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Combination of Conventional Geophysical Methods for Sounding the Composition of Rock Glaciers in the Swiss Alps

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ABSTRACT

The composition of rock glaciers was sounded by a combination of conventional geophysical methods near the lower limit of mountain permafrost in the Swiss Alps. P-wave velocity, direct current (DC) resistivity and year-round ground surface temperatures were measured on 32 talus-derived rock glaciers. Subsurface P-wave velocities differ significantly between non-vegetated (probably active/inactive) rock glaciers and vegetated (probably relict) rock glaciers. DC resistivities reflect structural differences in the rock glaciers (e.g. bouldery or pebbly, ice-cemented or highly ice-rich) rather than thermal differences (i.e. frozen or unfrozen). The combination of these methods provides reliable thermal and structural information on subsurface deposits. In addition, mean annual ground surface temperatures are equally good indicators of the distribution of permafrost as bottom temperatures of snow. Copyright © 2006 John Wiley & Sons, Ltd.

KEY WORDS: rock glacier; mountain permafrost; seismic; DC resistivity; ground temperature; Swiss Alps

INTRODUCTION

Geophysical soundings have been extensively used to investigate the internal structure of rock glaciers because direct observation is difficult due to their blocky materials and inaccessible locations. In particular, direct current (DC) electrical resistivity measurements have been commonly applied to indicate subsurface structure (e.g. Fisch *et al*., 1977; King *et al*., 1987; Haeberli and Vonder Mühll, 1996). In addition, the recent use of two-dimensional DC resistivity measurement has revealed structure in greater detail (e.g. Vonder Mühll *et al*., 2000; Ishikawa *et al*., 2001). However, Hauck and Vonder Mühll (2003) have argued that such tomographical methods include ambiguity in the model inversion and interpretation of DC resistivity. Thus, the interpretation of DC resistivity in rock glaciers and related terrain requires further improvement, despite the difficulty of eliminating ambiguity in the model inversion without direct observation. One way to reduce such ambiguity is to combine DC resistivity and P-wave velocity measurements (Hauck and Vonder Mühll, 2003).

This study aims to reassess the merit of combining simple, conventional methods for sounding rock glacier structure. Both DC resistivities and P-wave velocities were measured on 32 talus-derived rock glaciers lying near the lower limit of mountain permafrost in the Swiss Alps. Non-tomographical methods were used to reduce equipment weight and operational time. Special attention was given to the effects of spatial variations in lithological and thermal conditions. The results were compiled for three types of rock glacier: bouldery rock glaciers having an active layer of matrix-free boulders, pebbly rock glaciers consisting of matrix-supported pebbles and cobbles, and densely vegetated rock glaciers that mostly represent relict bouldery rock glaciers (see Ikeda and Matsuoka, 2002; Matsuoka *et al*., 2005; Ikeda and Matsuoka, 2006 for detailed classification). The first two types lack

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vegetation on their upper surfaces. Although data on geophysical properties have been obtained extensively from bouldery rock glaciers (e.g. Barsch, 1996; Haeberli and Vonder Mühll, 1996; Vonder Mühll *et al*., 2002), there is a paucity of data from pebbly and vegetated (relict) rock glaciers. P-wave velocities were combined with bottom temperatures of winter snow cover (BTS) and mean annual ground surface temperatures (MAST) to extend empirical methods to detect permafrost, such as the BTS method (e.g. Haeberli, 1973; Haeberli and Patzelt, 1982).

STUDY AREA

Measurements were undertaken in two mountainous regions in eastern Switzerland (Figure 1). The main study area is the Upper Engadin (Figure 1A), where extensive high mountain slopes engender the widespread occurrence of contemporary permafrost (e.g. Haeberli *et al*., 1992; Keller *et al*., 1998). The second area is Davos (Figure 1B), where mountain slopes at lower elevations were subject to permafrost development during previous cold periods.

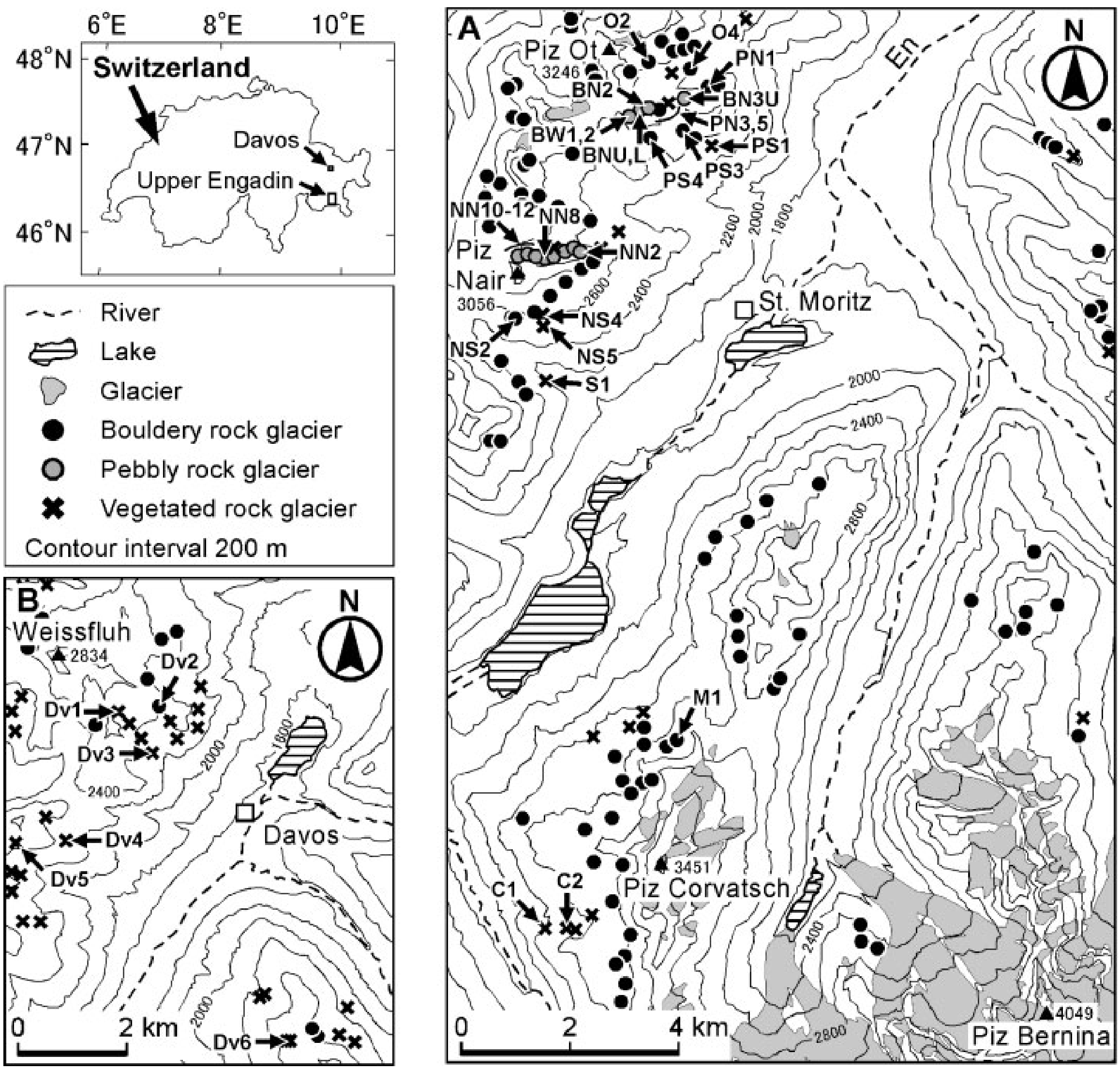

Figure 1 Location of the study areas, shown with the distribution of rock glaciers. (A) Upper Engadin. (B) Davos. The sounded sites are marked by arrows and identifiers (see text for explanation of identifiers).

Rock glaciers studied in the Upper Engadin were coded by the initials of adjacent peaks and slope aspects representative of exposed lithologies (Ikeda and Matsuoka, 2002, 2006). The names of the peaks employed are Ot (O—mainly consisting of granite/granodiorite), Padella (P—massive limestone/dolomite), Büz (B—shale and massive limestone/dolomite), Nair (N—shale and conglomerate), Murtèl (M—gneiss and schist) and Chüern (C—porous limestone and schist). The southern face of Nair is composed of conglomerate, whereas the northern face comprises shale partly embedding chert. Reflecting these lithological differences, Nair South (NS) rock glaciers are bouldery and Nair North (NN) rock glaciers are pebbly. Padella South (PS) bouldery rock glaciers include abundant pebbles and cobbles because calcareous conglomerate boulders derived from the southern face of Padella are largely shattered *in situ* after deposition. S1 rock glacier located in the Suvretta Valley originates from a small gneiss rockwall exposed solely on the flanks of this valley. In the Davos area, the rock glaciers studied are mainly composed of gneiss or amphibolite clasts.

Table 1 summarizes the location, morphology and composition of the rock glaciers studied. Bedrock lithology affects the surface structure of non-vegetated rock glaciers, so that massive limestone, conglomerate and most crystalline sources produce bouldery rock glaciers, whereas shale produces pebbly rock glaciers (Ikeda and Matsuoka, 2006). Vegetated rock glaciers lie in relatively warmer locations (Ikeda and Matsuoka, 2002).

The length and width of rock glaciers in Table 1 are those of the upper surfaces (i.e. the steep marginal slopes are excluded), and indicate the available extent of sounding profiles without large topographical gaps. The lengths and widths show that the rock glaciers are relatively small, so that only a few soundings on each rock glacier tend to be representative of the internal structures as a whole. In addition, all of the rock glaciers are talus-derived, regardless of the surface material. Further details about the rock glaciers studied in the Upper Engadin (photographs, large-scale maps, topographical profiles, etc.) can be found in Ikeda and Matsuoka (2002, 2006).

The sounded sites include two active rock glaciers where internal structures were directly observed: the Murtèl (M1) bouldery rock glacier (e.g. Haeberli *et al*., 1988; Arenson *et al*., 2002) and the upper lobe of the Büz North (BNU) pebbly rock glacier (Ikeda *et al*., 2003; Ikeda and Matsuoka, 2006). The M1 rock glacier has an approximately 30 m thick ice-rich layer, underlying a 2.5–3 m thick layer of openwork boulders and overlying frozen coarse debris down to 50 m deep near the sounding site. The upper half of the ice-rich layer is pure ice, whereas the lower half includes silt, sand and gravels in a highly ice-supersaturated structure. The bottom of the ice-rich layer consists of a 4 m thick layer of frozen sand and silt. The mean annual temperatures are −1.5 to −2°C in the ice-rich layer. The BNU rock glacier has ice-saturated or slightly supersaturated pebbles and cobbles embedded in sandy/silty matrix below a 2–2.5 m thick pebbly active layer. The frost table of the BNU lay about 1 m deep during the sounding periods and the temperature below the permafrost table was close to 0°C throughout the year.

METHODS

An easily portable seismograph, the McSEIS-3 (manufactured by Oyo, Japan), was used for refraction seismic soundings on 30 sites. This seismograph can accumulate up to 99 signals to improve the signal-to-noise ratio. A 4 kg sledgehammer was used to produce seismic pulses. To make a pair of travel time curves (i.e. forward and reverse arrival times measured on one survey line), seismic receivers (geophones) were fixed at both ends of a survey line and shot points were shifted. The shot points were placed at 2 or 2.5 m intervals near the receivers and elsewhere at 5 or 6 m intervals. Survey lines were 40–64 m long. The P-wave velocities of two or three layers and the depth of layer boundaries were determined using the intercept time method (Palmer, 1986). The reciprocal method (Palmer, 1986) was also employed to obtain more accurate P-wave velocities for the second layer by eliminating anomalies caused by irregular ground and refracting surfaces. The latter method was available only when a pair of travel time curves indicated a two-layer structure with the layer boundary at a relatively shallow depth (1–5 m deep in this study). The soundings were carried out from mid-July to mid-August 2000–03.

Ground surface temperatures were monitored automatically with data loggers on 25 rock glaciers. Miniature data loggers, Thermo Recorders TR-51 and 51A (T & D, Japan) and multi-channel data loggers, EXL-007-8A and LG-C1 (LOG Electronics, Japan) were used for the monitoring. These loggers recorded surface temperatures under the uppermost 2–5 cm of clasts at 1 h intervals with a resolution of 0.1°C, except for the EXL-007-8A logger which took measurements at 2 h intervals. The recorded data were calibrated by fitting the constant temperature below the wet snow cover in spring to 0.0°C to minimize the error near 0°C to within ±0.1°C. From the six-year

Table 1 Morphology and composition of sounded rock glaciers.

Site[1]	Vegetation[2]	Slope aspect (°)	Frontal elevation (m)	Length (m)	Width (m)	Frontal slope ht. (m)	Surface gradient (°)	Lithology[3]	Active layer structure[4]
Bouldery rock glaciers									
M1	NV	335	2630	270	150	15	12	GN, ST	OB
NS2	PV	195	2450	300	105	90[a]	13	CG	OB
O2	NV	160	2755	610	150	30	15	GR	OB
O4	PV	135	2570	230	210	40	10	GR	OB
PN1	NV	50	2535	540	130–380	5[b]	12	LS	OB
PN5	NV	25	2705	110	150	25	8	LS	OB
PS3	PV	210	2665	290	90–270	13	8	LS	BP/OB
PS4	PV	160	2680	280	190	25	13	LS	OB/BP/VS
Dv2	PV	205	2405	40	80	14	15	GN	OB
Pebbly rock glaciers									
BNU	NV	35	2800	50	130	10	25	SH, LS	PM
BNL	NV	35	2765	100	105	<1[b]	20	SH, LS	PM
BN2	NV	40	2730	45	120	20	18	LS, SH	BP[d]
BN3U	PV	340	2640	90	230	4[b]	28	SH, LS	PM
BW1	NV	270	2790	110	100	5[b]	23	SH, LS	PM
BW2	NV	270	2810	90	90	13	28	SH	PM
NN2	NV	5	2605	90	65	14	30	SH, CG	PM
NN8	PV	35	2705	240	100	10	17	SH	PM
NN10	NV	35	2735	245	100	7	20	SH, CG	PM
NN11	NV	20	2755	290	90	na[c]	15	SH, CG	PM
NN12	NV	30	2770	225	200	na[c]	18	SH, CG	PM
Densely vegetated rock glaciers									
C1	DV	310	2360	290	110	30	14	LS, ST	VS
C2	DV	340	2530	90	180	5–12	9	LS, ST	VS
NS4	DV	150	2565	430	140–300	35	9	CG	VS/OB
NS5	DV	170	2445	220	170	90[a]	15	CG	VS
PN3	DV	25	2650	80	110	10	7	LS	VS
PS1	DV	135	2455	180	140	35	16	LS	VS
S1	DV	25	2185	130	120	25	23	GN	VS
Dv1	DV	60	2460	60	210	3	11	GN	VS
Dv3	DV	130	2240	290	280	25	17	GN, ST	VS
Dv4	DV	50	2290	40	120	4	11	GN	VS
Dv5	DV	190	2410	110	70	5	13	GN?	VS
Dv6	DV	210	2480	50	80	10	11	AM	VS/OB

[1]First letter is the initial of adjacent peak except Dv (Davos); second letter refers to slope aspect (see text). [2]NV = not or sparsely vegetated (probably active); PV = partly vegetated (probably inactive); DV = densely vegetated (probably relict). [3]GN = gneiss; ST = schist; CG = conglomerate; GR = granite/granodiorite; LS = limestone/dolomite; SH = shale; AM = amphibolite. [4]OB = boulders without matrix; BP = boulders and pebbles/cobbles; VS = vegetated soil; PM = pebbles/cobbles with sandy/silty matrix. [a]The frontal height significantly overestimates the thickness of the rock glacier, because the slope lies on a 60 m high valley wall. [b]The frontal height underestimates the thickness of the rock glacier, because the upper surface inclines toward the front. [c]Artificially modified front. [d]A thin openwork bouldery layer covers matrix-supported pebbles and cobbles.

records of the ground surface temperature obtained in the study area (see Ikeda and Matsuoka, 1999, 2002; Matsuoka and Ikeda, 2003; A. Ikeda, unpublished PhD thesis, University of Tsukuba, 2004), the average temperatures in March 2000 and 2004 are proposed as representative of BTS values under normal snow condition (i.e. 1–3 m snow cover throughout late winter to spring and much thinner at the beginning of the winter). In addition to BTS, MAST values which are average temperatures for 365 days from early August 1999 and late July 2003 were determined.

One-dimensional DC resistivity soundings were performed on 26 sites from late July to early August 1999 and 2002 with the SYSCAL R1 resistivity meter

(Iris Instruments, France). The setting of the electrodes followed the Schlumberger array. About 20 apparent resistivity values were measured on a 200 m long profile. Modelled resistivity curves fitting with measured values were calculated with WinSev6 software (W_GeoSoft, Switzerland). First, the resistivities and boundary depths of subsurface layers were manually determined by fitting standard and auxiliary curves. Second, the initial values were automatically adjusted to measured resistivities by the least squares method (Koefoed, 1979).

'Equivalence' accompanying such an inverse modelling precludes a single solution (e.g. Koefoed, 1979). Thus, the precise resistivity and thickness of a subsurface layer can never be solved by DC resistivity methods alone. However, different apparent resistivity curves must result from different resistivity distributions within the ground. Taking these points into account, this study adopted a minimum number (three to five) of subsurface layers according to the convexity and concavity of measured resistivity curves. The restriction on layer numbers was intended for modelling the typical three layers of active rock glaciers: the uppermost unfrozen debris layer several metres thick, the middle creeping ice-rich layer several tens of metres thick and the underlying frozen or unfrozen debris layer (e.g. Haeberli *et al*., 1998; Humlum, 2000; Arenson *et al*., 2002), or its permafrost-degrading structure accompanied by soil development on the surface (e.g. Ikeda and Matsuoka, 2002). By comparing a number of resistivity stratigraphies obtained from the same manner of modelling, this study discusses structural differences that result in differing resistivity curves. In particular, one or more orders of difference in resistivity between layers at a similar depth can distinguish subsurface structures because the theoretically equivalent solutions could not compensate for such a difference without presuming extremely thin layers.

RESULTS AND INTERPRETATION

P-wave Velocity

Two to three velocity layers were identified for each pair of travel time curves (Table 2). A sharp break in P-wave velocities was observed in 26 pairs of travel time curves. The break results from the large contrast in velocity between the upper layer (330–780 m s^{-1}) and the lower layers (1900–4400 m s^{-1}), which represents the layer boundary between the upper unfrozen debris and lower frozen debris or bedrock. The other four curves lack such a break and have low P-wave velocities (360–950 m s^{-1}) over the whole sounded depth, except for an intermediate velocity (1600 m s^{-1}) below 7–12 m in PN3.

The calculated thicknesses of the low velocity layers are less than 5 m on six bouldery (M1, NS2, O4, PN1, PN5, PS3) and seven pebbly rock glaciers (BNU, BNL, BN3U, BW1, BW2, NN2, NN12). Most of these rock glaciers have steep frontal or lateral slopes higher than 10 m (Table 2), which indicate the absence of bedrock at least within 5 m depth. A frontal slope lower than 5 m in four pebbly rock glaciers (BNL, BN3U, BW1, NN12) underestimates the thickness of the rock glaciers, because either the upper surface gradually inclines toward the front or the frontal slope is artificially buried. The longitudinal profiles of these rock glaciers indicate that the sediments are 10 m or more thick (Ikeda and Matsuoka, 2006). Since the upper surface of the high velocity layer is too shallow to be bedrock, the second layer with a high P-wave velocity probably corresponds to frozen debris or ice. The calculated depths of the permafrost table in M1 and BNU rock glacier coincide with directly observed depths to within $\pm$ 1 m. In contrast, the thicknesses of low velocity layers are 9 m or more on seven vegetated rock glaciers (C1, NS5, PN3, PS1, S1, Dv3, Dv4). This can be attributed to the absence of permafrost or the possible presence of relict permafrost below a thick supra-permafrost talik.

On the other seven rock glaciers, the thicknesses of low velocity layers range from 5 to 9 m. In these cases, the thicknesses of bouldery rock glaciers (PS4, Dv2) are much thinner than the apparent debris thicknesses indicated by the heights of the steep frontal slopes. The NN8 pebbly rock glacier was also estimated to be 20 m thick from the longitudinal profile, although the steep (>30°) part is only 10 m high (Ikeda and Matsuoka, 2006). In contrast, the thicknesses of low velocity layers in the vegetated rock glaciers (C2, Dv1, Dv5, Dv6) approximate or exceed the apparent thicknesses indicated by frontal slopes (3–12 m high). The former is likely to indicate the presence of permafrost below a supra-permafrost talik and the latter the absence of permafrost within the thin rock glaciers. In general, the thickness of the low velocity layer distinguishes non-vegetated rock glaciers from vegetated ones.

These results and interpretations generally agree with those of previous studies where a high velocity layer (2400–4000 m s^{-1}) is regarded as an indicator of frozen conditions where it underlies a thin low velocity layer (250–1500 m s^{-1}) in a rock glacier (e.g. Haeberli and Patzelt, 1982; Barsch, 1996). In five

Table 2 P-wave stratigraphy, bottom temperatures of winter snow cover (BTS) and mean annual ground surface temperatures (MAST) of rock glaciers near the lower limit of mountain permafrost. Also displayed are the length of the sounding profile (AB), location of the sounding profile (LC) and height of the steep (30–45°) frontal slope (FS). Two tests were each performed on BNU, BNL and BW2 rock glaciers.

Site	P-wave stratigraphy[1]					AB	LC[2]	FS	BTS/MAST	
	First layer		Second layer		Third layer				1999–2000	2003–04
	V (m s^{-1})	D (m)	V (m s^{-1})	D (m)	V (m s^{-1})	(m)		(m)	(°C)	(°C)
Bouldery rock glaciers										
M1	350	1.3 ± 0.3[a]	3700			40	M	15		−6.7/−1.3
NS2	560	2.8 ± 0.8[a]	2700			50	M	90[d]	−5.2/0.4	−4.0/1.3
O4	650	2.7 ± 0.6[a]	2100			50	M	40	−4.7/0.9	−2.8/1.8
PN1	550	2.6 ± 0.5[a]	4300			55	M	10[e]	−3.7/−0.1	
PN5	650	1.5 ± 0.7[a]	3800			60	M	25	−4.7/−1.4	−3.8/−0.7
PS3	330	3.1 ± 0.3[a]	4400			60	M	13	−3.5/0.9	−3.1/0.9
PS4	380	8.8–9.2	2200			60	M	25	−1.9/0.4	−1.6/0.7
Dv2	360	4.8–6.2	2300			50	L-U	14		−1.7/1.3
Pebbly rock glaciers										
BNU-a	390	2.5 ± 0.6[a]	2800			50	L-U	10	−2.0/0.4	−2.2/0.7
BNU-b	370	1.6–4.2	2900			60	L	10	−2.0/0.4	−2.2/0.7
BNL-a	370	2.0 ± 0.3[a]	3000			50	U	8[e]	−1.2/0.9	−1.3/1.1
BNL-b	470	4.4–5.7	3400			40	L	<1[f]		−2.9/0.8
BN3U	350	2.4 ± 0.3[a]	2900			42	U	4[f]		−2.4/0.5
BW1	330	1.6 ± 0.4[a]	2600			40	L-U	5[f]		
BW2-a	350	2.3–5.3	2000			55	L-U	13	−6.4[h]/−1.0	−3.6[h]/0.7
BW2-b	330	2.7 ± 0.5[a]	2200			50	U	13	−6.4[h]/−1.0	−3.6[h]/0.7
NN2	420	4.6 ± 0.5[a]	2200			45	L-U	14		
NN8	450	4.8–8.6	3200			60	L-M	10[f]	−4.3/0.4	−3.4/1.0
NN12	340	2.1 ± 0.3[a]	3100			50	U	na[g]	−5.4/−0.5	−5.2/−0.4
Densely vegetated (probably relict bouldery) rock glaciers										
C1	360	2.5–3.3	670	>18[b]		45	M	30		
C2	410	4.7–6.3	2300			40	M	5–12		
NS5	400	1.8–4.7	730	17–22	3100	64	U	90[d]	−0.3/2.6	−0.1/3.0
PN3	480	1.5–1.8	700	6.7–12	1600[c]	48	M-U	10	−1.2/1.1	−0.4/1.3
PS1	410	2.2–4.2	950	>15[b]		40	M	35		−0.2/3.1
S1	410	2.6 ± 0.7[a]	700	>16[b]		40	M	25		−0.7/3.9
Dv1	340	2.2–3.0	780	7.2–11	3000	50	U	3		−1.0/2.0
Dv3	430	0.19–3.3	650	9.1–9.9	3200	50	U	5[e]		−0.1/2.7
Dv4	380	1.6–1.9	600	10	1900	50	L	4		−0.5/2.2
Dv5	530	4.7 ± 0.5[a]	3200			50	L-M	5		
Dv6	380	7.2–8.0	2000			50	L-U	10		

[1]V = P-wave velocity; D = depth of the layer base. [2]U = upper part; M = middle part; L = lower part. [a]Average depth and deviation calculated from the reciprocal method. [b]Minimum depth assuming that the velocity of the third layer is 2000 m s^{-1}. [c]Minimum depth is 16 m if the velocity of the fourth layer is 2000 m s^{-1}. [d]The frontal height significantly overestimates the thickness of the rock glacier, because the slope lies on a 60 m high valley wall. [e]Height of a side slope near the sounding profile. [f]The frontal height underestimates the thickness of the rock glacier, because the upper surface inclines toward the front. [g]Artificially modified front. [h]Snow cover is too thin to stabilize BTS.

non-vegetated rock glaciers (O4, PS4, Dv2, BW2, NN2), the P-wave velocity of the permafrost (2000–2300 m s^{-1}) is lower than the cited values but falls within the values for various types of permafrost (1500–4700 m s^{-1}) (Hunter, 1973). Such low velocities in permafrost may reflect honeycomb melting structures and/or large unfrozen water content caused by the ongoing inactivation of these rock glaciers. This is consistent with the relatively warm locations of O4, PS4 and Dv2 (see Table 1 for aspects and altitudes), thick active layers in PS4 and Dv2 (Table 2) and the low DC resistivity for rock-glacier permafrost in BW2 and NN2 (discussed below).

The conventional seismic method adequately indicates the presence of permafrost in rock glaciers thicker than 10 m, because the vertical difference in continuity and density between frozen and unfrozen debris appears to be much larger than lateral differences in similar material within 60 m of the soundings. In the following discussion, the presence or absence of permafrost is interpreted from seismic soundings.

BTS and MAST

Permafrost surveys using ground surface temperatures were checked against rock glaciers tested for the presence of permafrost by seismic soundings (Figure 2). The 8–10 m thicknesses of low P-wave velocity layers ($<2000\ \mathrm{m\,s^{-1}}$) distinguish non-vegetated (bouldery and pebbly) rock glaciers from vegetated rock glaciers. The former are probably active or inactive and the latter are relict.

Surface temperatures of most active and inactive rock glaciers decreased in the early winter under thin snow cover, and then remained nearly constant below −3°C or slowly increased towards the end of winter under thick snow cover (e.g. A. Ikeda, unpublished PhD thesis, 2004). As a result, average BTS values in March were lower than −2°C on 80% of the rock glaciers (Figure 2A). In contrast, BTS values on all of the relict rock glaciers, lacking permafrost at least in the uppermost 5 m, were −1°C or higher in late winter. Thus, BTS values of −1 to −2°C distinguish active/inactive rock glaciers from relict rock glaciers.

In winter 2004, about half of the active/inactive rock glaciers had BTS values higher than −3°C, which is the BTS value distinguishing ‘probable’ permafrost sites from ‘possible’ permafrost sites. The BTS values in 2004 were slightly higher than those in 2000, which may indicate that the snow cover was affecting the BTS method (Table 2), although the snow conditions were not as abnormal as those in previous years (2001, 2002, 2003) (*cf*. Hoelzle *et al*., 2003; A. Ikeda, unpublished PhD thesis, 2004). In addition, each active/inactive rock glacier with high BTS values has specific conditions which tend to maintain these values. The highest BTS value (−1.3°C on BNL) most likely originated from its exceptionally favourable location for snow accumulation (A. Ikeda, unpublished PhD thesis, 2004). Heat flow from a supra-permafrost talik, as indicated by thick (>5 m) low P-wave velocity layers in PS4 and Dv2, probably shifted BTS to the values slightly above −2°C. BTS values on BNU, BNL and BN3U pebbly rock glaciers (−1.3 to −2.9°C) were warmer

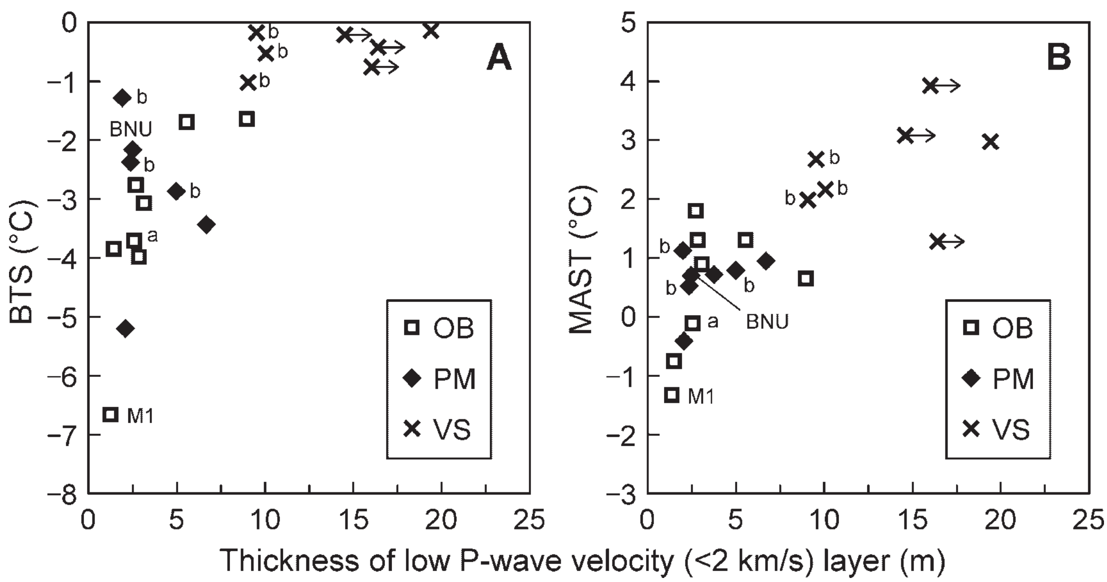

Figure 2 Relationship between the thickness of the low P-wave velocity layer ($<2000\ \mathrm{m\,s^{-1}}$) and ground surface temperature in 2003–04 on three types of rock glaciers: OB = bouldery type; PM = pebbly type; VS = vegetated type. The ordinate is given by (A) bottom temperature of winter snow cover (BTS) and (B) mean annual ground surface temperature (MAST). The point with an arrow indicates the minimum thickness. [a]Temperature data in 1999–2000. [b]The frontal slope is lower than 9 m. Values at two sites (M1, BNU) are individually indicated because the subsurface temperatures and structures were directly observed (see text).

than those on adjacent PN1 and PN5 bouldery rock glaciers. This is probably because the openwork active layers of bouldery rock glaciers promote convectional cooling, whereas the matrix-supported active layers of pebbly rock glaciers do not (Ikeda and Matsuoka, 1999; Matsuoka and Ikeda, 2003).

MAST is another indicator of permafrost (Figure 2B). The boundary MAST value between the presence and absence of permafrost was 1.5 to 2°C in 2003–04. MAST varied significantly from year to year, reflecting differing snow conditions and summer temperatures (A. Ikeda, unpublished PhD thesis, 2004). For example, MAST values in 2003–04 were 0.5 ± 0.5°C higher than in 1999–2000 (Table 2), reflecting the extremely hot 2003 summer and slightly thicker snow cover in 2004. Inevitably, the boundary MAST varies interannually. However, if the boundary MAST of each year is evaluated from the relationship shown in Figure 2B, the extensive use of miniature temperature loggers permits the establishment of an empirical rule for detecting permafrost from MAST. Such a rule would be particularly effective for locations where shallow snow invalidates the BTS method.

DC Resistivity

One-dimensional DC resistivity soundings produced resistivity stratigraphies characteristic of the three types of rock glaciers: bouldery rock glaciers have a resistivity of 10–50 kΩm on the surface and 10–500 kΩm in the subsurface (5–10 m deep), whereas pebbly and vegetated rock glaciers have resistivities of 1–5 kΩm on the surface and 0.5–10 kΩm in the subsurface (Table 3). The observed subsurface resistivities of the bouldery rock glaciers are within the range of previously reported values, also mostly obtained from bouldery rock glaciers (e.g. Haeberli and Vonder Mühll, 1996). Pebbly rock glaciers had significantly lower resistivities than values reported previously.

The resistivity of the surface layer indicates unfrozen openwork boulders on bouldery rock glaciers, unfrozen pebbles/cobbles filled with sandy/silty matrix on pebbly rock glaciers and soil on vegetated rock glaciers (Table 3). Four measurements on two bouldery rock glaciers (PS3-a, PS3-b, PS4-a, PS4-b) show low resistivity layers (1–5 kΩm) within the uppermost 0.7 m, because the electrodes were put into fine debris to decrease the contact resistivity. The relatively high resistivity (24 kΩm) on the BN2 pebbly rock glacier results from a thin openwork bouldery layer covering matrix-supported pebbles and cobbles. Two vegetated rock glaciers (NS4, NS5) have 1–3 m thick high resistivity layers (50–80 kΩm) near the surface (0.3–3 m deep), which indicate the presence of buried openwork boulders as confirmed by excavation at NS5.

Apparent resistivity typically increases from surface to subsurface layers in bouldery and vegetated rock glaciers and remains nearly constant in pebbly rock glaciers (Figure 3). The apparent resistivity of two bouldery rock glaciers (M1, PN5) significantly increases below the frost table estimated from seismic soundings, whereas that in two bouldery rock glaciers (NS2, PS3) increases from much deeper than the frost table. The former is probably active and the latter inactive (Table 1; see also Ikeda and Matsuoka, 2002, for classification of activity). The apparent resistivity of the vegetated rock glaciers increases from the beginning of the resistivity curves, which differentiates the shape of the curves from those of bouldery and pebbly rock glaciers. This is because vegetated soil is thinner than 1 m and is much wetter than the subsurface debris.

The apparent resistivities of the pebbly rock glaciers change only slightly across the frost table estimated from seismic soundings, except for BN2 and NN10 (Figure 3). The slightly higher resistivities below the frost table in five pebbly rock glaciers (BNU, BNL, BN3U, NN11, NN12) indicate the presence of permafrost along with high P-wave velocities. However, decreasing resistivity above 10 m depth in four pebbly rock glaciers (BW1, BW2, NN2, NN8) does not correspond to the presence of permafrost revealed by seismic soundings. The sharp decrease in resistivity in BN2 reflects clast size differences between the uppermost bouldery layer and the underlying pebbly layer. The resistivity curve of NN10 is similar in shape to that of active bouldery rock glaciers, but the apparent resistivities are an order of magnitude lower than those of the bouldery type.

The apparent resistivity clearly decreases in most bouldery rock glaciers below 10 m (M1, NS2, O2, PN5, PS4) and in half of the pebbly rock glaciers (BNU, BNL, BW1, NN11, NN12) below the same depths (Figure 3). The decrease probably indicates the boundary between ice-rich and less ice-rich layers or between frozen and unfrozen layers in rock glaciers. The absolute values of calculated resistivity and the upper depth for the low resistivity layer are somewhat ambiguous, however, because of inevitable modelling equivalence and lateral heterogeneity. Sounding profiles reaching the edge of non-vegetated rock glaciers show low apparent resistivity at the end of the resistivity curves, because the frozen core in each rock glacier is certainly lacking at the margins (see Tables 1 and 3 for surface extent and length of the sounding profiles).

Table 3 DC resistivity stratigraphy of rock glaciers near the lower limit of mountain permafrost. Also displayed are the surface material at the sounding site (SM), length of the sounding profile (AB) and thickness of the low P-wave velocity ($<2000\,\mathrm{m\,s^{-1}}$) layer (UF). Two tests were each performed on PS3, PS4 and NN8 rock glaciers.

Site	SM[1]	DC resistivity stratigraphy[2]							AB/2	UF
		First layer		Second layer		Third layer		Fourth layer		
		ρ (kΩm)	D (m)	ρ (kΩm)	D (m)	ρ (kΩm)	D (m)	ρ (kΩm)	(m)	(m)
Bouldery rock glaciers										
M1	b	23	2.9	170	22	1.1			100	1
NS2	b	52	8.9	84	24	8.0			64	3
O2	b	18	5.4	110	11	0.77			100	
PN5	b	18	2.2	410	18	0.18			100	2
PS3-a	bp[b]	4.6–30	1.5	9.2	6.5	190	18	7.2	100	3
PS3-b	bp[b]	3.8	0.52	10	11	37	54	6.3	100	3
PS4-a	v[b]	0.96	0.42	45	13	0.17			80	9
PS4-b	bp[b]	2.0	0.61	14	16	0.75			64	9
Pebbly rock glaciers										
BNU	p	0.60	0.48	2.1	6.6	3.7	16	0.90	50	2
BNL	p	5.7	0.30	1.8	3.1	3.7	42	0.26	100	2
BN2	pb	3.8	0.20	24	1.4	2.2			50	
BN3U	p	1.9	0.90	1.0	6.3	2.3	17	1.3	50	2
BW1	p	0.84	2.2	0.74	9.8	0.16	22	1.8	80	2
BW2	p	2.5	1.3	0.46	17	1.9			50	4
NN2	p	2.1	2.9	1.4	6.3	1.9			50	5
NN8-a	p	2.0	1.1	1.3	11	2.0			50	7
NN8-b	p	3.1	1.1	2.3	22	1.2			40	7
NN10	p	2.0	0.66	1.3	3.5	5.4			50	
NN11	p	3.7	2.7	6.4	26	0.62			50	
NN12	p	7.4	0.29	3.0	3.8	7.7	29	0.022	100	2
Densely vegetated (probably relict bouldery) rock glaciers										
C1	v	1.5	0.86	4.1	8.4	10			64	>18
C2	v	1.1	0.33	4.3	4.3	2.6	21	21	64	6
NS4[a]	v	2.9	0.27	81	1.4	23	8.4	1.4	64	
NS5	v	1.2	0.39	49	2.8	0.82			50	19
PN3	v	1.7	0.54	11	2.4	2.2	6.3	3.7	80	>16
PS1	v	0.16	0.48	1.4	5.7	2.0	24	0.88	64	>15

[1]b = boulders without matrix; bp = boulders and pebbles/cobbles; v = vegetated soil; p = pebbles/cobbles with sandy/silty matrix; pb = boulders covering matrix-supported pebbles and cobbles. [2]ρ = calculated resistivity; D = depth of the layer base. [a]Ikeda and Matsuoka (2002) regarded the rock glacier as inactive because of the low BTS values in 1999. [b]Electrodes were put on fine materials to decrease the contact resistivity, although boulders dominate on the rock glacier surface.

Directly observed structures in the M1 and BNU rock glaciers provide a constraint on the values of subsurface resistivity and show the great dependence on grain size and ice-supersaturation in permafrost. The highly ice-supersaturated permafrost in M1 results in an apparent resistivity of 170 kΩm. The calculated depth of the high resistivity layer indicates the basal depth of the ice-rich layer rather than that of the permafrost base. The ice-cemented layer below the ice-supersaturated layer has a resistivity of only 1 kΩm, although the value is much less reliable than that of the shallower part, mainly because of lateral heterogeneity. The ice-saturated debris at the melting point in BNU shows a resistivity of 2–3 kΩm, which is much lower than the typical values for permafrost in bouldery rock glaciers.

DISCUSSION

Permafrost in the studied rock glaciers probably has a mean annual temperature higher than −2°C, since the

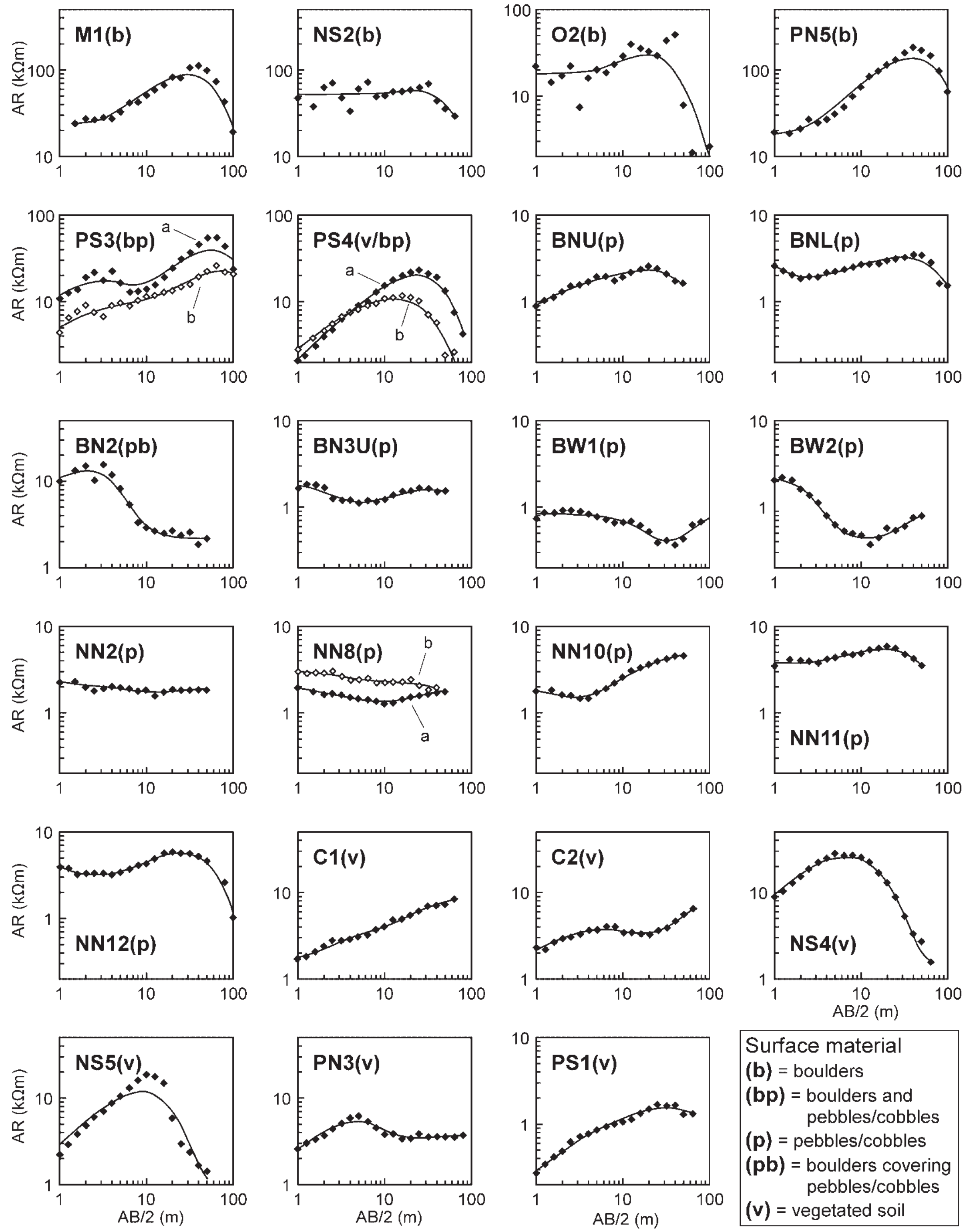

Figure 3 Apparent resistivity curves for bouldery, pebbly and vegetated rock glaciers. The abscissa is given by the half length of the sounding profile (AB/2) and the ordinate by apparent resistivity (AR). Note the different scale of apparent resistivity for each graph.

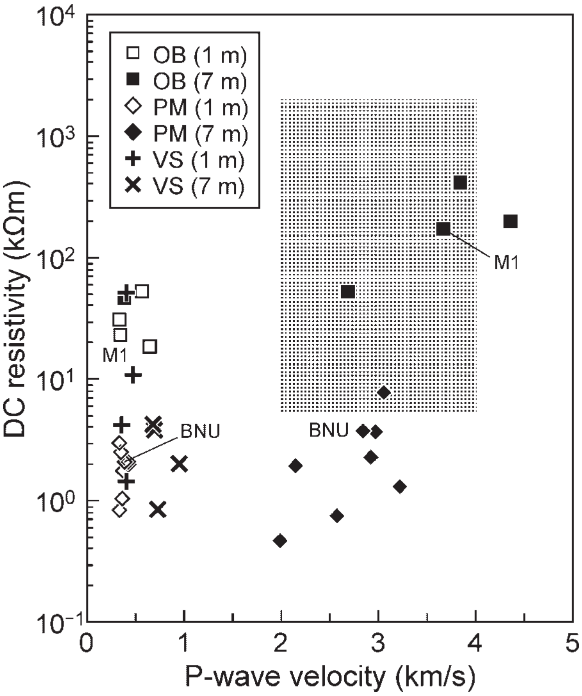

Figure 4 Relationship between P-wave velocity and DC resistivity at 1 and 7 m depth of three types of rock glaciers: OB = bouldery type; PM = pebbly type; VS = vegetated type. The shaded area indicates typical values for permafrost in (bouldery) rock glaciers (Haeberli and Vonder Mühll, 1996). Values at two sites (M1, BNU) are individually indicated because the subsurface temperatures and structures were directly observed.

lowest BTS and MAST values were recorded on the M1 rock glacier (Figure 2) where the permafrost temperature is mostly above -2°C (Arenson *et al*., 2002). Thus, the DC resistivity properties originate from warm permafrost.

Figure 4 displays the relationship between P-wave velocity and DC resistivity at near-surface (1 m) and core (7 m) depths for the three types of rock glaciers. The values of a thin vegetated rock glacier (C2) are eliminated from the figure because bedrock may exist below 5 m. The core P-wave velocity of the non-vegetated rock glaciers ($\geq 2000\,\mathrm{m\,s^{-1}}$), except for PS4 which may have a supra-permafrost talik, is much higher than the near-surface velocity (330–650 $\mathrm{m\,s^{-1}}$). In contrast, the difference in velocity between the near-surface (360–480 $\mathrm{m\,s^{-1}}$) and the core (670–950 $\mathrm{m\,s^{-1}}$) is small in vegetated rock glaciers. Where thick (>10 m) sediment is widely distributed, conventional seismic soundings are sufficient to elucidate local-scale permafrost distribution and to test an empirical rule using ground temperatures (see also Figure 2).

In contrast to P-wave velocity, DC resistivity has a wide overlapping range between the unfrozen surface (1–50 kΩm) and frozen subsurface (0.5–500 kΩm) near the lower limit of mountain permafrost (Figure 4). In particular, soil-rich pebbly rock glaciers often do not exhibit a difference in DC resistivity between surface and subsurface layers. No distinction in the subsurface resistivity is possible between pebbly and vegetated (probably relict) rock glaciers. In addition, the resistivity of openwork (unfrozen) boulders reaching up to 50 kΩm sometimes makes it difficult to distinguish an active layer from permafrost in bouldery rock glaciers. Thus, where a high resistivity layer (>100 kΩm) is absent, permafrost is difficult to identify by DC resistivity alone.

The differences in temperature between unfrozen and frozen layers do not always correspond to differences in DC resistivity (Figure 4). This contrasts with the observation that the DC resistivity of saturated sand and gravel sharply increases when ground temperatures fall below 0°C (Hoekstra and McNeill, 1973). However, the analogy from this result is misleading for the interpretation of DC resistivity in rock glaciers, because the active layer is dry rather than water-saturated (i.e. there is a difference in the water content between the active layer and permafrost).

The unfrozen water content in frozen debris depends on the total surface area of soil particles as well as on the temperature. Thus, the large difference in DC resistivity of permafrost between bouldery and pebbly rock glaciers partly reflects the material composition, because soil-rich pebbly rock glaciers hold more unfrozen water below the melting point than soil-poor bouldery rock glaciers (Figure 4). Similarly, the difference in DC resistivity between active/inactive and relict bouldery rock glaciers is attributed to material composition, because degradation of permafrost eventually results in weathering of clasts and subsequent production of fine debris, probably reducing the resistivity in rock glaciers (Ikeda and Matsuoka, 2002).

A wide variation in ice contents from less than 60% in ice-cemented debris (e.g. Fisch *et al*., 1977; Elconin and LaChapelle, 1997; Arenson *et al*., 2002) to nearly 100% in massive ice (e.g. Vonder Mühll and Holub, 1992; Arenson *et al*., 2002) has been reported for bouldery rock glaciers. Ice supersaturation increases the resistivity of permafrost because of a decrease in the continuity of unfrozen water films lapping soil particles (Fortier *et al*., 1994). Thus, resistivities much higher than those in openwork boulders (>100 kΩm) probably indicate the presence of a highly ice-supersaturated layer such as in the upper part of the M1 rock glacier (Haeberli *et al*., 1998).

Where permafrost is detected by other methods (e.g. seismic sounding), DC resistivity is a useful indicator of grain-size distribution and ice content (especially supersaturation) in warm permafrost (0 to -2°C) near the lower limit of mountain permafrost. Similarly, if an empirical rule using ground temperature (such as BTS) were tested by parameters more consistent than DC resistivity, DC resistivity would indicate subsurface structure less ambiguously if combined with the temperature parameter.

Differences in resistivity between same type rock glaciers may indicate thermal differences. The resistivity of permafrost decreases when the temperature increases, because unfrozen water content in permafrost mostly depends on temperature given similar material composition (e.g. Hoekstra and McNeill, 1973). For example, two inactive bouldery rock glaciers (NS2, PS3), characterized by relatively warm locations and high MASTs, have lower resistivities for permafrost than three active bouldery rock glaciers (M1, O2, PN5) (Tables 1–3; see Ikeda and Matsuoka, 2002, for detailed discussion). Similarly, the decreasing apparent resistivity of four pebbly rock glaciers (BW1, BW2, NN2, NN8) may result from permafrost at the melting point, as suggested by their relatively warm locations and/or thick active layers (Figure 3; Tables 1 and 2).

Two-dimensional resistivity methods are highly advantageous for displaying structure under rugged mountain slopes by modelling the lateral heterogeneity of the ground (e.g. Kneisel, 2004; Ikeda and Matsuoka, 2006). However, vertical and lateral extent of permafrost is still determined ambiguously by DC resistivity alone, because a number of equivalent solutions accompany inverse modelling (e.g. Koefoed, 1979) and changes in material composition mask thermal differences.

CONCLUSIONS

A combination of conventional seismic and DC resistivity soundings has the potential to reduce ambiguity concerning rock glacier subsurface thermal and structural information. Near the lower limit of mountain permafrost, the DC resistivity of rock glaciers reflects structural differences (e.g. bouldery or pebbly, ice-cemented or highly ice-rich) rather than thermal differences (frozen or unfrozen). Thus, where permafrost is detected by other methods (e.g. seismic sounding), DC resistivity indicates grain-size distribution and ice content.

DC resistivity significantly differs between bouldery (10–500 kΩm) and pebbly (1–10 kΩm) rock glaciers. Resistivities higher than 100 kΩm indicate the presence of a highly ice-supersaturated layer in bouldery rock glaciers, whereas resistivities of pebbly rock glaciers indicate soil-rich frozen debris with relatively high unfrozen water contents.

The wide, overlapping range of DC resistivity between non-vegetated and vegetated rock glaciers and between the surface and subsurface of rock glaciers demonstrates that the absence of permafrost may be difficult to predict on heterogeneous mountain slopes from DC resistivity alone.

Subsurface P-wave velocities differ significantly between non-vegetated rock glaciers ($\geq 2000\,\mathrm{m\,s^{-1}}$) and vegetated rock glaciers ($<1000\,\mathrm{m\,s^{-1}}$). Thus, P-wave velocities provide less ambiguous data than DC resistivity in sounding local-scale permafrost distribution where sediments thicker than 10 m are widely distributed. In addition, seismic soundings validate the BTS method and MAST as indicators of subsurface permafrost. The combination of these simple methods can enhance permafrost mapping in extensive inaccessible areas.

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