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Guide for the excursion

MOUNTAIN PERMAFROST AND SLOPE STABILITY IN THE PERIGLACIAL BELT OF THE ALPS

Meeting Ipa 1997 in Italy, Zürich-Bormio, August 22-27

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PREFACE

(F. Dramis & W. Haeberli)

Research on displacements of slope materials under periglacial conditions has a long tradition. Rock falls, debris flows, solifluction and permafrost creep have, indeed, been observed and studied for decades. Renewed attention and remarkable progress can now be observed, however, for three main reasons:
1. advanced technologies including core drilling, borehole measurements, geophysical soundings, continuous datalogging under harsh field conditions, detailed laboratory experiments and sophisticated numerical modelling help building up a much more precise process understanding than hitherto possible;
2. full treatment of all involved scales in time and space, including impacts on and effects from thick, ice-bearing permafrost layers help recognizing processes and phenomena of rock destruction and slope stability not investigated before and relating to important environmental and hazard aspects in cold-climate mountain topography;
3. observed glacier shrinkage and permafrost warming/degradation at secular to decadal time scales leads to the recognition by a wide public of the growth tendency, recent activation and sensitivity of periglacial mountain belts as related to effects of atmospheric warming.
Slope stability aspects involve time and depth scales ranging from daily cycles to millennia and from centimeters to hundreds of meters (fig. 1).

At daily and centimeter scales, surficial frost weathering by volume expansion and frost creep from pipkrake formation predominate. At the decimeter to meter scale, rock destruction, frost creep and gelifluction through ice segregation and subsequent thawing within seasonally frozen ground an permafrost active layers take place. At the scales of tens to hundreds of meters and decades to millennia, permafrost creep (rock glacier flow) and destabilization of large rock masses result. The field trip M4 leading from Zurich to the International Geomorphology Congress at Bologna illustrates characteristic examples of such aspects and documents the intensive research activity carried out by international and interdisciplinary research groups in the European Alps.

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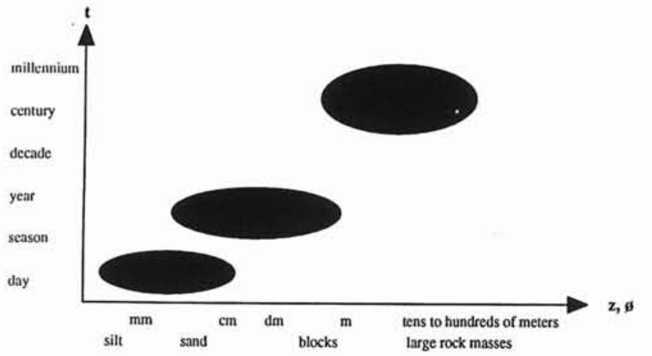

FIG. 1 - Time and depth scales of slope stability aspects.

One day of the field trip is devoted to a workshop on mapping, monitoring and modelling of mountain permafrost and related phenomena. Mapping, monitoring and modelling are fundamentally important elements of scientific research on the above-mentionned processes, indicating sensitive areas and situations in complex topography, enabling improved understanding of the multiple interactions and feed-back mechanisms related to freezing and thawing of ground materials on slopes, and establishing the data base for estimating impacts from potential future climate change scenarios on living conditions in cold mountain areas.

The field trip and workshop constitutes a central part of the activities undertaken by the Working Group on Mountain Permafrost of the International Permafrost Association and the Commission on Periglacial Environments of the International Geographical Union. Special thanks are due to M. Guglielmin, A. Kääb and D. Vonder Mühll for organizing the logistics and preparing the field guide/abstract volume. The Swiss Academy of Sciences (Sanw) helped with funding participation of international colleagues and young researchers.

Friday, 22. August 1997

PERMAFROST IN A SCREE SLOPE AT FLÜELAPASS (W. Haeberli)

In October 1972, twenty-five years ago, seismic refraction soundings had been carried out in the large scree slope at Schottensee, Flüelapass, as part of the first sytematic study on the distribution patterns of permafrost in the Alps (HAEBERLI, 1975). Already before, perennially frozen debris rich in ice had been encountered at the foot of the slope when excavations were made for highway construction. The scree slope is characterized by long-lasting to perennial avalanche snow at its base and by well developed solifluction lobes on the partly vegetation-covered upper part. Investigation of this easily accessible scree slope soon became a key to understanding the special aspects of permafrost formation on steep mountain slopes (fig. 2).

Four seismic profiles were measured: three profiles parallel to the lake shore at vertical distances of 0, 30 and 80

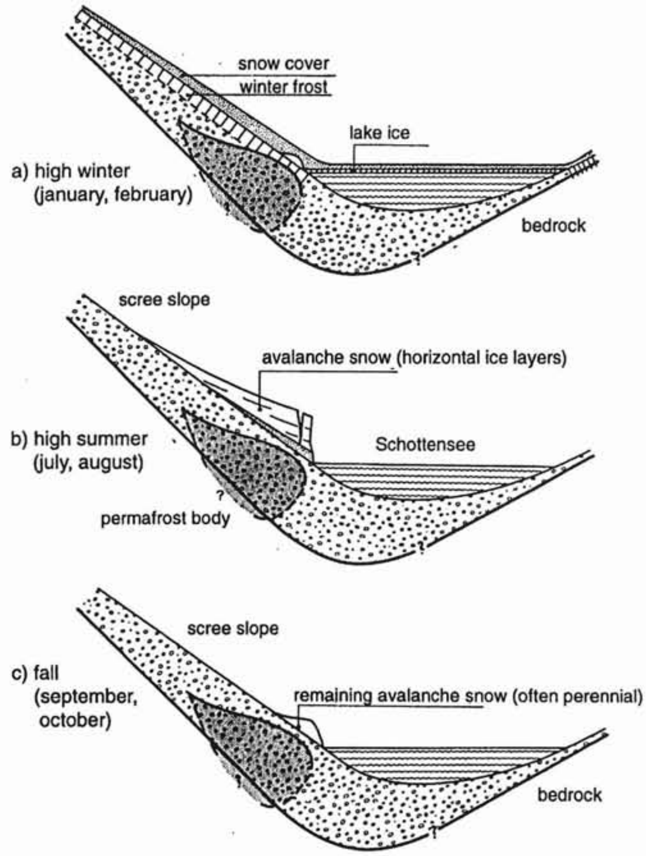

FIG. 2 - Permafrost in a scree slope at Flüelapass.

meters above lake level, and one profile from the lake shore upslope. The profile along the lake shore indicated maximum bedrock depth in the central part of the scree slope to be around 30 meters (cf. similar values determined later in other scree slopes by HAEBERLI & *alii*, 1990) and water-saturated debris beneath lake level to have P-wave velocities of about 2,200 m/s. In the profiles above lake level, permafrost with characteristic P-wave velocities of 3,300 to 3,700 m/s is present at all geophones but with active layer thickness increasing upslope from less than 4 m in the lower profile to maximum values of around 10 m in the upper profile. As revealed in the upslope profile, permafrost existence was limited to about the lower 100 meters of the scree slope with the upper scree slope being unfrozen. Interpretation of appearant bedrock velocities using a wavefront approach indicated that permafrost thickness increases downslope parallel to the decreasing active layer thickness. The possibility exists that permafrost reaches into bedrock in the lowermost part of the scree slope near the lake shore. Later radio-echo soundings at this site using a low-frequency monopulse device were described by KING & *alii* (1987).

The early investigation of the scree slope at Schottensee/Flüelapass helped understanding basic aspects of the specific microclimatic conditions on such Alpine slopes. The upper slope becomes snow-free because of snow evacuation by (wet) avalanches early in summer. The foot of the slope, on the other hand, remains covered by avalanche snow throughout the warm season and is protected against the influences from solar radiation and positive air temperatures. Sites below large avalanche slopes indeed belong to the coldest sites known in the Alps. This important microclimatic effect was included into the «rules of thumb» for predicting permafrost occurrences in the Alps (KELLER, 1992).

Saturday, 23. August 1997

PROJECTS ON PONTRESINA-SCHAFBERG
(D. Vonder Mühll)

In the slope immediately above Pontresina, permafrost and rock glaciers occurs. As part of a government research programme on avalanche protection in mountain permafrost, extensive geophysical soundings were carried out in 1989 (GEOTEST, 1990). In 1990, two boreholes (1/1990: 67 m and 2/1990: 37 m in depth) were drilled. Bedrock depth was estimated by the drilling crew at 16 m (1/1990) and 30 m (2/1990) below surface (VONDER MÜHLL & HOLUB, 1992). Additional geophysical surveys were performed 1992 in the vicinity of the drill sites (VONDER MÜHLL, 1993). Once a year, the long-term measurements in the borehole variables (temperature, horizontal and vertical deformation) are taken. For photogrammetric investigations using analytical aerophotogrammetry, photos from 1971, 1991 and 1994 were available and analysed (KÄÄB, 1996, see below). Within the Project CH-2000, which started in 1996, the interactions of avalanche defence structures and permafrost is studied with shallow boreholes (6 to 8 m; particularly 20 m.

Geophysics

The results of the seismic refraction and D.C. resistivity soundings can be summarized as follows: a continuos permafrost layer with P-wave velocities of typically 3000-4000 m/s and specific resistivities below 100 kΩm - in places even below 10 kΩm - occurs at shallow depth beneath the blocky rock glacier surface. Active layer thickness within the steep rock glacier front, where construction of avalanche protection structures were planned, increases to 5-10 m. This indicates that permafrost is presently degrading and that geotechnical properties of the steeply inclined loose sediments are changing accordingly. The same locality represents a potential starting zone for large-scale debris flows (VAW, 1990, 1991,1992; VONDER MÜHLL, 1993).

Permafrost Borehole 1/1990 and 2/1990

At Pontresina-Schafberg, no cores were taken during the drilling. In order to fill the borehole with water for borehole logging, great care must be taken to carry out all logs before the drilling penetrates the permafrost base (VONDER MÜHLL, 1996). Drilling 2/1990 at about 2740 m a.s.l., roughly 50 m behind the rock glacier front, shows that permafrost with mean annual surface temperatures near -0.5 °C just reaches bedrock at 30 m depth. Borehole 1/1990, completed in the eastern part of the cirque, penetrated 16 m of frozen sediments and then bedrock down to the bottom of the hole at 67 m. Mean surface temperature is about –1.7 °C and permafrost reaches a depth of about 70 m. The most remarkable features are the occurrence of a layer between 12 m depth and bedrock at 16 m, which has an ice content of about 80% by volume. This is indicated by the gamma and neutron logs, the small variability of P-wave velocities (about 3500-4500 m/s) within the frozen sediments (sonic log) and the decreasing electrical resistivity with increasing electrical resistivity and increasing depth.

Measurements of borehole temperature (fig. 3; STUCKI, 1995, VONDER MÜHLL & *alii*, in prep.) as well as of hori-

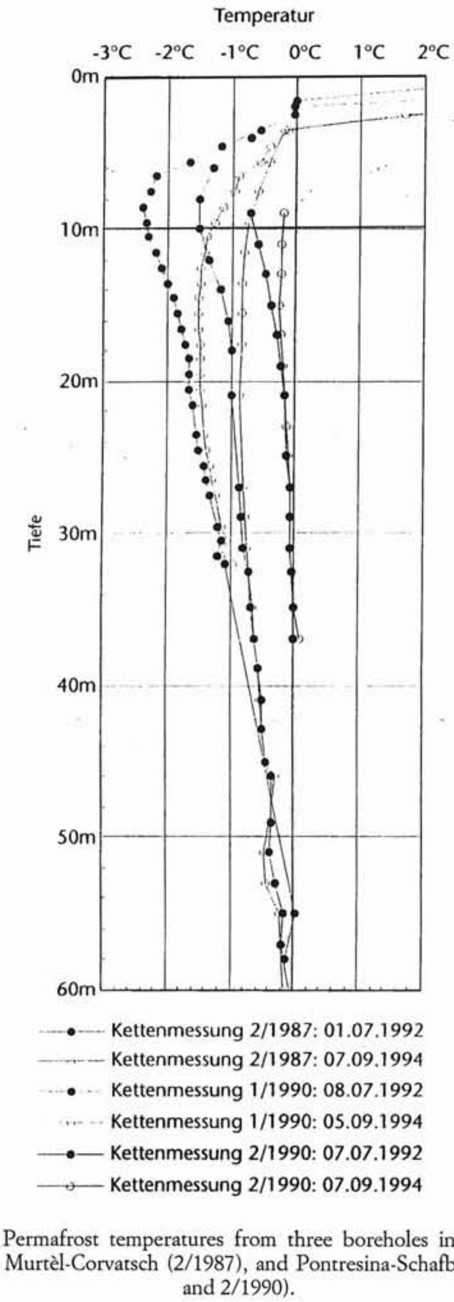

FIG. 3 - Permafrost temperatures from three boreholes in the Upper Engadin: Murtèl-Corvatsch (2/1987), and Pontresina-Schafberg (1/1990 and 2/1990).

zontal and vertical borehole deformation (fig. 4; WAGNER, 1996, HOELZLE & *alii*, in prep.) are part of long-term monitoring programmes on the evolution of Alpine permafrost. Temperature comparisons between Murtèl-Corvatsch and Pontresina Schafberg, which are about 10 km apart, show that the behaviour of the time series is parallel although the temperatures are quite different. Hence, the measured borehole temperatures represent at least a regional signal. For the deformation measurements of the boreholes, full contact with the formation was established only three years after the drilling. The movement values of borehole 1/1990 are about 4 cm/a (horizontal) and 0.7 cm/a (vertical settlement) and take place mainly at a depth between 11 and 15 m. In borehole 2/1990 the quality of the measuremets is not as good as in borehole 1/1990. Nevertheless, the horizontal displacements amount to approximately 1 cm/a, the vertical settlement is in the order of 0.2 cm/a). A distinct change in deformation behaviour is observed at depths of 20 m and 32 m.

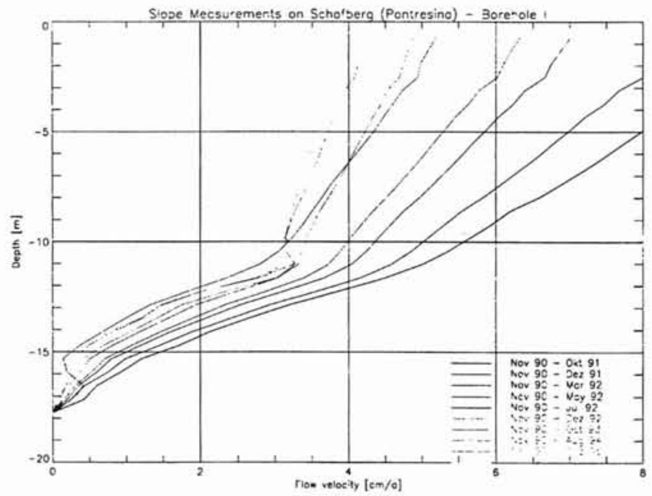

FIG. 4 - Flow velocity in borehole 1/1990 (Pontresina-Schafberg) between 1991 and 1996.

Photogrammetric analysis
(A. Kääb)

In addition to the yearly measurements of horizontal and vertical deformation at the boreholes 1 and 2, photogrammetric investigations were performed at Pontresina Schafberg using techniques of analytical aerophotogrammetry. Changes in surface elevation over the period from 1971 to 1991 have been determined by comparing two digital terrain models (Dtm) having a regular grid width of 25 m. Most measured changes are not significant, because they are lower than the accuracy of the elevation measurements (±2 cm/a). Larger surface lifting observed in the northern part of the site is assumed to be caused by scree accumulation from the rock wall delimiting the cirque in northern direction. Surface lowering in the northwest of the site below borehole 2 and just above the steep channel running to Pontresina may indicate surface erosion.

Surface velocities between 1971 and 1991 were determined by a special photogrammetric technique simultaneously comparing aerial photographs taken from different places as well as taken at different times (KÄÄB, 1996) (fig. 5). Measured surface movements of up to 10 cm/a can, in places, only be interpreted qualitatively, because their error range is around 2-3 cm/a and, therefore, some results are marginally significant. However, a comparison between the photogrammetric results and the measurements in the boreholes (4cm/a at borehole 1, ca. 1cm/a at borehole 2) shows quite good agreement. The coherent velocity field of the two rock glaciers points to a continuous presence of ice-saturated ground materials. Calculation of flow lines assuming a flow field which remains constant in time leads to an age estimate for the rock glacier surface of up to several thousand years.

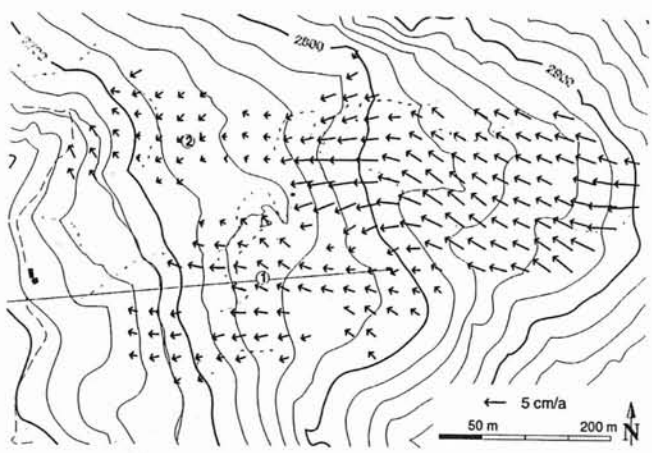

FIG. 5 - Average annual surface velocities 1971-1991 at Schafberg.

PERMAFROST PROJECT CH-2000: AVALANCHE DEFENCE STRUCTURES IN PERMAFROST TERRAIN (M. Phillips)

Experimental site Muot da Barba Peider, Pontresina, GR

The NW flank of Muot da Barba Peider was selected in spring 1996 as a Sfisar permafrost study site (fig. 6). The project aims at studying the interactions between avalanche defence structures and permafrost. The site was chosen, because it is typical of slopes requiring avalanche defence structures in permafrost terrain. The presence of permafrost in the slope was verified by Bts measurements and drilling in 1996. The slope is relatively steep (average 35°), oriented NW, and is located between 2870 and 2980 m. It was also chosen because experiments which could result in terrain instability do not endanger any settlements below. In addition, the rock glacier located at the foot of the slope (Fuora da l'Amd Ursina) has been studied extensively and data acquired there may help to understand the physical conditions reigning above.

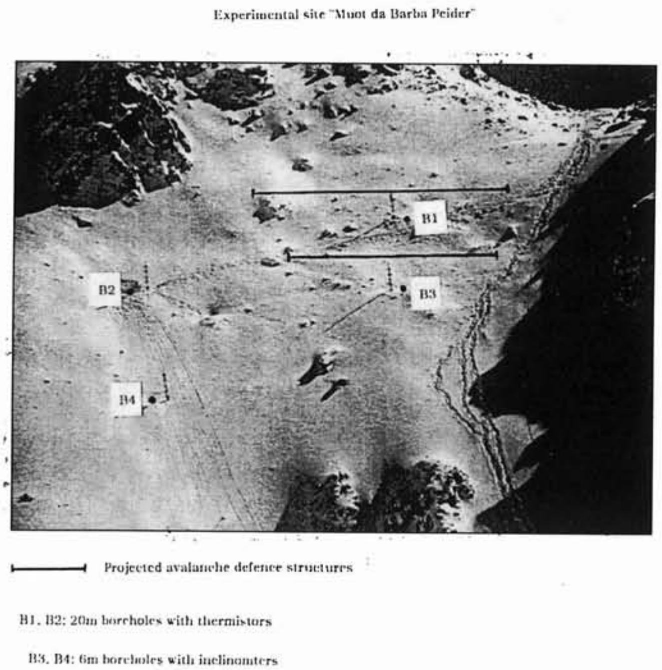

FIG. 6 - Location of the study site Muot da Barba Peider.

Geomorphology and geology

The slope is covered by a 2-3 m thick layer of loose scree (10-50 cm). The scree increases in size and quantity downslope and drilling reveals that sediment size diminishes vertically (see profile) due to the displacement of fine material by percolating water percolation. At around 2 m depth, there is an intermediate layer of frozen silt and clay. Ice content is approximately 10% in both this layer and the underlying bedrock. The latter is faulted Languard gneiss, highly subject to frost shattering when directly exposed. Rockfall due to this phenomenon is frequent on neighbouring rock faces in summer.

Snow cover

Due to the orientation of the slope, snow patches remain until early summer (June). Early snow falls in autumn often melt much later than on surrounding slopes or not at all, due to reduced solar radiation and cold ground temperatures. In winter, snow is strongly redistributed by wind, as the slope is located just beneath a ridge and is on the flanks of a cirque into which winds are funnelled from various directions. Natural slab avalanches (i.e. not triggered artificially) occur quite frequently and the fracture zone is near the top of the slope, where snow accumulations are most important. In spring, wet snow avalanches can also occur. Avalanche occurrence and snow distribution (temporal and spatial) are recorded 3 times weekly by an automatic camera located on the opposite slope.

Temperature and slope stability measurements

Four graduated snow stakes are visible near the top of the slope. Each stake marks the location of a borehole. The upper two are 20 m deep and were equipped with thermistors in autumn 1996. The lower two are 6m deep and contain slope inclinometer pipes. Ground surface temperatures are measured in the vicinity of these boreholes with univeral temperature loggers. Two other smaller boreholes (5 m) are also equipped with universal temperature loggers.

Avalanche defence structures will be erected in Summer 1997. Two systems, nets and bridges, are used for both rows. Different types of anchors are tested and equipped with thermistors. Boreholes drilled parallel to anchors are also equipped with thermistors. The temperatures of the steel components above the ground are continuously monitored.

These measurements contribute toward the development of guidelines for the construction of avalanche defence structures in permafrost and determination of their effects on the ground thermal regime and on slope stability.

MURAGL ROCK GLACIER
(D. Vonder Mühll)

The large rock glaciers in the Val Muragl have been described and investigated since the first half of the twentieth century (SALOMON, 1929; DOMARADZKI, 1951).

The Muragl I rock glacier is approximately 750 m long and 150 m wide. It has pronounced lobes, furrows, and ridges caused by compressive flow in its lower part. The lobes show a variety of pioneer plant associations with lichens and, hence, are most likely to reflect several generations of compression. The rock glacier flows from a small cirque into the main valley of Muragl. The flow direction turns sharply from NNE to NW. Muragl I is one of five rock glaciers investigated within a recently established permafrost monitoring program (HAEBERLI & *alii*, 1993).

BARSCH (1973) investigated the frontal part of rock glacier Muragl I by the refraction seismic method with a sledge hammer. At the front, the active layer thickness of up to 20 m indicated that the permafrost in the rock glacier is inactive. However, Barsch supposed that 80 m behind the front, the active layer thickness is about 4 m. In the frontal part, BARSCH & HELL (1975) measured a horizontal velocity of 21 cm/a and a vertical displacement of 10 cm/a.

Studies of the 1990s involve mapping of permafrost distribution, using the Bts method (HAEBERLI, 1992), seismic refraction, geoelectrical resistivity and Vlf resistivity soundings (VONDER MÜHLL, 1993), moreover, comparison of solar radiation patterns and permafrost occurrence (KELLER, 1994; HOELZLE, 1994) and long-term photogrammetric monitoring (VONDER MÜHLL & SCHMID, 1993; KÄÄB & GUDMUNDSSON, in prep.).

Seismic refraction surveys revealed a typical three-layer case: the active layer (2 m to 7 m thick) with a velocity of about 1000 m/s or less is underlain by a near-surface refractor with 3200 m/s to 3700 m/s representing the permafrost. With long profiles, bedrock (~5000 m/s) was reached at depths of about 50 m. D.C. resistivity soundings gave characteristic specific resistivities for the active layer of about 20 kΩm and 100 - 2000 kΩm for the ice-bearing,

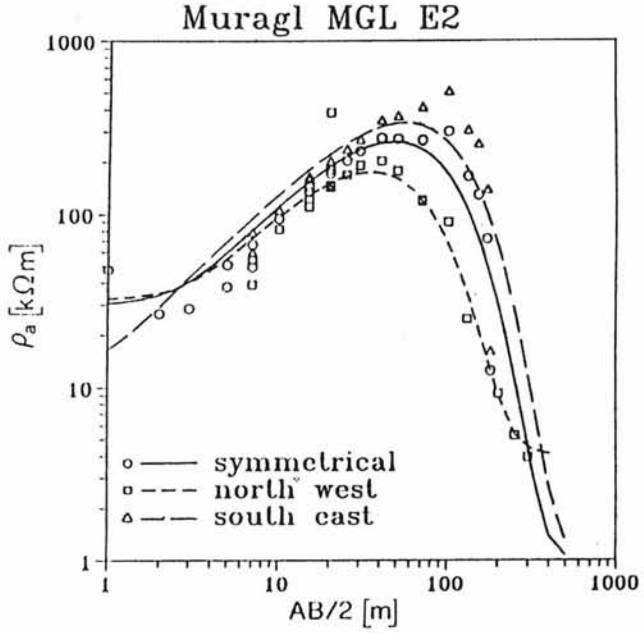

FIG. 7 - Sounding curve with Schlumberger (symmetrical) and Hummel array (north, south) on rock glacier Muragl with values which are typical for mountain permafrost (VONDER MÜHLL & SCHMID 1993).

highly resistive permafrost (fig. 7). The thickness of this high-resistivity layer may be up to 100 m and decreases from the upper part of the rock glacier to its lower part, probably reflecting effects from surface warming and changing ice composition along flowlines (HAEBERLI & VONDER MÜHLL, 1996). First results of the aerophotogrammetrical studies in the time period 1981/1985/1990 with an analytical plotter revealed that the permafrost body creeps with a velocity of 10-30 cm/a in a predominantly compressing mode. The overall flow direction changes drastically (about 90°) in the main valley (fig. 8). Variations in altitude

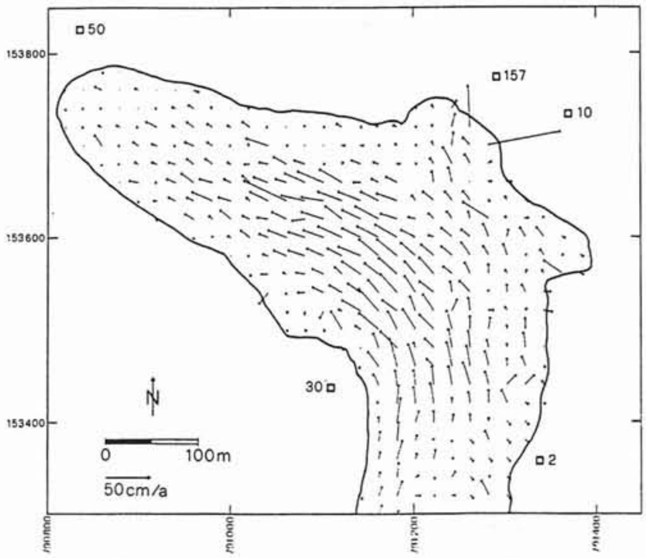

FIG. 8 - Flow velocity field from Muragl rock glacier (VONDER MÜHLL & SCHMID 1993).

and velocity between the two periods remain within measuring accuracy. The marginal and frontal parts of the rock glacier appear to slow down.

Calculations with the digital permafrost distribution program Permakart (KELLER, 1992) indicated that rock glacier Muragl I lies at the margin of the discontinuous permafrost area.

Permafrost occurrence around Piz Muragl as deduced from empirical rules of thumb and calculated with the program Permakart (KELLER, 1994) shows that permafrost distribution is strongly but not exclusively influenced by the pattern of incoming direct solar radiation. Active rock glaciers generally start from areas with probable permafrost occurrence and strongly reduced solar radiation, but their fronts in some cases reach sites with less favorable or even marginal permafrost conditions (HOELZLE, 1994).

Fluted moraines characterize the central part of the upper Muragl rock glacier, where no active permafrost exists. This shows that the small Muragl Glacier which extended onto the rock glacier surface during the Little Ice Age was partially warm-based and probably caused local degradation of the permafrost. The Muragl glacier forefield is presently being studied by Kneisel (KNEISEL & *alii*, 1997). In such cases of glacier-permafrost interactions, the thermal influence exerted by former (polythermal) glaciers on preexisting mountain permafrost and the supply of morainic material onto rock glacier surfaces may, in fact, be the essential processes.

Sunday, 24. August 1997

MURTEL - CORVATSCH
(D. Vonder Mühll)

The Corvatsch cable car leads from the belt of sporadic permafrost (below timberline) through the belt of discontinuous permafrost (middle station) to the lower limit of continuous permafrost near the summit station at 3'300 m a.s.l.. Mean annual air temperature at the Corvatsch summit station is –6.3°C and mean annual bedrock temperature is estimated at around –5°C. A permafrost thickness of more than 100 m can be expected. In the area around Piz Corvatsch, numerous glacial and periglacial phenomena such as fluted moraines and rock glaciers exist.

The active Murtèl-Corvatsch rock glacier has developed within a former cirque from perennially frozen, northwesterly exposed scree slopes between 2850 m a.s.l. and 2620 m a.s.l. It shows basically two flow types: extending flow in the upper part and compressing flow in the lower part of the rock glacier, which causes pronounced ogive-like transverse ridges (HAEBERLI & VONDER MÜHLL, 1996). The steep, approximately 20 m high front is largely free of vegetation and rests on permafrost-free granodiorite bedrock.

Borehole 2/1987

After several investigations (BARSCH, 1973, 1977 and KING & *alii*, 1987) a 60 m deep core drilling in the centre of rock glacier Murtèl-Corvatsch was performed in spring 1987 (HAEBERLI & *alii*, 1988). Borehole logging was done with some problems because the borehole could not be filled with water, hence only parts of the logs were run (VONDER MÜHLL & HOLUB, 1992). The borehole is equipped with a slope-indicator tube (horizontal deformation), magnetic rings (vertical deformation) and thermistors (temperature) for long-term monitoring. The main results are shown in fig. 9.

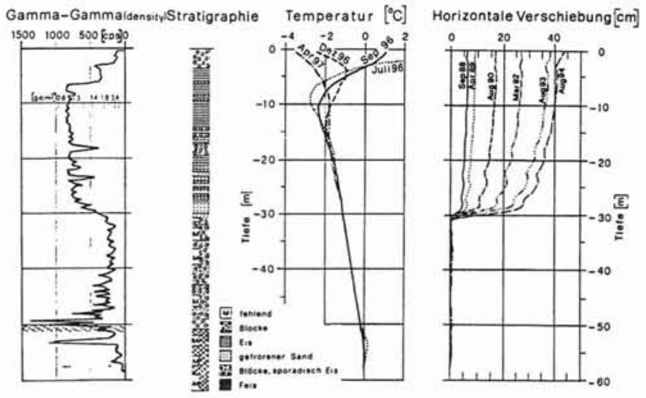

FIG. 9 - Main results of the drilling 2/1987 through rock glacier Murtèl-Corvatsch: density log, stratigraphy, temperture profile and deformation.

Structure: Beneath the permafrost table, the rock glacier is mainly built up by two layers: an upper one with an extremely high ice content (90-100% by volume) and a lower one consisting of coarse blocks with ice-filled pores (around 40% by volume) but almost completely without fine rock particles.

Deformation: The deformation of the borehole is determined by a slope-indicator tube (horizontal displacement) and magnetic rings (vertical displacement). According to the borehole deformation measurements and photogrammetrical analysis, the surface velocity at the drill site is about 6 cm/a. The main part (ca. 4 cm/a) takes place within the transition zone between the two main permafrost layers at a depth of 28 to 30 m with the upper (supersaturated) layer obviously undergoing steady-state creep and overriding the non-deforming (structured) lower layer (WAGNER, 1992, 1996). The slope indicator measurements could be performed until 1994. Since then, the slope device cannot pass the strongly bended borehole any more. Vertical deformation mainly occurs between 25 m and 30 m depth: the layer immediately above the shear horizon is shortened by roughly 1 cm/a. The resulting inclination of the flow vectors is about 10 to 15° and, hence, in the same range as the surface inclination of the compressing flow part of the rock glacier.

Temperature: The now ten year old temperature measurement series provides the following main results (VONDER MÜHLL & HAEBERLI, 1990; VONDER MÜHLL, 1993 & STUCKI, 1995):

– The mean annual surface at the surface is estimated at approximatly –2.5 to –3.0°C..

– Seasonal variations at a depth of 20 m are smaller than 0.1°C. There, a 'mid-term' temperature increase from –1.8 to –1.4°C occurred between 1991 and 1995.

– The uppermost 30 m rapidly warmed up by more than 1°C in 7 years at 11.6m depth (fig. 10). After 1994, within only two years, the temperature cooled down reaching again values similar to those in 1987. The variability of the observed permafrost temperatures is caused by several processes. 1) Long-lasting zero-curtains in autumn occur with early snow falls and reduce the length of the period with negative temperatures in the active layer. 2) Global radiation and air temperature influence temperatures mainly in summer time. 3) The snow cover thickness during winter time influencing the annual permafrost temperatures is another important factor: the regression between the snow cover thickness in November / December and the permafrost temperature near the permafrost table shows a high correlation. The strong cooling in 1994 - 1996 is due to the fact, that the Upper Engadin got very little snow and, consequently, the insulating snow cover was thin, enabling intensive cooling of the ground.

– Between 52 and 56 m depth, seasonal temperature variations around 0°C occurred every year and hint to an intra-permafrost groundwater flow (talik) which is active from June to September (VONDER MÜHLL, 1992). As a consequence this special boundary condition (–3°C at surface and 0°C at 52 m depth), a very high heat flow of about 150 mW/m^2 results.

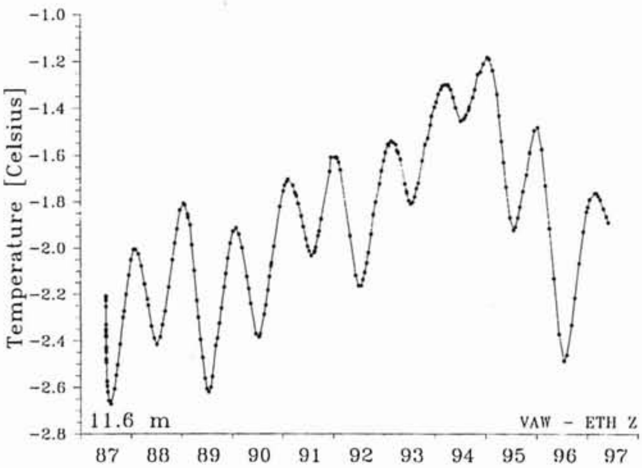

FIG. 10: Temperature at 11.6 m depth of borehole 2/1987 (Murtèl-Corvatsch).

Further investigations

Permafrost investigations in the Murtèl area continued in the years after the drilling:

– Several geophysical surveys (refraction seismics, D.C. resistivity soundings, ground penetrating radar Gpr, gravimetry and very low frequency resistivity Vlf-R) were applied to determine the spatial extent of the shear zone and bedrock depth of the rock glacier (VONDER MÜHLL, 1993; VONDER MÜHLL & KLINGELÉ, 1994). Results of the refraction seismics and of the gravimetry are shown in figure 11.

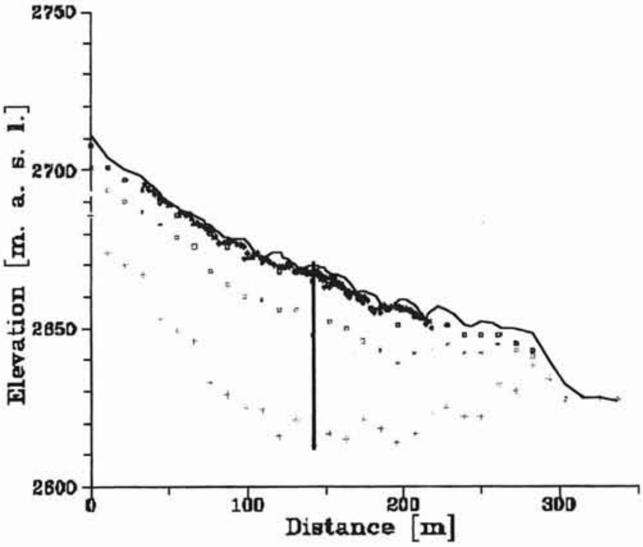

FIG. 11 - Permafrost table (refraction seismics), two layers with a ice content of 90 to 100% and the layer above the bedrock (+) in a cross section of Murtèl rock glacier with the borehole 2/1987 (VONDER MÜHLL & KLINGELÉ 1994).

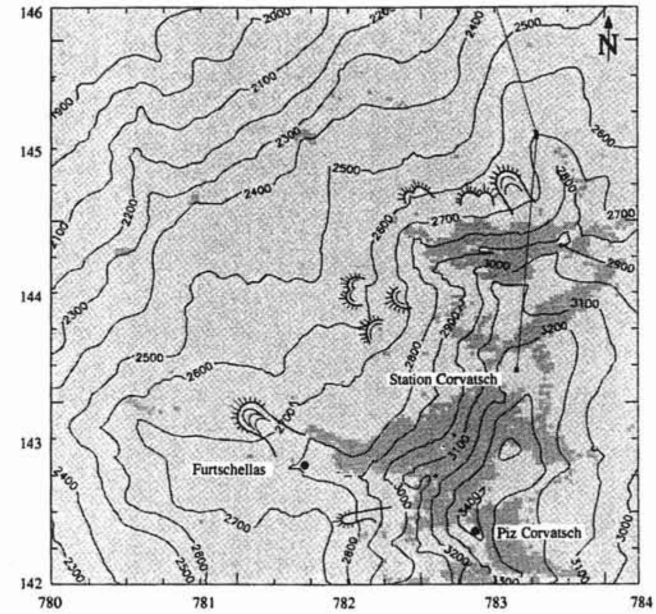

FIG. 12 - Supposed area with permafrost (light grey) and supposed permafrost distribution (dark grey) in the Corvatsch-Furtschellas area.

– The permafrost distribution in the area was determined using the Bottom Temperature of the winter Snow cover (Bts) as index (HOELZLE & *alii*, 1993). Continuous Bts measurements help to improve the understanding of the processes in mountain permafrost where the conditions change over short distances (see below).

– The spatial distribution of the permafrost can be modelled in two ways:

1. Empirical 'rules of thumb' (HAEBERLI, 1975) are implemented in a Geographical Information System (Gis) by KELLER (1992).

2. For each Bts-point the mean annual air temperature estimated from surrounding climate stations and the potential direct solar radiation (FUNK & HOELZLE, 1992) is calculated on the basis of a digital terrain model.A relation for the permafrost occurrence and these two climatic parameters results , which can be applied to other areas (fig. 12; HOELZLE, 1994, HOELZLE & HAEBERLI, 1995; in press).

– The characteristics of the snow cover are different in permafrost-free area and within the permafrost. Therefore, snow profiles were dug every two weeks during two winters (1989/90 and 1990/91) on the rock glacier (permafrost) and in the surrounding permafrost-free area. Figure 13 shows that different thermal regimes cause completely different behaviour of the snow cover in terms of thermal resistance, metamorphism, grain shape and hardness (KELLER & GUBLER, 1993; KELLER, 1994). Meltwater runoff from the snow cover in permafrost areas starts significantly later (roughly one month) than in permafrost-free areas. The pattern of areas becoming snow-free is observed by an automatic camera taking a picture every fourth day. With a Gis, the pattern is transformed onto a digital terrain model (BERNHARD, 1996). Moreover, funnels through the snow cover in the first part of the winter essentially influence the thermal regime at the surface of the rock glacier by directly coupling the air circulating in the coarse blocks of the active layer with the atmosphere (SUTTER, 1996; in prep.).

– Current photogrammetric analysis of large-scale aerial photographs from 1987 to 1996 shows surface changes in the order of few centimeters per year indicating predominance of subsidence and thaw-settlement of the whole

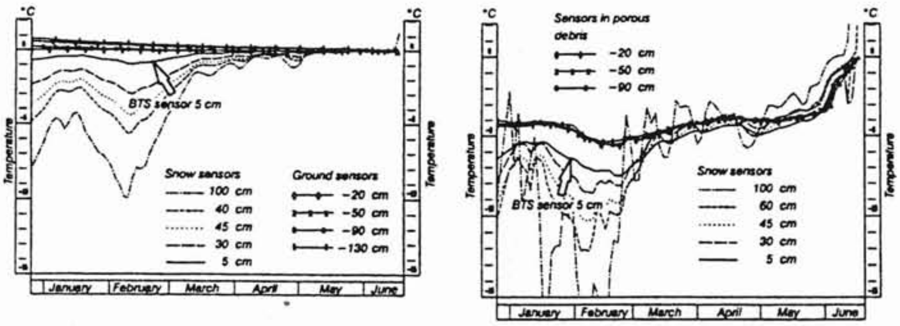

FIG. 13 - Snow and ground temperatures without (left) and in permafrost (right).

creeping permafrost. Comparison with earlier measurements completed by BARSCH & HELL (1975), who investigated the periods of 1932 - 1955 and 1955 - 1971, hints to decreasing surface kinematics.
– Measurements of the surface energy balance at the drill site were initiated in January 1997 (see below).

Increasing human activities in high mountain areas cause more and more contacts with the periglacial environment. Therefore, long-term investigation of Alpine permafrost conditions are essential in view to estimating potential reactions of this sensitive region to predicted climatic changes. The recently established permafrost monitoring network in the Swiss Alps (HAEBERLI & *alii*, 1993) is a first step to improve the basis for corresponding assessments.

ON THE USE OF MINIATURE TEMPERATURE DATALOGGERS FOR MAPPING AND MONITORING OF PERMAFROST IN HIGH MOUNTAIN AREAS. A TEST STUDY IN THE SWISS ALPS
(M. Hoelzle, M. Wegmann & B. Krummenacher)

Mapping and monitoring of mountain permafrost is an important tool to detect regional climate variability. The recently developed miniature temperature dataloggers (Mtd) open new possibilities of investigating such processes and of determining permafrost occurrences at low costs and even at remote sites with difficult access. The logger presently used in the Swiss Alps is based on a commercially available logger (circuit board) which is integrated in a newly developed logger case. This new tool is specially designed for rough field conditions.

Two case studies on and around rock glaciers in Switzerland were performed. One test site was in the Murtèl-Corvatsch area (Upper Engadin) and the other in the Furggentälti area (Bernes Alps). Here the results from the Murtèl-Corvatsch measurements are described.

Measurements were carried out in the Murtèl-Corvatsch area from the 7th of November 1994 until the 6th of May 1995 with an interval of 2.4 hours and have been recorded on four Mtds. On the rock glacier, two loggers were installed, one on top of a transverse ridge and one in a furrow. For comparison, one unit was installed in the permafrost-free zone in front of the rock glacier. All these thermistors were situated at the base of an undisturbed snow cover before the first snow had fallen. To show the effects of an artificially disturbed snow cover on the temperature development, the fourth set of readings was taken on a ski run at the base of a compacted snow cover. The two thermistors and where installed approximately 10 m apart from each other. The air temperature was measured at the borehole site on the rock glacier with a conventional data logger (CR10 - Campbell Scientific, Inc.). In addition, fresh snow heights from a standard measurement site (Swiss Federal Institute of Snow and Avalanche Research) at the Corvatsch cable car station where used. This station is situated at a distance of some 400 m from the rock glacier. In early winter (November to mid January) the temperatures are decreasing. Thereafter, a small increase until mid-February is observed. During the following two months, before the snowmelt starts, the BTS remains nearly constant.

The two sites on the rock glacier Murtèl were cooled differently between November and January. In comparison to the sinusoidal decrease in the furrow, the temperatures on the ridge decreases more or less continuously. The minimum in the furrow of –12°C (January 10th) is 3.5°C cooler and one month earlier than the lowest temperature on the ridge. After reaching a minimum, both curves behave in an almost identical way and reach a more or less constant temperature until the beginning of April for site and end of April for site. While the summerly temperature raise in the furrow is very abrupt, the temperature on the ridge increases one month earlier in a slow, continuous way. On May 16th the data loggers in the Murtèl area were collected. Thermistor on the ridge was covered by a 0.6 m thick snow cover, thermistor in the furrow was buried by a 3 m thick snow cover. Several ice lenses up to 10 cm thickness could be detected at site in the furrow.

With the continuous measurements, the assumptions of the conventional Bts method are verified. Important boundary conditions for Bts measurements are a sufficently thick, undisturbed snow cover and the adequate measurement time. In autumn before the snow cover is well developed, air circulation is still possible in the coarse active layer of the rock glacier and heat conduction through the thin snow cover is facilitated. At the end of the winter, melt water percolation disturbs the equilibrium Bts. At the base of an artificially compacted snow cover the temperatures were influenced throughout the whole winter by atmospheric variations.

SURFACE ENERGY BALANCE MEASUREMENTS AND SPATIAL NUMERICAL MODELLING OF DISCONTINOUS MOUNTAIN PERMAFROST
(M. Hoelzle, C. Mittaz & W. Haeberli)

For modelling of discontinuous permafrost distribution patterns in mid-latitude mountains, knowledge of energy exchange processes at the atmosphere/lithosphere boundary is of fundamental importance. A number of excellent energy balance investigations in the Arctic are available but their climatic and site-specific components are often quite different from mid-latitude high-altitude permafrost regions. Only a few studies are available in areas with mountain permafrost, such as the radiation measurements of HAPPOLT & SCHROTT (1992), SCHROTT (1991, 1992), LE DREW & WELLER (1978) and GAVRILOVA (1988).

Detailed energy balance measurements and spatial numerical modelling will be undertaken in Switzerland by the University of Zurich. Knowledge about mass and heat exchange processes which determine the distribution patterns and characteristics of high-mountain permafrost is still severely limited. Measurements concerning near-surface energy exchange processes are, therefore, initialized at one selected high-mountain site on rock glacier

Murtèl-Corvatsch. Results of these measurements will allow improved modelling and simulation of permafrost distribution patterns in different Alpine regions and for various climatic scenarios. The objectives are (1) to measure near-surface energy exchange processes at selected high-mountain permafrost sites in the Swiss Alps, (2) to improved the process understanding with respect to permafrost formation, existence and disappearance, and (3) to further develop modelling and simulation of permafrost distribution patterns in different mountain regions and for various climatic scenarios based on the most influencial climatic factors as derived from the energy balance study.

The following instruments are installed on the measurement tower at the Murtèl-Corvatsch permafrost-drilling site:

One net radiometer for measuring solar and far-infrared radiation balance (Kipp & Zonen), three anemometers for wind speed and wind direction (Young), four temperatures and humidity sensors (Rotronic). The data are stored by a Campell Scientific CR10x-Logger. In addition ground temperatures are measured with precisely calibrated thermistors (Fenwal and YSI) in the 58 m deep borehole. Therefore, all fluxes for determining the energy balance (radiation, sensible and latent heat, ground heat flux) can be calculated.

First results are presented in fig. 14. This figure shows the different radiation data measured since January 1997. The radiation balance is negative from January to the beginning of March, then the radiation balance becomes positive. This fact is caused by the shortwave radiation which increases strongly when the sun is able to shine directly to the measurement tower.

WEATHERING AND ROCKWALL INSTABILITY IN MURTEL REGION
(N. Matsuoka)

Weathering processes were monitored on a north-facing rockwall (MW site) behind the Murtèl I rock glacier. The rockwall is about 200 m high and composed of schist and phyllite. Weathering and subsequent rockfalls control the debris input into the rock glacier. Based on the volume of the rock glacier debris, BARSCH (1977b, cf. also HAEBERLI & *alii*, in preparation) estimated that the rockwall has retreated at a rate of some 1 to 2 mm yr^{-1} during the Holocene.

Although many field observations have suggested that bedrock weathering is intensified during the period with freeze-thaw oscillations (e.g. RAPP, 1960; FAHEY & LEFEBURE, 1988; MATSUOKA, 1990), rock breakage has never been monitored directly on alpine rock slopes. The present monitoring system permits a concurrent recording of rock fracturing and temperature. Rock fracturing was recorded with a crack extensometer, which can measure separation of joints reaching the rock surface. Rock fracturing at depth cannot be detected. Rock temperatures at 0, 10 and 40 cm deep were also monitored using thermistor probes. Data were recorded hourly in data loggers. Rock surface temperature was also monitored on a ridge (MR site), 50 m east of MW site.

Table 1 lists the mean annual near-surface temperatures and freeze-thaw frequency. The mean temperatures below –1°C are indicative of the presence of subsurface permafrost at both sites. Despite the northern exposure, the location favorable for sunshine provided large diurnal ranges on the ridge (MR site), resulting in high freeze-thaw frequency (ca. 100 cycles yr^{-1}). In contrast, diurnal ranges

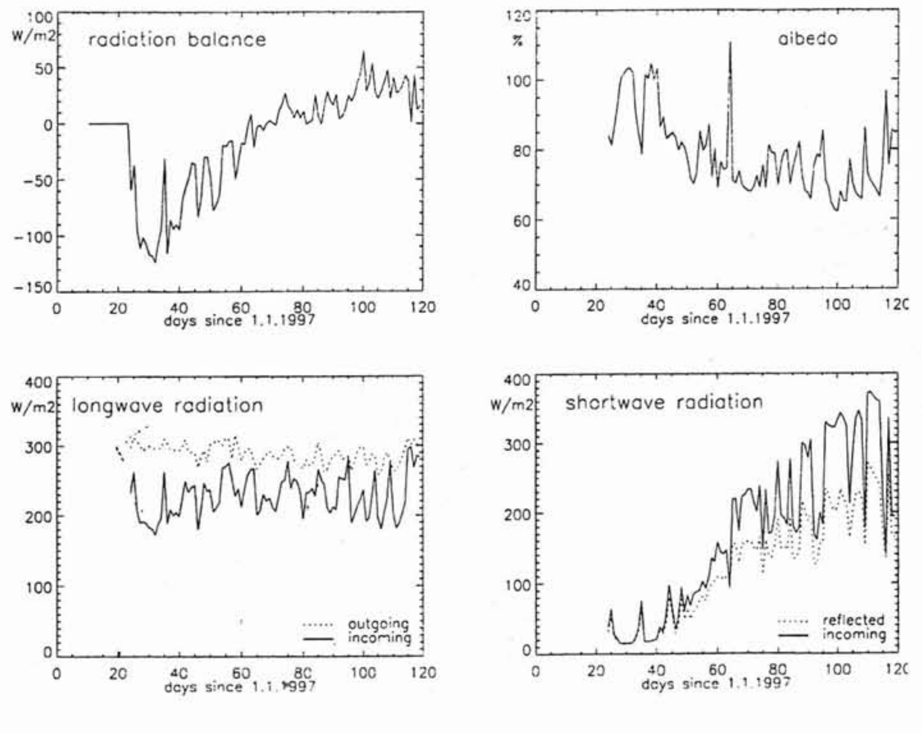

FIG. 14 - Different radiation data measured on Murtèl since January 1997.

of the surface temperature were no more than 10°C on the rockwall (MW site). The small diurnal ranges and late-lying snow cover combined to decrease freeze-thaw frequency to less than 60 cycles at the rock surface. Diurnal freezing and thawing rarely penetrate deeper than 50 cm and only the annual cycle can reach 1 m or deeper.

TABLE 1 - Summary of rock temperature measurements

Site	Depth (cm)	Mean annual rock temperatures (°C)		Number of freeze-thaw cycles (¹)	
		94/95	95/96	94/95	95/96
MR	0	−1.2	−2.0	95	99
MW	0	−1.8	−2.3	25	56
MW	10	−1.9	−2.4	12	38
MW	40	−2.1	−2.5	4	30

The types of the freeze-thaw cycles appear to control the character of frost weathering. The diurnal freeze-thaw action prevails on ridges and south-facing rockwalls, where rock debris smaller than the cobble size is produced from the densely-shattered bedrock. Such small debris is considered to reflect shallow, but frequent diurnal freeze-thaw cycles. In contrast, the north-facing rockwalls are dominated by the annual cycle. The heat conduction equation indicates that the annual freeze-thaw penetration reaches about 5 m deep in rockwalls (cf. MATSUOKA, 1994). The repetition of annual cycles may eventually release large rock blocks, which indeed constitute the scree slopes and rock glacier below the rockwall. Rock avalanches reaching 10 m deep or more can not be interpreted by the annual frost weathering but may be affected by millennial freeze-thaw activity.

Fig. 15 shows concurrent data on rock surface temperatures and joint widening at MW site. Joint widening are represented by the mean daily values, and thereby diurnal fluctuation is canceled. A large extension, 0.6 mm in total, occurred preceding the meltout of the rockwall in the early summer of 1995 (Event B). The joint widening began following the rapid rise in surface temperature reaching 0.0°C, indicating that prior to the extension the rock surface partially melted under the thinning snow. At this time, the near-surface rock temperature lay below 0.0°C. Hence, the joint is considered to have expanded when the meltwater infiltrated into the joint space and froze. A cooling, which was recorded from 23 to 25 June, would have enhanced the joint widening. Joint closing was initiated when the rock surface temperature rose to 0.0°C again. During the thawing period of 1996, no expansion comparable with Event B was encountered. Thawing of the snow cover was probably too rapid to permit the intrusion of meltwater and resulting freezing expansion.

The monitoring highlights that the moisture availability plays a vital role in frost weathering. Significant expansion occurred when the rock joint was presumably saturated with water at a subzero temperature. This observation agrees with the laboratory simulations that freezing expansion in a joint is most rapid just below the freezing point (DOUGLAS & *alii*, 1987; MATSUOKA, 1995). In contrast, the temperature data exhibit the occurrence of some freeze-thaw cycles with a large subzero temperature and a long freezing phase, but few of these cycles were accompanied by joint widening (fig. 15). Such inefficiency is probably due to the lack of moisture.

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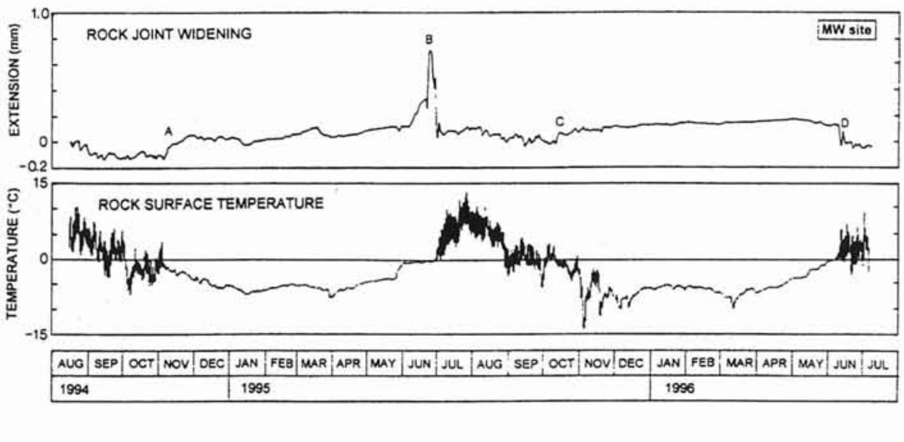

FIG. 15: Rock joint widening and rock surface temperatures at MW site. Widening is given by mean daily values and temperatures by daily ranges.

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PERIGLACIAL PHENOMENA, PERMAFROST DISTRIBUTION AND GLACIAL EVOLUTION IN FOSCAGNO AREA, LIVIGNO (SO), ITALY
(M. Guglielmin)

Since 1989, studies have been carried out in the Foscagno area (GUGLIELMIN, 1989b) and, more generally, in the Livigno area (GUGLIELMIN & TELLINI, 1992, 1994; GUGLIELMIN & *alii*, 1994), aimed principally at a comprehension of periglacial phenomena. The whole excursion area is located inside the territory of the future Natural Livignese Park, the landscape of which is mainly characterized by rock glaciers. Owing to interest in the rock glacier La Foppa I, on the part of both public authorities and the Universities of Parma and Milan, an automatic station was installed in September 1993 for the recording of nivologic and climatic data and for the monitoring of the temperatures of the active layer and of the more surficial portion of permafrost. A small alpine hut has been built close to the station to house the researchers.

Geographical characteristics of the M. Foscagno area

The area of M. Foscagno is located between the Livigno Valley and Upper Valtellina and includes part of the Inn basin on the NW and Adda basin on the SE (fig. 16). The two separate basins, divided by the Monte Foscagno-M. Forcellina group are connected by the Foscagno Pass (elevation 2294 m a.s.l.).

The two most interesting areas from a geomorphological point of view, and especially for periglacial phenomena, are the Foscagno Valley and the 'La Foppa' cirque in Val Vallaccia (Vallacia Valley).

The Foscagno Valley (named 'La Vallaccia' on I.G.M. topographic maps), is located just south of the Foscagno Pass between M. Corno (elevation 2986 m a.s.l.) on the north side and M. Forcellina (elevation 3090 m a.s.l.)-M. Sattaron (elevation 2735 m a.s.l.) on the southern one. This valley is a small, narrow glacial valley, which trends NE-SW. In the upper part, it is divided into two cirques: a major one (the Forcellina cirque) exposed to the NE and a secondary one (the Sattaron Cirque) exposed to the N.

FIG. 16 - Location of the area.

The glacial cirque named 'La Foppa' is a suspended cirque on Vallaccia Valley, which trends NNW-SSE, parallel to Livigno Valley, and is located between the Foscagno Pass to the east and the Eira Pass (elevation 2210 m a.s.l.) to the west.

Geological and climatological constraints

La Foppa cirque and Foscagno Valley belong entirely to the Campo-Languard nappe (Upper Austroalpino), which in this area, is composed of two metamorphic complexes. The first one, named «paragneiss with andalusite complex» (GUGLIELMIN, 1989a, NOTARPIETRO 1990) consists mainly of medium-high metamorphic grade paragneiss and orthogneiss, whereas the second one is named «micascistic-paragneissic complex» and consists mainly of finely-textured paragneiss of low-medium metamorphic grade. The rocks comprising the first metamorphic complex outcrop generally above an elevation of 2600-2700 m a.s.l., whereas the rocks comprising the second complex characterize the slope at lower elevations. The top of M. Forcellina is constituted of a little granitic-granodioritic pluton of Late Hercinyan age (GUGLIELMIN & NOTARPIETRO, 1992).

Near the study area, the only station with a statistically representative record is that of Livigno-San Rocco (elevation 1865 m a.s.l.), located in Livigno Valley, 7 Km northwest of the glacial cirque La Foppa. The mean annual air temperature (Maat) measured at this station is 2.6°C, with a maximum mean monthly value (12.4°C) in July and a minimum (–7.4°C) in January. Precipitation is concentrated in July and August, with a total annual amount of 720 mm.The snow cover is almost settled on 64 cm from January to March, while there are 170 days of permanence of the snow cover. Applying the thermal gradient proposed by BELLONI & *alii* (1993), that amounts to 0.6°C/100 m, the altitude of the isothermic lines of -1°C and of –2°C are 2465 and 2630 m, respectively. From the preliminary analysis of the first data from La Foppa cirque station (elevation 2660 m a.s.l.), the mean annual air temperature (Maat) proved to be –0.31°C. The mean annual soil surface temperature (Maat) is of 0,58 °C in correspondence of lower part of La Foppa I rock glacier, while on the near Late Glacial moraine ridge the Maat is 1.9°C (fig. 17).

Moreover, the snow cover is almost continuous here for more than 270 days every year.

VEGETATION PATTERNS IN PERIGLACIAL AREAS OF FOSCAGNO VALLEY AND LA FOPPA CIRQUE
(N. Cannone)

From the point of view of vegetation, the areas of La Foppa cirque and Foscagno Valley lie within different alti-

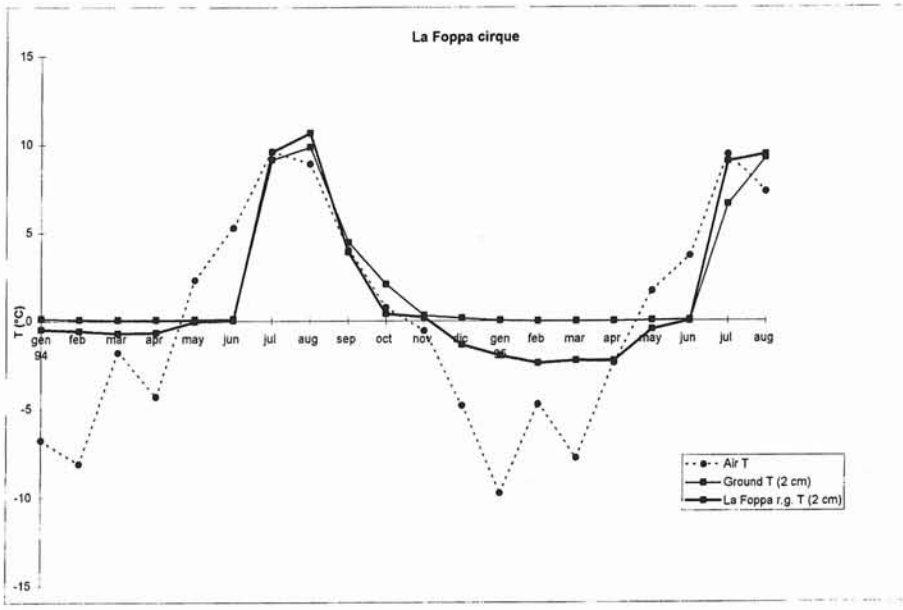


FIG. 17 - Plot Diagram of ground surface temperatures (depth of 2 cm) on frontal part of La Foppa I rock glacier (2650 m a.s.l.); of the Late Glacial moraine (2670 m as.l.) buried by the same rock glacier and the air temperature. It is possible to note the effects of permafrost also on the ground surface temperatures, especially during the winter season, when in the permafrost-free area the temperature remain at the melting point (0°C).

tude lines ranging from the lower Alpine belt (2200-2400 m), characterized by alpine shrub, *Rhododendro-Vaccinietum* and grasslands consisting of *Aveno-Nardetum* and *Festucetum halleri,* to the medium Alpine belt (2400-2600 m), with dwarf shrubs of *Loiseleurietum alpinae* and altitude grasslands consisting *Caricetum curvulae* (climax vegetation) and *Curvulo-Nardetum*, to the upper Alpine belt (2600-2800 m) with a non-continuous *Caricetum curvulae*, snow patch vegetation consisting of *Salicetum herbaceae* and *Polytrichetum sexangularis*, scree slope vegetation of *Oxyrietum digynae* and rocky location vegetation of *Androsacetum vandellii*, to the nival areas (2800-3000 m) with fragmented and poor facies of *Caricetum curvulae,* scree slope vegetation consisting of *Androsacetum alpinae*, and the nival areas with mosses and lichens.

The vegetation of the La Foppa I and Foscagno-Vallaccia rock glaciers, has been studied by CANNONE & *alii* (1995). The vegetation is present in very poor and fragmented facies because of the harshness of the environment. From these analysis, the importance of the vegetation is clear in terms of the reconstruction of the dynamics of periglacial phenomena, for the mapping of permafrost areas and for the reconstruction of the relations between glacial and periglacial events in areas without significant morphological evidence.

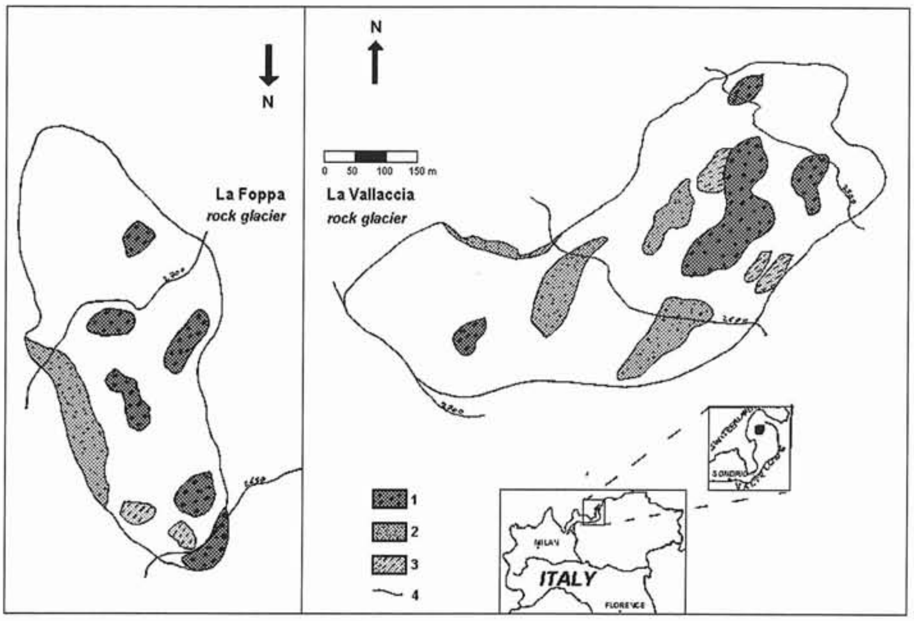
Vegetation studies are useful for analyzing the degree of activity of rock glaciers (differentiating active rock glaciers from uncertain and inactive rock glaciers) and to highlight areas with differing degrees of activity within the same rock glacier. More specifically, on the Foscagno-Vallaccia rock glacier, the most active parts proved to be the central area and the southern area, respectively, in addition to the frontal zone. In the rock glacier La Foppa I, the areas identified as having greater surficial movement were two in the central part of the rock glacier, besides the frontal area.

The reconstruction of periglacial dynamics through vegetation analysis is principally carried out using phytosociological associations for inactive rock glaciers and rock glaciers of uncertain activity (CANNONE & *alii*, 1997). For active rock glaciers, this is carried out using both the distribution of different facies of a particular phytosociological associations and, above all, the distribution of particular indicator species of surficial movement (such as *Geum reptans, Oxyria digyna, Linaria alpina*). Inside the rock glacier, the latter are distributed following dishomogeneous patterns with respect to the surficial texture and the normal altitude zones. Permafrost mapping inside active and/or uncertain rock glaciers is possible through the analysis of the distribution of flora species of surficial movement and of high soil moisture levels indicators (such as *Ranunculus glacialis, Leucanthemopsis alpina, Saxifraga bryoides)*. For our purposes, the comparison with the distribution of mosses and the snow-patch species proved to be a useful tool.

In active rock glaciers, (particularly La Foppa I and Foscagno-Vallaccia) (CANNONE & *alii*, 1997), the presence of particular facies of the association *Oxyrietum digynae* which is abundant with indicator species of surficial movement and of high soil moisture has shown good correspondence with the areas with probable or possible permafrost identified by geophysical or geothermal methods (Bts or Stg).

The analysis of the degree of evolution of the different phytosociological associations can be used for the reconstruction of the evolution of glacialism. More specifically, in the case of the Foscagno-Vallaccia rock glacier, the presence of less developed vegetational facies in the upper part is probably due to the neoformation of a little glacier during the Lia, whereas, on the La Foppa I rock glacier, the inverted altitude zones of the vegetation in the upper part indicates the absence of glacial events connected to the Lia (fig. 18).

FIG. 18 - Permafrost map obtained by the integration of PCA and stress indicator species mapping. Legend: 1) permafrost (active layer <2.5 m); 2) snow bank; 3) vegetation absence; 4) r.g. boundary.



GLACIAL AND PERIGLACIAL EVOLUTION OF THE STUDY AREA (M. Guglielmin)

Glacial evolution during the Late Glacial period was quite similar for the Foscagno Valley and La Foppa cirque. In both areas, which are exposed to the North, with altitudes of the tops located respectively between 2700-3090 m and 2600-3133 m, there are moraine ridges attributable to three different glacial events of the Late Glacial Period. Moreover, in the Foscagno Pass area there are older glacial and fluvioglacial deposits. These three different glacial events show comparable Elas. in both Foscagno valley and Vallaccia valley (Ela: Equilibrium Line Altitude; computed according to the AAR method).

On this basis, the three different glacial events were represented on the simplified geomorphological map of fig. 19 (reduced from GUGLIELMIN & TELLINI, 1997).

In Foscagno valley, the older glacial events (Foscagno I phase) reach an elevation of 1990 m a.s.l. at the frontal moraine ridges, whereas the Foscagno II phase reaches 2060 m a.s.l. at the front. The third Late Glacial event is expressed by the clearly visible left-lateral moraine form, at 2370 m in the principal cirque, and by the deposits located above 2400 m in the Mount Sattaron cirque. It is feasible that the two glacial streams were already divided in this period.

In Vallaccia Valley, the older Late Glacial event named Vallaccia I reaches the elevation of 1930 m a.s.l. at the front, whereas the second one (Vallaccia II) reaches the elevation of 2000 m a.s.l. with its front. The younger Late Glacial event in Val Vallaccia is recorded by the frontal moraine ridges of various lateral cirques, among which La Foppa reaches the elevation of 2200 m a.s.l. During this period, the Val Vallaccia glacier was completely broken up into several cirque glaciers with small tongues.

The Holocene evolution of these areas is still subject to discussion. In Foscagno Valley, there are Holocene deposits and moraine ridges in the Forcellina cirque above the altitude of 2730 m a.s.l. and in the Sattaron cirque, above 2510 m a.s.l. More specifically, the glacial deposits of the Forcellina cirque are located in the southern part of the principal cirque. The glacial deposits in this cirque are completely shapeless and almost completely covered by scree slope. At the altitude of 2800 m a.s.l., there is a small moraine ridge that is probably attributable to the Little Ice Age. In the Sattaron cirque, there are small moraine ridges that at the front reach an elevation of 2570 m a.s.l.

In Val Vallaccia, there are Holocene glacial deposits in La Foppa cirque where a small lateral moraine ridge is visible and reaches an altitude of 2685 m a.s.l. Such deposits also appear in the fourth glacial cirque of Pizzo Filone, where two small fronts reach altitudes of 2630 and 2710 m a.s.l., respectively. The Little Ice Age is recorded in the second cirque on Pizzo Filone, where moraine deposits are present between 2790 and 2850 m a.s.l.

A comparison of the Ela of different glacial events in the two valleys studied showed that phases Trepalle-Foscagno I, Vallaccia I-Foscagno II and La Foppa-Forcellina-Sattaron, are attributable to the same glacial events. The difference between the Ela of the three Late Glacial events identified and the Ela of the Lia (Ela 1850) in Upper Valtellina and Livigno area, was calculated and a comparison was made between these values and those known in literature on the Alpine and bordering areas (BURGA, 1988). It is thus possible to refer these three events approximately to the Daun, Egesen I e Egesen II stadia, respectively

The historical documentation of the glaciation in this area is limited to one small glacier located in the Forcellina cirque and to three small glaciers on northern slope of Pizzo Filone. All these glaciers were recorded until 1932 and in particular, it was noted that the small glacier of Forcellina was located between 2650 and 2800 m as.l., with a surface area of 7 ha.

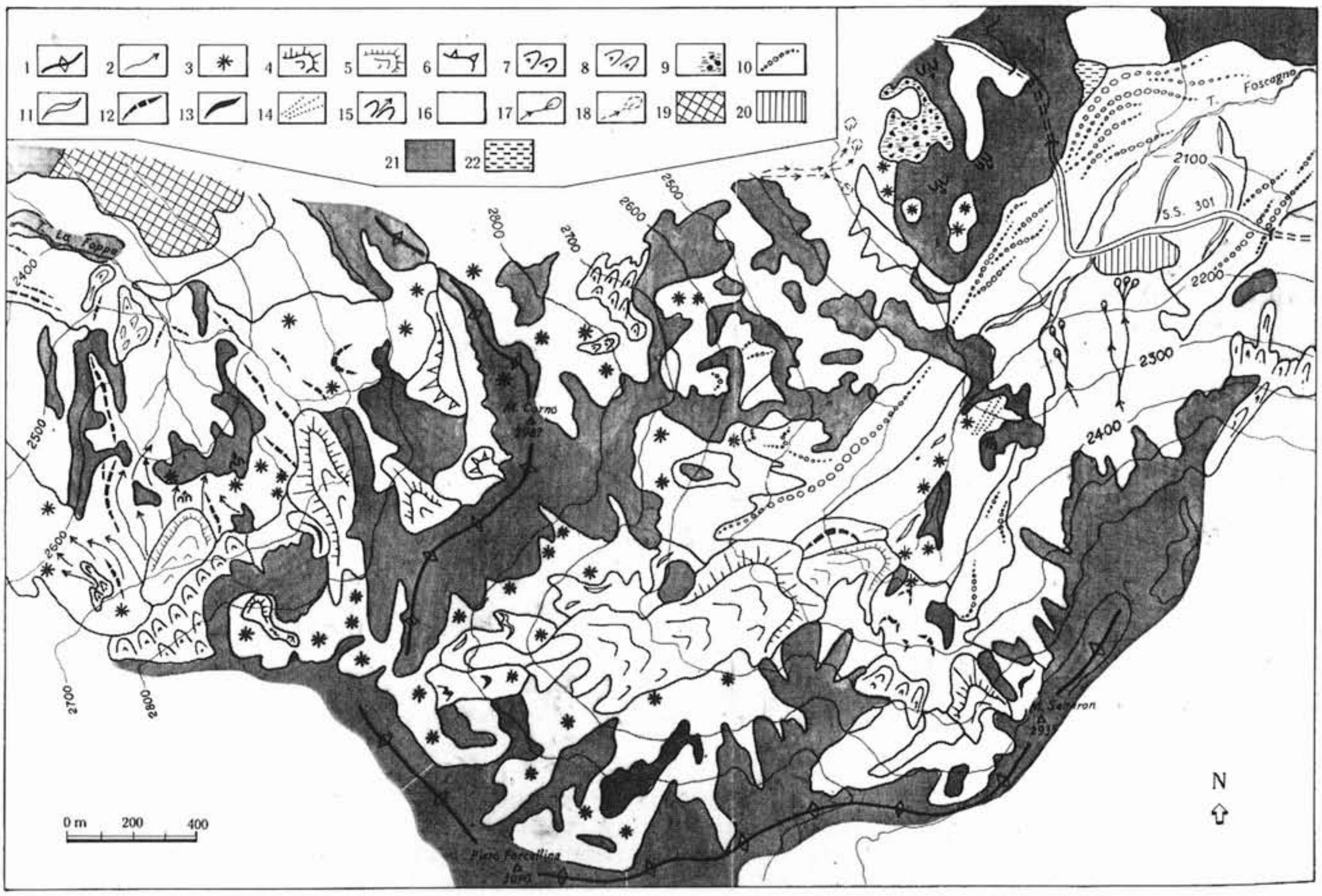


FIG. 19 - Simplified geomorphological map of the area.
Legend: 1) sharp ridge; 2) boulder stream; 3) patterned ground; 4) active rock glacier; 5) inactive rock glacier; 6) active protalus rampart; 7) active gelifluction lobes; 8) inactive gelifluction lobes; 9) pre-Late Glacial Foscagno I fluvioglacial deposits; 10) Late Glacial Foscagno I moraine deposits and ridges; ; 11) Late Glacial Foscagno II moraine deposits and ridges; 12) Late Glacial Forcellina; Sattaron and La Foppa moraine deposits and ridges;13) Holocene moraine deposits and ridges; 14) Flute moraine, 15) roches moutonnées with directions of the striations; 16) Holocene scree slope; 17) active debris flow; 18) inactive debris flow; 19) collapsed slope (Creep Mass Rock); 20) anthropic* deposits; 21) undifferentiated metamorphic rock outcrops.

The upper part of Foscagno Valley is characterized by the Foscagno rock glacier, located at an altitude between 2450 and 2730 m. The upper sector limits are not defined and it is a complex rock glacier, composed of various lobes, with a composite surficial morphology and texture characteristics. On the left slope, at 2510 m, it is possible to observe a small lateral lobe, with a rather fine texture, many concentric ridges and funnel-shaped depressions, partially filled by following materials from debris flow or from unconstrained surficial waters. In the central part of the rock glacier, there are three lobes showing only slight relief, with some concentric ridges in the left sector, whereas in the right sector, it is not possible to recognize any evident surficial forms and the texture is generally finer. In the frontal part of the rock glacier, there are two partially overlapped lobes.

The La Foppa cirque is characterized by periglacial landforms with five rock glaciers, of which four are active and one of uncertain activity or inactive. Among other periglacial active landforms, gelifluction lobes, patterned grounds (sorted and unsorted polygons, sorted striped soils) and ploughing blocks are more widespread. More specifically, the gelifluction lobes cover a large part of the La Foppa crest. Near the inactive rock glacier there are two large inactive stone circles (diameter> 5 m), which are probably attributable to the Late Glacial period. At the foot of M. Corno peak, there are two wide active protalus ramparts. The diameter of the active polygonal soils ranges from 30 to 100 cm and they are located above 2700 m a.s.l.

The fronts of the most important rock glaciers, named La Foppa I and La Foppa II, reach altitudes of 2635 and 2650 m a.s.l. (CALDERONI & *alii*, 1993). The first one is a tongue rock glacier, with a length of about 600 m and a maximum width of 150 m. The surficial morphology of the La Foppa I rock glacier is characterized by many ridges and furrows which are especially thick in the frontal part of the body. The second one is a lobate rock glacier of a triangular shape with an open-work surficial texture consisting in large boulders. This rock glacier presents some furrows and ridges, and a slight hollow profile.

In the Foscagno area outside of La Foppa cirque and Foscagno Valley, there are some additional rock glaciers, both active (Pizzo Filone I, II, III, Foscagno Pass) and inactive (Castelletto, Foscagno Pass, West Foscagno M.).

Some paleosoils buried by rock glaciers in the study area, have been dated using the ^{14}C dating method (CALDERONI & *alii*, 1993). The non-calibrated ages all indicated Holocene ages. The La Foppa I rock glacier is the youngest with a ^{14}C age of 1,120 ± 60 yrs BP, whereas the La Foppa II rock glacier is the oldest one with a ^{14}C age of 5,000 ± 70 yrs BP.

The inactive part of the Foscagno rock glacier is older than 2,700 ± 70 yrs BP and the active part is older than 2,200 ± 60 yrs BP. Considering the type of burial of these paleosoils and the time needed for aggradation of the permafrost before the beginning of creep downward, it is possible to suspect that the periglacial environment started at least 2,000 yrs before the radiometric dates indicated above. These data suggest a complete retreat of the glaciers at the end of the Late Glacial period in the Foscagno Valley, where only during the Lia could a new small glacier reform, whereas in the La Foppa cirque, the small Holocene moraine ridges are attributed to a slight glacial advance around 5,000 yrs BP.

PERMAFROST MAPPING
(M. Guglielmin)

The first step in permafrost mapping of the Foscagno Valley and La Foppa cirque, was to conduct a detailed geomorphological survey to detect active and inactive periglacial landforms. At the locations of these features, spring water temperature and other chemical and physical parameters were measured.

The temperature of spring waters found at the foot of active rock glaciers was always lower than 1.7°C, while those at the foot of inactive ones, range from 2 to 4°C, as does the temperature of others springs located at elevations above 2300 m as.l.

At the same time, some Stg (Shallow Temperature of Ground in summertime) measurements were carried out (GUGLIELMIN, in press). This method consists in digging shallow trenches where it is possible to measure the temperature of the ground at different depths, (generally, 5, 10, 20, 30 cm at the surface and the last one at the maximum depth possible). The temperature at a depth of 30 cm was correlated with altitude up to an elevation of 2500 m a.s.l. At higher altitudes, the temperature measured at ground locations with permafrost occurrence, appeared to be completely uncorrelated with altitude. This means that the temperature at the depth of 30 cm in these areas, reflects the mean annual temperature of the ground and hence these parameters can be used to map permafrost occurrence, in addition to Bts. In fact, according to the equation below, there is a good correlation between Stg measurements and Bts measurements .

$$STG_{30} = 6.68 - 0.56\ BTS$$

In the same areas, more than 300 measurements of Bts (Bottom Temperature of Winter Snow Cover) were carried out during the winter season, as shown in fig. 20.

According to HAEBERLI (1973), probable permafrost areas (Bts<–3°C) are continuous above the altitude of 2800 m a.s.l., whereas at lower elevations, the permafrost is limited to active gelifluction lobes and active rock glaciers.

The permafrost distribution appears completely different in the area between the La Foppa I rock glacier and the Foscagno rock glacier. In fact, the former has a central area with values representative of absence of permafrost (near 0°C), while in the latter, the same values characterize the upper area of the rock glacier. More than 30 Ves (Vertical Electric Soundings) were then carried out to detect the thickness of the active layer and, where possible, also the permafrost thickness. The thickness of the active

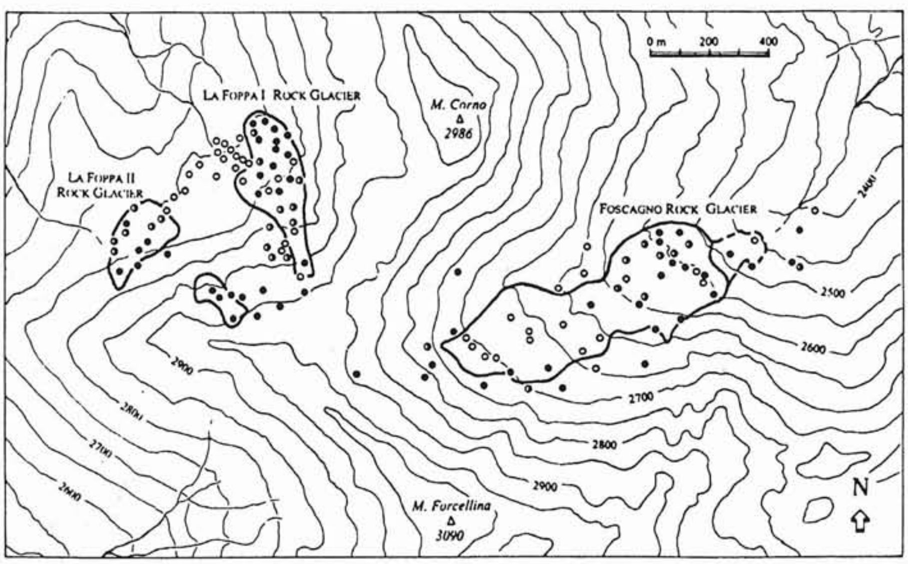


FIG. 20 - BTS measurements in the Foscagno valley and the La Foppa cirque.

layer in the active rock glaciers ranged from 0.8 to 5 m, with a mean value of 1.8 m. The mean permafrost thickness was 12 m, but many soundings did not reach the bottom of it because the soundings were not long enough.

In the inactive rock glaciers of the area, a relict permafrost core was generally found at a mean depth of 22 m.

Figure 21 shows the geological sections obtained along the flow direction of the Foscagno rock glacier (21a,b) and La Foppa I rock glacier. The two sections along Foscagno rock glacier point out the different conditions of permafrost degradation. In the section a-b the values of the electric resistivity indicate a permafrost less degradated (> 500 kΩm in the frontal and central part of the rock glacier) than in the a'-b' (the values > 500 kΩm are limited in the frontal part) section, probably for different conditions of insulation. Both sections show also that in the upper part of the rock glacier the permafrost is more degradated (21-65 kΩm) and thinner than in the frontal zone; this is, probably, the effect of thermal erosion by the small glacier that, during the Lia, was re-formed in the Forcellina cirque.

The conditions of permafrost in La Foppa I rock glacier appear more degradated, in fact the permafrost is generally thiner and with an lower electrical resistivity (23 to 90 kΩm) than in the Foscagno rock glacier. It is interesting the core of relict permafrost which occurs in the central part of the rock glacier and that probably represents a cold holocenic phase around 5,000 yrs BP. In the section 21c is possible to observe also the deep glacial hollow.

Geoelectrical prospecting carried out on the La Foppa II rock glacier indicated that the permafrost is probably very degraded, with a temperature probably nearing the melting point, because the electrical resistivity values were approximately equal to 10 kΩm (fig. 22). In this area, remote sensing techniques for permafrost mapping have been experimented with good results, as shown by the comparison between permafrost maps obtained through a supervised classification of multispectral Landsat TM images and some areas with permafrost identified by Bts and electrical soundings (83 % accuracy) (ANTONINETTI & *alii*, 1994).

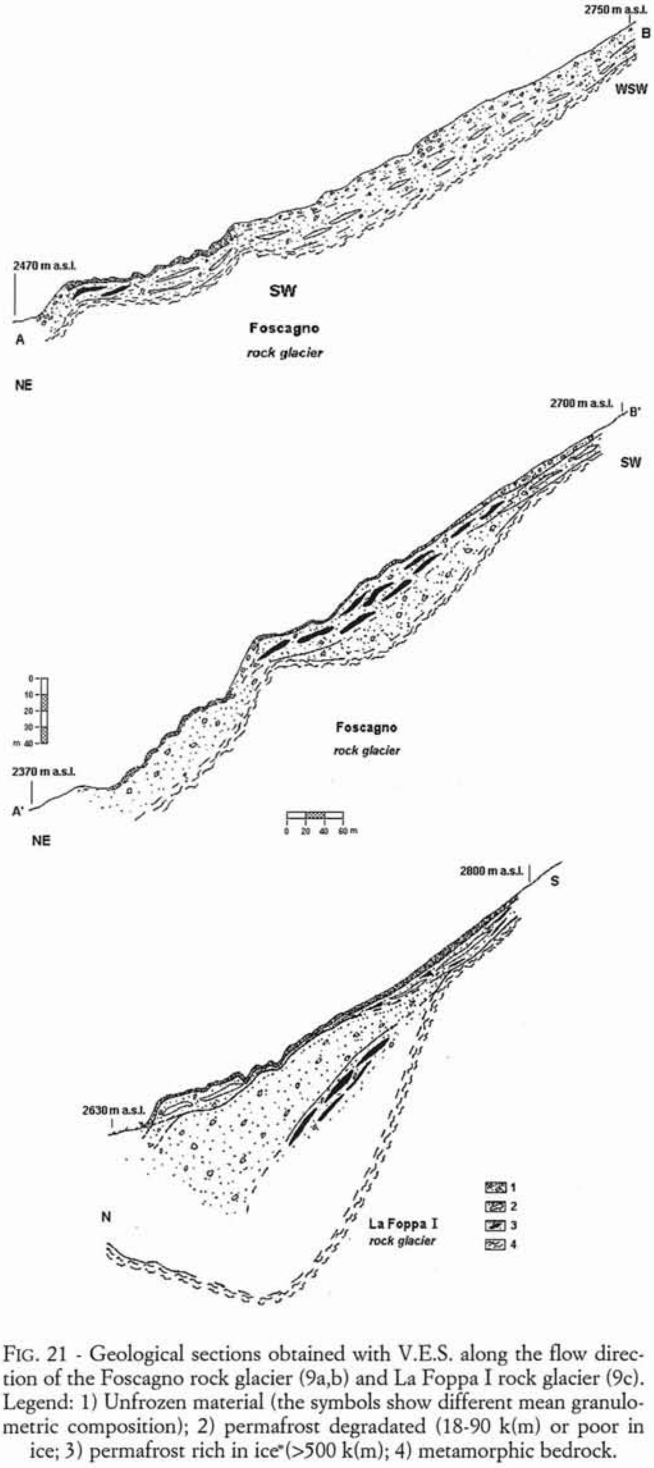


FIG. 21 - Geological sections obtained with V.E.S. along the flow direction of the Foscagno rock glacier (9a,b) and La Foppa I rock glacier (9c). Legend: 1) Unfrozen material (the symbols show different mean granulometric composition); 2) permafrost degradated (18-90 k(m) or poor in ice; 3) permafrost rich in ice*(>500 k(m); 4) metamorphic bedrock.

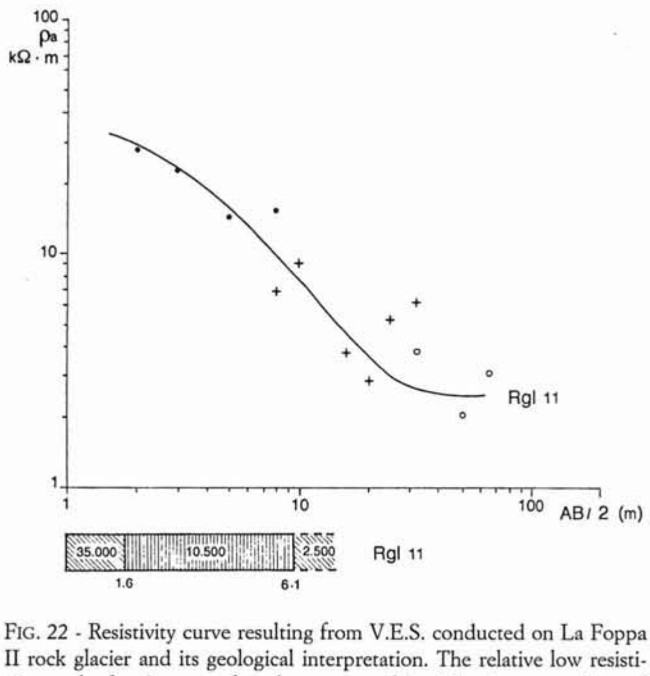


FIG. 22 - Resistivity curve resulting from V.E.S. conducted on La Foppa II rock glacier and its geological interpretation. The relative low resistivity results for the permafrost layer are explained by the strong thermal degradation of the permafrost (see the text).

Figure 23 shows the permafrost map of the Foscagno Valley and La Foppa cirque. The map was prepared on the basis of remote sensing techniques. As we can see, permafrost also occurs on some other scree slopes and moraine deposits above an elevation of 2500 m a.s.l., in areas not investigated by Bts and geophysical methods.

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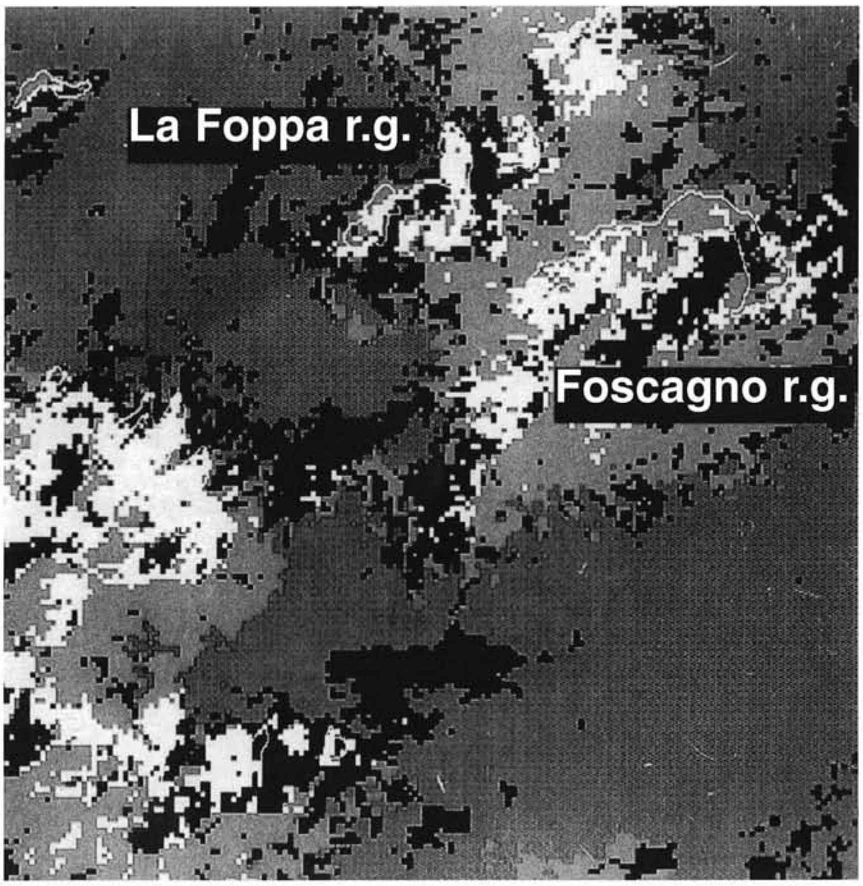


FIG. 23 - Permafrost mapping of the La Foppa cirque and Foscagno Valley with remote sensing images classified using the Fuzzy method. The higher probability-probable permafrost areas are represented in white.

Tuesday, 26. August 1997

PERIGLACIAL MORPHOLOGY IN VAL PISELLA
(C. Smiraglia & M. Guglielmin)

Val Pisella is situated in Valfurva (Upper Valtellina, Ortles-Cevedale Group, Italian Alps). It is a secondary valley, hanging on the main valley of Valfurva, which it joins with a north-south direction, enclosed between peaks: the Sasso della Manzina (2920 m) to the west, the Cime dei Forni (3240-3247 m) to the north and Monte dei Forni (3054 m) to the east (fig. 24).

The bedrock throughout the entire valley is formed of Paleozoic Bormio phyllites, which, in turn, consist of quartz phyllites with sericite and chlorite; garnet-bearing phyllites; and schists with chlorite. Quartzites and crystalline limestones are intercalated. The Bormio phyllites formation is cross-cut by quartz-diorites and porphyry dykes. The entire metamorphic bedrock of Val Pisella proves to be fractured and probably also faulted. The main fracture systems are oriented in two main directions: WNW-ESE, that is, approximately along the Valfurva axis, and NNW-SSE, particularly evident on the rock surfaces modelled by the ice.

The morphology of Val Pisella is essentially derived from the action of glacial and periglacial processes. At present, glacial morphogenesis is inactive. In fact, the last small glacier, the Ghiacciaio delle Cime dei Forni, contained within the eastern cirque of the valley head, has been considered as extinct since at least forty years (DESIO, 1967). Even the small glacieret that occupied the cirque bottom until the mid 1980s, has now disappeared.

The snow-, frozen ground- and frost-related processes are instead the processes that are now creating the most significant changes on the slopes. In the highest part of the basin, the intensity of the freezing-thawing action is markedly facilitated by the bedrock that is structurally subject to weathering. Large areas are covered by cryoclasts, which are generally found in cones bordering the base of the steepest slopes. Rock glaciers are numerous and they are distributed within two main altitude zones, a lower one ranging between 2550 and 2650 m (altitudes of the fronts) and an upper one ranging between 2825 e 3000 m. The former are situated in the cirques of the upper valley, whereas the latter are found in the intermediate-low sector of the left slope of the valley.

There are also numerous other examples of periglacial landforms at the level of meso- and microforms, such as hummocks, stone pavements, sorted stripes, ploughing blocks; the latter being very frequent on the debris cover in the lower Val Pisella. The other forms are found mainly in the upper valley, between the upper rock glaciers and the frontal moraines.

Climate

There are no meteorological stations located in Val Pisella. Therefore, the climatic data refer to the two nearest stations, the Forni station, located at 2120 m slightly further west from the Val Pisella outlet into Valfurva, and the S. Caterina Valfurva station, situated at 1740 m in altitude, about 5 km downstream from the previous one.

Some data recorded at the two meteorological stations may serve to illustrate the general climatic conditions:

	Forni	S. Caterina Valfurva
Mean annual air temperature	1.5 °C	3.7 °C
Mean air temperature in January	–6.8 °C	–6.3 °C
Mean air temperature in July	11.1 °C	12.9 °C
Days without thaw	81	106
Days with frost	125	56
Days without frost	160	203
Mean annual precipitation	807 mm	874 mm

Equilibrium line altitude (Ortles-Cevedale)roup)3050 m

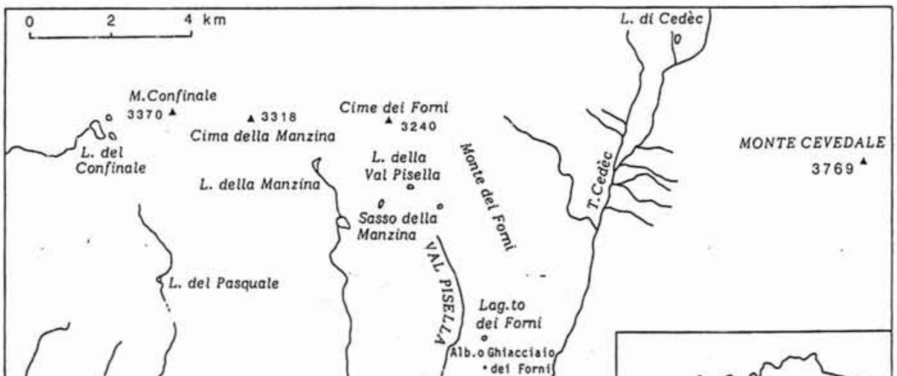


FIG. 24 - Geographical sketch of the Valfurva and Val Pisella area.

The rock glaciers in Val Pisella

The rock glaciers definitely constitute one of the principal and most interesting features of the periglacial morphology of Val Pisella. Six main forms were identified. The main topographical characteristics observed on the aerial photographs and by means of field surveys are presented in Table 2.

TABLE 2 - Main topographic parameters of the Val Pisella rock glacier

Location	Front elevation (m a.s.l.)	Max basin elevation (m a.s.l.)	Surface area (sq. m)	Max width (m)	Max length (m)	Slope	Aspect	State of activity
Cirque	2,875	3,040	23,750	125	250	12°	SE	Uncertain
Cirque	2,830	3,232	60,000	175	375	15°	S	Inactive
Cirque	2,925	3,232	45,000	175	350	20°	SE	Active
Valley	3,000	3,232	40,625	200	275	25°	S	Active
Cirque	2,825	3,100	100,000	200	500	19°	S	Active
Valley	2,650	3,000	43,125	225	250	16°	SW	Active

The three rock glaciers defined as active in the table are located in the two cirques which cut the upper sector of Val Pisella. They are connected by wide slopes covered by debris and are almost totally devoid of vegetation showing a classical morphology with ogive-shaped transversal ridges and abrupt fronts (with a steepness of up to 45°), where mostly fine material appears at the surface. The western and central rock glaciers have a lobate morphology, while the eastern one has a tongue which coming from the upper cirque and stretching down-valley (SMIRAGLIA, 1985).

Some morphological, topographical and geophysical observations have been carried out on these rock glaciers, starting from 1985, especially within the sphere of the work being done by the Gruppo Nazionale Geografia Fisica e Geomorfologia-Sezione Glaciologia (National Working Group on Physical Geography and Geomorphology - Glaciology Section) (fig. 25).

Temperature measurements, resistivity, pH measurements and total hardness of the water coming directly from rock glaciers held to be active and inactive, of direct meltwater from snow, springs and surface circulation water (GNGFG, 1987) were all recorded in order to obtain indications on the rock glacier activity and the presence of ice within the rock glacier.). The most significant results concerned the temperatures. In fact, waters from the rock glaciers held to be active due to morphological evidence and vegetation, showed temperatures well below 1°C, in some cases reaching values similar to those of snow water, but always different from the water circulating in the rocks (higher by more than 3°C) and the water circulating on the surface (temperature ranging between 3°C and 12°C). The waters from the rock glaciers held to be inactive have temperatures slightly lower than 2°C (fig. 26).

For the eastern Val Pisella rock glaciers, measurements of flow velocity were also recorded at 50 points distributed over the surface, from an altitude of 2848 m up to 2970 m (SMIRAGLIA, 1989). The measured values vary from a few centimeters up to a maximum of 40 cm per year, with the highest values recorded in the upper sector and at the center of the rock glacier. Towards the valley and the outer margins, the rate decreases and approaches zero at the terminus. In the belt of the ogive-shaped transversal ridges, the flow is compressing. In the upper part of the rock glacier, flow lines parallel to the longitudinal axis diverge approaching the front, where their trend is almost transversal, as is also revealed by the transversal trend of the long axis of many boulders.

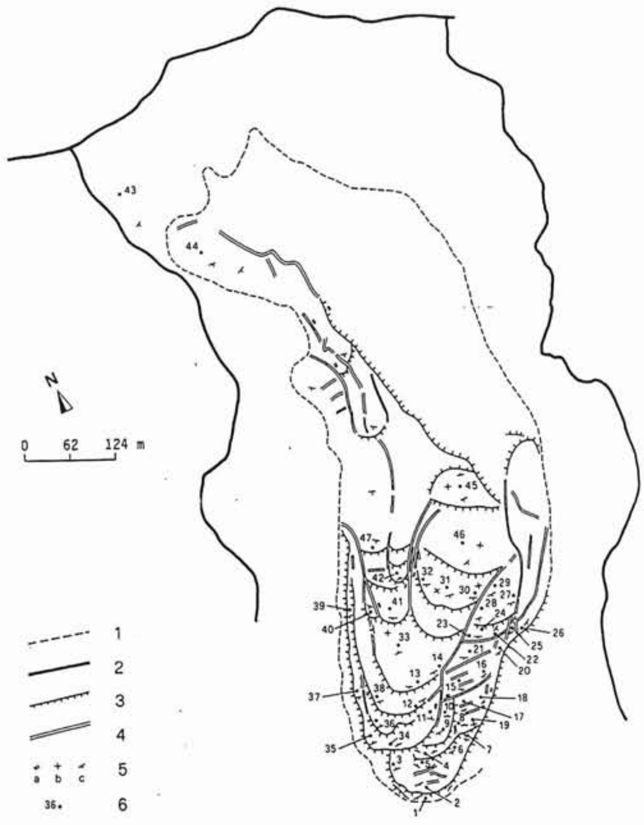


FIG. 25 - Morphological sketch of the eastern Val Pisella rock glacier: 1) edge of the rock glacier; 2) ridge; 3) scarp; 4) trenches; 5a) vertical boulder; 5b) horizzontal boulder; 6) measurent point (GNGFG, 1987).

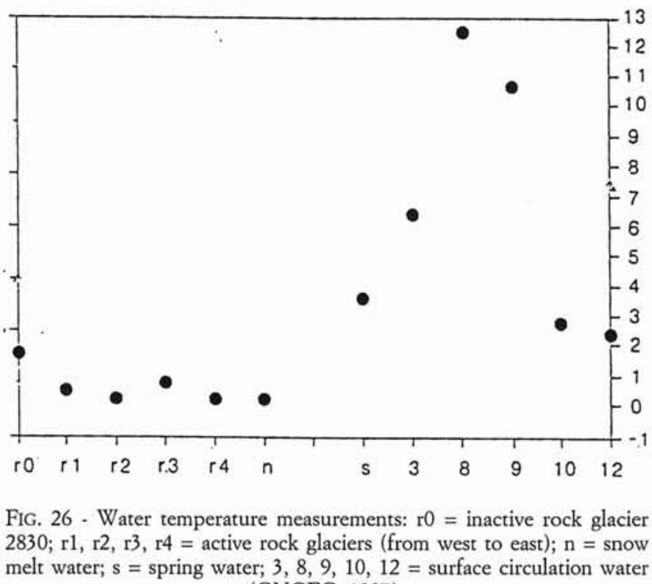
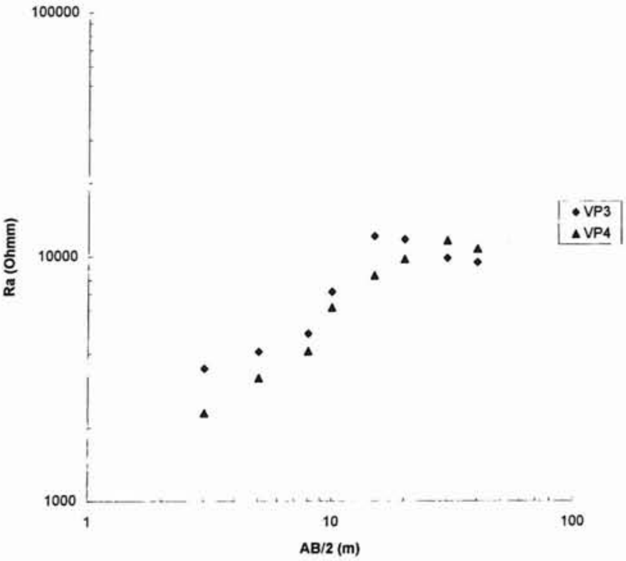


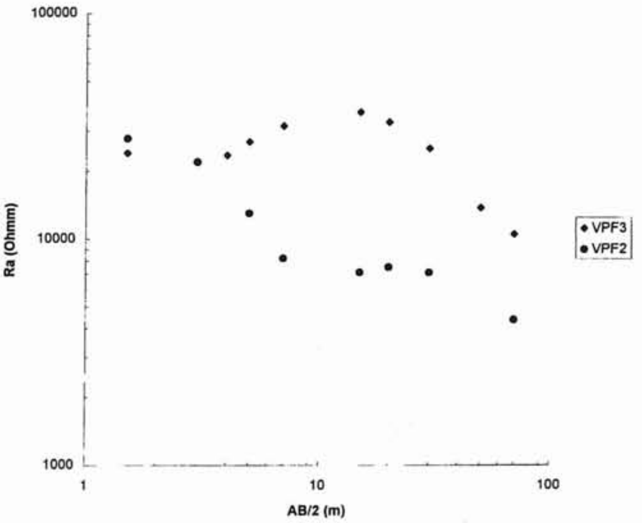
FIG. 26 - Water temperature measurements: r0 = inactive rock glacier 2830; r1, r2, r3, r4 = active rock glaciers (from west to east); n = snow melt water; s = spring water; 3, 8, 9, 10, 12 = surface circulation water (GNGFG, 1987).

On the rock glaciers of Val Pisella and on those found in the nearby Valle della Manzina, some geoelectrical surveys were carried out also in collaboration with some French colleagues (SMIRAGLIA & *alii*, 1994) in order to check the presence, type and thickness of the ice ground. Six vertical electric soundings (Ves) were conducted on the eastern rock glacier in Val Pisella. These Ves were distributed from the moraine in the upper part of the cirque down to the frontal part of the rock glacier; almost all were carried out with simmetrical Schlumberger quadripole trends in direction transversal to the flow of the rock glacier.

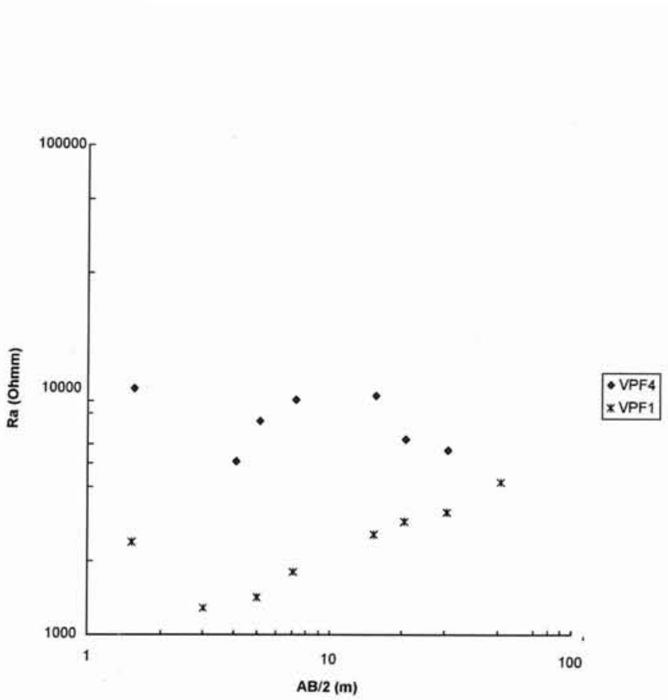
The VP3 and VP4 curves (fig. 27a) show a degradated permafrost (18-25 kΩm) with a thickness that increases towards the frontal part of the rock glacier (about 20 m).



a) Resistivity data obtained from V.E.S. on the eastern Val Pisella rock glacier (from a field survey of Cresnati).



b) Resistivity data obtained from V.E.S. on the upper part (VPF2) and on the frontal part of the eastern Val Pisella rock glacier (from a field survey of D. Fabre).



c) Resistivity data obatained from V.E.S. on the frontal part of the eastern Val Pisella rock glacier (VPF4) and on a moraine of the upper cirque (VPF1) (from a field survey of D. Fabre).

Fig. 27 - Resistivity data abtained from V.E.S. on the eastern Val Pisella rock glacier.

The figures 27b and 27c show the great difference that exist between the electric curves obtained on the moraine (VPF1) or in the upper part of the rock glacier (VPF2) (where only a thin layer of unfrozen diamicton occurs) and the curves obtained from the frontal part of the rock glacier (VPF4 and VPF3) that show the presence of degradated permafrost about 15-20 m thick (fig. 28).

The general conditions of thermal degradation of the permafrost which are still present in the rock glaciers of Val Pisella and those in Valle della Manzina, (SMIRAGLIA, & *alii*, 1994) conflicts with the temperature levels that could be obtained by applying a gradient of 0.64 °C/100 m

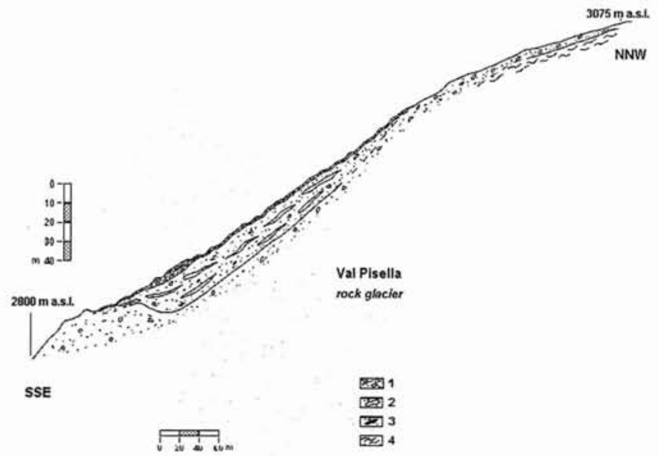


FIG. 28 - Geological section obtained along the flow direction of the eastern Val Pisella rock glacier: 1) = unfrozen material; 2) = degradated or poor in ice permafrost; 3) rich in ice permafrost (not present); 4) metamorphic bedrock.

for the zone falling between 2800 and 3000 m, starting from the mean annual temperatures from the Forni and S. Caterina Valfurva stations. With the data from the former station, at the altitudes indicated, the mean annual temperatures would range between – 2.8 and – 4.1 °C, whereas with the data from the latter, the values would range between – 3.7 and – 4.4 °C. It is well known that the annual isotherm of – 2 °C is considered to be a fundamental climatic parameter for determining the limits of rock glacier activity. In this specific case, the fundamental parameter seems to be the aspect, which is clearly southern for all of the rock glaciers, and which must have markedly raised the altitude limit along which the rock glaciers should retain appreciable amounts of permafrost.

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