Importance of seasonal and annual layers in controlling backscatter to radar altimeters across the percolation zone of an ice sheet

Importance of seasonal and annual layers in controlling backscatter to radar altimeters across the percolation zone of an ice sheet

Julian B. T. Scott,[1] Peter Nienow,[2] Douglas Mair,[3] Victoria Parry,[2] Elizabeth Morris,[4] and Duncan J. Wingham[5]

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[1] Radar altimeters are one of the main tools for measuring elevation changes across the Antarctic and Greenland ice sheets and larger ice caps. A ground-based radar was deployed in autumn 2004 and spring 2006 in the percolation zone of the Greenland Ice Sheet. This radar is a high bandwidth system operating in the Ku band, the same frequency as several satellite altimeters. Measurements were made over an elevation range of 1795 to 2350 m, along with snow pit and shallow core studies. These measurements demonstrate the spatial and temporal variations in the backscatter. Relative strengths of surface and volume reflections change dramatically between spring and autumn and there is also high spatial variability across the percolation zone. The extent of percolation will affect elevation estimates made by radar altimeters. **Citation:** Scott, J. B. T., P. Nienow, D. Mair, V. Parry, E. Morris, and D. J. Wingham (2006), Importance of seasonal and annual layers in controlling backscatter to radar altimeters across the percolation zone of an ice sheet, *Geophys. Res. Lett.*, *33*, L24502, doi:10.1029/2006GL027974.

1. Introduction

[2] The past, current and future contribution of the world's ice sheets and ice caps to sea-level rise is of major interest worldwide [*Alley et al.*, 2005; *Dowdeswell*, 2006]. One of the main methods for monitoring the mass balance of these ice masses is by calculating their elevation change using satellite radar altimeters [*Johannessen et al.*, 2005; *Zwally et al.*, 2005]. The calculation of surface elevation by radar altimeter is sensitive to ice sheet topography and to radar returns from beneath the surface, termed volume backscatter [*Wingham*, 1995; *Legresy and Remy*, 1997]. The volume backscatter is highly variable both spatially and temporally. This temporal variation has seasonal and interannual components, which are difficult to account for at the satellite data level. Observations of changes in volume backscatter from ground and airborne radar measurements are needed in conjunction with ground measurements of surface and near-surface structure to determine the causes of temporal and spatial changes. Such measurements will help improve elevation change estimates based on interpretation of repeat backscatter returns received by satellite radar altimeters.

[3] Previous ground based backscatter measurements in the percolation zone of ice sheets have been limited spatially and temporally [*Jezek et al.*, 1994; *Zabel et al.*, 1995]. Backscatter measurements in the percolation zone of Greenland prior to summer melt demonstrated that a large backscatter return was received from ice layers situated at, and below, the previous summer melt horizon [*Jezek et al.*, 1994]. This return was observed to be stronger than the surface return. More recently a Ku band airborne radar altimeter employed in the dry snow zone of Greenland has proved that internal layers can be observed to a depth of around 10 m [*Hawley et al.*, 2006]. They interpret these layers as annually occurring isochrones.

[4] The work presented here formed part of the CryoSat validation activities of 2004 and 2006. Radar measurements were made between points identified as T03 and T12, and 5 km past T12, along the Expedition Glaciologique au Groenland (EGIG) line [*Fischer et al.*, 1995]. This transect spanned a distance of 106 km with an increase in elevation of over 550 m (Table 1) and is situated within the percolation zone of the Greenland Ice Sheet [*Benson*, 1962; *Jezek et al.*, 1994]. Accompanying shallow firn cores and snowpits characterized the snow and firn stratigraphy, including density. The radar measurements were made in the Ku band (centre frequency around 13 GHz) for relevance to satellite radar altimeters. Field campaigns took place in the spring and autumn of 2004 and the spring of 2006. They enabled a characterization of the effects on radar backscatter from the major spatial and seasonal changes in the snowpack caused by summer melting and refreezing.

2. Methods

[5] In spring 2006 snowpits were dug to the previous end of summer melt surface where the density of the firn increases significantly at sites from T03 to T07 (Table 1). It was then possible to retrieve a core from the base of these pits to a depth of around 3.5 m from the surface. Densities of each stratigraphic layer were determined by weighing known volumes, both in the snowpits and from the core. In autumn 2004 it was possible to retrieve core from the surface and both the core and pits were logged from the surface.

[6] A step-frequency radar based on a network analyser was used for ground-based measurements; the system is described in detail by *Scott et al.* [2006]. The radar was mounted on a sledge with antennas on the end of a horizontal beam, pointing vertically down to give a zero incidence angle at the surface. The antennas were positioned

[1]British Antarctic Survey, Cambridge, UK.
[2]School of Geosciences, University of Edinburgh, Edinburgh, UK.
[3]Department of Geography and Environment, School of Geosciences, University of Aberdeen, Aberdeen, UK.
[4]Scott Polar Research Institute, University of Cambridge, Cambridge, UK.
[5]Centre for Polar Observation and Modelling, University College London, London, UK.

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Table 1. Locations of Radar Measurements

Site	Latitude	Longitude	Elevation, m	Spring 2006[a] Upper Layer	Spring 2006[a] Lower Layer	Autumn 2004[a]
T03	N69°46′27.9″	W47°43′52.2″	1795	2.30	2.02	NA
T05	N69°51′04.6″	W47°15′11.2″	1942	2.25	2.09	2.11
T07	N69°56′28.7″	W46°48′03.2″	2050	2.26	2.17	NA
T12	N70°10′32.9″	W45°20′40.4″	2350	2.32	2.32	NA

[a]Electromagnetic velocity, $\times 10^8$ m s^{-1}. NA, not available.

1.25 m above the snow surface so that the surface return was in the antenna far field. The frequency step used was from 9 GHz to 17 GHz with a step size of 909 kHz.

[7] Measurements presented here were made with the radar stationary for improved quality and each measurement is an average of five static shots; the radar was moved from 5 to 20 cm between each shot by pulling the sledge. This averaging will reduce any speckle noise. It will also reduce the effect of any strong reflectors that are not spatially extensive, such as ice blobs and any small areas giving a high return because of constructive interference.

[8] The velocity needed to convert travel time to depth was obtained by placing a metal reflector into the snowpack at known depths at T05 in autumn 2004 [*Scott et al.*, 2006]. The same method was used at T12 in spring 2006 where the electromagnetic velocity was found to be constant to a depth of 2 m. Therefore this velocity was used at T12, assuming the additional effect on velocity from firn densification not to be appreciably greater for the next few meters. Between T03 and T07, the much greater increase in density of the snowpack with depth in spring, caused by going from an upper layer of winter accumulated snow to a lower layer of icy firn below the previous summer melt surface, required a two stage electromagnetic velocity model with a high velocity near the surface and a lower velocity below the previous summer melt surface. Between T03 and T07, springtime electromagnetic velocities were calculated using the average densities in the respective layers. These were obtained from core and snowpit measurements. The empirical relationship of *Glen and Paren* [1975] was then applied to give electromagnetic velocity and this has been found to work well in previous studies at T05 [*Scott et al.*, 2006]. The velocities used at all locations are given in Table 1.

[9] Calculation of backscatter power was performed using the method presented by *Scott et al.* [2006] that accounts for the effect of spherical spreading. Each result was then normalized by setting the highest power to zero decibels. This allows the relative contribution of reflectors at different depths to be determined more easily between measurements sites.

3. Results

3.1. Spring to Autumn

[10] One of the greatest and most rapid changes to the snowpack in the percolation zone is caused by summer surface melting, percolation and refreezing at depth. The change this causes in the radar backscatter is demonstrated by the comparison of measurements taken in autumn 2004 with those in spring 2006 at T05 (Figure 1). These measurements were made prior to the onset and after the cessation of summer melt, with most of the snowpack well below −5°C. A strong reflection occurs at the surface in both spring and autumn (Figure 1). In the autumn this surface reflection is often observed to be a double reflection with similar magnitude returns from both the actual surface and thin, shallow ice layers at depths up to 20 cm [*Scott et al.*, 2006]. The backscatter return power then decreases rapidly with depth. However, in the spring there is a return at depth, which is often stronger than the surface return (Figure 1, marked A). This return was observed in all spring measurements from T03 to T12. The precise knowledge of electromagnetic velocity in the last year's snow, determined as outlined above (and by *Scott et al.* [2006]), enabled the location of the leading edge of the strong reflection at the previous summer surface. This summer melt surface was easily and unambiguously identifiable from logging of the stratigraphy within snowpits and from ablation stakes left in situ over the past two years. Other prominent reflectors (Figure 1, marked B) also appear between the surface and the previous summer melt surface. These reflectors occur at a similar depth to icy windcrust layers (Figure 2).

3.2. Spatial Variation

[11] At T03, the lowest measurement point, nearest to the ice sheet margin, percolation features, such as ice layers and lenses, were observed throughout the snowpack below the

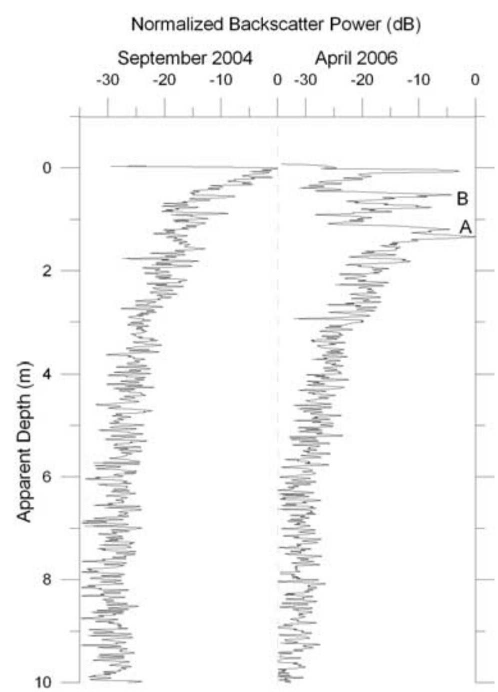

Figure 1. Normalized radar backscatter power return with depth at T05 in September 2004 and April 2006. Centre frequency 13 GHz, bandwidth 8 GHz. The 2005 end of summer surface is marked A. Strong reflectors, believed to be icy windcrust layers, are marked B.

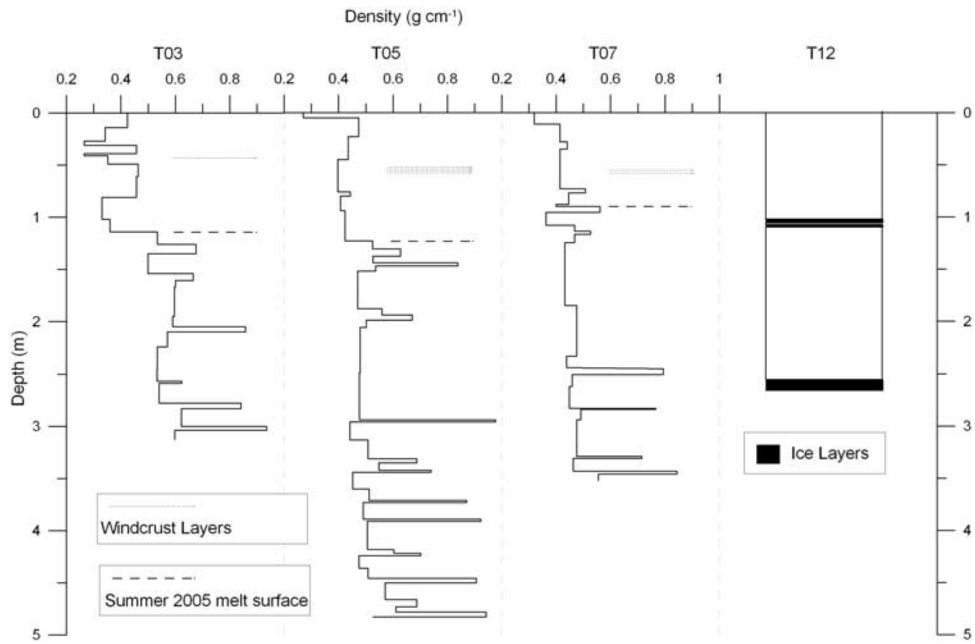

Figure 2. Density – depth profiles from shallow ice cores and pits at T03, T05 and T07, in May 2006. Icy windcrust layers and the summer 2005 melt surface are marked. Ice layers logged in a pit 10 m from T12 in April 2006 are shown.

summer melt horizon (Figure 2). At T05 observations from shallow cores and snowpits in spring and autumn 2004 and spring 2006 demonstrate that percolation features do not always reach the previous summer melt surface. There is also little additional densification, due to percolation, observed beneath the previous end of summer surface. This suggests that most of the percolating water refreezes within the last year's snow. Passing through T07 and T12 the depth that percolation features reach decreases considerably. Some thin ice layers are visible at T12 with surface melt layers refreezing at, or very near to, the end of summer melt surface in 2005, with little percolation observed.

[12] Radar backscatter results across the percolation zone in spring 2006 show certain similarities with a strong surface reflection and a distinctive reflection from the previous summer 2005 surface (Figure 3). Reflections from within the winter snowpack occur between these features and are probably due to buried icy windcrusts that can be several millimeters thick. At T07 (Figure 3) a sharp rise can be seen in the return from a depth of around 2.4 m (A

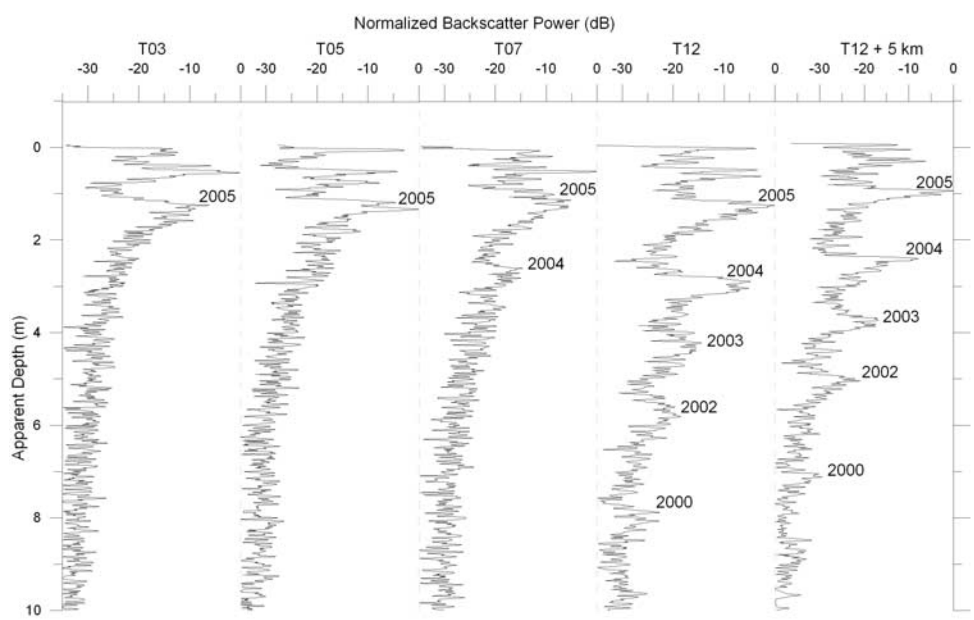

Figure 3. Normalized radar backscatter power return with depth across the percolation zone from T03 to T12, from measurements in April and May 2006. Centre frequency 13 GHz, bandwidth 8 GHz. Interpreted positions of summer melt surfaces are marked with dates. These surfaces correlate well with the surfaces tracked along EGIG from a borehole near summit using neutron probe density profiles.

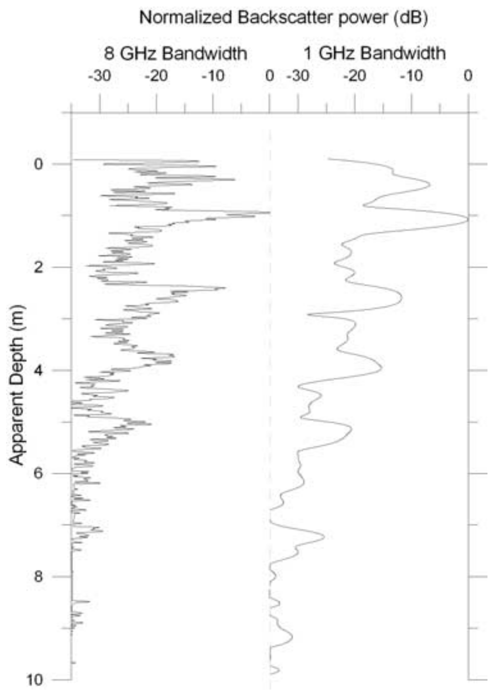

Figure 4. Normalized radar backscatter power return with depth at T12 + 5 km along EGIG (towards summit) in April 2006. Comparison of a measurement at a centre frequency of 13 GHz, bandwidth 8 GHz with a measurement at a centre frequency 13.5 GHz, bandwidth 1 GHz.

similar feature occurs at T06, approximately half way between T05 and T07). Measurements with ablation stakes left in 2004 show that this return is around the depth of the summer 2004 surface. At T12 this return is a major peak with a backscatter power similar to the surface return. Snowpit stratigraphy (Figure 2) along with ablation stakes identified the leading edge of this reflection as originating from an ice layer at or immediately below the 2004 summer surface. There are also additional peaks appearing at greater depths at T12 (Figure 3). At T12 + 5 km up EGIG these deeper reflections become sharper.

[13] Pits and cores sampled at T05 in spring 2006 show that percolation features from the summer melt of 2005, such as ice layers, do not always reach the end of summer 2004 surface. For example it can be seen that there is a section from around 2 to 3 m depth that displays no major densification due to percolation in a core taken at T05 (Figure 2). At T07 the core showed no evidence of summer 2005 percolation reaching the previous summer 2004 surface, with a stretch of low-density firn from around 1.2 to 2.5 m deep. A pit at T12 demonstrated that the 2005 summer melt did not percolate far into the snowpack with two very close ice layers, 3 cm and 2 cm thick, evident below the 2005 summer melt surface at a depth of just over 1 m (Figure 2). At T12 a 10 cm thick ice layer was present at the 2004 summer melt surface at around 2.5 m deep.

3.3. Resolution

[14] Airborne radar altimeter over-flights were made along the EGIG line in the spring and autumn of 2004 using the ASIRAS radar [*Hawley et al.*, 2006; V. Helm et al., Winter accumulation in the percolation zone of Greenland measured by airborne radar altimeter, submitted to *Geophysical Research Letters*, 2006, hereinafter referred to as Helm et al., submitted manuscript, 2006]. This is intended to be a proxy for satellite measurements. With a much smaller footprint than a satellite altimeter the results can be more closely compared to ground based results. ASIRAS operates at a bandwidth of 1 GHz with a centre frequency of 13.5 GHz. The satellite altimeters ERS-1 and 2 and the planned CryoSat-2 operate at even lower bandwidths (e.g. 320 MHz [*Wingham et al.*, 2006]). This means that the resolution of these systems is much less than the ground radar. The ground radar collects data in the frequency domain so it is easy to obtain measurements at any bandwidth within its frequency step. A comparison of 1 GHz to 8 GHz bandwidth measurements made by the ground-based radar 5 km up EGIG from T12 is shown in Figure 4. It can be seen that the annual layering observed at a bandwidth of 8 GHz is still distinctive at 1 GHz. It is also worth noting that the surface return is not resolvable from the shallow layers within the snowpack at 1 GHz although at many locations the shallow layers can be resolved at this bandwidth.

4. Discussion

[15] In the middle of the percolation zone, at around 2000 m elevation along this transect, the autumn backscatter is dominated by the surface return with a strong contribution from shallow ice features [*Scott et al.*, 2006], where after the backscatter power reduces rapidly with depth. Winter snow accumulation on top of this end of summer surface creates a double return with the actual surface providing a lower power return than the buried end of summer surface (Figures 1 and 3). One reason for the stronger return at depth may be the decimeter scale surface roughness. Ice layers at depth, particularly around the summer melt surface appear much flatter than the surface. More analysis is being done on the variation of the backscatter power with roughness. However, the surface backscatter power was seen to reduce by several decibels over rougher surfaces. The double return is further complicated by strong returns from icy windcrusts within the snowpack. It is likely that the lower resolution of the airborne ASIRAS radar means that the surface return is combined with these layers making the tracked surface in spring 13 cm lower than the actual surface (Helm et al., submitted manuscript, 2006). A similar situation could prevail in autumn due to shallow ice layers. Further work is being done to investigate the spatial extent of these windcrust layers and their backscatter power, to estimate their contribution at the satellite footprint scale. Thicker, denser windcrust layers are often due to surfaces exposed to wind and radiation for long periods of time and may be spatially extensive. However, the ASIRAS radar is able to track the spatially extensive summer melt surface and from this an estimate of the winter accumulation can be made (Helm et al., submitted manuscript, 2006).

[16] Moving up the percolation zone from higher to lower melt regions, there is a point where summer densification features from several years become resolvable (Figure 3). The resolution of such annual layers is dependent on the strength of individual melt years and the subsequent depth of percolation prior to refreezing. The summer 2004 melt surface becomes clearly resolvable at T06 and above but can

only be identified in some measurements at T05. Multiple summer melt surfaces become clearly resolvable at and above T12. This is also the point where ASIRAS begins to resolve annual layering within the snowpack (B. Hawley, personal communication, 2006). It appears these annual layers start to be observed around T12 and can be tracked past T21 [*Hawley et al.*, 2006], 70 km up EGIG. Even at T21 (2737 m elevation) thin ice layers were observed in pits dug in 2004. It is likely that the ice layers continue up the ice sheet becoming progressively thinner. The absence of a backscatter power peak between 5 and 7 m at T12 + 5 km, that would correspond to a summer 2001 melt surface (Figure 2), may indicate that during summers when there is little or no surface melting these features are not detectable. Another reason for the lack of a distinct peak may be due to a strong melt event in 2002 [*Steffen et al.*, 2004]. This could have caused surface meltwater to percolate most, or all, of the way down to the previous summer surface. Subsequent refreezing would then create strong radar reflectors that would obscure the 2001 summer surface reflection. This process would occur more frequently at lower elevations.

[17] The results presented here demonstrate that when modelling the radar altimeter return from the percolation zone of an ice sheet it is important to be aware of the significant spatial, seasonal and interannual variations that can occur. A surface slope (5.3 m per kilometer along the 106 km transect), along with a lower range resolution, causes a satellite radar altimeter to compound returns from the surface and internal layers together. If a greater proportion of the return signal was to originate at or near the surface it could be retracked as a false elevation increase. Calculation of elevation changes usually take into account seasonal variations. However, *Steffen et al.* [2004] show that recent melt events are extending up the ice sheet to include higher elevations. Strong reflecting layers nearer to the surface may appear in the satellite data as an elevation increase. This may even account for some of the calculated increase in ice thickness observed at high elevations [*Johannessen et al.*, 2005; *Zwally et al.*, 2005].

5. Conclusions

[18] Ground based radar measurements have helped to verify the ability of 1 GHz bandwidth Ku band radar altimeters to track annual layering into the high percolation zone as well as in the dry snow zone as suggested by *Hawley et al.* [2006]. Further, the high bandwidth of the ground measurements have enabled identification of the upper two reflectors responsible for this annual layering as thin ice layers coinciding at or just below the end of summer melt surfaces of 2005 and 2004. We believe that the ground measurements suggest the limitation on tracking the layers, as measurements continue lower down the percolation zone, is the depth percolation features reach within the previous year's snowpack. If the percolation passes through most of the last year's snow it will become impossible to distinguish the previous summer melt surface from the percolation features above it.

[19] Results given here have also confirmed the findings of *Jezek et al.* [1994], showing a strong return from the end of summer melt surface. This return can be used to track winter accumulation in the percolation zone as demonstrated by Helm et al. (submitted manuscript, 2006). However, the high bandwidth results presented here and by *Scott et al.* [2006] demonstrate that errors can arise because of the reflections being a combination of the surface and buried windcrust or ice layers in spring and autumn respectively.

[20] Models of volume backscatter can be created and applied to satellite retracking algorithms using the layering demonstrated in these results. Estimates of the potential elevation errors caused by strong seasonal and interannual variations in the radar altimeter reflection, across a percolation zone of varying extent, can be made. If the percolation zone continues to move up the ice sheet [*Steffen et al.*, 2004; *Ngheim et al.*, 2005] it could affect the elevation changes being measured by satellite altimeters, by placing strong reflecting ice layers near to the surface. This effect could cause an overestimation in surface elevation increase and must be modelled to improve the confidence in current mass balance estimates.

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D. Mair, Department of Geography and Environment, School of Geosciences, University of Aberdeen, Elphinstone Road, Aberdeen AB24 3UF, UK.

E. Morris, Scott Polar Research Institute, University of Cambridge, Lensfield Road, Cambridge CB2 1ER, UK.

P. Nienow and V. Parry, School of Geosciences, University of Edinburgh, Drummond Street, Edinburgh EH8 9XP, UK. (peter.nienow@ed.ac.uk)

J. B. T. Scott, British Antarctic Survey, High Cross, Madingley Road, Cambridge CB3 0ET, UK.

D. J. Wingham, Centre for Polar Observation and Modelling, University College London, Pearson Building, Gower Street, London WC1E 6BT, UK.