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ELECTRICAL D. C. RESISTIVITY SOUNDINGS WITH LONG PROFILES ON ROCK GLACIERS AND MORAINES IN THE ALPS OF SWITZERLAND

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With 11 Figures

SUMMARY

During construction work for the Grande Dixence power installations extensive electrical D. C. resistivity soundings, partially with long profiles, were carried out on two active rock glaciers and several neoglacial moraines in the "Combe de Prafleuri" (near Lac des Dix, Val d'Hérémence) and in the region of "Kintole" and "Chessi" (near the Taeschhorn, Mischabel group, Mattertal) within the Walliser Alpen (Switzerland). These soundings, although made about 20 years ago, furnish important geophysical information about the internal structure of alpine rock glaciers.

Resistivity values in temperate and slightly cold alpine glacier ice or buried snowbank ice are known to be on the order of 10 MΩm to more than 100 MΩm. On the other hand the resistivity values measured within the two active rock glaciers of the present study range from about 10,000 to about 300,000 Ωm. Similar resistivity values are known from perennially frozen sand and gravel in the Arctic as well as in the Alps, while unfrozen debris usually give values of the order of 1000 to 5000 Ωm in high alpine regions. Thus, it is shown that the two rock glaciers of the present study consist of perennially frozen debris and not of glacier ice nor of buried snowbank ice as it is often suggested. This interpretation was confirmed at the rock glacier in the "Combe de Prafleuri": an excavation several meters deep on the side of the rock glacier showed frozen debris very rich in ice. Only within the head region of the rock glacier in the "Chessi", which was still glaciated during the last century, some glacier ice was observed. This thin glacier ice seems to be embedded in thick frozen sediments. Even in this case the glacial contribution to the rock glacier formation is therefore probably restricted to the addition of some ice and morainic material on to the rock glacier surface.

The mean thickness of the high-resistivity permafrost-layer is on the order of 20 to 30 m with a maximum value of about 70 m in the "Chessi". Because the marked resistivity rise in frozen sand and gravel is known to take place at a temperature somewhat below 0° C, the thickness of the high resistivity layer only gives a minimum value for the real permafrost thickness. The mean permafrost temperature at the surface of the rock glaciers can be roughly estimated at about − 1 to − 2° C.

Below the high-resistivity permafrost-layer a thick low resistivity layer (1000 to 4000 Ωm) was encountered in all cases. It is suggested that this low resistivity layer represents, at least partially, unfrozen material, and that the rock glaciers therefore are not frozen to the bedrock in all places. Bedrock resistivity (gneiss of the Pennine nappes) was 2000 to about 12,000 Ωm and maximum bedrock depth in both cases markedly exceeded 100 m. The astonishing rock glacier thickness is due to a pronounced overdeepening in the bedrock relief of the former cirques where the rock glaciers originate. Although the order of magnitude of the electrically determined bedrock depth is probably quite correct, an accurate determination of the rock glacier thickness by geophysical soundings (seismic refraction and wide-angle reflection, geoelectrical resistivity soundings, gravimetry) seems to be difficult without better information about the thermal conditions within the permafrost. Deep drilling would be necessary to obtain reliable figures.

The neoglacial moraines in the immediate vicinity of the perennially frozen rock glaciers were not frozen (1000 to 5000 Ωm) as prooven by extensive excavations in the "Combe de Prafleuri". This observation confirms that active rock glaciers occur within the belt of discontinuous permafrost of the Alps.

ELEKTRISCHE GLEICHSTROM-WIDERSTANDSSONDIERUNGEN MIT LANGEN AUSLAGEN AUF BLOCKGLETSCHERN UND MORÄNEN DER SCHWEIZER ALPEN

ZUSAMMENFASSUNG

Während der Bauarbeiten für die Kraftwerksanlagen der Grande Dixence wurde eine große Zahl von elektrischen Gleichstrom-Widerstandssondierungen, z. T. mit langen Auslagen, auf zwei aktiven Blockgletschern und einigen neuzeitlichen Moränen durchgeführt. Obwohl diese Sondierungen in der „Combe de Prafleuri" (beim Lac des Dix, Val d'Hérémence) und in der „Kintole" und im „Chessi" (beim Taeschhorn, Mischabelgruppe, Mattertal) schon rund 20 Jahre zurückliegen, liefern sie wichtige geophysikalische Informationen über den inneren Aufbau von alpinen Blockgletschern.
Widerstandswerte in temperiertem und leicht kaltem Gletschereis der Alpen wie auch in begrabenem Schneefleckeneis liegen bekanntlich in der Größenordnung von 10 bis über 100 MΩm. Die in den beiden hier beschriebenen Blockgletschern gemessenen Widerstandswerte liegen dagegen zwischen etwa 10.000 und 300.000 Ωm. Ähnliche Widerstandswerte sind von dauernd gefrorenen Sanden und Kiesen sowohl in der Arktis wie auch in den Alpen bekannt, während ungefrorener Schutt in den Alpen meist Werte zwischen etwa 1000 und 5000 Ωm ergibt. Dies legt den Schluß nahe, daß die beiden Blockgletscher aus dauernd gefrorenem Schutt (Permafrost) bestehen und nicht etwa aus Gletscher- oder Schneefleckeneis, wie immer wieder angenommen wird. Am Blockgletscher in der „Combe de Prafleuri" wurde diese Interpretation bestätigt: Eine mehrere Meter tiefe Aufgrabung zeigte tatsächlich gefrorenen und sehr eisreichen Schutt. Nur gerade innerhalb der Wurzelregion des Blockgletschers im „Chessi" wurde etwas Gletschereis beobachtet. Dieses dünne Gletschereis scheint in gefrorenen Sedimenten von beträchtlicher Mächtigkeit eingebettet zu sein. Sogar in diesem Falle eines Blockgletschers, der eindeutig mit einem Gletscher oder Gletscherchen in Verbindung ist oder zumindest war, scheint die Rolle des Gletschers darauf beschränkt zu sein, etwas Eis und Moränenmaterial auf die Oberfläche des Blockgletschers abzulagern.
Die mittlere Mächtigkeit der hochohmigen Permafrostschicht liegt bei etwa 20 bis 30 m mit einem Maximum von rund 70 m im „Chessi". Da der markante Widerstandsanstieg in gefrorenen Sanden und Kiesen erst bei leicht negativen Temperaturen einsetzt, gibt die Mächtigkeit der hochohmigen Permafrostschicht nur einen Mindestwert für die tatsächliche Permafrostmächtigkeit. Die mittlere Permafrosttemperatur an der Oberfläche der beiden Blockgletscher kann roh auf etwa − 1 bis − 2° C geschätzt werden.
Unter der hochohmigen Permafrostschicht wurde in allen Fällen eine dicke niederohmige Schicht beobachtet (1000 bis 4000 Ωm). Vermutlich stellt zumindest ein Teil dieser Schicht ungefrorenes Material dar und die Blockgletscher sind deshalb nicht überall am Felsbett angefroren. Die Widerstände im Felsuntergrund (Gneis der Penninischen Decken) lagen zwischen 2000 und 12.000 Ωm und die maximale Tiefe des Felsuntergrundes war in beiden Fällen weit größer als 100 m. Die erstaunliche Blockgletscherdicke steht im Zusammenhang mit starken Übertiefungen im Felsbettrelief der ehemaligen Kare, aus denen die Blockgletscher herausfließen. Obwohl die Größenordnung der elektrisch bestimmten Felstiefe einigermaßen korrekt sein dürfte, scheint eine genaue Bestimmung der Blockgletscherdicke mit Hilfe geophysikalischer Verfahren (Refraktionsseismik, Weitwinkelreflexion, Geoelektrik, Gravimetrie) schwierig zu sein. Bessere Information über die thermischen Bedingungen im Permafrost oder Tiefbohrungen wären hier notwendig.
Die neuzeitlichen Moränen in der unmittelbaren Nachbarschaft der dauernd gefrorenen Blockgletscher waren nicht gefroren (1000 bis 5000 Ωm), was durch umfangreiche Grabarbeiten in der „Combe de Prafleuri" bestätigt wurde. Dies bestätigt die Vermutung, daß aktive Blockgletscher in den Alpen innerhalb der Höhenstufe mit diskontinuierlichem Permafrost auftreten.

INTRODUCTION

During construction work for the Grande Dixence power installations (Val d'Hérémence, Walliser Alpen, Switzerland, cf. Link 1970) extensive electrical resistivity soundings were carried out on two active rock glaciers and several neoglacial moraines. The measurements were made during the time from 1952 to 1960 by the office of W. Fisch sen. and jun. on the behalf of the Grande Dixence power company for practical purposes and the results were never published with the exception of very

short notes in Fisch and Fisch (1967) and Röthlisberger (1967a). Thus, the measurements are not new at all and more adequate equipment and techniques could certainly be used today. On the other hand the measurements reported in the following are the first resistivity soundings on rock glaciers and furnish very important geophysical information about the internal structure of alpine rock glaciers. An earlier publication of these measurements would have strongly simplified the still vehement discussion about the characteristics and the origin of rock glaciers. Therefore, the data material was worked up and re-interpreted by W. Haeberli in view of todays advanced glaciological knowledge. It is hoped that the interesting results of these consumptive measurements and observations may be a consolation for a heavily altered landscape at Prafleuri.

A. COMBE DE PRAFLEURI

GENERAL SITUATION

The "Combe de Prafleuri" is a small tributary valley of the uppermost Val d'Hérémence, immediately west of the Grande Dixence dam (Landeskarte der Schweiz 1:25.000, Blatt 1326 „Rosablanche"). At the upper end of this 4 km long valley a small flat glacier (Glacier de Prafleuri, B 74/18 in Müller et al. 1976) descends from the summit of the Rosablanche (3336.3 m. a. s. l.) to about 2860 m. a. s. l. (fig. 1). Neoglacial morainic material was deposited on a terrace-like bedrock shoulder in the northern part of the valley at about 2800 to 2900 m. a. s. l. and in the form of beautiful ridges in the central part of the valley down to about 2630 m. a. s. l. A big active rock glacier (about 1 km long) descending from the E to N facing slopes of Le Miroir (3129 m. a. s. l.) and Col de Mourti (2908 m. a. s. l.) to about 2620 m. a. s. l. lies immediately south of the former glacier tongue. The surface of this rock glacier was not glaciated during the last century as can be seen on the Topographischer Atlas der Schweiz, Blatt 527 "Lourtier", Revision X. Imfeld 1877, and in places is intensely lichen covered. Bedrock consists of gneiss of the Bernhard nappe, Pennine nappes (Carte spéciale N° 93: Carte Géologique de la région du Grand Combin, E. Argand 1905—1920).
For the construction of the Grande Dixence dam it was planned to use debris of both morainic and rock glacier material. After a first set of geoelectric soundings to calculate the volume of removable debris in 1952, the removal of the morainic ridges in the central part of the valley began. In 1956, when further debris supply was needed, a number of supplementary soundings were carried out in the moraines of the terrace-like shoulder in the northern part of the valley and on the rock glacier surface. Since 1956 a single probe prospecting configuration was used with one fixed current electrode in about 3 to 4 km distance from the operation area, one fixed potential electrode in about 2.5 to 2.8 km distance, one semi-fixed current electrode on the operation line and one moved potential electrode on the operation line. With respect to the more common Schlumberger configuration, the irregularities of the resistivity curves are smoothed out and better values for the lowermost layer — the bedrock — are obtained with this configuration. On the other hand the Schlumberger configuration was simultaneously measured in all profiles for the first 100 to 200 m to determine more precisely the near surface resistivity. The interpretation of the curves followed the common procedure as described e. g. by Bentz (1961, Bd. 1, p. 746ff.). About 3000 master curves were calculated or deduced from Schlumberger master curves. With Leclanché dry cells up to 720 Volt were applied at the current electrodes. Brass electrodes were used and the contact of the

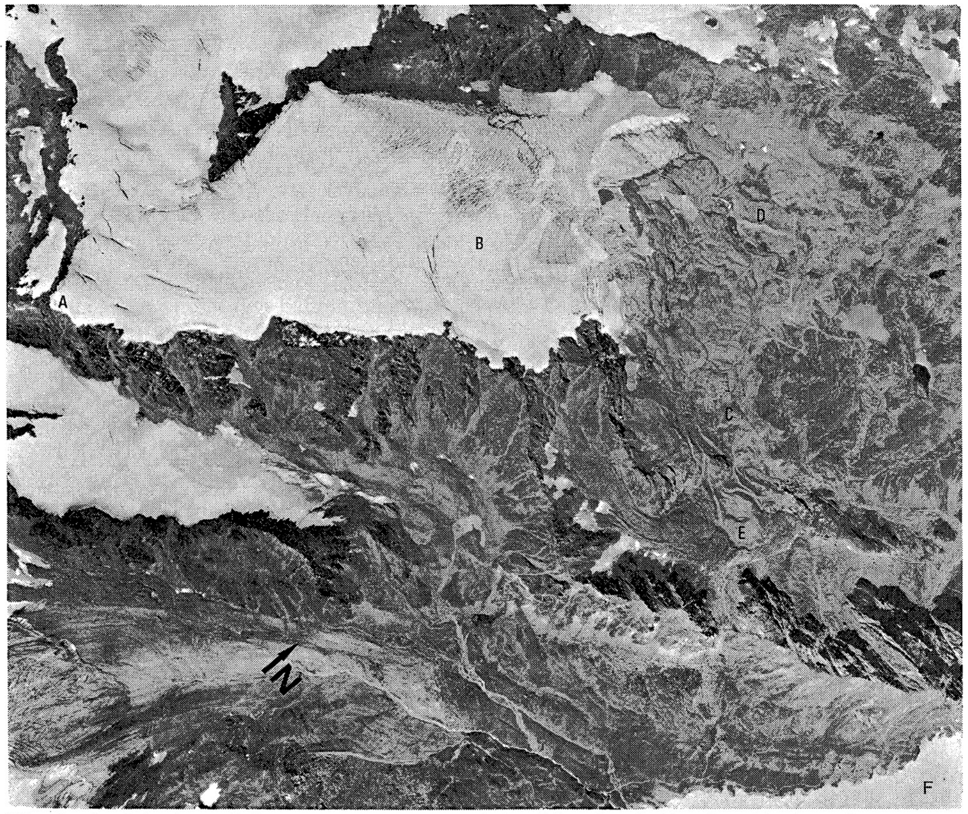

Fig. 1: Aerial view of the uppermost part of the "Combe de Prafleuri" after the excavations executed for the construction of the Grande Dixence dam. A = summit of the Rosablanche (3336.3 m a. s. l.), B = Glacier de Prafleuri, C = area of the former neoglacial morainic ridges in the central part of the valley, D = area of former neoglacial and older material on the terrace-like shoulder, E = active rock glacier, F = Lac des Dix. Luftaufnahme der Eidgenössischen Landestopographie, 13. August 1971.

electrodes in fine material or big boulders was improved by the use of sponges and brackwater. Potentials were measured by means of a compensation technique.

MEASUREMENTS AND FIELD OBSERVATIONS

The first soundings executed during the year 1952 in the region of the neoglacial morainic ridges in the central part of the valley (profiles A, B, C, D, E and F in fig. 2) revealed the expected feature of a more or less classical glacial trough within the bedrock relief with a maximum bedrock depth of about 80 m below surface in the lower section of profile A (at 2650 m. a. s. l.) and about 60 m in the middle of profile D (at 2735 m. a. s. l., fig. 6). Profile F followed a smaller secondary trough (mean bedrock depth about 50 m with 20 and 70 m as extremes). A very steep walled rock spur already visible at the surface seemed to separate the two troughs, which are well marked by the system of neoglacial morainic ridges at the surface (fig. 2 and 3). Resistivity values ranged from 1,000 to 6,000 Ωm in the neoglacial morainic material with a slight tendency towards higher values in the upper part of the

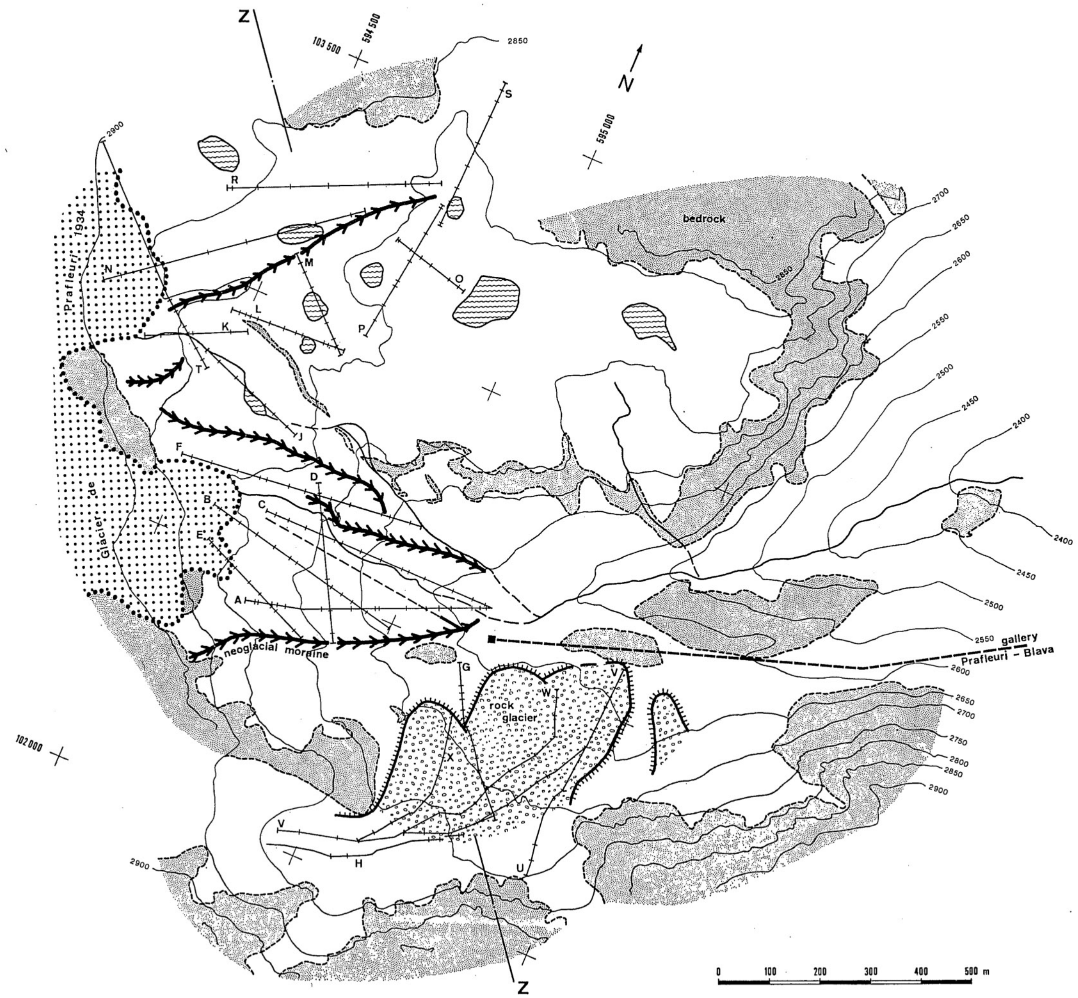

Fig. 2: Situation in the upper part of the "Combe de Prafleuri" before any excavation was made (topography from 1934). A, B, C ... = electrical resistivity soundings, Z—Z = summary cross section (cf. fig. 7), equidistance of the isohypses: 50 m.

profiles (towards the glacier) and near the surface of each profile. Bedrock resistivity was within the range from 2,000 to 12,000 Ωm with a clear tendency towards higher values in the near glacier region. Two profiles (G and H) also were measured on the surface of the rock glacier. Here, a high resistivity layer was encountered near the surface in both profiles. Resistivity values in this layer ranged from 15,000 to 50,000 Ωm on profile G and from 135,000 to 280,000 Ωm on profile H. The thickness of this high resistivity layer was about 20 to 30 m on both profiles. Low resistivity values occured near the surface (1,000 to 5,000 Ωm) to a depth of 1 to 3.5 m and below the high resistivity layer (1,900 to 3,200 Ωm). Bedrock was not reached in this case, because of the unexpected high resistivity of the near surface layer (spread too short).

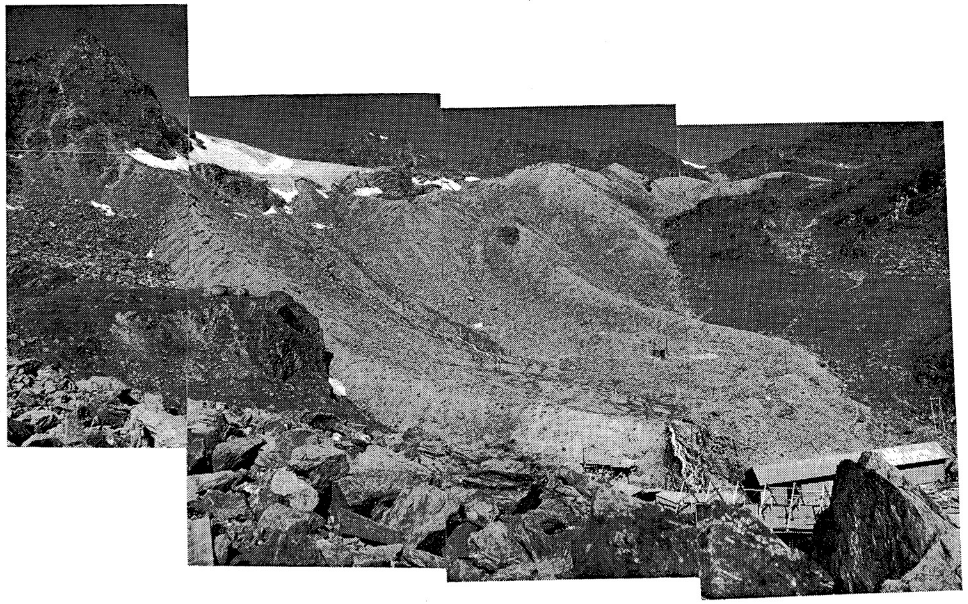

Fig. 3: The neoglacial morainic ridges of the Glacier de Prafleuri in the central part of the valley prior to the excavations. Foto by W. Fisch 1952.

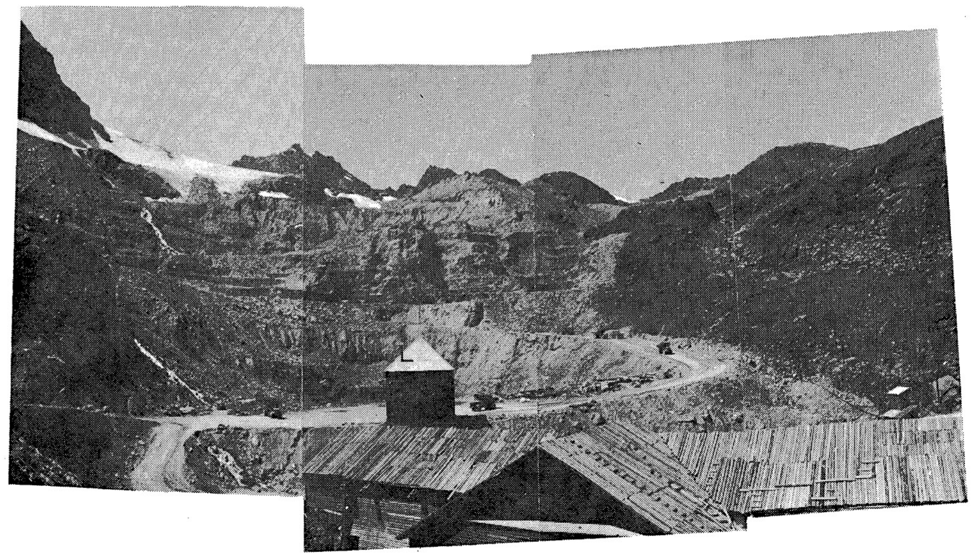

Fig. 4: About the same view as fig. 3, but after the excavations. Note the steep bedrock walls. Foto by W. Fisch 1956.

During the following years, the excavation of the morainic material in the central part of the valley (fig. 3 und 4) confirmed the bedrock depth determinations from the soundings (real bedrock depth within $\pm 10\%$ of the expected value). No dead ice nor frozen ground was encountered in this region, while an excavation at the left side of the rock glacier near profile G to a depth of about 10 m revealed angular debris cemented by interstitial ice beneath a thin layer of unfrozen coarse blocky debris at the surface (fig. 5). The ice content was very high — the debris fragments were embedded in ice — and the size of the cristals was very small (in a written note from W. Fisch sen. the ice cristals are described as "... far too small to be of glacial origin"). Thus, it was clear that the high resistivity layer detected in the profiles G and H was perennially frozen ground. Another interesting result was that the rock spur separating the troughs of the profiles A to C and F respectively had a nearly vertical wall in the downvalley direction as expected from the soundings. A nearly vertical wall also was excavated at the isolated rock spur near profile G in the direction of the neoglacial moraines (fig. 2 and 4). Another secondary trough with important bedrock overdeepening was crossed by the tunnel connecting the area of debris removal with the dam site (gallery Prafleuri - Blava on fig. 2). At a distance of about 500 m east of the rock glacier bedrock depth was seen to be greater than 50 m and a strongly overdeepened secondary trough was expected to exist in the region of the rock glacier.

Fig. 5: Excavation at the left side of the rock glacier in the "Combe de Prafleuri" near the resistivity profile G: small cristal size frozen ground, very rich in ice was observed below a thin unfrozen surface layer. Foto taken by W. Fisch 1956.

The need for further debris supply made it necessary to complete the soundings on the rock glacier and to make new measurements in the region of the terrace-like shoulder in 1956. These latter soundings were executed partially within neoglacial morainic material and older morainic debris (profiles J to P and R to T on fig. 2). The bedrock relief revealed by the soundings and in places by the following excavations was a relief of rounded hummocks with a number of small depressions. Maximum bedrock depth was observed within profile N (3 depressions, 50 to 60 m) and profile L (neoglacial morainic ridge, 40 to 50 m). Resistivity values in the neoglacial moraine were about 2,000 to 6,000 Ωm (profiles N and R) and slightly lower in the older morainic material (1,000 to 3,000 Ωm, profiles M, O, P, S, fig. 6). The profiles N and R showed a two layer system within the neoglacial morainic material with a lower-resistivity layer beneath a higher-resistivity layer. Because the resistivity values of the lower-resistivity layer correspond well with the values of the older morainic material as observed in the profiles M, O, P and S, it is supposed that the layer of neoglacial morainic material with a thickness of about 10 to 30 m overlies a somewhat thicker layer of older morainic material. Bedrock resistivity was 2,000 to 7,000 Ωm and the resistivity of water saturated sediments within small lakes was on the order of 800 to 1,200 Ωm.
In 1956 the longer spread of the profiles U, V, W and X on the rock glacier surface now allowed bedrock depth determinations in this region. On profile V — the longest of all profiles — the resistivity of the permafrost layer ranged from 10,000 to 195,000 Ωm with a clear tendency towards higher values in the upvalley direction (fig. 6). The thickness of the same layer was calculated to be about 15 to 40 m, while

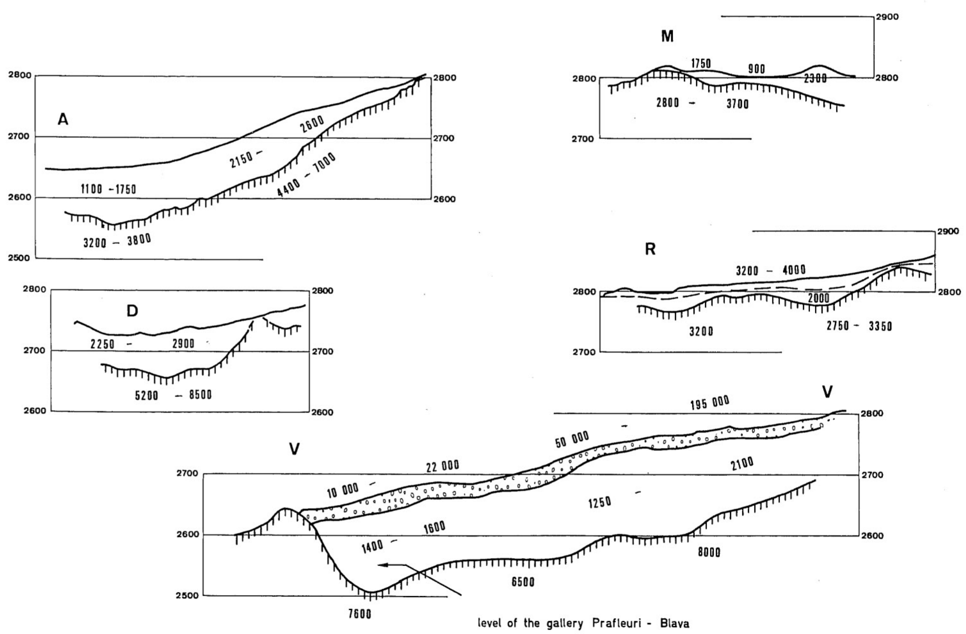

Fig. 6: Some examples of resistivity profiles from the "Combe de Prafleuri" (cf. fig. 2). Average resistivity values are given in Ωm, altitudes is in m a. s. l. Vertical scale not exaggerated!

bedrock depth reached an astonishing value of 130 to 160 m except in the frontal part of the rock glacier. This profile also showed that a low resistivity layer (1,400 to 2,200 Ωm) was present between the permafrost layer and the bedrock (6,500 to 8,000 Ωm). The profiles W and X clearly confirmed these results. The resistivity within the permafrost layer was 16,000 to 90,000 Ωm on profile W and 22,000 to 250,000 Ωm on profile X, while maximum bedrock depth reached 180 m in profile X and even 190 m in profile W. Profile U possibly showed a thin permafrost layer beginning at the rock wall (about 50,000 Ωm, only several meters thick) and reaching the profile V with a 25 m thick 8,000 to 9,000 Ωm layer, but the interpretation seems to be very ambiguous here. All these profiles (U, V, W and X) revealed a thick low resistivity layer with a mean resistivity value of about 1,500 Ωm between the high-resistivity permafrost-layer and the bedrock.

INTERPRETATION

The results of all soundings and excavations are summarized in a profile crossing the whole valley from SE to NW (Z—Z in fig. 2, fig. 7). The surface topography of this profile does not take into account modifications due to excavations. The resistivity values are taken from the different soundings crossed by this summary profile and are valid for the uppermost layer of about 10 to 20 m but not for the uppermost 1 to 5 m. Resistivity ranges for bedrock, water saturated sediments, older moraine, neoglacial moraine and perennially frozen ground (rock glacier ice) are compiled

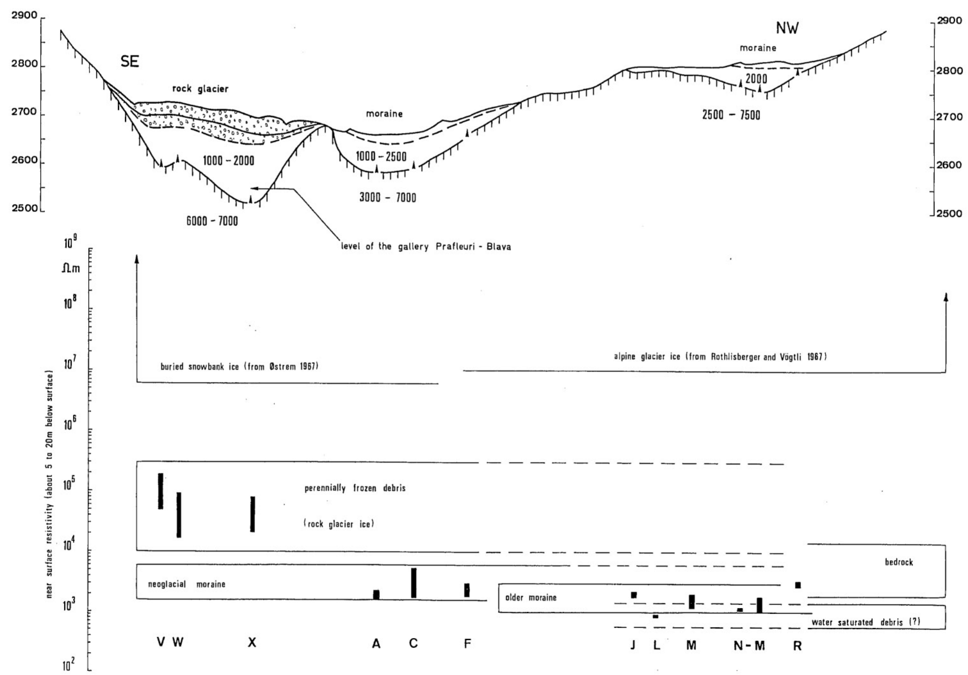

Fig. 7: Summary cross section through the upper part of the "Combe de Prafleuri" (see fig. 2 for position) with near surface resistivity values taken from the profiles A, B, C . . . Vertical scale not exaggerated!

from all data available from the soundings. Resistivity values for alpine glacier ice are taken from Röthlisberger and Vögtli (1967, p. 619) and values for buried snowbank ice from Østrem (1967, p. 646, laboratory measurements).
From this figure it is quite evident that the rock glacier material differs from the morainic material of the other parts of the valley: the rock glacier is perennially frozen, while the moraines are not. Furthermore it is noteworthy that no glacier ice nor buried snowbank ice was detected in the whole set of soundings (the profiles B, E, K, N and T did not reach the glacier, which had melted away in this region during the warm period between 1934 and 1952/56 respectively, fig. 2). As will be discussed in a later section, the thickness of the high-resistivity permafrost-layer only gives a minimum value for the real permafrost thickness and therefore the real thickness of the permafrost layer may possibly be 50 to 60 m instead of 20 to 30 m or even more. The low resistivity layer between bedrock and permafrost within the rock glacier could be interpreted as older morainic material or (at least partially) as water saturated sediments. The deepest point of the bedrock valley floor is within the secondary trough, which is occupied and masked by the rock glacier. It is interesting that the deepest point of this valley is only about 30 m deeper than the point, where the gallery Prafleuri - Blava crossed the same debris filled valley at a distance of about 750 m in the downvalley direction. Thus, the depth determination — although not controlled by excavations — seems to be quite realistic. Because the rock glacier is at a place, where no glacier existed in the middle of the last century and because alpine glaciers during postglacial time never reached a much greater extention than in the middle of the last century (Patzelt and Bortenschlager 1973), it seems that the rock glacier probably never was in contact with a real glacier. It may be surprising that the trough occupied by this rock glacier is clearly deeper than the troughs in the middle and in the northern part of the valley, which certainly were glaciated several times during postglacial glacier advances. The uppermost part of the "Combe de Prafleuri" seems to have an asymmetric cross section in the bedrock relief with the deepest depression in the southernmost part — an observation, which is also made in the region of "Kintole" and "Chessi", another alpine valley of similar character.

B. KINTOLE — CHESSI

GENERAL SITUATION

"Kintole" and "Chessi" are two places in the middle section of a small and unnamed tributary valley of the uppermost Mattertal (Landeskarte der Schweiz 1:25.000, Blatt 1328 „Randa"). The Wildibach, which reaches the Mattervispa between Randa and Taesch, drains the waters of the Kingletscher (fig. 8). This glacier descends in two separate streams from the north to west facing slopes of the Dom (4545.4 m. a. s. l.) and the Taeschhorn (4490.7 m. a. s. l.) to about 2900 to 3000 m. a. s. l. (B 55/8 and 9 in Müller et al. 1976). Neoglacial morainic ridges are well developed for the two formerly joined glacier tongues in the northern part of the valley between about 2400 and 3100 m. a. s. l. A much smaller (cirque) glacier must have existed in the extremely well shaded and steep walled cirque between the Kinhorn (3752 m. a. s. l.) and the Leiterspitzen (3268 m. a. s. l.) at about 2850 to 3100 m. a. s. l., as can easily be seen on the Topographischer Atlas der Schweiz, Blatt 533 „Mischabel", Revision X. Imfeld 1878/79. This glacier or glacieret (B 55/10 in Müller et al. 1976) was reduced to a greater perennial snowbank during the 20. century. Its extension during its maximum stage of neoglaciation seems to be difficult to delineate, because

Fig. 8: Aerial view of the region of "Kintole" and „Chessi" with the two tongues of the Kingletscher. 1 = Neoglacial moraines of the Kingletscher, 2 = the formerly glaciated cirque named "Chessi", 3 = the active rock glacier originating from the "Chessi", 4 = Another active rock glacier originating from another cirque, which was probably not glaciated during whole holocene period (5). A, B, C ... = electrical resistivity soundings, X — X = position of the summary cross section (cf. fig. 12).

morainic ridges cannot be detected in its surrounding from field observations or air foto interpretation. In front of this former cirque glacier a long (1.5 km) active rock glacier nearly parallels the neoglacial morainic ridge at the left side of the formerly joined Kingletscher tongues. Connected with this rock glacier in the head region, another rock glacier occupies the southernmost part of the valley. This second rock glacier originates in another cirque at the foot of the Leiterspitzen, which, very

probably, was not glaciated during the last century and therefore probably also not glaciated during the whole postglacial time period. The fronts of the two active rock glaciers facing northwest lie at about 2500 m. a. s. l. Bedrock consists of gneiss of the Bernhard nappe, Pennine nappes (Geologischer Atlas der Schweiz 1:25.000, No. 43 „Randa“, P. Bearth 1934—1960).
To collect the waters from the western side of the Mischabel group for the Grande Dixence power scheme, a water gallery had to be made in the region of "Kintole" and "Chessi" at an altitude of about 2500 m. a. s. l. Bedrock depth determinations by means of geoelectrical resistivity soundings were carried out to guarantee that the planned gallery would be within solid bedrock at every place. Four profiles were measured in the region of the neoglacial moraines and three profiles on the surface of the active rock glacier parallelling the left neoglacial moraine and in the area formerly occupied by the small cirque glacier within the "Chessi". The same measuring techniques were used as described for the soundings in the "Combe de Prafleuri".

MEASUREMENTS AND FIELD OBSERVATIONS

All soundings (A, B, C, D, G, F and H on fig. 8) were carried out during the summer of 1960. Profile A on the right neoglacial moraine of the former Kingletscher tongue revealed in its lower part a 10 to 20 m thick morainic layer (2,600 to 4,200 Ωm) covering the 6,000 to 9,000 Ωm. bedrock. In the upper section of the profile above about 2660 m. a. s. l. an overdeepening of the bedrock was observed (fig. 9). Here, the thickness of the 2,200 to 3,000 Ωm morainic layer amounts to 55 to 70 m. A somewhat smaller overdeepening was also found in the upper section of profile B between 2640 and 2720 m. a. s. l. (bedrock depth about 70 m below surface again). In the lower section of this profile, bedrock depth is mostly between 0 and 10 m., but at the uppermost end of profile B a short supplementary profile (VII, not indicated on fig. 8) revealed the presence of another overdeepening within the bedrock (60 to 70 m), which would not be expected from the bedrock topography at the surface in the immediate vicinity of this profile. The profiles C and D ran within the basin between the neoglacial lateral moraines and gave — as expected — only shallow bedrock depth values. The morainic debris cover is somewhat thicker in the region of profile D (up to 30 m in the middle section) than of profile C (3 to 12 m with a maximum of about 20 m in the lowermost section) and the resistivity values of this layer, which may be (partially) water saturated from the main outflows of the Kingletscher, range from 1,000 to 2,500 Ωm, while bedrock resistivity is on the order of 5,000 to 10,000 Ωm (fig. 9).
The profiles G, F and H were nearly parallel to the main longitudinal axis of the bis active rock glacier which is in immediate contact with the left neoglacial moraine of the Kingletscher (fig. 8). As expected from the experience with the rock glacier in the "Combe de Prafleuri" a high resistivity layer was found near the surface over the whole length of profile G (20,000 to 90,000 Ωm with a thickness of about 15 to 30 m, fig. 9 and fig. 10). Below this high-resistivity near-surface layer a very thick low resistivity layer (2,900 to 4,000 Ωm) was observed and bedrock depth amounted to an astonishing value of 210 m at the uppermost end of this profile. At about 2750 m. a. s. l. a pronounced "riegel" (bedrock depth 135 m) probably corresponding with the quite similar feature in the middle of profile A (fig. 9) was found. Even much higher resistivity values for the near surface layer (below the unfrozen debris cover of some 1 to 3 m thickness) were observed over the whole length of profile F (up to 300 MΩm, thickness about 20 m, fig. 9 and 10). Once more, a very thick low

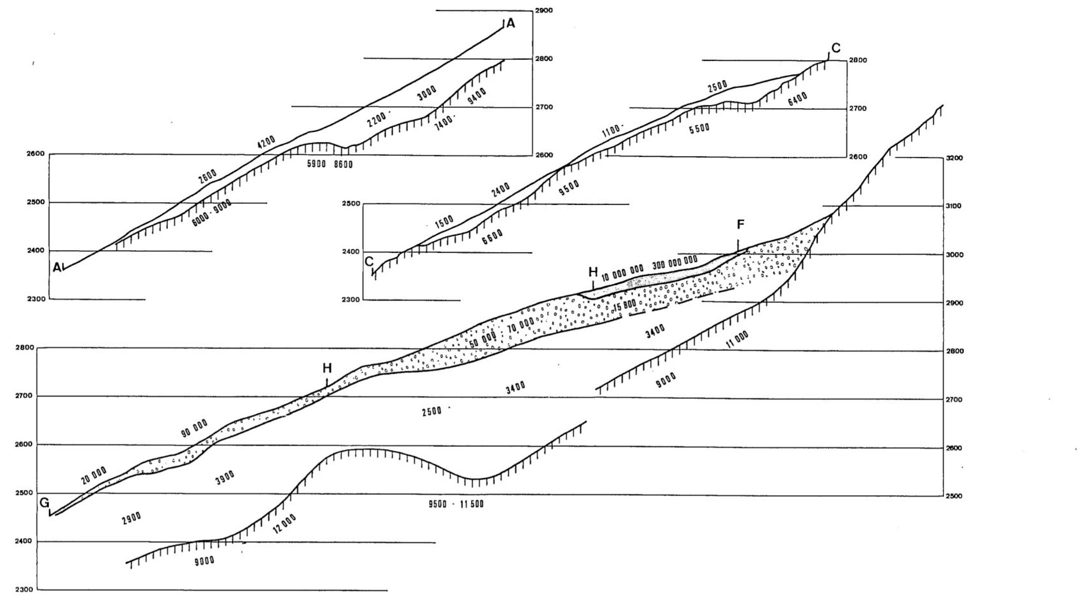

Fig. 9: Some examples of resistivity profiles from the region of "Kintole" and "Chessi". Average resistivity values are given in Ωm, altitude is in m a. s. l. Vertical scale not exaggerated!

resistivity layer (3,400 to 3,900 Ωm) could be observed below this layer and bedrock depth amounted to about 220 m near the intersection of the profiles F and G. The characteristics of the near surface layer with extremely high resistivity were soon clarified by the observation of some glacier ice in the lowermost section of profile F. At the base of the thin glacier ice, which was seen in a small outcrop, there was still sort of a subglacial channel (water outlet). A most interesting sounding could be carried out within this channel (fig. 10): the resistivity curve measured at the base of the near-surface glacier-ice layer clearly revealed the presence of a high resistivity layer (15,800 Ωm) of about 60 to 65 m thickness underneath, thus showing that the buried glacier ice of some 10 to 20 m thickness rests on a thick layer of perennially frozen debris. The great resistivity difference between the extremely high-resistivity glacier-ice layer at the surface and the low resistivity layer at depth respectively makes it impossible to detect the permafrost layer with intermediate resistivity between these two layers from soundings made at the surface of the glacier ice.
Profile H interconnecting the profiles G and F and nearly paralleling the lower section of profile F clearly confirmed these results. Again, a high-resistivity near-surface layer (50,000 to 70,000 Ωm) overlaied a thick low resistivity layer (2,500 to 3,400 Ωm). The thickness of the high resistivity layer was calculated to be on the order of 55 to 70 m, which is in good agreement with the 60 to 65 m found below some meters of glacier ice as mentionned above in profile F. Bedrock depth in the middle of profile H was found to be on the order of 330 m. At the intersection of the profiles H and F a discrepancy in bedrock depth of about 50 to 60 m exists (fig. 9), bedrock depth below the buried glacier ice being smaller than below permafrost without glacier ice. The 60 to 65 m thick high-resistivity permafrost-layer

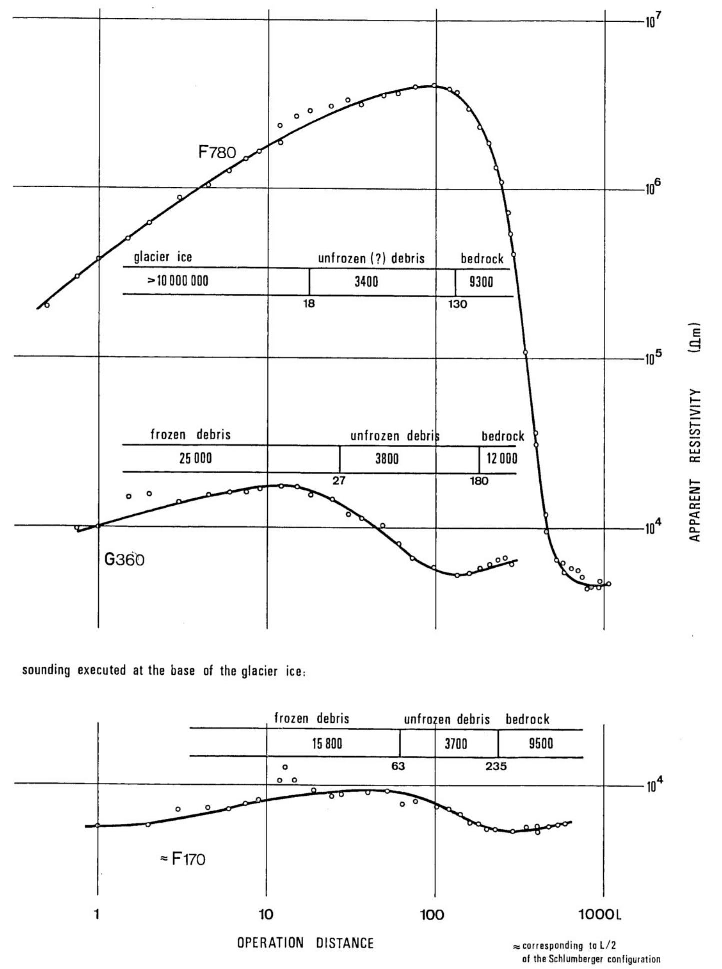

Fig. 10: Three examples of resistivity curves measured at the surface of the rock glacier in the "Chessi". Note broken scale for apparent resistivity!

found at the base of the glacier ice, however, was not taken into account for the bedrock depth determinations made from the soundings, which were carried out at the surface of the glacier ice. Therefore, it is possible that the bedrock depth value determined below the glacier ice is somewhat too small. On the other hand it may be seen from this example that bedrock depth determinations by means of electrical resistivity soundings can be somewhat ambiguous.

INTERPRETATION

As in the case of the "Combe de Prafleuri" the results of all soundings within the region of "Kintole" and "Chessi" are summarised in a profile (X—X in fig. 8) crossing the whole valley from south to north (fig. 11). Again, the resistivity values are taken from the different soundings crossed by this summary profile and are valid for the uppermost layer of about 10 to 20 m but not for the uppermost 1 to 5 m. Resistivity ranges for bedrock, neoglacial morainic material and perennially frozen

debris (rock glacier ice) are compiled from all data available from the soundings within the studied region, and restitivity ranges for alpine glacier ice and buried snowbank ice are taken from Röthlisberger and Vögtli (1967) and Østrem (1967) respectively.

The features observed in this region are very similar to the features observed within the "Combe de Prafleuri". The system of neoglacial moraines seems not to be perennially frozen nor to contain any dead glacial or buried snowbank ice. On the other side the rock glacier obviously consists of perennially frozen debris of quite important thickness (at least 15 to 70 m, probably even more as indicated on fig. 12) and carries on its back and in the uppermost head region a thin layer of buried glacier ice (or buried snowbank ice, if the very small cirque glacier or glacieret delivering

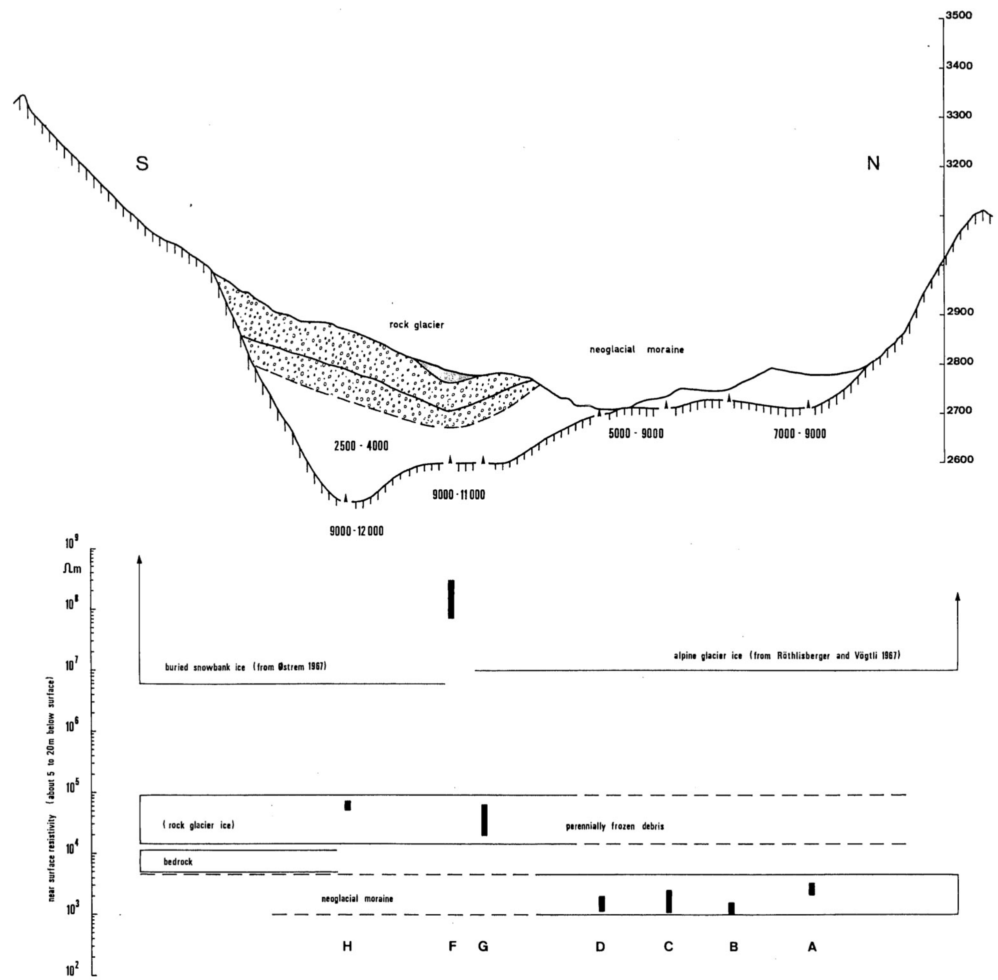

Fig. 11: Summary cross section through the region of "Kintole" and "Chessi" (see fig. 8 for position!) with near surface resistivity values taken from the profiles A, B, C ... Vertical scale not exaggerated!

this ice is considered as a perennial snowbank). Two conclusions can be drawn from these observations: a) as the rock glacier in the "Combe de Prafleuri" the rock glacier within the "Chessi" exists in the belt of discontinuous permafrost as was suggested for all rock glaciers of the Alps (Haeberli 1975 and 1976) and b) dead glacial ice is very rare within morainic material from the last century (or even centuries), if no permafrost conditions are present.

Another interesting feature is the asymmetry of the valley cross section. Again — as in the case of the "Combe de Prafleuri" — the greatest bedrock depth is found in the southernmost part of the valley. In this part of the valley the influence of holocene glacier activity certainly was very small, whereas it certainly was much stronger in the central and northern part of the valley marked by the neoglacial morainic system. The detection of the strong overdeepening in the region of the rock glacier had the consequence that the gallery of the Grande Dixence was constructed somewhat more to the east as formerly planned.

DISCUSSION

Three points of special interest shall be discussed in the following:

a) the resistivity and the characteristics of the near surface layer of active rock glaciers,
b) the internal layering of active rock glaciers as indicated by the vertical change of electrical resistivity and
c) the problems of accurate bedrock depth determinations and the observed bedrock relief.

a) The presence of ice within active rock glaciers is known since the early time of rock glacier research (Brown 1925, Capps 1910) and more recently was clearly confirmed by a great number of excavations (e. g. Johnson 1967, Lliboutry 1965, Potter 1972), seismic refraction measurements (Barsch 1973, Haeberli and Patzelt, in preparation) and indirect heat flow measurements (Haeberli 1973, Haeberli and Patzelt, in preparation). On the other hand the discussion, weather this ice is of glacial origin or not (interstitial ice = permafrost) is still open (cf. the discussion given by Smith 1973). The permafrost (interstitial ice) hypothesis represented e. g. by Wahrhaftig and Cox (1959) or Barsch (1969, 1971) is mostly based on the simple fact that the major part of all rock glaciers never were in contact with a real glacier and rock glaciers, which are still active today, are found within regions, where permafrost obviously occurs (Haeberli 1975). Based more on airfoto interpretation and morphogenetic speculations than on reliable field observations the glacial ice hypothesis is supported until today (Guiter 1972, Klaer 1974, Whalley 1974), whereas certain authors believe in both possibilities (Outcalt and Benedict 1965, Washburn 1973, White 1976).

Electrical resistivity determinations in temperate as well as in slightly cold alpine glacier ice were summarised by Röthlisberger (1967a) and Röthlisberger and Vögtli (1967). The observed resistivity was always higher than 10 MΩm (10^7 Ωm). Only in very cold arctic ice sheet ice, which is not very likely to be present in this context, much lower resistivity values were found by various authors (e. g. Hochstein 1967, Meyer and Röthlisberger 1962, Thyssen 1976, Vögtli 1967). Laboratory measurements, although perhaps somewhat difficult to interprete, showed that the resistivity of snowbank ice embedded within permafrost is on the same order as the resistivity of alpine glacier ice (Østrem 1967, cf. Østrem 1964, p. 330). Resistivity values in perennially frozen ground of the Arctic are much lower. (Barnes 1965, McKay 1969).

The recent compilation given by Hoekstra and McNeill (1973) shows that the resistivity of saturated sand and gravel is on the order of about 1,000 Ωm in the unfrozen and about 10,000 to 100,000 Ωm in the frozen state. About the same result was found in the Alps at the Upper Theodulgletscher (near Zermatt), where the resistivity of a perennially frozen neoglacial moraine (sand and gravel cemented by interstitial ice and ice lenses) amounted to 150,000 Ωm. The presence of permafrost and the absence of buried glacier ice in this case was prooven by core drillings and excavations down to the bedrock (Geotest AG, Zollikofen, unpublished data, cf. fig. 3 in Haeberli, in press). Thus, the resistivity of glacier ice (or perennial snowbank ice) and of perennially frozen ground (interstitial ice) respectively differs at least by a factor of 30 to 100. If this difference is real — there is no argument against it today —, resistivity determinations are very appropriate to discriminate between glacier (snowbank) ice and frozen ground respectively and may solve the oldest and most intriguing problem of rock glacier research.

If one takes the rock glacier within the "Combe de Prafleuri" — a rock glacier, which most probably never was in contact with a real glacier throughout the whole holocene time period — as representative for the permafrost (interstitial ice) hypothesis and the rock glacier within the "Chessi" — a rock glacier, which certainly was and probably still is in contact with a real glacier or glacieret — as representative for the glacier ice hypothesis respectively, the following conclusion can be made: perennially frozen ground (interstitial ice or perhaps also ice lenses/segregated ice) is present in both cases, but buried glacier (snowbank) ice only in one case and even in this case in an area of limited extent and only at the rock glacier surface. Thus, one can postulate that the presence of perennially frozen ground is very probably the "conditio, sine qua non" for the existence of an active rock glacier, while the presence of buried glacier or snowbank ice certainly is not! The data of the present study strongly suggest that even in the case of the rock glacier within the "Chessi" the presence of glacier ice is somewhat accidental (simple coincidence of the topographic position?) and the glacial contribution to the rock glacier formation is very probably restricted to the addition of some ice and morainic material on to the rock glacier surface.

b) In the "Combe de Prafleuri" as well as in the "Chessi", a two layer structure can be observed within the active rock glaciers. As a first approximation, one may take the lower boundary of the near-surface high-resistivity layer as the lower boundary of the vertical permafrost extent. In a second step, however, two facts must be taken into account:

- the marked resistivity rise in frozen sand and gravel only takes place at about −1°C (Hoekstra and McNeill 1973: fig. 1, p. 518). As the permafrost temperature rises with increasing depth as a function of the geothermal heat flow, some low resistivity permafrost at a temperature near 0°C may not be seen from electrical resistivity soundings.
- the resistivity of frozen materials seems to increase with decreasing temperature (Hoekstra and McNeill 1973: fig. 1, p. 518). Again, as the permafrost temperature increases with depth, one should expect the resistivity to decrease with depth. As in the case of permafrost between glacier ice with an extremely high resistivity and low-resistivity unfrozen-ground, a medium resistivity permafrost of say 20,000 Ωm would probably be masked by the great resistivity difference between high-resistivity cold permafrost of say 200,000 Ωm at the surface and low-resistivity unfrozen-ground at depth.

The correct permafrost thickness can be calculated, if the mean permafrost temperature and the temperature gradient (or the heat conductivity of the frozen material and the geothermal heat flow) are known (Gold and Lachenbruch 1973, Shumskii 1964). Not one of these parameters exactly is known for any rock glacier untill today. Therefore, only very rough estimates of the permafrost thickness and the mean permafrost temperature can be made from the electrical data of the present study by estimating the temperature gradient. If one assumes that (1) the temperature gradient in ice-rock mixtures is somewhat steeper than the average value of about 0.03 °C/m, because of the low heat conductivity of ice, that (2) the temperature gradient is reduced by about a factor of 2 following the topographic reduction of the geothermal heat flow at the site of the studied rock glaciers near the mountain crests (cf. Röthlisberger 1967 b, p. 89) and (3) no paleoclimatic corrections must be applied to the geothermal heat flow, then a value of about 0.02 to 0.03 °C/m may not be too far from reality. With this temperature gradient and a critical temperature of -1 °C for the marked rise of the resistivity in frozen ground the permafrost temperature (mean annual surface temperature) of the rock glaciers can be estimated to be on the order of about -1 to -2 °C for the rock glacier in the "Combe de Prafleuri" and for the lower part of the rock glacier in the "Chessi" and about -2 °C or even somewhat colder for the upper part of this latter rock glacier. These temperature estimates are in good agreement with the values measured in permafrost and even in glacier ice at the same altitude within the Alps (Haeberli 1976). Based on the same assumptions as discussed above some 30 to 50 m of permafrost thickness are added (dashed line) to the electrically determined permafrost thickness in fig. 7 and fig. 11 and the corrected permafrost thickness then becomes about 50 m for the rock glacier in the "Combe de Prafleuri" and for the lower part of the rock glacier in the "Chessi", while it becomes 100 m or even more in the upper part of this latter rock glacier. Certainly, assumption (3) is not too far from reality for temperature estimates in the uppermost permafrost layers, but on the other hand it is not correct at all for estimates of the real permafrost thickness. Better estimates of the real permafrost thickness should be made on the basis of (colder) surface temperatures during the last centuries: the latent heat of fusion of ice causes a long response time of the permafrost thickness to changed surface temperature conditions! Therefore, not only the electrically determined but also the "corrected" permafrost thickness values (using a stady state condition of the geothermal heat flow) only give minimum values for the real permafrost thickness. The maximum possible permafrost thickness could be on the order of 100 and 200 m respectively (cf. Patzelt 1977, p. 93).

In both rock glaciers of the present study the bedrock depth is very probably greater than the permafrost thickness. The two rock glaciers are not frozen to the bedrock in all places, but only where the steep bedrock side walls in the head region are obviously in direct contact with the rock glacier permafrost. The resistivity values observed between the permafrost and the bedrock are at the upper limit of the resistivity range known from groundwater. On the other hand, the possibility that there may exist a thick subpermafrost groundwater layer cannot be strictly excluded. The presence of such a groundwater layer could perhaps enable the rock glaciers to slide. This interpretation would not be in contradiction with the seasonal velocity fluctuations observed at the surface of other rock glaciers in the Alps (Barsch and Hell 1976). Another point open to discussion is, weather the unfrozen (and possibly water saturated) sediments at depth take part in the rock glacier flow (cf. the suggestions of Lliboutry 1977 for the Glaciar Hatunraju). Both, basal sliding as well as the movement of unfrozen debris at depth, would reduce the effective contribution of permafrost creep, which already seems to be smaller than the creep of temperate glacier ice by about an order of magnitude.

c) The bedrock depth values of nearly 200 m below the rock glacier in the "Combe de Prafleuri" and more than 300 m below the rock glacier in the "Chessi" seem to be quite astonishing. One may state that bedrock depth determinations by means of electrical resistivity soundings are somewhat ambiguous, but on the other hand, in the complicated case of rock glaciers, this is true for other geophysical soundings too. Strong overdeepenings are hardley detected by seismic refraction or wide-angle reflection as was done at the rock glaciers Macun and Murtel (Barsch 1971, Barsch and Hell 1976). Furthermore, the possible presence of an unfrozen layer with low seismic velocity at depth ($V_1 > V_2$!) makes the interpretation of refraction seismic profiles difficult. For accurate gravimetric measurements and interpretations the density (ice/water content) of the frozen and unfrozen material should be known. Thus, only deep drilling combined if possible with deep permafrost temperature measurements could lead to a better interpretation of the geophysical soundings. If one compares the bedrock depth values of the present study with the bedrock depth values obtained by seismic refraction and wide-angle reflection (alltogether are somewhat ambiguous as discussed above), the electrically determined values seem to be far too high. On the other hand the rock glaciers Macun (bedrock depth about 80 m, Barsch 1971) and Murtel (bedrock depth about 50 to 60 m, Barsch and Hell 1976) are much smaller than the rock glaciers described in the present study. If one plots the mean bedrock depth against the rock glacier length for these four cases (the only known bedrock determinations on rock glaciers!), a nearly linear relation results. Certainly, this phenomenon could be accidental and it is somewhat difficult to understand the processes which may lead to such a relation.

From the experience in the "Combe de Prafleuri", where the bedrock depth determinations could be controlled by tunneling (gallery Prafleuri - Blava, fig. 7) and excavations, it is estimated that the electrically determined bedrock depth values are correct within at least $\pm 20\%$. The fact that even much steeper bedrock wall inclination than observed in the present study is known from various glaciated or formerly glaciated valleys (e. g. Bindschadler et al. 1977, many examples in the Alps) and strong overdeepening is a quite common feature in glacial troughs (e. g. Thyssen and Ahmad 1969) also suggests that at least the order of magnitude of the bedrock depth determinations reported in this study is correct. Even the asymmetry of the valley bedrock relief seems not to be uncommon (e. g. Aric and Steinhauser 1977). This asymmetry of the bedrock topography may be the primary cause for the heavy debris accumulation in the area of the rock glaciers, but the rock glacier permafrost may also have had a conservative function with respect to the processes of debris erosion during late glacial and holocene time.

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