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Snow Research in the Dronning Maud Land within the Finnish Antarctic Research Program in 1989–2014

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Abstract

Snow research has been carried out at the Finnish Antarctic station Aboa (73°02.5'S 13°24.4'W) and in its surroundings in the Dronning Maud Land since it was established in 1989. In 1999 the work expanded to cover the spatial variability of snow and its properties in the snow and ice surface layer (0–10 m) along a 500 km long transect passing Aboa from the shelf edge in the Weddell Sea to the edge of East Antarctic Plateau. The topics have been the mass and heat balance of snow, remote sensing, stratigraphy of snow cover, physical and chemical properties of snow, solar radiation budget, and snow on nunataks. Net accumulation of snow has been typically 200–300 mm of water equivalent per year, increasing towards the oceanic boundary. The distance from the ocean is the most predominant factor controlling the variations in the snow properties. Aerosol chemical composition, physical properties and processes from their origin to deposition have been studied mainly in Antarctic summers at Aboa.

Keywords: snow, Dronning Maud Land, stratigraphy, mass balance, solar radiation, snow chemistry, nunataks

1 *Introduction*

Antarctica is a key component in the Earth climate system, acting as a large heat sink. The continent is cold due to several factors. In its geographical location, the annual level of incoming solar radiation is low due to the long, dark winter and low solar elevation during the polar day. Terrestrial radiation losses are large due to the low cloudiness and atmospheric moisture content. The westerly wind zone and the Antarctic Circumpolar Current transport air and ocean waters clockwise around the continent at 50–60°S and limit the transport of warm air and water from lower latitudes further south. The mean elevation of the Antarctic continent is close to 3000 m above sea level, and over vast parts the adiabatic lapse rate places the surface temperatures around 30°C lower than they would be at the sea level. Additionally, the large-scale atmospheric circulation brings cold descending air close to the geographic pole, where the air begins to flow down the ice sheet in katabatic winds.

The climatic conditions maintain the snow and ice cover that blankets the Antarctic continent up to 98% of the surface area. These vast snowfields maintain the high reflectance, with albedo as much as 90%. Snowfall constitutes the input of mass into the Antarctic ice sheet (only a very small fraction of precipitation comes down in liquid phase). Snow builds up on the ice sheet, transforms into ice deeper down, and flows towards the surrounding ocean. The mass loss of the ice sheet is due to calving of icebergs and basal melting of the ice shelves, but mass is also lost due to sublimation and runoff, the former being more significant. Due to the remote location of Antarctica, the snow cover there represents 'almost pure snow'. Sources of impurities are primarily the ocean around the continent and nunataks. In addition, primary production takes place in supraglacial and epiglacial lakes, and cyanobacteria communities form an impurity source for the atmosphere (*Kyrö et al.*, 2013). In all the resulting impurity levels in snow are very low.

Ten meters thick surface layer of the ice sheet in the Dronning Maud Land (DML) has been one of the main research topics within the Finnish Antarctic Research Programme (FINNARP) since 1999. This surface layer, corresponding to net snow accumulation over 10–100 years, has a key role in many climatic and environmental questions in Antarctica. Snow is at the atmosphere–ice sheet interface, where interactive processes take place. Snow properties readily respond to changing environmental conditions, and becoming buried by new snow, individual snow layers record the past climate. To relate data obtained from the deep ice cores to environmental and climatic conditions, we need more information about how the parameters describing the conditions when the snow has originally been accumulated are preserved in the snow surface layer. These parameters include the chemical and physical properties of snow, temperature variability in snow, and snow mass balance. Chemical properties of the snow cover can tell about the characteristics of atmospheric transport in the region, and they can influence on the physical properties of snow. For example, the snow reflectance is highly influenced by deposition of the atmospheric black carbon on the snow. Signals received from the snow cover by satellite remote sensing are sensitive to the snow properties. Snow albedo, for example, depends on the liquid water content and grain size of snow in an interactive manner.

This paper presents an overview of the snow research done in the western Dronning Maud Land, mainly in the vicinity of the Finnish *Aboa* station in its first 25-year period, 1989–2014. The main themes have been the snow mass balance, stratigraphy of snow cover, surface heat balance, solar radiation transfer, and snow chemistry. The seasonal and perennial snow on nunataks has also been examined. Sampling and manual work has been performed in the summer season, while automated snow stations have taken care of the all-year data collection.

2 *FINNARP snow research in DML*

2.1 *FINNARP*

Finland joined the Antarctic Treaty in 1984, and the consequent research work was initiated in 1988. A year later the Finnish Antarctic research station *Aboa* was established (Fig. 1). The station is located in the western Dronning Maud Land on Basen nunatak, on bare ground, at 73°02.5'S 13°24.4'W, elevation 485 m above the sea level. Basen is located near the grounding line of the ice sheet. It is the most northern nunatak of the Vestfjella mountain range, which is aligned approximately parallel to the coast. The top of Basen is 584 m above the sea level. The closest nunataks are Plogen (25 km distance) and Fossilryggen (50 km distance). Finnish Antarctic research was first financed by the Ministry of Trade and Industry, and since 1998 the Academy of Finland has been the responsible funding organization. The funding has been based on 3–4-year projects. The logistics has been taken care initially by the Finnish Institute of Marine Research, fused in 2009 to the Finnish Meteorological Institute. Fieldwork has covered the sector 10–15°W in the Dronning Maud Land from the shelf edge in the Weddell Sea (70°S) 500 km south passing Aboa, to 75°S.

Fig. 1. Research station Aboa in Basen nunatak. Plogen nunatak, 25 km from Basen, is seen in the background.

Climate in the research area is dominated by three factors: 1) Strong seasonal cycle, with polar night and day 2 ½ months each; 2) Occurrence of katabatic winds generated by descending cold dry air from the southern polar plateau in the large-scale atmospheric circulation; and 3) Eastward circulating cyclones around the continent bringing heat and moisture to the coastal margin, which is subjected to snow storms throughout the year. *Kärkäs* (2004) reported weather statistics from the Aboa automated weather station for years 1989–2001. The mean annual air temperature was -15.0°C with the standard deviation of 0.9°C. The air temperature experienced a pronounced seasonal cycle: the mean summer temperature (DJF) was -6.4°C and the mean winter temperature (JJA) was -20.9°C. The range of monthly mean air temperatures was from -21.9°C (August) to -5.2°C (January). In a nearby (10 km) automatic weather station (AWS-5) on the ice sheet, at 300 m lower altitude than Aboa weather station, the mean annual air temperature was -17.1°C (*Reijmer*, 2001). Thus inversion showed clearly up in the mean air temperature, which increased from the surface of the ice sheet upward to Aboa weather station.

Also the wind speed had a seasonal cycle in *Kärkäs* (2004). The highest seasonal average was in winter (JJA), 8.6 m s^{-1}, while in summer (DJF) the mean was 5.9 m s^{-1}; the range of monthly mean wind speeds was from 5.4 m s^{-1} (January) to 9.1 m s^{-1} (June). Basen nunatak affects the local wind field, making the directional distribution of winds more strongly peaked than in the nearby ice sheet. The geometry directs the winds along north-easterly and south-westerly directions, with the mode in north-northeast (30°). Only strong katabatic wind events showed clear anomalies in the wind direction, changing the wind to southerly directions. Although Basen influences on the climatic conditions, the weather station provides a good reference for the surrounding area.

2.2 Snow research

The area of snow research (Fig. 2) covers a sector at Aboa meridian starting from the coastal margin across 50–100 km wide Riiser-Larsen ice shelf. Basen nunatak lies near the grounding line at the end of a sub-glacial peninsula, and then the ice sheet elevates gently further south to the next rock outcrop of Fossilryggen. Thereafter, at about 150 km further south the ice sheet ends beneath Heimefrontfjella mountain range, which is the barrier toward the East Antarctic Plateau. Several blue ice areas exist in vicinity of the mountain range, and FINNARP projects have also carried out extensive research campaigns in this region (*Sinisalo and Moore*, 2010; *Sinisalo et al.*, 2004; *Grinsted et al.*, 2003; *Sinisalo et al.*, 2003b). The two topographic highs are the *Kvitkuven* near the ice shelf edge and the *Högisen*, around 50 km southwest from Basen.

Snow became one of the first topics in Finnish Antarctic research in DML. In the first expedition, austral summer 1988/1989, a study on the vertical snow temperature distribution was conducted from two 10 m deep boreholes and from one 2 m deep snow pit to examine the stabilization of temperature during borehole measurements and to determine the mean annual temperature over the ice shelf (*Seppälä,* 1992). In the following year snow conditions were examined in the Weddell Sea as a part of sea ice research (*Granberg and Leppäranta*, 1999). Specific snow projects “Seasonal snow in Antarctica I and II” were funded for 1999–2005 (*Rasmus et al.*, 2003; *Kanto et al.*, 2007) and “Evolution of snow cover and dynamics of atmospheric deposits in the snow in the Antarctica" for 2009–2012 (*Järvinen et al*., 2012). Three PhD theses (*Kanto,* 2006; *Rasmus,* 2009; *Järvinen*, 2013) and three MSc theses (*Kärkäs*, 2000; *Koistinen*, 2007; *Mattila*, 2007) have been produced from these snow research projects. An analysis of summer snow cover in the Aboa region was presented by *Vihma et al.* (2011). More recently, seasonal and perennial snow cover on nunataks has been examined by *Leppäranta et al.* (2013a).

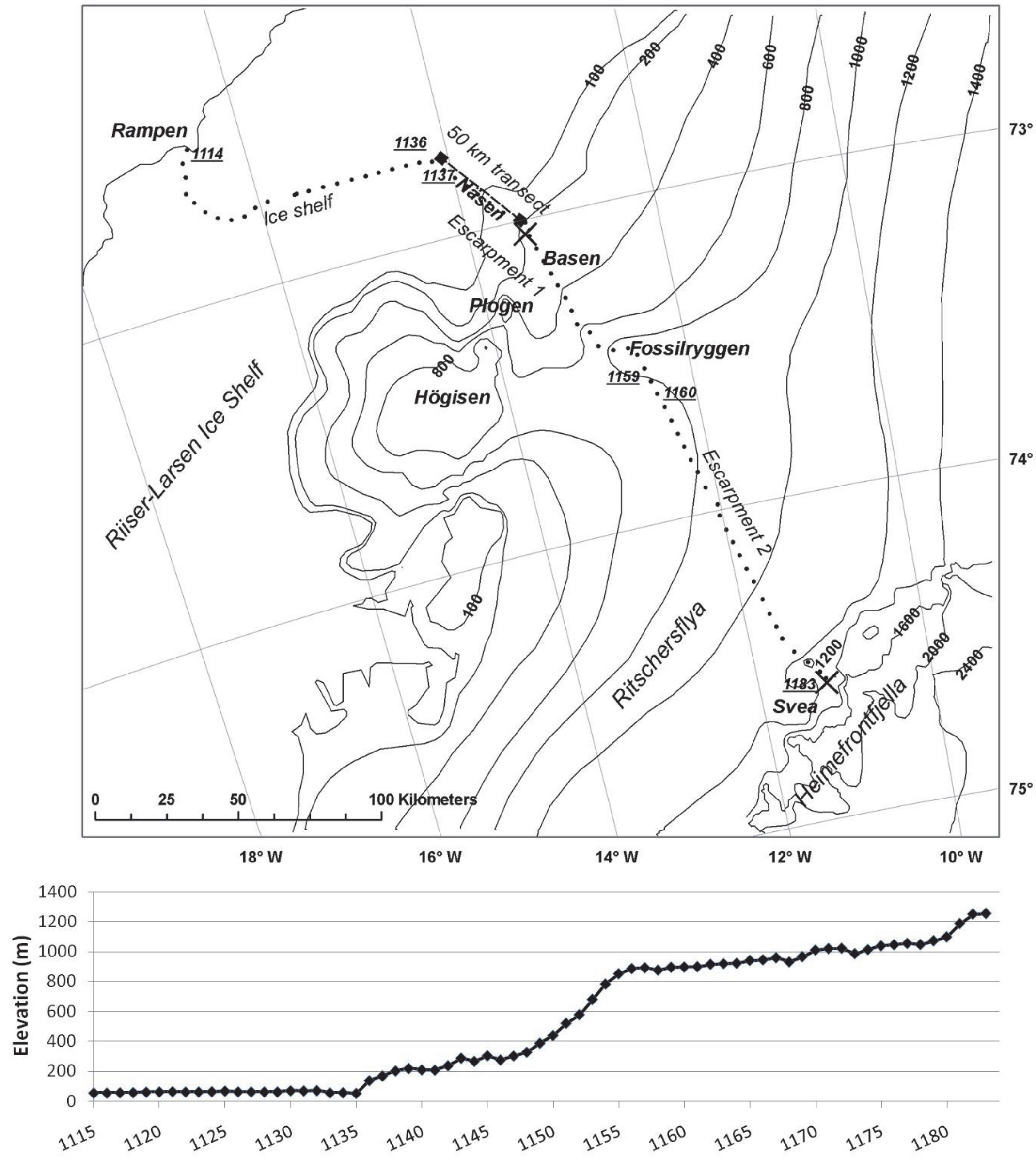


Fig. 2. a) Map over the research area (top). The north–south study section with locations of two snow observations sites (black dots) are also shown (*Järvinen et al.*, 2013). These observations are reported . b) Surface elevation variability along the research profile (dotted line in Fig. 2a) from the edge of the ice shelf to the edge of the East Antarctic Plateau.

Snow cover has been also of major interest in other FINNARP projects. Parameterization of snow albedo has been studied by observations for climate models (*Pirazzini*, 2004, 2008) and in atmospheric boundary-layer investigations (*Vihma et al.,* 2009).

Antarctic moisture budget was examined by *Tietäväinen and Vihma* (2008) based on ERA-40 reanalysis data. Snow accumulation near gravimetric stations with sensitive instruments needs to be accounted for, when the gravimetric method is used to examine change in the mass balance of the ice sheet (*Koivula and Mäkinen*, 2003; *Mäkinen et al.*, 2007; *Ruotoistenmäki and Lehtimäki*, 2009). Snow accumulation rate is also of interest when studying the dynamics of blue ice areas, and in the effort of extracting climatic records from these areas (*Grinsted et al.*, 2003; *Sinisalo et al.*, 2003b; *Sinisalo et al.*, 2004; *Sinisalo,* 2007; *Sinisalo and Moore, 2010*). Also the structure and biology of epiglacial and supraglacial lakes is related to the local snow conditions (*Keskitalo et al.*, 2013; *Leppäranta et al.*, 2013b). Comparison of atmospheric aerosol samples and chemical samples from snow pits brought new light to the formation of aerosols, their transport, origin and deposition (*Kärkäs et al.*, 2005b; *Kanto*, 2006). The information can further be used in interpretation of deep ice cores.

3 Mass balance of snow

3.1 Snow accumulation and sublimation

The mass balance of the snow layer is given in snow water equivalent (SWE), which equals the volume of water per area corresponding to the mass of snow (*M*). Net accumulation is due to local exchange with atmosphere and transport, written as:

$$\dot{M} = P - E - R + T \tag{1}$$

where *P* is precipitation, *E* is sublimation, *R* is runoff, and *T* is transport due to snowdrift. The dimension of Eq. (1) is length per time (normally in millimetres per day, month or year). Runoff water may be re-used in snow metamorphosis or can flow to open land or to ocean. In our research area, local runoff existed only in nunataks.

Automatic snow stations were deployed for all-year records of net accumulation by *Granberg et al*. (2009) in 1999–2000 in the research section from the shelf edge to the plateau. The normal level was 150–220 mm SWE in a year; the minimum was 52 mm SWE in the polar plateau at 75ºS, and the maximum was 897 mm SWE in Högisen dome due to snowdrift. These automatic stations consisted of thermistor strings, and the vertical temperature gradient was used to determine the snow surface (*Granberg et al.*, 2009). Autumn and winter snow storms accounted for most of the annual net accumulation, which occurred in events spaced 3–5 weeks apart (Fig. 3). The automatic station program was re-started in 2009 using two sites (at Aboa and near the shelf edge) with successful outcome (*Järvinen et al*., 2013), and then the annual accumulation was 345 mm SWE at Aboa.

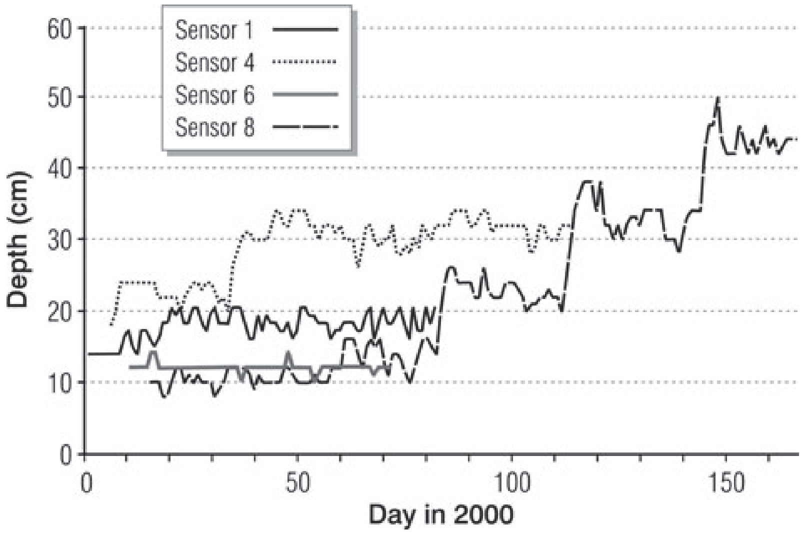


Fig. 3. Snow depth time series from a snow station experiment showing net accumulation of snow in four sites along the study section in the year 2000 (*Granberg et al.*, 2009).

Accumulation rate has also been determined with other methods in the research region. An analysis based on oxygen isotopes from three years (1999, 2000 and 2003) showed that annual accumulation varied between 162 and 355 mm SWE, when the polar plateau was excluded (*Kärkäs et al.*, 2005b). It was strongly correlated with distance to the sea, decreasing towards the interior of the continent due to reduced effect of circumpolar low-pressure systems. Within individual sites the inter-annual variation showed a large range from 19 mm to 145 mm SWE and differences between four sampling pits at roughly 100 m distances apart were 5 to 19 mm SWE. This suggests that small-scale processes (characteristic length scale below 100 m) are important in the spatial variability of the accumulation rate. Net accumulation rate was monitored from snow pits (*Kärkäs et al.*, 2002; Fig. 4). Snow pit and stake measurements in the vicinity of Aboa (Fig. 1) showed that the accumulation rate might vary by 200 mm SWE within a distance of one kilometre in this region (*Sinisalo et al.*, 2003a).

Snow patches on Basen nunatak were monitored using snow stakes in summer (*Leppäranta et al.*, 2013a). During 2004–2010 perennial snow patches lost snow on average about 15 mm SWE annually. The mean net accumulation rates in Scharffenbergbotnen valley in Heimefrontfjella (Fig. 1) were measured to range only a few kilometres apart over 6- and 10-year periods from -138±43 mm SWE on blue ice to 45±161 mm SWE on the accumulation area, respectively (*Sinisalo et al.,* 2003b; *Grinsted et al.,* 2003).

Sublimation has been determined using snow stakes and estimated from atmospheric surface layer measurements in summer (*Leppäranta et al.*, 2013a):

$$\frac{dh_w}{dt} = \frac{\rho_s}{\rho_w} \cdot \frac{dh_s}{dt} = \frac{\rho_a}{\rho_w} C_E \left(q_0 - q_a\right) U_a \tag{2}$$

where h_w is the surface layer SWE, h_s is the thickness of snow, ρ_a, ρ_s and ρ_w are the densities of air, snow and water, respectively, t is time, C_E is the turbulent transfer coefficient for water vapour, q_a and q_0 are the specific humidities in air and at the surface, respectively, and U_a is the wind speed. Here h_w and h_s are taken with respect to an arbitrary, fixed reference. The summer level of sublimation has been found to be 0–5 mm SWE per day, averaging to 1–2 mm per day, but it is likely that in winter the level would be much less because of the very low air temperatures.

Fig. 4. Profiling the complex dielectric permittivity using the snow fork in a snow pit. The measurements provide estimates for the density and liquid water content of snow (*Sihvola and Tiuri*, 1986).

3.2 Snow density

Snow density is needed for mass budget evaluation from snow stakes and for estimation of snow accumulation rate with the oxygen isotope method. Actually, nearly all the methods to estimate the accumulation rate require snow densities to be accurate. Snow density also has a strong influence on the thermal conductivity of snow, makes a strong contribution to the mechanical and electric properties of snow, and influences the transport of chemical constituents in snow.

Most of the density observations during FINNARP expeditions have been conducted in 50–200 cm deep snow pits (Fig. 5). This surface layer corresponds to 1–5 years' accumulation and is strongly subjected to wind driven processes. As a result, layers with heterogenic structure within the surface snow cover appear. *Kärkäs et al.* (2005b) reported mean density in the top meter from three measurement seasons as 394 kg m^{-3} with standard deviation of 26 kg m^{-3}. The measurement set consisted of 17 snow pits well scattered in the research area covering the ice shelf, coastal margin, polar plateau as well as the ice rises of Högisen and Kvitkuven. Spatially, snow density in the ice shelf and in the coastal margin displayed no correlation with distance to the coast or elevation. The polar plateau and the topographic rises showed lower densities by magnitude of 10 kg m^{-3}. They were also the only sites displaying a clear increase of density with depth in the top layer. Snow accumulation and differences in the wind conditions were likely the controlling factors for these differences.

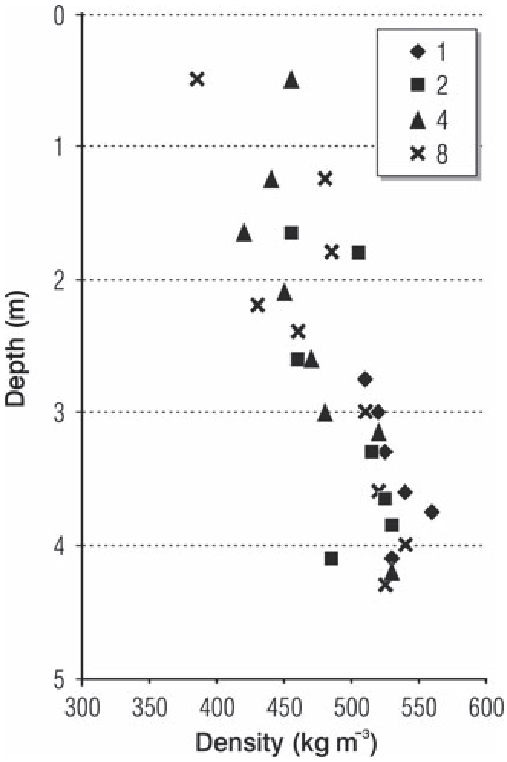


Fig. 5. Snow density profiles in the top 5-m layer measured in the research area snow sites in austral summer 2000 (*Granberg et al.*, 2009). The profiles are typical for the region.

The density of the top 50 cm layer was also observed along the traverses over the ice shelf and the coastal margin, excluding the polar plateau and local topographic highs, with 5 km intervals from shallow snow pits (Fig. 4). The density at the surface is quite homogenous averaging to 396 kg m^{-3} with standard deviation of 6 kg m^{-3} (*Mattila*, 2007).

Snow density increases with depth due to snow metamorphism, especially due to denser arrangements of snow grains and pressure of the overlying snow. Snow density profile is taken often in the shape suggested by (*Schytt*, 1958):

$$\rho = \rho_0 + [1 - \exp(-\mu z)](\rho_i - \rho_0) \tag{3}$$

where ρ_0 is density at the surface, ρ_i is ice density in the deeper ice sheet, μ is the inverse vertical length scale of density increase, and z is vertical co-ordinate positive downward. This equation is the solution of the linear differential equation:

$$\frac{d\rho}{dz} = \mu(\rho_i - \rho) \tag{4}$$

FINNARP results suggest that the surface density is ρ_0 = 395 kg m^{-3} and the vertical length scale is μ^{-1} = 10 m (*Granberg et al.*, 2009). The parameter $\mu$is determined by snow metamorphosis and therefore should depend on the climatic conditions.

4 Stratigraphy and properties of snow

4.1 Stratigraphy

Snow accumulation and metamorphism induce stratigraphy into snow cover (*Colbeck et al.*, 1990; *Colbeck*, 1991). In the present study area, annual accumulation ranged from 300 mm SWE at the shelf edge to 52 mm SWE in the polar plateau. The higher rate over the ice shelf increases the thicknesses of different snow layers, as observed in the snow pits (*Kärkäs et al.*, 2005b). The ice shelf goes through stronger melting phase in summer, displayed in stratigraphic profiles as 1–5 cm thick ice lenses. Further inland the thickness of these lenses is just a few millimetres. Stratigraphy plays an important role in the heat and water vapour fluxes within snow, creating another characteristic feature for the surface snow, namely the abundance of depth hoar layers associated with ice or high-density layers and strong temperature gradients (*Colbeck*, 1991). The ice layers also act as strong backscatterers for radar signals.

4.2 Physical properties of snow

The grain size distribution was determined from a sample set of over 10,000 snow grains by *Kärkäs et al.* (2005b) (Fig. 6). The results showed an exponential decrease with distance to the shelf edge. The median grain size was 1.0 mm (the size was determined from photographs as the longest diameter of the grain). Most (73%) of the measured grains were well rounded, supporting the assumption that Antarctic snow cover consists to a large extent of small and well-rounded grains. Grain size and shape have also great significance for remote sensing. The smaller the grains are, the larger will the light reflectance be. Faceted and cup-shaped surface and depth hoar crystals produce higher backscatter than well-rounded crystals.

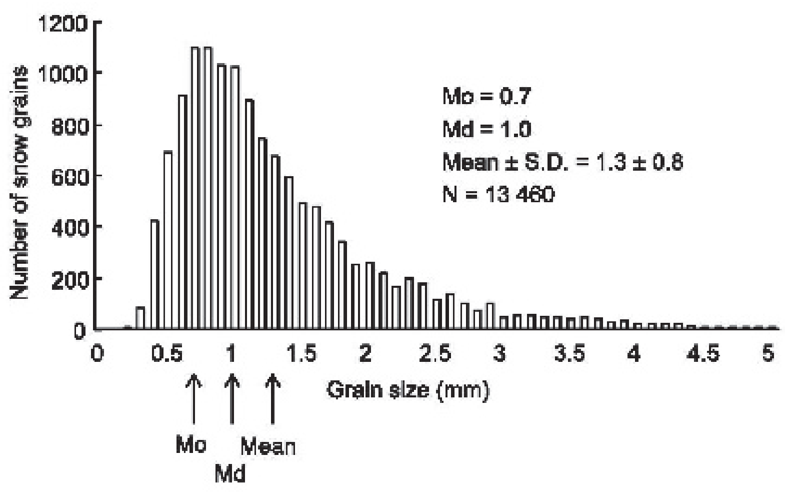


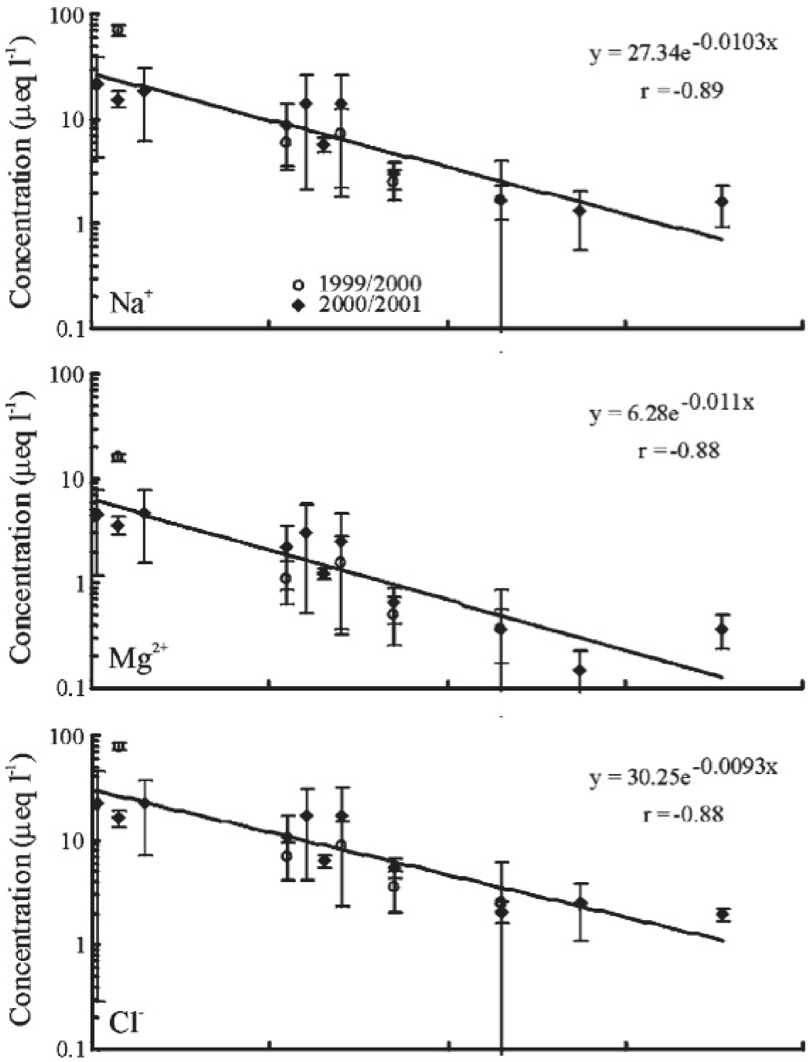
Fig. 6. Grain size distribution in the research area (*Kärkäs*, 2005b).

In our research area, snow is homogeneous, grains are well rounded, and the level of impurities is low. Therefore simple material models for the physical properties and processes work well. In particular, this kind of snow applies well for estimating the density and liquid water content from the complex electric permittivity, ε (see *Sihvola and Tiuri*, 1986). The 'snow fork' manufactured by Toikka Ltd. has been used for these measurements (Fig. 4). The density of snow, determined from the real part of ε, agreed rather well with direct measurements and provided a fast method for the density profiling. There was, however, a small systematic bias. Liquid water content is expected to come out from the real and imaginary parts of ε; however, no accurate reference measurements are available at our station for calibration but qualitative information and first-order comparisons look realistic.

4.3 Chemical properties of snow

Chemical analysis of the snow cover in the DML has included oxygen and hydrogen isotopes and particulate impurities or snow sediments. The proportions of the isotopes provide information of annual layers and therefore support stratigraphy analyses. There were very low but observable levels of impurities due to atmospheric deposition. The main sources were the Weddell Sea, providing marine salts, nunataks, and melt-ponds. Atmospheric deposition and interpretation of chemical signals from snow and ice cores requires detailed knowledge of the processes affecting the accumulation rate of the compounds. Extensive work on snow pit chemistry has been conducted for major ionic compounds by *Kärkäs et al.* (2005a) and *Kanto* (2006). Several studies of the chemical composition and physical properties of aerosols have also been conducted in austral summers at Aboa (*e.g., Koponen et al.,* 2003; *Virkkula et al.,* 2006b; *Virkkula et al.,* 2009; *Kyrö et al.*, 2013). It is important to know how does the chemical composition of particles change when marine aerosols are transported inland and how this is related to the coastal – inland gradient of the snow chemistry.

Air masses are transported to the western DML from the Weddell Sea mainly by cyclonic activity. Over the coastal margin the deposition of particles is through precipitation events, whereas over the plateau the dominant mechanism is dry deposition, *i.e.* deposition of particulate and gaseous components onto the surface in the absence of precipitation. Together with snow accumulation, the deposition related to precipitation is negatively correlated with distance from the coast. Exponential relationships were found for ionic concentrations as a function of the distance (Fig. 7) with high correlation for the following components of ions: sodium (Na^+), magnesium (Mg^{2+}), chloride (Cl^-) and sulphate (SO_4^{2-}). Correlation also existed for non-sea-salt sulphate ($nssSO_4^{2-}$) and methanesulphonic acid (MSA, $CH_3SO_3^-$), which is believed to be of marine biogenic origin (*Kärkäs et al.*, 2005a).


Concentration (μeq l⁻¹)
100
10
1
0.1
y = 27.34e^-0.0103x
r = -0.89
Na+
1999/2000
2000/2001
y = 6.28e^-0.011x
r = -0.88
Mg2+
y = 30.25e^-0.0093x
r = -0.88
Cl-

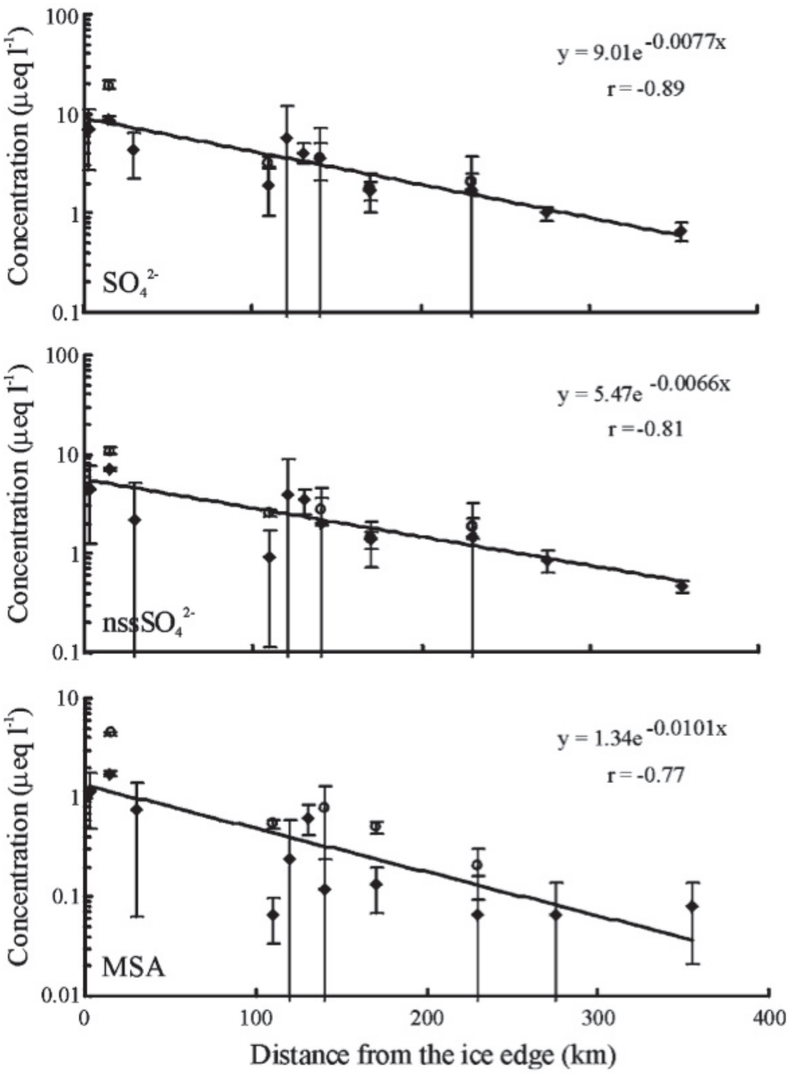


Fig. 7. Relationship between the most common ions in the surface snow pack and distance to the coast (*Kärkäs et al.*, 2005a).

5 *Heat budget of snow cover*

5.1 *Snow temperature*

Temperature and temperature gradient are the main drivers of post-depositional processes in the surface snow layer. The first snow study within the FINNARP-program, including temperature measurements in the research area, was executed during austral summer 1988–1989 (*Seppälä*, 1992). The main purpose was to investigate the time lapse needed to record temperatures from boreholes, when the surrounding snow was heated during the drilling of the borehole (Fig. 8). The research also produced an estimate of the mean annual temperature (the temperature at 10 m depth is normally taken as an estimate for the mean annual temperature). In the ice shelf it was found to be -18.3°C. The annual means in Aboa and German station Neumayer at the shelf edge were -15.0°C and -16.0°C, respectively, in 1989–2001 (*Kärkäs*, 2004). The standard deviations of the annual means were 0.9°C at Aboa and 0.8°C at Neumayer, Aboa was warmest site although its altitude is highest; this is because in winter the vertical temperature distribution shows inversion. Borehole temperature at 10 m depth at lower elevation about 9 km from Aboa was -17°C (*Sinisalo et al.,* 2003a). A cold layer was found from two sites roughly 5 m deep displaying the propagation of the cold wave from the previous winter.

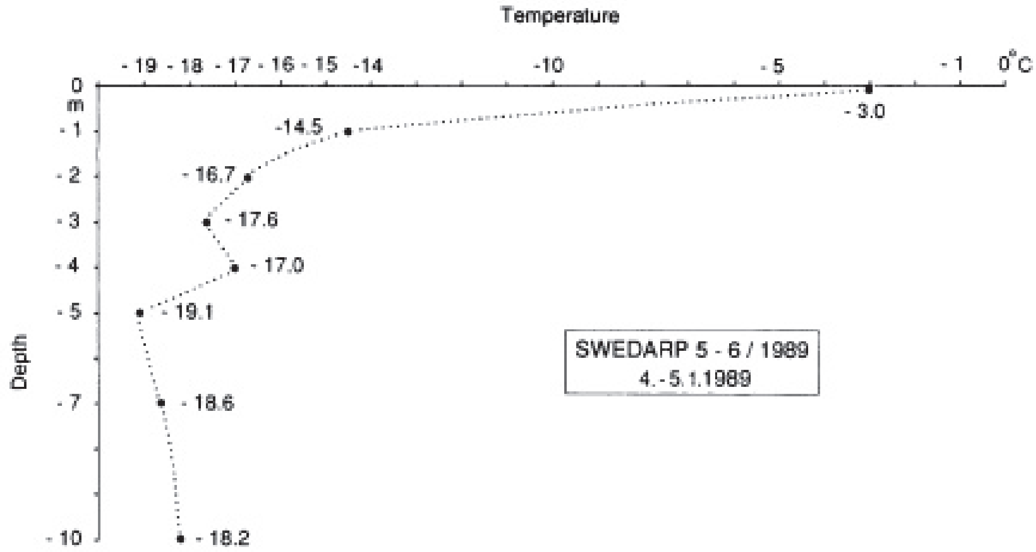


Fig. 8. Temperature distribution obtained from a 10 m borehole in Riiser-Larsen ice shelf, 70 m above the sea level on January 4–5, 1989 (*Seppälä*, 1992), indicating an annual mean temperature of about -18°C at the drilling site.

Snow temperature profiles were also measured from the 1-m and 50-cm snow pits in 2003–2004 and 2004–2005 expeditions (*Kärkäs et al.*, 2002; *Kärkäs et al.*, 2005b) and later by *Vihma et al.* (2011). The diurnal temperature cycle penetrated down to 30 cm. Below 30 cm the mean temperature gradient was –0.05°C m^{-1}, whereas the diurnal temperature cycle at the surface lead to stronger temperature gradients closer to surface to drive faster snow metamorphism.

An experiment was conducted to monitor the surface layer temperature regime and the accumulation rate through a set of nine thermistor strings deployed in the research area for one year of measurements during 2000 (*Granberg et al.*, 2009). Although there were some problems with the instrumentation due to the harsh conditions, the data suggest that cooling of the surface starts gradually from south ice and that correlation found with surface slope would infer that sublimation of snow in the katabatic wind regime is an important factor in the process. Automatic snow stations were repeatedly used in 2010 (*Järvinen et al.*, 2013). The power spectra revealed two distinct temperature cycles in the snowpack at the depth of 54 cm: one-day cycle, and approximately a ten-day cycle (synoptic timescale).

5.2 Radiation balance and turbulent fluxes

The most obvious climatic influence of the snow cover in the Antarctic continent is due to its high reflectance of solar radiation. Small changes in albedo have large effects on the radiation balance and consequently on the surface heat budget. At the level of 90%, reduction of albedo by 10% units (increase by 5% units) will double (half) the surface layer absorption of solar radiation.

Spectral reflectance and attenuation of solar radiation inside the snow cover and the supraglacial lakes have been examined at Aboa station and close to South African station SANAE4 (*Rasmus,* 2001; *Rasmus et al.*, 2003; *Leppäranta et al.,* 2012; *Järvinen et al.*, 2013). The measurements showed that reflectance and attenuation depend on solar altitude, cloudiness and snow grain size, with larger grains producing larger attenuation coefficient (Fig. 9). The albedo model by *Wiscombe and Warren* (1980) was used by *Rasmus* (2009) to calculate the spectral albedo.

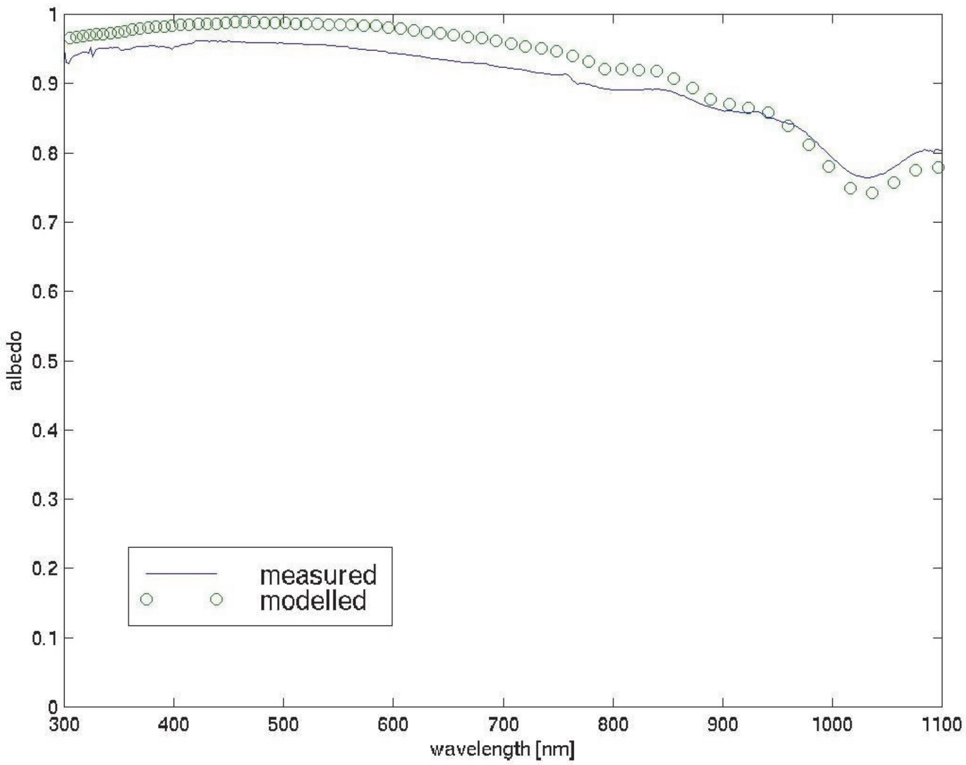


Fig. 9. Albedo measured at Aboa and according to model outcome (*Rasmus*, 2009). The model by *Wiscombe and Warren* (1980) based on Mie scattering was used.

Albedo has also been studied in other areas of Antarctica (*Pirazzini*, 2004, 2009) in order to enhance the albedo parameterization in weather and climate models. The results have been included into the ECMWF global weather forecast and climate model and produced improvements in the outcome. Depending on the site, the effect of snow properties or cloud cover dominated the seasonal albedo variability (*Pirazzini*, 2004). Snow metamorphosis is difficult to describe in climate models. Yet, more in-depth research on the snow properties, especially on the grain size and grain geometry as well as the macroscopic surface geometry is required. *Järvinen et al.* (2013) reported that the estimated average depth, at which the irradiance was 1% of the downwelling irradiance at the surface, was 50 cm.

6 *Remote sensing of snow*

A natural approach for areal snow observations in the DML is satellite remote sensing. Remote sensing signals over snowfields depend primarily on – and can provide information on – the surface roughness, grain size and liquid water content. Apparent optical properties, such as albedo, also depend on the illumination conditions. There are also indications of correlation between SAR backscatter and surface mass balance. Although remote sensing is a tempting method for large observation areas of difficult access, major problems are present in the interpretation of the imagery. This needs evidence from the ground to calibrate the system. FINNARP snow remote sensing has involved satellite optical and radar methods and a helicopter-borne survey of reflectance in the optical and near infrared bands.

Potentials of optical and near-infrared remote sensing for snow mapping were examined in FINNARP 1999 expedition (*Rasmus et al*., 2003). Nadir radiance was measured from a helicopter, and flight tracks covered a 400-km long section from the shelf edge to the station Svea. A weak dependency of reflectance with snow grain size was found but in general the snow cover was quite homogeneous leaving not much variability to detect. Instead, reflectance depended primarily on the cloudiness since that is the primary factor to modify the illumination conditions. *Laine* (2008) examined albedo in 1981–2000 over the Antarctic ice-sheet and sea-ice surface. The ice sheet work was based on Advanced Very High Resolution Polar Pathfinder data. He could recognize slightly positive spring–summer albedo trends.

Summer melting was monitored using ASAR (Advanced Synthetic Aperture Radar) instrument onboard ENVISAT (Environmental Satellite) orbiter of the European Space Agency (ESA) (*Mattila*, 2007). Surface melting was clearly observed throughout the ice shelf and also over the lower parts of the coastal margin (Fig. 10). Attempts were also made to infer the free water content of the surface snow using the SAR-backscattering model of *Koskinen et al.* (2000). The analysis showed that the SAR signal quickly saturates after the initial stages of surface melting, and therefore the free water content cannot be solely inferred from SAR. As the snow conditions change from the ice shelf toward the interior ice sheet, the SAR backscattering signal shows different temporal evolution.

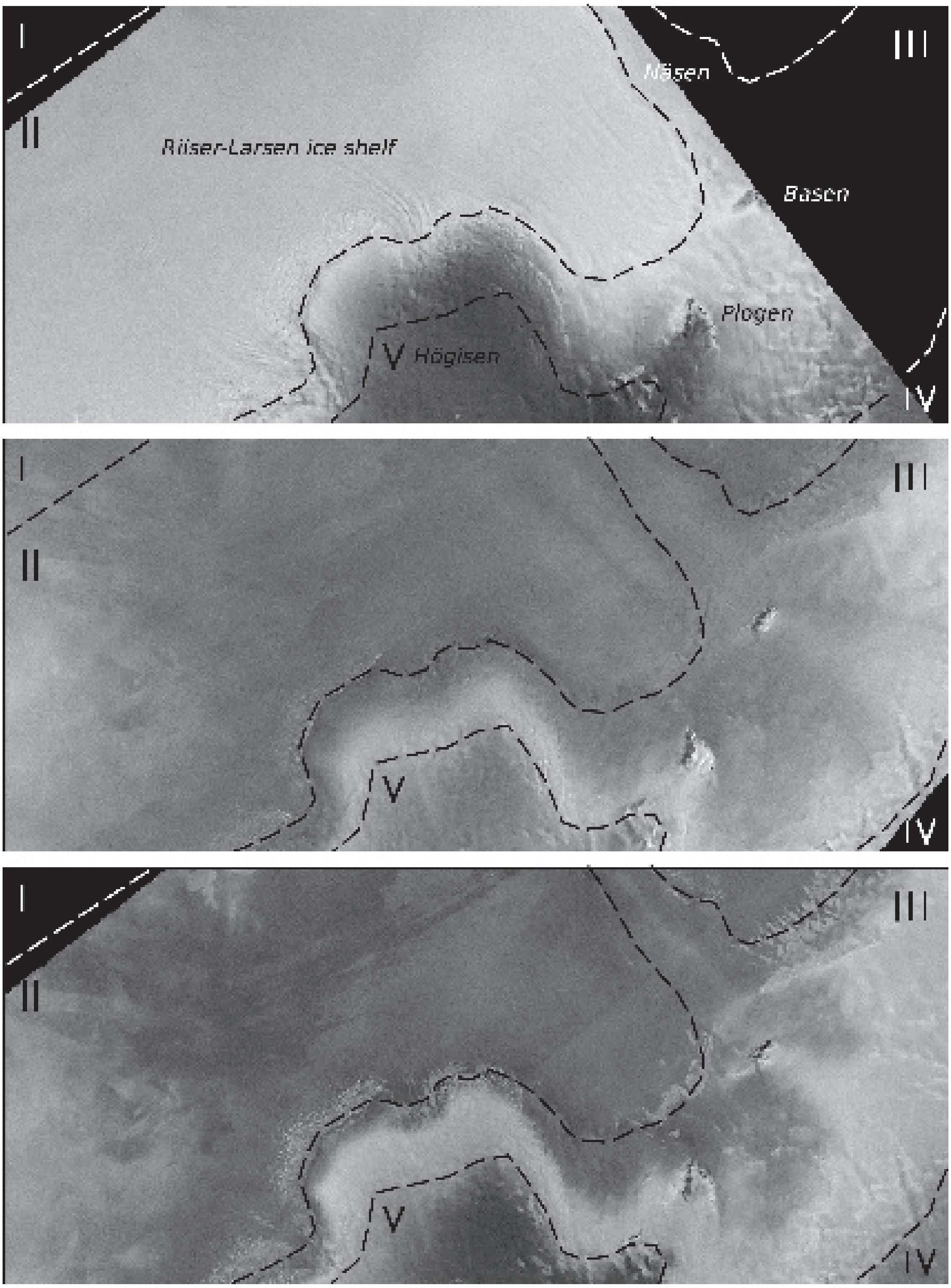


Fig. 10. Change the microwave backscattering signal due to increasing wetness of the snow surface due to the solar radiation over the summer season monitored by ENVISAT/ASAR image series (*Mattila*, 2007). @ European Space Agency.

7 *Final remarks*

Snow research has been carried out in the FINNARP programme in the western Dronning Maud Land, based in the research station Aboa, since 1988. In 1999 snow research expanded to cover the spatial variability of physical and chemical properties of the snow in the surface layer (0–10 m) from the shelf edge up to 400 km towards south. The research topics have been the mass and heat balance of snow, solar radiation budget, remote sensing, stratigraphy of snow cover, and physical and chemical properties of snow.

The snow mass budget consists of precipitation and sublimation. The net gain is positive but sublimation is a significant fraction of precipitation. Heat balance shows a stable annual cycle in the active surface layer. Snow patches in the Basen nunatak have been also studied for their mass and heat balance. The regional and temporal variability of snow properties have been measured and mapped in summer expeditions. Automatic snow stations have been deployed for the winter season. There are reasons to believe that the large-scale spatial variability is well represented in the dataset, but a question remains for the scale dependency for observations of different snow properties. For remote sensing the variability of snow properties within the spatial resolution of the instrument is crucial and needs further investigations. Also, understanding of the small-scale variability would be important for modelling the snow cover thermodynamics and for coupling the surface snow layer with the atmospheric boundary layer.

There are number of processes in the snow pack that are still inadequately understood. Such are penetration and re-freezing of meltwater and related processes of distribution of chemical components inside the snow, especially over the ice shelves. Also in the first eleven years of the life history of Aboa regular research was performed at the station, and it would be a challenge to try to expand the snow database to the summers before 2000. The next main aims are to expand ground observations through remote sensing and mathematical modelling for understanding the annual cycle of snow cover in the Dronning Maud Land. Presently there is no ongoing snow-dedicated research project at Aboa, but for coming new efforts the database collected in 1999–2011 is extensive and provides an excellent reference to understand possible future changes in the regional characteristics of the snow cover.

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References

Colbeck, S., 1991. Layered character of snow covers. *Reviews in Geophysics*, **29**, 81–96.

Colbeck, S., E. Akitaya, R. Armstrong, H. Gubler, J. Lafeuille, K. Lied, D. McClung and E. Morris, 1990. The international classification for seasonal snow on the ground. International Commission on Snow and Ice, Wallingford.

Doake, C.S.M., D.A. Hodgson, D.E. Sugden, S.J. Roberts, J.A. Smith, M.J. Leng and C. Bryant, 2005. Breakup and conditions for stability of the northern Larsen Ice Shelf, Antarctica. *Nature*, **391**, 778–780.

Granberg, H.B. and M. Leppäranta, 1999. Observations of sea ice ridging in the Weddell Sea. *Journal of Geophysical Research*, **104**(C11), 25,735–25,745.

Granberg, H.B., P. Cliche, O.-P. Mattila, E. Kanto and M. Leppäranta, 2009. A snow sensor experiment in Dronning Maud Land. *Journal of Glaciology*, **55**(194), 1041–1051.

Järvinen, O., M. Leppäranta and J. Vehviläinen, 2012. The surface layer of the ice sheet in Dronning Maud Land. Data report. *Report Series in Geophysics* **71**. Department of Physics, University of Helsinki. Helsinki, Finland.

Järvinen, O., 2013. Annual cycle of the active surface layer in western Dronning Maud Land, Antarctica. *Report Series in Geophysics* **74**. PhD thesis (geophysics). Department of Physics, University of Helsinki. Helsinki, Finland.

Järvinen, O. and M. Leppäranta, 2013. Solar radiation transfer in the surface snow layer in Dronning Maud Land, Antarctica, *Polar Science*, **7**(1), 1–17.

Järvinen, O., M. Leppäranta and J. Vehviläinen, 2013. One-year records from automatic snow stations in western Dronning Maud Land, Antarctica. *Antarctic Science*, **25**(5), 711–728.

Kanto, E., 2006. Snow characteristics in Dronning Maud Land, Antarctica. *Report Series in Geophysics* **49**. PhD thesis (geophysics). Department of Physics, University of Helsinki. Helsinki, Finland.

Kanto, E., M. Leppäranta and O.-P. Mattila, 2007. Seasonal snow in Antarctica. Data report II. *Report Series in Geophysics* **55**. Department of Physics, University of Helsinki. Helsinki, Finland.

Kärkäs, E. 2000. Kesäaikaisen lumipeitteen fysikaaliset ominaisuudet Kuningatar Maudin maalla, Etelämantereella [Physical properties of snow cover in summer in Dronning Maud Land, Antarctica]. MSc thesis, 92 p. Department of Geophysics, University of Helsinki. Helsinki, Finland.

Kärkäs, E., 2004. Meteorological conditions of the Basen nunatak in western Dronning Maud Land, Antarctica, during years 1989–2001. *Geophysica*, **40** (1–2), 39–52.

Kärkäs, E., H.B. Granberg, C. Lavoie, K. Kanto, K. Rasmus and M. Leppäranta, 2002. Physical properties of the seasonal snow cover in Dronning Maud Land, East-Antarctica. *Annals of Glaciology*, **34**, 89–94.

Kärkäs, E., K. Teinilä, A. Virkkula and M. Aurela, 2005a. Spatial variations of surface snow chemistry during two austral summers in western Dronning Maud Land, Antarctica. *Atmospheric Environment*, **39,** 1405–1416.

Kärkäs, E., T. Martma and E. Sonninen, 2005b. Surface snow properties and stratigarphy during the austral summer in western Dronning Maud Land, Antarctica. *Polar Research* **24**(1–2), 55–67.

Keskitalo, J., M. Leppäranta and L. Arvola, 2013. First records on primary producers of epiglacial and supraglacial lakes in the western Dronning Maud Land, Antarctica. *Polar Biology*, **36**(10), 1441–1550.

Koistinen, E., 2007. Antarktikan ilmastohistoria [Climate history of Antarctica]. *Report Series in Geophysics* **54**. Department of Physics, University of Helsinki. Helsinki, Finland.

Koivula, H. and J. Mäkinen, 2003. Geodetic activities at Finnish Antarctic research station Aboa. *Presented in The Fifth International Antarctic Geodesy Symposium, AGS'03*, Lviv, Ukraine.

Koponen, I.K., A. Virkkula, R. Hillamo, V.-M. Kerminen and M. Kulmala, 2003. Number size distributions and concentrations of the continental summer aerosols in Queen Maud Land, Antarctica. *Journal of Geophysical Research*, **108** (D18), 4587, DOI:10.1029/2003JD003614.

Koskinen, J., J. Pulliainen and M. Hallikainen, 2000. Effect of snow wetness to C-band backscatter-modelling approach. *Technical Report*, **41**. Helsinki University of Technology, Laboratory of Space Technology.

Kyrö, E.-M., V.-M. Kerminen, A. Virkkula, M. Dal Maso, J. Parshintsev, J. Ruíz-Jimenez, L. Forsström, H.E. Manninen, M.-L. Riekkola, P. Heinonen and M. Kulmala, 2013. Antarctic new particle formation from continental biogenic precursors. *Atmospheric Chemistry and Physics*, **13**, 3527–3546.

Laine, V., 2008. Antarctic ice sheet and sea ice regional albedo and temperature change, 1981–2000, from AVHRR Polar Pathfinder data. *Remote Sensing of the Environment*, **112**, 646–667.

Leppäranta, M., O. Järvinen and E. Lindgren, 2013a. Mass and heat balance of snow patches in Basen nunatak, Dronning Maud Land in summer. *Journal of Glaciology*, **59**(218), 1093–1105.

Leppäranta, M., O. Järvinen and O.-P. Mattila, 2013b. Structure and life cycle of supraglacial lakes in the Dronning Maud Land. *Antarctic Science*, 25(3), 457–467.

Mattila, O.-P., 2007. Summer surface melting in a coastal region of Dronning Maud Land, Antarctica, detected from ENVISAT ASAR imagery. MSc thesis (geophysics). Department of Physics, University of Helsinki. Helsinki, Finland.

Mäkinen, J., M. Almavict, K. Shibuya and Y. Fukuda, 2007. Absolute gravimetry in Antarctica: Status and prospects. *Journal of Geodynamics*, **43**, 339–357.

Pirazzini, R., 2004. Surface albedo measurements over Antarctic sites in summer. *Journal of Geophysical Research*, **109** (D20118), 1–15.

Pirazzini, R., 2008. Factors controlling the surface energy balance over snow and ice. *Finnish Meteorological Institute, Contributions* **75**, PhD thesis (meteorology). Department of Physics, University of Helsinki. Helsinki, Finland.

Pirazzini, R., 2009. Challenges in snow and ice albedo parameterizations. *Geophysica*, **45**, 41–62.

Rasmus, K., 2001. The albedo of Antarctic snow. *Report Series in Geophysics,* **46**. *Proceedings of the Symposium on Sea Ice Climate and Marine Environments in the Okhotsk and Baltic Seas—The Present Status and Prospects*, Archipelago Research Institute, Seili, Finland, 10–13 September 2001.

Rasmus, K., 2009. Optical studies of the Antarctic glacio-oceanic system. *Report Series in Geophysics* **62**. PhD thesis (geophysics). Department of Physics, University of Helsinki. Helsinki, Finland.

Rasmus, K., H.B. Granberg, K. Kanto, E. Kärkäs, C. Lavoie and M. Leppäranta, 2003. Seasonal snow in Antarctica. Data report. *Report Series in Geophysics* **47**. Department of Physics, University of Helsinki. Helsinki, Finland.

Reijmer, C. H., 2001. Antarctic meteorology - A study with automated weather stations. *PhD thesis*, Utrecht University. Utrecht, The Netherlands.

Rott, H., P. Skvarca and T. Nagler, 1996. Rapid collapse of Northern Larsen Ice Shelf, Antarctica. *Science*, **271** (5250), 788–792.

Ruotoistenmäki, T. and J. Lehtimäki, 2009. Geophysical and geodetic studies of bedrock, permafrost and continental ice in Queen Maud Land, Antarctica. *Geophysica*, **45**(1-2), 63–76.

Scambos, T.A., J.A. Bohlander, C.A. Shuman and P. Skvarca, 2004. Glacier acceleration and thinning after ice shelf collapse in the Larsen B embayment, Antarctica. *Geophysical Research Letters*, **31** (L18402), 4.

Schytt, V., 1958. The inner structure of the ice shelf at Maudheim as shown by core drilling. *Norwegian-British-Swedish Antarctic Expedition, 1949–1952. Scientific Results* **4**, *Glaciology* 2, 115–151. Norwegian Polar Institute, Oslo.

Sihvola, A. and M. Tiuri, 1986. Snow fork for field determination of the density and wetness profiles of a snow pack. *IEEE Transactions in Geoscience and Remote Sens*ing, **24**(5), 717–721. doi: 10.1109/TGRS.1986.289619

Seppälä, M., 1992. Stabilization of snow temperature in Dronning Maud Land, Antarctica. *Geografiska Annaler A*, **74**(2/3), 227–230.

Sinisalo, A. and J. Moore, 2010. Antarctic blue ice areas (BIAs) –towards paleoclimate interpretation. *Antarctic Science*, **22**(2), 99–115, doi:10.1017/S0954102009990691.

Sinisalo, A., A. Grinsted, J.C. Moore, E. Kärkäs and R. Pettersson, 2003a. Snow-accumulation studies in Antarctica with ground-penetrating radar using 50, 100 and 800 MHz antenna frequencies. *Annals of Glaciology*, **37**, 194–198.

Sinisalo, A., J.C. Moore, R.S.W. van de Wal, R. Bintanja and S. Johnsson, 2003b. A 14 year mass-balance record of a blue-ice area in Antarctica. *Annals of Glaciology*, **37,** 213–218.

Sinisalo, A, A. Grinsted and J.C. Moore, 2004. Dynamics of the Scharffenbergbotnen blue-ice area, Dronning Maud Land, Antarctica. *Annals of Glaciology*, **39**, 417–423.

Tietäväinen, H. and T. Vihma, 2008. Atmospheric moisture budget over Antarctica and Southern Ocean on the basis of ERA-40 reanalysis. *International Journal of Climatology*, **28**, 1977–1995, doi: 10.1002/joc.1684.

Van den Broeke, M., 2005. Strong surface melting preceded collapse of Antarctic Peninsula ice shelf. *Geophysical Research Letters* **32** (L12815), 4.

Vihma, T., J. Launiainen and R. Pirazzini, 2009. 20-years of Finnish research on boundary-layer meteorology and air-ice-sea interaction in the Antarctic. *Geophysica*, 45(1-2), 7–26.

Vihma, T., O.-P. Mattila, R. Pirazzini and M.M. Johansson, 2011. Spatial and temporal variability in summer snow pack in Dronning Maud Land, Antarctica. *The Cryosphere*, 5, 187–201, doi:10.5194/tc-5-187-2011

Virkkula, A., I.K. Koponen, K. Teinilä, R. Hillamo, V.-M. Kerminen and M. Kulmala, 2006b. Effective real refractive index of dry aerosols in the Antarctic boundary layer. *Geophysical Research Letters*, **33**, No. 6, L06805, 10.1029/2005GL024602

Virkkula, A., E. Asmi, K. Teinilä, A. Frey, M. Aurela, H. Timonen, T. Mäkelä, A. Samuli, R. Hillamo, P.P. Aalto, S. Kirkwood and M. Kulmala, 2009. Review of Aerosol Research at the Finnish Antarctic Research Station Aboa and its Surroundings in Queen Maud Land, Antarctica. *Geophysica*, **45** (1-2), 163–181.

Wiscombe, W.J. and S.G. Warren, 1980. A model for the spectral albedo of snow, I: Pure snow. *Journal of Atmospheric Sciences*, **37**, 2712–2733.