Application of field geophysics in geomorphology: Advances and limitations exemplified by case studies

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Received 30 December 2005; accepted 27 December 2006
Available online 10 May 2007

Abstract

During the last decade, the use of geophysical techniques has become popular in many geomorphological studies. However, the correct handling of geophysical instruments and the subsequent processing of the data they yield are difficult tasks. Furthermore, the description and interpretation of geomorphological settings to which they apply can significantly influence the data gathering and subsequent modelling procedure (e.g. achieving a maximum depth of 30 m requires a certain profile length and geophone spacing or a particular frequency of antenna). For more than three decades geophysical techniques have been successfully applied, for example, in permafrost studies. However, in many cases complex or more heterogeneous subsurface structures could not be adequately interpreted due to limited computer facilities and time consuming calculations. As a result of recent technical improvements, geophysical techniques have been applied to a wider spectrum of geomorphological and geological settings. This paper aims to present some examples of geomorphological studies that demonstrate the powerful integration of geophysical techniques and highlight some of the limitations of these techniques. A focus has been given to the three most frequently used techniques in geomorphology to date, namely ground-penetrating radar, seismic refraction and DC resistivity. Promising applications are reported for a broad range of landforms and environments, such as talus slopes, block fields, landslides, complex valley fill deposits, karst and loess covered landforms. A qualitative assessment highlights suitable landforms and environments. The techniques can help to answer yet unsolved questions in geomorphological research regarding for example sediment thickness and internal structures. However, based on case studies it can be shown that the use of a single geophysical technique or a single interpretation tool is not recommended for many geomorphological surface and subsurface conditions as this may lead to significant errors in interpretation. Because of changing physical properties of the subsurface material (e.g. sediment, water content) in many cases only a combination of two or sometimes even three geophysical methods gives sufficient insight to avoid serious misinterpretation. A “good practice guide” has been framed that provides recommendations to enable the successful application of three important geophysical methods in geomorphology and to help users avoid making serious mistakes.

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Keywords: Ground-penetrating radar; DC resistivity; Seismic refraction; Landforms; Internal structure

1. Introduction

Recently, the use of geophysical techniques has become increasingly important in many geomorphological studies. Without the application of geophysics our knowledge about subsurface structures, especially over...
larger areas, remains extremely limited. During the last decade the use of geophysical techniques has become a new and exciting tool for many geomorphologists. One reason for the increasing interest in geophysical field methods is certainly related to technical innovations as increased computer power and the availability of lightweight equipment allows for relatively user-friendly, efficient and non-destructive data gathering. However, the correct handling of the geophysical instruments and subsequent data processing are still difficult tasks and the methods often require advanced mathematical treatment for interpretation. It should be pointed out that without close collaboration between geomorphologists and geophysicists the accurate and effective use of geophysical techniques and their geophysical and geomorphological interpretation are often very limited. In addition, the correct description and interpretation of geomorphological settings and thus, the choice of adequate and meaningful field sites are not an easy task.

In most studies and textbooks, interdisciplinary aspects combining geophysics and geomorphology are poorly addressed. Many textbooks, in their description of geophysical surveying applications, focus on the exploration for fossil fuels and mineral deposits, underground water supplies, engineering site and archaeological investigation (e.g. Reynolds, 1997; Kearey et al., 2002). Although several potential applications for geophysical methods exist, many of them have not yet been fully integrated into geomorphological research. The most common geophysical applications are currently focusing on permafrost mapping, sediment thickness determination of talus slopes, block fields, alluvial fans and, increasingly, on the depth and internal structures of landslides (Hecht, 2000; Tavkhelidze et al., 2000; Hauck, 2001; Hoffmann and Schrott, 2002; Hauck and Vonder Mühll, 2003; Israil and Pachauri, 2003; Kneisel and Hauck, 2003; Schrott et al., 2003; Bichler et al., 2004; Sass et al., 2007-this issue). Other landforms such as karst and colluvia are comparatively rarely investigated (Hecht, 2003). Currently, the most common geophysical methods in geomorphological research are ground-penetrating radar, DC resistivity, and seismic refraction (Gilbert, 1999). Thus, the paper focuses on typical applications of these methods.

Each geophysical technique is based on the interpretation of contrasts in specific physical properties of the subsurface (e.g. dielectric constant, electrical conductivity, density). The type of physical property to which a particular geophysical method responds determines and limits the range of applications. As non-geophysicists cannot generally be aware of all limitations and pitfalls, there is a need to develop a set of the most suitable recipes combining geophysical methods. These methods can then be adjusted to particular environmental conditions and landforms. Studies that compare the application of various interpretation tools and discuss the combined/composite applications of geophysical field methods for a particular landform are still rare (Schrott et al., 2000; Schwamborn et al., 2002; Otto and Sass, 2006; Sass, 2006a,b).

Thus, the objectives of the present paper are:

(i) to show and assess the advances and limitations of different geophysical methods with the focus on applied geomorphological research;
(ii) to demonstrate the application of these methods in different natural environments and for distinctive landform types, and
(iii) to illustrate the advantages of combined and composite techniques leading to a set of recommendations for the application of field geophysics in geomorphology.

As the paper focuses on the potential application of three geophysical methods in different environments, the site descriptions have been reduced to a necessary minimum.

2. Ground-penetrating radar (GPR)

2.1. Principle and geomorphic context

Ground-penetrating radar is a technique that uses high-frequency electromagnetic waves to acquire information on subsurface composition. The electromagnetic pulse is emitted from a transmitter antenna and propagates through the subsurface at a velocity determined by the dielectric properties of the subsurface materials. The pulse is reflected by inhomogeneities and layer boundaries and is received by a second antenna after a measured travel time. In order to calculate depth, wide-angle reflection and refraction (WARR) or common mid-point (CMP) measurements have to be performed. These signal travel time measurements are made with a stepwise increase in the distance between the two antennas. From the distance/travel time diagram, the propagation velocity of the radar waves in the subsurface can be derived.

The common mode of GPR data collection is fixed-offset reflection profiling (Jol and Bristow, 2003). In this step-like procedure, the antennas are moved along a profile line and the measurement is repeated at discrete intervals resulting in a 2-D image of the subsurface. The possible working frequencies can range from 10 MHz to 1 GHz depending upon the aim of the investigation. Higher frequencies allow higher spatial resolution of the ground information, but lead to a lower penetration depth.
This has important implications in geomorphological applications. Knowledge about the geomorphological context (expected maximum depth and grain-size composition of a sediment body) is essential for choosing the appropriate frequencies. A comprehensive “good practice guide” for the application of GPR in sediments is provided by Jol and Bristow (2003).

The maximum depth of investigation depends mainly upon the dielectric constant ($\varepsilon$) and the electrical conductivity ($\sigma$) of the subsurface. A higher ground water and/or clay content (high $\varepsilon$, high $\sigma$) leads to a stronger attenuation and, therefore, a markedly reduced penetration depth. However, very pure groundwaters that have a low conductivity (e.g. glacial meltwater, bogs) are characterized by relatively low levels of GPR signal loss. Again, geomorphological expertise about possible water and clay layers can help to avoid disappointing measurements. On dry and electrically high-resistive debris, a penetration of between 30 and 60 m can be achieved (e.g. Smith and Jol, 1995). Sandy sediments are also favourable for GPR measurements at depths of between 15 and 30 m. Due to the strong attenuation in materials that have a high electrical conductivity the penetration depth in wet, silty and/or clayey sediments diminishes rapidly to, for example, less than 5 m in silt (Doolittle and Collins, 1995); in clayey soil the application of GPR may be altogether impossible (see Table 1). This is, however, only a very rough guide, because the penetration depth depends upon the device and antenna frequency used. The vertical resolution of GPR data is a function of frequency and propagation velocity. With higher velocities, resolution decreases and vice versa. As a rule of thumb, in the medium velocity range (0.1 m/ns) the resolution is approximately 1 m using 25 MHz antennae, 0.25 m using 100 MHz and 2.5 cm using 1 GHz.

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2.2. Advantages and disadvantages of GPR

Even when low-frequency 25 and 50 MHz antennas are used, GPR still provides a better spatial resolution than standard geophysical techniques. The survey speed even in rough terrain is relatively high and several hundred metres per day are possible. Steep and rocky slopes limit the progress of the survey, because low-frequency antennas have large geometrical dimensions and are difficult to handle. Recently developed, so-called, Rough Terrain Antennas (RTAs) may facilitate data gathering, however, conventional antennas are still necessary for CMP measurements or small object detection.

The available antenna frequencies allow for a broad variety of possible applications. However, the extremely variable penetration depth requires careful assessment of the subsurface parameters in the study area in order to minimize the risk of error. The electrical conductivity of the soil provides a rough indicator of potential target depth. From the authors’ experience, the use of GPR is not promising if the soil resistivity is lower than ca. 50–100 Ωm.

The application of GPR is subject to further restrictions. As the reflectivity at a layer boundary is determined by the contrast in the dielectrical properties of the subsurface units, no distinct reflection is found when this contrast is low. Small-scale spatial differences in water content and/or grain-size composition may yield stronger reflections than the target of the investigation (e.g. the bedrock surface). This problem may be overcome by using the radar facies of the sediments for interpretation. Different sediment units and bedrock yield typical reflection patterns that can be derived from, for example, reference profiles. The reconnaissance of these patterns significantly facilitates the interpretation of the radargram.

A portion of the energy transmitted by an unshielded antenna is emitted into the air and may be reflected by features above the surface. These air wave reflections cannot always be unequivocally distinguished from ground information and may severely affect the data quality. This makes the application of GPR particularly difficult in wooded terrain where, depending on the frequency used, each tree may act as a single reflector, leading to very noisy or altogether useless data. Although the effect of air wave reflections may be reduced using sophisticated filter algorithms (e.g. Van der Kruk and Slob, 2004), measuring in forested areas is not advisable. The use of shielded antennas is only possible for higher frequencies (> 100 MHz). Taking into consideration the major restrictions arising from “clayey or silty subsurface” and “wooded terrain” it is clear that GPR shows its potential particularly in arctic or alpine areas above the tree-line and where there is limited soil development. However, shallow subsurface investigations are also possible in fluvial deposits and even in peat, when the electrical conductivity of the groundwater is low.

2.3. Examples of application

There is a broad range of successful applications of GPR in geomorphological studies. These include the detection of buried structures, assessment of internal sediment structures and estimation of depth to bedrock. Various types of sediments have been investigated for geomorphological purposes (Bristow et al., 2000).

The internal structures of floodplain deposits and deltaic sediments have been visualized by, for example, Leclerc and Hickin (1997), Jol (1996) and Büker et al. (1996). Buried fluvial channels have been detected by Roberts et al. (1997) and Loope et al. (2004). The most frequently used antenna frequency for comparative studies is 100 MHz. The penetration depth is dependant upon clay and water content, but usually ranges from 10 to 20 m.

Loose sediments in arctic and alpine areas have also been the subject of GPR measurements. Lønne and Lauritsen (1996) and Overgaard and Jakobsen (2001) have investigated internal deformation structures of push-moraines and Berthling et al. (2000) clearly detected internal structures as well as the bedrock base of rock glaciers. The working groups used 50 and 100 MHz antennas and achieved a penetration depth of up to 30 m. Sass and Wollny (2001) and Sass (2006a) achieved a penetration depth of up to 50 m on talus slopes using 25 MHz antennas. They found surface-parallel structures in the debris body and evidence for moraine material at the base of the talus. Studies of sediment structures have also frequently been carried out in bogs. Vökel et al. (2001) investigated the subsurface structure of buried periglacial slope deposits, while e.g. Holden et al. (2002) determined the position and depth of subsurface piping.

The application of GPR on landslides has repeatedly been tested but with limited success. Wollny (1999) investigated the near-surface moisture distribution at 10 landslide areas and gained valuable information from only three sites. Bruno and Mariller (2000) and Wetzel et al. (2006) used GPR for detecting the vertical extension of landslides but did not reach the slip surface even with the use of low-frequency antennas. However, internal slide structures such as rotational features can be mapped when the slide surface is comparatively dry (e.g. Sass et al. (this issue) in displaced limestone blocks). Bichler et al. (2004) obtained very good results, distinguishing seven different facies of loose sediments. Performing GPR measurements at landslide sites is only worthwhile when comparatively...
coarse and dry deposits (debris, displaced blocks) superimpose the silty or clayey material of the slip surface. However, detailed assessment of the near-surface propagation velocity allows a rather accurate estimate of the soil moisture content (Topp et al., 1980) which is of interest for landslide investigation and many more geomorphological questions.

Another possible field of application is the investigation of permafrost features. The active layer thickness has been determined, for example, by Arcone et al. (1998) and Hinkel et al. (2001). Moorman et al. (2003) provided instructive pictures of typical reflection patterns in frozen and unfrozen ground. Although the presence or non-presence of permafrost is more difficult to establish with GPR than electrical resistivity techniques, GPR is superior in detecting spatially confined structures such as ice wedges (Hinkel et al., 2001). The thickness and the internal structure of glacier ice (unfrozen water content, cavities) have also been the target of many GPR investigations (e.g. Moorman and Michel, 2000).

The investigation of quasi point-shaped or linear buried structures in high resolution is the aim of many studies in the relatively new field of geo-archaeology that is closely related to geomorphology (Baker et al., 1997; Fuchs and Zöller, 2006). Leckebusch (2003) has provided a detailed description of the GPR method for archaeological purposes with numerous examples. The working frequencies for these applications are usually rather high (=200 MHz); the target depth is usually between 1 and 5 m.

2.3.1. Case study: localization of buried structures

The aim of this GPR application was to locate a Roman road buried under 1.5 to 3 m of peat and fluviatile sediments in the Murnauer Moos, Upper Bavaria (Sass et al., 2004). This example primarily highlights the potential for archaeological studies (Fuchs and Hruska, 1996). However, from a geomorphological perspective the Roman road can be used as a time marker for the evolution of the overlaying peat and interbedded clay-rich sediments. The road had been examined at an archaeological excavation nearby. Thus, the structure of the road (logs lying on a gravel layer) was known. The objective of the measurements was the quick localization in the surroundings of the excavation.

As a result of the high water content of the peat, the propagation velocity derived from WARR measurements was very low (0.45 m/ns). However, because of the low mineralization of the bog water the penetration depth using 200 MHz antennae was up to 5 m. The road was quickly and clearly located in a number of cross profiles (Fig. 1). Thus, the measurements provided valuable information on the straight-line structure. The cross-profile presented (Fig. 1a) shows the radar reflection of the Roman road at a profile distance of between 3 and 7 m. The shallow depression in the middle of the road (as observed at the excavation site), caused by the weight of the vehicles, can be clearly recognized. The longitudinal profile (Fig. 1b) illustrates the linear structure of the road. The main reflection is caused by the artificial gravel layer which obviously shows a distinct dielectrical contrast to...
the surrounding peat. The V-shaped structures below (e.g. at profile distances of between 10 and 14 and between 80 and 100 ns) are the hyperbolic reflection patterns of singular logs lying at right angles to the course of the road. From this typical pattern the road can be unequivocally distinguished from geological reflectors.

In a series of cross profiles, the road was located even in positions where borings failed to detect the structure, probably due to the advanced rotting of the logs. However, alluvial layers near the surface considerably reduced the penetration depth due to high attenuation in the clay. In the vicinity of some shallow alluvial cones, the penetration depth was reduced to less than 1 m which made the detection of any structures impossible.

2.3.2. Case study: talus sediment thickness

Fig. 2 shows the 50 MHz longitudinal section of a talus slope in the Kühtai area (Central Alps, Austria) at an elevation of between 2400 and 2600 m. The profile stretched from a moraine ridge at the foot of the talus upwards (inclined approx. 30°) to the gneiss/mica-schist rockwall (inclined at 55 to 60°). In the upslope parts of the profile (approximately 70 to 130 m), the debris is rather fine-grained (5 to 20 cm) and gradually gets coarser towards the base of the slope, where boulder-sized clasts prevail.

The bedrock surface is recognizable as a distinct reflector (highlighted by the thick dashed line) revealing a shallow basin (see Fig. 2). The part of the talus body close to the rockwall (between 60 and 110 m) is characterized by surface-parallel, slightly undulated and partly overlapping reflection patterns probably indicating stacked sediment layers. No comparable layers can be observed near the foot of the talus (0–60 m). This points to fluvial distribution that is restricted to the vicinity of the moderately steep rock face. The maximum thickness of the debris is 12 m in the middle of the profile. In the deeper subsurface, thin lines are visible which are slanting diagonally to the left. Considering the inclination of the profile, these lines represent reflectors dipping at an angle of between 60 to 70° to the south (left side). This agrees with the dipping of the gneiss–micaschist layers of the adjacent rockwall. The occurrence of airborne (rockwall) reflections is improbable because of the position and angle of the reflectors, together with the low inclination of the rockwall above.

3. 1-D and 2-D DC resistivity

3.1. Principle and geomorphic context

Geoelectrical sounding provides a one-dimensional vertical profile of the electrical resistivity distribution with depth. Resistivity measurements are conducted by applying a constant current into the ground through two “current electrodes” and measuring the resulting voltage differences at two “potential electrodes”. From the current and voltage values, an apparent resistivity value is calculated. 1-D surveys use only four electrodes. Generally, the two potential electrodes remain in position in the middle of the profile, while the current electrodes move stepwise to both sides (Schlumberger array). The wider the electrodes are apart, the deeper the electrical field penetrates into the ground. Thus, a depth profile under the approximate middle of the section is created. This array has been widely used in geomorphic research (Etzelmüller et al., 2003). 2-D arrays represent a further development of the 1-D technique, using 50 or more electrodes at a time. A micro-controller unit automatically switches between numerous electrode configurations, thus creating a 2-D pseudosection of the subsurface. The inversion of the gathered data using sophisticated computer programs produces a 2-D section of the subsurface resistivity. The Res2Dinv-software provided by Loke and Barker (1995) is the most commonly used in geosciences. The principle of operation is based on a stepwise iteration process which tries to minimize the deviation between the measured apparent resistivity and
the simulated apparent resistivity values calculated from a subsurface model. The 2-D section can also be conducted in a Schlumberger array, which provides particularly good resolution for lateral inhomogeneities. The Wenner array shows a good signal–noise ratio and is favourable for the detection of horizontal layers, while the Dipole–Dipole array is favourable for the delimitation of spatially confined objects in the shallow subsurface. For a comprehensive description of these configurations we refer to geophysical textbooks (Milsom, 1996; Reynolds, 1997; Kearey et al., 2002).

The maximum amount of *a priori* information on the geomorphological context should be obtained (e.g. layer thickness, bedrock type and expected resistivity value) before starting the inversion of the raw data. The *a priori* estimate (e.g. maximum resistivity value of an expected type of sediment) can be set as a fixed parameter and helps to improve the model. The required information can be derived from test profiles of, for example, bedrock or sediment units of known composition.

### 3.2. Advantages and disadvantages

A great advantage of the method is the high variability of electrode spacing and configuration. The distances between the electrodes can range from some centimetres to several tens or hundreds of meters, allowing a penetration between the electrodes can range from some centimetres to tens of electrode spacing and configuration. The distances

![Image](https://example.com/image)

... (Beauvais et al., 2003). However, the two main groups of current applications in geomorphology comprise the detection and characterization of permafrost and the investigation of landslides. Employing 2-D resistivity profiling for the detection of permafrost is highly advisable because of the very strong electrical contrast between ice and almost all other common substrates. Evidence for frozen ground in permafrost areas using 1-D resistivity profiling has been found, for example, by Assier et al. (1996) and Ishikawa and Hirakawa (2000). The active layer thickness in mountain permafrost has been assessed (e.g. by Gardaz, 1997), while Isaksen et al. (2000) have measured the thickness of the ice layer of rock glaciers in Svalbard. 2-D resistivity sections performed, for example, by Kneisel (2003) and Kneisel and Hauck (2003) clearly show the lateral extension of ice lenses in alpine rock glaciers and scree slopes. A range of possible applications in permafrost areas has recently been presented by Kneisel (2006).

Numerous papers have dealt with the assessment of the depth and the detection of the structures of landslides. The target depth of the investigations can be quite variable. The slip surface of a very shallow landslide was detected by Wetzel et al. (2006) using 2-D profiling to a depth of between 10 and 15 m, while Perrone et al. (2004) achieved a penetration depth of 80 m using survey lines of up to 600 m. Agnesi et al. (2005) reached a penetration depth of almost 200 m in a 1-D survey of a deep-seated landslide in Italy. Bichler et al. (2004) combined numerous 2-D profiles to a quasi three-dimensional picture of the landslide subsurface. Mauritisch et al. (2000), Godio and Bottino (2001) and Bichler et al. (2004) combined the geoelectric measurements with further methods (electromagnetic profiling, seismic refraction) to improve the interpretation of the combined data. Suzuki and Higashi (2001) demonstrated the effectiveness of 2-D resistivity for long-term investigations monitoring the infiltration of rainfall into a landslide body in numerous time slices.

Kneisel (2003) provides some further examples for the investigation of depth to bedrock such as the assessment of the thickness of aeolian sediments. Sass (2006a,b) and Otto and Sass (2006) applied Electrical Resistivity Tomography (ERT) to the measurement of...
talus thickness. A rather unusual application is the very small-scale monitoring of water movement during freezing, as performed by Sass (2004).

3.3.1. Case study: detection of frozen rock

2-D resistivity profiling was used for the monitoring of freeze–thaw cycles in different rock types at various test sites using 50 small electrodes inserted neatly into drilled holes with a spacing of 4 cm (Sass, 2003, 2004). The profile line extended diagonally over a jointed limestone outcrop. During the night, temperatures dropped to $-4.0\, ^\circ C$ at the surface and $-2.6\, ^\circ C$ 15 cm beneath the surface. 2-D resistivity measurements were carried out hourly using the Wenner array.

Fig. 3a shows the resistivity profile of the unfrozen rock. In Fig. 3b, the frozen areas of the rock are visible as confined patches of very high resistivity ($\geq 15\, k\Omega m$). In the course of freezing, near-surface ice formations occurred over the entire survey line, while the water content in the deeper parts of the rock slowly increased due to displaced pore water. The comparatively high RMS error of 15.9% in the inversion model is caused by the extremely sharp resistivity contrasts in the vicinity of the frozen areas. A minimum temperature of $-2.6\, ^\circ C$ was reached at a depth of 15 cm at 06:30. However, the entire rock body below approximately 10 cm remained unfrozen, which points to supercooling due to hydrostatic pressure under the superficial ice formation. The measurement highlights the generation of hydraulic pressure as a possible agent of frost weathering.

3.3.2. Case study: internal structure of a landslide

The Schliersberg slope is situated at the northern edge of the Bavarian Alps. Flysch and Ultrahelvetikum rock series prevail (sandstone and claystone) which are very prone to landslide processes. The most prominent geological feature is the overthrust fault of the Flysch nappe (here represented by comparatively stable sandstone) on the underlying clayey Ultrahelvetikum series. The slope is rather steep in the area of outcropping sandstone (up to 40°) and only moderately inclined in the clayey rock series (20–25°). A very active combined landslide has developed in the Ultrahelvetikum series which has caused considerable damage to the spruce forest and forestry roads. The slope beside the active slide shows obvious signs of former rotational slide movements on the surface such as ridges and damming wetness in hollows but is obviously stable today. Extensive 2-D resistivity profiling was carried out in order to highlight the structure and depth of the slide. The longitudinal profile presented in Fig. 4 runs parallel to the active slide at the currently stable slope, while the cross profile extends across both the inactive and active zones.

The contrast between the electrical properties of these series is clearly to be seen in the upper part of the longitudinal profile. The Flysch overthrust fault dips steeply to the south. Further down the slope the structure of a multiple rotational slide can be recognized. The landslide blocks consist of displaced, partly mingled sandstone and mudstone. The slip surface is at the top of the native mudstone bedrock. A step in the longitudinal
profile of the slope (64–80 m) is obviously due to a resistant layer of the Ultrahelvetikum bedrock. The cross profile shows the comparatively dry, inactive rotational blocks on the left, while the active combined landslide on the right consists of wet, clayey substrate and displays a very low resistivity. The conchiform slip surface is highlighted by the white dashed line. A displaced rotated block is embedded in the landslide near the surface. The 2-D profiles aid the recognition of the type of movement and clearly show the zones of rotational and translational features possibly associated with changing depths to the slip surface.

4. Seismic refraction

4.1. Principle and geomorphic context

The principle of seismic refraction is based on elastic waves travelling through different subsurface materials, such as sand, gravel, and bedrock, at different velocities. The denser the material, the faster the waves travel. A prerequisite for the successful application is that each successive underlying layer of sediment or bedrock must increase in density and therefore, velocity. Two types of seismic waves can travel through the subsurface: longitudinal or primary (P) waves which are characterized by deformation parallel to the direction of wave propagation, and transverse or secondary (S) waves where particle motion takes place at right angles to the direction of wave propagation.

The propagation of seismic waves through layered ground is determined by the reflection and refraction of the waves at the interface between different layers. When the wave reaches the interface some energy of the wave is refracted into the deeper layer, while the reflected wave transmits the energy back into the overburden layer. The angle of incidence of the reflected wave is equal to the angle of reflection, independent of the seismic velocities of the layers. However the angle of refraction follows Snell’s law:

\[
\sin \theta_i \sin \theta_r = V_1 / V_2
\]

where \( \theta_i \) is the angle of incidence, \( \theta_r \) is the angle of refraction and \( V_1 \) and \( V_2 \) the seismic velocities of the upper and lower layer respectively. In the case of critical refraction, where \( \theta_i = \theta_r = 90^\circ \) the refracted wave travels parallel to the interface with a velocity \( V_2 \). The critical refraction is given by the ratio of the velocities \( V_1 \) and \( V_2 \):

\[
\sin \theta_{c,1,2} = V_1 / V_2.
\]

Since \( \sin \theta_{c,1,2} \leq 1 \) it follows that \( V_1 \) has to be smaller than \( V_2 \). Therefore the prerequisite in refraction seismics is an increasing seismic velocity with depth. The critical refracted wave that travels along the interface produces oscillation stresses, which in turn generate upward moving waves, known as head waves. Since the propagation velocity \( V_2 \) is faster within the lower layer, at a certain distance from the shotpoint (crossover

Fig. 4. Longitudinal and cross sections of the Schliersberg landslide in the Bavarian Alps (Germany). The longitudinal profile highlights the rotational structure of the slide, while the cross profile shows the difference between the active and inactive areas of the slope.
distance) these head waves reach the surface (geophones) faster than the direct wave providing the first arrivals at the geophones. In seismic refraction studies, it is the first arrivals of the P waves that are utilized.

The seismic velocity of P waves is dependent on the elastic modulus and the density, \( \rho \), of the material, through which the seismic waves propagate.

Based on the raypath geometry described above and the structure of the layered underground, it follows that the travel times for the seismic waves can be used to obtain certain time–distance plots. In the case of a layered underground with planar interfaces, the first arrival times lie on a number of clearly defined straight-line segments. The number of the segments corresponds to the number of the underground layers. The slope of the straight line segments is proportional to the reciprocal velocity of the layers. Travel time anomalies are caused by irregularities of the interfaces and varying velocities within the different layers.

The arrival of a seismic wave is detected by geophones, which are placed firmly in the ground using a spacing of between 1 and 5 m. For example, 24 geophones (with a 24 channel seismograph) combined with a spacing of 4 m results in a spread of 92 m. If the expected depths of the interfaces are shallower than 30 m the geophone spacing may be reduced (\( \text{i.e. between 1 and 3 m} \)) to gain a higher resolution record of the underground. To detect a layer within the time–distance plot the geophone spacing must be smaller than the difference between two following crossover distances. At the crossover distance the direct wave is overtaken by a refracted wave. Beyond this offset distance the first arrival is always a refracted wave. In refraction surveying, recording ranges are chosen to be sufficiently large enough to ensure that the crossover distance is well exceeded in order to detect a sufficient number of first arrivals of refracted waves. In cases of thin subsurface layers, where the distance between crossover points is small, a narrow geophone spacing (between 1 and 3 m) is necessary. This in turn may result in a lower penetration depth due to a reduced total spread.

The most common seismic source in geomorphological studies is a 5 kg sledge hammer which generates seismic waves by hitting a metal plate placed on the ground. To improve the signal to noise ratio the sledge hammer “shot” is usually repeated several times (typically between 5 and 10 times) at the same source spot. The resulting seismograms of each shot are stacked (summed) manually or automatically to obtain a single seismogram per shot location.

Generally, a sledge hammer is sufficient to measure the first arrivals along offset distances of approx. 50 m and seldom more than 90 m. This corresponds to penetration depths of 10 to 30 m and is suitable for many landforms. More powerful sources (\( \text{e.g. drop weights, explosives} \)) must be used for longer surveys with larger penetration depths. To apply seismic surveys in geomorphic studies it is highly recommended to use instruments with several single light units rather than to use an all-in-one (heavy) seismograph.

4.2. Advantages and disadvantages

The accurate first onset detection of seismic waves is very often a difficult task. In coarse grained surface conditions (\( \text{i.e. on talus slopes and debris cones} \)) and hummocky topography (\( \text{i.e. on rock glaciers or rockfall deposits} \)) difficulties may arise from the poor coupling of geophones to the surface. On talus slopes with larger block sizes directly on the talus surface, it is advisable to improve the coupling of the geophones by removing the upper layer of larger boulders and pushing the spike of the geophone into fine-grained material underneath. Coupling to larger boulders or to compact bedrock may be established by drilling into the rock and plunging the spike within the drill hole.

In the vicinity of torrents and under strong wind or rainfall, the detection of seismic waves may be impossible due to strong noise in the seismic record. Special attention should be drawn to the obtained velocities of materials which are common in particular landforms (\( \text{e.g. talus deposit, till, rock glacier} \)). The large range of observed velocity values, spanning from \( \sim 400 \text{ m/s} \) (loose debris) up to \( 6500 \text{ m/s} \) (compact rocks) is generally conducive to the application of refraction seismics, as large velocity contrasts between the underlying materials are necessary. However, ranges of P wave velocities for rocks and sediments can overlap significantly. For instance, seismic velocities of dolomite and limestone can vary between 2000 and 6500 m/s depending on the grade of fracturing and weathering, while, for example, glacial sediments compacted by overlying ice can possibly range from 1500 to 3500 m/s. Consequently, there is no unambiguous relationship between certain subsurface units and the linked P wave velocities. As a result, it is not always possible to identify subsurface materials simply based on seismic velocities. To differentiate glacial till (without ice) from frozen ground or solid rock, cross checks with other geophysical methods and/or evidence from borehole logging are needed. Generally, landforms consisting of similar sediments but deposited by different geomorphic processes cannot be distinguished (Hoffmann and Schrott, 2003). Another very common disadvantage in seismic refraction is the “hidden layer” problem, which may lead to incorrect interpretations (Pullan and Hunter, 1990; Hecht, 2001). The hidden layer is caused by a
Fig. 5. (A) Seismograms of forward and reverse shots and intercept models of the refraction sounding on a talus cone (velocities of refractors derived from the slopes); (B) Intercept time model (straight line) and wavefront inversion model of the above mentioned sounding; (C) Refractors of the network ray tracing analysis and tomography using the same data set. The colours indicate the velocities (red < 300 m s\(^{-1}\), blue 300–800 m s\(^{-1}\), green/yellow 800–1200 m s\(^{-1}\), orange > 1200 m s\(^{-1}\)).
sandwiched layer with lower velocities between higher velocity units of refracted first P wave arrivals. Although a hidden layer produces head waves it does not give rise to clear first arrivals. Pullan and Hunter (1990) report an underestimation of bedrock of up to 25% due to a hidden layer.

4.3. Example of application

4.3.1. Case study: coalescing talus cone

Seismic refraction was carried out on a moderately steep (21°) talus cone in an alpine valley in the Bavarian Alps. For the survey, a 24-channel Bison Galileo seismic system and a 5 kg sledge hammer, as an impact source, were used. The applied geophone spacing was 3 m resulting in a profile length of 69 m. The intercept time method, the wavefront inversion method in combination with network ray tracing and the seismic tomography method were applied in order to interpret the seismic data and to show the differences among themselves (Hoffmann and Schrott, 2003). The intercept time method was only applied to forward and reverse shots, resulting in the apparent velocities and intercept times of the different layers.

Wavefront inversion (WFI) modelling was performed using the following steps: (i) reading of travel times of first arrivals of the direct and refracted P waves, (ii) combining travel times of each record to a single travel-time curve, assignment of the travel times to the corresponding layers in the subsurface model, and (iii) inversion of the travel times using the wavefront inversion algorithm.

For network ray tracing a start model (typically an existing WFI model) was used to calculate synthetic travel times and tomography analysis was applied to give automatic adaptation of synthetic travel time data to real data. Tomography analysis allows the modelling of a heterogeneous subsurface structure with a minimum of knowledge about it. Therefore, we used tomography models as background information for the modification of the WFI models in order to obtain a better correspondence of the synthetic and measured travel times. A detailed description of the sophisticated analysis of the subsurface structure including wavefront inversion method, network ray tracing and refraction tomography is described in Hoffmann and Schrott (2003).

It is likely that the investigated landform consists of a mixture of different types of sediment sources (rockfall deposits, moraine). From the seismic data alone, it is impossible to differentiate the internal structure because debris covered moraine and consolidated rockfall deposits probably have similar velocities. Thus, the main objective of this case study is the assessment of depths to bedrock and of the variation of values obtained from different interpretation tools. One of the most interesting findings of this case study is the varying depth to bedrock (or perhaps highly consolidated till) with regard to the interpretation tools (see Fig. 5). The mean depths of the refractors based on the most sophisticated network ray tracing model are 1.6 m (first refractor) and 10 m (second refractor), whereas the less reliable intercept method gives mean depths of 6 and 27 m for the first and second refractor, respectively. In this case, the sole application of the intercept method would have caused a serious error in the data interpretation.

Based on the P-wave velocities of the model layers using the above mentioned methods and on geomorphic evidence of some exposures (loose debris on top and rockfall and debris-flow deposits underneath) on the flanks of the coalescing talus cone we interpret the model shown in Fig. 5 as follows:

First layer: very loose limestone debris accumulated on top of the debris cone.
Second layer: most likely rockfall and debris flow material with a higher degree of compaction and a higher content of fines (compared to first layer).
Third layer: probably bedrock (limestone). Remnants of morainic deposits, however, are also possible. To distinguish between fractured bedrock and till an additional geophysical method (GPR or DC resistivity) should be applied. The velocity of P waves as a physical property is in this case not sufficient to enable interpretation of the third layer. The alternative assumption of till would result in a greater depth to bedrock.

5. Combining and adjusting geophysical methods to geomorphological problems

The various geophysical field methods rely on the evaluation of different physical properties and it is, therefore, essential that the most appropriate technique is applied to a given geomorphological problem (Milsom, 1996). Seismic refraction is sensitive to density contrasts and, thus, suitable for differentiating loose sediments and bedrock. With GPR it is possible to acquire information on internal sediment structures and thus, it can be successfully used to delimitate sediment units of different origin. 2-D resistivity is highly appropriate for investigations in permafrost environments and can deliver high-resolution subsurface data in loamy and/or
wooded terrain whereas GPR frequently fails to deliver any data at all.

Thus, each of the geophysical methods presented has a great potential for aiding clarification of geomorphological problems (Table 2). Nonetheless, every method has its drawbacks and limitations. The main reason for incomplete or false conclusions lies in a lack of contrast between the physical properties of the subsurface layers, particularly between compacted or water-saturated sediments and bedrock. With regard to the common problem of differentiating loose debris from the bedrock base, it can be stated that (1) there might be no distinct dielectrical contrast and thus, no radar reflection at the bedrock surface, (2) the electrical resistivity of debris can vary by orders of magnitude, which means that contrasts within the sediment body may exceed and mask the contrast to the bedrock, and (3) the seismic velocities of compacted sediment such as basal moraine may overlap the velocity range of bedrock, thus producing potential errors in the detection of the sediment base. To overcome these

<table>
<thead>
<tr>
<th>Physical property</th>
<th>Ground-penetrating radar</th>
<th>2-D DC resistivity</th>
<th>2-D seismic refraction</th>
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<tr>
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<td>Resistivity</td>
<td>Elastic moduli, density</td>
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<td>Rock/soil moisture distribution</td>
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Note: the indication of suitable application may be incorrect under specific or extreme local conditions (e.g. wet, dry and blocky).

5.1. Composite application on talus slopes

In the small Tegelberg cirque in the northern European Alps, a relict talus slope is interlocked with lake sediments at the cirque floor. The aim of the investigations was to assess the depth to bedrock and to detect further internal structures within the talus body in order to provide an estimate of the rate of Holocene backweathering. Three geophysical techniques were applied. Fig. 6 shows the combined interpretation of the results (Sass, 2006b).

- The 25 and 50 MHz GPR measurements were performed using a RAMAC GPR (MALÅ Geosystem). The profiles show a distinct reflector between the debris and the silty and clayey deposits of the cirque floor. A weaker reflector, that was traced upwards, was initially thought to mark the bedrock surface (white dashed line in Fig. 6A). The upper part of the debris body showed two different reflection patterns: an upper layer characterized by surface-parallel stratification and an unbedded layer below (separated by dotted line in Fig. 6A). The interface between these units was marked in the lower part of the talus by a distinct reflector.

- The 2-D resistivity measurements (GeoTom device, 88 electrodes, spacing 2 m) also displayed a distinct contrast between talus and cirque floor (Fig. 6B). Near the middle of the profile, a weak resistivity contrast at a depth of approximately 15 m was assumed (white dashed line) and this matches the GPR reflector mentioned above. However, the 2-D resistivity method was at the limit of its penetration depth.

- Seismic refraction measurements revealed that the P wave velocities near the supposed talus base were still much too low (800 to 2000 m/s) to indicate dolostone bedrock. The transition to higher velocities (3500 m/s) was found almost 10 m deeper than supposed.

Thus, it was concluded that a thick layer of basal moraine forms the base of the talus. According to field evidence and to the stacked radar facies, the Holocene debris is mainly due to episodic mudflow activity; it was marked off from the GPR reflection patterns at an average depth of 7 m. Within the accessible depth range of between 7 and 10 m, a number of percussion drillings successfully verified the geophysical results (Fig. 6C).
A combined application of seismic refraction and DC resistivity sounding using 24-channel Bison equipment and an Abem Terrameter respectively, was carried out on a basaltic block field in a German upland near Bonn. A geophone spacing of 3 m was applied resulting in a profile length of 69 m. For the 1-D resistivity sounding we applied the symmetrical Schlumberger configuration with a total profile length of 100 m (AB/2 = 50 m). From the measurements, field curves were established by plotting the apparent resistivity against the distance between the current electrode and the profile centre (AB/2) on a log–log scale (see Fig. 8). In this study the profiles were interpreted using the software RESIXPlus (®Interpex Lim). For both soundings a 3-layer model was assumed based on the synthetic curve which describes the resistivity variation with depth. The algorithm then calculates a curve giving the best fit of the synthetic curve to the field data (see dots in Fig. 8) using an iterative process (inversion technique). No constraints were manually applied to the inversions.

The 2-D interpretation of the seismic refraction data was performed with the generalized reciprocal method (GRM), which is a direct inversion technique which uses travel-time data from both forward and reverse shots (Palmer, 1981). This provides the geometry of sub-surface refractors (Fig. 7). In order to compare the two data sets an identical profile location was used. Geophones and electrodes were placed parallel to contour lines in the centre of the steep (30°) block field. The determination of the overall depth of the block field was of major interest. The seismic refraction reflects two refractors at a depth of approximately 2 and 14 m on average (Fig. 7). Along the profile the second refractor varies between 10 and 16 m. According to the obtained P wave velocities of >2800 ms\(^{-1}\) it was assumed that this was the depth-to-bedrock. This coincides well with the DC resistivity data showing, at similar depths, a significant increase in the apparent resistivity with typical values of between 10\(^3\)
and $10^4 \ \Omega \text{m}$ for basalt (Fig. 8). With regard to the upper layers and internal structure of the block field complementary results were obtained from both methods. Whereas the refraction data show a first refractor at a depth of approximately 2 m, that is evident from an increase in velocity of the P waves of more than $500 \text{ ms}^{-1}$, the resistivity shows only a slight increase at the same depth in the data from the survey in Summer. In Winter, however, the resistivity is significantly higher at the uppermost 2 m which may reflect an existing seasonal frozen layer and also dryer conditions (Fig. 8). An interesting feature is the drastic decrease in resistivity in both soundings just above the bedrock. This is probably caused by larger interstitials between the basalt blocks,
interflows and accumulated downwashed clay minerals which may lead to very low resistivities. This interpretation, however, remains speculative and requires further field evidence. In particular a two-dimensional sounding would help to improve the interpretation. Both methods delivered similar results for estimations of the depths to bedrock.

6. Conclusions

Geophysical techniques have rendered it possible to obtain many new and stimulating solutions for geomorphological problems during the last decade. Geophysics, as used by geomorphologists, should become a common tool within geomorphology, but should not marginalise the geomorphological question. Thus, there is a need for studies coupling the geophysical background with geomorphic approaches in different natural environments and for different landforms.

A number of studies have provided new insights in many landform and process related topics, for example, active layer, internal structure of rock glaciers and talus slopes, depth and extension of landslide bodies and sediment thickness. As previously pointed out, each technique is sensitive to contrasts in certain physical properties which in turn reflect different sediment characteristics (see Table 2). Ground-penetrating radar has been successfully applied in permafrost studies, depth to bedrock, and for soil moisture measurement, but is particularly suitable for assessment of internal sedimentary structures (especially in sediment units with size fractions larger than silt). In these environments, radar wave reflectivity is determined by contrasts in water content that are directly influenced by grain size composition.

2-D resistivity measurements are particularly promising in areas where there are strong contrasts in electrical resistivity such as permafrost and landslides. It is probably best suited for isolated bodies such as permafrost (Kneisel et al., 2000) or ice lenses and landslide structures. Sledge hammer seismic refraction is particularly recommended for shallow applications (<30 m) in environments with mostly horizontal bedding and strong differences in P wave velocities such as dry talus slopes with loose debris above bedrock.

Further investigations using integrated approaches are required. Only an appropriate combination of differing methods allows for much more sophisticated data interpretation. Attention should be paid, not only to the thickness, but also to the internal structure of loose deposits — a task that can only be can be achieved in sufficient accuracy with the combination of several methods. Furthermore, a multi-method approach allows for cross-checking of results and determination of the suitable methods for achieving valuable and reliable results in a particular environment.

The consideration of some general recommendations (“best practice guide”) may help to avoid incorrect and unsuccessful geophysical applications and interpretations in geomorphic research and to improve significantly the accuracy of gathered data:

1. Before using field geophysics, attempt to obtain as much information as possible about expected subsurface conditions such as lithology, soil thickness, groundwater, and depth-to-bedrock as this may significantly influence the measurement strategy. For GPR, DC resistivity and seismic refraction techniques the information may influence the choice of antenna frequency, electrode array and seismic refraction respectively.

2. For reliable interpretation of survey results, available knowledge of, for example, depth of a particular sediment, hollows, ice lenses and depth of groundwater at geomorphological and geological sites is essential. This a priori information may significantly improve post-processing in terms of creating adequate starting models or defining boundary conditions such as expected depth to bedrock and type of bedrock.

3. Always ask the question: Is the geophysical model geomorphologically and/or geologically plausible?

4. Do not strictly adhere to the measurement strategy. Whenever possible, vary the applied frequency or electrode configuration. The results are often surprising.

5. Never trust in a single method. Combine — whenever possible — two or three methods. Cross-checking is essential to increase the trustworthiness of the data interpretation.

6. A geophysical method never provides a unique solution to a particular geomorphic situation, but may help to improve the interpretation.

7. Check and validate results with, where possible, investigations of wells, exposures, borehole—logs or drillings.

8. If there are not any exposures at the study site (which may be, in fact, the very reason for your application of geophysics), try to obtain test profiles under more or less known subsurface conditions somewhere in the vicinity of the site. Knowing the physical properties or reflection patterns of typical sediments in your study area will significantly facilitate your interpretation.
9. Find a geophysicist who is interested in geomorphological approaches and discuss critical data and modelling strategies during the post-processing.

Acknowledgments

The authors are grateful to numerous students from the Universities of Augsburg and Bonn for their support in the field. Special thanks to Thomas Hoffmann for critical discussions on an earlier draft of this paper. We gratefully acknowledge many valuable comments by two anonymous reviewers and a thorough revision by one anonymous referee that substantially improved the manuscript. Jonathan Mole’s English corrections are greatly appreciated. We are also grateful to all the members of the research programme “Sediment Cascades in Alpine Geosystems” (SEDA) for co-operative field work. Financial support from the Deutsche Forschungsgemeinschaft is greatly appreciated (grant Schr 648/1-3).

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