Rock Permafrost Geophysics and Its Explanatory Power for Permafrost-Induced Rockfalls and Rock Creep: A Perspective

M. Krautblatter
Institute of Geography, University of Bonn, Germany

Abstract

Rockfalls and rock creep in permafrost-affected bedrock are an increasing hazard in high mountain areas. Besides temperature measurements and physically-based temperature modeling, geophysics in permafrost rocks provides a new methodology for investigating spatial and temporal patterns of permafrost rocks. First examples of 2D electrical resistivity tomographies and refraction seismic tomographies in permafrost rocks are displayed as well as 3D electrical tomographies. ERT time-lapse inversion routines allow for a direct comparison of subsequent time-sections and provide insight into temporal phenomena of heat propagation and permafrost aggradation and degradation. This article aims to show that in the case of ice-filled discontinuities and hydrological pressures geophysical results can potentially be linked to parameters that control rock stability.

Keywords: ERT; ice-mechanics; permafrost; refraction seismics; resistivity; rock creep; rockfalls.

Introduction

Degrading permafrost in rock walls is hazardous, partly due to the amount of potential energy that is released in case of instabilities (Harris et al. 2001). In 2002, the Dzhimarai-khokh rock/ice avalanche (Russian Caucasus) detached approximately 4 million m³ from a 1 km wide starting zone and caused more than 140 casualties (Haeberli, 2005, Kääb et al. 2003). Even smaller permafrost rockfalls, such as the 2003 Matterhorn rockfall, are considered major hazards in densely populated high mountain areas (Gruber et al. 2004). Inventories show that the frequency of these rockfalls has considerably increased in the warm 1990s and was boosted by the hot summer of 2003 (Schoeneich et al. 2004). Moreover, slow rock creep in permafrost rocks causes significant damage especially to tourist infrastructure in high mountain areas.

Besides temperature logger data, borehole information, and rock temperature modeling approaches, the geophysical applications described here provide a new tool for the spatial and temporal analysis of rock permafrost. In some cases, information on the thermal state of permafrost reveals sizeable information for stability consideration (Davies et al. 2001), but changing hydrological properties of ice, such as water content, may also play a vital part in decreasing resistive forces of ice-contained rock masses (Gruber & Haeberli 2007). This paper combines a review of existing geophysical techniques that are applicable in permafrost rocks and a perspective on how these can contribute to understanding mass movements in permafrost-affected bedrock in future. It will address three questions:

(1) What geophysical methods can be applied in permafrost rocks? (2) What properties do they detect? (3) What is their explanatory power for permafrost-induced mass movements?

Investigation Sites

Methods described in this article were tested at three investigation sites: the "Steintälli," a N-S exposed crestline (Matter/Turtmann Valleys, Switzerland) at about 3070–3150 m a.s.l. with slaty paragneiss (see Fig. 1); the North Face of the Zugspitze limestone summit (Wetterstein Mountains, Germany/Austria) at about 2800 m a.s.l., and the gneissic Gemsstock crestline (Switzerland) at 2900 m a.s.l. in collaboration with Marcia Phillips. Figure 1 shows a typical arrangement of a 2D-ERT in steep, permafrost-affected bedrock. More detailed site characteristics can be sourced from Krautblatter and Hauck (2007) and Gude and Barsch (2005). Problems associated with the comparison of different field sites and extrapolation of results are discussed in Krautblatter & Dikau (2007).

Figure 1. Electrical resistivity measurement at the Matter/Turtmann Valleys crestline, 3150 m a.s.l., Switzerland. 41 large steel screws serve as electodes along each transect.

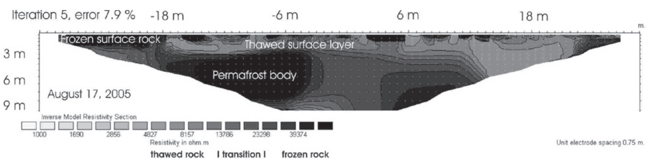

Figure 2. ERT in an east-facing rock wall recorded at August 17, 2005, at the Matter/Turtmann Valleys crestline, 3130 m a.s.l. Switzerland).

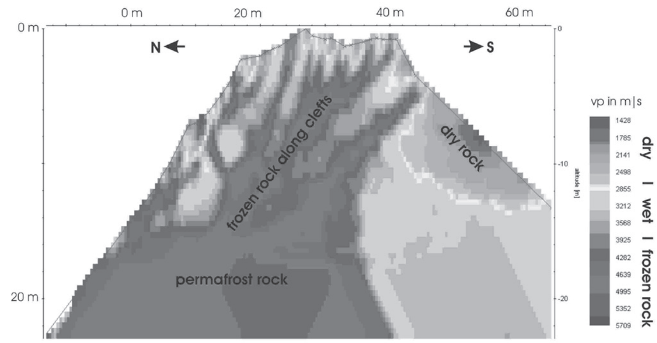

Figure 3. RST – north-south transect 9 (September 20, 2006), measured at the Matter/Turtmann Valleys crestline, 3150 m a.s.l.

Geophysical Methods for Rock Permafrost and Detectable Properties

Surface-based geophysical methods represent a cost-effective approach to permafrost mapping and characterisation (Harris et al. 2001). Hauck (2001) provided a systematic comparison of different geophysical methods for monitoring permafrost in high-mountain environments. However, the application of geophysical methods to permafrost rock walls just began in 2005 (Krautblatter & Hauck 2007). This section will give a quick overview of methods that have successfully been applied to permafrost rocks in the last three years. Data used for Figures 2, 6, and 8 were previously published in Krautblatter & Hauck (2007) and are described there in detail.

Electrical resistivity tomography (ERT)

ERT is a key method in permafrost research, as freezing and thawing of most materials are associated with a resistivity change that spans one order of magnitude, which is, in turn, easily detectable. The first approach to derive spatial information from rock faces by ERT was applied by Sass (2003). In subsequent studies he provided further evidence that ERT measurements are capable of measuring the degree of rock moisture (Sass 2005) and temporal and spatial variations of freeze and thaw limits (Sass 2004) in rock faces. These ERT measurements were confined to the monitoring of the upper weathering crust (centimeter- to decimeter-scale) of non-permafrost rock faces. Krautblatter and Hauck (2007) extended this method to a decameter-scale and applied it to the investigation of active layer processes in permafrost-affected rock walls.

Arrays with centimeter-long steel screws as electrodes were drilled into solid rock (see Fig. 1) and were measured repeatedly with high voltages (mostly 10^2–10^3 V) to improve the signal to noise ratio. A detailed survey of hardware and software adaptations and a systematic discussion of error sources is provided by Krautblatter and Hauck (2007). Errors associated with different ERT-arrays were assessed along with the impact of topography and other geometric error sources (Krautblatter & Verleysdonk 2008a). The Res2DInv software was chosen, as it is capable of topographic correction and "real" time-lapse inversion of subsequent measurements. To cope with high resistivity gradients, inversions models with mesh size smaller than the electrode distance and robust inversion routines provide better results. Resistivity values that correspond to the transition between frozen and thawed rock were measured repeatedly at the rock surface along different arrays and yielded evidence that the transition occurs between 13 and 20 kΩm for the gneissic rocks at the Steintälli (Krautblatter & Hauck 2007) and are in the same range as those established for carbonate rocks at the Zugspitze by Sass (2004) and our own measurements (Krautblatter & Verleysdonk 2008b). Figure 2 shows an ERT that was measured close to Figure 1 at an east-facing part of the rock crestline between Matter and Turtmann Valleys. It shows a 3 m thick thawed surface layer of rock above a constantly frozen permafrost layer; a plunge in temperature following August 14th is evident due to frozen rock close to the surface in all parts of the transect. Resistivity–temperature patterns of rock samples of all field sites are currently tested in a freezing chamber in the laboratory.

The relation between measured resistivity and rock temperature is straightforward. For temperatures below the freezing point, resistivity (p) depends mainly on unfrozen water content until most of the pore water is frozen. In Alpine environments resistivity can be calculated based on a reference value p_o as an exponential response to the temperature below the freezing point (T_f) according to McGinnis et al. (1973):

$$\rho = \rho_0 e^{b(T_f - T)} \tag{1}$$

The factor b in Equation (1) determines the rate of resistivity increase and can be derived empirically (Hauck 2001, Hauck 2002). Short-term changes in resistivity can be attributed to changes in pore water content and temperature, while changes in porosity and water chemistry can be neglected over daily to monthly measurement intervals in low-porosity rocks. Due to the exponential response of resistivity to temperatures below 0°C, freeze-thaw transitions correspond to an increase in resistivity by one order of magnitude and are thus a very sensitive method for detecting the state of rock permafrost close to 0°C. On the other hand, deeply frozen bedrock (below -5°C) along the measured transects causes problems for the electrode contact.

Refraction seismic tomography (RST)

The application of refraction seismics in permafrost studies is based on the interpretation of refracted headwaves that indicate the transition of a slower, unfrozen top layer

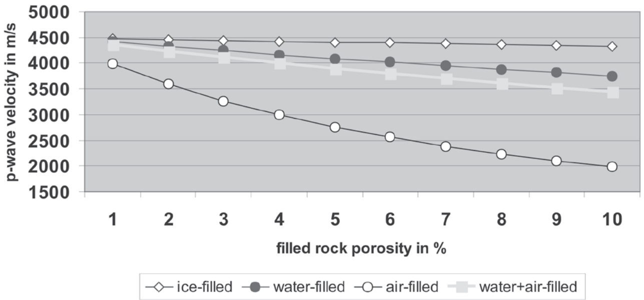

Figure 4. Estimation of P-wave velocities of rock with different porosity and pore content.

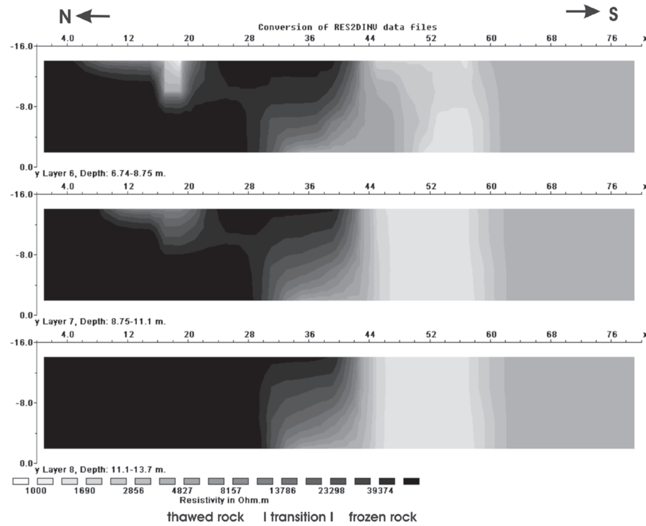

Figure 5. 3D-ERT cross sections at three different depths cutting the Turtmann/Matter Valley crestline N-S. Measured with ca. 1000 datum points from ca. 200 electrodes at September 5–9, 2006.

to a frozen layer with faster P-wave propagation below. Recent approaches apply tomographic inversion schemes (Otto & Sass 2006) often based on high-resolution datasets (Maurer & Hauck 2007, Musil et al. 2002). Seismics are also applied to determine 2D and 3D rock mass properties and potential instabilities (Heincke et al. 2006). Preliminary results from Krautblatter et al. (2007) indicate that refraction seismics are applicable for permafrost detection in solid rock walls, even if they provide less depth information than ERT measurements. On the other hand, it appears that they resolve small-scale features such as ice-filled clefts in more detail. For instance, Figure 3 shows a cross-cut through the E-W trending Steintälli crestline at Transect 9. Thoroughly frozen rock aligns along ice-filled discontinuities indicating the good thermal conduction (k = 2.2 W/(m*K)) without latent buffers in the readily frozen ice in clefts.

In practice P-Waves were stimulated with a 5 kg sledgehammer. Per transect, 24 drilled geophone positions in bedrock and 40 marked and fixed shot positions were arranged in line with the ERT transects so that RST and ERT were measured simultaneously (Krautblatter et al. 2007). A direct comparison of ERT and RST showed that frozen high-resistivity rocks in the ERT typically have P-wave velocities significantly above 4000 m/s (see section below) while wet and dry rock masses indicate a significantly slower propagation. As P-wave velocities of frozen and thawed rock differ only by a few hundred m/s in velocity, it is important to define the geometry of shot and recording position exactly, which was done using a high-resolution tachymeter. High P-wave velocities in rock necessitate high temporal resolution of geophone signals. Surface waves are not decelerated by a soft surface layer, such as soil, and thus often disturb signals recorded by geophones close to the shot position. We applied REFLEXW, Version 4.5 by Sandmeier Scientific Software, with model settings, such as initial P-wave velocity assumption, adjusted to bedrock conditions.

Air-, water-, and ice-filled pores in rock lead to significantly different attenuation of P-wave velocities. This is especially true for air-filled porosity. Figure 4 shows theoretical P-wave velocity for different pore-fillings and rock porosities derived from mixing laws. However, it appeared in simultaneous ERT and RST measurements that carefully conducted RST can resolve the difference between water and ice-filled rock even in low-porosity (2%–3%) bedrock, and that velocity differences are larger than expected from mixing laws for certain rock porosities. This could be due to the fact that melting in low-permeability rocks leads to a small gas content in pores to compensate the ice-water volume reduction or that the seismic velocities provide a more integral signal that includes ice in discontinuities in the rock mass.

The third dimension: 3D ERT and RST

ERT and RST can be conducted in 3D. Figure 5 shows three horizontal sections cut at depths of 7–9 m, 9–11 m and 11–14 m with N-S orientation that indicate a sharp divide between frozen rock to the north and thawed rock to the south at meter 44. Problems that arise when conducting three-dimensional geophysics in permafrost rocks are time-consuming measurements (ca. one week of uninterrupted measurements), the necessity of highly precise topographic information, and the required high resolution due to the enormous resistivity/seismic velocity gradients present in permafrost rock systems. Moreover, traditional 3D arrays (e.g., Pole-Pole or Dipol-Dipol ERT) result in bad signal to noise ratios (Krautblatter & Verleysdonk 2008a) and electrode/geophone spacing and arrays must be adjusted to local conditions. Therefore, the first 3D ERT and 3D RST array in permafrost rocks, which was conducted in 2006, relied on a very close (2 m) arrangement of geophones (120), electrodes (205), and shot positions (200) (Fig. 6) (Krautblatter et al. 2007) and is based on Wenner-arrays that yield much better signal to noise ratios than Pole or Dipole-type arrays.

The fourth dimension: Time-lapse ERT

The installation of permanent electrodes and modeling of subsequent resistivity datasets within the same inversion routine (so-called time-lapse inversion) allows for a direct assessment of spatial and temporal permafrost variability (Hauck 2002, Hauck & Vonder Mühll 2003). Figure 6 shows the freezing of the previously thawed surface rock up to 3 m

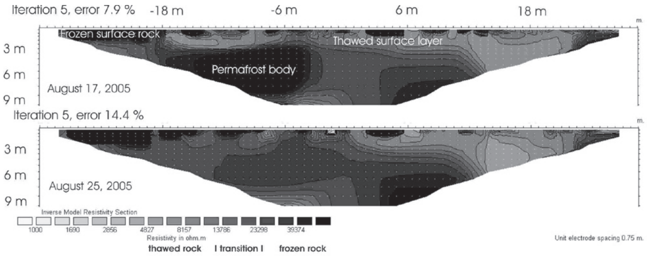

Figure 6. Freezing of surface rock from August 17 (top) to August 25, 2005 (bottom) due to a severe drop in air temperature recorded at the Steintälli E-transect (3130 m a.s.l., Matter/Turtmann Valleys, Switzerland).

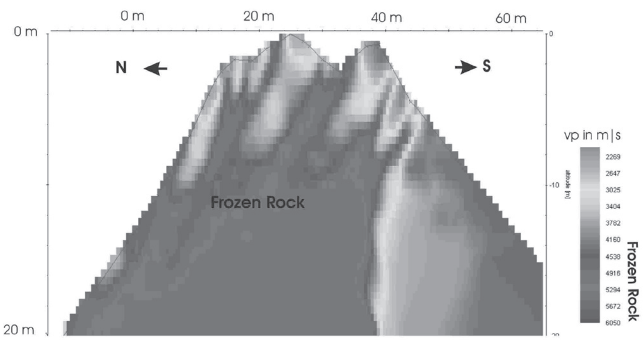

Figure 7. RST – north-south transect 7 (August 31, 2006) of the Matter/Turtmann Valleys crestline, 3150 m a.s.l. Mention the disposition of the frozen discontinuity zones as possible detachment zones with daylighted bedding.

depth following a plunge in temperature after August 14, 2005. While time-lapse routines for ERT are already in place, time-lapse routines for RST are still more difficult to perform.

Time-lapse inversion of subsequent measurements provides insights into short-term and long-term freeze-thaw thermal regimes (Krautblatter & Hauck 2007), response times (Krautblatter & Verleysdonk 2008b), changes in hydrological rock conductivity, and permafrost aggradation and degradation (Krautblatter 2008). Moreover, changes in subsequent time sections can definitely be attributed to changes in rock moisture or the state of freezing, while changes in geological properties can be ruled out for short-timescales.

Explanatory Power for Permafrost-Induced Mass Movements

We define permafrost-induced mass movements as those whose kinematical behavior is at least partly influenced by ice mechanics and permafrost hydrology. The most common types are rockfalls and rock block creep. These are usually explained (1) by a reduction of the resisting force, e.g., shear-strength in ice-filled clefts (Davies et al. 2001, Davies et al. 2000) or (2) an increase in the driving force, e.g., hydrological pressure (Fischer et al. 2007).

Ice-filled discontinuities

Figure 7 shows a cross-cut through the E-W trending Steintälli crestline. Geometrical position, orientation, and

Table 1. Geophysically detectable properties of permafrost rocks

ERT	Space-referenced integral tomography of frozen and thawed rock and hydrological conductivity at all measured depths. Temperature estimation (0° to -5°C) in combination with laboratory measurements (McGinnis).
RST	Space-referenced integral tomography of air-, water-, and ice-filled rock porosity. Exact positions of the uppermost freezing/thawing front and dominant air-, water-, and ice-filled rock discontinuities.
3D-measurements	3D spatial information on the freezing/melting front, hydraulic conductivity, and the persistence/ importance of discontinuity zones.
Time-lapse inversions	Development of heat fluxes, the permafrost system (aggradation/degradation), and the hydraulic system over time.

persistence of ice-filled clefts in the upper 10 m can be well detected in RST surveys. It is assumed that ice-filled discontinuities react according to stress-strain behavior of weight-loaded polycrystalline ice. The deformation of ice at constant stress is characterized by four phases: (1) elastic deformation that is followed by permanent deformation, first at a decreasing rate (2) primary creep, then at a constant rate (3) secondary creep, and finally at an increasing rate (4) tertiary creep (Budd & Jacka 1989). Mostly secondary creep and tertiary creep occur at speeds relevant for mass movements. The flow relation for secondary creep relates the shear strain rate ε_{xy} to the shear stress τ_{xy},

$$\varepsilon_{xy} = A\tau_{xy}^{n} \tag{2}$$

where A depends mainly on ice temperature, anisotropic crystal orientation, impurity content, and water content, and n increases at shear stresses greater than 500 kPa (Barnes et al. 1971). Crystal orientation, impurity content, and shear stresses remain more or less constant over short timescales. In contrary, ice temperature and water content in mass movement systems are subject to major annual and interannual changes. Thus, for temperatures above -10°C, A can be approached by

$$A = A_0 \exp(-\frac{Q}{RT}) \approx A_0 \exp(-\frac{16700}{T}) \tag{3}$$

where A_0 is independent of temperature, R is the universal gas constant, and Q is the activation energy (Weertman

1973) and A_{0t} for tertiary creep

$$A_{0t} = (3.2 + 5.8W) * 10^{-15} (kPa)^{-3} s^{-1} \tag{4}$$

can be related to the percentage water content W. It must be stressed that the water content strongly decreases with temperature. Paterson (2001) states -2°C as the lowest temperature at which the effect of water in the ice is relevant for the stress-strain behavior.

Equations (3) and (4) show that both ice temperature and water content play a dominant role in the mechanical behavior of ice-filled clefts at temperatures close to -0°C. Assuming moderate water content of 0.6%, the creep rate at 0°C is three times the rate at -2°C (Paterson 2001), which has serious effects on displacement rates and factors of safety considerations of mass movements.

As has been shown above, ERT and refraction seismics are highly susceptible to water/ice content inside the rock system just below the freezing point. Thus, the susceptibility range of seismic velocity and resistivity (ca. -0°C to -5°C) corresponds to the most important changes in ice-mechanical properties. This means the values temperature and water content, which are relevant for stability considerations in well-jointed permafrost rocks, are targeted by ERT and refraction seismics and combined interpretation strategies.

Hydrological pressure

Figure 8 shows light-colored, low-resistivity cleft water zones percolated by glacial meltwater that were observed to persist over several years and to limit the spatial extension of permafrost bodies (Krautblatter 2008). While pressure effects only have a small effect on the stress-strain behavior of ice itself (resisting force) (Weertman 1973), the reduction of applied normal stress and the increase in shear stress (driving force) may play a key role in preparing and triggering mass movements (Fischer et al. 2007, Terzaghi 1962). According to Wegmann (1998), permafrost degradation and aggradation in rocks in response to altered hydraulic conductivity occurs at all depths and quickly responds to annual melting patterns. He could also show that rock deformation quickly responds to spatial changes in permafrost rock conductivity.

Unfrozen cleft zones can easily be detected at the surface with RST and with ERT measurements possibly up to the maximum depth of the applied array (e.g., 80 m at the Zugspitze, 400 m Schlumberger-array). As shown in Figure 8, resistivity in water-filled cleft zones and frozen rock typically differs by 1–2 orders of magnitude and is, thus, easily detectable even at greater depths (Krautblatter & Hauck 2007). This opens up a whole range of new possibilities e.g., for the investigation of rock permafrost hydrology (Wegmann 1998), glacier-permafrost inter-connectivity (Moorman 2005), and rock deformation processes that are closely linked to freeze-thaw processes by latent heat transfer in clefts (Murton et al. 2006, Wegmann 1998).

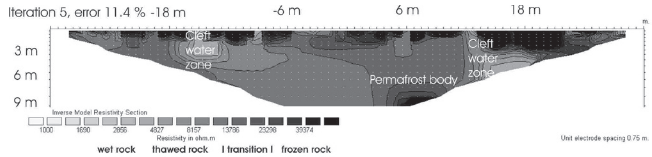

Figure 8. ERT of Transect NE (September 13, 2005). Mention the persistently thawed deep- reaching cleft water zones.

Conclusion

Resistivity monitoring may provide indications on temperature changes and water saturation, while refraction seismics help to gain insight into discontinuity zones and exact geometric properties of instable bodies. Repeated time-sections reveal interannual, annual, and multiannual time-patterns as well as response times, the fourth dimension of rock permafrost systems.

For permafrost-induced mass movements, with secondary and tertiary creep of ice close to -0°C, three highly-variable forces play a key role in unbalancing resisting and driving forces. The resisting force of ice-creep in clefts is mainly controlled by (1) temperature and (2) water content in the ice. Due to the laws of electrolytic conductivity, resistivity values assessed by ERT react sensitively to both parameters, and water content is a key control for P-wave velocity. The highly variable driving force, (3) hydrological pressure, is well detectable in ERT time-sections as pore and cleft space supersaturation lead to a plunge in electrolytic resistivity. However, many other anisotropic factors distort ERT and seismic measurements, and further field and laboratory experience is needed for the allocation of their influence and for the “suppression” of such noise.

Acknowledgments

This work has been funded by the German Research Foundation as part of the graduate school 437 “Relief.” Special thanks to C. Hauck, J.-C. Otto, R. Dikau, M. Moser, S. Wolf, S. Verleysdonk, and all who participated in the fieldwork.

References

Barnes, P., Tabor, D. & Walker, J.C.F. 1971. Friction and creep of polycrystalline ice. *Proceedings of the Royal Society of London, Series A*, 324(1557): 127-155.

Budd, W.F. & Jacka, T.H. 1989. A review of ice rheology for ice-sheet modeling. *Cold Regions Science and Technology* 16: 107-144.

Davies, M.C.R., Hamza, O. & Harris, C. 2001. The effect of rise in mean annual temperature on the stability of rock slopes containing ice-filled discontinuities. *Permafrost and Periglacial Processes* 12(1): 137-144.

Davies, M.C.R., Hamza, O., Lumsden, B.W. & Harris, C. 2000. Laboratory measurements of the shear strength of ice-filled rock joints. *Annals of Glaciology* 31: 463-467.

Fischer, L., Huggel, C. & Lemy, F. 2007. Investigation and modeling of periglacial rock fall events in the European Alps. *Geophysical Research Abstracts* 9(08160).

Gruber, S. & Haeberli, W. 2007. Permafrost in steep bedrock slopes and its temperature-related destabilization following climate change. *Journal of Geophysical Research - Earth Surface* 112(F2): F02S13.

Gruber, S., Hoelzle, M. & Haeberli, W. 2004. Permafrost thaw and destabilization of Alpine rock walls in the hot summer of 2003. *Geophys. Res. Lett.* 31(13): L15054.

Gude, M. & Barsch, D. 2005. Assessment of geomorphic hazards in connection with permafrost occurrence in the Zugspitze area (Bavarian Alps). *Geomorphology* 66: 85-94.

Haeberli, W. 2005. Investigating glacier-permafrost relationships in high-mountain area: historical background, selected examples and research needs. In: C. Harris & J.B. Murton (eds.), *Cryospheric Systems: Glaciers and Permafrost.* London: Geological Society Special Publication, 29-37.

Harris, C., Davies, M.C.R. & Etzelmüller, B. 2001. The assessment of potential geotechnical hazards associated with mountain permafrost in a warming global climate. *Permafrost and Periglacial Processes* 12(1): 145-156.

Hauck, C. 2001. Geophysical methods for detecting permafrost in high mountains. *Mitt. Versuchsanst. Wasserbau, Hydrologie und Glaziologie, PhD-thesis ETH Zürich* 171: 1-204.

Hauck, C. 2002. Frozen ground monitoring using DC resistivity tomography. *Geophys. Res. Lett.* 29, 2016, doi: 10.1029/2002GL014995: 12-1.

Hauck, C. & Vonder Mühll, D. 2003. Inversion and interpretation of two-dimensional geoelectrical measurements for detecting permafrost in mountainous regions. *Permafrost Periglac. Process.* 14(4): 305-318.

Heincke, B., Maurer, H., Green, A.G., Willenberg, H., Spillmann, T. & Burlini, L. 2006. Characterizing the fracture distribution on an unstable mountain slope using shallow seismic tomography: 2-D versus 3-D approaches. *Geophysics* 71(6): B241-256.

Kääb, A., Wessels, R., Haeberli, W., Huggel, C., Kargel, J.S. & Khalsa, S.J. 2003. Rapid aster imaging facilitates timely assessments of glacier hazards and disasters. *EOS* 13(84): 117, 121.

Krautblatter, M. 2008. Permafrost aggradation in the rock crest "Steintälli" (Valais, Switzerland) as a multiannual response to cool summers recorded by a three-year monitoring of rock permafrost by 2D/3D ERT and refraction seismics. *Geophysical Research Abstracts* 10: A-10496.

Krautblatter, M. & Dikau, R. 2007. Towards a uniform concept for the comparison and extrapolation of rock wall retreat and rockfall supply. *Geografiska Annaler* 89A(1): 21-40.

Krautblatter, M. & Hauck, C. 2007. Electrical resistivity tomography monitoring of permafrost in solid rock walls. *Journal of Geophysical Research-Earth Surface* 112. doi:10.1029/2006JF000546.

Krautblatter, M., Hauck, C. & Wolf, S. 2007. Geophysical 2D and 3D-monitoring of permafrost in rock walls. *Geophysical Research Abstracts* 9: A-09884.

Krautblatter, M. & Verleysdonk, S. 2008a. Rock wall permafrost monitoring with high-resolution 2D-ERT: lessons learnt from error estimates and a comparison of Wenner, Schlumberger, Gradient and Dipole-type arrays. *Geophysical Research Abstracts* 10: A-10383.

Krautblatter, M. & Verleysdonk, S. 2008b. Thawing regime and response time of present-day and relict permafrost revealed by monthly geophysical monitoring (Zugspitze, German/Austrian Alps). *Geophysical Research Abstracts* 10: A-10455.

Maurer, H. & Hauck, C. 2007. Instruments and methods: geophysical imaging of alpine rock glaciers. *Journal of Glaciology* 53(180): 110-120.

McGinnis, L.D., Nakao, K. & Clark, C.C. 1973. Geophysical identification of frozen and unfrozen ground, Antarctica. *2nd Int. Conf. on Permafrost, Yakutsk, Russia*, 136-146.

Moorman, B.J. 2005. Glacier-permafrost hydrological interconnectivity: Stagnation Glacier, Bylot Island, Canada. In: C. Harris & J.B. Murton (eds.), *Cryospheric Systems: Glaciers and Permafrost*, London: Geological Society Special Publication, 63-74.

Murton, J.B., Peterson, R. & Ozouf, J.-C. 2006. Bedrock fracture by ice segregation in cold regions. *Science* 314: 1127-1129.

Musil, M., Maurer, H., Green, A.G., Horstmeyer, H., Nitsche, F., Vonder Mühll, D. & Springman, S. 2002. Shallow seismic surveying of an alpine rock glacier. *Geophysics* 67(6): 1701-1710.

Otto, J.C. & Sass, O. 2006. Comparing geophysical methods for talus slope investigations in the Turtmann valley (Swiss Alps). *Geomorphology* 76(3-4): 257-272.

Paterson, W.S.B. 2001. *The Physics of Glaciers.* Oxford: Butterworth & Heinemann.

Sass, O. 2003. Moisture distribution in rock walls derived from 2D-resistivity measurements. *Z. Geomorph. N.F.*, Suppl.-Bd. 132(51-69).

Sass, O. 2004. Rock moisture fluctuations during freeze-thaw cycles: preliminary results from electrical resistivity measurements. *Polar Geography* 28(1): 13-31.

Sass, O. 2005. Rock moisture measurements: techniques, results, and implications for weathering. *Earth Surf. Process. Landforms* 30(3): 359-374.

Schoeneich, P., Hantz, D., Vengon, M., Frayssines, M., Deline, P., Amelot, F. & Savary, J. 2004. *A New Rlpine Rockfall Inventory.* Swiss Geoscience Meeting, Lausanne.

Terzaghi, K. 1962. Stability of steep slopes in hard unweathered rock. *Geotechnique* 12: 251-270.

Weertman, J. 1973. Creep of ice. In: E. Whalley, S.J. Jones & L.W. Gold (eds.), *Physics and Chemistry of Ice*. Ottawa: Royal Soc. of Canada, 320-337.

Wegmann, M. 1998. *Frostdynamik in hochalpinen Felswänden am Beispiel der Region Jungfraujoch - Aletsch.* Ph.D. Thesis. Mitt. Versuchsanst. Wasserbau, Hydrologie und Glaziologie, ETH Zürich 161: 1-144.