RADAR GLACIER ZONES AND THEIR BOUNDARIES AS INDICATORS OF GLACIER MASS BALANCE AND CLIMATIC VARIABILITY

Frank Rau, Matthias Braun, Matthias Friedrich, Frank Weber & Hermann Goßmann
University of Freiburg, Department of Physical Geography
Werderring 4, D-79085 Freiburg, Germany
Phone: +49-(0)761-2033550, Fax: +49-(0)761-2033596, e-mail: frank.rau@ipg.uni-freiburg.de

ABSTRACT

The measurement of glaciers mass balance and the derivation of meteorological parameters by means of remote sensing techniques are of particular scientific interest for glaciological and climatological studies. Using synergistic approaches, SAR-data complement field studies in remote and inaccessible areas and provide information with a high temporal and spatial resolution. However, the operational application of the existing SAR-archives for monitoring purposes require a more detailed understanding of the spatial and temporal evolution of the observable radar glacier zones. Backscatter modelling based on snow pit data gathered on the Antarctic Peninsula is used to investigate typical radar returns from the upper and lower boundary lines of the frozen percolation zone. The lower one is a clearly identifiable limit between bare ice and highly metamorphosed, coarse-grained snow. As it is impossible to identify the year of origin of this snowpack, the term firn line is suggested. Only occasionally the late-summer firn line position coincides with the equilibrium line. In the absence of field measurements, firn line altitude can be regarded as a first approximation to the ELA, although it might lead to an overestimation of glaciers mass balance. The upper boundary, the dry snow line, is described as a temporarily and spatially very persistent feature in SAR-imagery. Its position is very stable in the time range of years to decades. Positional shifts of this line provide information about the occurrence of singular high temperature events leading to enhanced metamorphism of the snow and accumulation conditions in the highest areas of the glaciers. Additionally, model results confirm the assumption that a threshold of –8 dB is appropriate to delimit the upper boundary line of the frozen percolation radar zone. This facilitates direct monitoring of the dry snow line.

INTRODUCTION

The large-scale and year-round monitoring of snow cover dynamics in polar regions has become feasible in recent years through the all-weather operating capabilities of orbital synthetic aperture radar (SAR) satellites. Starting with the launch of the first spaceborne SAR-instrument on board the SEASAT-satellite in 1978, radar remote sensing has proven to be a major tool for snow cover and glacier investigations. The recent generation of active microwave remote-sensing satellites provides coverage with a high temporal and spatial resolution unhampered by the frequent cloud cover or the limited day light during the winter season. In addition, unlike visible imagers, the radar signal penetrates the snow surface, which provides scientists with a tool capable of sensing internal properties of the snowpack. Consequently, remote sensing techniques using SAR-data complement field measurements and facilitate further interpretation of the observed backscatter signatures (e.g. SMITH et al., 1997; PARTINGTON, 1998; BINDSCHADLER, 1998; BRAUN et al., 2000; RAU et al., 2000).

Depending on the prevailing snowpack conditions during image acquisition, different snow zones on a glacier are identifiable in the SAR-images. These zones are attributed as radar glacier zones and can be classified by their backscatter characteristics and their elevational positions with respect to each other. A major subdivision comprises (1) a dry snow radar zone (DSRZ), (2) a frozen percolation radar zone (FPRZ), (3) a wet snow radar zone (WSRZ) and (4) a bare ice radar zone (BIRZ). Of particular scientific interest for glaciological and climatological studies is the measurement of the mass balance of the glacial systems and the derivation of actual and recent meteorological parameters. Hereby, the boundaries between the radar glacier zones provide information on both: (1) the transient firn line separating BIRZ and WSRZ (summer) or FPRZ (winter) documents the current spatial extension of the ablation zone. In particular, the position of the transient firn line at the end of the ablation season is often regarded as the equilibrium line altitude (ELA), which is a key parameter for the glaciers mass balance, and (2) the dynamic wet snow line between the actual WSRZ and FPRZ approximately coincides with the position of the actual 0°C isotherm and (3) the dry snow line skirting the DSRZ is an indicator of singular extreme melt events impacting also the uppermost altitudes (RAU et al., 2000).

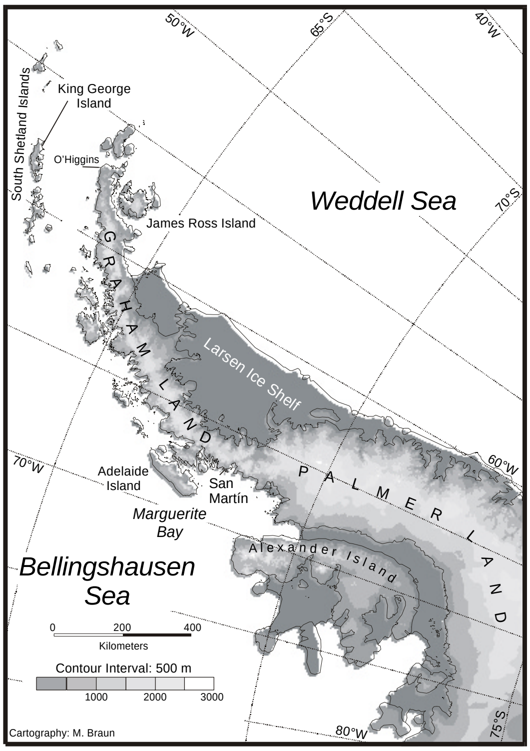

Fig. 1: The Antarctic Peninsula with place names mentioned in the text.

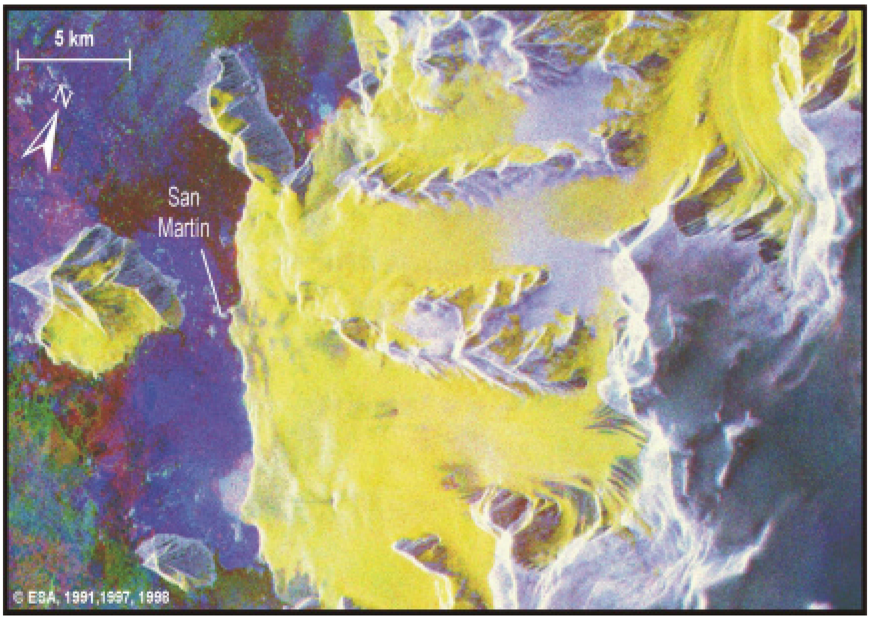

Fig. 2: Multitemporal SAR-composite from the inner Marguerite Bay, Antarctic Peninsula (12 Oct. 1991 (ERS-1), 13 Jul. 1997 and 08 Feb. 1998 (ERS-2)). The WSRZ during the summer 1997 appears in yellow, the FPRZ in light grey and the DSRZ in dark grey colours.

The operational utilisation of the existing SAR-archives for monitoring purposes and the combination of glacier parameters derived from remote sensing techniques with field measurements however require a more detailed understanding of the spatial and temporal evolution of the radar glacier zones. Furthermore, ambiguities in terminology between remote sensing and field glaciologists have to be avoided if synergistic approaches are to meet the expectations of both. As a contribution to the outlined topics, we are investigating in this study the temporal and spatial persistence of the radar glacier zones and their boundaries with respect to snow accumulation and snow metamorphism by means of backscatter modelling. In particular, we are focusing on the SAR-derived transient firn line as an indicator of glaciers mass balance and on the dry snow line as an indicator of climatological variability. The study is based upon current investigations on the Antarctic Peninsula (Fig.1), although the results might be applicable to other glaciers and ice sheets world-wide.

BACKSCATTER MODEL s°

In order to link the SAR-data with the snow pit measurements a backscatter model was developed, which facilitates the simulation of the radar return signal from a multi-layer snowpack depending upon numerous snow and sensor parameters (stratigraphy, density, grain size, liquid water content, surface roughness, frequency, polarization, incidence angle; FRIEDRICH, 1996). The contribution of subsurface snow or ice layers to the backscattered signal takes the extinction of all overlying layers into account. Hereby, volume scattering is calculated using the Rayleigh formula (ULABY et al., 1986), while surface scattering is modelled after Kirchhoff (stationary phase approximation: ULABY et al., 1986; scalar approximation: DRINKWATER, 1989).

Model evaluation was carried out by comparing modelled backscatter coefficients with measured values from SAR-data acquired on 12 October 1991 (ERS-1), 4 November 1996, 13 July 1997, 8 February 1998, and 19 November 1999 (ERS-2) over glaciers near the Argentine research station San Martín (Marguerite Bay, 68° S, 67° W; Fig. 1). Using the ESA SAR-Toolbox software, normalized backscatter coefficients (s°) were calculated for all images (after LAUR et al., 1997). In accordance to the suggestions of MEADOWS et al. (1999), the correction of the ERS on-board Analogue to Digital Converter saturation was applied. The images were co-registered and resampled to a common pixel resolution of 12.5 m. A median filter (kernel size 5*5 pixel) was applied to reduce image speckle. Finally, the images were merged and subset as a layer stack into a multi-band image (Fig. 2). Modelling of the FPRZ was based upon snow pit data (1994/95, 1996 and 1998) from the area, while snow pit and ice core data (THOMPSON et al., 1994) served as input for the simulated backscatter coefficients of DSRZ. Due to the SAR-inherent image speckle, it is impossible to identify a single snow pit in the images. Therefore, homogeneous areas covering the corresponding snow pit locations were selected as areas of interest to calculate averaged backscatter coefficients from each input image (frame size 100 * 100 pixel). The obtained value ranges indicate an excellent agreement between modelled and measured values in the DSRZ. The modelled radar returns from the FPRZ are slightly lower than the measured values, but generally show good accordance. The value ranges of the snow parameters used as input for the backscatter model are taken from snow pit measurements obtained during several field campaigns to the inner Marguerite Bay, the South Shetland Islands and James Ross Island (Antarctic Peninsula). Calculations were realized assuming constant incidence angles of 23° and a rough surface (rms slope = 0.5). However, as modelling was restricted to dry snow conditions, the influence of both parameters on the simulated backscatter values were negligible.

RADAR GLACIER ZONES

SAR-images of glaciers and ice sheets show a typical sequence of alternating dark and bright signatures, which originate from specific backscatter mechanisms in accordance to the prevailing snow cover parameters such as snow density, liquid water content, grain size, stratigraphy, and surface roughness. However, as the spatial and temporal evolution as well as the delimitation of these snow zones identifiable in SAR-images do not necessarily coincide with the characteristics of the classical glaciological snow zones (BENSON, 1962; PATERSON, 1994), they should be referred to as radar glacier zones (FORSTER et al., 1996; SMITH et al., 1997; RAU et al., 2000; BRAUN et al., 2000). Their formation is determined by metamorphic processes within the snowpack driven by the meteorological conditions prior to and during image acquisition. Consequently, they are dynamic on a time scale of days to weeks and show remarkable inter-annual variations.

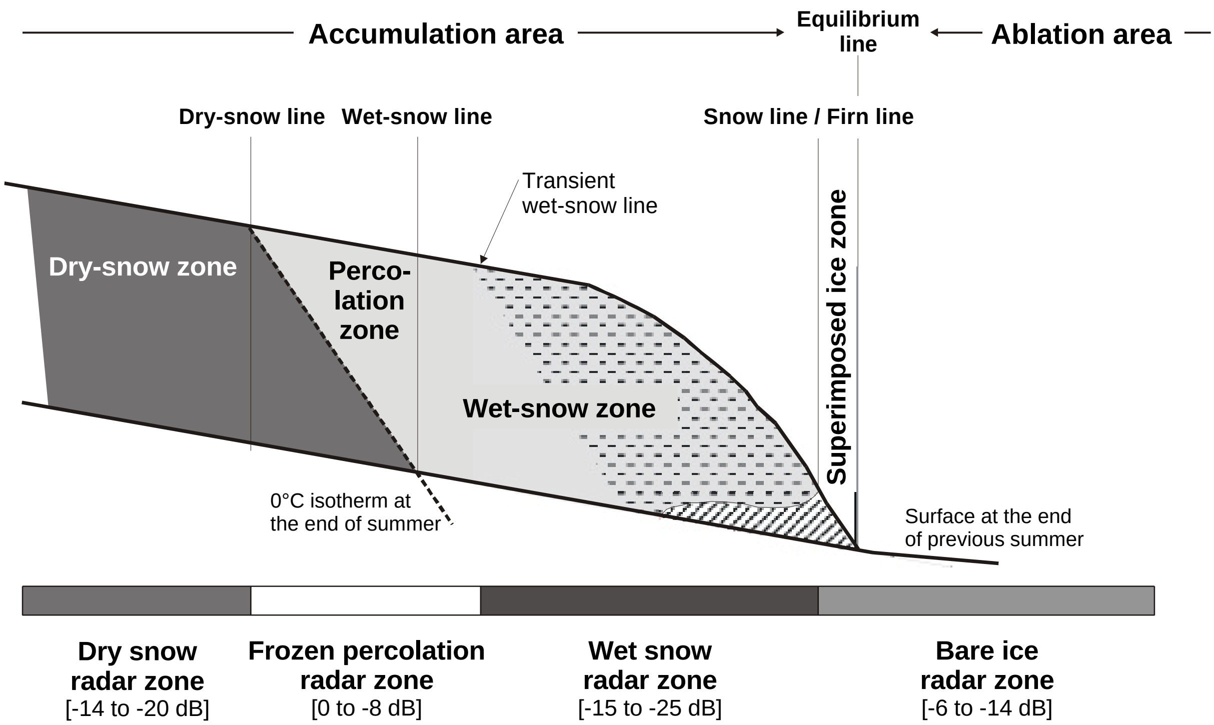

Fig. 3: Glacier snow zones and corresponding radar glacier zones with typical backscatter values (C-band, VV; modified after Paterson, 1994)

In this study, we refer to the classification scheme proposed by RAU et al. (2000) comprising dry snow, frozen percolation, wet snow and bare ice radar glacier zones (Fig. 3). Hereby, the DSRZ is restricted only to the highest areas, in which the temperatures never rise above freezing point. The absence of melt events and the predominating dry snow metamorphism result in small grain sizes and the absence of ice layers. Due to the high penetration depth and dominating volume scattering, the DSRZ is characterised by low backscatter values. Frequent or occasional melt-freeze-cycles lead to the formation of numerous subsurface ice bodies and large grain sizes in the snowpack of the FPRZ. While in a dry and frozen state, both ice layers and large snow grains act as strong scatterers of the radar beam. This results in high backscatter values from the FPRZ. During the ablation season, melting increases the liquid water content in the snowpack. As the radar signal is almost completely absorbed by liquid water, the penetration depth is reduced to the uppermost centimetres. Depending on the surface roughness, scattering on smooth surfaces leads to a dark appearance of a melting snow cover in a SAR-image. In the BIRZ, surface scattering causes a relatively strong backscatter signal in comparison to the WSRZ.

THE FIRN LINE IN SAR–IMAGERY – IS THE ELA DETECTABLE?

The boundary line separating the BIRZ and the adjacent wet snow or FPRZ is often regarded as the actual transient snow line of a glacier. At the end of the ablation period or during the winter, this well marked line (Fig. 4) is often interpreted as an approximation of the ELA (e.g. BINDSCHADLER, 1998), which is of special interest for the monitoring of climatic variations and glacier mass balance studies. The derivation of this valuable parameter by means of remote sensing techniques is of particular importance in remote areas such as the polar regions where regular field measurements are not feasible. Recent studies (e.g. BRAUN et al., 2000; KÖNIG et al., this volume) however outlined the need for a more precise definition of the feature observable in SAR-imagery.

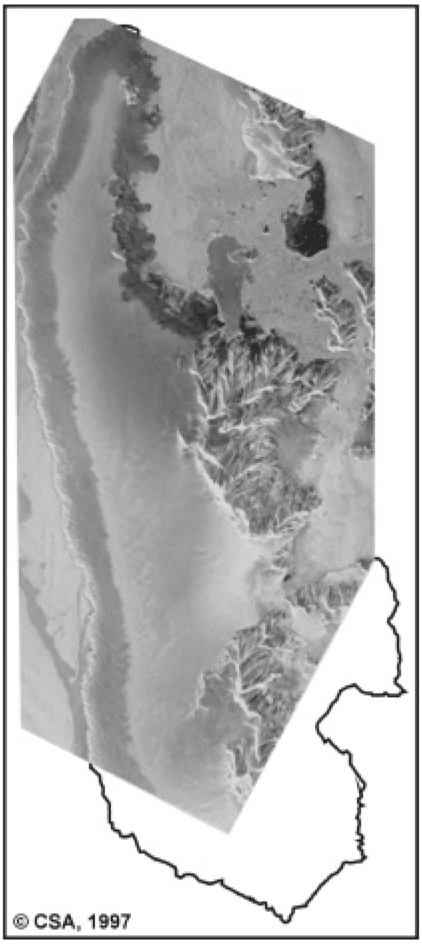

Fig. 4: Adelaide Island (Radar-sat, 10 May 1997). The dark BIRZ skirts the FPRZ of the islands interior. The boundary is represented by the late-summer firn line.

During high temperature periods, the snowpack of the lowermost accumulation zone is affected by intense wet snow metamorphism. Hence, the snow conditions at the end of the summer are characterised by large grain and cluster sizes, abundant ice layers and pipes, high densities and low profile depths. As model results indicate, high backscatter returns (–0.5 to –7.0 dB) from such a snowpack in a frozen state are predominantly caused by volume scattering from the snow crystals and clusters. The contributions from ice bodies in the snow are negligible. This is in good agreement with field data, where similar high backscatter values from the frozen percolation zone resulted from snowpacks with abundant ice layers as well as from snowpacks without ice layers or pipes. Typical snow properties with variable densities and grain sizes were used to model the backscatter returns in dependence of the depths of the metamorphosed snow cover overlying the glacier surface. As demonstrated in Fig. 5, a shallow coarse-grained snow cover of only 0.2 m depths rises backscatter values above a threshold of –8 dB, which can be regarded as a lower limit for the FPRZ. It can therefore be stated, that the radar-bright frozen percolation zone is delimited from the radar-dark bare ice zone by a sharp and well pronounced boundary, where backscatter values increase abruptly. However, it is impossible to deduce on the particular time of accumulation of the coarse-grained snowpack forming the lower limit of the snow covered area. For example, in years with a negative mass balance, firn from previous years might be exposed at the upper boundary of the BIRZ. In order to avoid ambiguities with the classical glaciological terminology, it is therefore suggested to apply the term firn line to the lower boundary line of the FPRZ (see also KÖNIG et al., this volume).

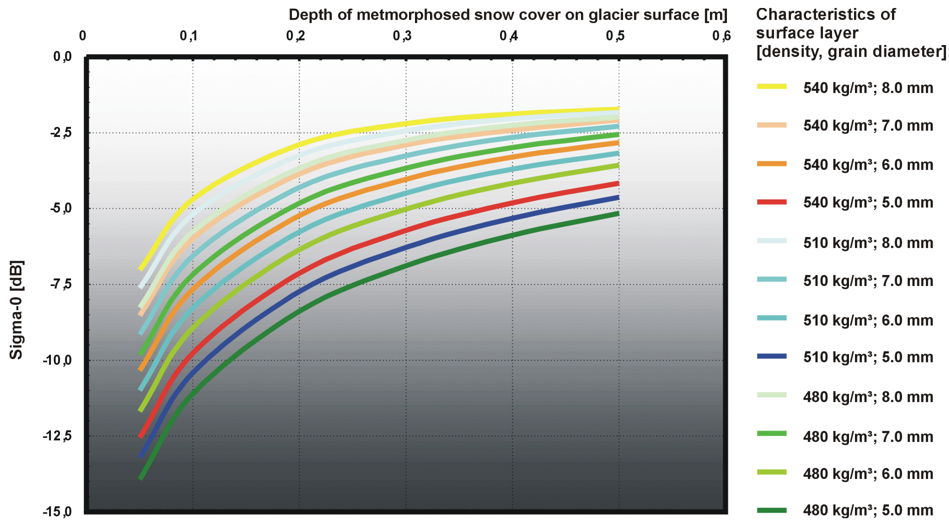

Fig. 5: Backscatter values from a coarse-grained snowpack covering the bare glacier surface in dependence from density and grain size. Model results indicate clearly, that a already shallow coarse-grained snowpack, as it is representative for the lower parts of the FPRZ at the end of the ablation season, leads to high radar returns above –8 dB.

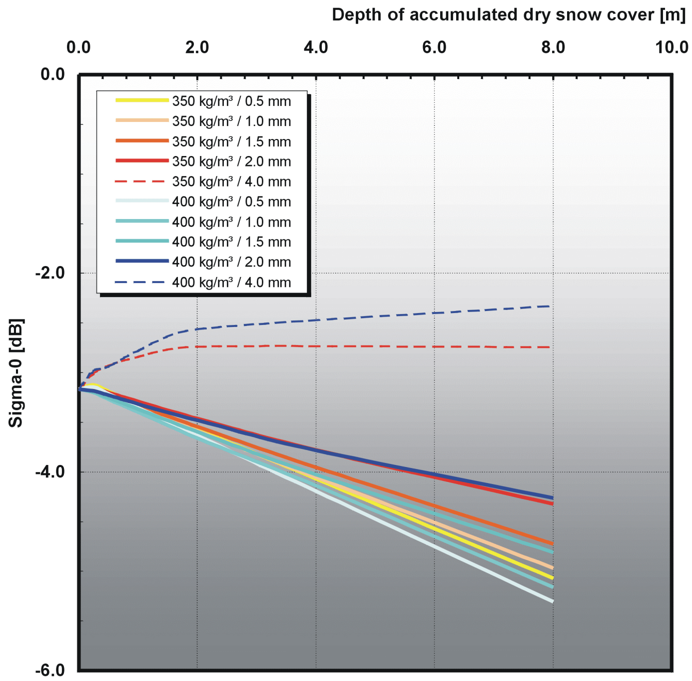

Fig. 6: Modelled backscatter values from a highly reflective snowpack of the FPRZ covered by a dry snowpack with variable densities and grain diameters. Even a dry fine-grained top layer of several meters thickness appears transparent in the SAR-images as the backscatter values are only slightly altered. Increasing grain sizes in the upper layer due to metamorphism however significantly increase the radar return.

To investigate the temporal persistence of this firn line visible in SAR-imagery under an accumulating wintry snow cover, the backscatter return from a highly reflective snowpack from the FPRZ was overlaid by a homogeneous fine-grained snow characterised by typical densities ranging between 350 and 400 kg/m³. Such a snow layer is almost transparent for the radar beam and the backscatter characteristics are dominated by the underlying coarse-grained snow or firn (Fig. 6). An accumulation of up to 8 m of dry and fine-grained snow only insignificantly decreases the backscatter values by –2 dB. However, an increase of the grain size due to metamorphism driven by occasional or frequent high temperature events during the winter will result in an abrupt increase of the s°-values. As the fresh snow cover is homogeneously distributed over both, the former BIRZ and FPRZ, such metamorphic processes would consequently hide the firn line supposing an extensive FPRZ. Particularly maritime regions such as the western side of the Antarctic Peninsula are often affected by this kind of mid-winter thawing events and it might therefore be difficult to detect the firn line position on a glacier. An adequate timing of the image acquisition in the early accumulation season, when the firn is completely refrozen and a probable fresh snow cover has not yet undergone any metamorphic processes, is recommended. Additionally, supporting meteorological data allow to draw conclusions about probable snow metamorphism prior to image acquisition and therefore permit further interpretation of the SAR-images.

These findings are of major importance for the detection of ELAs using SAR-imagery. Only in years with a positive mass balance of the glacier, the firn line at the end of the summer coincides with the position of the ELA. On the contrary, in years which are characterised by a negative mass balance, firn from previous accumulation seasons might be exposed, leading to the assumption of a too low ELA position. Consequently, deriving ELA from SAR-imagery might lead to an overestimation of glaciers mass balance. It should be mentioned here that the probable existence of a superimposed ice zone induces further uncertainties in the determination of the ELA from SAR imagery. Nevertheless, the firn line position provides climatological information and can be regarded as an approximation of the ELA. Especially in remote regions where no direct field measurements are available, mapping the firn line offers a possibility to support mass balance monitoring. It therefore seems to be necessary to initiate further studies to investigate the relation between firn line altitude (FLA), ELA and the glaciers mass balance.

THE TEMPORAL AND SPATIAL PERSISTENCE OF THE DRY SNOW LINE

The transition between the frozen percolation and the DSRZ has been regarded as an indicator of climatological variabilities providing information about the occurrence of singular high temperature events affecting the uppermost areas of the glacial systems (WUNDERLE, 1996; RAU et al., 2000). However, for a long-term monitoring of these a more detailed understanding of the temporal and spatial evolution of this transition zone is required. Firstly, it is of great interest to learn which processes lead to an increase of the backscatter values from approximately –14 to –19 dB in the DSRZ to values higher than –8 dB as typically found in the FPRZ. Secondly, interpretation of SAR-imagery requires knowledge of the temporal persistence of such a snowpack identifiable as a FPRZ under accumulation conditions, which are typical for the DSRZ.

Data from two snow pits were taken as a basis for modelling the alteration of the backscatter values as a consequence of metamorphism of the uppermost snow layers due to the impact of a high temperature event. The first snow profile can be regarded as representative for the DSRZ. It is characterised by small grain sizes and low densities resulting in a s°-value of –17.6 dB. The second one shows a higher backscatter value of –13.8 dB due to higher grain sizes and higher densities, which are characteristic of the transition zone stretching between the frozen percolation and the DSRZ. The metamorphism of the uppermost snow layer was simulated by gradual increase of grain size, density, and depth of the altered surface layer (Fig. 7). Model results indicate clearly that the backscatter coefficient is mainly influenced by increasing grain sizes of the snow crystals and clusters, while an increase of density contributes only to a minor degree to the radar return of the snowpack. Both snow profiles show a similar increase of s°-values exceeding a threshold of –8 dB only with grain sizes larger than 3 mm, densities around 500 kg/m³, and depths of the metamorphosed surface layer over 0.75 m. These value ranges might only be expected as a consequence of an extraordinary strong meteorological event impacting the snow cover of the transition zone between frozen percolation and DSRZ or, rarely, the DSRZ itself. An upward position shift of the transition line between both radar glacier zones has therefore to be attributed to the rare occurrence of meteorological conditions, which favour the rapid and profound metamorphism of the surface snow layer. Advection of warm and humid air masses, as frequently recorded on the Antarctic Peninsula even during the winter month, or singular rainfall events, as reported for the James Ross Island Ice Cap (pers. comm. Simões, 2000), would consequently give reasonable explanations for the significant increase of backscatter values from these areas.

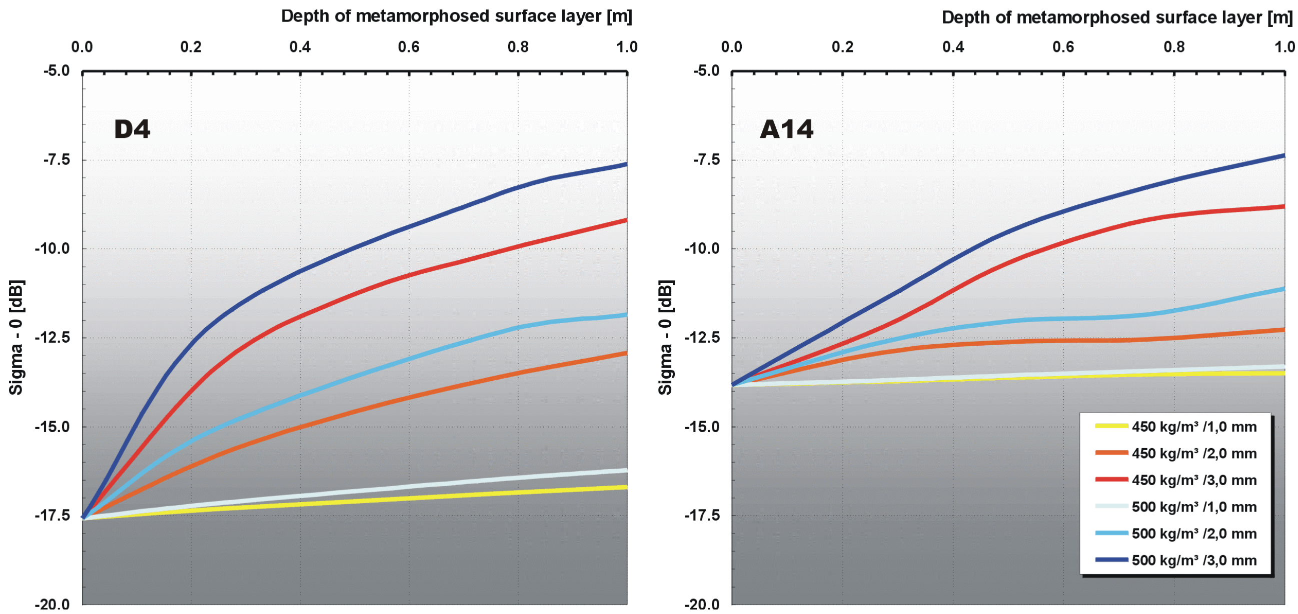

Fig. 7: A metamorphosed snow layer covering a snowpack of the DSRZ (left) or of the transition between FPRZ and DSRZ (right) increases the backscatter values only above a threshold of –8 dB, if grain diameters increase significantly. As such, an upward shift of the dry snow zone has to be a consequence of an extraordinary strong high temperature event.

A downward shift of the position of the dry snow line is a consequence of the accumulation of fine-grained dry snow upon a coarse-grained snow cover. To investigate the temporal evolution of backscatter values resulting from such an accumulating dry snow surface layer on a snowpack from the FPRZ, two snow profiles with a coarse-grained surface layer (grain size diameter 3 mm) resulting in s°-values of –4.4 and –7.4 dB, respectively, were used to initialize the model. Based on annual accumulation rates ranging between 200 and 800 mm/a as reported for the central Antarctic Peninsula (PEEL, 1992; TURNER et al., 1998) and a mean density of the uppermost snow layer of 400 kg/m³, a dry snow cover was overlain by a coarse-grained layer. Dry snow metamorphism within the snowpack due to gravity, wind as well as temperature and moisture gradients was simulated using a simple approach based on constant density and grain size relationship with profile depths. The initial coarse-grained surface layer subsequently was buried by dry snow maintaining the grain size, while density was adjusted to coincide with the corresponding depths (Fig. 8).

The results displayed in Fig. 9 demonstrate clearly that the backscatter values from such a snow cover remain almost constant above –8 dB during the first years. During this phase, the radar return originates predominantly from volume scattering of the large snow crystals and clusters of the metamorphosed layer, while the contribution of the fine-grained and dry layers on top of it is almost negligible. Depending upon the accumulation rate, and therefore on the thickness of the overlaying dry snow layers, s°-values decrease sharply, when the coarse-grained layer reaches a level corresponding to the penetration depth of the actual snowpack. The time range for this process varies between 8 years (500 mm/a net accumulation) and 20 years (200 mm/a net accumulation) for both modelled snow profiles. The displayed results (Fig. 9) furthermore support the assumption that the backscatter value of –8 dB is a well-suited threshold for delimiting the FPRZ from the adjacent transition towards the DSRZ (PARTINGTON, 1998; RAU & SAURER, 1998).

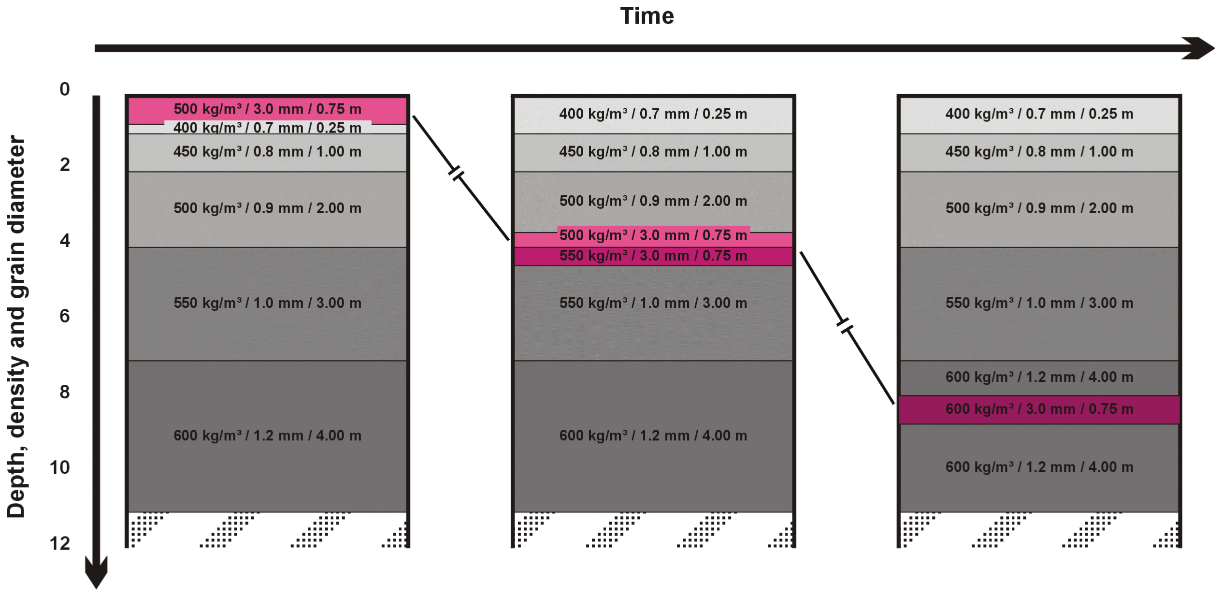

Fig. 8: *Dry snow accumulation on a metamorphosed coarse-grained snow layer. Metamorphism within the snowpack due to gravity, wind as well as temperature and moisture gradients was simulated using a simple approach based on constant density and grain size relationship with profile depths. Typical values for each level are displayed.*

As a consequence, both the FPRZ and the DSRZ near their boundary zone are temporarily and spatially very persistent features in SAR-images. Only the impact of an extraordinary high temperature event seems to be a reasonable impact causing an upward position shift of the transition line. A downward shift of this boundary line is an indicator of constant accumulation during several years revealing stable conditions and the absence of any high temperature event at these altitudes. Hence, analysing the dry snow line with multi-temporal SAR-imagery allows the detection of such singular extreme climatic events posterior to their occurrence and enables the monitoring of periods with constant accumulation conditions.

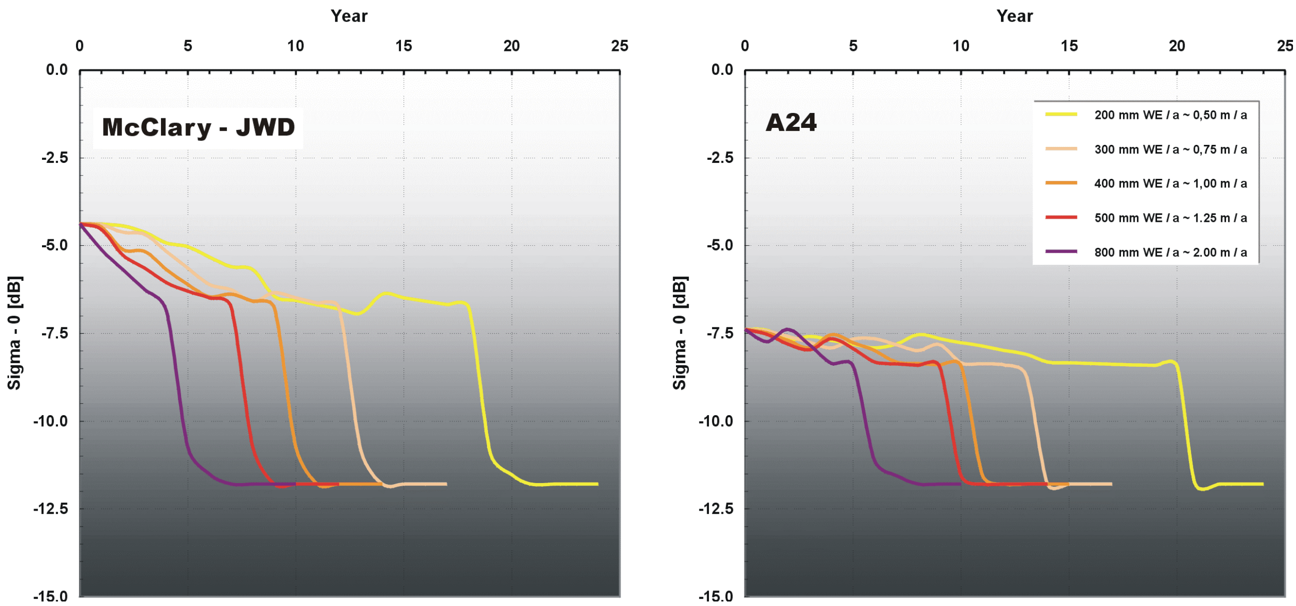

Fig. 9: *Accumulation of dry fine-grained snow on the upper boundary of the FPRZ. When the accumulated top layer exceeds the penetration depths of the snowpack, an abrupt decrease of the backscatter values leads to a downward position shift of the dry snow line.*

CONCLUSIONS

SAR remote sensing techniques have found a wide range of applications in glaciology in recent years. Using synergistic approaches, they complement field studies in remote and inaccessible areas and provide information with a high temporal and spatial resolution. Such a combination of results derived from field studies and remote sensing techniques however emphasises the need for precise definitions and a common terminology.

Based on ground truth data gathered during previous field campaigns to the Antarctic Peninsula, the backscatter characteristics of a dry and frozen snow cover on glaciers were modelled focusing on the lower and upper boundary lines of the FPRZ. The first one, often clearly identifiable in winter SAR-images, has been described as the transition between bare ice and metamorphosed coarse-grained snow. As it is impossible to identify the year of origin of this snowpack, it is suggested to apply the term firn line rather than snow line to this feature. Analysing the position of the late summer firn line position provides information about the glaciers mass balance. It has to be stressed, that the firn line position only occasionally coincides with the position of the equilibrium line, but in the absence of field measurements it can be regarded as a first approximation to it. Furthermore, FLA serves as a crucial input and verification parameter for glacier melt modelling. Special attention has to be drawn to the timing of image acquisition as any metamorphism of the snow, e.g. due to high temperatures or rain during the winter season, might hide the late-summer firn line. Hereby, concurrent meteorological data can help to reconstruct the meteorological conditions prior to and during the image acquisition facilitating further interpretation of the SAR-data.

The dry snow line forms the upper boundary of the frozen percolation zone. Its position is very stable in the time range of approximately 10 to 20 years. Thus, it provides information about the occurrence of singular meteorological events and the accumulation conditions in the higher areas of the glaciers. Additionally, model results confirm the assumption that a threshold of –8 dB is appropriate to delimit the upper boundary line of the FPRZ. This facilitates direct monitoring of the dry snow line.

With the SAR-data archive, which has been built up in recent years and the continuation of polar-orbiting SAR-missions in the future, the operational large-scale monitoring of glaciers and ice sheets has become feasible. It is therefore recommended, that a co-ordinated survey network for glacier monitoring is installed, where mass balance parameters and boundary positions of the different radar glacier zones will be monitored on a regular basis.

ACKNOWLEDGEMENTS

This research was supported by the Bundesministerium für Bildung und Forschung (BMBF) within the programme 'Dynamic Processes in Antarctic Geosystems' (contract # 03PL016A), by ESA projects 'Monitoring of Dynamic Processes in Antarctic Geosystems' (contract # AO2.D149) and ‘Mid-term monitoring of snow cover characteristics by ERS-SAR imagery on the Antarctic Peninsula’ (contract # AO3.D196). The authors would like to thank the Instituto Antártico Argentino (Buenos Aires), the Alfred-Wegener-Institut für Polar- und Meeresforschung (Bremerhaven), and the crew from San Martín for support in respect to logistics, field equipment and assistance.

REFERENCES

Benson, C. S. (1962): Stratigraphic studies in the snow and firn of the Greenland ice sheet. SIPRE Res. Rep. 70; US Army Corps of Eng., Hannover (N. H.).

Bindschadler, R. (1998): Monitoring ice sheet behaviour from space. Rev. Geophys., 36 (1), pp. 79-104.

Braun, M.; F. Rau; H. Saurer & H. Goßmann (2000): The development of radar glacier zones on the King George Island ice cap (Antarctica) during the austral summer 1996/97 as observed in ERS-2 SAR data. *Ann. Glaciology*, 31.

Drinkwater, M. R. (1989): LIMEX '87 ice surface characteristics: implications for C-band SAR backscatter signatures. *IEEE Transactions on Geoscience and Remote Sensing*, 27(5), pp.501-513.

Forster, R. R.; B. L. Isacks & S. Das (1996): Shuttle imaging radar (SIR-C/X-SAR) reveals near surface properties of the South Patagonian Icefield. *J. Geophys. Res.,* 101 (E10), pp. 23,169-23,180.

Friedrich, M. (1996): Der Rückstreukoeffizient sigma° für Schnee und Eis für aktive Fernerkundungssysteme im GHz-Bereich - Abhängigkeit und Berechnung. Unpubl. master thesis, Univ. of Freiburg.

König, M.; J. G. Winther; N. T. Knudsen & T. Guneriussen (this volume): Equilibrium- and firnline detection with multipolarization SAR – first results. Proc. Workshop EARSeL - Special Interest Group ‘Land Ice and Snow’, June 16-17, 2000, Dresden.

Laur, H.; P. Bally; P. Meadows; J. Sanchez; B. Schättler; E. Lopinto & D. Esteban (1997): Derivation of the backscattering coefficient s° in ESA ERS SAR PRI products. ESA/ESRIN, ES-TN-RS-PM-HL09, Issue 2, Rev. 5b.

Meadows, P.; H. Laur & B. Schättler (1999): The calibration of ERS SAR imagery for land applications. EOQ No. 62, pp. 5-8.

Partington, K. C. (1998): Discrimination of glacier facies using multi-temporal SAR data. *J. of Glaciology*, Vol. 44, No. 146, pp. 480-488.

Paterson, W. S. B. (1994): The physics of glaciers (3rd edition). Pergamon Press.

Peel, D. A. (1992): Spatial temperature and accumulation rate variations in the Antarctic Peninsula. In: Morris, E. M. (ed.): The contribution of Antarctic Peninsula Ice to Sea Level Rise. BAS Ice and Climate Special Report No. 1. BAS, Cambridge, pp. 11-15.

Ramage, J. M. & B. L. Isacks (1998): Seasonal changes in Alaskan Radar Glacier Zones. Proc. Final RADARSAT Application Development and Research Opportunity (ADRO) Symp., 13-15th of October 1998, Montreal.

Rau, F. & H. Saurer (1998): Investigations into snow cover dynamics on two glaciers in the central Marguerite Bay (Antarctic Peninsula) using ERS and RADARSAT SAR imagery. Proc. Final RADARSAT Application Development and Research Opportunity (ADRO) Symp., 13-15th of October 1998, Montreal.

Rau, F.; M. Braun; H. Saurer; H. Goßmann; G. Kothe; F. Weber; M. Ebel & D. Beppler (2000): Monitoring multi-year snow cover dynamics on the Antarctic Peninsula. *Polarforschung* 67 (1/2), pp. 27 – 40.

Smith, L. C.; R. R. Forster; B. L. Isacks & D. K. Hall (1997): Seasonal climatic forcing of alpine glaciers revealed with orbital synthetic aperture radar. J. of Glac., Vol. 43, No. 145, pp. 480-488.

Thompson, L. G.; D. A. Peel; E. Mosley-Thompson; R. Mulvaney; J. Dai; P. N. Lin; M. E. Davis & C. F. Raymond (1994): Climate since AD 1510 on Dyer Plateau, Antarctic Peninsula: evidence for recent climate change. *Ann. Glaciology*, 20, pp. 420-426.

Turner, J.; S. Leonhard; T. Lachlan-Cope & G. J. Marshall (1998): Understanding Antarctic Peninsula precipitation distribution and variability using a numerical weather prediction model. Ann. Glac. 27, pp. 591-596.

Ulaby, F. T.; F. T. Moore & A. K. Fung (1986): Microwave remote sensing: active and passive (Vol. III). From theory to applications. Artech House, Norwood.

Wunderle, S. (1996): Die Schneedeckendynamik der Antarktischen Halbinsel und ihre Erfassung mit aktiven und passiven Fernerkundungsverfahren. Freiburger Geographische Hefte, Nr. 48, University of Freiburg.