DETECTION OF MOUNTAIN PERMAFROST BY
COMBINING HIGH RESOLUTION SURFACE AND
SUBSURFACE INFORMATION – AN EXAMPLE FROM THE
GLATZBACH CATCHMENT, AUSTRIAN ALPS

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ABSTRACT. Permafrost distribution in mid-latitude moun-
tains is strongly controlled by solar radiation, snow cover
and surface characteristics like debris cover. With decreas-
ing elevation these factors have to counterbalance local
positive air temperatures in order to enable permafrost
conditions. We combine high resolution surface data derived
from terrestrial laser scanning with geophysical infor-
mation on the underground conditions using ground pen-
etrating radar and electrical resistivity tomography and
ground surface temperature data in order to understand the
effects of surface characteristics on permafrost distribu-
tion in an Alpine catchment, Austrian Alps (Glatzbach,
47°2′23.49″ N; 12°42′33.24″ E, 2700–2900 m a.s.l.).

Ground ice and permafrost is found above an elevation of
2780 m a.s.l. on north-east facing slopes in 2009, previous
studies detected permafrost at the same site at 2740 m a.s.l.
in 1991. Analysis of surface roughness as a proxy for grain
size distribution reveals that the lower boundary of discon-
tinuous and sporadic permafrost is lowered on rough
surfaces compared to fine-grain zones. At the same location
modelled potential summer solar radiation in coarse grain
zones is reduced by up to 40% compared to surfaces of fine
grain sizes. The mostly patchy permafrost distribution at
the Glatzbach can therefore be attributed to local surface
cover characteristics, particularly regolith grain size and its
influence on solar radiation. We conclude that the analysis
of ground surface characteristics using very high resolu-
tion terrain data supports the assessment of permafrost
in Alpine areas by identifying rough surface conditions
favouring permafrost occurrence.

Key words: Austrian Alps, electrical resistivity tomography,
ground penetrating radar, mountain permafrost, solar radia-
tion, surface roughness, terrestrial laser scanning

Introduction

Permafrost distribution in mid-latitude mountains is strongly controlled by solar radiation, snow cover and surface characteristics like debris cover (Hoelzle 1994; Keller 1994). With decreasing elevation these factors have to counterbalance positive air temperatures in order to enable permafrost conditions. Mountain terrain is often characterized by highly variable geological, geomorphological, climatic and topographic conditions. Mountain topography has a great variability at a wide range of scales between regolith grains to entire moun-
tain ranges. Snow distribution, solar radiation, surface material composition, vegetation cover and other factors influence ground thermal regimes. The factors that are controlled by local topography show a large spatial variability and their influences on ground thermal conditions should be considered in the investigation of permafrost at a local scale. Therefore, subsurface information, like temperature or resistivity, needs to be linked to surface characteristics.

Spatial distribution of mountain permafrost is generally acquired using geophysical techniques, as well as measurements of Bottom Temperature of Snow Cover (BTS) and Ground Surface Tempera-
ture (GST) (Hoelzle 1992; Keller 1994; Ishikawa and Hirakawa 2000; Hauck and Kneisel 2008). Analysing the influence of regolith characteristics on ground thermal conditions has been limited to point measurements in many previous studies, mainly in boreholes and at single blocks (Hanson and Hoelzle 2004; Gruber and Hoelzle 2008; Phillips et al. 2009). On a locale scale, at the lower limit of permafrost, Nyenhuis (2006) investigated
permafrost distribution in a hanging valley in the Turtmanntal, Swiss Alps and considers a mean block size of <50 cm to be unfavourable for permafrost occurrence. A spatial approach to quantify surface characteristics in mountain regions with permafrost occurrence is however necessary to fully understand this factor (Etzelmüller et al. 2001; Nyenhuis 2006). The combination of high resolution subsurface information with detailed surface data can help to understand heterogeneous ground ice distribution in alpine environments, especially at the lower boundary of discontinuous permafrost.

Controlling surface characteristics, mainly grain/block size, on permafrost distribution have been acknowledged by several authors, most of them working on rock glaciers (Ishikawa 2003; Hanson and Hoelzle 2004; Nyenhuis 2006; Lambiel and Pieracci 2008). Block fields which results from rock fall activity, or aggregated by periglacial creep, are considered as accumulators of cold temperatures (Gorbunov et al. 2004). High porosity and low thermal conductivity lead to a significantly different thermal subsurface regime compared to fine-grained surface material. A high surface roughness of blocky material prevents or retards a closed and deep snow cover. Holes in the snow and boulders that stick out of the snow cover enable an increased temperature exchange between the atmosphere and the ground compared to snow covered areas. Consequently, cold winter air is accumulated between coarse blocks. Several processes have been described from blocky surface material that lead to ground cooling for example the Balch effect (cold air sinks down due to higher density), free convection, the chimney effect or the reduction of insulation effect of the snow cover (Wakonigg 1996; Harris and Pedersen 1998; Ishikawa 2003; Gorbunov et al. 2004; Delaloye and Lambiel 2005; Phillips et al. 2009). Gruber and Hoelzle (2008) showed that ground cooling results from a preservation of low temperatures through winter and summer due to the low thermal conductivity of coarse debris. Apparently, coupling between atmosphere and ground is stronger in the cold season than in summer allowing cold temperatures to penetrate the blocky layer through free convection (Herz 2006). These findings suggest that regolith composition is one of the governing factors for discontinuous, sporadic and isolated permafrost distribution in the European Alps.

Topography impacts on the accumulation and duration of snow. Snow deposition and redistribution is mainly influenced by wind conditions, while duration of snow cover depends on snow depth, solar radiation and the local energy balance (Grunewald et al. 2010). Snow cover affects ground temperatures in two ways: Thin early winter snow cover (<15 cm) removes energy from the ground by increased long wave emissivity at the snow surface and increased heat flux from underground (Zhang 2005). This phenomena leads to intense ground cooling in autumn, termed autumn-snow effect by Keller (1994), which is preserved by a deeper snow cover in winter. Thick snow cover represents a thermal insulator due to its low thermal conductivity that shields energy exchange between the ground and the atmosphere (Keller 1994; Ishikawa 2003; Keller and Tamás 2003; Zhang 2005). Snow conditions consequently influence the local permafrost occurrence and can be responsible for local uncertainties in regional permafrost models.

Furthermore, terrain roughness resulting from coarse boulders generates highly variable conditions of shading and solar radiation impact. Consequently, area-wide data on surface material needs to be included in permafrost distribution models, especially with increasing model resolution. High resolution surface data from airborne or terrestrial laser scanning can be used to quantify surface characteristics (Heritage and Milan 2009). The morphometric variable surface roughness describes the variability of surface changes and can be taken as surrogate for grain size composition. Boulders and blocks stick out of the surrounding surface and create an irregular surface which causes high roughness, while small and equally sized grain compositions create smooth surfaces. Roughness is frequently analysed in studies of fluvial dynamics where gravel bed characteristics are studied in detail (Heritage and Milan 2009; Hodge et al. 2009). Computation of surface roughness is scale-dependent according to the size of the object of interest (Grohmann et al. 2011). With respect to grain size (>cobbles, i.e. >128 mm), very high resolution surface data is required that can only be provided by Terrestrial Laser Scanning (TLS) delivering highly accurate data at cm resolution. Various mathematical solutions exist to quantify surface roughness from digital elevation data. An overview and a comparison of different techniques is provided by Grohmann et al. (2011).

In this study we combine geophysical information using Ground Penetrating Radar (GPR) and Electrical Resistivity Tomography (ERT) with high
resolution surface data derived from TLS and GST data, in order to assess and understand local permafrost distribution at the Glatzbach test site, Austrian Alps (Fig. 1). Permafrost occurrence shows a strong correlation with rough surface conditions and reduced modelled solar radiation values. Additionally, the observed permafrost distribution is compared to data on permafrost occurrence from the same catchment published by Rennert (1991) showing an upward shift of around 50 m of the lower permafrost boundary between 1991 and 2009.

Study site
The Glatzbach catchment (47° 2′ 23.49″ N; 12° 42′ 33.24″ E) is located within the central Austrian Alps (Fig. 2a) south of Austria’s highest peak (Großglockner, 3798 m a.s.l.). We have focused our investigation on a north-east facing slope at the northern limit of the catchment (Fig. 2b). Within an elevation range between 2700 and 2900 m a.s.l. over a distance of 500 m, the study site covers the highest slopes and ridges in the north-western part of the Glatzbach catchment. The slope has a mean inclination between 25° and 45° and an overall straight profile with small concave and convex features. As a consequence of the regional lithological setting, the rather shallow topography of the Glatzbach catchment stands out from the surrounding valleys, which depict the typical high mountain imprint of Pleistocene glacial erosion with deep cirques and hanging valleys. The area is situated at the border between the crystalline rocks of the Penninic Tauern window (mainly gneiss and schist) and the Schober Mountains a Palaeozoic formation of mica schist. Lithology at the Glatzbach is part of the “Bündnerschist”, dominated by quartz- and calcareous phyllite, quartzite, as well as dolomitic- and calcareous marble (Höck and Pestal 1994). Phyllites are especially prone to physical and chemical weathering producing fine-grained regolith. In areas dominated by quartzite and dolomitic marble, blocky debris covers the surface. The surface and lithology of the investigated slope is characterized by fine-grained, platy debris of phyllite less than 10 cm in diameter in northern parts and of blocky debris larger than 30 cm in diameter resulting from rockfall events of the dolomitic marble outliers toward the southwest. Towards the southern boundary low bedrock steps exist resulting from dipping of the Penninic rocks towards southwest. Central parts of the catchment hold vegetation covered soils, mainly cambric and calcic cryosol and regosol and shallow podsol of up to 1 m thickness. Numerous periglacial landforms have developed in the Glatzbach area including solifluction lobes and earth hummocks (Rennert 1991). Solifluction has been studied and monitored intensively in the last two decades (Jaesche 1999; Jaesche et al. 2002; Jaesche et al. 2003; Stingl et al. 2010). Climate data for the Glatzbach site have been collected since 1997 by the University of Natural Resources and Life Sciences, Vienna, Austria. A mean annual air temperature between 1997 and 2010 of −1.4°C was recorded at an elevation of 2650 m a.s.l. (Fig. 3). Total annual precipitation is around 1120 mm recorded between 1987 and 1997 (Jaesche et al. 2003) with an assumed 70% of snow precipitation (Böhm et al. 2008). The site is located at the assumed lower boundary of discontinuous permafrost in this region (Ebohon and Schrott 2008), due to its relatively low elevation between 2700 and 2900 m a.s.l.
Previous permafrost investigation around the Glatzbach catchment

The Glatzbach catchment in the central Hohe Tauern range is among the few locations in the Austrian Alps where permafrost conditions and periglacial processes are studied outside of rock glaciers. Here, investigations started more than 20 years ago (Stingl 1971; Veit 1993; Jaesche et al. 2002; Jaesche et al. 2003). Ground movement on a NE-facing slope measured between 1985 and 2008 at two solifluction lobes located at 2650 and 2700 m a.s.l., respectively, reached values up to 200 cm per year with mean values around 2–20 cm per year, (Jaesche 1999; Jaesche et al. 2003; Stingl
et al. 2010). Previous investigation on permafrost distribution and subsurface conditions has been performed around the Glatzbach catchment within a Diploma project in 1991. The unpublished thesis by Rennert (1991) documents the permafrost detection using sediment cores and pits (at 12 locations), BTS-temperatures (36 locations), continuous measurements of ground temperature at different depths up to 2m (12 locations) and geophysical surveying (refractions seismic at 42 locations and 1-D resistivity sounding at 11 locations). The lower boundary of permafrost was determined at 2715 m a.s.l. within the Glatzbach catchment. The study documents a heterogeneous permafrost distribution at the site attributed to locally changing ground surface conditions affecting permafrost occurrence. Based on these observations Rennert (1991) concludes that two main factors control the permafrost conditions in this area: local aspect and regolith characteristics.

Methods, field survey and data
Permafrost detection at the Glatzbach site is based on geophysical field measurements from winter and summer 2009. A total of 15 GST data loggers have been installed at the study site logging ground surface temperature at an interval of 1-hour since 2008 (Fig. 2b). The loggers have been buried within the upper 5 cm of the ground for shielding of direct radiation. We used new UTL-1 and UTL-3 loggers (Universal Temperature Logger, Geotest) with a factory provided accuracy of +/−0.1°C (www.utl.ch). The locations of most of the GST loggers can be viewed on Fig. 2b. For the analysis we used only two loggers that are in close proximity to the geophysical surveys. Snow depths at the two logger sites used have been measured in mid-March 2009 using BTS probes and are also observable in the GPR soundings (Figs 4 and 5).

Ground penetrating radar was applied on the snow covered ground in March 2009. Five GPR profiles were measured with length between 60 and 140 m (Fig. 2b). We used a MALA RAMAC system with a 250 MHz shielded antenna system. The continuous measurement was applied with a station spacing of 5 cm triggered by a string and using a sample frequency of 2553 MHz. Within a time window of 204 ns a total of 32 stacks were obtained. Applying this measurement configuration we can expect a vertical resolution of about 20 cm and a lateral resolution of about 50 cm (for permafrost environment, v ~0.13 m ns⁻¹ at 1 m depth (McQuillin et al. 1984; Trabant 1988). GPR raw data were analysed using REFLEXW software applying standard filter routines, migration and topographic correction (see Appendix 1). Due to
Fig. 4. Radargram of GPR profile 4 (GPR4). (1) Dense pattern of reflections indicating ice lenses (velocity 0.15–0.16 m ns\(^{-1}\), hatched zone), (2) unfrozen zone (velocities 0.11–0.12 m ns\(^{-1}\)), (3) dense reflection patterns at the surface interpreted as seasonal frost layer (velocity 0.15–0.22 m ns\(^{-1}\), stippled zone), (4) snow cover above slope surface, (X) crossover with GPR5.

Fig. 5. Radargram of GPR profile 5 (GPR5). (1) Dense pattern of reflections indicating ice lenses (velocity 0.15–0.16 m ns\(^{-1}\), hatched zone), (2) unfrozen zone (velocities 0.11–0.12 m ns\(^{-1}\)), (3) snow cover above slope surface, (X) crossover with GPR4 and GPR1 (from left). (The seasonal frost layer is less well developed underneath the snow cover compared to GPR4 and cannot be distinguished in the radar data.)

Fig. 6. GPR3 radargram with parabolas representing single objects smaller than the GPR footprint in the given depth. Numbers represent propagation velocities in m ns\(^{-1}\).
the use of the shielded antenna electronics no Common Midpoint (CMP) or Wide Angle Reflection and Refraction (WARR) measurements were made. Because of the vertically and horizontally heterogeneous ground velocity and the lack of accurate CMP and WARR measurements a detailed data migration was not applied. However, in order to assess the penetration depth, we performed an overall migration using a single velocity of 0.15 m ns^{-1} for GPR4 only (Fig. 5). The data interpretation is based on the propagation velocity of the radar waves. The following velocity measurements are based on hyperbola analyses. Winter conditions, subsequent closed snow cover and low subsurface water content, facilitated the application of GPR and significantly improved the data quality. The measurements generated high-resolution, low noise data that also contain the snow depth and the accurate position of the slope topography. In order to compensate for possible uncertainties when using just one indirect prospection technique, the application of additional geophysical methods is recommended (Otto and Sass 2006; Schrott and Sass 2008). Consequently, in July 2009 subsurface resistivity was recorded at two locations using a GeoTom MK8E1000 multi-electrode resistivity system with 25 electrodes and 4 m electrode spacing resulting in profile lengths of 96 m each. ERT was analysed with the Res2DInv software package. However, an exact overlap of the GPR and ERT profiles was not possible due to the different measurement times.

In order to estimate surface roughness and regolith grain size, a 20 cm digital elevation model (DEM) was generated applying TLS. We used a RIEGL LMS-Z620 laser scanner with a factory-provided accuracy of 10 mm and a maximum range of 2000 m. The calculated DEM results were generated by two scans positions, each containing more than 9 million points. The scans were referenced using four fixed reflectors positioned by differential GPS (GPS positioning accuracy: vertical <0.7 m, horizontal: <0.8 m, standard deviation between tie points (scan position 1) and control points (scan position 2): 0.02 m, standard deviation between local coordinates and global coordinate system: 0.39 m). Point data was filtered using a 2.5D minimum filter applying a 20 × 20 cm analysis window in RiSCAN Pro software. This filter separates the lowest values within the analysis window and generated a data set of 1.6 million points that were homogeneously distributed throughout the study area. Point data were exported as x,y,z and interpolated using the topo2raster function (TOPOGRID algorithm) in ArcGIS with a 20 × 20 cm cell size. All spatial data analysis and visualisation was performed with ArcGIS. Serving as a proxy for regolith grain size, surface roughness was quantified using the standard deviation of residual topography (Haneberg et al. 2005; Grohmann et al. 2011). Residual topography results from the difference between the original and a smoothed DEM. We applied a low-pass filter using a 3 × 3 window in order to generate a smoothed DEM. Residual topography represents local variation between the highest and the lowest point in the neighbourhood without being influenced by locale slope. Solar radiation of the study site was modelled using ArcGIS Solar Analyst for the summer period between 15 June and 15 September. We used a sky map of 40 m, simulated a standard overcast sky using diffusion portion of 0.3 and transmissivity of 0.5 for the calculation (ArcGIS settings). The time period was chosen in correspondence to comparable studies showing that during the summer months, when snow cover is reduced completely or to a great extent at this elevation, albedo is low and solar radiation is the most important energy input affecting subsurface conditions (Hoelzle 1994; Schrott 1994).

Results and interpretation

Ground penetrating radar

In total 5 GPR profiles have been acquired in winter 2009 (Fig. 2b). All profiles show a similar pattern of reflections and energy propagation velocities with signals returning from up to 14 m below the snow surface. Beneath the snow cover the topography of the slope is clearly represented by the first strong reflector (Fig. 6). The slope at this location is concavely shaped in the upper section with a snow cover of up to 4 m, followed by a straight section in the lower part, where snow depth decreases to 0.2–1 m (Fig. 4). At the ground surface, all profiles show multiple reflections beneath the snow cover hinting at a boulder surface with voids. Propagation velocity within this upper layer is between 0.18 and 0.23 m ns^{-1} indicating the presence of ice and snow in these voids (cf. Fig. 4). This zone has a thickness of upper 1–2 m which increases where the snow cover decreases (dotted area, marked with 3 in Figs 4 and 5). Due to the propagation velocity that indicates ground ice, we interpret this first layer as seasonal frost at the
surface. The depth of this layer corresponds with previous thermal measurements by Jaesche et al. (2003) who observed winter frost up to depths of about 80 cm at the Glatzschneid. Locally, this dense reflection pattern extends to deeper zones along the slope (hatched area, marked with 1 in Figs 4 and 5). Here, propagation velocities are 0.13–0.16 m ns⁻¹, representing typical values for ground ice (Musil et al. 2002; Berthling and Melvold 2008). These zones are interpreted as active permafrost zones, where seasonal frost reaches the permafrost table. Other areas where energy propagates with velocities between 0.09 and 0.11 m ns⁻¹ (Fig. 6, marked with 2 in Figs 4 and 5) and few reflections are observable are interpreted as unfrozen zones. These are located below the layer of seasonal frost. No distinct boundary between regolith and bedrock can be observed. We assume that due to the high resolution of the GPR measurement the rough bedrock surface is apportioned into diffraction hyperbolas of single edges and cavities and not represented by a continuous reflector. In profile GPR4 two zones of dense reflectors and higher velocities are visible, one in the upper section and another starting at around 90 m downslope of a convex part of the section (Fig. 4). GPR5 profile is placed across the slope starting on the blocky material and running towards southeast which consists of finer grain sizes at the surface (Fig. 5). Reflection patterns are comparable to the downslope profiles revealing a zone of dense reflections from the surface into deeper parts of the slope.

**Electrical resistivity tomography**

In summer 2009 we carried out electrical resistivity soundings close to the zone where GPR has been applied in the previous winter. Two perpendicular profiles have been placed across the coarse grain zone of the slope (Fig. 2b). ERT1 runs downslope from W to E, crossing GPR4 and GPR5 at its lower end. Inversion results show higher resistivity in the central part of the section with resistivity values up to 80 000 Ω m, with its maximum about 5 m below the surface (Fig. 7). This zone of higher resistivity is observable in the upper section of ERT2 as well. However, here maximum values do not exceed 40 000 Ω m. Downslope of this zone in ERT2 resistivity strongly decreases below 5000 Ω m. We interpret resistivity values >10 000 Ω m as ground ice occurrence, following general observations by previous studies (Hauck and Kneisel 2008). Regions of resistivity below 10 000 Ω m can be regarded as non-frozen regolith; lower resistivity values (<1000 Ω m) suggest moisture within the slope, probably due to rainy weather conditions, snow/frost melt or subsurface drainage. The measurement was taken in early summer, before the maximum thickness of the active layer is reached. Since seasonal frost is assumed to penetrate only up to a maximum of 1 m into the ground (see above), we are convinced to identify permafrost in the deeper ground and not remains of seasonal frost. At the time of measurement, active layer thickness in profile ERT1 is of a few decimetres while in profile ERT2 a maximum of 2 m of the surface shows resistivity below 10 000 Ω m and appears to be unfrozen (Fig. 7).

**Ground surface temperature records**

Assuming a thick winter snow cover (>1 m), which has not been constantly monitored, we can interpret the GST data similar to BTS measurements, looking at GST below snow in late winter: Five out of 15 loggers have measured mean temperatures below −3°C in late winter with minor temperature changes in February and March, indicating permafrost conditions based on the BTS principle. Another two loggers recorded mean temperatures between −2 and −3°C in this period, which can be interpreted as possible permafrost occurrence, while the remaining loggers measured temperatures between 0 and −2°C and do not allow for permafrost interpretation. Loggers L1 and L2 have been placed in late August 2008 in close proximity to the GPR survey lines (Fig. 2b) documenting the ground surface temperature development in winter 2008/2009 (Fig. 8). Unfortunately, the local climate station failed between late-October 2008 and late-March 2009, preventing a comparison of GST and air temperatures throughout the winter.

GST dropped to almost constant 0°C between 16 September and 18 October (Figs 8a, b). October 2008 ended in a mild period with positive daytime air temperatures around 5°C and around 0°C at night. The 0°C constant indicates the first shallow, short term snow cover in early autumn, where latent heat release generates the so called zero curtain effect. GST dropped sharply below −5°C at L1 in mid-November, while at L2 temperature slowly decreased towards −5°C not until mid-January (Fig. 8a, b). Winter GST at L1 reveals short term fluctuations between −2.5 and −8.5°C (Fig. 8a). In contrast, L2 shows fewer fluctuations...
until end of December and a slower decrease of temperatures between -0.5 and -3°C (Fig. 8b). From early February 2009 to April 2009 both curves depict symmetrical GST fluctuations, however, with deeper absolute values at L1. Logger L1 was covered by approximately 1.5 m of snow, while L2 had at least 3–4 m of snow cover at that time. As of the beginning of April, snow melt produces the zero curtain effect producing a constant GST of 0°C. Snow melt lasted until 25 May and 21 May at L1 and L2, respectively. L1 was positioned on a coarse grain surface, while surface material underneath L2 is finer. To conclude, we interpret both logger locations as permafrost sites due to  

Fig. 7. Intersecting ERT profiles at Glatzach site. Profile lengths of both measurements were 96 m with a 4 m electrode spacing. The bold black line marks the 10,000 Ωm contour line.
mean temperatures below \(-3°C\) in late winter (February-March). Higher surface roughness at site L1 may have prevented a continuous snow cover in early winter leading to strong downcooling, larger temperature changes and lower maximum temperatures compared to L2, where winter temperatures more slowly adapt to subsurface thermal conditions. However, we acknowledge that due to the large fluctuation within the data we cannot exclude ventilation effect and influence of air temperature on the GST and thus GST data does not provide a highly certain permafrost indicator here.

**High resolution terrain analysis**

In order to analyse influences of surface characteristics on the permafrost distribution we quantified morphometric variables and potential solar radiation for the study site using a 20 cm DEM.
Quantification of surface roughness using the standard deviation of residual topography within an analysis window of 60×60 cm displays the location of boulders when compared with digital aerial photographs (compare Fig. 2b). Roughness values are around 0.01–0.02 m in smooth, fine-grained terrain, whereas coarse grained terrain shows values between 0.03 and 1 m (Fig. 9a). Solar radiation input on the study slope during the summer period (15 June–15 September) was modelled between less than 20 kWh m⁻² on the steeper inclined upper parts of the slope and more than 50 kWh m⁻² on southern slopes and flat areas (Fig. 9b). The impact of large boulders on the energy input is clearly reflected in the modelled values of solar radiation. In areas with large boulders, hence high surface roughness, solar radiation values vary between around 50 kWh m⁻² at the illuminated sides of boulders and less than 30 kWh m⁻² on the shaded side. Due to local shading effects of large boulders, modelled energy input is reduced by up to 40% on rough surfaces compared to surfaces covered by fines (Fig. 9b).

**Discussion**

Based on the field data we could detect and accurately localize permafrost occurrence at the Glatzbach site on the northeast facing slope above 2780 m a.s.l. However, GPR4 shows that some sporadic ice lenses can also occur below this zone. High resolution terrain analysis indicates that small scale variations of the surface are responsible for the permafrost occurrence. Figure 10 depicts a comparison of solar radiation, slope, roughness and snow depth along profile GPR4. The curves of surface roughness and solar radiation vary almost diametrically opposed indicating the negative influence of large boulders on the energy input. In contrast, surface roughness and slope angles show similar patterns. Modelled solar radiation along the profile ranges from 20 to 50 kWh m⁻² for the summer period. In the underlying GPR image, two zones of dense reflections patterns indicate the occurrence of subsurface ground ice. Their location corresponds well to areas of lower solar radiation and increased surface roughness at the surface. The upper zone of low solar radiation is influenced by a steep inclination, the general orientation towards NE and shading effects by the proximate steep rock wall towards southwest (Fig. 9b). The influence of shading effects can have significant impact on local permafrost occurrence in mountain regions (Schrott 1994; 1996). With increasing distance from the ridge solar radiation increases towards a first plateau of high energy between 53 and 62 m where solar radiation surpasses 50 kWh m⁻². The observed zone of reflections in GPR4 corresponds also to a convex part of the slope with greatest snow depths. At 80 m, ground ice occurrence falls into a second zone of reduced solar radiation and increased surface roughness (Fig. 10).

Modelling the solar radiation at this resolution enables to identify local surface influences on the distribution of permafrost. Single large boulders as well as a cluster of larger blocks on a slope reduced overall solar input through local shading effects, leading to cooler surface conditions in shaded areas and higher insolation on sun-facing parts of the boulders. This effect may either add to the influence of slope angle and aspect, or may counterbalance them. With respect to permafrost formation, decreased solar radiation on coarse-grained surfaces may lower the limit of permafrost compared to locations of similar elevation and aspect with fine-grain regolith. In case of GPR4 we observed that surface conditions enable the occurrence of isolated ground ice about 10 m lower compared to the potentially continuous occurrence in the upslope area. However, it needs to be verified to what extend the other potential influences on permafrost caused by surface boulders affect the permafrost distribution here. We can mostly exclude air circulation between the rocks since large amounts of fine material fill the voids. On the other hand, large blocks that stick out of the winter snow surface may influence the local conditions here as well. Additionally, it needs to be analysed to what extend increased insolation on the one side of the blocks counterbalances the shading effect on the other side. This would probably require some modelling to estimate this effect and is surely depending on the size of the block.

**Comparing permafrost distribution between 1991 and 2009**

In 1991 Rennert mapped permafrost distribution at the Glatzbach catchment (cf. Fig. 9b). Based on 1D-electrical resistivity, refraction seismic, BTS and ground temperature measurements, and pits he estimated the lower limit of active permafrost at 2740 m a.s.l. on the easterly and north-easterly part of the slope (Rennert 1991). Based on our investigation we detect permafrost at elevation
Fig. 9. (a) Map of surface roughness expressed as standard deviation of residual topography (3 × 3 window). Coarse blocks stick out in green compared to low smooth surfaces in light brown. (b) Map of solar radiation input between 15 June and 15 September (kWh m⁻²) based on a 20 cm DEM. Lowest values are modelled for the steep rock faces oriented in northern directions and for the block field in the centre of the slope. Highest values are observed on south facing slopes.
above 2790 m a.s.l. limited to steep parts of the north-easterly exposed slope and on surfaces characterized by coarse regolith (Fig. 9b). Even though we cannot judge the accuracy of the previous study by Rennert, we can assume a rise of permafrost of around 50 m in elevation between 1991 and 2009. Between 1997 and 2010 mean annual air temperature rose from \(-2.1^\circ C\) (1997–2003) to \(-1.2^\circ C\) (2004–2010).

Conclusions
The Glatzbach site in the central Eastern Alps of Europe can be considered as an area with sporadic and discontinuous permafrost occurrence. Ground ice and permafrost is found above an elevation of 2800 m a.s.l. on slopes facing north-east. The strongly varying pattern of permafrost distribution at the Glatzbach results from local surface cover characteristics, mostly influenced by regolith grain size and variation of solar radiation. Analysis of surface roughness as a proxy for grain size distribution using very high resolution (grid resolution 20 × 20 cm) terrestrial laser scan data reveals that the lower boundary of discontinuous and sporadic permafrost is lowered on rough surfaces compared to fine-grain zones. At the same location modelled potential summer solar radiation in coarse grain zones is reduced by 40% compared to surfaces of fine grain sizes and between the illuminated and the shaded side of the boulders. This effect adds up to previously considered influences of coarse blocks on ground thermal conditions including snow cover development and air circulation that in combination lead to localized ground cooling and permafrost conditions despite positive air temperatures.

The study shows that a combination of high resolution geophysical information with very high resolution surface information significantly supports the understanding of local permafrost conditions in alpine terrain. Digital elevation data derived from laser scanning, especially from terrestrial surveys provides sufficiently detailed surface data in order to analyse local regolith conditions. Modelling of permafrost conditions on large scales therefore should include the analysis of surface conditions in order to consider surface influences on the local permafrost distribution. It has to be mentioned however, that special attention is required concerning accurate positioning of the surface and subsurface data in order to provide precise data overlap. This includes the application of GPS for surveying of geophysical profiles.

Fig. 10. Comparing surface roughness, solar radiation, slope and snow depth at 15 March 2009 (thick dashed line) along profile GPR4. Curves of solar radiation, surface roughness and local slope have been smoothed using a running average of 3 m for better visualization. Permafrost areas are represented by dense reflection patterns in the radargram (compare to Fig. 4).
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**Appendix 1**

*Filter routines applied to GPR radar (in REFLEXW):*

- subtract-mean (dewow) / 4 / 0 / 0 / 0 / 0 / 1 / 1 / 1 / 2800
- static correction / 0 / 0 / 0 / 0 / 0 / 0 / 0 / 0 / 0 / 2800
- energy decay / 0.6 / 0 / 0 / 0 / 0 / 0 / 0 / 1 / 1 / 2800
- bandpass frequency / 100 / 150 / 450 / 500 / 0 / 0 / 1 / 1 / 2800
- background removal / 0 / 196.5943 / 0 / 140.4622 / 0 / 1 / 1 / 2800
- fk migration (Stolt) / 1 / 0.15 / 0 / 196.5943 / 0 / 0 / 0 / 0 / 0 / 2800
- correct 3D topography / 0 / 0 / 0 / 0 / 2844.52 / 0.3 / 0 / 0 / 0 / 0