ± 6 cm w.e. for 2003/2004 was calculated (where amount of accumulation rate is derived from elevations of ALS surface and ASIRAS-LSS and density is taken from snowpack measurements). This figure corresponds well with the average accumulation rate derived from field measurements (approx. 61 cm w.e.) in the nine snowpits.

[10] Analysis of the depth of LSS derived from ASIRAS along the test profile indicates a bimodal distribution with mean layer thickness of 1.30 m and 1.50 m, with 0.05 m and 0.03 m standard deviation, respectively (Figure 3c). If snow depths were unimodal, a difference in two way travel time corresponding to an apparent 0.2 m thickness variation could only be explained by a variation in snow density of about 60%. However, this variability in density is not observed in the winter snowpack near T05 where spatial variability of mean density is around 5%. Therefore, the difference in layer thickness in Figure 3a must be real and reflects a difference in the accumulation rate. Along the test profile, two clusters of surface gradients (derived from the ALS-DEM) are present, resulting from local ramp-like topography consisting of a slope (between 1920–1940 m) and a plateau (Figure 1). A strong correlation is found between surface gradient and accumulation rate (Figure 4), with higher accumulation rate in the 'plateau' area east of T05 (Figure 1) and lower accumulation rate at the slope. These two slope regimes finally cause the bimodal distribution in Figure 3c. Hence, the main influence on localized (<10 km) variations in winter accumulation rate are topo-

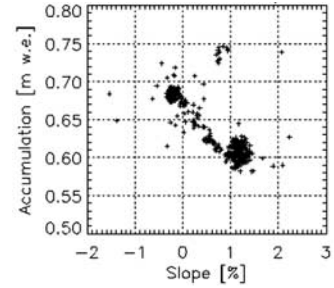

Figure 4. Dependency of the accumulation rate from the surface slope around T05. The accumulation rate is determined from the difference between ALS surface and ASIRAS last summer surface.

graphic undulations which affect snow distribution through drifting induced by strong katabatic winds.

[11] Along the test profile we observe the power response of the LSS to be twice as high as the surface response (Figures 2a and 2b), indicating a very strong density gradient between internal layers. The total backscatter is dominated by electro-magnetic wave scattering from large scatterers in the LSS, which has the effect of an almost complete loss of power for layers beneath the LSS. This indicates that measurements must be made annually, when studying temporal variations in winter accumulation rate changes in the percolation zone. The timing of measurements is also limited to a brief window before the onset of summer melt since a small amount of moisture in the surface snow significantly increases the damping factor of the electro-magnetic wave.

5. Summary and Conclusion

[12] Airborne and ground-based studies in the percolation zone of the Greenland Ice sheet reveal a clear density change between the late winter snowpack and the surface of previous summer melt. The strong density gradient at this interface generates the dominant ASIRAS echo that exceeds the surface signal. The development of a modified re-tracker enables the identification of both layers from which we can determine the surface elevation and the depth of the winter snowpack at larger spatial scales. In combination with field measurements of snow density (which can also be estimated for a late winter snowpack prior to the onset of melt), ASIRAS can therefore be used to derive winter accumulation rate. This approach could easily be extended to determine winter accumulation rates in percolation zones across poly-thermal Arctic glaciers and coastal areas of Antarctica. The technique would provide improved input data for estimates of winter accumulation for ice sheet mass balance. Therefore it could be of particular value given predictions of warming in the Arctic which would substantially increase the extent of Greenland's percolation zone [*Drinkwater et al.*, 2001]. Furthermore, the findings suggest that snow thickness on homogeneous sea ice could be determined, which is a critical parameter when estimating the sea ice thickness from freeboard measurements [*Laxon et al.*, 2003]. However, we suggest from the results shown in Figure 3c the minimal detectable snow thickness to be 0.3 to 0.4 m. Our results are therefore of value where measurements of winter accumulation rate are needed and could be greatly extended with a successful CryoSat 2 mission.

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