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Part I

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Electrical methods

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1.1 Introduction

Commonly used electrical methods in applied geophysics comprise direct-current (DC) electrical measurements, self-potential measurements (SP) and induced-polarisation methods (IP), including spectral-induced polarisation (SIP).

The SP method is based on passive measurements of natural electrical potential differences in the ground, which are often negligible in periglacial areas as electrically conductive materials or water flow have to be present to generate distinct SP patterns. A cryospheric example measuring subglacial drainage conditions with SP is given by Kulessa *et al.* (2003).

The IP and SIP methods are based on actively induced polarisation effects in the subsurface, which require polarisable material to be present. Again, these effects are usually small in frozen environments, which is why all three methods have seldom been used in periglacial research to date. A review concerning SP and IP methods is included in the review paper concerning the application of geophysical methods in permafrost areas by Scott *et al.* (1990). Further details on these techniques are given in Weller and Börner (1996) and Slater and Lesmes (2002).

In contrast, DC resistivity methods utilise distinct changes in the electrical resistivity within the subsurface, and constitute one of the traditional geophysical methods that have been applied in permafrost research. Since a marked increase of the electrical resistivity occurs at the freezing point, these methods are expected to be most suitable to detect, localise and characterise structures containing frozen material. Based on the number of scientific publications in the past decade and the large variety of applications, the tomographic variant of the method (electrical resistivity tomography, ERT) is maybe the most universally applicable method for research in periglacial permafrost-related mountain

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environments (in combination with another geophysical method, if possible, e.g. Hauck and Vonder Mühll 2003a, Kneisel and Hauck 2003). Due to the recent development of multi-electrode resistivity systems and commercially available two-dimensional inversion schemes for data processing, this method is comparatively easy to apply even in very heterogeneous mountain and arctic terrain. As for most geophysical techniques, the obtained resistivity model is not unambiguous and depends strongly on data quality, measurement geometry and the choice of inversion parameters.

In this chapter the measurement principles of DC resistivity soundings (also called vertical electrical soundings, VES) and ERT are introduced, including data acquisition and processing as well as a discussion of various pitfalls in resistivity inversion and interpretation concerning the detection and characterisation of subsurface materials in periglacial environments.

1.2 Measurement principles

Resistivity surveys are conducted by injecting a direct electrical current (I) into the ground via two current electrodes (A and B in Figure 1.1). The resulting voltage difference (ΔV) is measured at two potential electrodes (M and N). The overall purpose of resistivity measurements is to determine the subsurface resistivity distribution. From the current (I) and voltage difference values (ΔV) the resistivity ρ is calculated using

$$\rho_a = K \frac{\Delta V}{I}, \tag{1.1}$$

where K is a geometric factor that depends on the arrangement of the four electrodes. This calculated resistivity value is not the 'true' resistivity of the subsurface, but a so-called 'apparent resistivity' ρ_a, which equals the 'true' (or specific) resistivity only for a homogeneous subsurface. For heterogeneous

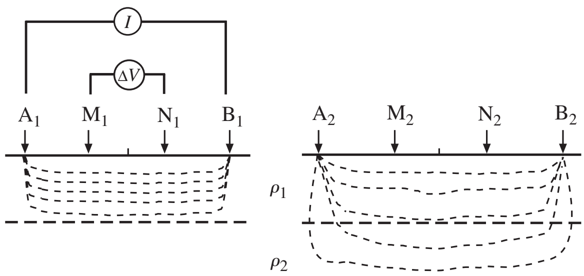

Figure 1.1. Conventional four-electrode configuration in resistivity surveys.

Table 1.1. *Range of resistivities for different materials*

Material	Range of resistivity (Ω m)
Clay	1–100
Sand	$100–5 \times 10^3$
Gravel	$100–4 \times 10^2$
Granite	$5 \times 10^3–10^6$
Gneiss	$100–10^3$
Schist	$100–10^4$
Groundwater	10–300
Frozen sediments, ground ice, mountain permafrost[a]	$1 \times 10^3–10^6$
Glacier ice (temperate)	$10^6–10^8$
Air	infinity

[a] Kneisel (1999)
Compiled mainly after Telford *et al.* (1990) and Reynolds (1997)

resistivity distributions in the ground the resistivity can be derived from the measured apparent resistivity values using inversion methods implemented, for instance, in commercially available software programmes (see below).

The basic principle for the successful application of geoelectrical methods in geomorphology/quaternary geology is based on the varying electrical conductivity (= 1/resistivity) of minerals, solid bedrock, sediments, air and water, and consequently their varying electrical resistivity (Table 1.1). Resistivity surveys give an image of the subsurface resistivity distribution. Knowing the resistivities of different material types, it is possible to convert the resistivity image into an image of the subsurface consisting of different materials. However, as a consequence of overlapping resistivity values of different materials, this conversion might be non-unique. The resistivity of rock, for example, depends on water saturation, chemical properties of pore water, structure of pore volume and temperature. The large range of resistivity values for most materials is thereby due to varying water content. Resistivity values for frozen ground can vary over a wide range (from between 1 and 5 kΩ m to several hundred kΩ m or even a few MΩ m: e.g. Hoekstra and McNeill 1973, Haeberli and Vonder Mühll 1996, Kneisel 1999, Ishikawa *et al.* 2001, Hauck and Vonder Mühll 2003a, Marescot *et al.* 2003, Kneisel 2006, Kneisel *et al.* 2007). Apart from the host material (lithology and textural characteristics of the frozen material), the resistivity depends on the ice content, the temperature, and the content of impurities. The dependence of resistivity on temperature is closely related to the unfrozen water content; as in most earth materials, electrical conduction takes place through ionic transport in the liquid phase.

When the distance between the current electrodes (A, B) is increased, a larger penetration depth is obtained (as indicated in Figure 1.1) yielding more information about the deeper sections of the subsurface. The penetration depth d depends on the measurement geometry and is limited by the maximum electrode spacing. It may be estimated using a formula by Barker (1989) with $d = 0.17L$ for the so-called Wenner array (see below), with L being the distance between the outer (current) electrodes.

1.2.1 Measurement configuration and array types

Vertical electrical soundings (VES)

For electrical resistivity surveys different array types are used. In traditional one-dimensional DC resistivity soundings (or vertical electrical soundings, VES) the symmetrical Schlumberger array is applied (see Chapters 5, 7, 8 and 9). In Schlumberger surveys the distance between the outer current electrodes is increased logarithmically to obtain a sounding curve with maximum penetration depth, while the distance between the potential electrodes remains mostly constant. For the interpretation of one-dimensional data the assumption is made that the subsurface consists of horizontal layers and that the resistivity changes only with depth but not horizontally. The obtained resistivity values are interpreted as a one-dimensional layered model of the subsurface using standard software packages.

Wenner profiling

As another classical survey technique, the resistivity profiling method is used for obtaining lateral changes in the subsurface resistivity. In this case, the Wenner array is applied where the spacing between the electrodes remains fixed, but all four electrodes are moved simultaneously for each reading. Wenner profiling is used to obtain information about lateral changes, but not about vertical changes in resistivity.

Electrical resistivity tomography (ERT)

From the description above, the limitations of Schlumberger sounding and Wenner profiling surveys are evident. The assumed horizontal layering as well as the assumption that resistivity changes only with depth but not horizontally will not always be valid in practice. On heterogeneous ground conditions, the interpretation of one-dimensional soundings can be difficult, as lateral variations along the survey line can influence the results significantly. The sounding curve produces an average resistivity model of the survey area. For some studies this might not be problematic and the results obtained from one-dimensional soundings are sufficient. However, individual anomalies will not show explicitly

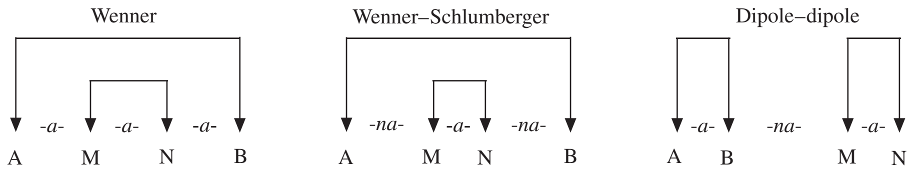

Figure 1.2. Schematics of the most commonly used array geometries in ERT surveys.

in the results. Two-dimensional resistivity tomography (ERT) overcomes this problem using multi-electrode systems and two-dimensional data inversion yielding a more accurate model of the subsurface (see Chapters 6, 8, 9 and 10). ERT requires multiple resistivity measurements with various electrode spacings along a profile line (2D) or on a two-dimensional grid (3D). The most commonly used measurement geometries in ERT surveys are the Wenner, Wenner–Schlumberger and Dipole–dipole arrays (see Figure 1.2).

Array types

In Wenner surveys, the two outermost electrodes (A and B) are used as current electrodes while the potential difference is measured at two electrodes in between (M, N). Potential-electrode spacing increases as current-electrode spacing increases, with equal distances between all electrodes for each measurement. The Wenner configuration has a moderate investigation depth and good resolution for horizontal structures that change with depth. Since the total number of measurements required is smaller than for other configurations, the time to complete a survey is comparatively short; however, less information for the subsurface is obtained than from other arrays. The Wenner–Schlumberger array is a combination of the Wenner array and the Schlumberger array with constant potential-electrode spacing but increasing current-electrode spacings leading to a better depth resolution compared to the Wenner configuration. The number of measurements is larger than for a Wenner survey but smaller than for a Dipole–dipole array. The Wenner–Schlumberger configuration is useful for horizontal and vertical geomorphological structures and can be the best choice as a compromise between the Wenner and Dipole–dipole arrays. The Dipole–dipole array comprises two dipoles formed by the current electrodes on one side and the potential electrodes on the other side. The current and potential spacings are the same and the spacing between them is an integer multiple (n) of the distance (a) between the current and the potential electrodes (Figure 1.2). This array type has a better horizontal resolution, but shallower investigation depth than the Wenner array. Furthermore, its signal-to-noise ratio is smallest and the required number of readings to complete a survey is largest of all three presented configurations.

A comprehensive evaluation of the characteristics of the specific arrays is given in a resistivity tomography tutorial by Loke (2004), together with useful information for the conduction of resistivity surveys and data inversion.

1.3 Data acquisition

Compared to vertical electrical soundings, which typically involve 10 to 20 readings, ERT imaging surveys consist of 100 to several hundred measurements. Depending on the quality of the recorded voltages, the measurements for each electrode configuration have to be repeated until the variance is less than a predefined threshold. Acquiring a full ERT data set with about 40 electrodes requires between 0.5 and 1.5 hours (depending on the configuration). The time for data acquisition depends on the number of measured electrode combinations (often called quadripoles) and on the number of repetitions of a single combination.

For the acquisition of the apparent resistivity data sets, multi-electrode systems are commonly used. These systems automatically measure the apparent resistivities for a series of electrode combinations for a given array geometry. Using 40 equally spaced electrodes with a spacing of 5 m and a Wenner array results in a data set of 190 apparent resistivities, a survey line of 195 m length and a penetration depth of about 30 m.

Choice of an appropriate electrode configuration is dependent on the difficult surface conditions associated with mountain regions. Since the maximum current injected into the ground can be quite low, the geometrical factors of the electrode configurations may be critical (Telford *et al.* 1990). For this reason, often Wenner or Wenner–Schlumberger configurations are employed, even though Dipole–dipole configurations may provide superior lateral resolution (Loke 2004). This characteristic is described in more detail in Section 1.5.3.

Further details on different array geometries are given, for instance, in Telford *et al.* (1990) and Reynolds (1997). Applications of different arrays to various geomorphological studies are described in Kneisel (2003, 2006).

1.4 Data processing

Data processing consists basically of applying an appropriate inversion algorithm to the observed apparent resistivity data set to determine the specific resistivity values on a two-dimensional x–z model grid. During the past few years, two- and three-dimensional inversion algorithms for resistivity data have been developed and applied successfully in many environmental and archaeological applications (e.g. Johansson and Dahlin 1996, Mauriello *et al.* 1998, Ogilvy *et al.* 1999, Olayinka and Yaramanci 1999, Daily *et al.* 2004, Günther *et al.* 2006).

The large number of ERT applications in recent years was partly due to the comparatively new availability of 2D and 3D resistivity inversion software like RES2DINV/RES3DINV, which performs a smoothness-constrained inversion using finite difference forward modelling and quasi-Newton inversion techniques (Loke and Barker 1995, 1996). The inversion results in a 2D or 3D specific resistivity model section as opposed to the so-called pseudosections obtained by analysing the apparent resistivities alone. In addition, topography may be incorporated in the inversion, which is an important factor in mountainous glacial and periglacial terrain. Even though there is more than one resistivity inversion software package commercially available, we will explain the exemplary inversion procedure using RES2DINV.

Prior to data processing with the inversion software, it is recommendable to check the data set for abnormally high or low resistivity values. If these values can be attributed to measurement errors and/or bad electrode contact (see Section 1.5.1), they should be excluded manually. The observed apparent resistivity data sets can then be inverted using either the least-squares or the robust inversion scheme (i.e. use of ℓ_2- or ℓ_1-norm for data and model space, respectively; Loke *et al.* 2003). Robust inversion is usually chosen over smooth inversion wherever sharp layer boundaries are expected, as they are reproduced better than with the more smearing least-squares norm.

By default, a homogeneous earth model is used as the starting model, which is obtained by calculating the average of the logarithm of the measured apparent resistivity values. From this resistivity model is calculated a set of apparent resistivities that would be observed in the field if the resistivity model represented the real resistivity distribution in the subsurface. In an iterative algorithm the optimisation method then tries to reduce the difference between the calculated and measured apparent resistivity values by adjusting the resistivity of the model blocks in the resistivity model. A measure of this difference is given by the root-mean-square (RMS) error. By using different starting models the reliability of the inversion results can be tested (Marescot *et al.* 2003).

The least-squares equation is given as

$$\boldsymbol{p} = (\boldsymbol{J}^{\mathrm{T}}\boldsymbol{J} + \lambda \boldsymbol{C}^{\mathrm{T}}\boldsymbol{C})^{-1}\boldsymbol{J}^{\mathrm{T}}\boldsymbol{g} \tag{1.2}$$

where $\boldsymbol{p}$ is the model perturbation vector, $\boldsymbol{J}$ is the matrix that includes the sensitivities of the data points with respect to a particular model parameter, $\boldsymbol{g}$ is the discrepancy vector, which contains the differences between measured and calculated apparent resistivities, and $^{\mathrm{T}}$ denotes the transpose of a matrix (Loke and Barker 1995). The matrix $\boldsymbol{C}$ acts as a flatness filter to minimise the underdetermined components of the inversion problem and force the inverted models to be smooth. The parameter λ specifies the weighting between data constraints and a-priori information (i.e. the assumed smoothness of the subsurface). Equation (1.2) is solved iteratively (by repeatedly updating $\boldsymbol{g}$ and $\boldsymbol{J}$) until the RMS of the discrepancies $\boldsymbol{g}$ does not alter significantly after an inversion step and/or it becomes smaller than the measurement accuracy.

In RES2DINV the user can specify the parameter λ, which is called the damping parameter. The higher the damping parameter the smoother the resulting resistivity model, but the weaker the model is constrained by the data set and the larger the RMS error. The lower the damping parameter the noisier the model, but the stronger the data constrain, corresponding to a small RMS error. However, the best model from a geomorphological or geological perspective might not be the one with the lowest possible RMS (see Section 1.5.3). Thus, it is essential to consider the local geomorphological setting in performing the interpretation. This enables unrealistic images of the subsurface structure to be excluded. In order to analyse the results in terms of, for instance, permafrost distribution and characterisation, the final resistivity model has to be interpreted and its reliability assessed. In the following, examples are shown with typical problems associated with resistivity inversion and interpretation.

1.5 Periglacial applications and particularities

1.5.1 Data acquisition

Application of geoelectrical surveys in periglacial environments often implies one major problem, which is the coupling between the electrodes and the sometimes heterogeneous and rocky ground surface. This problem can often be resolved by adding water in the immediate vicinity of the electrodes, by attaching sponges soaked in salt water and/or installing extra electrodes in parallel to the electrodes (see also Chapter 6). Experience has shown that a sufficient supply of water is more important than extra addition of salt. Electrodes should be long (0.4–1 m) and should be firmly positioned between blocks with maximum contact to the ground. Where larger rocks or rock faces are present, small holes can be drilled into the rock using metallic pins or screws as electrodes (Sass 2003, Krautblatter and Hauck 2007). Electrically conductive fluid can be used to further enhance the electrode contact.

The obtained contact resistances depend strongly on the surface conditions and can be as high as several hundred kΩ m. Anomalous bad electrode contacts, which may significantly influence the inversion results, are characterised by alternating high and low values in the apparent resistivity pseudosection (so-called W-shaped anomalies). If a faulty electrode is the cause for the