USING A MULTI-YEAR DATA ARCHIVE OF ERS SAR IMAGERY FOR THE MONITORING OF FIRN LINE POSITIONS AND ABLATION PATTERNS ON THE KING GEORGE ISLAND ICE CAP (ANTARCTICA)

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ABSTRACT

Multi-year SAR data is used to study the seasonal dynamic of the snow melt patterns on the ice cap of King George Island, Antarctica. The snow cover of the entire ice cap becomes wet during summer months, however, in the highest elevations frequent refreeze and melt cycles are observed. Bare ice areas form in the lowest elevations of the ice cap. These findings are in good agreements with energy balance studies and results of a ground penetrating radar survey. Backscatter values in intermediate elevations show a marked rise at the end of the ablation season. This fact is attributed to an increasing surface roughness on a wet metamorphosed snow cover at the end of the ablation season. The influence of thresholds on the determination of the extent of the bare ice radar zone in summer and winter imagery as well as in ascending and descending orbit is investigated. As a consequence, the border of the bare ice radar zone have been mapped using thresholds of –14 dB in summer and –6 dB in winter imagery. It could be shown, that thresholds of –12 dB and –8 dB, respectively, do not change significantly the extent of this radar glacier zone. Firn line altitudes were mapped for several consecutive years. A low inter-annual variation of this line with elevations of about 200 m a.s.l. was observed. The maximum altitude of the firn line was recorded in the mass balance year 1996/97with about 250 m a.s.l. Comparing these values to equilibrium line positions of previous measurements indicate an upward shift in the equilibrium line altitude since the 1970's.

INTRODUCTION

The Antarctic Peninsula and the adjacent South Shetland Islands have been identified as a region with a significant warming trend (e.g. King 1994, Smith *et al.* 1996, Skvarca *et al.* 1998). On the northern Antarctic Peninsula and the South Shetland Islands, snow and ice melt are important mass balance parameters of the small glaciers and ice caps (Bintanja 1995, Braun and Schneider 2000, Braun *et al.* 2001-a). Moreover, melt water has been identified as a probable source for the destabilisation and spectacular disintegration of the ice shelves in that region (Hulbe 1997). On King George Island itself, extended glacier retreat has been observed by several authors (e.g. Braun and Goßmann in press, Park *et al.* 1998, Simões *et al.* 1998, Wunderle 1996) and attributed to changes in glacier mass balance. Therefore, more detailed spatially distributed information on glacier mass balance parameters such as ablation patterns and firn line positions are required,

a) as input and verification parameters for glacier melt models,
b) to improve the monitoring of short-term glacier mass balance changes,
c) to increase the accuracy of long-term predictions of glacier mass balance changes.

Due to the operation of the German Antarctic Receiving Station at the Chilean base O'Higgins, it is now possible to use a multi-year data record of ERS-1/2 SAR imagery for glaciological studies on the Antarctic Peninsula. This time series, starting in 1991, enables the monitoring of the seasonal evolution and inter-annual variation of ablation patterns on the Antarctic ice cap of King George Island as already demonstrated for single years by Braun *et al.* (2000) and Wunderle (1996).

In this study, we utilize ERS SAR data from King George Island to complement glacier melt modelling based on micro-meteorological ground observations, to observe inter-annual variations of the snow cover dynamics and firn line positions on the ice cap and to support the interpretation of a comprehensive ground penetrating radar survey (Pfender 1999).

STUDY SITE

With about 1250 km², King George Island is the largest of the South Shetland Islands. 93 % of the island are ice covered. The islands major ice cap ranges between sea level and about 700 m a.s.l. (Figure 1). More than 30 percent of the island is located in an elevation range below 250 m a.s.l. Low gradient slopes prevail on the north-western part of King George Island, whereas steep fjord-like inlets intersect the smooth surface morphology of the ice cap on the southern side. A detailed outline of the island's topography can be found in Simões *et al.* (1999) and Braun *et al.* (2001-b).

Due to its location in the southern hemisphere west-wind zone, King George Island is subject to a highly maritime climate. The mean annual air temperature (1944-1999) at Bellingshausen Station is – 2.9 °C. Positive air temperatures are recorded throughout all summer months at lower elevations. Snow melt events also occur frequently during winter (Rachlewicz 1997). Generally, advection of warm humid air masses from northerly directions lead to the highest snowmelt rates (Braun *et al.* 2001-a).

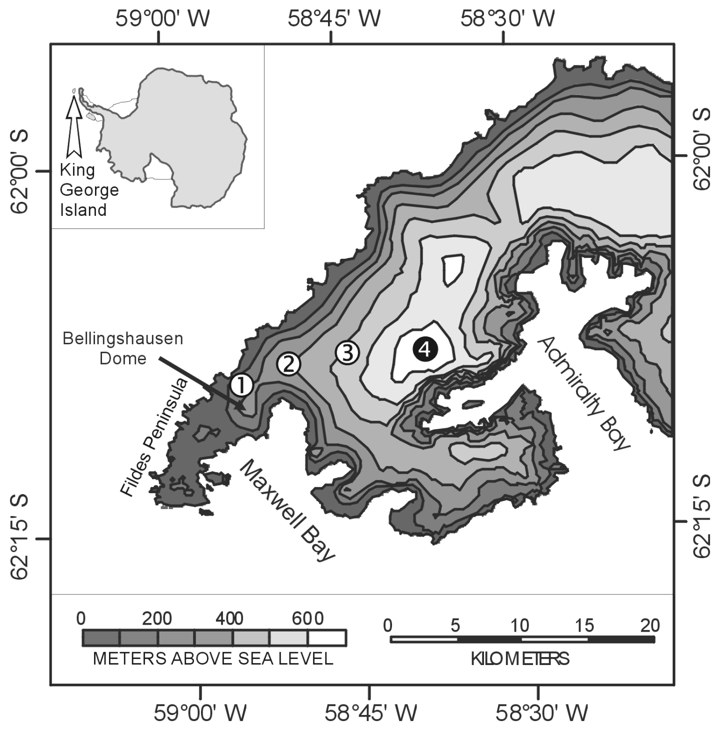

Figure 1: Sketch map of King George Island. Locations of the areas of interest (AOIs) are indicated with numbers

As a consequence of the smooth surface morphology, relief-induced distortions in the SAR imagery are small on the north-western part of the ice cap. Furthermore, changes in the size of the ablation area using satellite remote sensing are more easily detected on moderately inclined surfaces than on steep slopes (Bindschadler 1998). Therefore, the present study will focus on this area of King George Island.

DATABASE AND DATA PROCESSING

For the present investigation, 40 ERS-1/2 scenes were analysed using data from both ascending and descending orbits. The imagery covers the time period between July 1992 and November 1999 (Figure 2). Data preparation was realized using processing chains in the ESA SAR Toolbox. From all images backscatter values were calculated using the algorithm based on Laur *et al.* (1998). In this procedure a compensation for losses during the analogue-digital conversion and a correction for the replica power variation was included. Subsequently, the images were co-registered using a master image from 18 February 1998 for the descending and an image from 15 July 1997 for the ascending orbit.

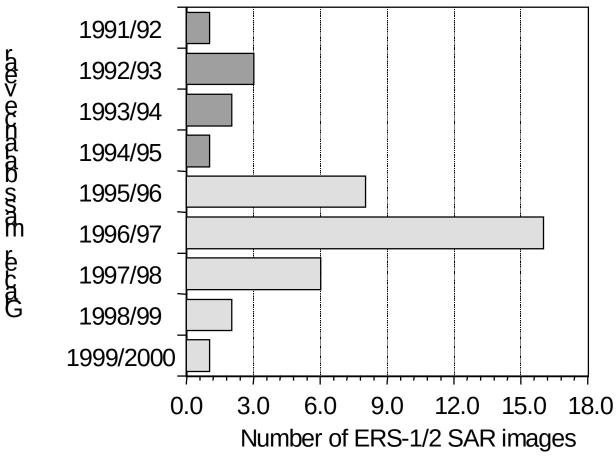

Figure 2: Distribution of ERS 1/2 SAR imagery used in this study in the different mass balance years. Dark grey bars denote ERS-1, light grey ERS-2 scenes.

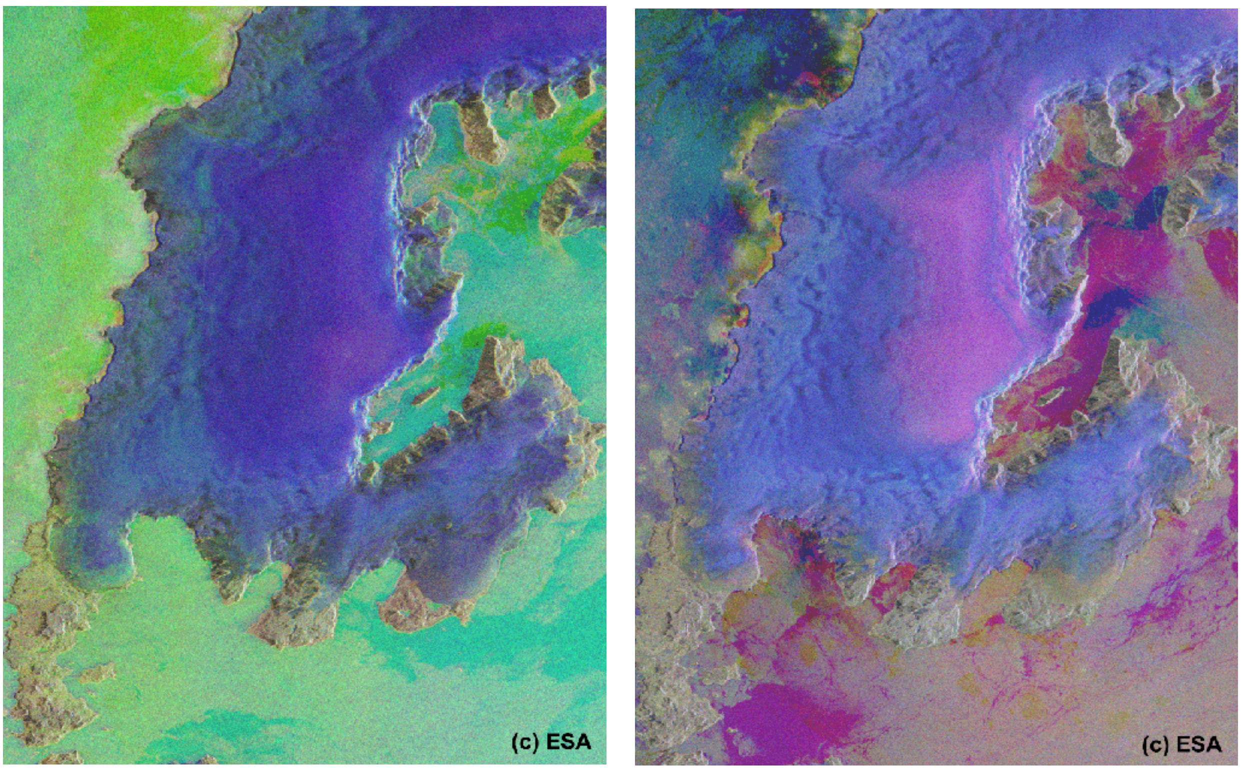

Figure 3: Multi-temporal colour composites of the mass balance years 1996/97 (left) and 1997/98 (right). The imagery was rescaled to a value range of –26 to +6 dB and stretched linearly for this figure. Image identification and colour channel association are summarised in Table 1 and Figure 4.

The timing of the melt onset in November can only occasionally be determined using SAR data due to the limitation in the operation of the German Antarctic receiving station (GARS) in O'Higgins. However, for almost all mass balance years at least one winter scene and several summer images are available. Multi-temporal colour composites for the various years were composed using images with acquisition dates in winter, mid summer and end of summer.

Areas of interest (AOIs) covering about 3000 pixels are defined on the ice cap along an altitudinal profile ranging from 85 m a.s.l. to about 650 m a.s.l. The locations of the AOIs are shown in Figure 1. For all AOIs averaged backscatter values were calculated. In order to determine the firn line positions, thresholds were applied to separate bare ice from the wet snow and the frozen percolation radar zone respectively. Finally, the firn line positions were digitised and superimposed on a digital elevation model (Braun *et al.* 2001-b) to determine the mean elevation of the firn lines. Since the imagery was not corrected for topography, this approach bears errors due to relief induced distortions. However, the errors resulting due to this fact were considered minor. The terminology of radar glacier zones used in this paper is extensively discussed in Braun *et al.* (2000) and Rau *et al.* (2000).

RESULTS AND DISCUSSION

SNOW COVER DYNAMICS AND ABLATION PATTERNS

On King George Island , snow melt starts generally in November and lasts until March (Wen et al 1998) in lower elevations. Figure 3 shows two multi-temporal colour composites generated from ERS-2 imagery of the mass balance years 1996/97 and 1997/98 (image identification is given in Table 1). The colour scheme proposed by Partington (1998) was applied using a winter (July/October) image in the blue, a mid-summer image (January/February) in the red and a late summer image (February) in the green colour channel. The relationship between the backscatter signal on the different dates of image acquisition and colours in the multi-temporal SAR image is given in Figure 4. Generally, blue colours dominate on the ice cap in both images. High backscatter typical for a frozen percolation radar zone (> -6 dB) during winter acquisition and low backscatter (< -15 dB) from a wet snow radar glacier zone lead to this colour developing during summer months. The lowest elevations of the ice cap near Bellingshausen Dome or Potter Peninsula reveal green to brown colours. An increase of the backscatter signal from values of the former wet snow radar zone (< -15 dB) to about –12 to –8 dB indicates the appearance of the bare ice radar zone in the summer imagery in these areas. It has to be mentioned, that the image from 17/02/1997 does not show the real extent of the bare ice zone on that date, since snowfall prior to that date partially covered this radar glacier zone (Braun *et al.*, 2000). Purple colours are formed by high sigma-0 values (frozen percolation radar zone) in at least the winter and mid summer images and lower values in the late summer image. This indicates that occasional refreezing in the late summer image occurs. Similar results on the snow melt dynamics have been presented for the mass balance year 1992/93 by Wunderle (1996). Energy balance studies performed during December and January 1997/98 on various locations on the ice cap (Braun *et al.* 2000) and concurrent ground surveys confirm these findings.

Season	**Winter**	**Early summer**	**Late summer**
Color channel	Blue	Red	Green
1996/97	19/10/96	29/01/97	17/02/97
1997/98	07/07/97	02/02/98	18/02/98

Table 1: Image identification and associated colour channels as used in Figure 3.

To further determine inter-annual variations of the ablation patterns, mean backscatter coefficient within 4 AOIs were calculated from all images and plotted in relation to the month of the mass balance year (Figure 5).

AOI-1 represents the lowest elevations of the ice cap (85 m a.s.l.). This AOI displays a large scatter of the backscatter coefficients with values ranging between –16 and –6 dB. The sigma-0 values of the winter imagery show a very homogeneous behaviour grouping around –10 dB, which is the typical value range of a bare ice radar zone (Rau *et al.*, this volume). A probably existing fine-grained winter snow pack overlying the bare ice is transparent for the radar beam. Similarly, a large amount of the values in the summer months and of the spring imagery show values in this range. Only in November and January values below –15 dB can be found in AOI-1. These low values are caused by a wet snow radar zone covering the bare ice at that time.

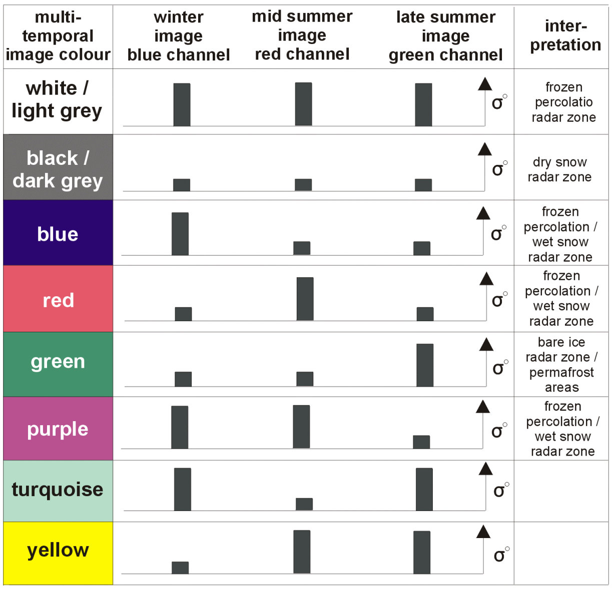

Figure 4: Colours scheme for the multi-temporal SAR images (modified after Partington 1998). Vertical bars indicate backscatter coefficients. The right column gives reference to the radar glacier zones.

AOI-2 and AOI-3, located in an intermediate elevation range (250 and 400 m a.s.l.), show an almost identical distribution of the sigma-0 values – high backscatter during winter months indicating a frozen coarse grained snow pack. The low sigma-0 values during summer months are due to the high liquid water content of the snow pack which causes the absorption of the radar signal. With time a continuous increase of the backscatter values in this wet snow radar zone is observed. This phenomenon shows up in all mass balance years. Braun *et al.* (2000), Ramage and Isacks (1998) as well as Smith *et al.* (1997) have reported that increasing surface roughness of a wet snow cover might be responsible for the increased backscatter values. This situation is a consequence of the continuos snow metamorphism at the end of the mass balance year. Therefore, it might also be the reason for the increasing sigma-0 values in the wet snow radar zone to the end of the ablation season on King George Island.

The backscatter signal from AOI-4 (650 m a.s.l.), which is located at the highest elevation of the ice cap, shows high values in the winter and early spring images. These backscatter coefficients are typical for a frozen, strongly metamorphosed snow pack of a frozen percolation radar zone. However, a large scatter of the sigma-0 values can be found during the summer months. This is attributed to frequent melt and refreeze cycles that lead to a changing backscatter signal according to the actual liquid water content of the snow pack. This interpretation is supported by a comprehensive ground penetrating radar (GPR) survey (Pfender 1999). In the lower elevations a high amount of internal water inclusions and the formation of a pronounced water table at the firn ice transition was observed, whereas in the highest parts of the ice cap lesser water inclusions were recorded. First results from distributed energy balance modelling also show considerably less energy available for melt and frequent refreezing of the snow at these altitudes. In the highest elevations of the ice cap, inter-annual variation of the snow melt dynamics as revealed by SAR imagery are highest, reaching typical values of the frozen percolation radar zone (-4 to –6 dB) in the summer months of the mass balance years 1992/93 and 1997/98.

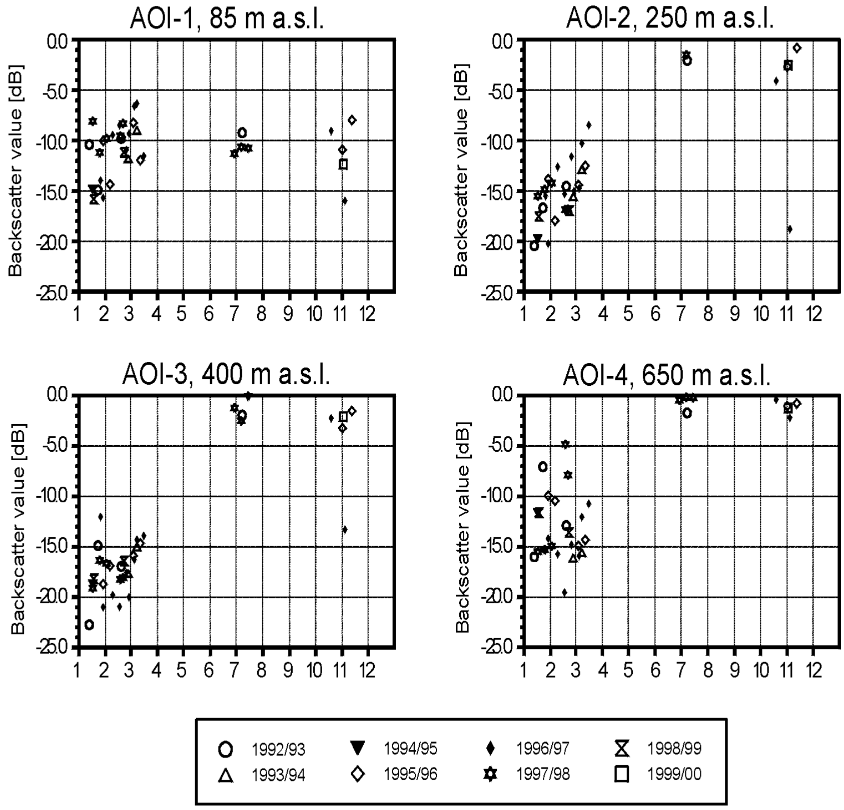

Figure 5: Backscatter values in the four areas of interest (AOIs) of all analyzed images (1992-1999). The mass balance seasons are distinguished by different symbols. AOI1 is located in the lowermost part of the island, AOI-2 and AOI-3 in intermediate elevations while AOI-4 consists of the highest part of the main ice cap of King George Island

Balance year	Location	ELA [m]	Reference
1968-69	G1 (DI)	275	Orheim & Govorukha (1982)
1969-70	G1 (DI) / Bellingshausen Dome (KGI)	290 / 140	Orheim & Govorukha (1982)
1970-71	G1 (DI) / Bellingshausen Dome (KGI)	325 / 170	Orheim & Govorukha (1982)
1971-72	G1 (DI) / Rotch Ice Dome (LI)	370 / 140	Orheim & Govorukha (1982)
1972-73	Stenhouse Glacier, Admiralty Bay (KGI)	< 100	Curl (1980)
1972-73	G1 (DI) / Rotch Ice Dome (LI)	330 /170	Orheim & Govorukha (1982)
1973-74	G1 (DI) / Rotch Ice Dome (LI)	280 /150	Orheim & Govorukha (1982)
1985-86	Bellingshausen Dome (KGI)	150	Jiawen et al. (1995)
1985-89	Nelson Island	110	Jiawen et al. (1995)
1987-88	Admiralty Bay (KGI)	300-350*	Simões et al. (1999)
1991-92	Bellinshausen Dome (KGI)	140	Wen et al. (1994)
1992-93	Hurd Dome (LI)	235	Vilaplana & Pallos (1994)
1965-93	Hurd Dome (LI)	ca. 200	pers. comm. Pourchet 1996

Table 2: Equilibrium line altitudes in the region of the South Shetland Islands as reported by various authors.

*DI: Deception Island, KGI: King George Island, LI: Livingston Island, * denoted value signifies an approximation of ELA by the transient snow line position revealed by analyzing satellite data*

Firn Line Positions

The firn line, separating the bare ice radar zone from the frozen percolation and wet snow radar zone, respectively, has been identified as a first approximation of the equilibrium line altitude (ELA) (Rau *et al.*, this volume). Moreover, firn line positions (FLP) are a key parameter for snow-melt modelling as the distinctions in surface albedo of snow and firn lead to a considerable differentiation of the melt rates. Extracting the FLPs from SAR data on a multi-year basis therefore enhances the knowledge on glacier mass balance considerably. However, a quantitative determination of the firn line altitudes has only rarely been undertaken.

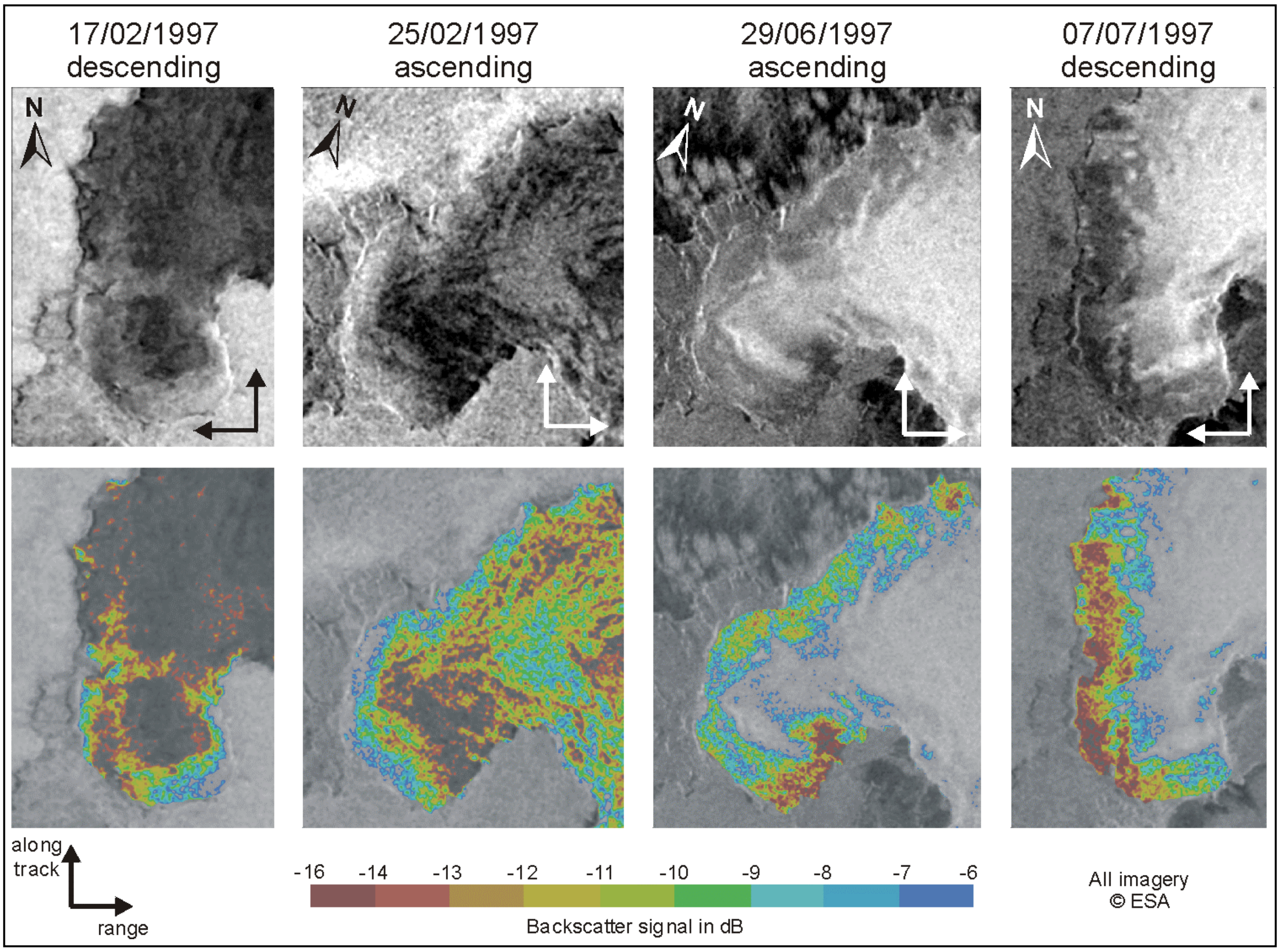

Figure 6: Differences in the determination of the firn line using winter and summer imagery from the ascending and descending orbit for a subset of Bellingshausen Dome, King George Island. In the upper row, the original ERS-2 SAR images are shown, whereas the lower row shows the classified bare ice radar zone using a value range between –16 and –6 dB.

The backscatter values for the bare ice radar glacier zone published in the literature are distributed over a large range (-14 to –6 dB). In Figure 6, the variations of the classification of the bare ice radar zone for the same mass balance year on a subset of Bellingshausen Dome are shown using late summer and winter imagery from both, ascending and descending orbit. The backscatter values are classified in a colour scheme between –16 and –6 dB. For this site, information from our own ground surveys in the months December to January of the mass balance years 1995/96, 1997/98 and 1999/2000 are available. Furthermore measured equilibrium line altitudes have been published for several years from the South Shetland Islands (Table 2). The small ice cap of Bellingshausen Dome addressed to in the following covers an elevation range from 35 m a.s.l. to 250 m a.s.l..

In order to enable an objective outline of the firn line in all images, the value range between –6 to –16 was classified (Figure 6). These thresholds were varied to determine the persistence of the classified area in relation to the threshold value, but no considerable changes in the FLP could be determined. The classified bare ice zone agrees well with ground observations from the previous and the following year. In the ascending as well as in the descending orbit, the boundary between bare ice and the frozen percolation radar zone is marked by values of about –8 to -6 dB in the winter imagery, whereas in the summer imagery the radar zones are separated by the lowest values of –14 to -16 dB (Figure 6). This indicates that for winter and summer acquisition dates different thresholds have to be applied to discriminate the bare ice radar zone.

Both images from the descending orbit show good agreement regarding the classified bare ice areas. Only in the northern part, minor changes are caused by a snowfall event before 17/02/1997, which partially covered the bare ice. The classification in the ascending image from 26/06/1997 coincides with the results of the descending orbit. However, the ascending imagery from 25/02/1997, marking the maximum bare ice extent in this season shows only very little agreement with the other imagery. The maximum elevation of firn line in this image at about 250 m a.s.l. agrees with measurements from the GPR survey in the following year by Pfender (1999). Consequently, it has to be supposed that the winter images from 1997 do not show the real extent of the bare ice zone at the end of the former ablation season. Rachlewicz (1997) stated that advection of warm humid air masses can even in winter lead to a complete disappearance of the snow cover at lower elevations. It might be the case that, prior to the image acquisition in June 1997, such an advection event melted completely the autumn snow cover in the lower elevations of Bellingshausen Dome and metamorphosed the snow pack in the higher parts. Cold temperatures in the following days and during image acquisition led to the typical backscatter values of a frozen percolation radar zone in the upper and a bare ice radar zone at the lower elevations.

Glacier mass balance year	**Date of Image acquisition**	**Orbit**	**Firn line altitude [m]**
1991/92	08/07/1992	Desc	160
1992/93	04/03/1993	Desc	200
1993/94	08/03/1994	Asc	200
1995/96	12/03/1996	Asc	200
1996/97	25/02/1997	Asc	250-270
1997/98	21/02/1998	Desc	180-200
1998/99	22/02/1999	Desc	200-220

Asc: Ascending orbit; Desc: Descending orbit

Table 3: Altitudes of the transient firn line as revealed from ERS SAR imagery for Bellingshausen Dome on King George Island.

This example demonstrates that timing of image acquisition is of major relevance for the determination of the FLP. However, for regions like King George Island where days with air temperatures above 0°C occur in almost all winter months, the use of late summer imagery is inevitable for the determination of the FLP. Hence, for the following analysis of FLPs late summer images from ascending and descending orbits were preferred.

The analysis of the entire set of ERS SAR images reveals no large inter-annual variations of the firn line altitude (Table 3). Generally, the values range between 180 and 220 m a.s.l. in almost all mass balance years of the 1990's. The highest altitude of the firn line could be observed for the mass balance year 1996/97 with about 250 m a.s.l. As stated by Braun *et al.* (2000), summer air temperatures in this year were well above the long-term average and snowmelt started already in early November. The ELA measurements for the mass balance year 1991/92 on Bellingshausen Dome (Table 2) are in good agreement with observed firn line altitude in the radar imagery. The small differences may be caused by the location of the point measurements and that ELA includes superimposed ice, whereas in the FLA determination this parameter is not considered. Using the derived firn line altitude as an approximation of the ELA and comparing them with the ELA values of table 3, a rise of the ELA/FLA of about 50 m seems to have taken place since the 1970's.

CONCLUSIONS

Analysis of multi-year ERS SAR data enables the monitoring of ablation patterns and inter-annual ablation variations of snow melt dynamics. The highest elevations of King George Island are subject to frequent melt refreeze cycles, whereas in the lower elevations a wet snow and bare ice zone form during summer months. In an example, the differences in the classification of the bare ice zone depending on acquisition date and orbit were demonstrated. In winter images, the bare ice radar zone is separated by sigma-0 values of about –6 to –8 from the frozen percolation radar zone, whereas in summer scenes, values of about –16 to –14 separated wet firn and bare ice. Therefore, different thresholds have to be applied in summer and winter imagery for the analysis of a multi-year data archive. The strong influence of winter melt events on the determination of FLPs from winter imagery have been demonstrated. As a consequence, the use of late summer or early winter images are recommended for regions with frequent advection of warm air masses during winter months. The firn line positions in the 1990's around 200 to 220 m a.s.l. as determined from the SAR imagery revealed low inter-annual variations. A highest firn line altitude of about 250 in the mass balance year 1996/97 was observed. However an increase in comparison to ELA measurements from the 1970's and 1980's could be identified. The results on snow melt dynamics and FLP are in good agreement with results of energy balance modelling (Braun *et al.* 2001-a).

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