Geophysical methods for detecting permafrost in high mountains

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presented by
Christian Hauck
Dipl. Met., Universität München
born 28 May 1970
in München, Germany

accepted on the recommendation of
Prof. Dr. A. Ohmura, examiner
Dr. D. Vonder Mühll, co-examiner
Dr. H. Maurer, co-examiner
Prof. Dr. M. Davies, co-examiner

2001
4.4.5 Conclusions ........................................................................... 74

4.5 Time-lapse DC resistivity tomography to monitor freezing and thaw¬
ing processes in permafrost areas ................................................. 76

4.5.1 Introduction ......................................................................... 76

4.5.2 Installation of a fixed electrode array and data acquisition ...... 77

4.5.3 Synthetic modelling of time-lapse DC resistivity tomography 
data ......................................................................................... 78

4.5.4 Quasi-monthly resistivity measurements at Schilthorn ........ 80

4.5.5 Analysis of the data in terms of temperature and unfrozen 
water content ............................................................................. 86

4.5.6 Conclusion ........................................................................... 89

5 Electromagnetic induction methods .............................................. 91

5.1 Introduction ........................................................................... 91

5.2 Theory and instruments ........................................................ 92

5.2.1 Frequency-domain systems (EM-31, GEM-300) ................. 92

5.2.2 Time-domain systems (PROTEM) ....................................... 97

5.3 Results .................................................................................. 99

5.3.1 Conductivity mapping using a priori information (EM-31) .... 99

5.3.2 Investigation of the spatial variability at the PACE drill sites 
(EM-31) ..................................................................................... 103

5.3.3 Active layer mapping using the 'optimal' frequency (GEM-300) 103

5.3.4 Electromagnetic soundings to determine the depth of per¬
frost base (PROTEM) .............................................................. 104

5.4 Conclusions ........................................................................... 109

6 Refraction seismics ................................................................. 111

6.1 Introduction ........................................................................... 111

6.2 Theory and tomographic inversion ........................................ 112

6.2.1 Elastic waves .................................................................... 112

6.2.2 Critical refraction and head wave ...................................... 112

6.2.3 2-D refraction tomography ................................................ 113

6.3 Synthetic modelling to optimise data acquisition and data inversion . 115

6.3.1 Determining the minimal number of sources and receivers ... 115
6.3.2 Choice of the initial velocity model ..................... 116
6.3.3 Regularisation parameter ................................. 119
6.3.4 Convergence ............................................. 120
6.4 Field example — Val Bever ............................... 122
   6.4.1 Data acquisition .................................. 124
   6.4.2 Results ............................................ 124
6.5 Conclusions .............................................. 127

7 Thermal methods ............................................ 129
   7.1 Bottom temperature of snow cover (BTS) ............... 129
   7.2 Passive microwave radiometry .......................... 130
      7.2.1 Introduction .................................. 130
      7.2.2 Theory ....................................... 130
   7.3 Setup of the field measurement and data acquisition system .......... 132
   7.4 Field example — Schilthorn ............................ 133
   7.5 Conclusions ......................................... 135

8 Discussion .................................................. 137
   8.1 General mountain permafrost prospecting ............... 139
   8.2 Mapping the lateral variation of mountain permafrost .... 142
   8.3 Determining the vertical extent of mountain permafrost (sounding) ... 145
   8.4 Active layer studies .................................. 148
   8.5 Characterising the physical properties of mountain permafrost .......... 150
      8.5.1 Val Bever: ice anomalies ....................... 150
      8.5.2 Schilthorn: bedrock anomalies .................. 151
      8.5.3 Juvvashøe: air cavern anomalies ................ 153
   8.6 Permafrost monitoring ................................ 153

9 Conclusion and outlook .................................... 157

Bibliography ................................................. 160

Acknowledgements ........................................... 170
A Field results

A.1 Janssonhaugen ........................................ 172
A.2 Hiorthfjellet ........................................ 173
A.3 Tarfala ................................................. 174
A.4 Juvvasshøe ............................................ 176
A.5 Schilthorn ............................................. 180
A.6 Zermatt ............................................... 183
A.7 Val Bever ............................................. 184
A.8 Murtèl ................................................ 186
A.9 Stelvio ................................................ 188
A.10 Foscagno ............................................. 189
A.11 El Veleta ............................................. 189
Seite Leer / Blank leaf
Abstract

Five different geophysical techniques were tested for application for permafrost studies in high mountain regions within the PACE project (Permafrost and Climate in Europe). The aim was to determine suitable methods to detect, characterise, map and monitor permafrost in the context of possible slope instabilities resulting from climate induced thawing of permafrost bodies. Measurements were conducted on various field sites including pure bedrock and unconsolidated permafrost sites, such as rock glaciers and moraines. The applied geophysical techniques include DC resistivity tomography, refraction seismic tomography, frequency-domain and time-domain electromagnetic induction methods and passive microwave radiometry.

2-dimensional DC resistivity tomography turned out to be a suitable method for a number of permafrost related questions, such as detecting permafrost, mapping its horizontal extent, estimating the ice/unfrozen water content, determining the permafrost base for shallow permafrost occurrences and monitoring seasonal variations in the active layer. 2-dimensional refraction seismic tomography is equally well suited for these targets, except that interpretation of the results is more difficult, since the velocity contrasts of frozen and unfrozen ground is smaller than those of electrical resistivity. To remove interpretational ambiguities both methods should be combined.

Laboratory studies using a miniature DC resistivity tomography system were used to visualise freezing and thawing processes and to estimate the total water content at the field site. Permafrost monitoring on a monthly basis was conducted using DC resistivity tomography and a fixed electrode array. Resistivity changes were related to borehole temperature changes and the evolution of the unfrozen water content could be determined at different depths. As the unfrozen water content is a key parameter concerning slope instabilities induced by thawing permafrost, this approach is considered an important contribution to future monitoring studies.

Measurements with electromagnetic induction methods include conductivity mapping with frequency-domain electromagnetic instruments (EM31, GEM300) and transient electromagnetic soundings (PROTEM). Due to the small conductivity values encountered and the heterogeneity of most permafrost field sites in high mountain environments, conductivity mapping using the EM-31 and GEM-300 has to be combined with other geophysical techniques, for example DC resistivity. Once, a permafrost occurrence is detected by a DC resistivity survey, its extent can easily be mapped with a conductivity meter, which is light-weight and therefore suitable to map large areas in short time by a single person. Electromagnetic soundings using the time-domain instrument PROTEM can be used to determine the depth of the...
permafrost base. The maximal investigation depth depends on the transmitter coil dimensions and the upper layer resistivity.

Passive microwave radiometry is used in winter to determine the bottom temperature of the snow cover (BTS) on a larger "footprint" than with the commonly used BTS-probes. As for conventional BTS surveys, an at least 0.8–1.0 m thick snow cover has to be present for some time in order to relate the temperature signal to permafrost occurrences. The microwave radiometer (11 GHz) can be mounted on a sledge or may be used from aboard a helicopter.

Even though the applicability of each method was tested separately in this study, it is recognised that in most cases an additional method is needed to unambiguously interpret the results in terms of permafrost prospection. A typical example in DC resistivity surveys is the differentiation between isolated rock, ice and air occurrences, each resulting in anomalously high resistivity values. In combining DC resistivity with refraction seismic surveys this ambiguity can be resolved as the seismic velocities of the three materials are markedly different.
Zusammenfassung


Elektromagnetische Induktionsverfahren messen die elektrische Leitfähigkeit im Boden und werden unterschieden in 'frequency-domain'-Methoden (EM-31 und GEM-300), die mit konstanter Frequenz laterale Veränderungen der Leitfähigkeit aufspüren und 'time-domain'-Methoden (PROTEM), die durch zeitaabhängige Messun-

Passive Mikrowellen-Radiometrie ist ein neues Verfahren zur Bestimmung der Bodentemperatur der Schneedecke (BTS). Hierbei wird über einen grösseren räumlichen Ausschnitt gemittelt, als dies mit herkömmlichen Sonden möglich ist. Wie auch bei konventionellen BTS-Sondierungen, wird eine zeitlich konstante mindestens 0.8-1 m dicke Schneedecke benötigt, um die gemessene BTS-Temperatur zur Permafrostverbreitung in Beziehung setzen zu können. Das Radiometer kann auf einem Schlitten oder an Bord eines Helikopters installiert werden.

Trotz der guten Ergebnisse einzelner Methoden für die Permafrosterkundung, ist in den meisten Fällen die kombinierte Anwendung mehrerer Verfahren notwendig, um verlässliche Resultate zu erhalten. Ein typisches Problem in der Gleichstromgeoelektrik ist die Unterscheidung zwischen isolierten Fels-, Eis- und Lufteinschlüssen im Boden, welche alle zu extrem hohen Widerstandswerten führen können. Durch die Kombination von Gleichstromgeoelektrik und Refraktionsseismik können diese Anomalien korrekt zugeordnet werden, da die sich seismischen Geschwindigkeiten von Fels, Eis und Luft deutlich unterscheiden.
Chapter 1

Introduction

Permafrost is a temperature phenomena defined as lithospheric material with a temperature below 0°C continuously for more than one year. Approximately 20% of the Earth's land surface is underlain by permafrost (Brown, 1997). Of these, a substantial part may be characterised as mountain permafrost, which is located in high mountain environments with special morphological features like rock glaciers or ice-cored moraines (e.g. Washburn, 1979, Guodong and Dramis, 1992). Mountain permafrost usually refers to mountain regions that are predominantly underlain by permafrost, while it is absent in adjacent lowlands, e.g. the European Alps. This is in contrast to polar permafrost, which blankets the landscape at all elevations, e.g. the North-American or Siberian Arctic (Harris and Corte, 1992).

The localisation and characterisation of permafrost in high mountain areas has been recognised as a major goal in construction work due to dangers for buildings and foundations from frost and thaw action (e.g. Haeberli, 1992). Recently, the interest in mountain permafrost has increased substantially, due to possible dangers for civilisation from an increased frequency of slope instabilities resulting from climate induced thawing of permafrost bodies. The occurrence of ice — especially in the case of permafrost — fundamentally influences the geotechnical properties of the ground. The presence of ice in frozen soils and rock joints can contribute to maintaining the stability of mountain slopes and rock walls. Upon thawing, the loss of strength due to melting of the ice matrix might combine with an increasing supply of water, leading to stability problems within the slope. Rock falls may be caused by a weakening of ice-bonded joints due to a rise in rock temperatures (Davies et al., 2001). There is evidence that the largest amplitude of temperature increase due to climate change by greenhouse forcing will occur in the belt of cold mountains. This prediction is based on general circulation and local area models (e.g. Ohmura and Beniston, 1996, Rotach et al., 1997). Mountain slopes in the belt of discontinuous permafrost with ground temperatures close to the freezing point are therefore particularly vulnerable to climate changes.
1.1 The EU-project PACE

In December 1997 the EU-project PACE (Permafrost and Climate in Europe) started aiming at understanding the processes which lead to enhanced permafrost thawing in the context of climate change (Harris et al., 2001a). The core of the project consists of a series of at least 100 m deep boreholes in permafrost regions along a north-south transect from Spitsbergen/Svalbard to the Sierra Nevada/Spain.

The PACE project is organised in 6 work packages:

1. **Thermal monitoring**, including drilling, instrumentation and data processing of the PACE boreholes. Furthermore, inverse temperature modelling to determine paleoclimatic information from the borehole temperatures were conducted (e.g. Isaksen et al., 2001a).

2. **Geophysical surveys** (including this study), responsible for mapping the permafrost distribution around the boreholes and the evaluation of geophysical key parameters for detecting and characterising mountain permafrost (e.g. Vonder Mühll et al., 2001, Hauck et al., 2001). Furthermore, geophysical techniques to monitor climatically-induced changes in permafrost distribution and character have been evaluated.

3. **Geomorphological mapping**, focussing on periglacial phenomena at different scales in the vicinity of the borehole locations, including the compilation of a mapping handbook.

4. **Numerical modelling**, including GIS-based modelling of the permafrost distribution around the PACE field sites (e.g. Gruber and Hoelzle, 2001, Hoelzle et al., 2001). High-resolution energy balance measurements at most of the PACE sites were used to verify the models (Mittaz et al., 2000).

5. **Geotechnical centrifuge modelling**, for simulating thaw consolidation and thaw-induced mountain slope instability. Centrifuge-scaling laws allow both rapid and repeatable modelling of thaw-related slope processes, providing the capability of measuring the effects of constitutive soil properties, ice contents, thermal regime and surface slope geometry, and thereby ultimately developing predictive models (e.g. Harris et al., 2000, Davies et al., 2000, 2001).

6. **Geotechnical hazard prediction**, integrating the results of all work packages in order to provide guidelines for geotechnical hazard assessment of degrading mountain permafrost in the context of global warming (Harris et al., 2001b).

1.2 Geophysics in the study of permafrost

Most permafrost related questions call for effective methods to detect, characterise, map and monitor mountain permafrost on various scales. On larger scales (> several square-kilometres) statistically and physically based numerical models are used.
to determine the permafrost distribution and to predict changes due to changing climate (e.g. Keller, 1992, Hoelzle, 1996, Beniston et al., 1997, Mittaz et al., 2000, Hoelzle et al., 2001). However, verification of the models is only possible through detailed field measurements on a smaller scale. Apart from geomorphological mapping and direct temperature measurements in boreholes, geophysical methods are the most powerful techniques for detecting and mapping permafrost. They are often the only means applicable, since geomorphological mapping is limited to special permafrost features, such as rock glaciers, and the drilling of boreholes is costly and often logistically impossible.

Geophysical techniques have been used to study permafrost and characterise areas of permanently frozen ground for many years, but they were mostly applied in polar permafrost regions (e.g. Hoekstra and McNeill, 1973, Timofeev et al., 1994). Seismic, electromagnetic and electric methods were applied successfully especially for exploration and engineering purposes (for a review see Scott et al., 1990). Due to the flat and generally uniform surface characteristics in the plain North-American and Siberian Arctic, the permafrost distribution is quite homogeneous and does not change much on horizontal scales less than a few kilometres. In addition, the surface consists mainly of peat and other materials with a high unfrozen water content. These surface characteristics facilitate the application of most geophysical techniques, since sufficient coupling between sensors and the ground is guaranteed. It is especially important for electric methods, which use direct contact to insert electric current into the ground. In contrast to that, permafrost occurrences in the European mountains are highly variable and depend strongly on altitude, incoming radiation, local climate and geology. The heterogeneous permafrost distribution calls for methods which are able to map the shallow subsurface at scales between a few metres and 1 km. Furthermore, at higher altitudes the surface is often free of vegetation and consists of loose debris with poorly developed soils. Consequently, the unfrozen water content is low, rendering contact of geophysical sensors to the ground more difficult. Surface characteristics are highly variable, thus often excluding the application of commonly used plane-layer approximations for the processing of geophysical data.

Nevertheless, on mountain permafrost the combination of 1-dimensional (1D) DC resistivity soundings and refraction seismics has been standardly applied for many years (Barsch, 1973; Fisch et al., 1977; King et al., 1987; Evin and Fabre, 1990, King et al., 1992; Vonder Mühll, 1993; Vonder Mühll and Schmid, 1993; Wagner, 1996). These methods have been particularly effective in determining the depth of the permafrost table and the approximate thickness of the permafrost. In special cases (rock glaciers and ice-cored moraines) the origin of the permafrost ice (sedimentary or congelation/metamorphic, Haeberli and Vonder Mühll, 1996) and/or the ice content could be estimated using DC resistivity and gravity (Vonder Mühll and Klingele, 1994). In addition, ground penetrating radar (GPR) was used to map the internal structure of the uppermost 20 to 30 m of rock glaciers (Vonder Mühll et al., 2000, Isaksen et al., 2000b, Berthling et al., 2000). However, most surveys were conducted on special morphologic permafrost features, such as rock glaciers, and only a few geophysical studies have been reported from bedrock permafrost sites.
1.3 Objectives and structure of the thesis

The aim of this study was to investigate the feasibility of various geophysical methods and state-of-the-art processing schemes for permafrost studies in high mountain environments. Several types of permafrost occurrences will be examined. The study was conducted within the work package 2, geophysical surveys, of the PACE project. In the course of the project a set of geophysical surveys was conducted at each PACE site. The objectives of the present study are as follows:

- Evaluation of a variety of geophysical techniques (including DC resistivity, refraction seismics, electromagnetic conductivity mapping, electromagnetic soundings and passive microwave radiometry) for addressing specific permafrost-related questions in high mountain areas, such as:
  - General prospecting of mountain permafrost occurrences
  - Mapping the lateral distribution of mountain permafrost
  - Determining the vertical extent of mountain permafrost occurrences
  - Active layer studies
  - Determining the spatial representativeness of the borehole results in the vicinity of the PACE drill sites.
  - Characterising the physical properties of mountain permafrost occurrences in terms of unfrozen water/ice content
  - Monitoring the temporal evolution of mountain permafrost

- Compiling a catalogue of advantages and disadvantages of each method for the application in mountain permafrost studies

In the case of DC resistivity and refraction seismics, state-of-the-art 2-dimensional data inversion techniques were applied. Furthermore, laboratory DC resistivity studies were conducted to determine fundamental relationships between electrical resistivity, temperature and unfrozen water content of partially frozen material.

In order to keep the focus on the evaluation of the applicability of the various geophysical methods, only a selection of the conducted surveys are presented in detail. All other survey results are included in the appendix. The outline of the thesis is as follows: In chapter 2 a short field site description is given together with a complete list of surveys conducted at each site. After a brief introduction of geophysical survey and interpretation techniques in chapter 3, the individual geophysical methods are introduced in chapters 4 to 7 and analysed by showing examples from selected field surveys. In chapter 4 the DC resistivity tomography method is evaluated including laboratory experiments and repeated measurements at a permafrost monitoring site. In chapter 5 three electromagnetic induction instruments, the EM-31, the GEM-300 and the PROTEM, are applied to different permafrost-related questions. The refraction seismic tomography method is evaluated in chapter 6 and the passive microwave radiometry is introduced in chapter 7. In chapter 8 the applicability of the individual methods for addressing the above permafrost-specific questions is discussed in order to yield comprehensive guidelines for using geophysical methods to investigate permafrost. Finally, a summary and outlook is given in chapter 9.
Chapter 2
Field sites

The geophysical surveys were conducted at different mountain permafrost test sites, including every PACE drill site (see Fig. 2.1). Mountain permafrost may display many different morphologies with specific physical properties. Rock glaciers usually have a high ice content compared to bedrock sites with comparatively low ice content. Permafrost occurrences in Svalbard or mainland Scandinavia can be quite different from occurrences in the Alps or in the Sierra Nevada, Spain. Consequently, a geophysical technique may be ideal for a survey on a certain field site, whereas it may not be applicable to another. In this study geophysical surveys were conducted on a large variety of different permafrost morphologies, including rock glaciers, moraines, debris covered slopes, bedrock and even forest slopes. A short description of each field site is given in this chapter and a comprehensive list of all field sites and the methods applied is shown in Table 2.1.

2.1 Janssonhaugen/Svalbard

Location and climate

The northernmost PACE drill site is situated at Janssonhaugen (78°11’N, 16°28’E at 270 m asl), on a low sandstone hill in the middle of the valley bottom of Adventdalen, some 15 km east of Longyearbyen/Svalbard (Fig. 2.2). The mean annual air temperature (MAAT) at Janssonhaugen is −7.8°C, calculated from data of the meteorological station at Svalbard airport during the period 1976–1998. The average precipitation is around 300–500 mm a⁻¹ and snow thickness is very low due to strong redistribution of snow by wind (Isaksen et al., 2000a). In Svalbard, permafrost covers the entire land area with depths varying between 200 and 450 m (Liestøl, 1980).

Geological setting

The bedrock in the area is Lower Cretaceous, dominated by fine-grained sandstone with a high content of silt and interbeds of shale (Dypvik et al., 1991). Results
Figure 2.1: Map of the field site locations.
<table>
<thead>
<tr>
<th>Site Description</th>
<th>Type</th>
<th>Depth [m]</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Altitude [m asl]</th>
<th>DC resistivity</th>
<th>EM mapping</th>
<th>PROTEM sounding</th>
<th>Refraction seismic</th>
<th>Radiometry</th>
</tr>
</thead>
<tbody>
<tr>
<td>Janssonhaugen (P)</td>
<td>bedrock</td>
<td>102</td>
<td>78°11'N</td>
<td>16°28'E</td>
<td>270</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td></td>
<td>X</td>
</tr>
<tr>
<td>Hiorthfjellet</td>
<td>rock glacier</td>
<td>–</td>
<td>78°15'N</td>
<td>15°47'E</td>
<td>440</td>
<td>X</td>
<td>X</td>
<td></td>
<td></td>
<td>X</td>
</tr>
<tr>
<td>Tarfala (P)</td>
<td>bedrock</td>
<td>100</td>
<td>67°55'N</td>
<td>18°38'E</td>
<td>1550</td>
<td>X</td>
<td>X</td>
<td></td>
<td></td>
<td>X</td>
</tr>
<tr>
<td>Juvvasshøe (P)</td>
<td>bedrock</td>
<td>129</td>
<td>61°41'N</td>
<td>8°22'E</td>
<td>1894</td>
<td>X</td>
<td>X</td>
<td></td>
<td></td>
<td>X</td>
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<td>Schilthorn (P)</td>
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<td>101</td>
<td>46°33'N</td>
<td>7°50'E</td>
<td>2900</td>
<td>X</td>
<td>X</td>
<td>X</td>
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<td>Zermatt, Stockhorn (P)</td>
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<td>100</td>
<td>45°59'N</td>
<td>7°49'E</td>
<td>3410</td>
<td>X</td>
<td></td>
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<td></td>
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<tr>
<td>Zermatt, Gandegg</td>
<td>moraine</td>
<td>–</td>
<td>45°58'N</td>
<td>7°43.5'E</td>
<td>3030</td>
<td>X</td>
<td></td>
<td></td>
<td></td>
<td>X</td>
</tr>
<tr>
<td>Val Bever</td>
<td>forested valley slope</td>
<td>–</td>
<td>46°33.5'N</td>
<td>9°51.5'E</td>
<td>1800-1880</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td></td>
</tr>
<tr>
<td>Murtèl</td>
<td>rock glacier</td>
<td>58</td>
<td>46°26'N</td>
<td>9°49.5'E</td>
<td>2670</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td></td>
<td>X</td>
</tr>
<tr>
<td>Stelvio (P)</td>
<td>rock glacier</td>
<td>–</td>
<td></td>
<td></td>
<td>2700-2760</td>
<td>X</td>
<td>X</td>
<td></td>
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<tr>
<td>Stelvio (P)</td>
<td>bedrock</td>
<td>100</td>
<td>46°31'N</td>
<td>10°29'E</td>
<td>3000</td>
<td>X</td>
<td>X</td>
<td>X</td>
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<tr>
<td>Stelvio (P)</td>
<td>ski run</td>
<td>–</td>
<td></td>
<td></td>
<td>2730-2830</td>
<td>X</td>
<td>X</td>
<td></td>
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<td></td>
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<tr>
<td>Foscagno</td>
<td>rock glacier</td>
<td>24</td>
<td>46°29'N</td>
<td>10°12.5'E</td>
<td>2520-2630</td>
<td>X</td>
<td>X</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>El Veleta (P)</td>
<td>bedrock</td>
<td>114.5</td>
<td>37°03'N</td>
<td>3°22'E</td>
<td>3380</td>
<td>X</td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>Corral del Veleta</td>
<td>rock glacier</td>
<td>–</td>
<td>37°03'N</td>
<td>3°22'E</td>
<td>3100</td>
<td>X</td>
<td></td>
<td></td>
<td></td>
<td>X</td>
</tr>
</tbody>
</table>
of the mineral quantification of cuttings from the PACE borehole showed fairly homogeneous bedrock with a high and stable quartz concentration, interpreted as sandstone (density 2280 kg m$^{-3}$, porosity 25%) (Isaksen et al., 2000a). Only two samples at 40 m and 92 m depth contained high percentages of clay minerals. The homogeneity of the bedrock was further confirmed through visual inspection of the borehole using a video camera, which showed little caliper variations. The vegetation cover at the drill site is very sparse and the debris cover above the bedrock is thin and dominated by a patchy pattern of morainic material and in-situ weathered bedrock.

**Borehole results**

The 102 m deep PACE borehole was drilled in May 1998. Figure 2.3a shows the temperature profile from Janssonhaugen recorded on 11. May 1999. The smooth profile suggests an absence of any geothermal disturbance factors like variable snow thicknesses, 3-dimensional temperature effects due to topography and geologic heterogeneity. By using a linear extrapolation and a mean gradient derived from the borehole measurements, the permafrost thickness is estimated to be around 220 m (Isaksen et al., 2000a). From Fig. 2.3b it is seen that the depth of the Zero Annual Amplitude (ZAA) is at 18 m. Maximum depth of the active layer in summer 1998 was 1.5 m.

**Geophysical techniques applied**

Around Janssonhaugen the following geophysical surveys were conducted in April 1999:

- DC resistivity tomography to determine changes in the permafrost distribution around the borehole.
- Electromagnetic (EM-31) mapping to determine changes in the active layer.
- Electromagnetic (PROTEM) soundings at the drill site to determine the depth of the permafrost base.

The study at Janssonhaugen was conducted in co-operation with Ketil Isaksen (Department of Physical Geography, University Oslo, Norway) and Daniel Vonder Mühll (VAW-ETH Zürich, Switzerland).

**2.2 Hiorthfjellet/Svalbard**

**Location and climate**

The active rock glacier Hiorthfjellet (78°15'N, 15°47'E at 440 m asl) is situated 5 km north-east of Longyearbyen and some 15 km north-west of the PACE drill site at
Figure 2.2: Map (published by Norsk Polarinstittutt) and picture of the Janssonhaugen drill site. The location of the borehole, its cemented lid can be seen in the picture, is marked by a cross. All geophysical surveys in this area were conducted close to the borehole. Picture taken by C. Mittaz/C. Naguel.

Janssonhaugen (Fig. 2.4). It is a typical tongue-shaped, talus rock glacier, which is confined by a large bowl in the Hiorthfjellet mountainside. The MAAT at the rock glacier is estimated to be around —9°C and average annual precipitation is assumed to be 300–500 mm (Isaksen et al., 2000b). As for the PACE drill site at Janssonhaugen, the permafrost thickness is assumed to be between 200 and 450 m.

Geological setting

The investigation area is composed of tectonically undisturbed bedrock from the Tertiary, alternating between sandstone, siltstone and shale (Major and Nagy, 1972). The surface material of the rock glacier consists of weathered material of this rock type, comprising massive, light-coloured sandstone and fine grained grey-green sandstone. Large blocks are seldom seen and the main part of the material is composed of rocks with a length and width between 0.2 and 0.5 m (Isaksen et al., 2000b).
Figure 2.3: Temperature profile of the borehole at Janssonhaugen, Svalbard. (a) Temperature profile from 11.5.1999 and (b) seasonal variation of temperature in the uppermost 20 m (Isaksen et al., 2000a).

Previous studies

The rock glacier was investigated extensively by Isaksen et al. (2000b), measuring surface velocity vectors and determining the inner structures using DC resistivity and ground penetrating radar. The results of the surface velocity measurements show an average horizontal velocity of approximately 0.09 m a\(^{-1}\). DC resistivity soundings showed an active layer thickness between 0.7 to 1.5 m with average resistivities of 15-25 kΩm. Underneath, an ice-rich layer with resistivity values of 100–900 kΩm with maximal thickness of 25–35 m was found. Resistivity values for the bedrock underneath the rock glacier were determined to be 6 kΩm.

Geophysical techniques applied

The following geophysical surveys were conducted during a fieldwork campaign in April 1999:

- Four electromagnetic (EM-31) survey lines along and across the rock glacier to delineate the boundaries of the ice body within the rock glacier.
- Transient electromagnetic soundings (PROTEM) at the rock glacier tongue to determine the depth of the permafrost base.
The study at Hiorthfjellet rock glacier was conducted in co-operation with Ketil Isaksen (Department of Physical Geography, University Oslo, Norway).

2.3 Tarfala/Sweden

Location and climate

The PACE drill site (67°55′N, 18°38′ E at 1550 m asl) near Tarfala Research Station is located on Tarfalaryggen, a north-south oriented mountain ridge east of the Kebnekaise massif. The borehole is situated at the saddle point between Tarfalapakte (1806 m asl) and Tarfalatjåarro (1622 m asl) (see Fig. 2.5). The estimated MAAT for the drill site and Tarfala Station is between −6 to −7°C and −3.9°C, respectively, and annual precipitation is approximately 500 mm (Isaksen et al., 2001a). During a normal winter the snow cover thickness is 0.2–0.3 m. Permafrost is assumed to be continuous with maximal thickness of more than 100 m (e.g. King, 1976, 1984, Kneisel, 1999). The lower limit of permafrost on northern and southern slopes in
the Tarfala region is estimated to be around 920 m asl and 1420 m asl, respectively (King, 1984).

**Geological setting**

Bedrock in the area consists of metamorphic rocks from micaschists to gneisses and is mostly exposed. At the drill site about 4 m of weathered material overlay the firm bedrock. Cracks filled with unconsolidated material extend to a depth of 10–15 m (Isaksen et al., 2001a). Vegetation is scarce and is dominated by lichen and mosses.
Borehole results

The 100 m deep borehole was drilled in March 2000, showing a temperature of \(-2.8^\circ\text{C}\) in 100 m depth. Linear extrapolation of the borehole temperatures suggests a permafrost thickness of 350 m (Isaksen et al., 2001a). No data on seasonal temperature variation are available yet.

Geophysical techniques applied

The following geophysical surveys were conducted in the area around the Tarfala drill site in spring 1999 and summer 2000:

April 1999

- A 3 km long electromagnetic (EM-31) and BTS profile was recorded reaching from Tarfala station (1130 m asl) to the drill site (1550 m asl) and further to a small ridge in Tjeurolako (1430 m asl) (see Fig. 2.5). The primary goal was to determine the suitability of using EM-31 as proxy for BTS data in regions with homogeneous surface conditions.

August 2000

- DC resistivity tomography profiles near the drilling site and at Tarfala station (conducted by D. Vonder Mühll) to determine changes in the permafrost distribution around the borehole.

The study near Tarfala station was conducted in co-operation with Ketil Isaksen (Department of Physical Geography, University Oslo, Norway), Peter Jansson (University of Stockholm) and Daniel Vonder Mühll (VAW-ETH Zürich, Switzerland).

2.4 Juvvasshøe/Norway

Location and climate

The PACE drill site at Juvvasshøe (61°41'N, 8°22'E at 1894 m asl) is located in Jotunheimen, southern Norway. It is situated on top of the northernmost extent of the Galdhøppigen massif, Norway's highest mountain top (see Fig. 2.6). The MAAT is estimated as \(-4.5^\circ\text{C}\) (Aune, 1992). Mean precipitation is about 800 mm a\(^{-1}\) (Østrem et al., 1988), and due to the exposed location and strong winds, snow thickness is very small throughout the winter (Isaksen et al., 2001a). Earlier investigations (e.g. King, 1982, Ødegård et al., 1996) suggested a permafrost thickness of the order of less than 200 m, which was corrected to be at least 300 m during the PACE investigations (see borehole results). Extensive BTS measurements showed that the permafrost distribution in this region depends mainly on altitude. It is likely to reach altitudes between 1490 m and 1410 m along the northern slopes (Isaksen et al., 2001b).
Figure 2.6: Map (published by Statens Kartverk 7-91) and picture of the Juvvasshøe drill site. The picture shows the surface characteristics along the permafrost transition profile (here the DC resistivity line, see text). The viewing angle of the photo is indicated on the map. The location of the borehole (site A and marked by the cross), the long survey line on the northern slope, the permafrost transition area (site B) and the non-permafrost site (site C) are marked on the map. Picture taken by D. Vonder Mühll.

**Geological setting**

The greater part of the bedrock around the drill site consists of light to medium-grey gneiss composed predominantly of plagioclase feldspar and quartz (Isaksen et al., 2001a). From the results of the borehole logs, the weathered layer extends to 3–4 m where firm bedrock was located. From the logs, as well as from a visual inspection through a borehole camera, a high frequency of fissures and fractures was determined. These fractures are ice-filled as seen from high-resistive peaks in the resistivity log (around 8 and 18 m). The surface at the drill site consists of small to medium size debris originating from weathered bedrock with no or little vegetation cover.

**Previous studies**

Former DC resistivity soundings at Juvvasshøe showed contradicting results. King (1982) found a permafrost thickness of about 50 m at the Juvvasshøe plateau and resistivities for frozen bedrock of 100–150 kΩm as opposed to 20–25 kΩm for unfrozen bedrock. Ødegård et al. (1996) estimated the permafrost thickness to be 130–170 m with resistivities of 40–90 kΩm of the frozen and 10–25 kΩm of the unfrozen bedrock. Resistivities of unfrozen debris were estimated as 5–10 kΩm (Ødegård et al., 1996). Refraction seismic surveys at Juvvasshøe showed an upper layer P-wave velocity of
less than 800 ms$^{-1}$ with a thickness of 1.3 m, a 5 m thick middle layer with a P-wave velocity of 3500 ms$^{-1}$ and a P-wave velocity of the bedrock of 6000 ms$^{-1}$ (King, 1984).

**Borehole results**

The 129 m deep borehole was drilled in August 1999. A set of geophysical borehole logs and X-ray diffraction analysis (XRD) of the borehole cuttings was performed. Figure 2.7 shows the temperature profile from Juvvasshøe recorded at 29th February 2000. Temperature is $-2.2^\circ$C at 129 m depth. The small geothermal gradient indicates thick mountain permafrost. By using a simple linear extrapolation and a mean gradient derived from the borehole, the permafrost thickness on Juvvasshøe is estimated as 380 m (Isaksen et al., 2001a). Seasonal temperature variation showed the depth of the Zero Annual Amplitude (ZAA) around 16 m and the maximal depth of the active layer as 2.15 m (Isaksen et al., 2001a).

**Geophysical techniques applied**

Around Juvvasshøe the following geophysical surveys were conducted in August 1999:

- Characterisation of the permafrost occurrence around the PACE borehole using two 160 m long orthogonal DC resistivity tomography profiles, one 60 m
refraction seismic tomography profile and electromagnetic (EM-31) mapping (110 m x 110 m, grid size 5 m x 5 m).

- Determination of the altitudinal limit of permafrost along the northern slope using a 4 km long electromagnetic (EM-31) profile reaching from the drill site (1890 m asl) down to non-permafrost areas (1300 m asl), a 640 m long DC resistivity tomography profile and four 60 m refraction seismic tomography profiles.

- Determination of base-line resistivity and P-wave velocity values for non-permafrost areas around Juvvasshøe using a 160 m DC resistivity tomography profile and a 60 m refraction seismic tomography profile.

The study around Juvvasshøe was conducted in co-operation with Ketil Isaksen, Bernd Etzelmüller, Johan Ludvig Sollid (Department of Physical Geography, University Oslo, Norway), Nick Russill (Terradat Ltd, Cardiff, Wales) and Daniel Vonder Mühll (VAW-ETH Zürich, Switzerland).

2.5 Schilthorn/Switzerland

Location and climate

The Schilthorn (2970 m asl) is located in the Bernese Oberland in the Northern Swiss Alps. It is part of a massif situated at the transition between the Prealps in the north and the principle chain of the Bernese Alps. The Schilthorn is an east-west striking crest with north and south facing slopes. The PACE drill site is located on a small plateau on the north facing slope (46°33’N, 7°50’E at 2900 m asl, see Fig. 2.8). The MAAT at the top of Schilthorn is estimated to be -4°C and annual precipitation varies from about 1200 mm in the valley bottom (796 m asl) up to about 2700 mm on the top (Imhof et al., 2000). Due to the high amount of precipitation and additional snow input through wind transport, the snow cover persists usually from October to June. The high amount of precipitation and the exposed setting of the Schilthorn massif, as the northernmost chain of the Bernese Alps, lead to the heaviest glaciation of the Alps, leaving little space for periglacial phenomena (Imhof, 1996). However, above 3300 m asl, permafrost in the relatively rare unglaciated parts can be considered as continuous (Haeberli, 1975).

Presence of permafrost has been found at the summit when the facilities for the cable car were built 30 years ago. During the construction of the buildings, several ice lenses with a thickness up to 1 m were encountered. In 1998 temperature readings in a 12 m deep construction shaft at the summit were close to 0°C, but not negative (Imhof, personal communication). Measurements of the bottom temperature of the snow cover (BTS) and DC resistivity soundings did not provide clear indication concerning the presence and the distribution of permafrost in the Schilthorn area (Imhof et al., 2000).
Geological setting

The Schilthorn consists of dark, micaceous shales with interbedded ferruginous quartzose sandstones (Glockhaus formation) that weathers to form fine-grained debris (Imhof et al., 2000). This is in contrast to the crystalline rocks of the Central Alps, where most of geophysical investigations in Alpine permafrost have been made so far. Therefore, the physical and especially the electrical properties are quite different from other permafrost sites in the Alps (Vonder Mühll et al., 2000). The surface at the drill site consists of small to medium size debris originating from weathered bedrock with no vegetation cover.
Figure 2.9: Seasonal variation of the temperature profile in the 14 m borehole at Schilthorn/Switzerland.

Borehole results

To check whether the Schilthorn site is located in the belt of present permafrost, a 14 m deep borehole was drilled in mid-October 1998. The cuttings pulled to the surface were slightly wet to a depth of 5 m, and generally dry below. No ice layers were encountered during the drilling. Temperature readings showed an active layer thickness of almost 5 m, indicating that below the weathered layer the uppermost few metres of the bedrock thaw seasonally as well (see Fig. 2.9). At greater depths, temperature values decrease and reach $-0.7^\circ$C in 14 m. These permafrost temperatures are considerably higher than at sites with comparable altitude and exposition, which is probably due to the low albedo of the dark surface material. Furthermore, the thick snow cover acts as insulation cover against the negative winter air temperatures (Vonder Mühll et al., 2000).

In August 2000, two 101 m deep boreholes, one vertical and one with an angle of 60° to the vertical, were drilled 14 m to the west of the 14 m-borehole. This array allows two-dimensional temperature modelling of the Schilthorn crest. As the thermistor chains in the boreholes are not yet connected to a data logger, no continuous temperature readings are available at the moment. However, measurements taken shortly after the drilling showed a very small temperature gradient with depth and still marginally negative temperatures at 100 m depth of both drillings (Vonder Mühll, personal communication).
Geophysical techniques applied

The drill site at Schilthorn was used as one of the main field sites in this study. Therefore, a number of geophysical surveys with different goals were conducted between 1998 and 2000:

- Four 200 m long DC resistivity tomography and three 115 m long refraction seismic tomography profiles around the drill site (August 1998) to determine changes in the permafrost distribution around the borehole.
- Electromagnetic (EM-31) mapping of the area around the drill site (October 1998) to determine changes in the active layer.
- Seasonal DC resistivity tomography measurements approximately on a monthly basis to detect resistivity changes due to freezing and thawing using a 60 m fixed electrode array (from September 1999 to September 2000).
- Test study using passive microwave radiometry (January 1999) to determine the bottom temperature of the snow cover (BTS).
- Further DC resistivity and refraction seismic profiles in the Schilthorn region conducted by Karin Giesa (Diploma thesis, University of Zurich, Switzerland) and Jon Russill (Bsc, University of Cardiff, Wales).
- Electromagnetic soundings (PROTEM) at the drill site conducted by Susanne Hanson (University of Copenhagen, Denmark) to determine the depth of the permafrost base.

The study on Schilthorn was conducted in collaboration with Catherine Mittaz, Martin Hoelzle and Wilfried Haeberli (Department of Geography, Universität Zürich, Switzerland), Daniel Vonder Mühll (VAW-ETH Zürich, Switzerland) and Nick Russill (Terradat, Cardiff, Wales).

2.6 Zermatt/Switzerland

Location and climate

The PACE drill site at Stockhorn near Zermatt, Switzerland, is located in the Western Swiss Alps (45°59'N, 7°49'E at 3410 m asl, see Fig. 2.10). The site is situated on a small plateau on a mountain crest with a topography comparable to the drill sites on Schilthorn and El Veleta/Sierra Nevada. The MAAT at Gornergrat (some 300 m below the drill site, see Fig. 2.10) is given as −4.1°C. Estimates for annual precipitation range between 1500 and 2500 mm for the investigation area (King and Kalisch, 1998).

In addition to the Stockhorn site, permafrost related investigations were performed near Gandegghütte and were focused on an ice-cored moraine (see dashed blue lines...
in the detail map of Fig. 2.10). The north-south oriented moraine is the eastern lateral moraine of the Upper Theodul glacier at an altitude of 3050–3000 m asl. It was first investigated in 1974 to choose a location for a pylon of the cable car to the Klein-Matterhorn (3884 m asl). The cores of nearby boreholes revealed indeed the presence of massive ice inside the moraine, so that the foundations of the pylon had to be fastened into the bedrock underneath and to be protected against the creep pressure of the ice (Keusen and Haeberli, 1983).

**Geological setting**

At Stockhorn the surface cover is thin consisting of medium size debris with no vegetation. The surface of the moraine at Gandegg is blocky with some fine material and vegetation. To the east of the moraine a flat bedrock plateau is present consisting of Serpentinite with small cracks visible at the surface.
Previous studies

The results from a series of Schlumberger vertical electrical soundings across the moraine at Gandegg showed a thin (about 2 m) active layer with resistivities between 1.2 and 3 kΩm and frozen sediments with resistivities around 150 kΩm below (Keusen and Haeberli, 1983). The maximum thickness of the frozen moraine material was estimated to be between 20 and 30 m. The bedrock underneath consists of massive serpentinite with a resistivity of about 18–25 kΩm.

Geophysical techniques applied

In July 1998 the following geophysical surveys were conducted:

Stockhorn

- Two 200 m long DC resistivity tomography profiles across the drill site plateau to determine changes in the permafrost distribution around the borehole.

Gandegg

- Five 200 m long DC resistivity tomography profiles near Gandegghütte to detect and characterise the ice occurrence inside the moraine and to determine the permafrost distribution.

The surveys near Zermatt were conducted in collaboration with Thomas Herz, Martin Schlerf and Lorenz King (Department of Physical Geography, Universität Giessen, Germany).

2.7 Val Bever/Switzerland

Location and climate

Val Bever is a tributary of the main valley of the Upper Engadin, Eastern Swiss Alps. It is a trough-shaped, low altitude valley with valley bottom elevations between 1730 m and 1800 m asl (Fig. 2.11). Both the north- and south-exposed valley sides are wooded. The regional climate is rather continental with a fairly low amount of precipitation and a comparatively high temperature range. The MAAT at the station Bever (valley entrance) is around +1°C. Due to steep valley slopes and frequent temperature inversions mean temperatures in winter are very low. January and February temperatures at the meteorological station Bever (1710 m asl) are −10°C and −8°C, respectively, compared to −6.8°C and −5.8°C at the station St. Moritz (1825 m asl), which is located 115 m higher in the main valley of the Upper Engadin (after Schüepp, 1967).

Results from a GIS-based permafrost model (PERMAMAP, Hoelzle, 1996) suggest a probable permafrost occurrence on the north-exposed valley side, with a lower boundary at the foot of the slopes between 1800 m and 2000 m asl. On the south exposed, orographic left valley side, permafrost is improbable up to the mountain crest with altitudes between 2700 m to 2950 m asl (Kneisel et al., 2000).
Geological setting

The steep and rocky valley walls are incised by distinct rock couloirs and torrents, which formed part of the starting zones of several minor debris flows triggered during a thunderstorm in July 1995. Most parts of the scree slopes which occur below the rock walls are well covered with vegetation. However, large boulders are visible in the forest (Kneisel et al., 2000).
Previous studies

In a previous study, results from DC resistivity soundings and BTS measurements support the results from the model simulation suggesting a shallow permafrost occurrence below the timberline (Kneisel et al., 2000). Resistivities of the active layer were found to be around 10 kΩm with a thickness of 1–2 m. Underneath a 10–15 m thick layer showed resistivities between 30–90 kΩm representing the permafrost.

Geophysical techniques applied

The Val Bever was chosen as a test site due to its character as a low altitude permafrost occurrence with sporadic patches of frozen ground, as opposed to larger morphological permafrost features like rock glaciers or the PACE drill sites, where permafrost is distributed rather homogeneously. The field measurements were concentrated along two clearings on the north-exposed valley side where only small larch trees are present (see Fig. 2.11). The following geophysical surveys were conducted in July 1998:

- Three 200 m long DC resistivity tomography profiles to obtain detailed a priori information at one representative permafrost location in the target area (clearing I, see Fig. 2.11).
- A 100 m long refraction seismic tomography profile to confirm the presence of permafrost at one representative permafrost location in the target area (clearing I, see Fig. 2.11).
- Two 200 m long electromagnetic (EM-31) lines along the DC resistivity profile lines in clearing I to determine a permafrost signal for the EM-31.
- Electromagnetic (EM-31) mapping of the whole study area between the two clearings (in January 1999) to determine the permafrost distribution.
- Two 200 m long DC resistivity tomography profiles in clearing II and one 200 m long DC resistivity profile in the wooded part between clearing I and II to verify the EM-31 results.

The study in the Val Bever was conducted in co-operation with Christof Kneisel (Department of Physical Geography, University of Trier, Germany) and Felix Keller (Academia Engiadina, Samedan, Switzerland).

2.8 Murtèl/Switzerland

Location and climate

The rock glacier Murtèl-Corvatsch (46°26'N, 9°49.5'E at 2670 m asl), situated in the Upper Engadin valley, Eastern Swiss Alps, is one of the best investigated creeping permafrost bodies (Fig. 2.12). The active rock glacier has developed within a
Figure 2.12: Map and photo of rock glacier Murtel with the survey lines of the geophysical surveys. The borehole is marked with a red circle and the view angle of the photo is marked in green on the map. Survey line A marks the EM-31 and GEM-300 profile, line B marks the DC resistivity tomography and Cl to C5 are the locations of PROTEM soundings. Map: PK50 © Bundesamt für Landestopographie. Picture taken by F. Keller.

former cirque from perennially frozen, north-westerly exposed scree slopes between 2850 m asl and 2620 m asl. It shows basically two flow types: extending flow in the upper part and compressing flow in the lower part of the rock glacier, which causes pronounced ogive-like transverse ridges (Haeberli and Vonder Mühll, 1996). The MAAT at the drill site is around −1°C and annual precipitation is around 1000 mm (Mittaz, personal communication).
Geological setting

The surface of the rock glacier with its marked micro-topography of furrows and ridges consists of coarse, crystalline boulders without any fine material and very sparse vegetation. The size of the boulders can reach 3 m and elevation differences between furrows and ridges are up to 5–8 m. The steep, approximately 20 m high front is largely free of vegetation and rests on granodiorite bedrock.

Previous studies

After a 58 m core drilling in 1987, a series of investigations followed concerning borehole deformation, drill core analysis, snow cover characteristics, energy balance, borehole and surface temperatures, creep behaviour, hydrology, melt water analysis as well as a number of geophysical surveys (e.g. Haeberli et al., 1988, Vonder Mühll and Haeberli, 1990, Haeberli et al., 1998, Vonder Mühll et al., 1998, Vonder Mühll et al., 2000, Mittaz et al., 2000 and references herein). An energy balance station at the borehole was established in 1997 measuring air temperature, humidity, wind speed and direction at three different heights, as well as all components of the radiation balance (see Mittaz et al., 2000). The geophysical surveys included refraction seismics, DC resistivity soundings, gravimetry and ground penetrating radar (e.g. Vonder Mühll, 1993, Vonder Mühll and Klingele, 1994, Vonder Mühll et al., 2000, Lehmann et al., 2000). Resistivities of the ice body range between 50 kΩm and 2 MΩm and seismic P-wave velocities are between 3400 and 3700 ms\(^{-1}\). Resistivities of the uppermost, unfrozen layer are given as 5–15 kΩm (Vonder Mühll, 1993).

Borehole results

In 1987 a 58 m deep core drilling was performed at the transition from extending to compressing flow. Analysis of the drill core revealed two permafrost layers beneath a 3–4 m thick active layer: an upper layer with an extremely high ice content (90–100% by volume between 3 and 28 m) and a lower one consisting of coarse blocks with ice-filled pores (around 40% by volume), but almost completely without fine rock particles (Haeberli et al., 1998). The main zone of shearing (ca. 4 cm a\(^{-1}\)) is in the transition zone between the two permafrost layers, at a depth of 28 to 30 m, with the upper (supersaturated) layer obviously undergoing steady-state creep and overriding the non-deforming lower layer (Wagner, 1992). Bedrock was found at 52 m depth.

Figure 2.13 shows the temperature profile from 1. September 1997 as well as the seasonal variations in the uppermost 20 m. Active layer thickness is around 3–4 m and the zero annual amplitude is at 20 m depth. Between 52 and 56 m depth, seasonal temperature variations around 0°C occurred every year and hint to an intrapermafrost groundwater flow (talik), which is active from June to September (Vonder Mühll, 1992).
Due to the large number of earlier surveys, the drill cores and the detailed knowledge about temperatures and physical properties, the rock glacier is an ideal test site for applying new geophysical methods and evaluating their applicability to mountain permafrost studies. During this study a number of geophysical surveys were conducted at the rock glacier Murtèl site:

- Radiometric test surveys on and in front of the rock glacier using a passive microwave radiometer to estimate the BTS (in March 1998 and March 1999, lead by Hansueli Gubler/AlpUG).

- Electromagnetic surveys (EM-31 in April 1998 and GEM-300 in June 1999) along the rock glacier to determine changes in the active layer.

- A 160 m long DC resistivity tomography profile from the borehole across the tongue of the rock glacier to the non-permafrost area in front of it (July 1998).

- Five transient electromagnetic soundings (PROTEM) along a 200 m profile from the borehole to the tongue to determine the depth of the permafrost base (June 1999).
Figure 2.14: Map of the whole survey area and picture of the rock glacier near Stelvio pass. A marks the DC resistivity line on the rock glacier, B marks the location of the EM-31/GEM-300 survey line shown in Fig. 5.2, the rectangle C shows the EM-31/GEM-300 area map and D corresponds to the DC resistivity and seismic refraction line on the old ski run. The dashed lines near the ski run mark the two lateral moraines of the old Scorluzzo glacier. Picture taken by C. Hauck.

The study at Murtèl rock glacier was conducted in collaboration with Catherine Mit-taz, Martin Hoelzle and Wilfried Haeberli (Department of Geography, Universität Zürich, Switzerland) and Daniel Vonder Mühll (VAW-ETH Zürich, Switzerland).

2.9 Stelvio pass/Italy

Location and climate

The PACE drill site at Stelvio Pass (46°31'N, 10°29'E at 3000 m asl) is located in Northern Italy between the Ortler massif and the Valtellina valley (Fig. 2.14). The regional climate is continental, with annual precipitation ranging between 1200 and 1600 mm and a MAAT at 3100 m asl of −4°C (Guglielmin et al., 1994). Two additional permafrost sites in the area were investigated during the fieldwork in summer 1998 and summer 1999: a rock glacier on the northern slope of Monte Scorluzzo and an abandoned ski run at the same altitude 300 m to the east (see Fig. 2.14).

The old ski run is situated inside the area that the Scorluzzo Glacier occupied during the Little Ice Age. The frontal moraine reached down to 2705 m asl. By the retreat of the Scorluzzo Glacier since the Little Ice Age, a recent formation of permafrost seems possible at this site (see also Kneisel, 1999). In addition, construction work for the ski run, performed in 1991, and snow preparations may have changed the thermal regime of the ground (Haeberli, 1992). This effect, which is believed to be in favour for permafrost conditions, was already noted near Murtèl rock glacier/Switzerland (Hoelzle, personal communication).
Geological setting

The drill site is characterised by large outcrops of layered limestones and massive and layered dolostones. The surface is covered by a few centimetre thin layer of weathered material. The surface of the ski run is covered by more homogeneously distributed small scale debris due to the construction work. The rock glacier surface consists of medium scale boulders (up to 1 m) and little micro-topography (see Fig. 2.14). The vegetation is scarce at all three sites.

Previous studies

The rock glacier was already investigated in 1997 concerning permafrost and active layer thickness using vertical electrical soundings and vegetation mapping (Guglielmin and Cannone, personal communication). The results showed an active layer of 2–3 m and a permafrost thickness of 15–20 m in the main part. In the frontal part of the rock glacier a thicker active layer and a thinner permafrost body (around 10 m) was found.

Borehole results

The 100 m deep PACE borehole was drilled in May 1998. The drilled bedrock consists of alternating layers of limestone and massive dolomites. The ground ice content is locally high. The ice is trapped in narrow fractures and probably fills also wider fractures and small caves (Guglielmin et al., 2001). Temperatures are negative below 5 m depth throughout the year (see Fig. 2.15).

Geophysical techniques applied

The following geophysical surveys were conducted during two fieldwork campaigns in August 1998 and June 1999:

PACE drill site:

- 200 m long DC resistivity tomography profile to determine changes in the permafrost distribution around the borehole.
- Five transient electromagnetic soundings (PROTEM) along a 200 m profile around the drill site to determine the depth of the permafrost base.

Ski run:

- 200 m long DC resistivity tomography profile to detect and characterise possible permafrost occurrences.
- 200 m long electromagnetic (EM-31 and GEM-300) survey lines along the DC resistivity profile to compare the different methods and determine a permafrost signal for EM-31 and GEM-300 surveys.
Electromagnetic (EM-31 and GEM-300) mapping of the area occupied by the Scorluzzo Glacier during the Little Ice Age to map possible permafrost occurrences.

Rock glacier

- 200 m long DC resistivity tomography profile along the rock glacier to determine the extent of the ice-rich permafrost body.

The study at Stelvio Pass was conducted in co-operation with Mauro Guglielmin (Department of Geology, University of Roma III, Italy).

2.10 Foscagno/Italy

Location and climate

The active rock glacier Foscagno (46°29’N, 10°12.5’E between 2630 and 2520 m asl) is located in Valle di Foscagno in Northern Italy some 20 km west of the PACE drill site at Stelvio pass. Climate conditions are assumed to be similar to the Stelvio area. The rock glacier consists of a number of partially overlapping lobes, as seen in Fig. 2.16. The main active lobe reaches 2520 m at the front.
Figure 2.16: Map and photo of rock glacier Foscagno. The GEM-300 survey line is marked by the solid line. The borehole location is marked by a black cross. The black rectangle marks the GEM-300 mapping survey area. Map: PK50 ©Bundesamt für Landestopographie. Picture taken by C. Hauck.
Geological setting and borehole results

In June 1998 a 24 m deep borehole was drilled at the frontal part of the southern lobe of the rock glacier. Between 2.5 m and 14.5 m a core was collected for future analysis. The borehole stratigraphy shows an active layer thickness of 2.5 m, made of large boulders. The ice core consists of alternating layers of pure ice and frozen debris. From 14.5 to 23.5 m unfrozen sediments or frozen sediments with poor ice content are present. The bedrock, consisting of gneiss, is found in 23.5 m depth (Guglielmin et al., 2001).

Previous studies

The rock glacier Foscagno has been investigated extensively by Guglielmin et al. (1994) and Guglielmin (1997). DC resistivity soundings showed an active layer of 1–3 m with resistivity values around 10 kΩm. Underneath, resistivity values between 20 to 500 kΩm were found with a maximal thickness of around 20 m, indicating ice-rich permafrost (Guglielmin et al., 1994).

Geophysical techniques applied

The following geophysical surveys were conducted during a fieldwork campaign in June 1999:

- Electromagnetic (GEM-300) survey line across the frontal part of the rock glacier to determine differences in the active layer between the two main lobes (see Fig. 2.16).
- Electromagnetic (GEM-300) mapping of the frontal part of the southern lobe of the rock glacier to determine changes in the active layer.
- Transient electromagnetic soundings (PROTEM) near the drill site to determine the depth of the permafrost base.

The study at Foscagno rock glacier was conducted in co-operation with Mauro Guglielmin (Department of Geology, University of Roma III, Italy).

2.11 El Veleta/Spain

Location and climate

The southernmost PACE drill site is situated at El Veleta peak (37°03.4'N, 3°22'E at 3380 m asl) on the western tip of the Sierra Nevada range, Spain. The region is located in the cryomediterranean bioclimatic zone. To the north, under the steep walls of El Veleta, a small oval-shaped, north-west oriented cirque is located (see
In this former glacier basin, called the Corral del Veleta, buried ice occurrences are suspected (Gomez et al., 2001). To the east the peaks of Los Machos (3330 m asl) and El Mulhacen (3524 m asl), the highest peak of the Iberian peninsula, are located. The MAAT around 3100 m is presumed to be 0°C (Gomez et al., 2001).

Figure 2.17: Map of the locations of the geophysical surveys and picture of El Veleta peak. The borehole is marked with a cross. DC resistivity tomography and refraction seismic lines are shown in red with letters A to F. The view angle of the picture is marked in green on the map. Map published by Instituto de Cartografía de Andalucía, Junta de Andalucía. Picture supplied by the Spanish PACE group.

Geological setting

The lithology of the region consists of feldspathic micaschists, much affected by tectonics. The surface at the drill site is strongly exposed to frost wedging, with
gelification stone lobes visible in many parts of the area. The vegetation cover is very sparse.

Borehole results

A 114.5 m deep borehole was drilled in September 2000. The borehole is equipped with a PACE standard 100 m thermistor string. No data are available yet. In addition a small meteorological station was installed nearby, including sensors for temperature, humidity, snow depth and soil moisture.

Geophysical techniques applied

The following geophysical surveys were conducted during two fieldwork campaigns in September 1998 and March 1999 (lead by Rob McDonald, Terradat/Cardiff):

- Four 200 m long DC resistivity tomography profiles and three 115 m refraction seismic profiles in the Corral del Veleta to detect buried ice occurrences (September 1998).
- A 200 m long DC resistivity tomography profile and a 115 m refraction seismic profile at the top of El Mulhacen to detect permafrost (September 1998).
- Two 100 m long DC resistivity tomography profiles at the saddle point near Los Machos peak to detect permafrost (March 1999).

The study in the Sierra Nevada was conducted in co-operation with Rob McDonald (Terradat, Cardiff, Wales), Luis Miguel Tanarro, Enrique Luengo and David Palacios (Department of Physical Geography, Universidad Complutense Madrid, Spain), Miguel Ramos (Department of Physics, Universidad de Alcala de Henares, Spain) and Antonio Gomez (University of Barcelona).
Chapter 3

Geophysical methods

3.1 Principles of geophysical prospecting

Geophysical methods are used to gain information about the physical properties and the structure of the subsurface. In permafrost studies the properties of interest are temperature and ice content. Without a borehole, these properties cannot be observed directly. Therefore, the detection of permafrost from the surface depends on those characteristics that differentiate it from the surrounding media. In geophysical prospecting, methods based on variations in the elastic properties of rocks, variations in electrical conductivity and local changes in gravity, magnetism and radioactivity have been developed providing information about the nature of the structures below the surface. To establish a relationship between these properties and the permafrost conditions, some knowledge about the physical properties of permafrost is required.

3.2 Physical material properties of permafrost

The application of geophysical methods in mountain permafrost regions is related to changes of the physical properties of earth material mainly associated with freezing of incorporated water. The degree of change in the physical properties depends on water content, pore size, pore water chemistry, ground temperature and pressure on the material (Scott et al., 1990). For example, the finer-grained soils permit ice lens growth and have the largest ground-ice volumes. Saline pore water in sediments and interfacial interaction between the mineral matrix and water in clays account for large unfrozen water contents at subzero temperatures (Anderson and Morgenstern, 1973, Tice et al., 1981). When applied on permafrost, most geophysical methods detect parameters correlated to ice content. However, the results may not agree exactly with the permafrost distribution indicated by temperature alone, as negative temperatures do not always correspond to large ice contents. Fortunately, significant changes in properties of the Earth to which geophysical techniques respond usually occur at temperatures only slightly lower than 0°C.
The two commonly used geophysical parameters for differentiating between frozen and unfrozen material are the electrical resistivity and the seismic compressional wave velocity, which is related to the elastic properties of a material.

### 3.2.1 Electrical resistivity

A large variety of electric and electromagnetic techniques are based on changes of subsurface resistivity (see chapters 4 and 5). A marked increase in resistivity at the freezing point was shown in several field studies (e.g. Hoekstra et al., 1975, Seguin, 1978, Rozenberg et al., 1985, Benderitter and Schott, 1999) and laboratory experiments (Olhoeft, 1978, Pandit and King, 1978, King et al., 1988), as shown in Fig. 3.1a. For many soils, the resistivity increases exponentially until most of the pore water is frozen (McGinnis et al., 1973, Daniels et al., 1976, Pearson et al., 1983). The resistivity is reduced for saline pore waters, as the freezing point is depressed and the unfrozen water content at subzero temperatures is increased (Pandit and King, 1978). In section 4.4 results from laboratory experiments are presented, which analyse the relationships between resistivity, temperature and unfrozen water content in detail for typical mountain permafrost material.

There are different ways of electric current propagation in earth material. In most permafrost materials electrolytic conduction takes place, where the current is carried by ions in the pore fluids of the material. In poor conductors with few carriers, such as ice, a slight displacement of electrons with respect to their nuclei produces a dielectric polarisation of the material, leading to displacement currents (Telford et al., 1990). However, the dielectric mechanisms of ice can become important only for very high frequencies (in the MHz range, used in ground penetrating radar, GPR). Generally, the resistivity values depend mainly on the material type and the
unfrozen water content in the sample (Hoekstra and McNeill, 1973, Olhoeft, 1978). A list of resistivity values for common materials is shown in Fig. 3.2. Note, that the large range of resistivity values for most materials is due to varying water content (see e.g. Table 5.4. in Telford et al., 1990).

### 3.2.2 Seismic velocity

Seismic techniques make use of changes in the compressional- and shear-wave velocities of rocks and soils (see chapter 6). Upon freezing compressional- and shear-wave velocities of most materials increase sharply (see Fig. 3.1b). This increase is more pronounced the larger the porosity of the material (McGinnis et al., 1973). An increase in pore-water salinity reduces this effect near the freezing point, as less water become frozen (Pandit and King, 1978). Similar to resistivity, the increase in velocity upon freezing is closely related to the decrease in unfrozen water content (e.g. King et al., 1988, Leclaire et al., 1994). However, for rocks, there is an important difference between the behaviour of resistivity and seismic velocities at subzero temperatures. While the resistivity continues to increase even at very low temperatures, when the pore space is nearly filled with ice, seismic velocities reach a plateau, where further cooling produces very little change (Pandit and King, 1978, Pearson et al., 1983). This illustrates a fundamental difference in the mechanisms by which electrical and seismic signals are transmitted in rocks. Electric conduction
Figure 3.3: Range of compressional (or P-wave) velocities for different material types and rocks (compiled after Röthlisberger, 1972, Telford et al., 1990, Milsom, 1996).

takes place in the unfrozen portion of the pore water, so electrical properties remain sensitive to the amount of unfrozen water present, even if the unfrozen water content becomes very small. In contrast, seismic wave energy is transmitted primarily through the solid matrix, so once the pore volume is largely filled with ice, a further decrease in the already small unfrozen water content produces only a negligible change in velocity (Pearson et al., 1983).

This contrast may also be seen by comparing Figs. 3.2 and 3.3. While the resistivities of permafrost materials are very high with almost no upper bound, the compressional velocity of ice is confined to values around 3500 ms\(^{-1}\), which is less than the velocity of most rock types.

Because determining the compressional wave velocity requires only the analysis of the first arrival times, it is easier to determine than the shear-wave velocity. Due to its slower velocity shear waves are more difficult to identify and require the processing of a larger part of the seismogram, as well as the use of high-energy seismic sources. Consequently, the analysis of seismic velocities is focused on the compressional velocity in the remainder of this work.
3.3 Geophysical survey and interpretation techniques

There are several geophysical methods to determine subsurface structures by measuring the change of physical properties of the ground. The methods used in this study include DC resistivity, frequency-domain electromagnetics (FEM), time-domain electromagnetics (TEM), refraction seismics and microwave radiometry. The first three methods make use of changes in the electrical resistivity, as described in section 3.2.1. The refraction seismic method is based on changes in the seismic velocity (section 3.2.2) and the microwave radiometry method determines the temperature of the ground by measuring the emitted radiation in the microwave range.

Apart from differences in the measurement setup of the various methods, which are described in detail in the following chapters, there are common survey and interpretation techniques. Geophysical mapping determines lateral changes of a physical property independent of depth. Vertical soundings are used to determine the variation of a physical property with depth. Geophysical imaging is used to visualise the internal structure of the subsurface. Finally, tomographic inversion attempts to determine the full 2- or 3-dimensional character of the variation of the physical property in the subsurface.

3.3.1 Mapping

Mapping the lateral changes of a physical property is the simplest geophysical survey technique. It is suitable for many permafrost applications, e.g. mapping isolated ice lenses or determining the lateral extent of the permafrost distribution. Mapping surveys are conducted by traversing along a survey line (or grid), while keeping all other survey parameters constant, e.g. the geometry of the instrument sensors. Interpretation of data is mainly qualitative and no information about vertical changes can be obtained. However, survey speed is fast and data processing is usually very simple. In DC resistivity surveys, mapping is usually performed using the Wenner electrode array (see chapter 4), which is moved along a straight line. Changes in the measured apparent resistivity can be related to lateral heterogeneities in the subsurface material. In FEM surveys, a conductivity meter, like the EM-31 and the GEM-300, can be used to map the bulk conductivity of the uppermost few metres at walking speed (see chapter 5). TEM methods are working in the time-domain and by this conducting vertical soundings (see next paragraph). In refraction seismics, the simplest interpretation technique is to map lateral variations of the P-wave velocity or variations of the first subsurface layer thickness along a survey line. The radiometry method is another example for a mapping technique, as the lateral variation of the bottom temperature of the snow cover (BTS) is mapped, independent of vertical variations.
3.3.2 Vertical sounding

In contrast to mapping, vertical soundings attempt to determine the vertical variation of a physical property. For example, the rock glacier base or the depth of the permafrost table can be determined. Soundings are commonly conducted by keeping the measurement location constant, but changing the geometry of the instrument sensors. To interpret the data from such a survey, it is normally assumed that the subsurface consists of horizontal layers. A 1-dimensional model of the subsurface is used to interpret the measurements in terms of geology. DC resistivity soundings are commonly conducted using the Schlumberger array, even though other array types are applicable as well (see chapter 4). Electromagnetic soundings are usually performed using time-domain electromagnetic instruments, like the PROTEM system (see chapter 5). In addition to the measurement location, the measurement geometry is kept constant as well and signals from different depths are obtained at different time instances. In refraction seismics, velocity and thickness of different layers can be obtained by using different source-receiver distances. The radiometry method cannot detect vertical changes in temperature, except for variable frequencies. The most severe limitation of sounding methods is that the assumption of lateral homogeneity is rarely justified. A lateral variability of the measured property will cause changes in the measurement results, which might be misinterpreted for vertical changes of that property.

3.3.3 Imaging

More accurate models are obtained by allowing the physical property to change in the vertical, as well as in the horizontal direction. By this, the internal structure of the subsurface can be determined, e.g. internal layering of a rock glacier. To visualise data from a 2-dimensional imaging survey, each measurement is placed on a 2-dimensional plot according to the midpoint and the geometry of the instrument sensors used for the measurement. In 2D DC resistivity surveys, the results are sometimes presented as pseudosections, where the raw data are used to give rough visual impressions of the resistivity distribution. However, pseudosections give only a very approximate picture of the true subsurface resistivity, as the resulting resistivity contours depend on the array type used as well as on the true subsurface resistivity. The same principle can be applied to FEM methods, where the frequency is varied in order to get an additional depth resolution, and to TEM methods, where the location of the measurement centre is varied in order to get an additional horizontal resolution. Other examples for imaging techniques are reflexion seismics and ground penetrating radar (GPR). For both methods signals from the subsurface may be displayed in 2D or 3D representations without additional data inversion, e.g. a migration of the data. The difference between imaging techniques and the tomographic inversions described in the next paragraph is often small. However, in the case of DC resistivity (or electromagnetic methods) the difference between a pseudosection and a tomographic inversion result is large, as will be seen in chapter 4.
3.3.4 Tomographic inversion

To improve data interpretation from a 2-dimensional imaging survey, a tomographic inversion can be performed. Tomography (tomos (gr.) = slice) provides images of the interior of an object from measurements at its surface. For that reason the interior is divided into a number of model blocks, each having a different value of the respective physical property, e.g. specific resistivity or seismic P-wave velocity (see Fig. 3.4). The inversion method tries to find the model of the subsurface, whose response agrees best with the measured data. A least-squares inversion scheme is commonly used to determine the values of the physical property for each model block. This method is applied to the DC resistivity and refraction seismic data sets of this study (chapters 4 and 6), but is applicable to electromagnetic methods as well. The general theory of the inversion technique is explained in detail in the next section.

3.4 Inversion theory

The general aim of a geophysical inversion is to find the subsurface model, whose response fits the observed data best. A brief description of the posing and the solution of this so-called inverse problem is presented in this section. The terminology, the notation and the derivation of the equations are taken from Menke (1989).

3.4.1 Inverse and forward problem

Consider the problem of fitting a straight line to a set of observation data. The input data for the inversion consist of a set of observation data, $d$, which are used to determine specific model parameters, $m$. In this case, the model parameters are
the slope and the intercept of the line. A necessary condition for an inverse problem is that model parameters and data are in some way related. This relationship is called model and has to be known (or guessed) beforehand. Fitting a straight line to observation data is assuming a linear relationship between model parameters and data, that is a linear model.

The term inverse problem is used in contrast to forward problem, which is defined as the reverse process, where a data set is predicted on the basis of known model parameters. The connection between forward and inverse problem is given as follows:

Inverse problem:

observation data $\rightarrow$ model $\rightarrow$ estimates of model parameter $d \rightarrow f(d, m) \rightarrow m^{est}$

Forward problem:

model parameters $\rightarrow$ model $\rightarrow$ prediction of data $m \rightarrow f(d, m) \rightarrow d^{pre}$

Usually, solving the forward problem is the first step of solving the corresponding inverse problem. Most generally the relation between data and model parameters (= the model) can be written as:

$$f_i(d, m) = 0, \quad i = 1, \ldots, N \quad (3.1)$$

where $N$ is the number of equations, i.e. the number of data points. The purpose of inversion is to solve these equations for the unknown model parameters.

The simplest and best-understood inverse problems are those that can be represented with the explicit linear equation

$$d = Gm \quad (3.2)$$

where $G$ is called the data kernel. Here and in the following bold letters denote vectors and bold capital letters denote matrices. Using again the example of fitting a straight line to some observed data set, e.g. the change of temperature, $T$, with depth, $z$, at an investigation site, we can rewrite Eq. (3.1)

$$T_i = a + b z_i, \quad i = 1, \ldots, N \quad (3.3)$$

with $N$ the number of measurements. Here, the observation data $d$ consist of the values of $T_i$ at depths $z_i$, and the model parameters $m^{est}$ are the slope $b$ and the intercept $a$, respectively. Eq. (3.3) can be arranged in form of the matrix equation (3.2), giving

$$\begin{pmatrix} T_1 \\ T_2 \\ \vdots \\ T_N \end{pmatrix} = \begin{pmatrix} 1 & z_1 \\ 1 & z_2 \\ \vdots & \vdots \\ 1 & z_N \end{pmatrix} \begin{pmatrix} a \\ b \end{pmatrix}. \quad (3.4)$$
3.4.2 Solution of the inverse problem

One of the simplest methods for solving the linear inverse problem (Eq. (3.2)) is to minimise the sum of the squared prediction errors, or misfits,

\[
E = \sum_{i=1}^{N} e_i^2 = \sum_{i=1}^{N} (d_i - d_{i}^{\text{pre}})^2,
\]

which is called the least-squares method. In case of the line-fitting example above, this corresponds to minimising the sum of the squared distances between each observation data point and the corresponding point on the line. We therefore seek values of the model parameters \( m^{\text{est}} \) that lead to predicted data, \( d^{\text{pre}} \), minimising Eq. (3.5). In matrix notation this leads to

\[
E = (d - Gm^{\text{est}})^T(d - Gm^{\text{est}}),
\]

where \( T \) denotes the transpose of a matrix. Minimising \( E \) is done by setting the derivatives of \( E \) to zero with respect to the model parameter, \( m^{\text{est}} \). A square matrix equation for the unknown model parameters is obtained, given as

\[
G^T Gm^{\text{est}} - G^T d = 0.
\]

Presuming that the inverse \([G^T G]^{-1}\) exists, we get

\[
m^{\text{est}} = [G^T G]^{-1} G^T d
\]

which is the least-squares solution to the inverse problem \( Gm = d \). The details of this derivation can be found in Menke (1989).

3.4.3 Over-, under- and mixed-determined problems and damped least-squares solution

In the derivation of the above solution to the inverse problem, the question of existence and uniqueness of the solution has not been discussed. Only in the special case, where there are as many unknowns as independent data, a unique solution with zero prediction error exists, e.g. fitting a line through two points. If there are less data than unknowns, e.g. fitting a line to only one point, several solutions can have zero prediction error. The above mentioned inverse \([G^T G]^{-1}\) does not exist and the least-squares solution (3.8) fails. These cases are called purely underdetermined problems. On the contrary, if there are more data than unknowns, e.g. fitting a line to more than two independent data points, the problem is called purely overdetermined as no solution can yield zero prediction error.

In most geophysical applications mixed-determined problems are encountered, where some model parameters are overdetermined and some are underdetermined. Consider the schematic principle of a refraction seismic tomography experiment shown
Figure 3.5: Schematic plot of idealised tomography experiment with two model blocks. Thick lines denote ray paths and circles and crosses denote sources and receivers, respectively. (a) Velocity $v_1$ is overdetermined, velocity $v_2$ is purely underdetermined. (b) Velocity $v_1$ and $v_2$ are both even-determined. (c) The average velocity of $v_1$ and $v_2$ is overdetermined, but the individual velocities are both underdetermined, since each path has an equal length in either model block (after Menke, 1989).

In Fig. 3.4 and as explained in detail in chapter 6. The measured seismic traveltime data are inverted to yield the seismic velocity in each model block. The model equations for the traveltime data can only be evaluated along the ray paths of the waves, which are determined through the velocity distribution in the model blocks. Because of measurement geometry and velocity distribution, there may be model blocks through which several rays pass and others that have been missed entirely. This case is shown in Fig. 3.5, which represents an idealised seismic experiment, where source and receiver are placed at opposite ends of two model blocks. In Fig. 3.5a, the left model block is overdetermined since measurements are made along two different paths. On the other hand, the right model block is purely underdetermined, as it has been missed entirely by the rays. In Fig. 3.5b, both model blocks are even-determined, as there is only one solution for the velocity in each model block and the prediction error is zero. There may also be model blocks that have more data than unknowns, but cannot be individually resolved, because every ray that passes through one also passes through an equal distance of the other (Fig. 3.5c). These model blocks are also underdetermined, since only their mean velocity is determined.

In these cases further constraints are required to obtain an unambiguous solution to the problem. In principle, these constraints could be any relation of the model parameters, which could be included in the misfit function $E$ (Eq. (3.5)) by way of Lagrange multipliers. Most commonly the constraint of small model adjustments during the inversion (e.g. the so-called damping constraints of Marquardt (1970)) and/or constraints on the model smoothness are used. This is achieved by using an extended version of Eq. (3.2), i.e.
where $h = D\hat{m}$ is of the form

$$
\begin{pmatrix}
d' \\
h'
\end{pmatrix} = 
\begin{pmatrix}
G \\
D
\end{pmatrix} \hat{m},
$$

(3.9)

Hereby, $I$ is the identity matrix and $S$ denotes a Laplacian smoothing matrix. The parameter $\lambda$ controls the total amount of regularisation applied, and $\beta$ ($0 < \beta < 1$) determines the proportional weights of damping and smoothing (Lanz et al., 1998).

The regularisation factor $\lambda$ determines the relative weight of the additional constraints relative to minimising the prediction error. This concept will be explained in more detail in chapter 4 for the resistivity tomography and in chapter 6 for the refraction seismic tomography.

### 3.5 Non-linearity

Since for most geophysical applications the model functions are non-linear, a linearisation of the above equations has to be performed. This can be done by applying a Taylor expansion to the model function about the estimated model parameters, $m^{est}$. Ignoring terms of second and higher order, the linearised model equations are given as

$$
\Delta d = \hat{G} \Delta m,
$$

(3.11)

where $\Delta d = d^{obs} - g(m^{est})$, $\hat{G} = (\partial g/\partial m^{est})$ is the Jacobian matrix relating changes in the model data to changes in the model parameter and $\Delta m = m - m^{est}$ is the model parameter misfit (Menke, 1989). Eq. (3.11) must be solved iteratively with the new estimated model parameters $m^{est}$ determined from the data misfit $\Delta d$.

Non-linear effects can play a major role in the inversion process. Due to non-linearities not all model parameters can be determined equally well, as the determination of the model parameters may depend on the model parameters itself. For example in DC resistivity surveys, the injected current will predominantly flow in conductive regions of the subsurface, which will hinder the determination of the resistivity in the more resistive regions. If a highly conductive region is present at shallow depth along the whole survey line, all current will flow along these conductive flow paths and no information about deeper, more resistive layers can be obtained. In refraction seismic tomography inversions, the model equations are evaluated along seismic ray paths (see Fig. 3.4 and chapter 6) and the seismic velocity can only be determined of those model blocks, which are passed by a ray (c.f. Fig. 3.5). But the flow path of each ray depends on the velocity distribution itself. Consequently, some model blocks will not be passed by rays at all during inversion and their velocity cannot be determined.
3.6 Accuracy, resolution and equivalence

In order to estimate the quality of the inversion results the model resolution and accuracy may be analysed. Both quantities are strongly influenced by the number of model parameters, that is the number of model blocks in a tomographic inversion. If many model parameters are selected, the accuracy of these parameters may be low, whereas the resolution of the inversion result is high. If only a few model parameters are selected, the accuracy is high, but the resolution low. That means, there is a trade-off between accuracy and resolution in choosing the number of model parameters for a given data set.

In the absence of boreholes, geophysical techniques are usually restricted to viewing angles from the surface looking down. This geometrical constraint usually results in a decrease of the sensitivity of the model parameters to the data with depth. One possibility is to increase the block size with depth leading to fewer model parameters and therefore higher accuracy at larger depths. At shallow depth, where the sensitivity is usually largest, a higher resolution is often achievable.

Furthermore, there are limitations inherent to the physical methods. In refraction seismic tomography, the spatial resolution is connected to the flow paths of the rays. In model regions with a high ray coverage the resolution is large, whereas it is very low in regions with poor ray coverage. This leads to a sharp, but locally incomplete image of the subsurface. In DC resistivity tomography, the spatial resolution is smaller, but in contrast to refraction seismsics, reasonable estimates of all model parameters are usually obtained.

Finally, in DC resistivity uncertainty arises from the principle of equivalence, which states that two highly resistive bodies with different resistivities and thicknesses may give the same response, if the product of their thickness and resistivity values, $zp$, is the same. Furthermore, two highly conductive bodies will give the same response, if the ratio between their thickness and resistivity values, $z/\rho$, is the same (e.g. Telford et al., 1990).

These limitations lead to models, which are usually non-unique, requiring e.g. additional geophysical, geological or borehole information to reduce the range of plausible models.
Chapter 4

DC resistivity tomography

4.1 Introduction

The main physical parameter used in geophysical studies for differentiating between frozen and unfrozen material is the electrical resistivity or its reciprocal, the electrical conductivity (section 3.2.1).

From the various geophysical survey techniques shown in section 3.3, mainly 1-dimensional (1D) vertical electrical soundings (so-called Schlumberger soundings) have been used in resistivity surveys on mountain permafrost (e.g. King et al., 1987, Vonder Mühll, 1993, Ødegård et al., 1996, Wagner, 1996). However, typical targets in mountain permafrost include isolated bodies, like ice lenses or ice-cored moraines. These can often not be characterised adequately with vertical electrical sounding methods, which are only capable of providing layered Earth models (see section 3.3). For an investigation depth of 30 m the spacing between the outermost electrodes is around 200 m (Edwards, 1977). Therefore, 1D interpretation methods assume laterally homogeneous ground conditions over a distance of 200 m, which is rarely the case in mountain permafrost studies.

In some studies resistivity mapping using the Wenner array was used to map lateral variations in the permafrost distribution (e.g. Osterkamp and Jurick, 1980). In contrast to Schlumberger soundings, Wenner mapping may provide information on lateral changes of the subsurface resistivity, but cannot detect vertical variations.

A powerful alternative to sounding and mapping is 2-dimensional (2D) DC resistivity tomography, which is suitable for areas with moderately complex geology (Griffiths and Barker, 1993). In the last few years 2D and even 3D DC resistivity tomography has been widely used in a variety of applications in environmental and engineering studies (e.g. Daily et al., 1992, Griffiths and Barker, 1993, Binley et al., 1996, Johansson and Dahlin, 1996, Mauriello et al., 1998, al Hagrey and Michaelsen, 1999, Dannowski and Yaramanci, 1999, Ogilvy et al., 1999, Olayinka and Yaramanci, 1999).

2D resistivity tomography was applied on most of the field sites of this study (see Table 2.1). In order to check the reliability of the method on mountain permafrost,
Figure 4.1: Basic setup of DC resistivity surveys: a current is passed between electrodes A and B; by measuring the resulting potential between electrodes M and N the resistivity \( \rho \) of the underlying layers may be determined.

synthetic modelling and subsequent inversions have been performed as well (section 4.3). Additionally, laboratory studies using a miniature DC resistivity system were performed to model freeze and thaw processes under controlled conditions (section 4.4). Finally, the method was applied to monitor temporal resistivity changes at the Schilthorn test site, where the results could be related to temperature data from the PACE borehole (section 4.5). Additional case studies are discussed in chapter 8 and shown in the appendix.

4.2 Theory

In DC resistivity surveys the ground resistivity is measured by passing a current between a pair of grounded electrodes (A, B) and measuring the potential between a second pair of electrodes (M, N, see Fig. 4.1).

An equation for the potential distribution due to a point current source \( I_s \) located at point \( x_s \) at the surface can be derived from

**Ohm’s law:**

\[
j(x) = \sigma(x)e(x)
\]

(4.1)

and the

**divergence condition**

\[
\nabla \cdot j(x) = I_s \delta(x - x_s),
\]

(4.2)

where \( e \) is the electric field [in V/m], \( j \) is the current density [in A/m²], \( \sigma \) is the conductivity of the medium [in S/m] and \( x = (x, y, z) \) (Keller and Frischknecht, 1966). The time independent form of the first Maxwell equation, \( \nabla \times e = 0 \), implies the existence of a scalar electric potential:

\[
e(x) = -\nabla \Phi(x),
\]

(4.3)

which may be combined with Eq. (4.1) and Eq. (4.2) to give

\[
\nabla \sigma(x) \cdot \nabla \Phi(x) + \sigma(x) \nabla^2 \Phi(x) = -I_s \delta(x - x_s).
\]

(4.4)
4.2.1 Homogeneous half-space model

Assuming a homogeneous half-space Earth model, the first term on the left hand side of Eq. (4.4) vanishes and the potential caused by a current source located at $x = (0, 0, 0)$ is given by

$$\Phi(x) = \rho I_s \frac{1}{2\pi |x|},$$

(4.5)

where $\rho = \frac{1}{\sigma}$ is the resistivity and $|x|$ is the distance from the origin. Hereby, the boundary conditions $\Phi = 0$ for $|x| \to \infty$ and $\Phi \to \infty$ for $x = (0, 0, 0)$ are applied.

Since potential functions can be added arithmetically, the total potential at one observation point may be calculated by adding the potential contributions from each source. The potential difference between two potential electrodes, M and N, induced by a pair of current electrodes, A and B, is then given by

$$\Phi_M - \Phi_N = \Delta \Phi = \rho I_s \frac{1}{2\pi} \left( \frac{1}{AM} - \frac{1}{BM} - \frac{1}{AN} + \frac{1}{BN} \right),$$

(4.6)

where $AM$ denotes the distance between current electrode A and potential electrode M. Hereby, the minus sign for two of the distance terms arise since one of the current electrodes in a normal two-electrode current must have a negative sense of current flow compared to the other.

4.2.2 Two- and three-dimensional half-space model

When $\sigma$ is allowed to vary over a full 2D or 3D half-space Earth model the first term in Eq. 4.4 does not vanish. Integrating over volume $V$

$$\iiint_V (\nabla \sigma(x) \cdot \nabla \Phi(x) + \sigma(x) \nabla^2 \Phi(x)) \, dx \, dy \, dz = \iiint_V (-I_s \delta(x - x_s)) \, dx \, dy \, dz$$

(4.7)

and applying Green's theorem yields

$$\iint_S \sigma(x) \frac{\partial \Phi(x)}{\partial n} \, dS - \iiint_V \nabla \sigma(x) \cdot \nabla \Phi(x) + \iiint_V \nabla \sigma(x) \cdot \nabla \Phi(x) = I(x),$$

(4.8)

which simplifies to

$$\iint_S \sigma(x) \frac{\partial \Phi(x)}{\partial n} \, dS = I(x),$$

(4.9)

where $n$ is the unit vector normal to the surface.

Using the finite-difference discretisation of Dey and Morrison (1979) this leads to a matrix equation of the form
\[ G \Phi = I, \]  

(4.10)

where \( G \) is the conductance matrix consisting of the discretised conductivities and \( \Phi \) are the discretised potentials. The generally sparse conductance matrix \( G \) can be inverted using a sparse matrix solver to give the potentials over the whole 2D- or 3D model grid.

### 4.2.3 Geometric factor and apparent resistivity

The quantity that is actually measured in DC resistivity surveys is the potential difference between the two potential electrodes (M and N in Fig. 4.1). For a homogeneous Earth Eq. (4.6) can then be used to calculate the resistivity \( \rho \). Hereby, the terms can be rearranged to obtain

\[ \rho = K \frac{\Delta \Phi}{I}, \]  

(4.11)

where \( K \) is called the geometric factor combining the effect of electrode separation distances (Keller and Frischknecht, 1966).

If the subsurface is non-uniform, the so-called apparent resistivity \( \rho_a \) is determined from Eq. (4.11), which is a diagnostic quantity to some extent, but only in the case of homogeneous grounds it is equal to the actual resistivity.

### 4.2.4 Electrode array geometries

Eqs. (4.12) to (4.14) provide the geometric factors for the three electrode geometries used in this work (Fig. 4.2):

**Wenner:**

\[ K = \frac{2\pi}{a - \frac{1}{2a} - \frac{1}{2a} + \frac{1}{a}} = 2\pi a \]  

(4.12)

**Wenner-Schlumberger:**

\[ K = \frac{2\pi}{\frac{1}{na} - \frac{1}{a+na} - \frac{1}{a+na} + \frac{1}{na}} = \pi n(n + 1)a \]  

(4.13)

**Double-Dipole:**

\[ K = \frac{2\pi}{\frac{1}{a+na} - \frac{1}{na} - \frac{1}{2a+na} + \frac{1}{a+na}} = \pi n(n + 1)(n + 2)a \]  

(4.14)

where \( a \) is the electrode spacing distance and \( n \) is the multiple which determines the distance between the two current and two potential electrodes in the Wenner-Schlumberger and Double-Dipole array (Telford et al., 1990). As shown in Fig. 4.2, the Wenner array uses the two outer electrodes as current electrodes with the two potential electrodes between them. The spacings between the electrodes are always
Figure 4.2: Schematic plot of array geometries used in this study. The factor $a$ denotes the minimal electrode spacing, the multiple $n$ determines the distance between the two current and two potential electrodes for the Double-Dipole and Wenner-Schlumberger array.

multiples of the minimum spacing $a$. This array configuration minimizes the time used for a complete survey, as the number of individual measurements required is relatively small. The resolution is best for vertically layered media, but small scale lateral heterogeneities remain often unresolved.

The Double-Dipole array uses the current electrodes on one side, the potential electrodes on the other side, thus forming two 'dipoles'. The spacing of the dipoles ($a$ in Fig. 4.2) is always equal or smaller than the spacing between the dipoles (denoted as $na$, where $n = 1, 2, \ldots$). This configuration has a much better horizontal resolution but the penetration depth is smaller than for the Wenner array. Moreover, a larger number of measurements is needed.

The Wenner-Schlumberger array is a combination of a Wenner array and the Schlumberger sounding array, where the midpoint is kept fixed and electrode spacings are increased logarithmically (Pazdirek and Blaha, 1996). Like for the Wenner array, the outer electrodes are used as current electrodes and the inner to measure the potential. The spacing between the potential electrodes is kept constant ($a$ in Fig. 4.2), but the other spacings are increased ($na$ in Fig. 4.2) to enhance depth resolution, compared to the Wenner array. This results in a slightly increased number of measurements, but not as much as for the Double-Dipole array.
4.3 Application of DC resistivity tomography to permafrost studies

4.3.1 Data acquisition

For the acquisition of the apparent resistivity data in this study an ABEM Terrameter SAS 300 C and an ABEM Lund multi-electrode system were used. The system automatically measures the apparent resistivities for a series of electrode combinations for a given array geometry. In most experiments, 41 equally spaced electrodes with a standard spacing of 5 m were used, which resulted in survey lines of 200 m length. With the 41 electrodes available, 190 possible Wenner configurations could be implemented. Depending on the quality of the recorded voltages, the measurements for each electrode configuration were repeated up to 16 times. Acquiring a full Wenner array data set required about 1.5 hours.

Good electrical coupling between the electrodes and subsurface is essential for DC resistivity surveys. Mountain permafrost sites are usually extremely dry. They may include piles of large boulders at the surface. Therefore, it may be very difficult to establish reasonable electrical contact. The problem is resolved by attaching sponges soaked in salt water to the electrodes. This provides a good electrical contact, but decreases the resistivities in the immediate vicinity of the electrodes. Extensive tests have shown that this effect is only significant in a cylindrical area of a few centimetres radius around the electrodes and does not affect the inversion results (Vonder Mühll, 1993).

Measurements were conducted on most field sites as described in chapter 2. In this section, one example from the ice-cored moraine near Gaudegg (Zermatt/Switzerland) is presented. The field site is described in section 2.6 and the location of the survey line is shown in Fig. 2.10 (line A).

4.3.2 Inversion

The apparent resistivity data sets were inverted using the software package RES2DINV (Loke and Barker, 1995, 1996a). This program solves the tomographic inversion problem using a smoothness-constrained least-squares method (deGroot-Hedlin and Constable, 1990) and produces a 2-dimensional subsurface model from the apparent resistivity pseudosection. Using the approach outlined in section 3.4, the resistivity of the rectangular blocks (= the model parameters \( m \)) that will minimize the differences between the calculated and measured apparent resistivity values (= observation data \( d \)), can be determined. As the inverse problem is non-linear, the linearised version of Eq. (3.9) has to be used (analogous to Eq. (3.11)). The smoothness-constrained least-squares method as used in RES2DINV has the following form:

\[
\Delta m_i = (J_i^T J_i + \lambda_i S_i^T S_i)^{-1} J_i^T \Delta d_i, \tag{4.15}
\]
which is of the form of the general least-squares solution (Eq. (3.8)). Here, $i$ is the iteration number, $\Delta d_i$ is the discrepancy vector which contains the differences between logarithms of the measured and calculated apparent resistivity values, $\lambda_i$ is the regularisation factor and $\Delta m_i$ is the perturbation vector to the model parameters for the $i$th iteration (Loke and Barker, 1996a). The 2D flatness filter $S$ is used to constrain the smoothness of the perturbations to the model parameters to some constant value (Sasaki, 1992). Note, that in this case regularisation only consists of a smoothing constraint, without additional damping ($\beta = 0$ in Eq. (3.10)).

The inversion algorithm may be divided into five main steps:

- Calculation of the apparent resistivities, $\rho_{a,i}$ for an initial model with resistivities $\rho_i$
- Calculation of the discrepancy vector $\Delta d_i = \rho_{a,i}^{\text{est}} - \rho_{a,i}^{\text{obs}}$, where $\rho_{a,i}^{\text{est}}$ and $\rho_{a,i}^{\text{obs}}$ are the calculated and observed apparent resistivities, respectively
- Calculation of the Jacobian matrix $J_i$ of partial derivatives of the model parameters
- Solving the least-squares equation (4.15) to determine the model parameter perturbation vector $\Delta m_i$
- A new estimate for the model resistivity of each block is calculated by using

$$\rho_{i+1} = \rho_i + \Delta m_i,$$

which becomes the initial model for the next iteration

These steps are repeated until the algorithm converges or a predefined maximum number of iterations is reached. A homogeneous starting model is obtained by calculating the logarithmic average of the measured apparent resistivity values (Loke and Barker, 1995).

**Regularisation**

Eq. (4.15) tries to minimise a combination of two quantities, the difference between calculated and measured apparent resistivities, $\Delta d$, as well as the reciprocal of the model smoothness, the roughness. The regularisation factor $\lambda$ specifies the weighting between the two during the inversion process. The larger the regularisation factor, the smoother the model. As indicated by the subscript $i$ in Eq. (4.15), $\lambda$ changes during the iteration. In RES2DINV the user specifies an initial regularisation value $\lambda_0$ and a minimal regularisation factor $\lambda_{\text{min}}$. If an iteration has resulted in a reduction of the length of the discrepancy vector $\Delta d$, the regularisation factor is reduced by half until $\lambda_{\text{min}}$ is reached (Loke and Barker, 1996a). The length of the discrepancy vector is given as a root-mean-square (RMS) value, that is
Choosing appropriate regularisation factors for a given data set is an important task in order to get a realistic geological model. The initial regularisation factor $\lambda_0$ controls the RMS reduction in the first few iterations, whereas $\lambda_{\text{min}}$ is important to stabilise the inversion process. Loke and Barker (1996a) used a value of 0.2 for $\lambda_0$ and 0.04 for $\lambda_{\text{min}}$ for data sets containing a substantial amount of noise, whereas smaller values are used for delineating sharp boundaries of separated targets. $\lambda_{\text{min}}$ is usually set to about one-fifth the value of $\lambda_0$. Sensitivity studies using different ratios between $\lambda_0$ and $\lambda_{\text{min}}$ have shown that the inversion results depend primarily on $\lambda_{\text{min}}$, except for very noisy data sets.

To estimate the reliability of the results and to understand the behaviour of the inversion algorithm, a generic model of an idealised permafrost environment was constructed (Fig. 4.3a). A highly resistive (500 k$\Omega$m) anomaly, representing the ice-rich permafrost body, is present in a less resistive (25 k$\Omega$m) host material, representing unfrozen debris or bedrock. The boundaries of the anomaly were smoothed out to avoid sharp edge effects. Synthetic data were calculated using the forward modelling tool RES2DMOD (Loke, 1996) for a Wenner array (41 electrodes, 5 m spacing). Random errors were added to the synthetic data to make them comparable to the observed data, the relative accuracy of which was estimated to be approximately 5%. For comparison, Fig. 4.3b shows the resistivity model for the ice-cored moraine at Zermatt, as determined from the field data.

In a first set of inversion tests, the influence of the regularisation factors $\lambda_0$ and $\lambda_{\text{min}}$ was examined. In Fig. 4.4 the relationships between $\lambda_{\text{min}}$ and the RMS of the discrepancy vectors $\Delta d$ are displayed. After quickly reaching a minimum value, the curve for the synthetic data (Fig. 4.4a) shows a gradual increase of the RMS error.
with increasing $\lambda_{\text{min}}$. Even for quite large $\lambda_{\text{min}}$ values the RMS error is below the true 5% error level. This indicates that small $\lambda_{\text{min}}$ values tend to overfit the data, such that data errors are inverted to become structural features in the tomogram. As field data sets invariably contain some noise, this puts a lower limit on the regularisation factor that can normally be used (Sasaki, 1992, Loke and Barker, 1995). Indeed, only for the highest degree of smoothing is the corresponding tomogram free of artefacts (Fig. 4.5e), although the estimated resistivity contrast of the anomaly is too small. Hereby, the background resistivity (25 k$\Omega$m) is reproduced well, whereas the resistivity of the anomaly is too small (100 k$\Omega$m instead of 500 k$\Omega$m, see Fig. 4.3a). This is due to the equivalence principle (see section 3.6) and the large regularisation factor, which prohibits a sufficient perturbation of the initially low model resistivities (Loke and Barker, 1995). With decreasing $\lambda_{\text{min}}$ the shape of the anomaly as well as the anomaly contrast is imaged more accurately, but the tomograms show a number of artefacts (Figs. 4.5c and 4.5a). In addition to the high-resistive anomaly in the centre, positive and negative anomalies with smaller amplitude can be found in the whole model region due to an enhanced representation of the added 5% noise. However, the (true) high-resistive anomaly in the centre is modelled more accurately than in Fig. 4.5e, with an estimated resistivity maximum of 250 k$\Omega$m and 400 k$\Omega$m in Figs. 4.5c and 4.5a, respectively.

The RMS-$\lambda_{\text{min}}$ curve for the observed data (Fig. 4.4b) reaches the estimated error level of 5% at about $\lambda_{\text{min}} = 0.01$. The corresponding tomogram is shown in Fig. 4.5d. Within the uppermost 10 m, the resistivity distribution is fairly irregular with values of 10–100 k$\Omega$m. The significance of these variations can be appraised by considering tomograms with higher (Fig. 4.5f) and lower (Fig. 4.5b) smoothing values applied. Differences between the tomograms in Figures 4.5d and 4.5b are small, and even for the highly smoothed tomogram of Figure 4.5f, a similar resistivity distribution is observed within the uppermost 10 m of the earth.

At greater depth, a pronounced high resistivity (200 k$\Omega$m) zone is observed in the centre of all tomograms in Figures 4.5b, 4.5d and 4.5f. As expected, the resistivity contrast is lowest in Figure 4.5f, but its shape is similar in all tomograms. Underneath this highly resistive anomaly, resistivities decrease again reaching values $< 30$ k$\Omega$m at the bottom of the tomograms for all levels of smoothing. Because this feature is not so pronounced in Figure 4.5f, it is concluded that a lower resistivity zone is a necessary feature of the data, but that its properties are not well constrained.

Convergence

As a convergence criterion the change in RMS error from one iteration to the next, $e_i$, is set. This is given by

$$e_i = \frac{\varepsilon_i - \varepsilon_{i+1}}{\varepsilon_i},$$

(Loke and Barker, 1996a), where $\varepsilon_i$ and $\varepsilon_{i+1}$ are the RMS errors for the $i$th and $(i+1)$th iterations, respectively (Eq. (4.17)). Due to the homogeneous starting model
the RMS reduction is largest for the first iterations and decreases until convergence, as can be seen from Fig. 4.6 for synthetic and field data. The choice of the convergence criterion depends on the noise level in the data. For field data, which usually have a higher noise level, the convergence criterion was set to \( e_t < 5\% \). For synthetic data, with artificially added 5% Gaussian noise, \( e_t < 3\% \) was used throughout this study.

4.3.3 Choice of array type and model sensitivity

Choosing the most suitable array type for a given survey site is an important task prior to any field measurements. Due to their known characteristics two array types were selected for testing of their applicability in mountain permafrost studies: the Wenner and the Double-Dipole array (e.g. Telford et al., 1990). The Wenner-Schlumberger array introduced in section 4.2.4 was only applied in the laboratory experiments (see section 4.4.4).

The Wenner array is robust and sensitive to resolving horizontal structures, but poor in detecting narrow vertical changes. As the signal strength is inversely proportional to the geometric factor \( K \) (Eq. 4.11), it has one of the strongest signals among the commonly used array types (Telford et al., 1990). On the contrary, the Double-Dipole array is sensitive to horizontal changes in resistivity, but relatively insensitive to horizontal structures. It has a better horizontal resolution than the Wenner array. However, it is greatly affected by noise and has a shallower depth of investigation than most other array types. Possibly the most important disadvantage of the Double-Dipole array is the very small signal strength for measurements with comparatively large spacings between the two 'dipoles' (= large \( n \)-factors, Fig. 4.2). To use this array effectively, the resistivity meter should have a high sensitivity and
Figure 4.5: Tomographic inversions of synthetic data set (a, c, e) and data set recorded on the ice-cored moraine near Zermatt (b, d, f) for small, intermediate and large regularisation.
there should be good contact between electrodes and ground, which is difficult to achieve on the debris-covered surface of most mountain permafrost sites.

Figure 4.7 shows a comparison between model results for field measurements with the Wenner and Double-Dipole configuration using the same electrodes along a 200 m survey line on Schilthorn, Switzerland (line E in Fig. 2.8). Here, the permafrost resistivities are comparatively low (between 1 and 10 kΩm), due to 'warm' permafrost temperatures and a rather high clay content (see Vonder Mühll et al., 2000). The contact between electrodes and the ground was rather good, as the surface consists of small grained weathered material. The Wenner data were inverted using the standard regularisation factors as determined above, yielding a model with a fairly small RMS error of 6.9% (Fig. 4.7a). However, it was not possible to invert the Double-Dipole data with a RMS error of less than 35% (Fig. 4.7b), even though a whole range of regularisation factors and inversion parameters were applied. In comparing the model results for both arrays the best agreement is found in the uppermost 8 m. The poor inversion performance of the Double-Dipole array is most pronounced for deeper model layers, where the corresponding spacing between the two dipoles is large.

The smaller penetration depth of the Double-Dipole array can also be seen from the much smaller relative sensitivity values below 10 m depth (Figs. 4.7c and d). The relative sensitivity is a measure of model resolution indicating how strong the calculated apparent resistivities depend on a small change in one of the model parameters, that is in the resistivities of the model blocks. The relative sensitivity is defined as the sum over each column of the sensitivity matrix J_i in Eq. (4.15). It is a measure of the amount of current flowing through a particular cell. The higher the sensitivity value the more reliable is the model resistivity value. In Figs. 4.7c and d the relative sensitivity values are largest in the uppermost model blocks, as
Figure 4.7: Inversion results for two measurements using the same electrodes at Schilthorn, Switzerland (line E in Fig. 2.8): (a) Wenner array and (b) Double-Dipole array. The relative sensitivities are shown in (c) and (d), respectively.

the sensitivity function of a homogeneous Earth model has strong maxima near the electrodes and decreases strongly with distance to the electrodes.

In high Alpine environments, the surface material often consists of even coarser grained debris, compared to the above example. On these surfaces, good contact between electrodes and ground is much more difficult to achieve. Large amplitude signals will seldom be obtained. For this reason the Wenner array was used as the standard array for all surveys presented in this study.

4.3.4 Results

Based on the above considerations the final inversion result for the field data set from the ice-cored moraine near Gandegg was obtained as shown in Fig. 4.8a. The intermediate regularisation factor (Fig. 4.5d) was chosen, representing a compromise between data dependency and model smoothness.
Figure 4.8: Inversion results for data recorded on ice-cored moraine near Zermatt, Switzerland (line A in Fig. 2.10) for $\lambda_{\min} = 0.01$. (a) Tomogram, (b) pseudosection of observed apparent resistivity data, (c) pseudosection of predicted apparent resistivity data, (d) predicted pseudosection–observed pseudosection, and (e) relative sensitivity distribution. The relative sensitivity is the sum over each column of the sensitivity matrix $J$. 
The scattered resistivity values near the surface represent the active layer and sedimentary material with only low amounts of ice. A few patches with resistivities above 100 kΩm may indicate the presence of air-filled cavities or small ice lenses. The large high resistive block in the center of the tomogram is interpreted as the ice core of the moraine. This was further confirmed by refraction seismic survey results, which estimated the seismic P-wave velocity of the resistive block as 3500 m s⁻¹ (= the value for ice, see Fig. 3.3). Its vertical extent coincides well with the earlier results (Keusen and Haeberli, 1983). Low resistivities of around 10 kΩm at the bottom of the tomogram are interpreted as unfrozen material or material with a low ice content. In the eastern part of the tomogram, bedrock crops out (near line C, see Fig. 2.10). The maximum resistivities in this area reach values slightly below 100 kΩm, indicating the presence of fractures filled by interstitial ice within the bedrock. To the west, the survey line extends to the unfrozen part of the moraine with resistivities below 10 kΩm.

Further information on the consistency of the inversion results can be obtained by comparing calculated data based on the final resistivity model with the observed data. This is depicted in Figs. 4.8b and c in the form of pseudosections. The overall match between calculated and observed data is quite good, such that the difference plot in Fig. 4.8d does not exhibit any systematic patterns. However due to the principle of equivalence there exists a general nonuniqueness of resistivity inversion results independent of the agreement between calculated and observed data.

The relative sensitivity values for the inversion model are shown in Fig. 4.8e. As expected, less current flows through the more resistive parts in the center of the tomogram than through other regions. The highest values are observed in the lowermost part of the model, where the resistivities are relatively low. It should be noted that the existence of current in a particular cell is a necessary but not a sufficient condition to resolve the cell resistivity. For example, although there must be a substantial amount of current flowing at the bottom of the model, because there are no intersecting current paths, resistivities of adjacent cells in this lower region may trade off.

### 4.3.5 Conclusion

An application of 2D resistivity tomography to study mountain permafrost was presented for the first time. The aim was to evaluate the feasibility of this method to detect and characterise permafrost in high mountain environments. The particularities of this environment, e.g. the exceptionally high resistivity values up to several MΩm, the large resistivity contrasts and the usually bad electrical contact between the electrodes and the ground required special considerations for data acquisition and data inversion.

Key results of synthetic modelling studies and a field case include:

1. The choice of the regularisation parameter $\lambda_{\min}$ in RES2DINV is critical for the inversion results. It is advisable to consider tomograms computed with a range of $\lambda_{\min}$ values for a meaningful interpretation. The upper limit for $\lambda_{\min}$
is constrained by the estimated level of data errors, whereas the lower limit
is governed by the stability of the inversion. Only those features that appear
in the tomogram with the highest smoothing should be considered as real.
Detailed information on their shape and resistivity contrast may be extracted
from tomograms with decreased smoothness constraints.

2. Isolated high-resistive anomalies are generally resolved. However, the resistiv-
ity maxima tend to be underestimated by the inversion process.

3. For permafrost applications the Wenner array is superior to the Double-Dipole
array. A comparison in the field showed insufficient signal strength of the
Double-Dipole array to give reliable results. This is especially important in
regions with heterogeneous ground conditions and for poor electrode contact.

4. Results from a field study on the ice-cored moraine near Gandegg/Zermatt
delineated successfully the extent of the ice-core showing resistivity values of
100–300 kΩm. The maximum depth of the ice-core was estimated at 30 m.

The results presented in this section suggest a wide applicability of this approach
for permafrost studies, especially for laterally heterogeneous ground conditions.

4.4 Laboratory DC resistivity measurements

4.4.1 Introduction

The application of DC resistivity tomography showed encouraging results for ice-
rich permafrost environments, as seen in the former section. This is mainly because
electrical resistivity of unfrozen material is markedly lower than that of frozen mate-
rial, as the resistivity of water is 4–5 orders of magnitude lower than that of ice (see
Fig. 3.2). If the ice content of the material examined is high, resistivity tomographies
can readily be interpreted in terms of permafrost. However, for materials with a low
ice content, e.g. bedrock, this may become difficult as the difference between the
resistivity of the unfrozen and frozen material is small. In these cases, ambiguities
in interpreting the specific resistivity values in a geologic and thermal context may
occur. Changes in resistivity may be due to differences in temperature, ice/water
content or geologic variations. Additional geophysical methods (e.g. refraction seis-
mics, see chapter 8) or repeated, seasonal field measurements with a fixed electrode
array (section 4.5) help to interpret the resistivity variability of the subsurface.

In addition, laboratory experiments provide detailed information on electrical prop-
erties of frozen soils under physically modelled field conditions. The dependence of
resistivity on temperature and water content may be examined in detail for differ-
ent materials. Laboratory DC resistivity measurements with frozen material have
been reported by various authors (e.g. Østrem, 1967, Hoekstra and McNeill, 1973,
King et al., 1988). Hereby, the main focus was to evaluate the variation of resistiv-
ity with temperature for homogeneous samples of a particular material. However,
this approach does not represent field conditions, which are much more complex
due to non-uniform material, lateral and vertical variations and differences in the ice-content.

In this section the applicability of DC resistivity tomography for laboratory experiments is evaluated. The tomographic approach enables the study of electrical properties of frozen material under simulated field conditions. Lateral and vertical effects can be rebuilt in a miniature 3D-model and measurements can be made under controlled temperature conditions. Furthermore, transient processes like the advance and retreat of a freezing front may be visualised in terms of resistivity variations with time. The results from these laboratory studies will be used in section 4.5 to analyse seasonal DC resistivity field measurements at Schilthorn.

The main objectives of the laboratory experiments is divided into two parts:

1. **Determination of resistivity-temperature relationships in dependence on water content for different materials.** From these important material characteristics the unfrozen water content, the amount of freezing and finally the temperature of permafrost occurrences in the field may be estimated. The seasonal change in resistivity (section 4.5) can be related to temperature if the water content is known approximately. On the other hand the unfrozen water content may be estimated from resistivity and temperature measurements.

2. **Creating a series of resistivity tomographies during freezing and thawing of the samples, to monitor transient processes like the advance and retreat of freezing fronts.** In many laboratory and field studies, where resistivity measurements are conducted, the subsurface is considered laterally homogeneous and irregular freezing and thawing processes are often neglected. 2D resistivity tomographies monitor these transient processes and visualise lateral and temporal changes in resistivity during freezing and thawing.

### 4.4.2 Resistivity, temperature and unfrozen water content

Permafrost material in mountainous regions may be considered as a four-phase medium, consisting of ice, water, air and the rock matrix. The last is usually a composition itself. For saturated media this is reduced to three phases (water, ice and rock) which may be a good approximation for permafrost, except for the active layer of rock glaciers and other coarse-debris permafrost bodies, where air voids between large boulders are present.

**Archie’s law for partially frozen soils**

As pointed out in section 3.2.1, in most Earth materials electric conduction takes place through ionic transport in the liquid phase. A well known empirical relationship called *Archie’s law* (Archie, 1942) relates the resistivity of a 2-phase medium
(rock matrix, liquid) to the resistivity of the water, the porosity of the rock matrix and the water content:

$$\rho = a \rho_w \Phi^{-m} S_w^{-n},$$  \hspace{1cm} (4.19)

where $\rho_w$ is the resistivity of the water in the pore spaces, $\Phi$ is the porosity, $S_w$ is the fraction of the pore space occupied by liquid water and $a$, $m$ and $n$ are empirically determined parameters (e.g. Keller and Frischknecht, 1966).

In partly frozen materials, ionic transport still takes place in the liquid phase. Therefore, the resistivity of frozen material is not depending directly on temperature or ice content, but on the unfrozen water content, which can be substantial even at relatively low temperatures (Anderson and Morgenstern, 1973). Upon freezing, $S_w$ is reduced as ice begins to form, but $\rho_w$ is reduced as well, because of the migration of electrolytes in solution from the freezing water to that remaining liquid. For weak electrolytic solutions an increase in salinity is accompanied by an approximately inversely proportional decrease in resistivity, $\rho_w$ (Daniels et al., 1976). Assuming the pore space of the material was completely filled with water prior to freezing ($S = S_w = 1$ for temperatures above the freezing point), $S$ may be identified as the fraction of water remaining unfrozen at subfreezing temperatures (in the following just unfrozen water content). The resistivity of this ‘freezing water’, $\rho_{w,f}$ is then related to that initially before freezing, $\rho_{w,i}$ by

$$\frac{\rho_{w,f}}{\rho_{w,i}} = S.$$  \hspace{1cm} (4.20)

Eq. (4.19) can then be used to relate the resistivity of a partially frozen material $\rho_f$ to that unfrozen $\rho_i$ by

$$\frac{\rho_f}{\rho_i} = S^{1-n}.$$  \hspace{1cm} (4.21)

(Daniels et al., 1976).

King et al. (1988) conducted laboratory experiments on permafrost samples from the North-American Arctic, measuring both the resistivity and the fraction of the unfrozen water content, $S$. Using Eq. (4.21) they obtained estimates for the so-called saturation exponent $n$ for different unconsolidated permafrost samples. For sands, $n$ was estimated between 2 and 3, for silts between 2.5 and 4 and for clays between 5 and 8. Strictly speaking, the increase in resistivity due to a reduction in temperature below 0°C must also be taken into account (see next section). This effect will tend to reduce the estimates for $n$, typically by about 6% (King et al., 1988). For rocks a value of $n = 2$ (e.g. Keller and Frischknecht, 1966) is generally assumed.

**Resistivity-temperature relationships**

Resistivity-temperature relationships for temperatures near the freezing point have been determined in field (McGinnis et al., 1973, Seguin, 1978, King and Garg, 1980)
and laboratory studies (Harlan et al., 1971, Hoekstra et al., 1975, Olhoeft, 1978, Pandit and King, 1978, Pearson et al., 1983) to facilitate the use of resistivity as a proxy for temperature (see also Fig. 3.1a). In permafrost areas ground resistivity is much easier to determine than ground temperature, as the latter involves drillings in often remote and mountainous areas. The dependence of resistivity on temperature can be separated into two parts: for temperatures above and below the freezing point. At temperatures above the freezing point, a decrease in temperature changes the resistivity of the material only in so far as the resistivity of the pore water is changed. A decrease in temperature increases the viscosity of water, in turn decreasing the mobility of the ions in the water, which increases the resistivity. A relationship between resistivity \( \rho \) and temperature \( T \) (in °C) for temperatures above the freezing point is given by Keller and Frischknecht (1966):

\[
\rho = \frac{\rho_0}{1 + \alpha(T - T_0)}, \quad (4.22)
\]

where \( \rho_0 \) is the resistivity measured at a reference temperature \( T_0 \) (in °C) and \( \alpha \) is the temperature coefficient of resistivity, which has a value of about 0.025 per °C for most electrolytes (Keller and Frischknecht, 1966).

For temperatures below the freezing point resistivities increase exponentially due to the decreasing unfrozen water content, as described above. The results from different laboratory studies (e.g. McGinnis et al., 1973, Pearson et al., 1983) suggest an exponential relationship of the form

\[
\rho = ae^{-bT}, \quad (4.23)
\]

where \( a, b \) are constants and \( T \) is given in °C. Note, that the constant \( a \) is given by the resistivity \( \rho \) at \( T = 0 \). Substituting Eq. (4.23) into Eq. (4.21) the unfrozen water content \( S \) is given as

\[
S = \exp\left(\frac{b(T_f - T)}{1 - n}\right), \quad (4.24)
\]

where \( T_f \) is the temperature of the freezing point in °C (corresponding to \( S = 1 \)). Fig. 4.9 illustrates Eqs. (4.22)–(4.24) for typical values for \( b \) and \( n \). For \( n = 2 \) and \( T_f = 0°C \), Eq. (4.24) describes simply an exponential decrease of the unfrozen water content with decreasing temperature. The factor \( b \) controls the rate of the decrease and can easily be determined from Eq. (4.23) if resistivity data for different subzero temperatures are available. The parameter \( n \) controls the rate by which the resistivity increases as the unfrozen water content \( S \) decreases (see Eq. (4.21)). For \( n = 2 \), \( \rho \) increases at the same rate as \( S \) decreases; the higher \( n \) the higher the unfrozen water content at a given temperature.

### 4.4.3 Experimental procedure

The DC resistivity tomography experiments in the laboratory were conducted in close collaboration with N. Depountis, C. Harris and M. Davies from the Department
of Earth Sciences, University of Cardiff and the School of Engineering, University of Dundee. This group developed a set of miniature electrodes and cables for the use on the laboratory scale. This miniature DC resistivity tomography system was originally developed to monitor the migration of contaminants in soils in scaled centrifuge experiments (Depountis et al., 1999). During a visit at the University of Cardiff in June 1999, the system was used to monitor freeze and thaw processes in a water saturated sand sample. Due to the encouraging results, a similar miniature resistivity tomography system was built at VAW/ETH Zürich in December 1999.

The system includes a standard DC resistivity meter and a set of 24 miniature electrodes connected to a switchbox, which allowed rapid measurements with different electrode combinations. The general setup of the experiments is seen in Fig. 4.10, showing the system built at VAW. The resulting apparent resistivities were inverted using RES2DINV, as explained in detail in section 4.3.

Two different sets of experiments with different materials were conducted at the laboratories in Cardiff and Zürich:

1. DC resistivity tomography measurements for an unfrozen and frozen sand sample to visualise advance and retreat of the freezing front (Cardiff)

2. DC resistivity tomography measurements at various temperatures for gravel and fine material samples from Schilthorn and for different water contents to determine resistivity-temperature relationships (Zürich)
Resistivity tomographies to visualise the freezing front in a sand sample (Cardiff)

In all experiments at the University of Cardiff, a water saturated sand sample (50 cm x 20 cm x 18 cm in size) contained in a plastic box with transparent walls was used. In addition to the miniature electrode set at the surface another 24-electrode set fastened to a rigid, non-conductive frame was placed at the bottom of the box. The electrode spacing of all electrodes was 3 cm. The sand was saturated by filling the box with water from a hole near the bottom and by pouring it on the sample from the top. By this, the sample was not saturated homogeneously which was considered unimportant as the tests were primarily used to evaluate the applicability of the method. Cracks partly filled with water and partly filled with air were visible through the transparent walls of the box. Measurements were performed with a standard Campus resistivity imaging system, which is working similar to the ABEM Lund system described in section 4.3. Due to limitations of the miniature electrode system no cross-sample measurements between the two sets of electrodes could be made. The sample was frozen by using cold air through vortex tubes attached to the surface of the box.

Resistivity-temperature relationships for different materials and water contents (Zürich)

For the laboratory experiments at VAW/ETH Zürich only one set of electrodes was used. Standardly, 24 miniature electrodes with 3 cm spacing were installed at the surface. The miniature electrodes were made of stainless steel with an especially enlarged surface area to ensure electric contact with the frozen surface of the sample. As the measurements were conducted with a standard 4-channel geoelectric instrument (Oyo McOhm), the electrodes were connected to a manual switchbox which allowed rapid measurements of different electrode combinations.

The sample was contained in a 80 cm x 60 cm x 50 cm plastic box with a water outlet at the bottom and a punctuated tube across the bottom floor for water injection. Three temperature mini-loggers (UTL-1, see e.g. Hoelzle et al., 1999) were installed at different depths (see Fig. 4.10). The box containing the sample was placed in a cold room with the cables led outside to the switchbox and the geoelectric instrument. Measurements were recorded from the outside.

4.4.4 Results

Tomographies of sand samples (Cardiff)

Table 4.1 shows the various experiments conducted at Cardiff University. The first measurements were taken after an initial freezing period, which left parts of the sample unfrozen (partly-frozen state). During the night the sample was completely frozen and a second set of measurements was taken the next morning (frozen state).
Finally, the sample was thawed from the top by using heaters directed to the surface. Due to limited time, measurements were taken shortly after the thawing process started, with only the uppermost centimetres of the sample in unfrozen state (thawing state). In addition, measurements with different electrode array geometries (Wenner, Wenner-Schlumberger and Double-Dipole) were conducted to evaluate the suitability of each array for the laboratory measurements. Due to limited time, measurements could not be conducted with the top and base array for all arrays, as indicated in Table 4.1.

Measurement difficulties were encountered in getting proper contact between the surface electrodes and the frozen sample. This was partly due to the small pin size, so that most part of the pin was inside the frozen crust which developed on the surface during the freezing. This was solved using small quantities of warm water around the pins. The data were inverted with the software package RES2DINV as described in section 4.3.

Table 4.1: Laboratory experiments conducted with the sand sample (Cardiff, June 1999).

<table>
<thead>
<tr>
<th>Array Type</th>
<th>Partly-frozen</th>
<th>Frozen</th>
<th>Thawing</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wenner</td>
<td>top, base</td>
<td>top</td>
<td></td>
</tr>
<tr>
<td>Wenner-Schlumberger</td>
<td>top, base</td>
<td>top, base</td>
<td></td>
</tr>
<tr>
<td>Double-Dipole</td>
<td>base</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Figure 4.11: Inversion results for resistivity data set (Wenner-Schlumberger array) of the partly-frozen sand sample. The data sets of the surface and base array were inverted separately, using the same inversion parameters. The depth scale is given in cm from the surface. The range of resistivities representing frozen and unfrozen sand is indicated by the arrows.

Figure 4.12: Inversion results for resistivity data set (Wenner-Schlumberger array) of the frozen sand sample. The data sets of the surface and base array were inverted separately, using the same inversion parameters. The depth scale is given in cm from the surface. Due to bad electrode contact, some electrodes on the right hand side of the base array had to be excluded. The range of resistivities representing frozen and unfrozen sand is indicated by the arrows.
Partly-frozen state The inversion results for the sample in partly-frozen state is shown in Fig. 4.11 with horizontal and vertical scales in cm. From the results of the base array it is seen that the lowermost 5 cm are not frozen showing very low resistivities of 50 Ωm. This is due to drainage water and the migration of ions to the
still unfrozen water during the freezing process (see section 4.4.2). Both processes decrease the resistivity of the still unfrozen part of the sample. The frozen part shows specific resistivity values between 1-10 kΩm. The freezing front in the bottom part of the model is clearly visible. For the surface array the resistivity values are smaller and more spatial variability is seen. This is due to the water distribution in the sample, which was heterogeneous in the uppermost 5 cm, as additional water was put on the surface to enhance connectivity between electrodes and material. The patchy distribution froze irregularly and incomplete, leading to resistivity values between 100 and 500 Ωm. Due to these heterogeneities no distinct freezing front can be delineated.

Frozen state Figure 4.12 shows the inversion model results for the sample after one night of additional freezing. The results for the surface array show resistivities of up to 10 kΩm throughout the whole right part and most of the left part of the sample. This indicates almost complete freezing except for the uppermost layer and two patches in the lowermost part of the surface model (200 Ωm). The freezing front in the base array has moved slightly downwards, indicating that the unfrozen parts at the bottom become more and more frozen. The ability to monitor the propagation of the freezing front with the miniature DC resistivity tomography approach is one of the main results of this pilot study.

Comparison of different electrode arrays Up to now all measurement results shown were conducted using the Wenner-Schlumberger array, which is a good compromise between model resolution and measurement speed (see section 4.2.4). Figure 4.13 shows a comparison of measurements with the three electrode arrays shown in Fig. 4.2 for the base array of the frozen sample. The slightly smaller model depth for the Wenner array compared to the Wenner-Schlumberger array is not due to differences in the array configuration but was caused by bad electrode contact of this Wenner measurement, where some electrodes had to be excluded. However, the Double-Dipole array has always a significantly lower penetration depth (c.f. section 4.3). From Figs. 4.13a and b it is seen that the differences in the results for the Wenner and the Wenner-Schlumberger array are negligible. The results for the Double-Dipole array shows much more small-scale structure, but the magnitude of the values and the large-scale structure is comparable to the other arrays. Note also, that the RMS error is much higher for the Double-Dipole results than for the other arrays. It may be concluded that for vertically layered structures, like the propagation of freezing fronts, the Double-Dipole array gives not many additional insights compared to the other two arrays. The data quality is worse and the measuring time is longer. Hence, for laboratory models without dominant lateral heterogeneous structures, a Wenner or Wenner-Schlumberger array is favoured.

Thawing state Fig. 4.14 shows a comparison between the surface array results for the frozen and thawing state of the sample (Wenner array). The measurements of the thawing state of the sample were done shortly after the thawing process was initiated. The differences in the uppermost 2 cm are clearly visible. The resistivity
values for the thawed region are between 100 and 300 $\Omega$m as opposed to around 1 k$\Omega$m when still frozen. The thawing front is easy to determine. However, patches with a higher resistivity in the thawing compared to the frozen state can be identified as well (marked with the black rectangle). This may be due unfrozen water migrating to the thawing front leaving a dry and more resistive patch of material.

**Resistivity-temperature relationships for different materials and water contents (Zürich)**

The laboratory work at VAW/ETH Zürich (in collaboration with Carolyn Scheurle, VAW/ETH Zürich) was focused on sample material from the Schilthorn field site, in order to compare laboratory and field measurements presented in section 4.5. This material consists mostly of fine grained weathering products of dark limestone schists with a substantial clay content. As a coarse grained reference material, fine gravel with grain sizes between 2.35 mm and 3.1 mm was used. Four measurement series were conducted, each material in a 'dry', that is, without additional water, and in a saturated state. In each case the sample was frozen and thawed continuously. During that time, a full set of 84 resistivity measurements was performed at specific intervals.

**Resistivity-temperature relationships** Figure 4.15 shows the resistivity-temperature curves for the four experiments. The average apparent resistivity of the uppermost layer is plotted against the temperature of the uppermost thermistor. For temperatures above the freezing point the small increase of resistivity with decreasing temperatures is clearly seen. The results are in good agreement with Eq. (4.22) (dashed lines in Fig. 4.15). For temperatures below the freezing point, which is estimated from Fig. 4.15a for the Schilthorn material to be $-0.5^\circ$C, the resistivity increases exponentially with decreasing temperature. From these curves, the factor $b$ in Eq. (4.23) can be estimated for each of the four cases. For both materials the exponential increase is larger in the saturated experiment than for the material in "dry" state ($b = 0.735$ vs $b = 0.273$ for the Schilthorn material and $b = 1.6$ vs $b = 0.12$ for the coarse-grained material). This is in good agreement with previous studies, where the largest resistivity increase due to freezing was found for samples with comparatively high water content (Olhoeft, 1978, Seguin, 1978). Furthermore, comparing the two different materials it is seen, that the rate of increase is larger for the coarse grained material than for the finer grained Schilthorn material in the saturated state and lower in the dry state (Fig. 4.15). Again, this is probably due to the different total water contents. Due to the larger porosity, this is highest in the saturated coarse grained sample and lowest for the coarse grained sample in dry state, as the "dry" Schilthorn material still contains a substantial amount of water, due to the high clay content (e.g. Nobes, 1996). Larger increase rates in resistivity upon freezing for coarser grained samples compared to fine material was also found by Keller and Frischknecht (1966).

**Unfrozen water content** Using the values for $b$, the unfrozen water content can be calculated as a function of temperature for the different materials using Eq. (4.24).
Figure 4.15: Resistivity-temperature relationship determined in the laboratory for four different samples: (a) Schilthorn material saturated with water, (b) Schilthorn material in its initial "dry" state, (c) coarse grained material saturated with water and (d) coarse grained material in its initial "dry" state. An exponential relationship fitting the data below the freezing point (Eq. (4.23)) and Eq. (4.22) for the data above the freezing point is included for each case. Note, that the scale is different for the Schilthorn and the coarse grained material.

In Fig. 4.16 the evolution of the unfrozen water content $S$ with decreasing temperature is shown for the experiments with the saturated Schilthorn material (solid line) and the saturated coarse grained material (dashed line) for $n = 2$ (triangles) and $n = 3$ (squares). Starting with $S = 1$ at the freezing point the unfrozen water content for the Schilthorn material decreases exponentially and reaches 0.06 ($n=2$) and 0.24 ($n=3$) at $-3.9°C$, respectively. For silt and clays $n$ values between 3 and 5 are appropriate (King et al., 1988). From this it may be concluded that the unfrozen water content in the frozen Schilthorn material is still quite substantial at $-4°C$ ($>0.24$). In contrast, for the coarse grained material $S$ decreases rapidly (0.12 at $-1.3°C$ for $n = 2$). Using this value for $n$, as given by King et al. (1988), this results in a very small unfrozen water content for temperatures below $-1.5°C$. 
Evolution of $\rho$, $T$ and $S$ during freezing  Plotting $\rho$, $S$ and $T$ against time for the saturated Schilthorn material, the processes described qualitatively in section 4.4.2 are visualised (Fig. 4.17). Hereby, the direction of the $\rho$-axis is reversed to facilitate the interpretation. As the temperature is approaching the freezing point, the resistivity is increasing slightly, due to a diminished mobility of the ions in the pore water. At the freezing point the temperature curve flattens, as ice begins to form in the pore spaces. At the same time the unfrozen water content starts to decrease. Due to the migration of the ions from the freezing phase to the still unfrozen parts of the sample the freezing point is lowered and the temperature is still decreasing slightly. The resistivity is increasing only slowly, because the migration of the ions decrease the resistivity of the unfrozen water, which nearly cancels out the effect of the decrease of $S$. After more than half of the water is frozen, the temperature curve decreases faster (after ca. 12 hours) accompanied by a fast increase in resistivity, as less and less unfrozen water is available for electric conduction. Finally, the temperature curve flattens again accompanied by a sharp bend in the curve for the unfrozen water content. As the resistivity is very sensitive to further reductions of the unfrozen pore water, it is still increasing at this point.

4.4.5 Conclusions

A miniature DC resistivity tomography system has been tested for application in permafrost studies in the laboratory. The aim was to simulate field conditions in the laboratory to get a better understanding for interpretation of 2D resistivity inversion results of permafrost sections. The effects of spatial and vertical variations in water content, ground material and temperature can be simulated and investigated under controlled conditions and transient processes can be monitored.
Figure 4.17: Resistivity $\rho$ (broken line), temperature $T$ (dashed) and unfrozen water content $S$ (solid) as a function of time for the laboratory experiment with the saturated Schilthorn material. Note, that the $\rho$-axis is reversed. Above the freezing point, the resistivity increases slowly with decreasing temperature. Upon freezing, $S$ is reduced leading to an increase in resistivity, which is nearly cancelled out by the migration of ions to the still unfrozen pore water, which also lowers the freezing point. After nearly all water is frozen the resistivity increases rapidly and temperature decreases again until it is in equilibrium with the temperature of the cooling chamber.

Key results of experiments with different samples conducted at laboratories in Cardiff and Zürich include:

1. Monitoring the evolution of freezing and thawing fronts in partly frozen material is possible with the miniature DC resistivity tomography system. Furthermore, heterogeneities in the sample and patches with non-uniform water saturation can be delineated.

2. A comparison between three different electrode array configurations showed similar results for the Wenner and the Wenner-Schlumberger array, but significant differences for the Double-Dipole array. The Double-Dipole array, with comparatively smaller penetration depth and longer measurement duration, has a much better lateral resolution for small-scale structures. However, the data quality is usually worse and therefore the Wenner or Wenner-Schlumberger array is favoured for predominantly horizontally layered structures.

3. Resistivity-temperature curves show linearly increasing resistivities for decreasing temperatures above the freezing point and exponentially increasing resistivities for decreasing temperatures below the freezing point. The first is due to diminished mobility of ions in the pore water due to an increase in viscosity for decreasing temperatures. Below the freezing point the pore water
starts to freeze, thereby decreasing the unfrozen water content and increasing the resistivity. The results of the experiments show good agreement with theoretical predictions and previous studies.

4. Experiments with different sample material and water content showed largest resistivity increase rates upon freezing for samples with comparatively high total water content.

5. The unfrozen water content of a sample can be estimated as a function of temperature through repeated resistivity measurements.

The miniature DC resistivity tomography system has proven to be very efficient and has great potential for application in permafrost studies in the laboratory.

4.5 Time-lapse DC resistivity tomography to monitor freezing and thawing processes in permafrost areas

4.5.1 Introduction

In addition to using DC resistivity tomography to detect and map permafrost, repeated tomographic surveys may be used to monitor resistivity changes due to freezing and thawing processes, as described in the laboratory experiments. In the past, resistivity measurements in permafrost areas were usually conducted at only one time instance. The resulting resistivity model was then interpreted in terms of the assumed seasonal conditions in the area, depending on the time and date of the measurement. As sufficient electrode contact is difficult to obtain in winter, where the active layer is frozen and a thick snow cover can be present, measurements are usually conducted in the summer season. By this, only an incomplete view of the permafrost conditions is obtained, as short-term variations of the subsurface conditions in the uppermost 5–10 m may strongly influence the measurement results. Repeated measurements throughout the year are required in order to correctly relate resistivity variations to changes in temperature and ice content.

Monitoring of time-dependent processes (time-lapse) through repeated resistivity measurements has been used in hydrogeological tracer experiments in groundwater studies (e.g. Daily et al., 1992, Morris et al., 1996, Binley et al., 1996, Johansson and Dahlin, 1996, Barker and Moore, 1998, Schütze, 1999, Moore et al., 2000). These repeated measurements are usually conducted on a time-scale of hours or a few days to monitor the propagation of artificially induced tracers or natural rain and/or ground water. For permafrost, time-scales of interest are of the order of weeks to months, and the focus is on monitoring the freezing and thawing processes. In section 4.4, resistivity-temperature relationships have been obtained in laboratory experiments and transient processes, like the propagation of the freezing front, have been monitored. The same approach is now used in this section for field conditions
using a fixed electrode array at the Schilthorn drill site in the Swiss Alps. Temperature data from the PACE borehole allow to relate resistivity changes to changes in temperature. The fixed electrode array allows measurements throughout the year, independent of the snow cover thickness. This approach enables to distinguish between resistivity variations due to geology and variations due to temperature.

### 4.5.2 Installation of a fixed electrode array and data acquisition

The PACE drill site at Schilthorn has been described in detail in section 2.5. Permafrost temperatures measured in the borehole are comparatively warm, reaching $-0.7^\circ\text{C}$ in 14 m depth. Consequently, the unfrozen water content is high (cf. section 4.4) leading to comparatively low resistivity values, which are even more decreased through a rather high clay content (Vonder Mühll et al., 2000).

In order to choose a suitable location for the fixed electrode array, several geophysical surveys were conducted prior to the installation. Fig. 4.18 shows the resistivity tomogram of a 200 m long survey line across the drill site (line D in Fig. 2.8). Resistivity values are between 1 kΩm in the shallow subsurface near the borehole and to the east (regions I and II in Fig. 4.18) and 4 kΩm in the west and at larger depths (regions III and IV). From this and the results of a refraction seismic survey (see chapter 8) as well as direct observations at the surface it is concluded that regions III and IV are associated with firm bedrock, whereas regions I and II correspond to unfrozen weathered material.

As the main focus of the resistivity monitoring was the visualisation of the freezing and thawing processes in the active layer, a shallow high-resolution survey line at the Schilthorn drill site was chosen for the fixed electrode array. This line include the more weathered part around the borehole (region I) as well as the firmer bedrock in the west (region III). A set of 30 electrodes with a spacing of 2 m was installed in September 1999 between stations -10 and 48 in Fig. 4.18 (black rectangle). Due to a large perennial snow field at the eastern side of the drill site (see photo of the drill site in Fig. 2.8), the electrode array could not be centred around the borehole.
The installation setup is shown in Fig. 4.19. The 30 stainless steel electrodes were buried 1 m into the ground. Each electrode was connected to a cable via shrinking tubes, which prohibit oxidation of the cable connectors. The cables were connected to a manual switchbox, which is accessible throughout the winter, and were buried for safety reasons in case of avalanches. Resistivity surveys were made by connecting the resistivity meter to the switchbox for each of the selected electrode configurations. This setup allows measurements to be taken throughout the year regardless of the snow cover thickness. Throughout the first year, no problems were encountered with this configuration. Stable, reproducible data were obtained on every survey.

An OYO McOhm resistivity meter was used for data acquisition. Because of the high contact resistances and the presence of a large amount of geological noise the Wenner array was used for the measurements (see section 4.3). One survey consists of 120 data points and each measurement was repeated at least once depending on the variance of the measurements. A whole survey takes approximately 90 minutes. The measured apparent resistivities were inverted using RES2DINV as described in section 4.3. Hereby, a minimal regularisation factor $\lambda_{\text{min}} = 0.01$ and an initial regularisation factor $\lambda_0 = 0.05$ were used for all data sets.

4.5.3 Synthetic modelling of time-lapse DC resistivity tomography data

There are several possibilities of analysing the time series of tomograms with respect to the change in resistivity due to freezing and thawing. Especially, when large lateral resistivity differences due to geologic heterogeneities occur, small temporal resistivity changes are difficult to detect in the model inversion results. Furthermore, noisy data sets, e.g. due to bad electrode contact, tend to mask small resistivity changes. One possibility is to invert the data for the resistivity change between two
sets of measurements, in order to explicitly model the temporal change in resistivity (Schütze, 1999, Loke, 1999, Tsourlos and Ogilvy, 2000). This very promising approach came up only recently, but can not be done with a commercial software like RES2DINV, as the inversion code would have to be adapted. Yet, development of a new inversion routine for time-lapse resistivity tomography data was beyond the scope of this work. However, to reduce systematic inversion errors and to suppress noise, the ratio of two consecutive data sets may be used, as will be shown below using a synthetic data set representing an idealised freezing model.

Starting from a uniform resistivity distribution with a background value of 2 kΩm at time \( t_0 \) (Fig. 4.20a), Fig. 4.20b shows the synthetic model at time \( t_1 \) with a resistivity anomaly of 3 kΩm representing local freezing. In the synthetic modelling study presented in section 4.3, 5% Gaussian noise was added to the apparent resistivities to simulate field conditions. Gaussian noise, as used in the example above, is randomly distributed. This is usually a fair approximation of natural conditions in permafrost studies, when examining a single set of measurements. "Noise" in resistivity studies on permafrost may be introduced by small scale variations of geology, temperature and unfrozen water content or through variations in electrode contact. However, when using a fixed set of electrodes, some of these "noise components" will not change with time. These are mainly the geologic variations and the electrode contacts. This systematic noise can be filtered, using the ratio between the inversion results for two data sets. This results in an anomaly index \( A_n \) for each model block, that is

\[
A_n = \frac{\rho_n(t_1)}{\rho_n(t_0)},
\]

where \( \rho_n \) is the specific resistivity of each model block \( n \) for the two time instances \( t_1 \) and \( t_0 \). The anomaly index \( A_n \) is the amount of increase or decrease of specific resistivity over a time span \( t_1 - t_0 \). \( A_n \) equals 1 denotes unchanged resistivity values. Ideally, the ratio should be taken between the two sets of apparent resistivities and be inverted afterwards (see Schütze, 1999). However, this involves changing the inversion algorithm, which could not be done with the commercial software RES2DINV.

Consider the "unfrozen" and "frozen" model states shown in Fig. 4.20a and b. Assuming a situation where all "noise" is independent of time, the same 5% Gaussian noise distribution was added to both models. Considering the inversion results for both data sets (Figs. 4.20c and d) alone, it is difficult to determine the regions with increased or decreased resistivity values, that is the resistivity change between the model at times \( t_0 \) and \( t_1 \). By plotting the anomaly index \( A_n \), the shape of the freezing anomaly is reproduced quite accurately (Fig. 4.20e), as the systematic noise is filtered from the inversion results. Hereby, the RMS value given in Figs. 4.20c–e relate to the RMS difference between the respective "true" resistivity model and the inversion result. Again, the RMS difference for the anomaly index is much smaller than for both model states alone.
4.5.4 Quasi-monthly resistivity measurements at Schilthorn

Between September 15, 1999 and August 28, 2000 eleven sets of DC resistivity tomography measurements were conducted with the fixed electrode array at Schilthorn. The time span between measurements was roughly 1 month except for the thawing season 2000, where measurements were conducted every 2 weeks (June/July). Ground temperature was measured every 6 hours in the shallow PACE borehole (51/1998) at 17 different depths down to 14 m. An energy balance station measuring all components of energy fluxes from the atmosphere into the ground is located next to the borehole.

Seasonal evolution of the borehole temperatures

Fig. 4.21 shows the temperature evolution in the borehole and the dates of the resistivity measurements. The first electrode array measurement coincided with the summer temperature maximum at 0–2 m depth. A permanent snow cover was already present at the end of October 1999, which can be seen from the smoothed temperature curve at shallow depths. In late February the surface temperatures reached a minimum. Due to heat conduction, this minimum penetrated to larger depth. At 9 m depth, the temperature reached its minimum not before May. In the beginning of May positive air temperatures led to thawing of the snow cover and the upper subsurface layers. The temperature at shallow depth remained almost constantly close to 0°C during this period, due to the phase transition from ice to water. The resistivity measurement at June 6, 2000 was conducted just at the end of this so-called zero curtain effect. Afterwards, temperatures increased until they reached a maximum again in September 2000.

Analysis of the resistivity tomograms

Figure 4.22 compiles the inversion results for the eleven DC resistivity tomography measurements. The resistivity pattern in all tomograms is the same nearly throughout the year: lower resistivities (0.5–1.5 kΩm) in the eastern part of the profile, where the borehole is located, and larger resistivities (1.5–5 kΩm) in the more heterogeneous western part. The latter includes a low resistive weathered layer in the uppermost 1–2 m and a suspicious low resistive patch (marked with the circle) between two high resistive anomalies. The resistivity values correspond to the main structural features seen in Fig. 4.18 (black rectangle) with low resistivities in the weathered zone around the borehole (region I) and larger resistivities for the firmer bedrock to the west (region III). The boundary between these two layers is indicated by the black line in the tomograms in Fig. 4.22.

The largest resistivity changes can be seen between September and October (Figs. 4.22a and b), at the start of the freezing season. Maximum values in the upper layer were reached in April (Fig. 4.22f) and a large decrease was encountered between April and June (Figs. 4.22f and g), as thawing began, until the initial state was reached again at the end of August (Fig. 4.22k).
Figure 4.20: Synthetic modelling results for an idealised freezing model. (a) Resistivity model at time $t_0$ representing a homogeneous unfrozen host material (2 kΩm). (b) Resistivity model at time $t_1$ representing a local freezing anomaly (3 kΩm). (c) Inversion results for the homogeneous unfrozen model containing additional 5% Gaussian noise. (d) Inversion results for the freezing model containing the same artificial noise. (e) Anomaly plot, denoting the ratio between the inversion results in (d) and (c) (Eq. 4.25). The RMS values given in the plot relate to the RMS difference between the respective "true" synthetic model and the inversion result.
Analysis of the resistivity anomaly plots

In contrast to the absolute resistivity values shown in Fig. 4.22, Fig. 4.23 shows the resistivity changes due to freezing and thawing as the ratio between two subsequent readings (see Fig. 4.20e). The extent of the freezing front (blue colors) is clearly seen in the anomaly plot for the October/September measurements. It extends along the whole survey line and reaches a depth of 2 m (see arrow at the left hand side of the plot). Underneath, there are patches with $A_t < 1$, denoting decreasing resistivities. This can be due to several causes: firstly, when a water-saturated sediment starts freezing, electrolytes in solution migrate from the freezing water to that remaining unfrozen, thus decreasing the resistivity of the unfrozen part (King et al., 1988). Secondly and probably more important, the borehole temperatures at 4 m depth were positive and still increased during this period (see Fig. 4.21) representing the phase-lagged begin of the summer warming. Positive temperatures at this depth will lead to melting and hence a decrease in resistivity. Finally, layers with large resistivity values below layers with small resistivity values (or vice versa) can often be seen in model inversion results with extremely high resistivity contrasts. This is due to the equivalence principle explained in section 3.6. By this, an anomalously high (or low) resistivity value could produce an anomaly of the opposite sign below it, without changing the resulting apparent resistivities.
Figure 4.22: Inversion results for the eleven DC resistivity tomography measurements at Schilthorn, Switzerland. The shallow PACE borehole is located at station 10. The black line marks the transition between regions I and III in Fig. 4.18. The circle marks a low resistive zone discussed in the text.
Figure 4.23: Anomaly plots for two consecutive resistivity measurements shown in Fig. 4.22, respectively. $A > 1$ denotes a resistivity increase (by factor $A$) and $A < 1$ a resistivity decrease from one measurement to the next. The arrow in (a)-(e) marks the location of the freezing front. The black line marks the transition between regions I and III in Fig. 4.18. The circle marks a low resistive zone discussed in the text.
Freezing period

From October 1999 to February 2000 no large resistivity changes occurred, except for a small gradual increase in resistivity in the uppermost 1–2 m (Figs. 4.23b and c). This characteristic is due to the insulating snow cover, which arrived in October and prevented further large resistivity changes due to cold air temperatures. The insulating effect of the snow cover, which effectively decouples the subsurface thermal regime from the atmosphere, has been described by several authors (e.g. Goodrich, 1982). Fig. 4.24 shows a comparison between the ground-heat flux and the heat flux through the snow cover at the borehole during winter 1999/2000. Both fluxes were calculated as pure heat conduction based on the temperature gradients between the uppermost thermistors in the borehole and the uppermost thermistor and the snow surface, respectively (see Mittaz et al., 2000). Hereby, the temperature at the snow surface was obtained from the pyrgeometer measurements at the energy balance station. Both fluxes are negative throughout the winter indicating a cooling of the ground, however, the heat flux through the snow cover is much smaller (< 1 W/m²) than the ground-heat flux (5–10 W/m²). This indicates the small influence of atmospheric temperatures on the subsurface in the presence of a snow cover. Gradual freezing of the active layer occurs mainly through the trapped, cold October temperatures as seen by the gradual increase in resistivity between the October and February measurements. During this phase transition the temperatures remained close at 0°C, while the resistivities increased, as the unfrozen water content is diminished. From Figure 4.21 it is seen that this so-called zero-curtain effect started at the end of October and lasted until end of December at 1 m depth, until mid-January at 3 m and until beginning of February at 4 m depth. This gradual downward shift of the freezing front is clearly visible in the anomaly plots (black arrows), where the region of \( A_i > 1 \) gradually extends to deeper layers, reaching 4–5 m at the end of February and 6 m in April (Figs. 4.23b–c).

Melting period

Due to the melting snow cover, temperatures near the surface increased already in March 2000. In the beginning of May the temperature approached 0°C and melting of the active layer started. Again, temperatures remained almost constant at 0°C during the phase transition. But due to the thinning snow cover and positive air temperatures, this spring zero-curtain lasted much shorter. At the time of the first "summer" resistivity measurement (June 2000), most of the frozen water in the uppermost 2–3 m of the active layer had already melted. Together with additional water input by rain, this led to a wet soaked surface layer, decreasing the resistivity strongly near the surface (Fig. 4.23f). The patches with increased resistivity values at 3–4 m depth could be due to freezing anomalies, as the temperatures were still decreasing below 4 m depth (see Fig. 4.21), but are more likely due to inverse model artefacts or both, as described above. Between June and July 2000, temperatures increased at all depths down to 10 m with a corresponding resistivity decrease throughout the major part of the survey area. This decrease continued until end of August 2000, where the resistivity distribution of September 1999 was reached.
Figure 4.24: (bottom) Ground-heat flux and (top) heat flux through the snow cover at the Schilthorn borehole during winter 1999/2000. The fluxes are calculated from the temperature gradients between the uppermost thermistors in the borehole and between the uppermost thermistor and the temperature at the snow surface, respectively (Mittaz et al., 2000). The temperature at the snow surface was obtained from pyrgeometer measurements at the energy balance station. Data were kindly made available by Catherine Mittaz (Geographical Institute, University of Zürich, Switzerland).

again. The difference between the September 1999 and the August 2000 measurement is less than most individual monthly differences (Fig. 4.23k). By comparing the pattern of resistivity changes in Fig. 4.23 with the geologic structures obtained from Fig. 4.22 and marked by the black lines and circles, no obvious dependence of freezing and thawing processes on geology can be seen. The seasonal changes in resistivity appear to be more connected to lateral variability in water content and local freezing anomalies.

4.5.5 Analysis of the data in terms of temperature and unfrozen water content

The tomograms of the last section have been shown that seasonal changes in the subsurface resistivity distribution can be analysed qualitatively using repeated DC resistivity tomography measurements. In this section the resistivity variations around the borehole are compared with borehole temperatures, and estimates of the unfrozen water content of the subsurface are obtained. This analysis uses the principles and relationships introduced in section 4.4.2.

Fig. 4.25 shows a scatter plot of all resistivity data points versus each corresponding temperature value from the borehole. As seen in the laboratory results, the
resistivities increase slowly for decreasing but still positive temperatures. Below
the freezing point, which may be estimated from Fig. 4.25 to be about \(-0.2^\circ C\),
the resistivities increase exponentially with cooling. However, as the data originate
from different depths, the rate of increase is not uniform for all data points. Three
branches can be identified and corresponding exponential functions can be fitted to
the data (shown by the black lines in Fig. 4.25). The uppermost branch includes
the two deepest model depths (below 6 m), the central branch consists of the data at
1.6 and 4.3 m depth and the lowermost branch includes the model depths at 0.5 and
2.8 m. The factor \( b \) (Eq. (4.23)) was estimated as 2.422 (upper), 0.617 (central) and
0.244 (lower), respectively. Two other curves were added for comparison (dotted
lines): the line fitting the laboratory results for the saturated \((b = 0.735)\) and the
unsaturated Schilthorn sample \((b = 0.273)\) shown in section 4.4.4. The similarity
between the two laboratory curves and the results from the two lower branches,
corresponding to the active layer, suggests that differences in the amount of water
saturation may be the reason for the existence of the two lower branches in Fig. 4.25.
The data points of the uppermost branch most probably correspond to firm bedrock
below the active layer (5–6 m) and cannot be related to the laboratory experiments.

Using Eq. (4.24) the evolution of the unfrozen water content for the different model
depths can now be calculated analogous to the laboratory experiment in section
4.4.4. The parameter \( b \) was chosen for each depth according to the exponential
behaviour of the corresponding resistivity data in Fig. 4.25. Figure 4.26 shows the
results for two different values for the factor \( n \). As discussed earlier, the unfrozen
water content starts to decrease at the end of October. The minimum is reached
in February \((S = 0.6)\) and subsequently later at greater depth (beginning of June
at 8.7 m depth). The minimal value of \( S \) is smallest at larger depths \((0.2–0.3 below
6 m for \( n = 2 \)) and largest at intermediate depths \((0.6–0.8 at 2–4 m for \( n = 2 \))\), but
depends on the choice of parameters \( b \) and \( n \). The larger \( n \), the smaller the variations
of \( S \). At larger depths the evolution of \( S \) is nearly sinusoidal, corresponding to the
seasonal variation of ground temperature.
Figure 4.25: Resistivity-temperature relationship for the resistivity data taken from the inversion results shown in Fig. 4.22 and the borehole temperature data at Schilthorn, Switzerland (Fig. 4.21). The data are in good agreement with Eq. (4.22) (black dashed line) for temperatures above the freezing point, and split into three branches for temperatures below the freezing point. Each branch shows an exponential increase of resistivity with decreasing temperature, but at different rates (factor $b$ in Eq. (4.23)). The exponential increase rates of the laboratory measurements with the sample material from Schilthorn (saturated and dry state) are shown for comparison (green and orange dashed lines).

Figure 4.26: Evolution of the unfrozen water content $S$ at different depths at the Schilthorn drill site for the period September 1999 to September 2000. The values were calculated from the resistivity and temperature data shown in Fig. 4.25 using Eq. (4.24) for two different values for the saturation exponent $n$ ($n = 2$: triangles, $n = 3$: squares).
4.5.6 Conclusion

Time-lapse resistivity measurements at a mountain permafrost site have been presented for the first time. A set of eleven DC resistivity tomography measurements were performed between September 1999 and September 2000 using a fixed electrode array at Schilthorn, Switzerland. The resulting tomograms were analysed in terms of seasonal resistivity changes and were compared with borehole temperature data.

Key results from this unique data set include:

1. Temporal resistivity changes can be accurately determined using a fixed electrode array, even in high Alpine environments.

2. Maximum resistivity changes were observed in autumn (September-October), before a permanent snow cover has been established, and in late spring (May-June), when the thawing snow cover and additional water from precipitation greatly decreased the resistivity values in the active layer. During the winter, the snow cover effectively decouples the ground from atmospheric influences. The heat flux through the snow cover was less than 1 W/m², estimated from energy balance measurements. Consequently, only a small resistivity increase was obtained during winter, which was due to the trapped, cold October temperatures. From December to May the freezing front moved gradually downward, reaching 6 m in mid-April. After the start of the melting season the resistivities decreased again until the previous September values are reached again at the end of August 2000.

3. Synthetic modelling studies showed the advantages of using anomaly plots showing the ratio of the results for consecutive resistivity measurements in order to filter systematic noise. Through this procedure the resistivity change due to a change in temperature can easily be distinguished from vertical and spatial resistivity variations due to geology.

4. Resistivity-temperature relationships between the resistivity values at the borehole location and borehole temperatures show good agreement with theory and with the results from the laboratory experiments shown in section 4.4. The increase of resistivity with decreasing temperature is small and linear for temperatures above the freezing point and exponential for temperatures below. A freezing point depression to $-0.2^\circ$C was noted, probably due to a significant mineral content.

5. From these resistivity-temperature curves the degree of water saturation at different depths can be estimated. A comparison of the field data with the laboratory results for the Schilthorn sample material, showed a good correspondence between the saturated laboratory sample and the ground conditions at around 1.5 m and 4 m depth, whereas the dry laboratory sample corresponds to depth ranges at 0–1 m and 2–3 m.

6. The calculated temporal evolution of the unfrozen water content show a strong decrease during the winter months in the active layer and a quasi-sinusoidal behaviour below.
Chapter 5

Electromagnetic induction methods

5.1 Introduction

Similar to DC resistivity techniques, electromagnetic (EM) induction techniques measure the electrical resistivity, or its reciprocal, the electrical conductivity (in Siemens/metre or usually milli-Siemens/metre, mS/m), but with the difference that no direct (galvanic) contact with the ground is needed. This is a great advantage in high mountain environments, as getting sufficient electrical current into the ground is one of the largest problems in DC resistivity surveys. Furthermore, DC resistivity surveys in winter time are usually impossible to conduct, as the snow cover acts as an electrical insulator. Electromagnetic induction methods employ a magnetic field to induce the electrical current in the subsurface. Therefore, there is no need for direct electrical contact between the transmitter and the ground.

Electromagnetic induction methods have been used for a long time in studies of lowland permafrost, mostly in the Arctic regions (e.g. Hoekstra and McNeill, 1973, Sartorelli and French, 1982, Rozenberg et al., 1985, Harada et al., 2000). Only recently, these methods have been applied in mountainous regions and especially in the European Alps. This is partly due to the fact that apparent resistivities of mountain permafrost bodies are generally much higher than in the Arctic (up to 1 MΩm, compared to 1–10 kΩm in the Arctic, see Fig. 3.2) and measured apparent conductivities (1/resistivity) are therefore at the resolution limit of most EM instruments. Nevertheless, in this section it will be shown that electromagnetic induction techniques are very suitable, especially for mapping of shallow permafrost occurrences as well as to determine the permafrost base at larger depths. Three different instruments were used: the single-frequency conductivity meter EM-31 (Geonics), the multi-frequency conductivity meter GEM-300 (GSSI) and the time-domain EM system PROTEM (Geonics).

Measurements were conducted at various field sites including Hiorthfjellet rock glacier and Janssonhaugen on Svalbard, Tarfala/Sweden, Juvasshøe/Norway, Schilthorn, Val Bever and Murtel rock glacier in Switzerland and Stelvio pass and Foscagno rock glacier in Italy. The results were compared with data from nearby boreholes and supplementary geophysical surveys.
5.2 Theory and instruments

Electromagnetic induction is the basis of a number of geophysical techniques. It is based on the principle that each current-carrying wire is surrounded by circular, concentric lines of magnetic field. If bent into a small loop, the wire produces a primary magnetic dipole field, \( H_p \). In electromagnetic induction this dipole field is varied, either by alternating the current (operating in the frequency-domain (FEM method)) or by terminating it (transient methods, operating in the time-domain (TEM)). This time-varying magnetic field induces very small eddy currents in the Earth. The eddy currents generate a secondary magnetic field, \( H_s \), which may be sensed at a receiver loop at the surface. The more conductive the subsurface, the larger are the eddy currents and the larger is the measured secondary field, which in turn allows the ground conductivity to be determined.

In general, the secondary magnetic field is a complicated function of the ground conductivity, as well as instrument specific parameters, such as the geometry of transmitter and receiver or the operating frequency. For layered Earth the secondary magnetic field \( H_s \) response due to a vertical magnetic dipole has the following form (Ward and Hohmann, 1988):

\[
H_s = \frac{m}{4\pi} \int_0^\infty K(\lambda) \lambda^2 J_0(\lambda d) d\lambda,
\]

where \( m \) is the dipole moment, \( \lambda \) the horizontal wave number, \( d \) the transmitter receiver distance, \( J_0 \) the Bessel function of order 0 and \( K \) contains the so-called kernel function defined by the electrical properties of the subsurface. Under certain constraints this expression simplifies substantially so that ground conductivity can be calculated readily from the respective instrument response.

5.2.1 Frequency-domain systems (EM-31, GEM-300)

*Ground conductivity meter* like the Geonics EM-31 and the GSSI GEM-300 operate in the frequency-domain at fixed frequencies and consist of two co-planar coils acting as transmitter and receiver, respectively. They are connected by a rigid frame at a fixed distance \( d \). Conductivity meter are commonly used for mapping spatial conductivity variations in the survey area. Both, the EM-31 and the GEM-300 are easy to handle and can be carried by one person. The processing of the data is straightforward and not time consuming. Both instruments are robust and therefore well suited on steep mountain slopes.

For both instruments the measured parameter is the ratio between the secondary and the primary magnetic field at the receiver coil. For a vertical dipole configuration this ratio is given by (McNeill, 1980)

\[
\frac{H_s}{H_p} = \frac{2}{(\gamma d)^2} \{9 - [9 + 9\gamma d + 4(\gamma d)^2 + (\gamma d)^3]e^{-\gamma d}\},
\]

92
where
\[ \gamma = \sqrt{i \omega \mu_0 \sigma} \]
\[ \omega = 2\pi f \]
\[ f = \text{frequency (Hz)} \]
\[ \mu_0 = \text{magnetic permeability of free space} \]
\[ i = \sqrt{-1}. \]

This rather complicated expression may be simplified under the constraint of the so-called operation at low induction numbers, which will be explained in the following.

**Induction number**

The induction number, \( B \), is defined as the ratio between the intercoil spacing \( d \) and the electrical skin depth \( \delta \), which is the distance in a homogeneous half-space that a propagating plane wave has travelled when its amplitude has been attenuated to \( 1/e \) of the amplitude at the surface. The skin depth is given by
\[
\delta = \sqrt{\frac{2}{\omega \mu_0 \sigma}} \quad \text{(5.3)}
\]
and therefore
\[
B = \frac{d}{\delta} = d \sqrt{\frac{\pi f \mu_0 \sigma}{}}. \quad \text{(5.4)}
\]

If \( B \) is much less than unity (i.e. \( \gamma d \ll 1 \)) it can be shown that Eq. (5.2) reduces to
\[
\frac{H_s}{H_p} \approx \frac{i B^2}{2} = \frac{i \omega \mu_0 \sigma d^2}{4} \quad \text{(5.5)}
\]
which is called the low induction number approximation (McNeill, 1980). By taking the real part, or quadrature component, of the measured \( H_s/H_p \) ratio, the apparent conductivity of the ground can be calculated by
\[
\sigma_a = \frac{2}{\pi f \mu_0 d^2} \left( \frac{H_s}{H_p} \right)_{\text{quadrature component}} \quad \text{(5.6)}
\]
(McNeill, 1980).

**Exploration depth and instrument response**

The exploration depth of most instruments is limited by two factors: the electrical skin depth and the transmitter-receiver distance. Depending on the ratio between both factors (= induction number \( B \), Eq. (5.4)) the exploration depth is limited by either factor or both. For small values of \( B \) the exploration depth is controlled mainly by the transmitter-receiver distance, whereas it is limited by the electrical skin depth for large values of \( B \). Fig. 5.1a shows the electrical skin depth as a function of
resistivity and frequency. Comparing the values for \( \delta \) with the transmitter-receiver distances of the EM-31 \((d = 3.7\, \text{m})\) and the GEM-300 \((d = 1.6\, \text{m})\) it is seen that \( \delta \approx d \) only for very high frequencies and low resistivities.

Following these considerations, the exploration depth of the EM-31 and GEM-300 in permafrost environments is mainly limited by the transmitter-receiver distance. For the EM-31, which has a fixed frequency of 9.8 kHz, the maximal exploration depth is given as 6 m (McNeill, 1980). As the transmitter-receiver distance is fixed, measurements of conductivity variations with depth can only be obtained by varying the instrument height above the ground or by using different instrument polarisations (see McNeill, 1980). The same is true for the GEM-300, which is very similar to the EM-31, except that up to 16 different frequencies can be measured simultaneously. In typical permafrost environments, this additional information content does not improve much, since the exploration depth is still limited by the fixed transmitter-receiver distance as long as the instrument operates at low induction numbers (Geonics, Technical Notes TN-30 and TN-31). Only for high frequencies and low resistivity values, the exploration depth becomes a function of the skin depth and therefore of frequency (see Fig. 5.1a). Furthermore, Maurer et al. (2000) showed that in order to conduct FEM soundings, the transmitter-receiver distance rather than the frequency should be varied.

Nevertheless, choosing the appropriate frequency may optimise the signal strength for a specific target. The target signal, that is the instrument response to the presence of a change in the conductivity of the subsurface, is not equally strong for all frequencies, because changing the frequency changes the instrument response (McNeill, 1990, Maurer et al., 2000). The instrument response is closely related to the induction number, with a maximum depending on the ratio between frequency and target resistivity and on the target dimension (Keller and Frischknecht, 1966, Telford et al., 1990). The instrument response is given as the imaginary part (quadrature component) of the vertical magnetic field strength at some distance from a vertical
magnetic dipole. For a homogeneous half space Earth model, the behavior of the field strength can be calculated as a function of frequency and ground resistivity (Ward and Hohmann, 1988, p.211). From Fig. 5.1b it is seen, that although the response is increasing linearly with frequency at low frequencies, it depends on frequency and ground resistivity in a non-trivial way, exhibiting local minima at higher frequencies. Consequently, the survey results measured at different frequencies cannot be compared to each other quantitatively, as the signal strength is different for each frequency. Furthermore, at low frequencies the instrument response decreases with increasing ground resistivity, whereas at higher frequencies the instrument response increases with ground resistivity. Multi-frequency instruments may be used to conduct surveys at different frequencies to determine the 'optimal' survey frequency, that is the frequency, which response is best for a given target resistivity and target depth.

Even though the instrument readings show absolute conductivity values, relative conductivity values are used throughout this study. This is due to the zero error, that is the shift of the zero level of the conductivity meter, which becomes significant for very low conductivity values. Consequently, all the results presented in this study must be interpreted as relative conductivity variations against a background conductivity value.

Effects of local snow heterogeneities and instrument height

Apart from the electrical properties of the shallow subsurface, the response of an electromagnetic induction instrument like the EM-31 or the GEM-300 may be influenced in a number of detrimental ways: changing surface characteristics (ice, wet
Figure 5.3: Sensitivity of the EM-31 on instrument height and snow cover depth. (a) Schematic plot and (b) data from Schilthorn/Switzerland. As air ($\sigma_a$) and dry snow ($\sigma_s$) have smaller conductivities as the ground ($\sigma_g$), the bulk conductivity ($\sigma_{\text{total}}$) increases with decreasing instrument height. This is only true for dry snow as an increasing water content in wet snow would lead to an increase in the bulk conductivity (see Fig. 5.2).

snow, water, metallic objects), instrument drift, instrument height above ground, electrical power lines or atmospheric lightning in the vicinity. The influence of metallic objects, power lines or lightning are usually very pronounced and cannot be mistaken for a target signal from the subsurface, even though they may prohibit data acquisition. The influence of variations in instrument height or a locally heterogeneous snow cover are much smaller and can easily be mistaken for an actual permafrost signal.

Figure 5.2 shows an example of an EM-31/GEM-300 comparison near Stelvio Pass. The results of both instruments are very similar with a dominant conductivity maximum observed between station 150 and 200. This relative maximum coincided with a wet snow patch on the surface of an otherwise snow-free survey line. As unfrozen water has a much higher conductivity than the underlying material, this wet snow patch masks any possible permafrost occurrences underneath as the conductivity signal of the latter is much smaller. If the survey is conducted completely on snow, care has to be taken in order to ensure that the snow cover is homogeneous with respect to water content and thickness, as these heterogeneities may be misinterpreted as permafrost or lithological variations. In case of wet snow conditions, ground conductivity surveys will give highly erroneous results.

The dependence of the conductivity response on the snow cover is coupled with the dependence on differences in the instrument height above the ground. The apparent conductivity response as measured with a ground conductivity meter is a function of all specific conductivities in a 6 m column beneath the instrument. This includes the air layer between the instrument and ground surface as well as the snow layer, if
Figure 5.4: Drift experiment EM-31 at Juvvasshoe/Norway. Measurement duration was 30 minutes. The instrument was set on the ground and conductivities were measured every 5 seconds.

Present (see Fig. 5.3a). Different air layer and snow cover thicknesses will change the measured apparent conductivities even in the case of unchanged subsurface conductivity. Figure 5.3b shows the dependence of apparent conductivity on instrument height and snow cover from a study on Schilthorn, where measurements were made at different instrument heights above the snow cover and after removing the snow.

Instrument drift

Most ground conductivity meters are designed to have a constant zero level with time to within about 1 mS/m (McNeill, 1990). The accuracy of the zero level may be different for each instrument and can depend on age, battery charge, temperature etc. All EM-31 surveys presented in this work were conducted with the same instrument. The accuracy of the zero level was tested in a drift experiment where the instrument was put on the ground and measurements were taken continuously for 30 minutes (Fig. 5.4). There is a marked increase in conductivity (0.3 mS/m) during the first 15 minutes followed by periods of smaller decrease and increase. For most geophysical surveys this accuracy is sufficient as the target signal is usually much larger than the drift amplitude. However, in permafrost studies the permafrost signal can be very small, so instrumental drift may cause severe problems. To ensure good quality data, care has to be taken when measuring over long time periods. Especially if the air temperature is changing rapidly repeated measurements at fix points (as in gravimetric surveys) may be conducted in order to estimate the amplitude of the drift error.

5.2.2 Time-domain systems (PROTEM)

Instruments operating in the time-domain are used to perform electromagnetic soundings in order to determine the resistivity variations with depth. This is anal-
ogous to Schlumberger soundings in DC resistivity surveys. In contrast to FEM
instruments, most time-domain electromagnetic (TEM) instruments (e.g. the Geon-
ics PROTEM) use a transient transmitter current, which, while still periodic, is a
modified symmetrical square wave. During the current-on time a static magnetic
field is established in the ground. After every second quarter-period the current is
abruptly reduced to zero for one quarter-period, whereupon it flows in the oppo-
site direction. This induces eddy currents to flow in the immediate vicinity of the
transmitter, which propagate downwards. The hereby induced secondary magnetic
field is measured by the receiver in the transmitter-off periods. The response of
the subsurface in terms of the decaying amplitude of the secondary magnetic field
is recorded as a function of time and therefore of depth, because later responses
originate at greater depths. The quantity measured is usually a voltage value that
is directly proportional to the decaying magnetic field. To transform the data into
apparent resistivities (or conductivities) an asymptotic solution is used for the time
span of the measurements (McNeill, 1994). This so-called late-stage or late-time
approximation is given as (e.g. Fitterman, 1989)

\[
\rho_a(t) = \frac{1}{(400\pi)^{1/3}} \cdot \left( \frac{I r^2 S_r}{V(t)} \right)^{3/2} \cdot \left( \frac{\mu_0}{t} \right)^{1/2},
\]

where

- \( I \) = transmitter current
- \( r \) = transmitter loop radius
- \( S_r \) = receiver coil area
- \( V(t) \) = voltage response induced in the receiver coil
- \( \mu_0 \) = magnetic permeability of free space
- \( t \) = time since current turnoff.

In TEM studies different measurement configurations are utilised. Central loop
soundings, where the receiver coil is located at the centre of a large transmitter
coil (usually a square of 40 m x 40 m or 100 m x 100 m) have been conducted on
permafrost by e.g. Rozenberg et al. (1985) and Todd and Dallimore (1998). Harada
et al. (2000) used an outside configuration, where the receiver was placed outside a
60 m x 60 m transmitter loop to reduce the noise level through primary field effects.
On mountain slopes or rock glaciers such large transmitter loops are often practicallly
impossible to use, due to the blocky and irregular terrain. Therefore, a flexible eight-
turn 5 m x 5 m transmitter loop was used in this work, which allowed for reasonably
easy handling and a still sufficient penetration depth.

The size of the transmitter loop influences not only the quality of the data, but
also the penetration depth. In addition, the penetration depth is mainly limited
by the upper layer resistivity: the lower the resistivity, the smaller the penetration
depth. According to the PROTEM-47D operation manual, the maximum investiga-
tion depth can be approximately calculated by:

\[
z = 8.94 l^{0.4} \rho^{0.25},
\]

where \( z \) is the depth in m, \( l \) is the loop side length in m and \( \rho \) is the upper layer resistivity in \( \Omega m \) (Geonics, 1994). For a typical active layer resistivity of 5 k\( \Omega \)m
and the configuration used in this work a theoretical penetration depth of 330 m is estimated. Although the time for one sounding is short (a few minutes) it takes rather long time to set up the system.

The system is essentially designed for surveying in conductive material and is generally considered less well suited in high-resistive areas (McNeill, 1990). Harada et al. (2000) applied the method to permafrost mapping in Alaska and obtained promising results for permafrost resistivities as low as 1 kΩm. Todd and Dallimore (1998) encountered even lower permafrost resistivities of less than 300 Ωm along a transect in the Mackenzie River Delta, Canada. One of the aims of this work was to evaluate the applicability of the method for mountain permafrost studies, where resistivities may reach values up to 100 kΩm to 1 MΩm. Different sites with the same morphologic structure (rock glaciers) were compared and the depth of the permafrost base was estimated for various PACE drill sites.

5.3 Results

5.3.1 Conductivity mapping with the EM-31 using a priori information

Procedure

As mentioned in section 5.1, one major difference between mountain and polar permafrost is the much higher resistivity values of the former. Resistivity values in the permafrost regions of the European Alps range typically between a few tens of kΩm (low ice content) and 1 MΩm (massive ice) (Vonder Mühll, 1993, Haeberli and Vonder Mühll, 1996). In the Arctic, typical values are much smaller, between 1 and a few kΩm (e.g. Osterkamp and Jurick, 1980, Timofeev et al., 1994), see Fig. 3.2. This is due to the much more conductive overburden and host sediments in Arctic tundra regions. For conductivity measurements in mountain regions the high-resistive uppermost layer causes problems, because the conductivity values are close to the accuracy limit of the EM-31 instrument (0.1 mS/m). This leads to substantial noise in the data, as does the inhomogeneous overburden and the topography. To overcome these problems a three step procedure is used:

1. Obtain a priori information at one representative permafrost location in the target area. This was done by two-dimensional resistivity tomography and refraction seismic tomography.

2. Repeat the survey line with the EM-31 and compare the data with the resistivity values from step 1. With this a permafrost signal for the EM-31 can be determined.

3. Map the whole target area with the EM-31 and plot the results as deviations from the measured mean value.
This procedure is illustrated by an example from Val Bever, Switzerland. The resistivity tomography measurements were done as described in chapter 4 using 41 electrodes with a 5 m electrode spacing.

Results

The investigation site in Val Bever is described in detail in section 2.7. It is situated on a 35° steep north-facing slope well below the tree line (1800 m asl). Data acquisition with the EM-31 was difficult due to the steep forest slope. The site was chosen because results from a permafrost model based on potential solar radiation and mean annual air temperature (PERMAMAP, Hoelzle and Haeberli, 1995) predicted limited permafrost occurrence.

Following the steps outlined in the section above, resistivity and refraction seismic surveys were performed in a forest clearing on the western boundary of the survey area, where permafrost was suspected (clearing I, Fig. 2.11). Fig. 5.5a shows the resistivity tomography model, which results from a 200 m survey line along the forest clearing (line A in Fig. 2.11). There are values up to 80 kΩm in the uppermost 15 m of the ground. A refraction seismic survey along the same line (shown in section 6.4) yielded P-wave velocities between 3000 and 3500 m/s for this layer, which are typical values for ice-rich permafrost (see Fig. 3.3). Thus, the high resistive areas are interpreted as permafrost lenses above unfrozen material.

The lower half of this survey line (station 50 – 150) was repeated with the EM-31. A measurement was taken each metre and a 5-point low-pass filter was applied to the raw data. Fig. 5.5c shows a comparison between the EM-31 results and the apparent conductivity values for the corresponding depth range calculated from the DC resistivity raw data (Fig. 5.5a). To account for uncertainties in the EM-31 results due to instrument drift, error bars of ±0.15 mS/m were added to the conductivity values. Nevertheless, both data sets show good qualitative agreement. As the absolute numbers differ substantially, relative conductivity deviations from the mean were used to compare the different data sets.

A second example from a profile perpendicular to the first survey line is shown in Fig. 5.5b (line B in Fig. 2.11). Here, the ground is even more heterogeneous and the conductivity values are changing on a scale smaller than the 5 m electrode spacing of the DC resistivity survey. Consequently, the EM-31 data (Fig. 5.5d) show more small-scale structures than the DC resistivity data and the agreement between the two methods is less obvious than in the first example. The overall pattern is similar, but the discrepancies may indicate that this line is located in the transition zone between permafrost and non-permafrost areas (see Fig. 5.5e).
Figure 5.5: Results for a permafrost mapping survey in Val Bever, Upper Engadin, Eastern Swiss Alps. (a) Model results for a 200 m resistivity section along and (b) across a forest clearing on the western boundary of the survey area (clearing I, line A and line B in Fig. 2.11). (c) Comparison of conductivity data measured with the EM-31 and determined from the DC resistivity section in (a). (d) Comparison of conductivity data measured with the EM-31 and determined from the DC resistivity section in (b). (e) Conductivity anomaly map of the whole survey area as marked in Fig. 2.11. Negative conductivity anomalies, calculated as deviations from the global mean, are shown in blue delineating possible permafrost occurrences.
Figure 5.6: EM-31 conductivity map of the area around the borehole at (top) Juvvasshoe/Norway and (bottom) Schilthorn/Switzerland. The location of the boreholes are marked with a cross. The highly conductive region between x-stations 15 and 30 at Schilthorn corresponds to a buried metallic sewage pipe.

Figure 5.7: GEM-300 conductivity map of the frontal part of rock glacier Foscagno/Italy (see Fig. 2.16) for a frequency of (a) 19975 Hz and (b) 325 Hz. The flow direction of the rock glacier is marked by the long arrow. Whereas the conductivity increases towards the rock glacier front in the 19975 Hz results, no trend is visible in the 325 Hz results. Line I mark the transition line between the resistive and the conductive part of the rock glacier tongue. The resistive anomalies II–IV are either due to air-filled cavities or ice patches. The small arrows mark the location of the survey lines.
Fig. 5.5e shows an anomaly plot for the EM-31 data of the whole survey area. The area is 300 m x 100 m wide, bounded by two forest clearings to the east and west and by a road to the north (c.f. Fig. 2.11). The area slopes to the north with a difference in altitude of 70 to 90 m. The general striped pattern in north-south direction is due to a strong undersampling of the EM-31 data in this direction. Due to the thick forest and the steep slope, survey lines could only be directed in east-west direction, as indicated by the arrows in Fig. 5.5e. The survey line shown in Fig. 5.5a is located on the right hand side of Fig. 5.5e. The permafrost areas, that is, areas with anomalously low conductivity values, can be clearly delineated from this contour map. These are located mainly in the upper part of the survey area, with a more patchy distribution further down.

To verify these results, additional resistivity tomography measurements were performed. Examples are given in Hauck and Vonder Mühll (1999) and Kneisel et al. (2000) and are shown in Appendix A.7.

5.3.2 Investigation of the spatial variability at the PACE drill sites with the EM-31

In the previous section, the applicability of the EM-31 for mapping shallow mountain permafrost occurrences was shown, even if the contrast between the conductivity of the unfrozen and frozen ground is small. This method was consequently applied around most of the PACE drill sites in order to evaluate the homogeneity of the borehole environment in terms of geology and active layer thickness.

Figure 5.3.1 shows examples from the PACE drill sites at Juvvasshöe and Schilthorn, respectively. At both sites a map of apparent conductivities of the uppermost 6 m was constructed from the results of an EM-31 survey. The conductivities around the Juvvasshöe borehole are rather homogeneous (top panel), whereas the conductivities around the Schilthorn borehole show relative conductivity differences of up to 2 mS/m (bottom panel). The 14 m deep borehole at Schilthorn is situated in a region of comparatively low conductivities, with increasing conductivities towards the south. This corresponds to different degrees of weathering of the bedrock around the drill site with fine grained material in the vicinity of the borehole and firm bedrock outcrops to the south. As the survey was conducted in June, when the active layer was still frozen underneath the snow cover, the lower conductivity values around the borehole are due to a larger ice content in the fine-grained material than in the bedrock outcrops to the south. In contrast to Schilthorn, the drill site at Juvvasshöe is situated on a plateau in very homogeneous lithological surroundings, as confirmed by the relatively uniform conductivity values (differences less than 0.5 mS/m).

5.3.3 Active layer mapping using the 'optimal' frequency with the GEM-300

Conductivity measurements with different frequencies were conducted on several field sites in the Alps using the GSSI GEM-300. As pointed out above, differences
in the instrument response using different frequencies can be large, and wrong or misleading results can be obtained, when choosing an inappropriate frequency. In Fig. 5.7, a conductivity map of the SE-part of the frontal area of rock glacier Foscagno, a few kilometres west of Stelvio Pass (see Fig. 2.16), is shown for two different frequencies (19975 Hz and 325 Hz). Again, the striped pattern is due to undersampling in the flow line direction of the rock glacier, as indicated by the small arrows. Whereas the high frequency (19975 Hz, Fig. 5.7a) shows a sharp conductivity increase towards the rock glacier tongue (transition line I) as well as three resistive anomalies (II-IV), no features or trends are visible for the low frequency (325 Hz, Fig. 5.7b). This difference is assumed to be due to the stronger response signal for high frequencies in the presence of very resistive background material (see Fig. 5.1b). The increase in conductivity towards the tongue is most likely due to an increase in the active layer thickness, as shown by BTS measurements (Guglielmin, 1997). The resistive anomalies are either due to air-filled cavities or ice anomalies.

The above example demonstrates that care has to be taken when choosing the frequency for an EM-conductivity survey on mountain permafrost. The optimal frequency should be chosen according to the largest response signal depending on coil separation and target resistivity (Keller and Frischknecht, 1966). Due to non-linear behaviour of the conductivity response with frequency and depth, a possible utilisation of the different frequencies for obtaining information about conductivity variations with depth could not be adressed in this work. Clearly more research is needed, in order to make a full use of multi-frequency instruments with fixed transmitter-receiver spacing in permafrost studies.

5.3.4 Time-domain electromagnetic soundings (PROTEM) to determine the depth of permafrost base

Measurements with the Geonics PROTEM were conducted at different permafrost sites in Svalbard, Italy and Switzerland. The measurements served as a preliminary investigation to evaluate the applicability of the method for mountain permafrost studies. The principle aim was to determine the depth of the permafrost base. Due to technical limitations of the sounding instruments, that is the transmitter current cannot be stopped instantaneously and the first time gate recorded must occur a certain time after the termination, no information from the topmost few metres (e.g. the active layer) can be obtained. Furthermore, TEM data cannot resolve the absolute resistivity values of a resistive middle layer, but is very sensitive to the thickness of such a layer (Maier et al., 1995). This characteristic makes TEM soundings an ideal method to determine the permafrost base, even though the resistivity of the permafrost layer must be determined by other methods. Borehole data were available to verify the PROTEM results.

Standard measurements were conducted with the offset sounding measurement configuration, i.e. a 5 m x 5 m 8-turn transmitter cable located 10 m apart from the receiver coil. At all sites, several soundings with different parameter settings (such as gain factors and repetition times) were performed. Transmitter currents ranged between 1 and 3 amperes and two transmitter-waveform base frequencies (285 and
Figure 5.8: PROTEM sounding results for (a) Janssonhaugen, Svalbard, (b) Hiorthfjellet rock glacier, Svalbard, (c) Murtèl rock glacier, Switzerland and (d) Foscagno rock glacier, Italy. Apparent resistivities are displayed as functions of time, which is equivalent to depth in TEM soundings. Later times correspond to greater depths.

75 Hz, respectively) were used. The best results were generally obtained using the largest gain factors and repetition times.

Fig. 5.8 shows a comparison between typical sounding curves from the PACE drill site at Janssonhaugen/Svalbard and from the three rock glaciers Hiorthfjellet/Svalbard, Murtèl/Switzerland and Foscagno/Italy, respectively. Apparent resistivity values are shown as a function of time, which is equivalent to depth in TEM soundings. Except for the very early time gates, the two sounding curves from Svalbard (Fig. 5.8a and b) are very similar, showing decreasing apparent resistivities with slightly higher values at Janssonhaugen. Compared to the other two sounding curves from the Alpine rock glaciers (Fig. 5.8c and d) the apparent resistivity values are one order of magnitude higher. This is due to the markedly colder permafrost temperatures at Svalbard with very low unfrozen water contents and therefore much higher resistivities. Additionally, the permafrost base is much deeper in Svalbard (estimated as 200–450 m, see Isaksen et al., 2000b, and Liestøl, 1976) than in the Alps. Consequently, the bedrock beneath Hiorthfjellet rock glacier is frozen as well and no resistivity contrast is seen between rock glacier material and underlying bedrock.

In contrast to Hiorthfjellet rock glacier, the two Alpine rock glacier sites (Murtèl and Foscagno) show typical permafrost curves similar to data from vertical electrical soundings (Vonder Mühll, 1993, Guglielmin et al., 1994): low resistivities at early times (corresponding to shallow depth), maximal resistivities in the middle and de-
creasing resistivities at later times (corresponding to greater depths). Both sounding curves are quite similar, except for slightly larger maximal values for Murtèl and a higher noise level at Foscagno.

As seen from Eq. (5.8) the investigation depth depends on the upper layer resistivity for a given transmitter-receiver geometry. The measurements in Svalbard were made in April when the surface was frozen and upper layer resistivities were high. Measurements at the rock glaciers Murtèl and Foscagno were performed at the end of June shortly after the thawing of the snow cover in rather wet surface conditions. This led to rather low upper layer resistivities as can be seen from the data for the early time gates in Fig. 5.8. From Eq. (5.8) the theoretical investigation depth was estimated to be 300–400 m for the Svalbard sites and 100–150 m for the Alpine sites. However, due to the extremely resistive subsurface material the induced currents are very low throughout the whole investigation area. Consequently, the voltages observed at the receiver coil are small, and signal amplitudes from larger depths lie within the noise level of the instrument, thus reducing the effective investigation depth.

Layered Earth inversion was carried out using the forward-inverse modelling package TEMIX (Interpex Ltd, 1993). The inverse model is obtained through iterative adjustment of the model parameters in the least-square sense using a ridge regression algorithm (Inman, 1975). The starting model is supplied by the user. Alternatively, a smooth model inversion (with up to 19 model layers) can be chosen, where a homogeneous Earth is used as starting model.

The inversion results for the sounding at Janssonhaugen/Svalbard (Fig. 5.9) indicate a three-layer case: variable resistivities between 1–20 kΩm in the uppermost 10 m, followed by the main permafrost body with resistivities around 100 kΩm down to a depth of ca. 250 m. The range of these values, determined from equivalence models, is comparatively small yielding 30–300 kΩm and a depth of 230–270 m. Note, that due to the small signal amplitudes there is some doubt about the reliability of the model results at larger depths. Consequently, the third layer is poorly constrained having a range of specific resistivities over 4 order of magnitudes with a best-fit value of 10 kΩm. Fig. 5.9b gives the model results for the smooth inversion (using 19 layers) scheme. The results are very similar to the three-layer model shown in Fig. 5.9a with a resistivity maximum between 50 and 150 m depth and decreasing resistivities below. Due to the smooth nature of this inversion technique it is difficult to determine the depth of the permafrost base exactly.

This result can be compared to the temperature data from the PACE borehole (Isaksen et al., 2000a, see Fig. 2.3), which showed very homogeneous ground conditions from beneath the uppermost 10–20 m down to the bottom of the borehole at 100 m. From linear extrapolation of the thermal gradient the permafrost thickness was estimated to around 220 m, which is in good agreement with the result from Fig. 5.9a. However, the large range of resistivities calculated for the third layer in Fig. 5.9a and the smooth model curve in Fig. 5.9b indicate that in a case of small resistivity contrasts between the frozen and the unfrozen layer it remains very difficult to estimate the exact depth of the permafrost base.

Fig. 5.10 shows the model results for sounding curves obtained for the two Alpine
Figure 5.9: Inversion results for the TEM sounding at Janssonhaugen/Svalbard using (a) layered Earth modelling and (b) smooth model inversion. The left panels show measured (squares) and modelled (line) apparent resistivity values, the right panels show the calculated resistivity model. The red solid line marks the layered Earth model with the best data fit, the dashed lines mark equivalent models which fit the data similar well. They indicate the allowable range of the model parameters, specific resistivity and layer depth, respectively. Equivalence models are only available for layered Earth inversions (a). The permafrost thickness as estimated through linear extrapolation from temperature data (Isaksen et al., 2000a) is indicated as well (ISA 2000).
rock glaciers, Murtèl/Switzerland and Foscagno/Italy. Here, the frozen rock glaciers rest upon unfrozen or marginally frozen bedrock, leading to large resistivity contrasts between the high-resistive ice body and the comparatively low-resistive bedrock. As mentioned above, the resistivity of the ice-rich permafrost layer cannot be determined through the TEM data set, because the method is insensitive to the exact value of high-resistive middle layers. Consequently, the high-resistive middle-layer, representing the whole frozen rock glacier body, was set to 700 kΩm in the initial model in both cases, even though the ice content at Murtèl rock glacier is supposed to be larger than at Foscagno rock glacier. Furthermore, it was not distinguished be-
tween permafrost layers with high or low ice content. As can be seen from Figs. 5.10a and b the best fit models predict the depth of the permafrost base at 50 m for Murtèl rock glacier and 30 m for Foscagno rock glacier. The ranges are estimated to be 30–70 m for Murtèl and 20–35 m for Foscagno, which is rather small considering the high amount of geological noise in the data (especially at Foscagno). These results are in a very good agreement with the borehole data, which showed the permafrost depth at 52 m at Murtèl (Vonder Mühll, 1993, see Fig. 2.13) and 23.5 m at Foscagno (Guglielmin et al., 2001).

These results show the great potential of TEM soundings for determining the depth of the permafrost base at mountain permafrost sites. This is especially true for sites exhibiting a large contrast between the resistivity of the frozen and the unfrozen layer, as at most Alpine rock glaciers sites or for ice-cored moraines.

### 5.4 Conclusions

Three different electromagnetic instruments were evaluated for application in mountain permafrost terrain. With respect to measured electrical resistivity values, mountain permafrost differs from lowland Arctic sites, in that the former exhibit much larger resistivity values of up to 1–2 MΩm. The corresponding conductivities are therefore close to the resolution limit of most electromagnetic instruments. Surveys with three different electromagnetic instruments were conducted: conductivity mapping with the single-frequency Geonics EM-31 and with the multi-frequency GSSI GEM-300 and conductivity soundings with the time-domain Geonics PROTEM. Data from nearby boreholes and complementary geophysical surveys were used to verify the results.

Key results include:

1. Mapping the distribution of mountain permafrost with an EM-31 conductivity meter has been proven to be a reliable, fast, high resolution method for heterogeneous alpine terrain. Due to the small conductivity values encountered, results are mostly qualitative and a priori information from quantitative methods like two-dimensional DC resistivity tomography has to be used to relate the data to permafrost occurrences.
2. In the presented example of a steep, forested slope in Val Bever (1800 m asl) permafrost could be detected in the upper part of the survey area, in forest clearings and on isolated patches further down. Conductivity anomaly plots are well suited to locate the distribution of permafrost lenses on a scale of a few square-metres.
3. Sensitivity studies showed a strong (undesirable) sensitivity on the snow cover, especially for wet snow conditions. Furthermore, a significant instrument drift was found, which might be a problem in some permafrost applications due to a low signal-to-noise ratio. To avoid poor data quality long survey durations should be avoided.
4. Measurements with the multi-frequency conductivity meter GEM-300 showed a strong dependence of the instrument response on the induction number which depends on frequency, ground resistivity and transmitter-receiver spacing. For high-resistive materials like rock glaciers with a high ice content, high frequencies (19975 Hz with the GEM-300 or the single frequency 9800 Hz for the EM-31) are appropriate. It has still to be evaluated whether additional information can be gained from multi-frequency surveys using the GEM-300 other than choosing the optimal frequency for a specific target. Comparisons between the EM-31 and the GEM-300 show good qualitative agreement.

5. Results from the electromagnetic soundings with the PROTEM showed a great potential for determining the depth of the permafrost base at mountain permafrost sites. This is especially true for sites with a large contrast between the resistivity of the frozen and the unfrozen layer such as at most Alpine rock glaciers sites. Layered Earth inversion as well as smooth inversions were performed leading to resistivity models in the range of 10-200 m, depending on the transmitter loop dimensions and the upper layer resistivity.

6. Results from field studies showed a permafrost thickness of more than 200 m (best fit 250 m) at Janssonhaugen/Svalbard, 30–70 m (best fit 50 m) at rock glacier Murtel/Switzerland and 20–35 m (best fit 30 m) at rock glacier Foscagno/Italy. These estimates are in very good agreement with the temperature data from nearby boreholes, even though the noise level in the PROTEM data is rather high.

The results have shown that electromagnetic induction methods have a strong potential for application on mountain permafrost. However, each method has its own special field of application. The conductivity meters EM-31 and GEM-300 can only be used for mapping the shallow subsurface. The PROTEM will only give reliable results below a depth of 5-10 m, but may reach depths of more than 100 m. Future studies should include joint inversions of DC resistivity sounding data and TEM sounding data (Sandberg, 1993, Maier et al., 1995).
Chapter 6

Refraction seismics

6.1 Introduction

Seismic methods are the standard geophysical techniques with many applications, especially in mining and oil prospecting. Hereby, seismic reflection is by far the most commonly used method of all geophysical techniques. Most surveys are aimed at defining structures down to depths of thousands of metres using hundreds or even thousands of detectors. In contrast to seismic reflection, seismic refraction techniques are commonly used in near-surface applications down to depths of 100 m, mostly for environmental and engineering purposes. In permafrost applications, seismic reflection plays only a minor role due to its limitations in resolution at shallow depths. On the other hand, refraction seismic has a long tradition in permafrost studies (e.g. Timur, 1968, Röthlisberger, 1972, McGinnis et al., 1973, Pandit and King, 1978, King and Garg, 1980, Pearson et al., 1983, Zimmerman and King, 1986, King et al., 1988, King et al., 1992, Vonder Mühll and Schmid, 1993, Leclaire et al., 1994, Wagner, 1996, Carcione and Seriani, 1998, Carcione and Tinivella, 1998). Hereby the goal is to determine the P-wave velocity, which can be used as a complementary indicator to resistivity for the presence of frozen material. The method is especially useful to determine the active layer thickness, as the contrast for the P-wave velocity between active layer (400–1500 ms\(^{-1}\)) and permafrost body (2000–4000 ms\(^{-1}\)) is usually large.

In permafrost studies refraction seismic interpretation techniques were based largely on simple plane-layer models, commonly restricted to two or three layers. As pointed out in section 3.3, this approach may be of limited use for very heterogeneous ground conditions in Alpine environments. As seen for the DC resistivity technique, tomographic inversion schemes may be used for reliable 2D interpretation. Musil et al. (2001) successfully characterised the interior of a rock glacier using high-resolution seismic refraction data (120 geophones with 2.5 m spacing and shot points every 5 m) and a tomographic inversion scheme. The logistic effort and the costs of a survey of this scale are large, with more than 1 week of work with a field crew of 6 persons, explosives for a large number of shots and helicopter transport of several hundreds of kilograms of equipment. This is usually beyond the scope of standard fieldwork in mountain permafrost environments. Hence, the aim of this study was
to apply the principles of 2-D refraction tomography on a smaller scale. A set of 24 geophones, up to 12 shot points and a sledgehammer as source were used on several field sites to investigate the usefulness of the method for different permafrost environments. In this chapter the refraction seismic tomography technique is introduced using a synthetic data set and a field example from Val Bever. Further examples are presented in chapter 8.

6.2 Theory and tomographic inversion

6.2.1 Elastic waves

Seismic waves transmit energy by the vibration of rock particles and can be regarded as elastic in regions not too close to the seismic source. When a seismic source is triggered, several types of elastic waves are generated. The fastest wave, known as the primary wave, pressure wave or P-wave, is a longitudinal wave, where the rock particles oscillate backwards and forwards in the direction of energy transport (see section 3.2.2). Rock particles vibrating at right angles to the direction of energy flow create an shear wave (or secondary (S-) wave) with considerably lower velocity. In addition to these so-called body waves, surface waves can be generated. All these waves may carry a considerable proportion of the energy of the seismic source. However, in refraction surveys only the primary (P-) waves are analysed. As the P-wave is the fastest seismic wave, the first arrival time of the seismic waves at each receiver (= geophone, see Fig. 3.4) is used to determine the travelttime between source and receiver and, hence, to calculate the P-wave velocities of the subsurface.

The P-wave velocity depends on the density and the elastic constants of the material and is given by

\[ V_p = \sqrt{\frac{3K + 4\mu}{3\rho}}, \]  

(6.1)

where \( V_p \) is the P-wave velocity, \( \rho \) is the density, \( K \) is the bulk modulus and \( \mu \) the shear modulus (e.g. Telford et al., 1990).

6.2.2 Critical refraction and head wave

When a seismic wave hits an interface between different subsurface layers, one part of the energy is reflected at the angle of incidence, \( i \) (see Fig. 6.1a). The remainder is refracted at an angle given by Snell’s law

\[ \frac{V_1}{V_2} = \frac{\sin i}{\sin r}, \]  

(6.2)

where \( V_1 \) and \( V_2 \) are the velocities of the two layers and \( i \) and \( r \) are the angles of incidence and refraction, respectively (see Fig. 6.1b). If \( V_2 \) is greater than \( V_1 \) and if \( \sin i = \frac{V_1}{V_2} \) (the so-called critical angle), the refracted ray will travel parallel to the
interface at velocity $V_2$ ($r = 90^\circ$, Fig. 6.1b). Some of the wave energy returns to the surface as a head wave, leaving the interface at the critical angle $i_c$. As the refracted waves are travelling along the interface with velocity $V_2 > V_1$, they will eventually overtake the slower direct waves, despite the longer travel path.

### 6.2.3 2-D refraction tomography

Tomographic inversion schemes have been developed for different purposes, such as near-surface corrections for deep reflection data (e.g. Zhu et al., 1992), interpretation of crustal-scale seismic refraction data (White, 1989, Zelt and Smith, 1992) or for applications in the shallow subsurface, like delineating the geometry of landfills (Lanz et al., 1998). The tomographic inversion scheme used in this study was developed by Lanz et al. (1998) at the Geophysical Institute, ETH Zürich. It is based on the principles and general equations introduced in chapter 3. The derivation of the model equations described in the following are taken from Lanz et al. (1998).

#### Model equations

The time $t$ of a seismic wave traveling along a raypath $S$ through a 2-D medium can be written as

$$t = \int_S u(r(x,z)) \, dr,$$

where $u(r)$ is the slowness field (reciprocal of the velocity field) and $r(x,z)$ is the position vector. As seen in Fig. 3.4 this may be approximated with $m$ equidimensional cells or model blocks, each having a constant slowness $u_k(k = 1 \ldots m)$. Each recorded traveltime $t_i$, corresponding to one pair of source and receiver, can be written as
\[ t_i = \sum_{k=1}^{m} l_{ik} \hat{u}_k = L_i \hat{u}, \quad (6.4) \]

where \( l_{ik} \) denotes the portion of the \( i \)th raypath in the \( k \)th cell. Note, that Eq. (6.4) is of the form of Eq. (3.2), with the observation data \( t_i (i = 1 \ldots n) \), the model parameters \( \hat{u} \) and the data kernel \( L_i \). The forward problem consists of calculating the traveltimes for propagating wavefronts through heterogeneous media and determining the raypaths necessary for constructing \( L \). For solving the forward problem a fast finite-difference eikonal solver is employed (after Schneider et al., 1992).

**Solution of the inversion problem**

Even though the relationship between the known traveltimes \( t_i \) and the unknown slowness field \( \hat{u} \) is linear, the corresponding inversion problem is nonlinear, because the values of \( l_{ik} \) depend on \( \hat{u} \). Furthermore, the problem is mixed-determined, as the raypaths will not be evenly distributed throughout the model cells (chapter 3). Some cells will not be covered by a raypath at all and the corresponding slowness \( \hat{u}_k \) cannot be determined without the implementation of additional constraints. These additional constraints are used to

1. minimise the amount of deviation from the input model \( \hat{u}_0 \) (= damping, Marquardt, 1970) and
2. impose constraints on the model smoothness (Constable et al., 1987),

and can be combined as

\[
\begin{pmatrix}
\lambda \beta \hat{u}_0 \\
0
\end{pmatrix} = 
\begin{pmatrix}
\lambda \beta I \\
\lambda(1 - \beta)S
\end{pmatrix} \hat{u},
\]

where \( \hat{u}_0 \) is the initial model, \( I \) is the identity matrix and \( S \) denotes a Laplacian smoothing matrix (see Eq. (3.10)). The parameter \( \lambda \) controls the amount of regularisation applied, and \( \beta \) \((0 < \beta < 1)\) determines the proportional weights of damping and smoothing (Lanz et al., 1998).

As shown in section 3.4.3, Eq. (6.5) can be written as \( h = Du \) and may be combined with (6.4) to give

\[
\begin{pmatrix}
t \\
h
\end{pmatrix} = 
\begin{pmatrix}
L \\
D
\end{pmatrix} \hat{u}.
\]

This can be expressed in a more compact form as

\[
d = Gm.
\]

which is again of the form of the general inverse problem Eq. (3.2) and must be solved iteratively, due to its nonlinearity (see section 3.4). The solution of the inverse problem is then obtained by iteratively minimising the least-squares misfit between the observed and modelled traveltime data analogous to the procedure for the inversion of the DC resistivity data shown in section 4.3.2 (see Eq. (4.15)).
6.3 Synthetic modelling to optimise data acquisition and data inversion

In order to evaluate the applicability of the refraction tomography scheme to mountain permafrost studies, a series of synthetic modelling studies were conducted. Firstly, the question is addressed, if isolated permafrost occurrences can be delineated with a comparatively small number of receivers and source points. In addition, the influence of 2-D inversion parameters, like the choice of the initial model and the amount of regularisation, on the inversion results was determined. A synthetic data set was generated from an idealised permafrost model consisting of isolated patches of ice occurrences in an unfrozen host material. A gradient model with a P-wave velocity $V_p = 500\text{ms}^{-1}$ at the surface, which increases by $100\text{ms}^{-1}$ per metre to a maximal value of $4500\text{ms}^{-1}$ was used (Fig. 6.2). Two velocity anomalies ($V_p = 3500\text{ms}^{-1}$) were inserted representing the permafrost occurrences.

6.3.1 Determining the minimal number of sources and receivers

In refraction seismic surveys, the quality of the tomographic inversion results depends critically on the number of traveltime data available. The smaller the data set the more the inversion result is affected by noise, as each traveltime value gets a correspondingly larger weight. Furthermore, the spacing between the receivers determines the size of objects, which can be resolved during inversion.

To determine the minimal number of sources and receivers required for correctly delineating the two high-velocity anomalies in Fig. 6.2a, five experiments with different numbers of sources and receivers were conducted. Synthetic traveltimes were calculated by using the same fast finite-difference eikonal solver as in Lanz et al. (1998). Analogous to the procedure applied for generating synthetic DC resistivity data, 5% Gaussian noise was added to the traveltimes to simulate field conditions. The traveltimes were inverted using the inversion code by Lanz et al. (1998). The initial velocity model was chosen to be the "true" gradient model of the synthetic model.

Figure 6.2b–f shows the inversion results after 20 iterations for the five different experiments. The high-resolution experiment (f) with 120 geophones and source points is shown as reference model. In all four low-resolution experiments both anomalies were detected. However, in the experiment with the lowest resolution (Fig. 6.2b with 58 traveltimes) the shape and the lateral and vertical extent of the anomalies can not be determined correctly. This is slightly improved for the two medium-resolution experiments (Fig. 6.2c and d with 122 and 115 traveltimes, respectively). The lateral extent of the anomalies are smaller in experiment (c) than in (d), due to the smaller number of geophones, but are more pronounced, due to the higher number of traveltime data. In the experiment in Fig. 6.2e (276 traveltime data) the shape and the lateral and vertical extent of the anomalies are delineated quite well. The maximal P-wave velocities of the anomalies are too high in all experiments (up to $5000\text{ms}^{-1}$ instead of $3500\text{ms}^{-1}$). In the high-resolution
experiment the permafrost anomalies are modelled well, especially concerning lateral extent and upper boundary. The lower boundary of the anomalies is poorly resolved, as seen as well in all other cases. This is due to the decreasing ray density with depth and the concentration of the rays on the upper boundary of the high-velocity anomalies (see e.g. Fig. 6.3).

Comparing the results for experiments (e) and (f) it can be stated that enlarging the number of source points does not lead to a proportional increase in the quality of the inversion results. A quantitative measure of the quality of the inversion results is the model misfit, that is the difference between the modelled velocity values and the true model values of Fig. 6.2a. This should not be confused with the data misfit, which is the difference between modelled and observed data values. The model misfit for the different source-receiver geometries is given as RMS error in Table 6.1. Except for experiment (c), which has the largest RMS error due to an additional high-velocity anomaly at 60 m horizontal distance, the RMS error decreases with increasing number of traveltimes. In contrast to the low-resolution experiments, the difference in RMS is almost negligible between (e) and (f). This confirms that increasing the number of traveltime data above a certain level does not necessarily result in higher resolution in the results.

Based on these experiments all field measurements were conducted with 24 geophones and with as many source points as possible, which was often constrained by surface characteristics and time.

### 6.3.2 Choice of the initial velocity model

In the above examples, the "true" velocity gradient was chosen as the initial model. In real field experiments the background velocity gradient is obviously not known. Ideally, the initial velocity model should reflect the gross structure of the survey area. The choice of the initial model strongly influences the ray paths of the seismic waves, along which the model equations are evaluated (see section 3.5). Care has to be taken to ensure that raypaths through the initial model are sampling the entire depth range of interest. Consequently, velocities should increase monotonically with depth (Lanz et al., 1998). The dependence of the tomographic inversion results on the initial model is shown in Fig. 6.3. Three different gradient models were chosen (left column in Fig. 6.3): a small velocity gradient (10 ms⁻¹ per metre), the "true" velocity gradient (100 ms⁻¹ per metre) and a large velocity gradient (200 ms⁻¹ per metre). In all cases 24 geophones and 12 shot points were modelled and inversion results after 20 iterations are shown.

The results for the small gradient initial model show two clearly delineated anomalies, but without their vertical extent being correctly resolved. In comparison to
Figure 6.2: (a) Synthetic permafrost model and inversion results for five different source and receiver geometries. (b) 12 geophones with 10 m spacing and 5 source points, (c) 12 geophones (10 m spacing) and 12 source points, (d) 24 geophones (5 m spacing) and 5 source points, (e) 24 geophones (5 m spacing) and 12 source points and (f) 120 geophones and 120 source points. The high-resolution experiment (f) is shown for comparison. Circles and crosses denote source and receiver points, respectively.
Figure 6.3: Inversion results for the synthetic data set and three different initial velocity models. Circles and crosses denote source and receiver points, respectively. The residual RMS error of the final iteration is shown above the figures in the right column.
the model velocities in Fig. 6.2a, all velocities below 10 m model depth are much too small, indicating that they did not change much during iterations. This is confirmed by considering the calculated ray paths (right column). The rays, simulating the wave propagation, do not reach depths below 10 m, indicating that model cells at greater depths could not be perturbed substantially during the inversion. In contrast to that, ray coverage in the other two tomograms is extending down to 20 m, enabling the inversion algorithm to perturb the initial velocities. Due to the zones of anomalous velocity the ray coverage of the final iteration is not distributed uniformly. The absence of rays passing through certain parts of the final models does not necessarily mean that the low velocities in these regions are model artefacts. As the velocities did change in comparison to the initial model, rays must have passed through these regions during earlier iterations. However, strictly speaking, the reliability of the model results in these regions is diminished.

The results for the "true" and the large-gradient velocity model do not differ much, except for a slightly larger penetration depth of the latter. The root-mean-square error (here given as the RMS residuals between observed and calculated traveltimes) of both inversion results is the same (1.7 ms), but much smaller than for the small gradient model (3.1 ms). These results show that care has to be taken in choosing the appropriate initial model. If a priori information on the subsurface velocities is poor, the results suggest to overestimate rather than underestimate the initial velocity gradient to ensure sufficient ray coverage. This confirms results found by Lanz et al. (1998) in similar tests with synthetic data.

### 6.3.3 Regularisation parameter

Further important inversion parameter are the regularisation parameter, $\lambda$ and $\beta$ (Eq. (6.5)). The parameter $\lambda$ determines the amount of smoothing and damping applied during the inversion. A value of $\lambda = 1$ corresponds to equal weights for regularisation and data during the inversion process. For small values of $\lambda$ the underdetermined components of the inversion problem may become dominant, such that no unique solution can be found (see chapter 3). This can be especially problematic where the data coverage is low and underdetermined model parameters are numerous, as in many examples of this study. On the other hand, large values for $\lambda$ may lead to over-smoothed and/or over-damped models, the latter prohibiting the inversion algorithm to perturb the initial velocities. The parameter $\beta$ determines the proportional weight between smoothing and damping. Except for extreme cases, where a particularly smooth or rough velocity distribution was expected, this parameter was set to 0.5, forcing smoothing and damping constraints to represent 50% of the regularisation each.

Inverting the data for different values of $\lambda$ and plotting the resulting RMS errors (after 20 iterations) against $\lambda$ the range of regularisation factors leading to small RMS errors can be obtained (e.g. Herwanger et al., 2000). This is shown in Fig. 6.4 for the synthetic data set using again the true velocity gradient as initial model. For very large and very small values of $\lambda$, large RMS errors are observed. Appropriate $\lambda$ values can be chosen from within the broad minimum of the RMS data differ-
Figure 6.4: RMS difference versus regularisation parameter $\lambda$ for different inversions of the synthetic data set. A value of $\lambda = 2$ corresponds to smoothing and damping constraints with twice the weight of the data during the inversion process.

ences. The absolute minimum is found for $\lambda = 0.5$, corresponding to smoothing and damping constraints given half the weight of the data.

The influence of the amount of regularisation on the resulting tomograms is shown in Fig. 6.5 for $\lambda$ values of 0.1, 1 and 10, respectively, together with the corresponding ray path diagrams. For $\lambda = 0.1$ the small-scale structures are dominant in the tomogram (Fig. 6.5a). The high-velocity anomalies are difficult to delineate due to the enhanced representation of the artificially added noise. When using too much regularisation, all variations from the background geologic model are damped and/or smoothed out (Fig. 6.5e). This is the case for $\lambda = 10$, where an almost lateral homogeneous velocity model is obtained during inversion. This discrepancy would not become apparent by considering only the residual RMS error (Fig. 6.4).

### 6.3.4 Convergence

Figure 6.6 shows the evolution of the RMS error for subsequent iterations. The inversion process converges already after 5 iterations. After that, only minor reductions in RMS errors are observed. This is mainly because the initial velocity model for the synthetic data set was set very close to the 'true' velocity distribution. For field data sets, convergence is usually reached after 10–20 iterations. However, even though the initial RMS error is often much higher for field data sets than for the synthetic data set shown in Fig. 6.6, the final RMS errors do not differ much (see examples below).

A measure of the improvements gained during tomographic inversion is provided by plotting all observed and predicted traveltimes for the initial and the final model (blue and red circles in Fig. 6.7, respectively). The predicted traveltimes from the initial velocity model do not explain the broad scatter of the observed data (Fig. 6.7a), but those computed from the final model show much better agreement (Fig. 6.7b).
Figure 6.5: Final tomograms (after 20 iterations) and corresponding ray path diagrams derived from the synthetic data set for three different values of the regularisation parameter $\lambda$ (0.1, 1 and 10, respectively). Models with greater structural relief result from inverting the data with low values of $\lambda$ (a). Relatively smooth models are obtained when using comparatively large $\lambda$ values (e). For most field data sets shown in this study, minimal RMS errors are obtained when the overall regularisation was given the same weight as the data during the inversion ($\lambda = 1$, (c)).
Figure 6.6: Evolution of the RMS error (in ms) for subsequent iterations for the synthetic data set.

Plotting the residual differences against distance from the survey midpoint is used to detect systematic misfits in the final model (Fig. 6.8). In the initial model a general overestimation of the model velocities lead to positive residuals, whereas the residuals are distributed evenly around zero in the final model.

An additional measure of the improvements gained during inversion is provided by the spatial distribution of the traveltime residuals (Lanz et al., 1998). In Fig. 6.9 the traveltime residuals (predicted—observed) have been plotted in the form of pseudo-sections, in which the midpoints of the source-receiver pairs are displayed along the horizontal axis and the source-receiver distances are shown along the vertical axis. Since the distance between source and receiver is a rough measure of depth sensitivity, pseudo-sections yield crude information on the spatial distribution of residual errors. In theory, the traveltime residuals for the input 1D model approximately image the lateral locations of the velocity anomalies (Lanz et al., 1998), which is difficult to observe in Fig. 6.9a due to the sparse data set. However, the traveltime residuals for the final model are considerably smaller with no significant spatial correlations (Fig. 6.9b).

6.4 Field example — Val Bever

In this section a field example from Val Bever/Switzerland is presented. The aim of the survey was to determine, if permafrost is present at this low-altitude suspected permafrost site. Radiation-based modelling and BTS measurements suggested the presence of permafrost along the north-faced valley slopes (see section 2.7). DC resistivity tomography surveys showed isolated resistivity anomalies of up to 80 kΩm in the uppermost 10–15 m in a host material with resistivity values of less than 5 kΩm (c.f. section 5.3.1). As these values may originate from ice lenses as well as air-filled cavities near buried boulders, e.g. from a former rock fall or debris flow from the rock wall above the valley slope, seismic measurements were required to unambiguously interpret the anomalies.
Figure 6.7: Comparison between observed (blue) and predicted (red) traveltimes for (a) the initial and (b) the final model for the synthetic data set. Whereas the discrepancy between observed and predicted traveltimes is large in the initial model, it is reduced substantially after 20 iterations.

Figure 6.8: Histogram of residual traveltimes for (a) the initial and (b) the final model for the synthetic data set. Whereas the model velocities are generally overestimated in the initial model leading to positive residuals, the residuals are distributed evenly around zero after 20 iterations.

Figure 6.9: Residual (predicted–observed) traveltime plots. (a) Initial and (b) final residuals (after 20 iterations).
6.4.1 Data acquisition

The field site at Val Bever is described in detail in sections 2.7 and 5.3.1. The survey line was located in the western clearing at the same location as the DC resistivity profile (line A in Fig. 2.11).

A Geometrics EG&S seismograph system with 12 channels was used for data acquisition. The seismograph is relatively light-weight and can be carried by a single person. A seismic survey as presented here can be conducted by two persons. The time needed for the completion of one survey line depends nearly linearly on the number of shot points. To allow for acceptable resolution in accordance with the synthetic modelling results, two sets of 12 geophones were placed with 5 m spacing. An overlap of 3 geophones and a total number of 10 shotpoints enabled the data set to be inverted with the tomographic inversion algorithm. A sledgehammer was used as source. The coupling of the source to the ground was good due to firmly embedded rocks at the surface, which were used as base plate for the sledgehammer.

Due to low ambient noise levels high signal-to-noise ratios were achieved. The first arrival times of the P-waves were picked manually from the seismograms. Due to the comparatively small number of traveltime data, each trace could be analysed in great detail to minimise picking errors.

6.4.2 Results

The results of the refraction seismics survey line along the western clearing are shown in Fig. 6.10. A gradient model with a surface velocity of 500 ms\(^{-1}\), a maximal velocity of 5500 ms\(^{-1}\) and a gradient of 60 ms\(^{-1}\) per metre was used as initial model. As the inversion scheme by Lanz et al. (1998) only allows horizontally layered initial gradient models, topography was incorporated by using a rotated coordinate system. By this, initial velocity contour lines remain parallel to the surface. The regularisation parameter \(\lambda\) was set to 1. The inversion results show isolated anomalies with P-wave velocities between 3000-4500 ms\(^{-1}\) in a host material with velocities between 1000-2000 ms\(^{-1}\). P-wave velocities between 2500-4000 ms\(^{-1}\) are characteristic for permafrost. Combined with the DC resistivity results, these high-velocity anomalies can indeed be interpreted as ice lenses, because air-filled cavities would result in much smaller P-wave velocities (c.f. Fig. 3.3). The maximum velocities of the anomalies are slightly higher than for commonly found unconsolidated permafrost occurrences. This may be due to an inversion artefact, as seen from the synthetic modelling shown above. There, the "true" P-wave velocity of the anomaly (3500 ms\(^{-1}\)) was overestimated by up to 1000 ms\(^{-1}\) in the centre of the anomaly and underestimated near the edges (c.f. Fig. 6.5).

Fig. 6.10b shows the ray paths of the final iteration. The ray coverage is sufficient in the uppermost 10-20 m, including the anomalies, but sparse below. The presence of ray paths in the low-velocity zones between the anomalies indicate an absence of any high-velocity (bedrock) layers in the uppermost 25 m. The reliability of the inversion results can be further analysed by considering the residuals between observed and predicted travel times. Figs. 6.11a and b show a comparison between initial and
final residuals plotted as histograms against distance from the survey midpoint. A significant improvement is observed. Furthermore, the final residuals are centred around 0 ms, indicating no systematic misfits in the final model. Figs. 6.11c and d show a comparison of observed and predicted traveltimes for the initial and final model. Again, the predicted traveltimes are in good agreement with the observed traveltimes for the final model. Finally, Figs. 6.11d and e compare the spatial distribution of the residuals for initial and final model as pseudosections. In contrast to the initial model, the residuals of the final model are small with larger values in the lower, northern part of the profile.
Figure 6.11: Comparison of traveltime residuals for the initial and final model of the inversion results from Val Bever. (a) Histogram of traveltime residuals for the initial and (b) the final model. (c) Comparison of observed (blue) and predicted (red) traveltimes and for the initial and (d) the final model. (e) Residual (predicted−observed) traveltime pseudosection for the initial and (f) the final residuals.
6.5 Conclusions

A refraction seismic tomography scheme was applied for the detection of mountain permafrost. The algorithm was developed at the Institute of Applied and Environmental Geophysics, ETH Zürich, and originally used to delineate shallow landfills. As these localities are easy to access by motorised transport, field surveys are commonly conducted with a large number of geophones and shot points. In mountain permafrost environments limitations to the size of the spread are given by the often remote and difficult access to the field site and the difficult terrain. Standard refraction surveys in the past consisted of 12–24 geophones and 5 shot points.

Key results of synthetic modelling studies and a field case include:

1. The tomographic inversion scheme is indeed applicable for mountain permafrost studies, even for small data sets. The use of 24 geophones and 12 shot points showed good results for the delineation of isolated ice anomalies. Increasing the number of traveltime data does not lead to a proportional increase in the quality of the inversion results.

2. The lateral extent of isolated high-velocity anomalies is correctly resolved, but maximum velocities are overestimated and the base of the anomalies could not be resolved in detail.

3. The correct choice of suitable inversion parameters becomes more important for smaller data sets. Inversions with different regularisation parameters are necessary to distinguish real structural features from noise. Due to the non-linear nature of the model equations the choice of the initial model can strongly influence the inversion results. If a priori information about the subsurface velocity distribution is insufficient, initial velocities should be over-, rather than underestimated.

4. The reliability of the inversion results can be estimated by analysing the ray path coverage of the final iteration. Areas with low ray density correspond to low data sensitivity and results have to be analysed with care. Analysis of the residual differences between observed and predicted traveltimes give additional information about the quality of the inversion results.

5. Results from a field study at Val Bever, Switzerland, showed localised anomalies of 3000–4000 ms\(^{-1}\) along a low-altitude forest clearing. Previous results from a DC resistivity survey showed high-resistive anomalies along the same survey line. Due to the moderate P-wave velocities these high resistivities could clearly be attributed to ice occurrences, as opposed to air-filled cavities with much lower velocities.
Chapter 7

Thermal methods

In the foregoing chapters several indirect geophysical methods and their applicability on mountain permafrost were described. These methods do not measure temperature or ice content directly, which defines and characterises permafrost. Instead, other physical properties like resistivity or seismic P-wave velocity are measured, which may be related to the properties of primary interest, temperature and ice content.

One method, which utilises direct temperature measurements from the surface is the so-called BTS method (Bottom Temperature of the Snow cover, Haeberli, 1973). It has been applied as a standard method to delineate the distribution of mountain permafrost in the Alps and Scandinavia for many years. In the course of the PACE project, BTS surveys using standard thermistor probes were conducted at most field sites to map the permafrost distribution in winter. In addition, passive microwave radiometry was tested to map the BTS on a larger footprint. The radiometry measurements were performed in collaboration with Hansueli Gubler (ALPUG, Davos, Switzerland) and the Institute for Applied Physics of the University of Berne, which were primarily responsible for the work. The measurement principle and some first results are presented in this chapter to complement the indirect geophysical approaches of the foregoing chapters.

7.1 Bottom temperature of snow cover (BTS)

The BTS method, introduced by Haeberli (1973), uses the insulation effect of the winter snow cover to the underlying soil. When the snow cover is dry and about 1 m or more thick, the temperature at the bottom of the snow cover is effectively shielded from short-term variations in the surface energy balance and remains nearly constant. Then, the BTS is mainly controlled by heat transfer from the upper ground layers, which in turn is strongly dependent on the presence of permafrost. A BTS value of $<-3\,^\circ C$ is generally considered corresponding to a probable permafrost occurrence, a BTS of $>-2\,^\circ C$ to a probable non-permafrost site (e.g. Hoelzle, 1992, Hoelzle et al., 1999).
However, the BTS is not a physical constant but varies from one year to the next. Snow fall history of the winter can modify the BTS markedly (Vonder Mühll et al., 1998). The temperature ranges given above have been determined empirically for typical permafrost environments in the Alps and may not be generalised for other climatic and geologic environments. Due to the condition of a dry, permanent and at least 1 m thick snow cover, the method is often not applicable on many wind-blown sites, such as exposed mountain tops (e.g. El Veleta/Sierra Nevada, Janssonhaugen/Svalbard, Juvsassho/Switzerland). Permafrost sites with temperatures close to 0°C often have BTS temperatures in between the ranges given above, making permafrost mapping with the BTS method difficult to nearly impossible (e.g. Schilthorn/Switzerland, Val Bever/Switzerland). Finally, the commonly used thermistor probes measure the temperature at singular points underneath the snow cover, depending much on the surface characteristics. For heterogeneous surface conditions a high spatial resolution is needed to prevent erroneous and noisy data sets. This problem is addressed in the next section, where a radiometric approach is used to obtain larger spatial footprints for BTS measurements.

7.2 Passive microwave radiometry

7.2.1 Introduction

Hansueli Gubler (ALPUG/Davos), a partner of VAW/ETH Zürich within the PACE project, initiated the use of microwave radiometers to map permafrost based on the BTS method. In a preliminary investigation the X-band radiometer MIRA, kindly made available by Chr. Mätzler, Institute of Applied Physics, University of Berne, has been used to measure BTS and typical signatures at Alpine permafrost and non-permafrost sites. The radiometer has been built into a semi-automatic measuring and recording system that allows for direct readings of the apparent brightness temperature $T_B$ and the corresponding BTS. The system can be mounted on a sledge or aboard a helicopter. Test measurements were made at various typical Alpine permafrost locations in Switzerland including rock glacier Murtel, Schilthorn and rock glacier Muragl (Vonder Mühll et al., 2001). The long time goal of this investigation is to built a reliable snowmobile- or helicopter-based system for mapping permafrost.

7.2.2 Theory

Microwave radiometers measure the natural emission of electromagnetic radiation by objects within the viewing angle of the receiving antenna. According to Planck’s law a black body emits a given spectral distribution of electromagnetic radiation depending on the temperature of the object. According to Wien’s law radiation power peaks at a wavelength of about 10 μm for a body temperature of 270 °K. Microwave radiometers (5 to 10 Ghz) are sensitive at much higher wavelength (1 to 6 cm). For these very high wavelengths an approximation of Planck’s law states that the emitted power per unit wavelength is proportional to the physical temperature
Fig. 7.1: Configuration of the radiometry measurements (from Vonder Mühll et al. (2001)).

Temperature $T$ of a black body. Grey bodies emit less than an equivalent black body, therefore their emissivity $e(\lambda)$ is $< 1$. Using Kirchhoff’s law the brightness temperature $TB_p$ of a real body of a physical temperature $T$ is

$$TB_p(\lambda) = e_p(\lambda)T,$$

(7.1)

where $p$ denotes vertical ($v$) or horizontal ($h$) polarisation. $v$ and $h$ are used here in a more general way, $v$ denoting polarisation within the incidence plane and $h$ denoting polarisation rectangular to the incidence plane, or in parallel to the snow surface. In order to separate atmospheric effects from the snow-ground signal, the physical temperature $T$ is calculated from the measured brightness temperature $TB_p$ according to

$$T = \frac{TB_p - (1 - e_p)TB_S}{e_p},$$

(7.2)

where $TB_S$ is the brightness temperature of the sky as observed from the surface in the relevant direction.

Eq. (7.2) is based on the following approximation: The radiometer looking downward at an incident angle $\Theta$ receives direct radiation from the ground surface and the snow, $e_pT$, and sky radiation reflected by the snow cover, $r_p(\Theta)TB_S$ (see Fig. 7.1). For non-transmitting media (ground including snow cover) $r_p(\Theta) = 1 - e_p(\Theta)$. $T$ is the effective physical temperature of the ground-snow system. The totally received radiation is
\[TB_p = e_p T + (1 - e_p) TB_s.\]  \hspace{1cm} (7.3)

Equation (7.2) follows directly from Eq. (7.3). Microwaves at frequencies < 30 GHz usually penetrate dry, fine grained seasonal snow packs with little attenuation. The effective temperature \(T\) is therefore a good approximation for the BTS (Mätzler, 1994). Mätzler (1994) lists emissivities for different types of seasonal snow covers and bare grounds. At increasing frequencies the dependence of \(e_p\) on snow layering, water equivalence of the snow pack and type of ground gets more significant. Especially melt crusts and corse grained depth hoar layers may significantly reduce the apparent emissivity of the ground-snow object.

The contribution of reflected sky radiation can be minimized if the measurements are taken at vertical polarisation and if the incidence angle is equal to the Brewster angle \(\Theta_B = \arctan n\), where \(n\) is the index of refraction, \(n = \sqrt{\epsilon}\), and \(\epsilon\) is the real part of the permittivity. According to the Fresnel equations, radiation reflected from the surface of a dielectric material with \(n > 1\) under the Brewster angle is completely polarised in a plane rectangular to the incidence plane (called here horizontal polarisation, \(h\)). For typical densities for dry, late winter snow covers in potential permafrost areas the Brewster angle is calculated to be about 50°. Therefore, measurements should be taken at an angle of \(\Theta = 50^\circ\) and at vertical polarisation (Vonder Mühll et al., 2001).

### 7.3 Setup of the field measurement and data acquisition system

The MIRA 11.4 GHz radiometer is an X-band mini radiometer built by Chr. Mätzler, W. Amacher and A. Widmer at the Institute of Applied Physics, University Berne, Switzerland. The instrument size is 50 x 30 x 20 cm\(^3\), its weight 8.5 kg. The instrument is built into a rugged aluminium case. For the preliminary field measurements a sledge-based supporting structure for the radiometer has been built. The radiometer is fixed to two swivelling devices allowing independent selection of the incidence angle from ±30° to ±60° and of the polarisation, \(h\) or \(v\). The radiometer antenna axis is located 1.2 m above the snow surface.

Calibration measurements were made at the beginning and at the end of the measurements, as well as two times in between. To avoid extensive drifting of the calibration, the radiometer has to warm up for at least 15 minutes before the first calibration measurement is made. The remaining drift between two calibrations was small but still significant (−0.5°C to −0.7°C for \(TB\)). No attempt was made to correct for this drift because other uncertainties were assumed larger. Simultaneously, the BTS was measured with a conventional BTS gauge within the estimated footprint of the radiometer.
7.4 Field example — Schilthorn

During the winters of 1998 and 1999 measurements were made at different permafrost sites including Murtel rock glacier and Schilthorn. The goal of these measurements was to check the feasibility of the method to measure BTS. At Schilthorn, several measurements were conducted near the drill site. It was shown that ground characteristics play an important role: the method worked well if the soil topography was uniform. Based on these preliminary test surveys, a survey using the radiometer from aboard a helicopter was conducted at rock glacier Muragl/Switzerland in March 2000. The results are very encouraging and it was concluded that under particular snow conditions a BTS survey can be accomplished from aboard a helicopter in very short time (see Vonder Mühll et al., 2001, for details). Here, results are presented from the test survey at Schilthorn.

The radiometer survey line ran parallel to the longitudinal resistivity and seismic lines shown in Fig. 2.8, about 50 m to the west. The upper part of the slope is very steep (40°) with a rough microtopography, the lower part of the downslope measuring line is more gentle (30°) and homogeneous. The sledge with the radiometer and the operator were attached to a cable and lowered slowly (at about 0.5 m/s) downslope with an electric winch. The measurements were made during the movement of the sledge with vertical polarisation (ν), that is perpendicular to the slope, and an incidence angle of 50° with the radiometer looking sidewards. At the lower end of the 100 m long measuring line the sledge was turned around and the measurements were repeated while pulling the sledge upwards. The radiometer was looking roughly at the same stripe of snow moving downward and upward. Two runs were made. Run 1 was done on 27.3.98 with cloudy and very warm weather and a slight drizzle. The uppermost centimetres of the snow cover were wet. During the night the weather cleared up and the surface layer refroze without building a crust. Run 2 was done on 28.3.98 with fair and dry weather.

Fig. 7.2a shows the snow height and the BTS measured with a conventional thermistor probe along the survey line at Schilthorn. The location of the stations are given in distances from the ridge. In Fig. 7.2b the results of the radiometer measurements were averaged within sections of about 10 m. The error bars mark the maximum and minimum readings. In the upper, very steep part of the slope, the incidence angle could not be well controlled because the sledge was tumbling. Besides, the ground and the snow heights were very heterogeneous in this part of the slope so that the large scattering of the data in Fig. 7.2 is not surprising. The data from the lower part of the slope show rather good agreement between the two different BTS measurement methods. The averaged radiometer values are similar for the four different radiometer surveys showing that the results can be reproduced. Systematic errors are present, as seen from the systematic deviations of the four survey results, especially in the lower part of the slope. These are most probable due to the different snow and sky conditions during the two survey days and due to differences in the radiometer footprint between upslope and downslope surveys.

The snow distribution along the survey line is typical of a leeward slope, large accumulation close to the ridge that thins out to a very shallow snow pack some 30 m down from the crest. But also in the more gentle lower part of the slope snow
Figure 7.2: Results from a radiometer test study at Schilthorn along a profile from the ridge (left) to the drill site plateau (right) on the northern flank. (a) Snow height and BTS measured with a conventional thermistor probe. (b) Comparison of the averaged radiometer data from 4 surveys (coloured symbols) along the profile line with the BTS data obtained with the thermistor probe (solid line). The error bars give the maximal and minimal readings of the radiometer.

height variations were still significant due to small terraces on the rock/rock-debris slope. Furthermore, the snow surface within the footprints of the radiometer was partly distorted from walking, especially in the very steep upper part of the slope.

The two methods show the same trend for the BTS, although the scatter for the radiometer measurements is fairly large, especially in the upper part of the profile. As long as the sledge with the radiometer moves smoothly it looks like that averaging leads to fair results.
7.5 Conclusions

The results from the preliminary test studies using the passive microwave radiometer from aboard a sledge and a helicopter suggest that BTS can be measured with microwave radiometers (see Vonder Mühll et al., 2001). Measurements can be done quickly, and if snow conditions and the topography allows, large areas could be scanned from a moving snowmobile or from a helicopter. Rough terrain may significantly influence the measurements due to their strong dependence on polarisation and incidence angle. A radiometer at about 5 GHz instead of 11.4 GHz would further reduce the influence of the snow stratification on the apparent brightness temperature $TB$, but more detailed investigations are necessary to understand the influence of ground roughness and the snow stratification on the measurement signal.
Chapter 8

Discussion

In the foregoing chapters different geophysical methods for investigating in mountain permafrost were introduced. These chapters were focused on synthetic modelling studies to evaluate suitable measurement geometries and inversion parameters. The results from these modelling efforts were used in corresponding field cases to show the applicability of the respective method to specific permafrost problems. In this chapter, the advantages and disadvantages of each method concerning their applicability for specific permafrost problems are discussed in detail.

Permafrost problems amenable to solution by geophysical means fall into three major groups:

1. those defining the distribution
2. those defining the properties and
3. those defining the temporal evolution.

The problem of defining the permafrost distribution may be subdivided into definition of lateral extent (mapping) and vertical extent (sounding). In addition, determining the extent of the active layer is an important task in many environmental and engineering studies. Characterising the permafrost properties, such as ice content and material type, often involves the application of more than one geophysical method, as different permafrost (or non-permafrost) materials may exhibit the same value of a certain geophysical parameter. Finally, determining the temporal evolution of permafrost occurrences by geophysical means has rarely been done in the past, but may be one of the most important tasks for future studies. Under a changing climate the monitoring of soil and rock properties in permafrost regions may provide the necessary information regarding permafrost degradation and related natural hazards.

In the next sections each geophysical method is discussed in detail concerning its applicability to the permafrost problems listed above. The results are summarised in Table 8.1.
<table>
<thead>
<tr>
<th>Method</th>
<th>General permafrost prospection</th>
<th>Permafrost distribution</th>
<th>Characterising permafrost properties</th>
<th>Permafrost monitoring</th>
</tr>
</thead>
<tbody>
<tr>
<td>DC resistivity</td>
<td>![Applicable]</td>
<td>![Applicable]</td>
<td>![Non Applicable]</td>
<td>![Non Applicable]</td>
</tr>
<tr>
<td>FEM methods</td>
<td>![Applicable]</td>
<td>![Applicable]</td>
<td>![Applicable]</td>
<td>![FUTURE]</td>
</tr>
<tr>
<td>TEM methods</td>
<td>![Applicable]</td>
<td>![Applicable]</td>
<td>![Non Applicable]</td>
<td>![STUDIES!]</td>
</tr>
<tr>
<td>Refraction seismic</td>
<td>![Applicable]</td>
<td>![Applicable]</td>
<td>![Applicable]</td>
<td>![Non Applicable]</td>
</tr>
<tr>
<td>Radiometry/BTS</td>
<td>![Applicable]</td>
<td>![Applicable]</td>
<td>![Non Applicable]</td>
<td>![Non Applicable]</td>
</tr>
</tbody>
</table>

Table 8.1: Qualitative comparison of applicability of the geophysical methods used in this study.
Table 8.2: Examples of this study for the application of DC resistivity tomography to general permafrost prospecting.

<table>
<thead>
<tr>
<th>Location</th>
<th>Methodology</th>
</tr>
</thead>
<tbody>
<tr>
<td>Zermatt moraine (section 4.3.4, p. 60)</td>
<td>Detection of ice core</td>
</tr>
<tr>
<td>Murtel rock glacier (section A.8, p. 191)</td>
<td>Delineation of main ice body</td>
</tr>
<tr>
<td>Stelvio rock glacier (section A.9, p. 193)</td>
<td>Delineation of main ice body</td>
</tr>
<tr>
<td>Stelvio ski run (section A.9, p. 192)</td>
<td>Detection of newly formed permafrost</td>
</tr>
<tr>
<td>Val Bever (section 5.3.1, p. 101)</td>
<td>Detection of isolated ice patches</td>
</tr>
<tr>
<td>Juvvashhoe (section A.4, p. 177)</td>
<td>Detection of frozen bedrock</td>
</tr>
</tbody>
</table>

8.1 General mountain permafrost prospecting

Before considering the specific permafrost problems, the applicability of the methods for detecting mountain permafrost in general is discussed. Ideally, the presence of permafrost should be detected without using any a priori information, and independent of host material, ice content and dimension of the permafrost occurrence.

DC resistivity tomography (Table 8.2)

**Advantages** Judging from the results of this work the most universally applicable method is the DC resistivity tomography technique. Permafrost was successfully detected for several different permafrost environments including rock glaciers, ice-cored moraines, a low-altitude forested valley slope and bedrock (see Table 8.2). Because the method primarily depends on the unfrozen water content, it is very sensitive to temperature changes, even in the presence of very high resistivity values. The method is applicable under various surface conditions, provided that there is sufficient electrical contact between electrodes and the ground. As pointed out in chapter 4 this can be enhanced by using sponges soaked in salt water.

**Disadvantages** Due to the galvanic contact required between electrodes and ground, DC resistivity is usually only applicable on snow-free surfaces. Furthermore, the high contact resistances obtained between electrodes and ground often prohibit the injection of sufficient electrical current into the ground. The commonly used Wenner electrode array has limited lateral resolution and tends to smooth out horizontal small-scale variability. The Double-Dipole array with higher lateral resolution is usually non-applicable, as its signal-to-noise ratio is too small. The sensitivity of the inversion results to the input data set is low for highly resistive model regions. The resistivity of highly resistive anomalies tends to be underestimated by the inversion process. Finally, the applicability of the method depends on the contrast in resistivity between the frozen and unfrozen state of the material. For materials with very low water content, like in many rocks, this contrast can be very small.
Table 8.3: Examples of this study for the application of FEM methods to general permafrost prospecting.

<table>
<thead>
<tr>
<th>Example</th>
<th>Section/P.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stelvio (section 5.2, p. 95)</td>
<td></td>
</tr>
<tr>
<td>Tarfala (section A.3, p. 176)</td>
<td></td>
</tr>
<tr>
<td>Hiorthfjellet rock glacier (section A.2, p. 174)</td>
<td></td>
</tr>
<tr>
<td>Val Bever (section 5.3.1, p. 101)</td>
<td></td>
</tr>
<tr>
<td>Juvvasshøe (section A.4, p. 179)</td>
<td></td>
</tr>
</tbody>
</table>

FEM methods (Table 8.3)

**Advantages**  Conductivity meter as introduced in this study are generally lightweight and easy to handle instruments. Data acquisition is fast and data processing is simple. As a magnetic field is used to induce the electric current in the ground, no direct contact between the instrument and the ground is needed. Consequently, surveys can be made on snow-covered ground, providing the snow cover is dry and homogeneous.

**Disadvantages**  As explained in detail in chapter 5 the penetration depth of the FEM-instruments used in this study is limited to about 6 m and vertical resolution is sparse to absent. Furthermore, due to the very low conductivities observed in mountain permafrost environments, the absolute conductivity values are close to the resolution limit of the instruments and cannot be used as an indicator for the presence of permafrost. Furthermore, the instrument drift can be substantial compared to the permafrost signal and conductivity anomalies due to changing surface characteristics can mask any conductivity variations in the subsurface. Consequently, reliable detection of mountain permafrost by an EM-31/GEM-300 survey without additional information is limited to some special cases.

TEM methods (Table 8.4)

**Advantages**  In chapter 5 the results from the PROTEM soundings at various field sites have shown that deep permafrost occurrences can be detected with this method. As the TEM methods work in the time-domain, large penetration depths can be obtained with comparatively small measurement geometries at the surface. This is especially useful for deep soundings on geomorphological features with small lateral extent, e.g. rock glaciers and moraines. Similar to the FEM methods, no direct contact between instrument and ground is needed, facilitating the conduction of surveys on snow covered ground.

**Disadvantages**  Due to technical limitations no information about the uppermost 5–10 m can be obtained. Furthermore, spatial resolution is limited to the transmitter
Table 8.4: Examples of this study for the application of TEM methods to general permafrost prospecting.

<table>
<thead>
<tr>
<th>Location</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Janssonhaugen (section 5.3.4, p. 107)</td>
<td>Bedrock permafrost site with a gradual decrease in resistivity with depth</td>
</tr>
<tr>
<td>Hiorthfjellet rock glacier (section 5.3.4, p. 105)</td>
<td>Arctic rock glacier with gradual decrease in resistivity due to frozen bedrock underneath the ice body</td>
</tr>
<tr>
<td>Murtel rock glacier (section 5.3.4, p. 108)</td>
<td>Typical Alpine rock glacier with highly resistive middle layer</td>
</tr>
<tr>
<td>Foscagno rock glacier (section 5.3.4, p. 108)</td>
<td>Alpine rock glacier with thin ice body</td>
</tr>
</tbody>
</table>

Coil dimensions. For the configuration used in this study the resolution limit in the lateral dimension is around 50 m. Even though TEM methods are highly sensitive to the thickness of a resistive layer (see below), the sensitivity is poor for determining its resistivity value. Finally, as for all methods utilising electrical resistivity as geophysical parameter, a significant contrast between the resistivity of unfrozen and frozen material is needed to successfully delineate the permafrost occurrence.

Refraction seismic tomography (Table 8.5)

**Advantages** Similar to DC resistivity, refraction seismic tomography is well suited to detect mountain permafrost (chapter 6). Isolated permafrost occurrences can be detected with even higher resolution than in DC resistivity surveys (see section 8.5). Logistic effort and survey time are similar for both methods. In contrast to the electric and electromagnetic techniques refraction seismic make use of a different geophysical parameter, the P-wave velocity. In addition, the method provides more structural information about the subsurface than DC resistivity and inductive EM methods.

**Disadvantages** The higher resolution in seismic surveys is in contrast to the partly lower accuracy of the results, especially in model regions where ray coverage is poor during inversion (see sections 3.5 and 3.6). The P-wave velocity of high-velocity anomalies tends to be overestimated by the inversion process. Finally, the P-wave velocity is not as sensitive to temperature changes in partly frozen materials as resistivity, because seismic waves travel along the frozen part of the material, as opposed to the resistivity, which depends on the unfrozen part. In partly frozen material, a further temperature decrease will significantly change the resistivity, whereas the P-wave velocity will not change much, as soon as a continuous frozen matrix has been established.

BTS and radiometry (Table 8.6)

**Advantages** The BTS method is directly measuring a temperature signal of the ground, which does not have to be related to an additional geophysical parameter.
Table 8.5: Examples of this study for the application of refraction seismic tomography to general permafrost prospecting.

<table>
<thead>
<tr>
<th>Location</th>
<th>Application</th>
</tr>
</thead>
<tbody>
<tr>
<td>Val Bever</td>
<td>Detection of isolated ice patches</td>
</tr>
<tr>
<td>Juvvasshoe</td>
<td>Detection of the permafrost transition zone</td>
</tr>
</tbody>
</table>

Table 8.6: Examples of this study for the application of the BTS and radiometry method to general permafrost prospecting.

<table>
<thead>
<tr>
<th>Location</th>
<th>Application</th>
</tr>
</thead>
<tbody>
<tr>
<td>Schilthorn</td>
<td>Comparison of conventional BTS probing and the radiometry method</td>
</tr>
</tbody>
</table>

The method is fast, easy to use and no processing of the data is required. The radiometry method is measuring the BTS on a larger footprint. Preliminary studies show a great potential for mapping large areas from aboard a helicopter.

**Disadvantages** The largest restriction on the applicability of the BTS method is the condition of the presence of an undisturbed snow cover of at least 0.8–1 m thickness. The snow cover has to be present for about one month prior to the measurements, in order that BTS can be related to the permafrost conditions. In addition, the BTS is not constant from one winter to the next, depending on the snow fall history during the winter. Conventional measurements with probes are conducted at singular points underneath the snow cover, depending much on the surface characteristics. Furthermore, the method and the commonly used BTS ranges (>-2°C and <-3°C) have been calibrated based on measurements in the Central Alps and a generalisation to other mountain permafrost environments is uncertain. The radiometry method is not yet operational. The main problems still to be solved are the strong dependence on the measurement angle, the temperature and layering of the snow cover and the amount of cloudiness.

### 8.2 Mapping the lateral variation of mountain permafrost

**DC resistivity tomography** (Table 8.7)

**Advantages** Due to its 2-dimensional representation, the DC resistivity tomography is well suited to determine lateral variations of permafrost. The horizontal resolution can be increased by decreasing the electrode spacing. For data sets with a low amount of measurement noise, the representation of small-scale variations can be improved by using smaller regularisation factors in the inversion. In comparison to FEM data the information content is larger, because data are obtained and inverted in two dimensions.
Table 8.7: Examples of this study for the application of DC resistivity tomography to mapping lateral variations of permafrost.

<table>
<thead>
<tr>
<th>Location</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Juvvasshøe (section A.4, p. 181)</td>
<td>Mapping the altitudinal limit of permafrost and the permafrost transition zone</td>
</tr>
<tr>
<td>Stelvio ski run (section A.9, p. 192)</td>
<td>Mapping the lateral extent of isolated ice patches</td>
</tr>
<tr>
<td>Murtel rock glacier (section A.8, p. 191)</td>
<td>Mapping the lateral extent of the ice-rich permafrost body</td>
</tr>
<tr>
<td>Val Bever (section 8.5.1, p. 151)</td>
<td>Good example for the smoothing effect of the Wenner array in comparison to the higher lateral resolution of the refraction seismic results</td>
</tr>
<tr>
<td>Zermatt moraine (section 4.3.4, p. 57)</td>
<td>Example for the effect of different regularisation parameters for resolving small-scale lateral variability</td>
</tr>
</tbody>
</table>

**Disadvantages** Even though DC resistivity is a suitable method for delineating permafrost occurrences of limited extent, it is not feasible for mapping large areas. 3-dimensional survey techniques have been proposed recently (e.g. Loke and Barker, 1996b), but survey speed is slow and inversion routines can only process a comparatively small number of data points. Another possibility is to use several 2D survey lines along a 3D grid. However, as a 200 m DC resistivity survey line takes approximately 90–120 minutes to measure, survey areas of one to several square-kilometres are nearly impossible to cover with a reasonable logistic effort. On smaller scales, the smoothing effect of the Wenner array often prohibits detection of small-scale variability on the order of one to two electrode spacings.

**FEM methods (Table 8.8)**

**Advantages** Due to their mapping speed FEM instruments like the EM-31 and the GEM-300 are much better suited for covering large areas than 2-dimensional tomographic methods, even though only the average conductivity of the uppermost 6 m is measured and no vertical resolution can be obtained. In section 5.3.1, a procedure was proposed, combining the advantages of the high resolution DC resistivity tomography method with the mapping speed of the EM-31. By this, a survey area of several square-kilometres can be investigated in one day.

**Disadvantages** For long survey duration and small conductivity contrasts, the instrument drift may influence the survey results. Repeated measurements of survey lines and recalibration of chosen grid points may overcome this problem. On snow-covered ground, influences through variations in snow height and composition have to be taken into account. FEM mapping of permafrost variability can not be done without the use of additional information at certain representative locations.
Table 8.8: Examples of this study for the application of FEM methods to mapping lateral variations of permafrost.

<table>
<thead>
<tr>
<th>Location</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stelvio ski run (section A.9, p. 192)</td>
<td>Mapping the lateral extent of isolated ice patches over a larger area</td>
</tr>
<tr>
<td>Val Bever (section 5.3.1, p. 101)</td>
<td>Mapping the distribution of isolated ice patches on a forested valley slope</td>
</tr>
<tr>
<td>Juvvasshøe/Schilthorn (section 5.3.2, p. 102)</td>
<td>Mapping the lateral variability around the PACE boreholes to determine the representativeness of the borehole results</td>
</tr>
</tbody>
</table>

Table 8.9: Examples of this study for the application of TEM methods to mapping lateral variations of permafrost.

no TEM mapping studies have been conducted in this study

TEM methods (Table 8.9)

**Advantages** A series of PROTEM soundings along a profile line or grid may be combined to give a 2D- or even 3D-representation of the permafrost distribution. This approach has been successfully used on polar permafrost sites, where topographical influences and variations in the subsurface geology are small (e.g. Todd and Dallimore, 1998, Harada et al., 2000).

**Disadvantages** As no information about the uppermost 5–10 m can be obtained and spatial resolution is limited by the transmitter coil dimensions (see chapter 5), the application of PROTEM surveys for mapping lateral variability of mountain permafrost is limited. Furthermore, 2D and 3D inversion algorithms are not as highly developed as for DC resistivity or seismic techniques. Consequently, most 2D TEM data is shown in the form of pseudosections (see 3.3).

Refraction seismic tomography (Table 8.10)

**Advantages** Similar to DC resistivity, refraction seismic tomography is well suited to determine lateral variations of permafrost due to its 2-dimensional representation. In contrast to DC resistivity tomography the lateral resolution of refraction seismics is high, providing the ray coverage is sufficient. It can be further increased by decreasing the geophone spacing or increasing the number of shot points. For data sets with a low amount of measurement noise, the representation of small-scale variations can be improved by using smaller regularisation factors in the inversion.

**Disadvantages** Like DC resistivity, refraction seismic is not feasible for mapping large areas due to the large logistic requirements. A large number of shot points is needed to obtain a good ray coverage. Data processing is more time consuming than for DC resistivity, because the first arrival times have to be picked prior to the inversion process. For very large data sets, the picking can not be done manually and automatic picking schemes have to be employed.
Table 8.10: Examples of this study for the application of refraction seismic tomography to mapping lateral variations of permafrost.

<table>
<thead>
<tr>
<th>Juvvasshøe (section A.4, p. 181)</th>
<th>Mapping the permafrost transition zone</th>
</tr>
</thead>
<tbody>
<tr>
<td>Val Bever (section 8.5.1, p. 151)</td>
<td>Good example for the higher lateral resolution of the refraction seismic results compared to DC resistivity tomography</td>
</tr>
</tbody>
</table>

Table 8.11: Examples of this study for the application of the BTS/radiometry method to mapping lateral variations of permafrost.

no BTS/radiometry mapping studies have been conducted in this study

BTS and radiometry (Table 8.11)

**Advantages** In winter time, the BTS method is still the most important survey technique for mapping the permafrost distribution. It is the only method which determines the permafrost distribution from its thermal properties, that is without using resistivity or seismic velocity as reference parameter. The survey speed is very fast and no data processing is required. The passive microwave radiometry method introduced in this work aims to improve this approach in terms of measurement speed, e.g. through airborne measurements, and concerning spatial representativeness. As a larger footprint becomes averaged, the BTS results may lose its dependency on local heterogeneities

**Disadvantages** Because a thermal signal is measured, one cannot distinguish between different subsurface materials or material properties. As explained above, BTS surveys require a permanent 0.8–1 m thick snow cover for at least one month. The results are depending on the snow fall history and may vary from one year to the next.

8.3 Determining the vertical extent of mountain permafrost (sounding)

Determining the vertical extent of permafrost is one of the major tasks in permafrost studies. The classical approach is to use vertical electrical soundings, assuming vertical layering and laterally homogeneous ground conditions. As pointed out in section 3.3, at mountain permafrost sites this assumption is seldom valid, as ground conditions can be very variable on a small scale.

DC resistivity tomography (Table 8.12)

**Advantages** Due to its 2-dimensional representation DC resistivity tomography surveys overcome the problem of lateral heterogeneity, as shown in section 4.3. The
Table 8.12: Examples of this study for the application of DC resistivity tomography to determine vertical variations of permafrost.

<table>
<thead>
<tr>
<th>Location</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stelvio rock glacier (section A.9, p. 193)</td>
<td>Good example for successful detection of the permafrost base</td>
</tr>
<tr>
<td>Murtel rock glacier (section A.8, p. 191)</td>
<td>Example, where the permafrost base is deeper than the penetration depth</td>
</tr>
<tr>
<td>Zermatt moraine (section 4.3.4, p. 60)</td>
<td>Example, where the depth of the permafrost base is similar to the penetration depth</td>
</tr>
</tbody>
</table>

Table 8.13: Examples of this study for the application of FEM methods to determine vertical variations of permafrost.

<table>
<thead>
<tr>
<th>Methods</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>no FEM sounding studies have been conducted in this study</td>
<td></td>
</tr>
</tbody>
</table>

thickness of the active layer can be determined and for shallow permafrost occurrences the depth of the permafrost base can be delineated. In comparison to electromagnetic soundings the information content is higher, because the results can be inverted in a 2-dimensional way. To improve vertical resolution the Wenner-Schlumberger array can be used instead of the Wenner array used in this study.

**Disadvantages** In order to obtain deep sounding data, large electrode spacings have to be used. To obtain a good vertical resolution, the horizontal resolution has to be increased as well, leading to a large number of electrodes. Consequently, survey speed would be low and data processing requires large computer capacities. Synthetic modelling studies have shown that DC resistivity surveys tend to underestimate the thickness of a resistive middle layer.

**FEM methods (Table 8.13)**

**Advantages/Disadvantages** With the FEM instruments used in this study no significant vertical resolution can be obtained. As described in chapter 5 some information about vertical conductivity changes may be obtained by using different instrument heights above the ground, different instrument polarisations or different transmitter-receiver spacings. The last option can not be done with the EM-31 or GEM-300, but is possible for the Geonics EM-34, which uses two separate coils joined by a flexible cable. However, for deep electromagnetic soundings, TEM methods are much superior to FEM systems (e.g. Telford et al., 1990).

**TEM methods (Table 8.14)**

**Advantages** For delineating the base of deep permafrost occurrences transient electromagnetic soundings, as performed with the PROTEM was found to be the most promising technique. Measurement speed is fast and the equipment may be carried by 2–3 persons. The size of the measurement configuration is small compared
Table 8.14: Examples of this study for the application of TEM methods to determine vertical variations of permafrost.

<table>
<thead>
<tr>
<th>Location</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Janssonhaugen (section 5.3.4, p. 107)</td>
<td>Determining the depth of the permafrost base for a deep permafrost occurrence</td>
</tr>
<tr>
<td>Murtel rock glacier (section 5.3.4, p. 108)</td>
<td>Determining the base of an Alpine rock glacier</td>
</tr>
<tr>
<td>Foscagno rock glacier (section 5.3.4, p. 108)</td>
<td>Determining the base of a shallow Alpine rock glacier</td>
</tr>
</tbody>
</table>

to the penetration depth, yielding penetration depths of more than 100 m with a 5 m by 5 m multiturn cable. For comparison, a DC resistivity sounding with a Schlumberger array would need a cable spread, and therefore homogeneous ground conditions, of at least 1–2 km for a similar penetration depth. The results shown in section 5.3.4 were best for targets with large resistivity contrasts, e.g. Alpine rock glaciers.

**Disadvantages** The TEM method is poor at resolving the resistivity of an intermediate resistive layer, where DC resistivity methods are much superior. Consequently, joint interpretation or joint inversion of both data types would be very effective at reducing the deficiencies of either technique when used alone (e.g. Sandberg, 1993, Maier et al., 1995).

**Refraction seismic tomography (Table 8.15)**

**Advantages** Similar to DC resistivity the thickness of the active layer and the depth of the permafrost base can be delineated for shallow permafrost occurrences. The 2-dimensional representation increases the information content and the reliability of the interpretation. In case of vertical layering and increasing velocities with depth, the depth of the respective layers can be determined with great accuracy.

**Disadvantages** In addition to the logistic problems concerning long survey spreads and a large number of sources and receivers, seismic soundings in permafrost terrain are hampered by the difficulty to resolve a large low-velocity zone below a high-velocity zone. If the high-velocity zone, corresponding to the ice-rich permafrost layer, is laterally extending and velocity is decreasing with depth, all wave energy will become focused in this layer and no information about deeper, unfrozen layers can be obtained. Furthermore, ray density decreases with depth leading to uncertainties concerning the thickness of isolated high-velocity anomalies, as shown in the synthetic modelling studies in section 6.3.3.

**BTS and radiometry**

No depth resolution can be obtained with this method.
8.4 Active layer studies

The active layer, being the uppermost subsurface layer and subject to large seasonal changes in its physical properties, is the focus of many permafrost studies.

DC resistivity tomography (Table 8.16)

**Advantages** By reducing the electrode spacing, the active layer can be investigated in great detail. Especially through repeated surveys along the same lines, active layer processes like freezing and thawing can be visualised. By conducting laboratory studies using miniature electrodes, resistivity changes can be related to temperature changes or lateral and vertical differences in water content. Due to its high sensitivity to unfrozen water content, resistivity monitoring of phase transitions and of the flow paths of rain and melt water is highly effective.

**Disadvantages** For standard resistivity surveys as conducted in this study, an electrode spacing of 5 m is too large to resolve the active layer sufficiently. Decreasing the electrode spacing decreases the penetration depth for a constant number of electrodes. Consequently, simultaneously resolving the active layer and the permafrost body below requires a large number of electrodes and increases measurement and processing time.

FEM methods (Table 8.17)

**Advantages** Generally, FEM methods have a strong applicability to active layer studies, especially concerning the lateral variability. Due to their small penetration
Table 8.17: Examples of this study for the application of FEM methods to active layer studies.

<table>
<thead>
<tr>
<th>Location</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Murtel and Hiorthfjellet rock glacier (section A.8 and A.2, p. 187 and p. 174)</td>
<td>Examples for increasing conductivity values to the rock glacier front</td>
</tr>
<tr>
<td>Foscagno rock glacier (section 5.3, p. 102)</td>
<td>Active layer mapping of the frontal part, again showing increasing conductivity values to the rock glacier front</td>
</tr>
<tr>
<td>Foscagno rock glacier (section A.10, p. 194)</td>
<td>Detection of differences in the active layer thickness on two parts of rock glacier Foscagno</td>
</tr>
</tbody>
</table>

depth, the EM-31 and GEM-300 are very sensitive to changes in the active layer, as was shown in section 5.3. In contrast to DC resistivity and refraction seismic, FEM methods are easy to use on blocky terrain, like rock glaciers, as they do not need electrical contact with the surface. Consequently, they are well suited to investigate changes in the near subsurface of rock glaciers, where a significant increase in conductivity to the rock glacier front was noted (Haeberli and Vonder Mühll, 1996). Survey lines can easily be repeated due to the fast survey speed. Limited depth resolution may be obtained by conducting surveys with different instrument heights above the ground or different instrument polarisations.

**Disadvantages** As mentioned above, without using different polarisations or instrument heights, only the bulk conductivity of the active layer can be obtained. Consequently, it can not be distinguished between apparent conductivity changes due to specific conductivity variations and changes due to active layer thickness.

**TEM methods**

As explained above, the PROTEM transmitter current cannot be stopped instantaneously and the first time gate recorded must occur a certain time after the termination. Consequently, no information from the topmost 5–10 m can be obtained.

**Refraction seismic tomography (Table 8.18)**

**Advantages** Due to the high lateral resolution, refraction seismic tomography is very suitable for investigating the active layer in detail. In the shallow subsurface the ray coverage is large and high accuracy can be obtained without reducing the geophone spacing.

**Disadvantages** For DC resistivity surveys temporal variations in the active layer can be measured by repeating survey lines using fixed-electrode arrays (see section 4.5). In contrast to electrodes, which may consist of cheap steel rods (e.g. tent pegs), it is generally not feasible to leave geophones at fixed locations in the field for a long time. Consequently, difficulties arise in securing that the same source and receiver locations are used for repeated surveys.
Table 8.18: Examples of this study for the application of refraction seismic tomography to active layer studies.

| Juvvasshøe (section A.4, p. 181) | Detecting air-filled cavities, water and rock anomalies in the active layer |

BTS and radiometry

BTS studies can only be conducted in winter, when the active layer is frozen.

8.5 Characterising the physical properties of mountain permafrost

As explained above and in chapter 4, electric and electromagnetic methods are especially well suited to detect permafrost, because the electrical resistivity depends strongly on the unfrozen water content, which is highly dependent on the amount of freezing. However, for certain permafrost environments, interpretation of DC resistivity and EM results are difficult, as the measured resistivity values can be caused by several materials. Whereas the resistivity contrast between ice and unfrozen water is huge, it is small between ice, air and certain rock types, as all three nearly behave as an electrical insulator with very high resistivities (see Fig. 3.2). Furthermore, the resistivity values for frozen ground span a wide range from about 10 kΩm to a few MΩm, depending on the ice content. In contrast to that, seismic P-wave velocities in unconsolidated permafrost are bounded from above (2000–4000 ms⁻¹) and the P-wave velocities for ice and air are markedly different (see Fig. 3.3). On the other hand, the values for most common rock types (e.g. sandstone, limestone or granite) are similar to the values for ice, so the detection of mountain permafrost is often difficult by refraction seismic alone. From this it becomes clear that for some permafrost applications both methods have to be combined, in order to get unambiguous results in terms of permafrost delineation.

To illustrate this, three field cases are presented, where DC resistivity failed to unambiguously detect permafrost occurrences and refraction tomography was needed to correctly interpret the results. In all cases the goal was to decide, if high resistivity anomalies correspond to air-filled cavities, ice or bedrock anomalies.

8.5.1 Val Bever: ice anomalies

The first example presents survey results from Val Bever and was already discussed in detail in sections 5.3.1 and 6.4. The results of the refraction seismics survey line along the western clearing are shown together with the DC resistivity tomography results (Fig. 8.1). Isolated anomalies with P-wave velocities between 2500–4500 ms⁻¹ are present in a host material with velocities between 1000–1500 ms⁻¹. The locations coincide with the locations of the high-resistivity anomalies, except that the P-wave velocity anomalies are smaller and much more sharply delineated than the
resistivity anomalies. In chapter 4 it was shown that the Wenner electrode array has comparatively low resolution for detecting laterally varying structures and sharp edged anomalies are smoothed out during inversion. Accordingly, both survey results represent the same anomalies, which are delineated more clearly by the refraction tomography. Due to the high resistivity values and P-wave velocities between 2500–4500 ms\(^{-1}\), they can indeed be characterised as ice lenses, because air cavities would result in much smaller P-wave velocities (Fig. 3.3).

### 8.5.2 Schilthorn: bedrock anomalies

At the PACE drill site Schilthorn, subsurface temperatures are only slightly below 0°C and measured resistivities are low (1–5 k\(\Omega\)m). In section 4.5 the results from repeated DC resistivity tomographies are presented, showing that the resistivity of the active layer varies substantially throughout the year. However, the question remained whether locally high resistivity anomalies, as seen in Fig. 4.18 and as reported from several locations near the drill site (Russill, 2000, K. Giesa, personal communication), could be due to isolated ice occurrences at this permafrost site. As mentioned before, rock and ice anomalies may have similar resistivities, depending on the rock type and the ice content. In contrast to that, the seismic P-wave velocity of ice is bounded from above, that is if the velocity is much higher than 3000–4000 ms\(^{-1}\) (P-wave velocity for ice) the ice content cannot be high and anomalies are
Figure 8.2: Seismic inversion results (top panel) for the survey line across the drill site at Schilthorn in comparison to the DC resistivity tomography results (bottom panel). The location of the 14 m borehole is marked with the black line. Regions I–II correspond to features explained in the text.

Figure 8.3: Seismic inversion results (top panel) for the survey line at the non-permafrost site near Juvvasshoe, Norway, in comparison to the DC resistivity tomography results (bottom panel). The very low velocities (< 500 m/s) in the uppermost 1–2 m indicate the presence of air-filled cavities. Regions I–II correspond to features explained in the text.
Fig. 8.2 shows the inversion results for the refraction seismic survey line together with the corresponding DC resistivity profile. Both profiles were recorded in summer 1998, a few months before the 14m drilling was done. Again, the standard inversion parameter were used. The weathered layer in the uppermost 5 m (region I, c.f. sections 2.5 and 4.5) is clearly delineated showing P-wave velocities between 1000 and 2000 ms\(^{-1}\). Underneath, high-velocity anomalies (up to 7000 ms\(^{-1}\)) are found, roughly coinciding with the high-resistive anomalies in the DC resistivity results (region II). These results indicate the presence of bedrock anomalies, as the P-wave velocities are much too high to be caused by ice. Besides, the structures of the two tomograms are not identical. This may be a further hint to an absence of any large quantity of ice, which would be delineated clearly by both methods.

8.5.3 Juvvasshøe: air cavern anomalies

The third example addresses the problem of high resistivity values due to air-filled cracks or cavities, which may be misinterpreted for permafrost occurrences. Cracks or air-filled cavities can be found frequently on mountainous terrain, especially near large boulders and underneath a debris covered surface. Resistivities can be anomalously high, as air acts as a perfect electrical insulator. However, the seismic P-wave velocity in air is very low (330 ms\(^{-1}\)), therefore easy to contrast to the much higher velocities of rock and ice. The example shown in Fig. 8.3 originates from a non-permafrost site near Juvvasshøe, Norway (site C in Fig. 2.4). As explained in detail in Hauck et al. (2000) and shown in Appendix A.4, an extensive survey along the northern slope of the drill site at Juvvasshøe was conducted to delineate the transition area between permafrost and non-permafrost. A 640 m long DC resistivity profile showed a clear transition zone between high resistivities of up to several tens of kΩm and resistivities as low as 1 kΩm. However, anomalously high resistivity values were also found near the surface in the presumed non-permafrost areas (see blue region I in the DC resistivity tomogram in Fig. 8.3). To determine the cause for these anomalies, a refraction survey was conducted along the same profile line. From the results in Fig. 8.3 it is clearly seen that air-filled cavities have to be present, as the P-wave velocities in the uppermost 1–2 m are lower than 500 ms\(^{-1}\). Below, the values increase slowly to 1000–1500 ms\(^{-1}\), indicating the presence of unconsolidated, unfrozen material and water, which is in good agreement with the low resistivity values found at this depth (region II in Fig. 8.3).

8.6 Permafrost monitoring

In a changing climate the monitoring of the permafrost distribution, as well as the detection of changes in the physical properties of mountain permafrost is one of the most important tasks to assess the potential of an increased frequency of permafrost related natural hazards. Temperature monitoring in deep boreholes and monitoring of BTS temperatures on a larger spatial scale present the foremost tasks to be able
to detect permafrost degradation. However, to detect changes in the physical properties, like a decreasing ice content or increasing unfrozen water content, monitoring of geophysical parameters is essential. In this section the potential of each method for permafrost monitoring is discussed.

**DC resistivity tomography**

In section 4.5 results from repeated DC resistivity tomography measurements using a buried, fixed-electrode array were shown. Resistivity variations in the uppermost 10 m could be related to temperatures and the evolution of the unfrozen water content with time was determined. This approach is probably the most promising for future studies, especially in combination with energy balance measurements and/or other geophysical measurements on a seasonal basis (e.g. seismics). Once a relation between resistivity and temperature is established the resistivity evolution could be used as proxy for the temperature evolution at different depths. Hereby, not the absolute resistivity values, which depend on the material, are important, but the rate of resistivity increase/decrease or the differences between consecutive measurements, which depend solely on the amount of freezing/thawing, and thereby on temperature. As electric conduction mostly takes place in the unfrozen part of the pore water, resistivity is very sensitive to freeze and thaw processes, even for very low unfrozen water contents.

**FEM methods**

Using FEM methods for monitoring purposes is not as feasible as the DC resistivity approach described above, because no fixed measurement setup can be installed to accurately monitor conductivity changes. However, as FEM surveys are fast and can be conducted with minimal effort, measurements along the same survey lines may be repeated on a regular basis. Due to the strong sensitivity to instrument height and surface conditions, care has to be taken to ensure similar measurement conditions.

**TEM methods**

In a warming climate, permafrost is likely to thaw from the top down, that is, the most drastic changes may be seen at shallow depth. As TEM methods give no reliable results for the uppermost 5-10 m, their monitoring potential is believed to be small.

**Refraction seismics**

In principle, refraction seismics may be used for monitoring similar to the DC resistivity approach shown above. A fixed geophone array can be installed and measurements may be conducted along the survey line using different source points. Determining the velocity changes due to freezing and thawing would facilitate the
interpretation of refraction seismic survey data from mountain permafrost areas. Together with changes in resistivity, these velocity changes could be used to accurately determine changes in unfrozen water and ice content. In addition, laboratory studies as shown in section 4.4 for the DC resistivity technique could be performed to determine velocity-temperature relationships for typical permafrost materials.
Chapter 9

Conclusion and outlook

Five geophysical methods were tested for applicability on mountain permafrost. Hereby, the focus was on 2-dimensional mapping techniques, which included state-of-the-art tomographic data inversion algorithms to improve data interpretation. The geophysical methods included DC resistivity tomography, electromagnetic conductivity mapping (EM-31 and GEM-300), transient electromagnetic soundings (PRO-TEM), refraction seismic tomography and passive microwave radiometry. Their applicability was evaluated concerning specific permafrost related questions. The field data of this study was obtained within the PACE project, including surveys at all PACE drill sites and at additional test sites, where specific permafrost environments were present. In addition, a DC resistivity field monitoring system and a miniature DC resistivity laboratory system were developed for permafrost monitoring purposes and to determine the unfrozen water content in partially frozen material, respectively.

For the general prospecting of mountain permafrost occurrences DC resistivity tomography and refraction seismic tomography showed the best results, especially for heterogeneous ground conditions. Both methods are comparatively fast, easy to apply and commercial 2- and 3-dimensional tomographic inversion schemes are available. As the contrast in resistivity and seismic velocity between the unfrozen and frozen phase is largest for water (compared to rocks), the methods work best for permafrost sites with comparatively large ice/water contents. In DC resistivity surveys the necessity to obtain sufficient electrical coupling with the ground may lead to difficulties on dry, debris covered surfaces. Sponges soaked in salt water can be used to reduce the contact resistances. To remove interpretational ambiguities both methods should be combined.

Even though DC resistivity and refraction seismic are suitable for delineating permafrost occurrences of limited extent, they are not yet feasible for mapping the lateral distribution of mountain permafrost over large areas. A 200 m DC resistivity survey line takes approximately 90–120 minutes to measure, rendering survey areas of one to several square-kilometres impossible. Electromagnetic conductivity meter like the EM-31 and the GEM-300 are much better suited for this task, even though only an average conductivity of the uppermost 6 m can be obtained. In this work a procedure is proposed, combining the advantages of higher resolution of the
DC resistivity tomography method with the mapping speed of the EM-31. By this, several square-kilometres can be investigated in one day.

For delineating the vertical extent of permafrost occurrences the transient electromagnetic sounding system PROTEM was found to be most promising. The size of the measurement configuration is small compared to the penetration depth, yielding penetration depths of more than 100 m with a 5 m by 5 m multiturn cable. Again, the results were best for targets with large resistivity contrasts, e.g. rock glaciers.

In winter time, the BTS method is still one of the most important survey techniques for permafrost mapping, provided that a 0.8–1 m thick snow cover is present for the main part of the winter. The passive microwave radiometry method introduced in this work aims to improve this approach in terms of measurement speed (e.g. through airborne measurements), as well as spatial representativeness. As a larger footprint becomes averaged, the results may lose its dependency on local heterogeneities. However, keeping a constant measurement angle and reducing the influence of clouds and snow cover, are still substantial problems to be solved.

To characterise the physical properties of permafrost, e.g. to determine the material composition and the ice content, or to learn more about the origin of the permafrost or the processes involved, only a combination of several methods is promising. Especially refraction seismics should be applied as a standard complementary technique to DC resistivity wherever possible, as it utilises a completely different physical property. Furthermore, the combined application of two tomographic methods improves data interpretation in a tremendous way, as not only structural information can be compared, but also quantitative analyses in terms of indirectly determined physical properties seem possible.

For permafrost monitoring on a seasonal basis, repeated DC resistivity tomography measurements using a buried, fixed-electrode array were performed. Resistivity variations in the uppermost 10 m were related to temperatures and the evolution of the unfrozen water content with time could be determined. As the unfrozen water content is one of the key parameter in the context of slope instabilities induced by thawing permafrost, this approach is probably the most promising for future studies, especially in combination with energy balance measurements and/or other geophysical measurements on a seasonal basis (e.g. seismics). Furthermore, resistivity measurements may be used as proxy data for the temperature evolution at different depths, as resistivity is much easier to measure than temperature, the latter depending on costly drillings of boreholes.

This work was focused on applying state-of-the-art geophysical instruments, data acquisition schemes and data processing techniques to specific mountain permafrost related questions. Future work should focus on integrated studies using several geophysical methods in combination with data from other sources. These may include borehole temperature data, drill core analyses, measurements of the water content using time-domain reflectometry (TDR) or nuclear magnetic resonance (NMR), surface energy balance measurements and data from laboratory experiments. As shown in this study results from laboratory experiments can greatly enhance the interpretability of field geophysical measurements. In addition to the geophysical methods used in this study, other methods, such as ground penetrating radar (GPR),
which showed good results in delineating the internal structure of special permafrost occurrences, have to be evaluated concerning their general applicability on mountain permafrost. Cross-borehole and surface-to-borehole measurements could be used to improve data coverage in tomographic surveys. Field measurements should be repeated on a seasonal basis to reduce the influence of short-term variations in the near subsurface. Permanent measurement arrays are capable of monitoring the ground conditions with high spatial resolution, as seen in this study. Theoretical efforts should concentrate on inversion schemes, which are able to incorporate data sets from different methods. Joint inversions of refraction seismic and resistivity data are likely to be the next step to obtain a more quantitative picture of permafrost conditions in high mountain areas.
Bibliography


160


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169
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Appendix A

Field results

As the survey results obtained within the PACE project are too numerous to be discussed in detail in the main part of this study, a complete list is given here. Note, however, that in order to interpret the results correctly, a detailed analysis should be conducted, as shown for each method in the foregoing chapters. Here, only the main results are presented. Existing publications, where more detailed information is found, are referenced.

A.1 Janssonhaugen

Around Janssonhaugen the following geophysical surveys were conducted in April 1999:

- 2 DC resistivity tomography profiles to determine changes in the permafrost distribution around the borehole (Fig. A.1).
- Electromagnetic (EM-31) mapping to determine changes in the active layer around the borehole (Fig. A.2).
- Electromagnetic (PROTEM) soundings at the drill site to determine the depth of the permafrost base (see section 5.3.4).
Figure A.1: (a) East-west and (b) north-south profile across the drillsite at Janssonhaugen. Both profiles cross at station 60.

Figure A.2: EM-31 conductivity map of the area around the borehole at Janssonhaugen. The location of the borehole is marked by a cross.

### A.2 Hiorthfjellet

The following geophysical surveys were conducted during a fieldwork campaign in April 1999:

- Four electromagnetic (EM-31) survey lines along and across the rock glacier to delineate the boundaries of the ice body within the rock glacier (Fig. A.3).
- Transient electromagnetic soundings (PROTEM) at the rock glacier tongue to determine the depth of the permafrost base (see section 5.3.4).
Figure A.3: EM-31 survey results on Hiorthfjellet rock glacier. (a) Longitudinal profile from the upper part to the rock glacier front (line A in Fig. 2.4) and (b) upper (line D), central (line C) and lower (line B) cross profile. The black vertical lines mark the geomorphological boundaries between rock glacier (RG), transition zone (T) and bedrock (B), respectively.

Figure A.4: (a) North-south profile across the PACE drill site at Tarfala (B in Fig. 2.5) and (b) east-west and (c) north-south profile across the borehole at Tarfala station (A in Fig. 2.5).

A.3 Tarfala

The following geophysical surveys were conducted in the area around the Tarfala drill site in spring 1999 and summer 2000:
• DC resistivity tomography profiles near the drill site and at Tarfala station (conducted by D. Vonder Mühll) to determine changes in the permafrost distribution (Fig. A.4).

• A 3 km long electromagnetic (EM-31) and BTS profile was recorded reaching from Tarfala station (1130 m asl) to the drill site (1550 m asl) and further to a small ridge in Tjeurolako (1430 m asl) (see Fig. 2.5). The primary goal was to determine the suitability of using EM-31 data as proxy for BTS data in regions with homogeneous surface conditions (Fig. A.5). The survey was conducted on a cold snow surface in April 1999 along a west-east oriented survey line. The results of the BTS survey (Fig. A.5b) indicate permafrost (< −3°C) along the whole survey line. Furthermore, the BTS is decreasing with increasing altitude. In comparison to the apparent conductivity results from the first EM-31 survey, there is a quite good correlation between BTS and EM-31, except for the last 400 m, where the BTS is decreasing again. However, when repeating this line in the opposite direction a totally different result was obtained (Fig. A.5c). The differences are due to instrument drift, which is becoming more pronounced the longer the survey takes (in this case 1-2 hours). One can try to (very roughly) filter this error by taking the average of the two measurements (Fig. A.5d), which still indicates a relation between BTS and conductivity. Note however, that the drift rate is not constant with time (see Fig. 5.4), depending on external factors like air temperature as well. More accurate filtering is therefore difficult to apply.

More details can be found in Hauck et al. (2001).
Figure A.5: Comparison of the bottom temperature of the snow cover (BTS) and conductivity (measured with the EM-31) along a 2km long E-leading survey line near Tarfala station, Swedish Lapland. The survey line led from the slope above Tarfala station across the drill site to a plateau near Tjeurolako (see Fig.2.5). (a) topography, (b) BTS in °C, (c) relative conductivity results from the two EM-31 surveys and (d) averaged conductivity from the two EM-31 surveys.

A.4 Juvvasshøe

Around Juvvasshøe the following geophysical surveys were conducted in August 1999:

- Site A, drill site: Characterisation of the permafrost occurrence around the PACE borehole using two 160 m long orthogonal DC resistivity tomography profiles (Fig. A.6), one 60 m refraction seismic tomography profile (Fig. A.7) and electromagnetic (EM-31) mapping (110 m x 110 m, grid size 5 m x 5 m, see section 5.3.2).
Figure A.6: (a) West-east and (b) south-north profile across the drillsite at Juvvasshøe. The borehole is located at station 78 in (a) and 74 in (b). The crossing between the profiles is at station 64 in (a) and 80 in (b).

- Site B, transition profile: Determination of the altitudinal limit of permafrost along the northern slope using a 4 km long electromagnetic (EM-31) profile reaching from the drill site (1890 m asl) down to non-permafrost areas (1300 m asl) (Fig. A.8), a 640 m long DC resistivity tomography profile (Fig. A.9) and two 60 m refraction seismic tomography profiles (combined in Fig. A.10).

- Site C, non-permafrost area: Determination of base-line resistivity and P-wave velocity values for non-permafrost areas around Juvvasshøe using a 160 m DC resistivity tomography profile and a 60 m refraction seismic tomography profile (see section 8.5.3).

Figure A.7: (a) South-north refraction seismic profile across the drill site and (b) corresponding ray paths. Station 20 and 86 correspond to station 62 and 128 in Fig. A.6b, respectively.
An integrated approach to determine the altitudinal limit of permafrost at Juvvasshøe is presented including DC resistivity tomography, refraction seismic tomography and EM-31 surveys. All three methods are used in a complementary way according to their applicability to the respective objective. The EM-31 is used to locate permafrost occurrences on a larger survey scale. For more detailed studies on a smaller survey scale DC resistivity and refraction seismic tomographies are used.

The fieldsite at Juvvasshøe is described in detail in section 2.4. Measurements of the bottom temperature of the snow cover (BTS) suggested that the permafrost distribution in this region depends mainly on altitude and that it is likely to reach altitudes around 1450 m along the northern slopes (Isaksen et al., 2001b). Measurements were conducted along a 4 km long survey line from the PACE drill site at 1894 m asl (site A in Fig. 2.6), where permafrost is present to non-permafrost areas at 1300 m asl (site C in Fig. 2.6). Along this long transition survey line, site B was investigated in detail to detect the permafrost boundary. The surface cover around site B is irregular and consists mainly on vegetation and medium size debris with air voids in between (see photo in Fig. 2.6). These medium size blocks are underlying the parts with the vegetation cover as well, so that air filled voids are expected to be present in the whole area at small depths.

Figure A.8 shows the EM-31 results along the whole survey line. The conductivity of the first part of the line is generally around 4.5 mS/m, with small variations due to patches of coarse debris with air voids in between (lower conductivity), snow patches (low values, if dry snow) and wet peaty patches (high values). Between 1500 m and 1330 m the conductivity increases to values around 5.3 mS/m, which is interpreted as due to diminishing permafrost indicating a permafrost boundary between 1500 m and 1330 m. Based on these results, the area marked by the black arrows was investigated in more detail using DC resistivity and refraction seismic tomography.

Figure A.9 shows the inversion model results for a 640 m long permafrost transition DC resistivity profile, as marked by the black arrows in Fig. A.8 (site B in Fig. 2.6). The low resistivities (1–5 kΩm) at the bottom end of the profile around 1370 m are in good agreement with the DC resistivity and refraction seismic results from the non-permafrost site shown in Fig. 8.3, meaning that permafrost cannot be present at this altitude. The high resistivities (> 30 kΩm) at the upper end of the profile are very probable due to frozen rocks and weathered material. In between, the distribution of high-resistive patches is irregular, caused by frozen bodies or even ice lenses occurring in the transition zone between permafrost and non-permafrost. However, as described in the section above, large resistivity values may be due to several causes, e.g. ice, rock or air.

To determine the cause for these high resistivity values a 140 m long refraction seismic tomography line was conducted between stations 144 and 290 of the DC resistivity line (see Fig. A.10). As seen from the low velocity values in the uppermost 4–5 m, the alternately very high and very low resistivity values between stations 144 and 304 have to be caused by air-filled cavities and unfrozen water in the active layer, respectively. Below, there are high-velocity patches corresponding to the high-resistive patches in the DC resistivity results.

Between stations 170 and 208 resistivities are > 30 kΩm and velocities are between 3000–4000 ms\(^{-1}\), indicating ice occurrences. Between stations 208 and 255 resistivities become less (< 20 kΩm) and high-velocity patches of 5000–7000 ms\(^{-1}\) are found. These must be caused by rock, frozen or unfrozen, as such high values cannot be caused by ice alone. Downslope of station 255, the velocity values decrease again, exhibiting values ≤ 3500 ms\(^{-1}\).
Figure A.8: EM-31 survey from the borehole (1894 m, site A) to an altitude of 1300 m (non-permafrost, (1300 m, site C) along the dark line in Fig. 2.6.

and corresponding to the decreasing resistivity values. The permafrost boundary may be found near station 368, where the resistivity values decrease below 5 kΩm.

The location of this permafrost transition area is in very good agreement with the findings from the EM-31 survey and with results from earlier BTS surveys (Isaksen et al., in press). Besides, the results agree well with the estimates from earlier DC resistivity soundings in this area, which determined the resistivity of frozen bedrock as 50-150 kΩm, of unfrozen bedrock as 10-25 kΩm and of unfrozen debris as 5-10 kΩm (King, 1982, Ødegård et al., 1996).

More details can be found in Hauck et al. (2000) and Isaksen et al. (2001b).
A.5 Schilthorn

Around Schilthorn a number of geophysical surveys were conducted between 1998 and 2000:

- Four 200m long DC resistivity tomography (Figs. A.11 and A.12) and three 115m long refraction seismic tomography profiles (Fig. A.13) around the drill site (August 1998) to determine changes in the permafrost distribution around the borehole.

- Electromagnetic (EM-31) mapping of the area around the drill site (October 1998) to determine changes in the active layer (see section 5.3.2).

- Seasonal DC resistivity tomography measurements approximately on a monthly basis to detect resistivity changes due to freezing and thawing using a 60m fixed electrode array (from September 1999 to September 2000, see section 4.5).

- Test study using passive microwave radiometry (January 1999) to determine the bottom temperature of the snow cover (BTS) (see section 7.4).

More details can be found in Vonder Mühll et al. (2000) and Vonder Mühll et al. (2001)
Figure A.9: Inversion results for a 640m long DC resistivity line to map the permafrost transition zone.

Figure A.10: Refraction seismic tomogram for a survey between stations 144 and 290 of the DC resistivity survey line.
Figure A.11: Tomogram for the DC resistivity survey line B across the drill site at Schilthorn (see Fig. 2.8). The location of the 14 m borehole is marked by the black vertical line.

Figure A.12: Tomograms for three DC resistivity survey lines at the drill site at Schilthorn. (a) Line C, (b) line D and (c) line E (see Fig. 2.8). The location of the 14 m borehole is marked by the black vertical line in (b).
Figure A.13: Tomogram with ray paths for the refraction seismic results from line B, line D and line F at Schilthorn (see Fig. 2.8). The location of the 14m borehole is marked by the black vertical line. Note, that line F is located on the southern side of the Schilthorn crest, with presumed non-permafrost ground conditions.

A.6 Zermatt

In July 1998 the following geophysical surveys were conducted at Stockhorn and Gandegg:

- Two 200 m long DC resistivity tomography profiles across the drill site plateau at Stockhorn to determine changes in the permafrost distribution around the borehole (Fig. A.14).
- Five 200 m long DC resistivity tomography profiles near Gandegghütte to detect and characterise the ice occurrence inside the moraine and to determine the permafrost distribution (Fig. A.15).

More details can be found in section 4.3.
Figure A.14: Tomograms for two DC resistivity survey lines across the drill site at Stockhorn. (a) northwest-southeast oriented profile and (b) southwest-northeast oriented profile. The crossing between the two lines is marked by the arrows.

Figure A.15: Tomograms for five DC resistivity survey lines near Gandegghütte. (a) Across the ice-cored moraine (line A), (b) along the ice-cored moraine (line B), (c) on the bedrock plateau (line C), (d) along a narrow sediment-filled valley and (e) across the valley, the bedrock plateau and the moraine (line E) (see Fig. 2.10). Line crossings are indicated by the respective coloured arrows.

A.7 Val Bever

In Val Bever the following geophysical surveys were conducted in July 1998:
- Six 200 m long DC resistivity tomography profiles (Fig. A.16) within the area between the two clearings (see Fig. 2.11).

- A 100 m long refraction seismic tomography profile to confirm the presence of permafrost at one representative permafrost location in the target area (see section 6.4).

- Two 200 m long electromagnetic (EM-31) lines along the DC resistivity profile lines in clearing I to determine a permafrost signal for the EM-31 and electromagnetic (EM-31) mapping of the whole study area between the two clearings (in January 1999) to determine the permafrost distribution (see section 5.3.1).

Figure A.16: Tomograms for six DC resistivity survey lines at Val Bever. (a) Longitudinal profile in clearing I (line A), (b) longitudinal profile in clearing II (line D), (c) upper cross profile in clearing I (line B), (d) lower cross profile in clearing I (line C), (e) cross profile in clearing II (line E) and (f) cross profile in the forest between clearing I and II (line F) (see Fig. 2.11). In (g) the lines C, E and F where combined and jointly inverted. Line crossings are indicated by the respective coloured arrows.

Further details can be found in Kneisel et al. (2000) and Hauck and Vonder Mühll (1999).
A.8 Murtèl

At rock glacier Murtèl a number of geophysical surveys were conducted:

- Electromagnetic surveys (EM-31 in April 1998 and GEM-300 in June 1999) along the rock glacier to determine changes in the active layer (Fig. A.17).

- Five transient electromagnetic soundings (PROTEM) along a 200 m profile from the borehole to the tongue to determine the depth of the permafrost base (June 1999) (section 5.3.4 and Fig. A.17).

- A 160 m long DC resistivity tomography profile from the borehole across the tongue of the rock glacier to the non-permafrost area in front of it (July 1998) (Fig. A.19).

The DC resistivity survey on rock glacier Murtèl yielded surprisingly good results, considering the extreme surface conditions, and is therefore presented in more detail. Electrode coupling was especially difficult as the surface consists of up to 2 m high blocks without fine material. The survey line leads from near the borehole (station -80) downslope across the tongue of the rock glacier to the non-permafrost regions below (line B, see Fig. 2.12). Figure A.19a shows the specific resistivities of the inversion model with the borehole stratigraphy determined by the core analysis added on the left hand side for comparison. The sharp gradient at the tongue (near station 0) is clearly visible, delineating the extent of the ice body. The active layer thickness is between 3 and 5 m, which coincides well with stratigraphy from the borehole. The maximum resistivities of 2 MΩm are located in the depth range between 5 and 15 m, corresponding to the massive ice found in the drill-core. Underneath, the resistivity decreases again representing the decrease in ice content as the amount of sand and coarse-grained debris become larger (c.f. section 2.8). Differences between observed and calculated apparent resistivities are less than 5% for most of the data (Fig. A.19b–d). However, due to the highly resistive environment uncertainties concerning the absolute resistivity values of the ice-body are present. From Fig. A.19d it is seen that the region in the middle of the pseudosection, corresponding to the resistivity maximum, is underestimated by about 7.5%. This agrees well with the synthetic model results of section 4.3, where the resistivity maximum was underestimated in all cases. This underestimation of highly resistive structures is probably a combined result of using a homogeneous starting model and the nonuniqueness due to the principle of equivalence. Furthermore, the relative sensitivity values shown in Fig. A.19e indicate that large relative sensitivities are only present in the uppermost model layer and in the non-permafrost area below the tongue. Throughout the rest of the model, sensitivities are small due to the high resistivity values of the ice-body. As the injected current tends to avoid high-resistive regions, the specific resistivity values of these model blocks are not well constrained. However, due to the large resistivity contrasts, the shape of the ice occurrence and the order of magnitude of the corresponding resistivity values are modelled well and are in good agreement with the borehole results.

Further details can be found in Hauck et al. (in prep.).
Figure A.17: (a) GEM-300 conductivity profile along rock glacier Murtel/Switzerland for 4 different frequencies. The topography and the location of the borehole 1987 are shown for orientation (line A, Fig. 2.12). The profile starts 50 m upslope of the borehole 1987 (station 80) and leads across the rock glacier tongue (station 400). The spike at station 80 is due to the metallic installations near the borehole 1987. (b) EM-31 profile along the same survey line.

Figure A.18: Voltage response for PROTEM measurements at rock glacier Murtel. C1–C4 correspond to the locations shown in Fig. 2.12.
A.9 Stelvio

Around Stelvio pass a number of geophysical surveys were conducted during two fieldwork campaigns in August 1998 and June 1999:

- 200 m long DC resistivity tomography profile at the PACE drill site to determine changes in the permafrost distribution around the borehole (Fig. A.20).
- 200 m long DC resistivity tomography profile (Fig. A.21a) and electromagnetic (EM-31 (Fig. A.21b) and GEM-300 (Fig. A.22)) mapping of the area occupied by the Scorluzzo Glacier during the Little Ice Age to map possible permafrost occurrences.
- 200 m long DC resistivity tomography profile along the rock glacier to determine the extent of the ice-rich permafrost body (Fig. A.23).

The permafrost mapping survey of the area around the old ski run near Stelvio pass (area C in Fig. 2.14) represents a further example of the joint DC resistivity and EM-31 approach presented in section 5.3.1. The aim was to map newly formed permafrost occurrences inside a recently deglaciated area. The permafrost occurrences were detected by a DC resistivity tomography survey (line D in Fig. 2.14) as shown in Fig. A.21a. The approach described in section 5.3.1 was used and the EM-31 results are shown in Fig. A.21b. The low-conductive regions at the top and the bottom of the slope extend horizontally indicating a strong influence of topography on the permafrost distribution. To explain the larger conductivities present in the central part of the slope, two hypotheses may be formulated. Firstly, the two low-conductive zones may present permafrost occurrences of different origins. The upper part may represent newly formed permafrost due to the recent deglaciation, whereas the presence of long lasting snow patches at the foot of the slope, which keep the ground cold for a long time, may have favoured the formation of the lower permafrost occurrence (c.f. Haeberli, 1975). Secondly, both permafrost patches could be part of a larger permafrost occurrence along the whole length of the slope, but with geological differences in subsurface conditions. In this case the conductive central part of the survey area could correspond to firm bedrock at shallow depth, with a low ice content in comparison to the resistive patches at the top and the bottom of the slope.

As a further example for the application of DC resistivity tomography on rock glaciers, Figure A.23 shows the inversion results for a survey along the rock glacier near Stelvio Pass/Italy (c.f. Fig. 2.14). In contrast to Murtèl rock glacier (Fig. A.19), whose rock glacier base is much deeper than the exploration depth of 30 m, the bottom of the ice-rich permafrost can be determined quite accurately in this case from the strong decrease in resistivity at intermediate depths. The resistivity maximum of the inversion model (Fig. A.23a), corresponding to the ice body, is thinner and has lower values (up to 500 kΩm) than in case of rock glacier Murtèl. The measured and modelled apparent resistivities are in good agreement, except for the lowermost data points (Figs. A.23b–d). The relative sensitivities shown in Fig. A.23e are larger than in the example from rock glacier Murtèl, indicating that a significant part of the electric current is penetrating underneath the ice-rich part of the rock glacier, by this determining its depth. However, the sensitivity values are still very small in this depth region, leaving some uncertainty about the exact thickness of the permafrost. This was also seen in the synthetic modelling studies shown in section 4.3 and in results from Harada et al. (2000), which indicate that DC resistivity surveys rather underestimate the permafrost thickness.

Further details can be found in Hauck et al. (in prep.).
A.10 Foscagno

The following geophysical surveys were conducted during a fieldwork campaign in June 1999:

• Electromagnetic (GEM-300) survey line across the frontal part of the rock glacier (Fig. A.24) to determine differences in the active layer between the two main lobes (see Fig. 2.16).

• Electromagnetic (GEM-300) mapping of the frontal part of the southern lobe of the rock glacier to determine changes in the active layer (see section 5.3.3).

• Transient electromagnetic soundings (PROTEM) near the drill site to determine the depth of the permafrost base (see section 5.3.4).

Further details can be found in Hauck et al. (2001).

A.11 El Veleta

A number of geophysical surveys were conducted during two fieldwork campaigns in September 1998 and March 1999 (lead by Rob McDonald, Terradat/Cardiff):

• Four 200 m long DC resistivity tomography profiles in the Corral del Veleta (September 1998) and two 100 m long DC resistivity tomography profiles at the saddle point near Los Machos peak to detect permafrost (March 1999) (Fig. A.25).

• Three refraction seismic tomography profiles in the Corral del Veleta (September 1998). The data were processed by Terradat, Cardiff and cannot be reproduced here.

More details can be found in Gomez et al. (2001).
Figure A.19: Inversion results for the data recorded on Murtèl rock glacier (line B in Fig. 2.12) for $\lambda_{\text{min}} = 0.01$. (a) Tomogram and stratigraphy obtained from borehole 1987, (b) pseudosection of observed apparent resistivity data, (c) pseudosection of predicted apparent resistivity data, (d) predicted pseudosection–observed pseudosection and (e) relative sensitivity distribution. The relative sensitivity is the sum over each column of the sensitivity matrix $J$. 
Figure A.20: Tomogram for the DC resistivity survey at the PACE drill site at Stelvio pass. The location of the 14 m borehole is marked by the black vertical line.

Figure A.21: (a) DC resistivity tomography (line D in Fig. 2.14) and (b) EM-31 permafrost mapping survey (area C in Fig. 2.14) of the old ski run near Stelvio pass. The black line in (b) marks the location of the DC resistivity line.

Figure A.22: GEM-300 mapping survey (area C in Fig. 2.14) of the old ski run with different frequencies. The values are normalised deviations from the mean. The black line marks the location of the DC resistivity line.
Figure A.23: Inversion results for the data recorded on Stelvio rock glacier (line A in Fig. 2.14) for $\lambda_{\min} = 0.01$. (a) Tomogram, (b) pseudosection of observed apparent resistivity data, (c) pseudosection of predicted apparent resistivity data, (d) predicted pseudosection–observed pseudosection and (e) relative sensitivity distribution. The relative sensitivity is the sum over each column of the sensitivity matrix $J$. 

193
Figure A.24: GEM-300 conductivity profile across rock glacier Foscagno (in SE-direction) for 4 different frequencies. The black arrows mark the extent of a firn-covered trough (see Fig. 2.16). The conductivity variations along the SE-directed survey line are due to changes in micro-topography and active layer thickness. Overall, an increase in conductivity along the line is noted. These are in good agreement with results from BTS and DC resistivity soundings (Guglielmin, 1997), which showed a thicker active layer on the SE-part compared to the NW-part.

Figure A.25: Tomograms for six DC resistivity survey lines at Corral del Veleta and Lomos de los Machos (see Fig. 2.17). (a) Line A, (b) line B, (c) line C, (d) line D, (e) line E and (f) line F. Line A and B are located in non-permafrost areas in the Corral del Veleta, line C and D were conducted along and across the rock glacier and line E and F are located on the presumed permafrost site Lomos de los Machos.
Curriculum Vitae

Christian Hauck
Date of birth: 28. May 1970, in München
Germany

1976–1979 German School Brussels, Belgium (Primary School)
1980–1984 Gymnasium Bad Homburg (Secondary School)
1984–1989 Gymnasium Icking (Secondary School)
1989 Abitur, Gymnasium Icking
1991–1994 University of München (Physics/Meteorology)
1994–1995 University of Bergen, Norway (Meteorology, Oceanography)
Oct–Dec 1995 Alfred-Wegener-Institut Bremerhaven (Theoretical Oceanography)
1996–1997 University of München (Meteorology), part-time job at DLR Oberpfaffenhofen (Atmospheric Dynamics)
1997 Diploma in Meteorology, University of München (Department for Theoretical Meteorology)
1998–2001 PhD-student at VAW, ETH Zürich, Switzerland (Glaciology Department)