

# Composition and internal structures of a rock glacier on the strandflat of western Spitsbergen, Svalbard

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This article presents data on composition and internal structures of a large lobe-shaped rock glacier at Nordenskiöldkysten, western Spitsbergen, based on Ground Penetrating Radar (GPR) and direct current (DC) resistivity tomography measurements. This rock glacier has a more complex surface topography than most rock glaciers in Svalbard and is probably also older. The GPR data indicate a quite chaotic structural pattern, far more complicated than for other investigated rock glaciers in Svalbard. This pattern might be due to the anticipated complex history of the rock glacier, involving glaciation and relative sea level changes. The geophysical results indicate ice-rich conditions, especially below ridges. Overall, the resistivity values are decreasing towards the front of the rock glacier, probably due to dynamic effects. Also, the resistivity results indicate that the active layer is thicker in dumps than below ridges. Considerations of ice content combined with previous velocity measurements indicate that the apparent standstill of the rock glacier is due to the low-gradient situations in which the rock glacier flows onto the strandflat.

Keywords: DC resistivity, GPR, ice content, internal structure, rock glacier, Svalbard

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## Introduction

Rock glaciers are lobate or tongue-shaped creeping permafrost bodies flowing due to deformation of the ice within (Barsch 1992, 1996, Haeberli 2000). This ice may have multiple origins, including freezing of sub- and suprapermafrost groundwater, as well as burial of snowdrifts (Haeberli & Vonder Mühll 1996, Berthling et al. 1998, Isaksen et al. 2000). Rock glaciers are the cumulative expression of their entire history, reflecting their present and past environments (Frauenfelder & Kääb 2000). They have total ages of several thousands years (Barsch 1996, Haeberli et al. 1999), and thus have a great palaeoclimatic potential.

Lobate rock glaciers are quite common features along the western and northern parts of the Svalbard archipelago, amounting to several hundreds (Kristiansen & Sollid 1986, Sollid & Sørbel 1992). This is probably due to a combination of favourable high relief settings, often associated with the inner parts of the low-gradient strandflat (Holtedahl 1998) and early ice-free conditions, and thus the rock glaciers have had a long period of time for their formation (Sollid & Sørbel 1992, 2003). Tongue-shaped rock glaciers also occur in Svalbard. These are situated on mountain slopes, but they are not too common (Isaksen 2001). The rock glaciers in Svalbard were first described by Liestøl (1962) as ‘terrace-like talus formations’. He questioned a glacial origin and believed the rock glaciers contained talus deposits and ice originating from debris-covered snow. In the last decade the composition and internal structure of some rock glaciers in Svalbard have been investigated using geophysical methods such as Ground Penetrating Radar (Berthling et al. 2000, 2003b, Isaksen et al. 2000), DC resistivity soundings (Hoelzle 1992, Wagner 1996, Berthling et al. 1998, Isaksen

et al. 2000), seismic refraction soundings (Wagner 1996), and gravimetrical measurements (Vonder Mühll 1995). These methods have been validated in the Alps using borehole information (Vonder Mühll et al. 2000), but such evidence is unfortunately not available from rock glaciers in Svalbard so far.

Most lobate rock glaciers in Svalbard, such as on Prins Karls Forland (Berthling et al. 1998, 2000), have a quite simple surface topography and usually a depression behind their front. At Nordenskiöldkysten, western Spitsbergen (Fig. 1), some considerably larger and more complex lobe-shaped rock glaciers occur. These spread far out into the strandflat showing a pronounced sequence of transverse ridges, furrows and irregular forms (Eiken et al. 2001, Isaksen 2001, Kääb et al. 2002).

A large and complex lobe-shaped rock glacier at Nordenskiöldkysten is investigated in this study (Fig. 2). The rock glacier’s surface displacements have previously been investigated using digital photogrammetry on aerial photos from 1969 and 1990 (Eiken et al. 2001, Kääb et al. 2002). Although no significant velocities (above  $c.1 \text{ cm a}^{-1}$ ) were found, except in the southernmost talus slope where surface velocities between  $4\text{--}8 \text{ cm a}^{-1}$  were detected, the rock glacier is probably not dynamically inactive, but its surface velocities are below a significant level (Eiken et al. 2001, Kääb et al. 2002). This is based on the observations that most of the front has a sharp upper edge and a steep front slope of  $c.40^\circ$  (Chandler 1973), with less vegetation and lichen cover than the rest of the surface. Also, the sequential changes of the longitudinal ridge-and-furrow structures indicate a continuous, slow evolution. More important, the outermost part of the rock glacier exceeds the Late Weichselian Marine Limit (dated 11 ka B.P. by Landvik et al. (1987)) by

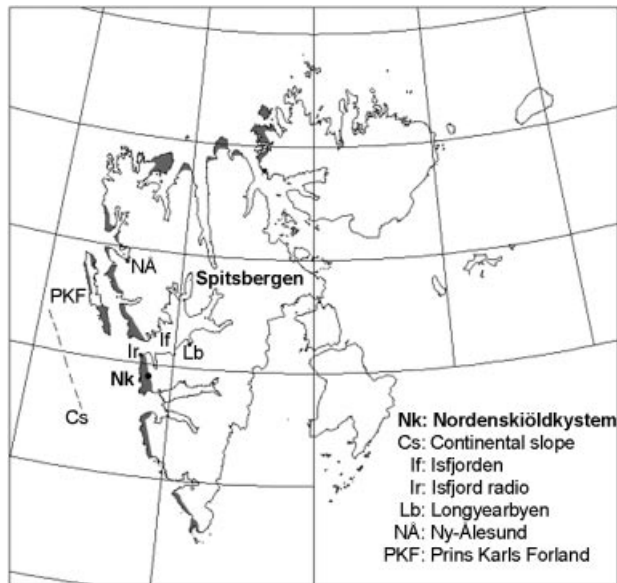


Fig. 1. Location of the study site in Svalbard marked with a dot. The grey areas broadly indicate occurrences of strandflat (modified after Sollid & Sørbel 2003). The rock glaciers in Svalbard referred to in the text are located near Longyearbyen, Ny-Ålesund, and on Prins Karls Forland.

c.20–50 m, indicating an advance rate of at least a few  $\text{mm a}^{-1}$  for the last 11 ka if deformation has been continuously slow (Eiken et al. 2001, Kääb et al. 2002). If advance rates have been this slow during the main parts of its history, the rock glacier is very old. The assumption that present-day velocities are roughly equivalent to the average velocities over the entire evolution of rock glaciers in Svalbard is supported by several authors (Isaksen et al. 2000, Berthling 2001, Isaksen 2001, Kääb et al. 2002). Based on length and an advance rate of  $0.5 \text{ cm a}^{-1}$ , Isaksen (2001) suggests a

total age of the rock glacier of more than 50 ka, given a roughly stable flow rate.

The aim of this study is to investigate the composition and internal structures of this complex rock glacier with the aid of GPR and DC resistivity tomography measurements and to compare the results with previous results, particularly from simpler lobe-shaped glaciers (Berthling et al. 1998, 2000) and tongue-shaped rock glaciers (Isaksen et al. 2000, Ødegård et al. 2003a) in Svalbard.

## Study site

The field site is located at the 40 km long Nordenskiöldkysten strandflat, south of Isfjord Radio, on the western part of Spitsbergen ( $77^{\circ}53' \text{N } 13^{\circ}54' \text{E}$ ) (Fig. 1). The strandflat is 12 km wide on land and continues several kilometres to the west offshore. A pattern of three well-developed raised marine shorelines occurs on the strandflat (Landvik et al. 1987), and ice wedge polygons are common.

Facing west, the rock glacier is found below the escarpment, which delimits the inner part of the strandflat. It is c.1 km wide and 300 m long, with an up to 40 m high and up to  $40^{\circ}$  steep front slope. The average slope above the front is only a few degrees. A visual inspection of the rock glacier indicates a complex of individual creep lobes over-riding older ones, and this is supported by the occurrence of embryonal rock glaciers higher up in the talus. Several smaller lobe-shaped rock glaciers with simpler morphology occur in the bottom of the nearby valley Orustdalen (Fig. 3), which enters the strandflat a few kilometres to the north of the investigated rock glacier. No intermediate forms between the two types of rock glaciers are found, indicating a marked difference involved in their evolution (Kääb et al. 2002).



Fig. 2. The investigated rock glacier, looking south. Note the complex surface of ridges, furrows and irregular forms. The sea is seen in the background of the picture (Photo: K. Isaksen, August 2000).



Fig. 3. Smaller lobe-shaped rock glacier in the nearby valley Orustdalen. Note the smooth surface topography of this rock glacier (Photo: K. Isaksen, August 2000).

#### *Past and present climate and permafrost conditions*

The meteorological station nearest the field site is Isfjord Radio (7 m a.s.l.) located 18.5 km to the north (Fig. 1). There, Mean Annual Air Temperature (MAAT) is  $-5.1^{\circ}\text{C}$  and Mean Annual Precipitation (MAP) is  $c.480$  mm (Norwegian Meteorological Institute, unpublished data). Cyclonic activity is very common during winter and causes unstable and stormy weather with prevailing winds from the northeast (Hanssen-Bauer et al. 1990). Snow accumulation is therefore mainly restricted to depressions. Measures and estimations of permafrost depths in Svalbard range from several tens of metres close to the coast to some hundreds of metres inland (Liestøl 1976, Isaksen et al. 2001). The area is thus quite probably underlain by permafrost far beneath the base of the rock glacier.

Different approaches (Birks 1991, Salvigsen et al. 1992, Svendsen & Mangerud 1997) indicate that MAAT was  $2\text{--}3^{\circ}\text{C}$  warmer during the early Holocene temperature optimum in western parts of Svalbard (O. Salvigsen, personal communication 2004), yet still sub-zero. Isaksen et al. (2001) showed that Mean Annual Ground Surface Temperature (MAGST) on Janssonhaugen, to the east of Longyearbyen, was  $0.8^{\circ}\text{C}$  warmer than MAAT. This effect is probably by far offset at the field site by the fact that coarse surface debris creates a colder microclimate than the adjacent fine-grained ground (Humlum 1997, Harris & Pedersen 1998). If covered by Late Weichselian ice, permafrost conditions probably still prevailed at the field site because warm-based ice mainly was restricted to the main valleys (Humlum et al. 2003). Palaeotemperature estimates indicate that central Prins Karls Forland (Fig. 1) has not been covered by warm-based ice for the last 70 ka (Andersson et al. 1999). Thus, it seems likely that the field site at Nordenskiöldkysten has experienced permafrost conditions for tens of thousands of years.

#### *Bedrock geology and Quaternary history*

The bedrock at the field site consists of metamorphic rocks of Late Precambrian (Hecla Hoek) age (Hjelle et al. 1986), and the general trend of the folding of the bedrock is N–S. The diameter of rock fragments in the rock glacier surface varies from a few centimetres to several metres in a spatial pattern, showing some zones mainly covered with small blocks and others with coarse material (Kääb et al. 2002). Overall, the block size decreases towards the front. At the front slope an upper coarse-grained layer covers more fine-grained material below.

Glacial erratics indicate that the area has been completely covered by ice during at least one Quaternary glaciation, but the exact age of such an event is not known. The Late Weichselian ice sheet probably reached the continental slope west of Isfjorden (Fig. 1) (Landvik et al. 1998). Considering such a great extent of the ice sheet slightly further north it seems unlikely that Nordenskiöldkysten was not covered, but the timing, extent and configuration of the Late Weichselian Barents ice sheet is uncertain and has been debated for several decades because of the lack of glacial landforms (Landvik et al. 1998). It has been speculated that the mountain range might have protected the field site from the ice flow, which followed the main valleys (Eiken et al. 2001). Considering the west coast of Spitsbergen, Andersson et al. (1999) conclude that during Late Weichselian the Barents Sea ice sheet drained through the main fjords, possibly leaving relatively large areas ice-free. If so, this might explain the distribution of the different rock glaciers in the area in which the smaller lobe-shaped rock glaciers with simpler morphology, located in the bottom of the valley, have formed after deglaciation, while the large complex rock glaciers have formed during much longer periods of time (Eiken et al. 2001). If ice-free during Late Weichselian, these

rock glaciers are important palaeoenvironmental indicators. However, they might also have been preserved by cold-based ice (Sollid & Sørbel 1992).

## Methods

### *Direct current (DC) resistivity tomography*

The purpose of electrical surveys is to determine the subsurface resistivity distribution by taking measurements on the ground surface. From these measurements, the true resistivity of the ground can be estimated. The existence of ice in the ground increases the resistivity markedly as the resistivity of ice is 4–5 orders of magnitude higher than water (Hauck 2001). However, ice of different origin reveal resistivities that change over several orders of magnitude (Haerberli & Vonder Mühl 1996). Thus, differences in resistivities on a rock glacier could be explained by differences in ice content as well as differences in the origin of the ice.

An ABEM Lund (Sweden) high-resolution two-dimensional resistivity system was used in this survey. This system consisted of 61 electrodes with a standard spacing of 2 m. A Wenner Alpha array was used, giving maximum penetration depth of roughly 26 m. Interpretations of the profiles and inversion of the apparent to true resistivity values was carried out using the computer programme RES2DINV. This programme subdivides the subsurface into a number of blocks and then uses a least-squares inversion scheme to determine appropriate resistivity value for each block (Loke & Barker 1996).

### *Ground Penetrating Radar (GPR)*

The GPR method utilizes high frequent electromagnetic (EM) pulses that are transmitted into the ground. When the signal reaches an object or a borderline between two materials of different relative dielectric constant, parts of the signal are reflected back to a receiver. The GPR system measures the traveltime from the transmitter to the object back to the receiver as well as the amplitude of the reflected signal. Knowing the velocity of the EM signal, the two-way traveltime (m/ns) can be converted to the depth (m) of the object.

In this survey, a RAMAC GPR system from Malå GeoScience was used. Antennas of 50 MHz centre frequency with a separation distance of 1 m were applied as these have given good results earlier (Berthling et al. 2000, 2003b, Isaksen et al. 2000). The depth conversion of the radar signals was estimated using a mean velocity of the EM energy of 140 m/μs. This value originates from velocity estimations from point scatter and agrees well with a CMP survey by Berthling et al. (2003b) on another rock glacier in Svalbard.

The profiles were processed using the software REFLEXW (Sandmeier scientific software, Karlsruhe, Germany). Because of the rapid attenuation of the radar signals with depth, an energy decay filter was applied. This filter applies a gaincurve in y-(time) direction on the

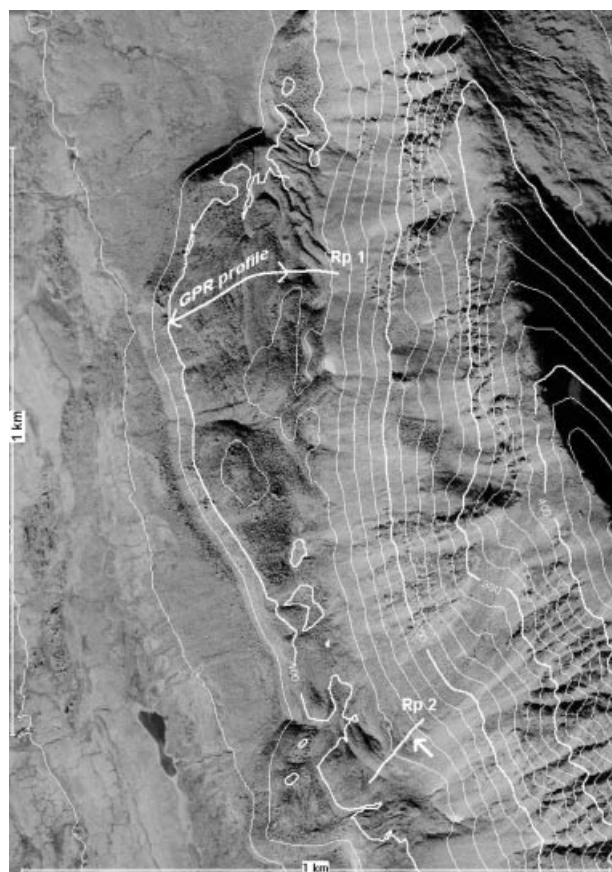


Fig. 4. Ortho-photo of the investigated rock glacier (contour intervals 25 m), showing the location of the resistivity tomography profiles (Rp 1 and Rp 2). The GPR profile covers the lower 225 m of Rp1. The arrow indicates the area where displacements have been found (Kääb et al. 2002). Note the marked zonation of grain sizes. Based on aerial photograph S90-6295 © Norwegian Polar Institute.

complete profile based on a mean amplitude decay curve which is automatically determined. The presented radar data has not been migrated. Thus, the reflectors probably seem steeper than they actually are. This effect is not regarded as crucial since only a few continuous reflectors occur on the radargram. Also, migration often introduces artificial incidents, so the quality of the radargram might deteriorate (Mauring et al. 1995).

## Results

Originally, eight resistivity tomography profiles and seven GPR profiles were obtained at the rock glacier (Farbrot 2002). However, only two longitudinal resistivity profiles and one longitudinal GPR profile are presented in this article (Fig. 4). The left-out profiles are in accordance with those presented, but reveal limited additional information. Resistivity profile 1 (Rp 1) is a longitudinal transect covering the northern part of the rock glacier (Fig. 5), whereas resistivity profile 2 (Rp 2) is a shorter longitudinal profile at the southern part (Fig. 6). The GPR survey presented is covering the lower 225 m of Rp 1.

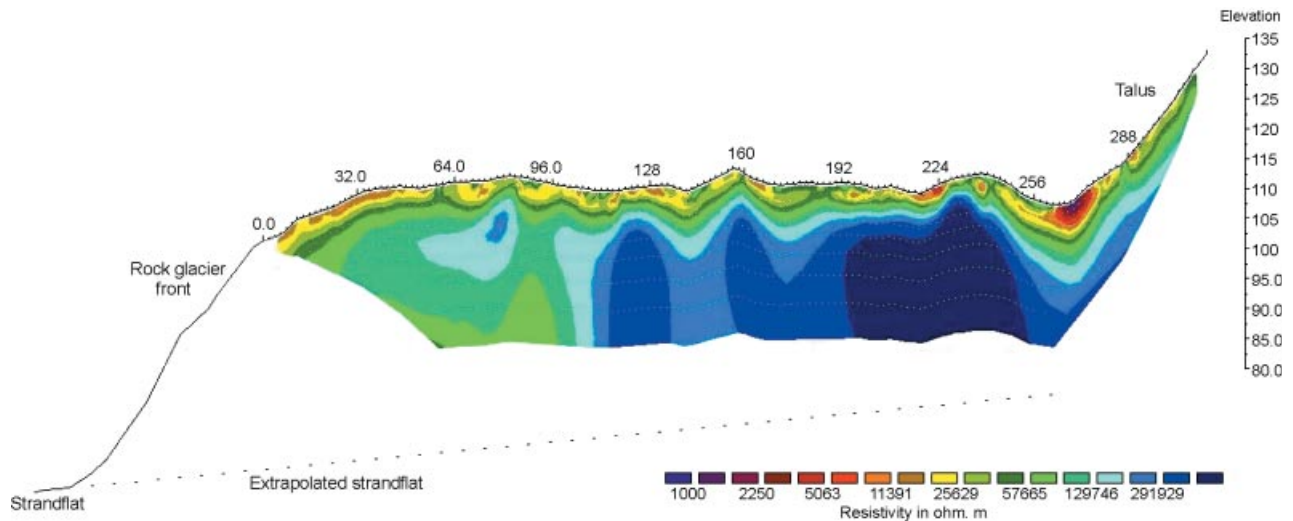


Fig. 5. Rp 1 covering the longitudinal transect of the northern part of the rock glacier (Fig. 4) with the overridden strandflat underneath indicated. High-resistive areas are associated with ridges, and the overall resistivity values decrease towards the terminus. The low-resistive area near the talus is due to moist conditions. By marking the transition between low and high resistivities, an approximate relative distribution of the active layer thickness is found (Fig. 7). Maximum penetration depth of the resistivity system is *c.* 26 m, so the bottom of the debris body is probably not encountered.

#### DC resistivity tomography

The profiles of the rock glacier (Figs 5 and 6) reveal resistivity values of some  $k\Omega m$  to a few tens  $k\Omega m$  in the near-surface part. The exceptions are some low-resistive patches due to moist conditions. Further down in the subsurface of the rock glacier the profiles reveal resistivity values of some tens to a few thousands  $k\Omega m$ .

Some interesting results are pronounced on Rp 1 (Fig. 5). First, it seems clear that the overall resistivity values

decrease from the back of the rock glacier to the front. The resistivity values of the lower part of the talus cone are lower than the values of the rock glacier. Second, some high-resistive areas with resistivities up to a few 1000  $k\Omega m$  occur. When considering the topography of the profile, it seems as though these areas are associated with ridges. Also, the low-resistive colours of the colour scheme (lilac–start of the green scale) are deeper beneath dumps indicating a thicker active layer. This is especially pronounced in the depression near the talus associated with a small puddle, but

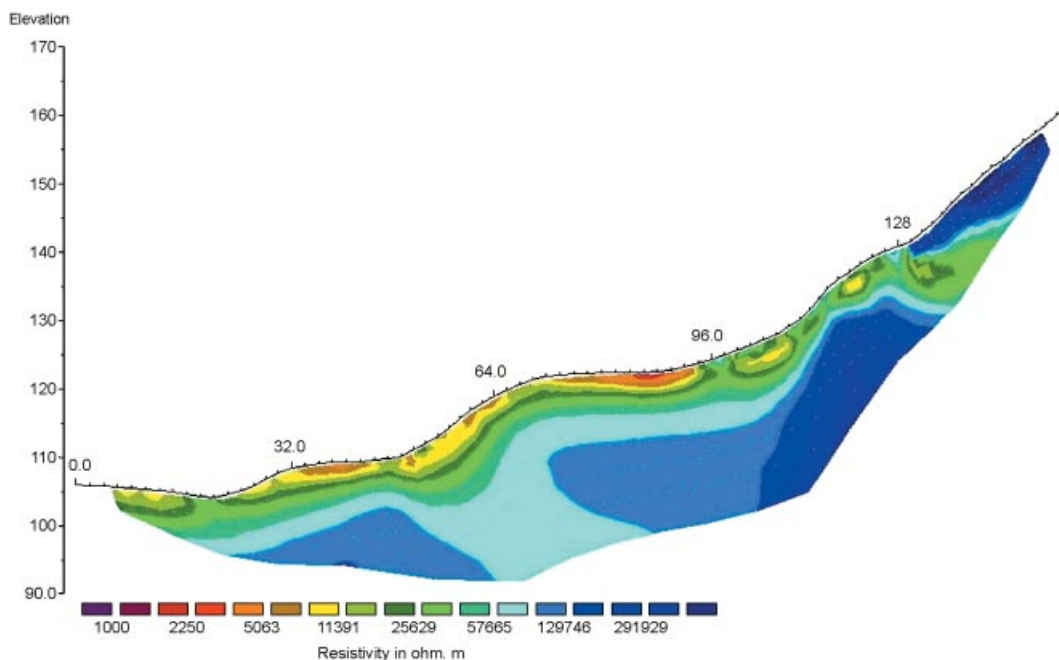


Fig. 6. Rp 2 covering a longitudinal transect of the southern part of the rock glacier. Note the high-resistive area in the upper part consistent with the only part of the rock glacier where displacements ( $4\text{--}8\text{ cm a}^{-1}$ ) have been found (Kääb et al. 2002). Elevation in metres (vertical axis).

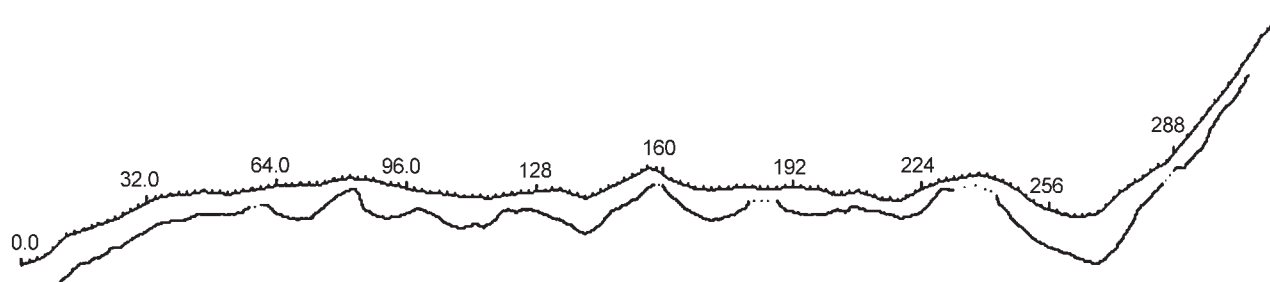


Fig. 7. Approximate relative variation in the thickness of the active layer along Rp 1, found by marking the transition between two colours in the transition between low and high resistivities in Fig. 5 (values expressed in metres).

dry dumps reveal apparently deeper active layers as well. As no borehole temperature data are present, no approximate resistivity values corresponding to the permafrost table are determined. However, by marking the transition between two appropriate colours in the transition between low and high resistivities, an approximate relative thickness distribution of the active layer is found (Fig. 7).

Compared with Rp 1, the southern part of the rock glacier (Rp 2) reveals lower resistivity values (Fig. 6). A striking feature is the high-resistive patch in the lower part of the talus cone in Rp 2. This area is consistent with the only part of the rock glacier where displacements of a few  $\text{cm a}^{-1}$  were found (Kääb et al. 2002). Above, a small embryonal rock glacier was observed.

#### GPR measurements

Generally, the radargram (Fig. 8) reveals information down to a depth of *c.* 25 m. The upper part reveals reflectors from the active layer. This layer appears too thick because the EM velocity of the active layer is far lower than the permafrost (Berthling et al. 2003b). A zone of no or only vague reflectors is found below the active layer. This layer disappears toward the front of the rock glacier. Further down, a zone of scattered reflectors is found. These are wavy and dipping in an apparently chaotic pattern. Also, a few continuous reflectors occur. Some vertical structures are due to noise and do not represent a subsurface boundary in this way.

## Discussion

### *Ice content of the rock glacier*

The investigated rock glacier reveals an apparent overall vertical pattern of two or three layers. The upper layer (a few metres thick) has low resistivities (mainly some tens  $\text{k}\Omega\text{m}$ ) associated with unfrozen medium- to large-sized blocks. Lower values due to moist conditions also occur, but high resistive areas are missing. This indicates a fairly closed matrix as air-filled cavities reveal high resistivities (Hauck & Vonder Mühll 2003). The second layer ( $>15$  m) reveals high resistivities of the order of some tens to a few thousand  $\text{k}\Omega\text{m}$ . These resistivity values are typical for rock glacier permafrost (King et al. 1987, Haeberli & Vonder Mühll 1996, Vonder Mühll et al. 2000, Hauck & Vonder Mühll 2003, Hauck et al. 2003), indicating an ice-rich frozen layer. High-resistive areas are probably associated with talus super-saturated with ice (Vonder Mühll et al. 2000, Hauck & Vonder Mühll 2003, Hauck et al. 2003). A layer of lower resistivities, though higher than the uppermost layer, occurs below the ice-rich layer at the outer 100 m of Rp 1. This is interpreted as frozen debris with lower ice content (Vonder Mühll et al. 2000). As the depth of the resistivity profiles is only *c.* 26 m, the third layer might not be encountered along the inner part of Rp 1, but might exist deeper down. If so, this agrees with core analyses from rock glaciers in the Alps in which the layer with highest ice content is found near the surface and the mineral content tends to increase with depth (Haeberli et al. 1998, Arenson et al. 2002). Resistivity

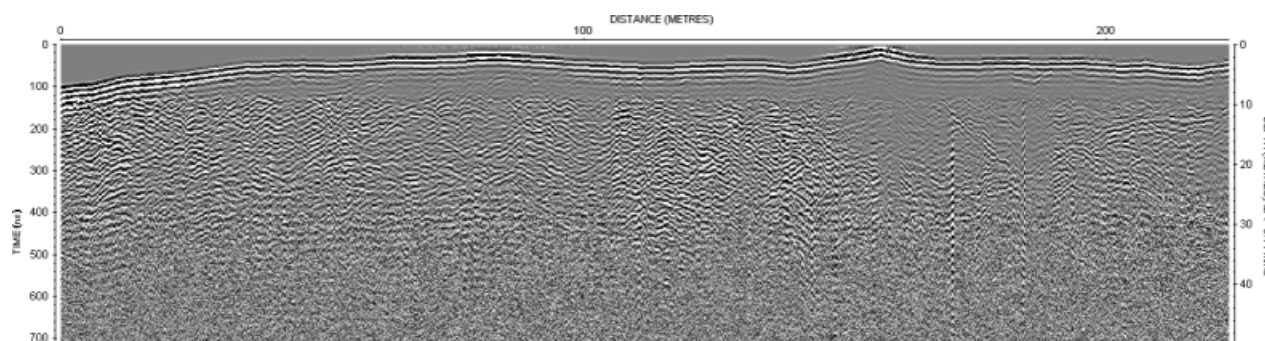


Fig. 8. Radargram covering the lower 225 m of Rp 1, showing a chaotic structural pattern with a few continuous reflectors. Below the active layer, a zone of limited reflectors occur. This is interpreted as a relatively homogenous ice layer and is diminishing towards the front. The marked vertical structures represent noise.

soundings on rock glaciers in Svalbard also seem to support this (Wagner 1996, Berthling et al. 1998, Isaksen et al. 2000). The basal, third layer is believed to originate from surface blocks falling from the front and later overridden by the advancing upper layers (Haeberli et al. 1998, Ødegård et al. 2003a).

### *Internal structures*

With the same GPR set-up, Berthling et al. (2000, 2003b) and Isaksen et al. (2000) revealed clear reflectors on lobe-shaped and tongue-shaped rock glaciers, respectively. The layered structure is believed to develop through burial of snow or ice-supersaturated permafrost by larger mass movement events. The reflectors of this rock glacier are interpreted in the same way. The general coarse blocks of the rock glacier surface indicate that large mass movement events regularly occur, although the possibility of smaller events of large rocks can not be excluded. Liestøl's (1962) observations of alternating layers of debris and dirty ice in an exposure in a rock glacier near Longyearbyen strongly suggests this to be a widespread phenomenon.

The investigated rock glacier reveals a quite chaotic structural pattern compared with GPR surveys revealing systematic systems of structures on tongue-shaped (Isaksen et al. 2000, Ødegård et al. 2003a) and less complex lobe-shaped rock glaciers (Berthling et al. 2000, 2003b) in Svalbard. These rock glaciers reveal quite smooth surfaces compared with the pronounced ridge-and-furrow structure of the investigated rock glacier. The processes involved in the formation of the complex surface topography of the investigated rock glacier may have altered original systematic structures, leading to a present chaotic pattern. Kääb et al. (2002) speculate that influence of glaciation and corresponding relative sea level changes might explain the complex surface of the investigated rock glacier. Also, regarding the assumed considerable age of the feature, it has probably been exposed to large climate variation, affecting geometry and dynamics.

The layer of no or only vague reflectors below the active layer has also been reported from other rock glaciers in Svalbard (Berthling et al. 2000, 2003b). It is emphasized by the gain function applied, but the fact that the layer is disappearing towards the front indicates a true layer as the energy decay filter applies equally to all traces. The limited reflectors imply an area with small contrasts in relative dielectric constant and is interpreted as a relatively homogeneous ice layer (Berthling et al. 2000, 2003b). The GPR results indicate greater depth of this layer below the most pronounced ridge of the profile where also the resistivity measurements indicate ice-rich conditions. Following this interpretation, the ice layer is disappearing towards the front.

### *The connection between ridges and high ice content*

Considering the apparent connection between ridges and high ice content of the subsurface, the formation of the ridges could be the key to this pattern. If the surface topography in some parts of the rock glacier is a result of differential debris production, then the high ice content of the ridges is

quite easily explained. As burial of snowdrifts towards the back of the rock glacier is supposed to be an important source for the ice (Liestøl 1962, Berthling et al. 1998, Isaksen et al. 2000), increased debris production through weathering and mass wasting would lead to increased burial and preservation of snow, thereby increasing the ice content. As noted above, pronounced surface topography is an exception on rock glaciers in Svalbard in contrast to rock glaciers in many other areas. This is likely due to the slowness of these features in Svalbard (maximum  $\approx 10 \text{ cm a}^{-1}$ ) (Sollid & Sørbel 1992, Isaksen et al. 2000, Kääb et al. 2002, Berthling et al. 2003a), thereby restricting the rate of compressive flow. On the other hand, in the Alps, for instance, compressive flow is believed to cause the ridge-and-furrow topography in the lower parts of the faster moving (compared to Svalbard), 'classic-shaped' rock glaciers (Haeberli & Vonder Mühll 1996, Kääb et al. 1998, Kääb & Weber 2004). The location and the probable high age of the investigated rock glacier indicate that the flow regime could have been affected by glaciation, relative sea level change, as well as climate variations (Kääb et al. 2002). However, under such conditions, debris production would probably change as well. The complex surface topography is probably due to more than just one process, but it can not be ruled out that differential debris production is at least partly responsible for it.

In the upper part of the permafrost, the higher ice content of the ridges could also be explained by downward migration of water into the frozen soil. At temperatures below  $0^\circ\text{C}$  thermal gradients control the hydrological regime of fine-grained frozen ground. Gravitational and hydrostatic forces, in comparison, are generally negligible (Williams & Smith 1989). During winter the snow thickness is greater in dumps than in ridges, leading to more intense cooling of the ridges. This would lead to greater temperature gradients during the melting season in ridges compared to dumps as indicated by the difference in active layer thickness (see section on distribution of the active layer thickness). If water percolates through the coarse upper layer, the water will refreeze in the anticipated closed matrix further down. From there, the anticipated greater thermal gradients below ridges will create suction of the water to a greater extent towards the ridges, thereby increasing the ice content below ridges if passing the permafrost table.

Downwards migration of water into permafrost is a well-documented phenomenon (Cheng 1983, Mackay 1983, Burn & Michel 1988). This process is generally quite slow (Williams & Smith 1989), but given the assumed high age of the rock glacier it seems quite probable that considerable thicknesses of segregation ice can develop. Also, the great temperature gradients found in this environment (Ødegård & Sollid 1993) support this possibility. However, the processes described can only explain high ice content in the upper part of the subsurface below ridges where winter cold waves occur, because the temperatures of the permafrost otherwise usually increase with depth.

### *Decrease in resistivities towards the front*

The decrease in resistivity values towards the front has been observed for several rock glaciers in Svalbard (Berthling et al.

1998, Isaksen et al. 2000). Haeberli & Vonder Mühll (1996) explain this pattern in the Alps by a change in the origin and thus resistivity of the ice. In the zones of longitudinal extension the active layer is stretched and thinned, causing melting of near-surface medium- to high-resistive ice. As the active layer is thinned, the permafrost base penetrates into unfrozen material at depth where groundwater freezes, causing low-resistive ice. On the other hand, in the longitudinal compression zones, flow-induced thickening of the active layer as well as the permafrost results in melting at the permafrost base and new, mainly low-resistive, ice formation at the permafrost table, although less than the quantity of ice that melts in the longitudinal extension zones. This model was adapted by Berthling et al. (1998) for high arctic rock glaciers, with an important consideration: since the permafrost base may be well below an arctic rock glacier, melting and freezing at the base of the permafrost will have no effect on the rock glacier.

As nearly no movement is found on the rock glacier at present, the extent of extension and compression is unknown. Also, the probable existence of different generations of creep lobes makes assumptions about velocity fields doubtful, but given the probable long-term low-velocity situation ( $\sim 0.5 \text{ cm a}^{-1}$ ) (Isaksen 2001), the model by Haeberli & Vonder Mühll (1996) does not explain the resistivity pattern found in the investigated rock glacier. The disappearance of the upper, relatively homogenous ice layer towards the front, as indicated by the GPR results, might very well be due to extension-induced melting. However, considering Rp 1 (Fig. 5), the resistivity decreases towards the front at all depths. Surface strain measurements indicate that the model by Haeberli & Vonder Mühll (1996) does not fit to Hiorthfjellet rock glacier, to the north-east of Longyearbyen, either (Ødegård et al. 2003b).

Ødegård et al. (2003a) suggest that the decrease in resistivity towards the front of the tongue-shaped Hiorthfjellet rock glacier is caused by limited shear strain, bringing the individual particles in closer contact. Deeper lying boundaries between alternating ice and debris layers will probably be blurred due to higher strain rates. The findings of Arenson et al. (2002) in the Alps, that high percentage of air voids within two rock glaciers at that site makes them compressible, might also hold for rock glaciers in Svalbard. If so, resistivities naturally decrease towards the front in step with high-resistive air voids disappearing and particles coming into closer contact.

Other explanations for the resistivity pattern of rock glaciers than those due to dynamic effects are possible. Isaksen et al. (2000) suggest that a climate change towards conditions gradually more favourable for precipitation could explain the resistivity pattern on tongue-shaped rock glaciers close to Longyearbyen. Considering Rp 1 (Fig. 5), it is possible that some of the resistivity anomalies (for example, 120 m from the front) are associated with certain climate conditions. However, the general trend with lower resistivities towards the front is more likely due to flow-induced changes in the electrical properties of the ice and debris mixture since rock glaciers of assumed different ages in Svalbard reveal the same overall resistivity pattern. An early Holocene, late glacial, and Weichselian start for development

of tongue-shaped, simple lobe-shaped, and the investigated rock glacier respectively, have been suggested (Sollid & Sørbel 1992, Berthling et al. 1998, Isaksen et al. 2000, Berthling 2001, Isaksen 2001, Kääb et al. 2002).

### *Distribution of the active layer thickness*

The resistivity results indicate that the active layer is thicker in both moist and relatively dry dumps than in ridges on the rock glacier. This is the opposite of what is found in the Alps. There, long-lying snow combined with mid-winter cooling by cold-air drainage in snow funnels is believed to cause thinner active layer in the furrows (Barsch 1996, Bernhard et al. 1998). Thickness and timing of the snow cover is a crucial parameter concerning microclimate (Goodrich 1982, Ling & Zhang 2003). The elevated surface of a ridge will generally have a thinner snow cover, allowing more intense winter cooling. Thus, if not compensated by long-lying snow in the furrows (as in the Alps), ridges will reveal thinner active layers than furrows. The difference between the sites is probably due to climatic differences. Svalbard receives less precipitation during winter than the Alps, (for example, annual precipitation of  $c.1000 \text{ mm}$  on the rock glacier Murtèl-Corvatsch, Swiss Alps (Hauck 2001) as opposed to MAP of  $480 \text{ mm}$  at Isfjord Radio), and the strong winds limit the snow cover to the dumps of the highly exposed rock glacier. Further, the microclimate in the Alps is highly dependent upon incoming solar radiation (Hoelzle 1996, Mittaz et al. 2000), whereby ridges receive a greater amount of radiation than furrows. The same holds for Svalbard, but aspect is a less important factor controlling the microclimate due to more common cloud cover (Hanssen-Bauer et al. 1990) and the fact that, in spite of the midnight sun, energy received by shortwave radiation in the summer is less because of the higher latitude.

### *The connection between ice content and measured velocities*

Generally speaking, movement of rock glaciers is due to creep of the permafrost. The deformation of such a solid is a function of the shear stress applied, the ice content, the thickness as an indicator of the normal stress, the temperature, the sizes and forms of grains and ice crystals, and the density of the ice (Barsch 1996). The two last factors are difficult to determine and are also probably of less importance. A non-linear effect of ice-debris temperature on the deformation rate, in which warmer permafrost reveals higher deformation rates assuming equal ice content (somewhat comparable to Glen's flow law for pure ice), is suggested by centrifuge modelling (Rea et al. 2000, Davies et al. 2001). However, temperature differences inside the rock glacier probably have little effect on the velocity distribution. The general pattern of slow-moving rock glaciers in Svalbard, on the other hand, might be partly explained by cold, less plastic, ice-debris mixtures (Kääb et al. 2002).

The only part of the rock glacier with detectable velocities ( $4\text{--}8 \text{ cm a}^{-1}$ ) (Kääb et al. 2002) coincides with a high-resistive area in the southernmost talus. It cannot be excluded that the high resistivities are due to air-filled cavities (Hauck

& Vonder Mühll 2003), but the lack of coarse-grained material at the surface gives no indication in this regard. More likely, this area is associated with high ice content. However, similar and even higher resistive areas occur in large parts on the northern side of the rock glacier, where no velocities are detectable. Thus, the ice content of the main rock glacier body is probably not the limiting factor, considering its apparent inactivity. Rather, the topographical setting affects the differences in velocities; the talus has a steep gradient and thus relatively high shear stress, while the main body flows onto an almost flat surface, the strandflat, leading to (nearly) dynamic inactivity. Also, differences in thickness might contribute to the differences in velocities. Similar results were indicated by Sollid & Sørbel (1992) based on theoretical considerations, and by Wagner (1996) based on geophysical measurements on less complex lobe-shaped rock glaciers near Ny-Ålesund. Considering the northern and southern talus, the overall gradient is similar (Figs. 5 and 6). However, the northern talus reveals smaller resistivity values, indicating smaller ice content, but also here differences in thicknesses might contribute.

## Conclusions

The investigated rock glacier is generally rich in ice. Distinct ice-rich areas along the length profile are associated with ridges. This pattern might be due to the ridges' possible origin in increased material production, thereby increasing burial and preservation of snow, combined with water movements into the permafrost along the assumed steeper temperature gradients of the ridges. The overall resistivity values decrease towards the front of the rock glacier. The reason for this is not certain, but it is likely due to dynamic effects.

A few continuous radargram reflectors are present, but the overall picture is a chaotic structural pattern in contrast to results from simpler lobe-shaped and tongue-shaped rock glaciers in Svalbard. This is most likely due to the apparent complex history of the investigated rock glacier.

Contrary to rock glaciers in the Alps, the active layer of the investigated rock glacier is thicker in furrows than on ridges. This is probably due to thin snow cover on ridges allowing intense winter cooling. The difference between the sites probably reflects differences in climate with respect to snow conditions and radiation.

The only part of the rock glacier with detectable velocities has high ice content (and a steep gradient). However, the main body probably reveals ice contents of the same order. Thus, the study indicates that the apparent standstill of the rock glacier is due to the low-gradient situation in which the rock glacier flows onto an almost flat surface, i.e. the strandflat.

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