The Altiplano (Central Andes)

High Conductivity Zone:
Interpretation and Modeling
of Magnetotelluric Data

Diploma Thesis

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Hiermit versichere ich,
dass ich die vorliegende Arbeit
selbständig und nur mit den
angegebenen Quellen und Hilfsmitteln
angefertigt habe.

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Abstract

In the study presented it was tried to find geological constraints for the interpretation of the high conductivity zone beneath the Altiplano, Central Andes. For this it was first necessary to investigate, which type of conduction mechanism and with it which type of conductive material could possibly cause a conductive anomaly like the observed one. For the anomaly examined in this study the conductive material was very likely to be melt. Then based on this general geological models were constructed that consist of conductive elements of adequate type. In this study dikes, extended melt zones, and shear planes were analyzed because these structures contain more or less large amounts of melt. From these models synthetic MT measurement data were produced via 3D forward calculation. These were then run with a 2D inversion routine as had been done with the originally measured data from the ANCORP profile. The inversion results were compared with the inversion result from the ANCORP profile to answer the question, which models could technically reproduce the high conductivity anomaly of the Altiplano. For the positive solutions it then was analyzed, which ones were likely to be present in the geological setting of the Central Andes. Geologic and tectonic constraints and data from other geophysical investigations were used to constrain the interpretation. Dikes were ruled out based on tectonic considerations. Large-scale magma chambers were excluded because of physical constraints. Seismic and heat flow data let the shear planes appear unlikely. The best interpretation of the high conductivity zone was possible with the concept of large-scale crustal migmatization. This approach was supported by the geologic and tectonic as well as by the geophysical data. It was demonstrated that only the combination of several different methods can produce a meaningful solution of problems in geosciences.
1 Introduction

The Andes are a more than 7000 km long mountain belt along the west coast of South America with maximum elevations of almost 7000 m. They formed in the course of the convergence of the South American plate and the oceanic plate (Aluk, Farallon, Nazca) west of it. The Altiplano (southern Bolivia) is part of the Central Andes and forms together with the Puna (northern Argentina) an extended high plateau. The Altiplano has a flat topography and an average elevation of 3700 m, whereas the Puna displays more relief with an average elevation of approximately 4500 m (Brasse et al., 2002) (Fig. 1.1). Magnetotelluric measurements (along the ANCORP profile (SFB 267)) revealed an extended zone of extremely high conductivity in the middle and lower crust beneath the Altiplano with conductivity values of 2 S/m and more. The top of the conductivity anomaly was detected at a depth of roughly 20 km. The lower boundary of the structure could not be resolved clearly. 2D-sensitivity analyses done by Schwalenberg (2000) implied a minimum thickness of the structure of 20 km and an integrated conduction (i.e. the product of conduction and thickness) of 20,000 S for the whole structure. Schilling and Partzsch (2001) estimated even 30,000 S overall conductivity and rather 30 km thickness. The EW extent of the anomaly observed in the Altiplano region is of the order of 160 – 200 km. The NS extent is not clear because the anomaly does not appear as prominent as in the ANCORP data and the profile at 22°S in all of the other MT-profiles (e.g. PICA). The lateral dimension of the whole HCZ may be considerably wider than the Altiplano region and continue far beneath the Eastern and Western Cordillera, which border on the Altiplano in the east and the west, respectively. According to Echternacht et al. (1997), the HCZ does not reach below the recent volcanic arc in the east.

Schwalenberg (2000) and Brasse et al. (2002) among many others analyzed the total conductance of the HCZ and the internal conductivity distribution. The conductivity anomaly did not show any pronounced anisotropy. Brasse et al. (2002) also tried to characterize the HCZ in terms of conduction mechanism and type of conductive material. They proposed that a large volume of partial melt probably with a layer of aqueous fluids on top could be responsible for the anomaly.
In this study I tried to interconnect petrologic and tectonic principles with geophysical data exemplarily for the ANCORP profile on the Altiplano plateau in the central Andes. For this I discussed conduction mechanisms relevant for geologic settings and introduced different theoretical models of geological settings that included regions of increased electrical conductivity. I then applied the models and their result to the real geological setting of the Altiplano and discussed the relevance of these models for the geological and tectonic setting that is characteristic for the Andes.

The fundamental concept of this work was to set up a model first and then analyze the results of the synthetic data in comparison to the results of “real” data instead of processing original data to develop a model of the geologic setting from them. The models were constructed as 3D models using the WinGlink program. For these 3D-models synthetic measurement data were calculated using a 3D forward routine of WinGLink. These data were 2D inverted and the result was compared to inversion results based on real data as have been produced by Schwalenberg (2000) and Brasse et al. (2002) among others. The aim was to investigate, which geological models can in principle reproduce the conductance pattern of the previous analyses and to suggest ways to interpret the conductivity anomaly.
Fig. 1.1 Satellite images showing the location of the Altiplano area. The stars indicate the MT stations, the lines indicate the location of the ANCORP and PICA profiles. Modified from http://www.amaps.com/spsamer.htm and https://zulu.ssc.nasa.gov/mrsid
2 Geologic development and recent setting

The enormous mountain belt of the Andes is the result of the convergence of oceanic plates (Aluk, Farallon, Nazca) subducting under an overriding continental plate (South America). During the convergence history the velocity and obliqueness of the subduction, and the dip of the subduction slab have changed considerably (Schwalenberg, 2000). This caused significant changes in the stress regimes and the deformation style of the continental margin. The volcanic front migrated eastward and produced four north-south-trending magmatic arcs, which are still prominent structures of the recent Andes (Fig. 2.1, Echternacht, 1998). The migration was caused by tectonic erosion, which permanently abraded the forearc at the continental margin (Patzwahl, 1998). The whole system of the Peru-Chile Trench including the continental margin migrated 200 km eastward since the Jurassic due to “erosive subduction”. The term “erosive subduction”

Fig. 2.1 Geomorphology map of the Central Andes. Units: CC Coastal Cordillera, LV Longitudinal Valley, PC Pre-Cordillera, PD PreAndean Depression, WC Western Cordillera, AP Altiplano, PU Puna, NH Neogene-Holocene Volcanism, EC Eastern Cordillera, SA Sub-Andean, CH Chaco (Andean Foreland) After Schwalenberg (2000) and Reutter & Goetze (1996)
describes a type of subduction zone, which is characterized by a very deep trench that has basically no sediments within the trench zone (no accretionary wedge). The trench moves towards the continent such that the continent is eroded on its margin (Patzwahl, 1998). The evolution of the Andes started in the Jurassic with the Andean Cycle. In the time between 200 and 120 Ma (Jurassic to Early Cretaceous) the westernmost magmatic arc, the Coastal Cordillera, formed by the subduction of the former Aluk plate while the marine backarc was composed of the area of the later Longitudinal Valley and Precordillera (Scheuber, 1994). The magmatism of the arc introduced mantle-derived mafic intrusive and volcanic rocks to the crust, such as basalt and basalt-andesites. Only in this arc substantial amounts of mafic crust are present (Lucassen et al., 2001). This arc represents the oldest preserved arc of the Andean Cycle (Echternacht, 1998). Several major sinistral faults (e.g. the Atacama Fault Zone) evolved parallel to the magmatic arc because of SSE – SE trending subduction of the Aluk plate (Fig. 2.2, Scheuber, 1994). Also extension perpendicular to the magmatic arc occurred. Horizontal shortening did not play an important role.

![Fig. 2.2 The configuration of the plates in the SE-Pacific during the Upper Jurassic and the Lower Cretaceous (from Scheuber, 1994, and references therein)](image)

Between 110 and 80 Ma (Middle to Late Cretaceous) the magmatic arc moved eastward to the position of the Longitudinal Valley. In this depression marine Jurassic and continental Cretaceous sediments and Cretaceous volcanic and plutonic rocks were formed. Isotope ratios of igneous rocks indicate that the magmas significantly were
contaminated with Pre-Mesozoic continental crust (Lucassen et al., 2001). According to Lucassen et al. (2001) the Pre-Mesozoic basement of the Central Andes was consolidated in the Paleozoic and consists of felsic gneisses, migmatites, granitic intrusions of Paleozoic and Mesozoic age. It has repeatedly been the source of partial melting since the late Proterozoic. Until 100 Ma ago the subduction of the Aluk plate had a strong south component. During the Middle Cretaceous the divergent margin of the Aluk and Farallon plate moved southward and thus now the Farallon plate subducted under the South American plate instead of the Aluk plate. The convergence direction of the Farallon plate had a strong northeast component, which changed the obliqueness of the convergence of the continental and the oceanic plate significantly (Fig. 2.3, Scheuber, 1994). As a consequence of this the extension of the crust ceased and first shortening occurred between 90 and 80 Ma (Peruvian Phase, Late Cretaceous) (Scheuber, 1994), and the magmatic activity paused (Patzwahl, 1998).

![Fig. 2.3](image1.png) The configuration of the plates in the SE Pacific during the Lower Cretaceous. The Aluk plate was subducted with a strong SE component, the Farallon plate was subducted with a NE component (from Scheuber, 1994, and references therein)

The Precordillera developed between 72 and 35 Ma. Like in the Longitudinal Valley the igneous rocks show strong influence of continental crust (Lucassen et al., 2001). The time of the Maastricht-Lower Eocene was characterized by tectonic quiescence and slight extension due to a very low rate of convergence. Extension and formation of half grabens were reported for the Altiplano for the time range of Eocene to Oligocene (Elger, 2003). During the Upper Eocene (approximately 48 Ma) the convergence rate increased
considerably from less than 3 cm/yr to more than 10 cm/yr (Patzwahl, 1998) and intense shortening caused the formation of extended anticlines parallel to the magmatic arcs and of dextral fault systems due to the north component of the subduction of the Farallon plate (e.g. Falla Oeste) (Fig. 2.4, Scheuber, 1994; Echternacht; 1998).

The recent plate configuration of the Nazca plate that subducts under the South American Plate and the Cocos plate that subducts under Central America developed during the Oligocene (27 Ma) from the Farallon plate (Scheuber, 1994). The convergence rate of the Nazca plate was previously relatively low, but has increased to 9 cm/yr since 20 Ma (Schwalenberg, 2000) (Fig. 2.5).

The Altiplano region has experienced two phases of intense shortening and uplift since approximately 29 Ma (Late Oligocene) (Scheuber, 1994). During the first phase (before 10 Ma) the crustal shortening was distributed between the magmatic arc and the Principal Frontal Thrust, in the Eastern Cordillera and the Altiplano. In this episode the

Fig. 2.4  The configuration of the plates in the SE Pacific during the Eocene. (from Scheuber, 1994, and references therein)

regions of the present Sub-Andean fold-thrust belt and the actual foreland basin together formed a broad shallow foreland basin. The eastern Cordillera deformed internally and subsidence within the Altiplano occurred, probably associated with west-vergent backthrusting of the Eastern Cordillera (Baby et al., 1997; Gubbels et al., 1993). The
crustal shortening in the plateau region is assumed to be distributed homogeneously within the upper and lower crust by pure shear mechanism until approximately 10 Ma. Estimates of the amount of crustal shortening vary widely between 100-155 km (Isacks, 1988) and 250 km (Lamb et al., 1997). Later (after 10 Ma) the deformation mechanism turned into simple shear mode when the deformation migrated eastward into the Sub-Andean range (Allmendinger & Gubbels, 1996). Tectonic quiescence in the Altiplano between 25 and 17 Ma seems to correlate with a period of intense shortening in the Eastern Cordillera (Müller et al., 2002).

Approximately 17 Ma ago bivergent thrust systems formed in the Western and Central Altiplano with locally high strain rates. The whole plateau was influenced by strong compression (Elger, 2003). The Western Cordillera is the most recent magmatic arc and has developed between 18 and 10 Ma (Miocene to Holocene) (Scheuber, 1994). Andesitic volcanic rocks and rhyolitic ignimbrites are the common rock types. Especially during the Neogene the deformation affected the magmatic arc as well as the backarc region (Altiplano-Puna Plateau, Eastern Cordillera and Sub-Andean). Around 10 Ma (Late Miocene), during the second phase of deformation, the region between the Pre-Andean Depression and the Eastern Cordillera has been shortened intensely and the Altiplano-Puna Plateau was lifted to higher elevations, but it was not deformed internally anymore (Elger, 2003). In this time the large scale process of underthrusting of the cold stiff Brazilian shield began and thickened the lithosphere beneath the eastern Altiplano. The upper crust of the Eastern Cordillera overthrust the Brazilian shield on a mid crustal detachment, which is assumed by Gubbels et al. (1993) to be the westward continuation of the Sub-Andean master décollement. The crust below this detachment within the Eastern Cordillera is believed to have deformed ductily. According to Allmendinger & Gubbels (1996) the deformation beneath the Altiplano changed into simple shear mode while the deformation beneath the Puna remained in the pure shear mode.

The uplift of the Altiplano was mainly based on deformation along a west-vergent thrust system at the western margin of the Altiplano plateau (Victor, 2000). There was no clear deformation front migrating eastward through the Altiplano plateau (Elger 2003). Geometric modeling done by Victor (2000) indicated the existence of a transcrustal ramp reaching down to 30 to 35 km of depth. Elger (2003) reported that the eastern Altiplano is
characterized by west-vergent thin-skinned structures and shallow to moderately dipping faults, which appear as widely spaced ramp anticlines with blind thrusts. A bivergent thrust system and NNE/SSW trending faults form the central Altiplano. In the western

![Map of the plates of the SE-Pacific](image)

Fig. 2.5 The recent configuration of the plates of the SE-Pacific (from Scheuber, 1994, and references therein). The numbers denote the rate of relative motion of the plates (mm/yr) (Kearey & Vine, 1996)

part of the central Altiplano west-vergent fault propagation faults occur that continue into the western Altiplano. It is assumed that a lower crustal weak zone was an a-priori feature of the Andean plateau, because of which the plateau has been flat in topography since its formation. This is also implied by the diffuse deformation pattern observed in the Altiplano (Elger, 2003). Between 10 and 2 Ma large amounts of ignimbrite were erupted at the Altiplano and Puna plateaux and formed the Altiplano-Puna Volcanic Complex (APVC) (Babeyko et al., 2002). The isotope ratios of the ignimbrites indicate a very strong involvement of the crust: 50 to 100% of crustal material were estimated for the ignimbrites’ composition (Lucassen et al., 2001, and references therein). The crustal thickness has increased tectonically since the Miocene and now reaches more than 60 km beneath the eastern Altiplano (Wigger et al., 1994). According to Haschke & Günther (2003) the crustal thickening was mainly based on tectonic processes and only to a minor part on magmatic thickening. Models run by Wdowinski & Bock (1994a,b) implied that
the plateau crust thickened until the plateau reached a critical elevation. Then it resisted further accumulation of shortening and the center of deformation migrated eastward. The Sub-Andean has evolved since around 3 Ma due to westward underthrusting of the Brazilian Shield beneath the Andean orogen and the easterly directed subduction of the Nazca Plate. Since the Miocene also the forearc has experienced new deformation such as reactivation of the Atacama fault and rising of the Coastal Cordillera (Scheuber 1994).

During the Andean orogeny in Oligocene till Pliocene time a large scale orocline formed in the Central Andes with its hinge between 18 and 20°S that is called the Bolivian orocline (e.g. Müller, et al. 2002; Roperch et al., 2000). According to many paleomagnetic analyzes (e.g. Rousse et al., 2003) the northern limb of the orocline rotated counterclockwise; the southern limb rotated clockwise, whereas only little rotation was obtained for the bend (Kley, 1999). The details of kinematics are still under discussion. Generally, differential tectonic shortening along the strike is assumed to have caused the rotations and also may have initiated or enhanced shear in the limb regions (Müller et al., 2002).

**Summary**

The Andes are the result of the convergence of an oceanic plate and a continental plate that started in the Jurassic. The deformation of the overriding continental plate was characterized by successive eastward migration of the volcanic front due to tectonic erosion. Convergence rate and angle, and subduction dip changed in the course of subduction. This had significant influence on the stress regime and deformation style. The Altiplano region has been shortened and uplifted in two steps since approximately 29 Ma. Especially around 10 Ma, in the second stage of deformation the Altiplano-Puna region was uplifted intensely. In this time the Brazilian shield began to underthrust the Andes from the east. A large scale bend called the Bolivian orocline between 18 and 20°S formed in the Central Andes during the Andean orogeny. Since the Miocene the crustal thickness has increased considerably and reaches now more than 60 km beneath the Altiplano. This thickness is thought to be mainly based on tectonic processes and only to a minor part based on magmatic thickening.
3 Conduction processes within the crust

Most of the upper earth’s crust is relatively low conductive. Precambrian shields are least conductive ($10^{-3}$-$10^{-5}$ S/m), regions of tectonic activity during the Phanerozoic show somewhat higher conductivity ($10^{-2}$-$10^{-3}$ S/m) (Haak & Hutton, 1986; Nesbitt, 1993). Even though Precambrian shields tend to contain fluids of higher salinity than the others, the higher fracture density of Phanerozoic crustal structures over-compensate the difference.

The conductivity is one of the physical properties of geologic materials with the widest range. It can be as high as $10^6$ to $10^8$ S/m for pure metals or as low as $10^{-16}$ S/m for diamond (Fig. 3.1). Minerals and rocks usually have conductivities between $10^{-2}$ and $10^{-5}$ S/m based on in-situ measurements and up to $10^{-16}$ S/m in laboratory measurements (Beblo, 1982).

![Fig. 3.1 Comparison of specific conductivity of material in S/m. Modified from Schilling (2003) and Beblo (1982).](image)

It is necessary to first examine the micro-scale features of conductive systems because conductivity takes place at grain scale and is based on the smallest members of a geologic setting: on electrons and ions. The conductance mechanism is an important aspect to decide, which type of geologic structure principally can explain a crustal conductivity anomaly.
3.1 Principles of conduction

Several conduction processes have been suggested to explain the high conductivity zones of the Andean crust. Before these different concepts are evaluated, a few basic comments and definitions will be given.

Conductivity of a system is the sum of all simultaneously occurring transport processes. Only one or very few mechanisms are assumed to predominate within a certain temperature range because they are much faster than the other processes (Partzsch, 1998, Gaillard, 2004). An electrical field applied to a conductive system causes a movement of charged particles that is a current flow:

\[ j = \sigma \cdot E = \frac{1}{\rho} \cdot E \]  

\( j \) is the current density vector, \( E \) the electrical field, \( \sigma \) the specific conductivity and \( \rho \) the specific resistivity with \( \rho = \frac{1}{\sigma} \). (\( \sigma \) is assumed not to depend on orientation (isotropy)\(^1\), i.e. \( j \parallel E \)). This equation describes conductivity processes in metals as well as in melts and saline fluids. The electrical conductivity defines how much charge (I) can be transported through a cylindrical body with the length \( l \) and the area \( A \) and with the given potential difference \( U \) (Partzsch, 1998):

\[ \sigma = \frac{I}{U} \cdot \frac{l}{A} \]  

(3.2)

In that way it reveals how many charge carriers in a system can be activated by an electrical gradient, and how mobile they are (Gaillard, 2004).

The crust is composed of a huge variety of rocks, each of which usually contains a certain amount of different minerals. These minerals can be separated into three basic groups of conductors (Labendz, 1999). Minerals with metallic lustre, ores, graphite and

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\(^1\) Anisotropy can occur for example when elongate minerals in a preferred orientation are present. It can be observed with measurements in two perpendicular directions. Under utilization of frequency dependent methods such as impedance spectroscopy both of the polarization directions diverge in resistivity and phase. In larger scale, in which complete structures rather than minerals cause anisotropy magnetotellurics reveals anisotropy (Labendz, 1999).
graphitic shales are considered as good conductors. Feldspar, quartz, halite, and sulfur are isolators. Silicate minerals and oxides are semiconductors. Based on this classification most of the earth’s crust is weakly conductive, and certain regions of increased conductance occur as conductive anomalies.

The conductivity of a system depends on the specific conduction mechanism and is sensitive to several factors (Partzsch, 1998; Gaillard, 2004):

- chemical composition of the rock/melt, mineral content, water content
- nature of the constituting phase (solid, liquid)
- texture of the rock
- condition of grain boundaries
- pore shape geometry
- kind and amount of fluid
- kind and amount of salts dissolved in saline fluids
- oxygen fugacity
- temperature
- pressure

3.2 Conductivity based on conductive minerals

Metallic ore minerals such as iron and copper sulphides or magnetite are minerals with a significant conductivity. Conductivity values of roughly $10^{-2}$ to more than $10^3$ S/m have been reported (Beblo, 1982). The conductivity is based on the existence of valence electrons, which are not bound in the chemical ion binding and on overlapping valence and conduction bands. The electrons easily can be separated from the atom and serve as charge carriers. If the minerals form ore bodies or are disseminated throughout a larger volume (e.g. as grain boundary conductor), they can cause local electrical anomalies (Li
et al., 2003). To do so they have to form interconnected networks throughout the volume. If they are dissipated they are not detectable as conductors. Since these anomalies are usually spatially restricted to certain geologic settings that cause the formation of the ore concentration, they are not likely to cause regional scale conductivity anomalies as they have been observed in the Andean system. ²

### 3.3 Conductivity based on graphite

Analyses of conductive zones, e.g. in the KTB, showed that layers of a connected network of graphite can cause local conductive anomalies within the middle crust with conductivities of roughly 1 S/m (ELEKTB-Group, 1997). Graphite internally forms layers, in which each carbon atom has three neighboring carbon atoms around it in a triangular arrangement. They are interconnected with these neighbors via electrons in a strong $\sigma$-bound whereas the fourth electron ($\pi$-electron) is delocalized in the p-orbitals, which covers the whole layer (Fig. 3.2).

![Graphite structure](image)

**Fig 3.2** The structure of graphite causes a strong anisotropy of the conductivity.

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² Ore bodies additionally cause specific gravity and magnetic anomalies that have not been observed in the Andes in connection with high conductivity zones.
Therefore the conductivity of graphite is very good parallel to the layers. Conductivity values range between $10^{-2}$ and $10^2$ S/m, in laboratory measurements even $10^4$ S/m have been obtained (Beblo, 1982). Perpendicular to the layers the conductivity is $10^4$ times weaker (Riedel, 1994).

Graphite layers on fossil shear zones turned out to be a fairly common feature. Graphite can for example be introduced into the crust by subduction of sediments and metasediments, e.g. graphite bearing black shale (Haak & Hutton, 1986; Brasse et al., 2002). The earth mantle is not likely to be a source of carbon because of extremely low flow rates of CO$_2$ (Rath, 2000). Isotope analyses of graphite originally located in great depths indicate mainly biogenetic origin or derivation from carbonate rocks for many of the carbon sources. Increase of temperature (>400°C) can transform black shale into graphite layers (Rath, 2000). The carbon can be precipitated as interconnected films of only a few millimeters of thickness along shear planes within a cataclasite. The term “cataclasite” indicates a deformation regime, in which pressure-sensitive fracturing, grain boundary sliding and rotation of rigid rock particles are the main deformation processes during the establishment of graphitized shear zones (ELEKTB-Group, 1997). The graphite on the shear planes could have been concentrated by pressure solution and other fluid-controlled processes, or by fluids that carried carbon onto shear planes when they invaded the system. Furthermore, metamorphosis of carbon rich sediments and destabilization of graphite at temperatures above 600°C can mobilize carbon, which forms C-H-O-gases and redistributes carbon within the lower crust (Rath, 2000). Extended graphite films could evolve when CO$_2$ and CH$_4$ bearing fluids interact with reducing minerals such as biotite or magnetite (Frost & Bucher, 1994 and references therein).

The observed high-conductivity zone below the Altiplano is of vast extent and located within the crystalline basement. The ELEKBT-Group (1997) has shown that graphite films can form at the depth of the brittle-ductile transition of the matrix. However, graphite only forms a highly conductive layer when its crystals are arranged in a high-degree crystalline order.
A stable interconnected network of graphite of the observed size could only be associated with one or a few very large shear zones of underthrusting settings. Tectonically large scale thrusting is very possible, since the Andes including the Altiplano plateau were and are subject to intense shortening. Especially the Brazilian Shield has underthrust the Andes from the east since several million years and easily could have formed large scale shear systems. Based on the fact that hydrocarbon reservoirs have been explored in the region the input of carbon into the shear system seems possible, too.

The possibility of conductivity based on graphite or amorphous carbon cannot completely be ruled out at this point.

### 3.4 Conductivity based on presence of melt

In regions of high tectonic activity strong geothermal activity is observed as well. High geothermal activity may lead to partial melting of the rocks. Thus, melts could at least contribute to the high conductivity of the middle and lower crust beneath the Altiplano. Furthermore, there are geophysical and petrologic evidences that magma source regions with temperatures of at least temporally more than 800°C are located within a laterally extensive zone with its top at 20 to 25 km of depth in the crust (Babeyko et al., 2002).

Questions to answer are therefore how and where the melt is formed and transported through the crust, how it is distributed within the crust by migration and how it is stabilized in the storage site. The partial melts can contribute to the increased conductivity in the middle crust (van Ngoc et al., 1986).

**The process of melting and melt migration**

Melting of rocks occurs because of changes in the pressure and/or temperature conditions, and because of variations in the water content of the system. Decreased
pressure lowers the melting temperature by several hundreds of degrees (decompression melting due to rapid uplift and unroofing).

The porosity in the middle and lower crust is very small (0.1% or less (Clemens & Vielzeuf, 1987)) so that the amount of free fluid in the system must be very small. This little fluid should be rapidly consumed by any melting process where initially fluid is present because of the relatively high solubility of water in silicic melt (>2 wt%). Thus, the process would switch to fluid-absent conditions where larger amounts of melt only can be produced if a significant amount of additional heat (e.g. from the mantle) is provided to reach a temperature of at least 850°C depending on source rock and pressure. Fusion under these conditions is based on breakdown of biotite, mica, and amphibole, which produces water-undersaturated melts (Clemens & Vielzeuf, 1987). Assuming a melt system beneath the Altiplano the conditions would very likely differ from the scenario given above in that it will contain a significant amount of water. The oceanic plate subducting under the South American plate carries water bound chemically in minerals and physically in the plate’s structure. The farther down the oceanic plate is subducted the more it is heated, and eventually the water is released from the plate. The water rises through the crust above the slab and increases the water content of the rock by mineral reactions that form hydrated minerals (Frank Schilling and Robert Trumbull, 2005, GFZ, personal communication). Water furthermore decreases the melting temperature of the rock, and especially in wet rocks the increase of pressure reduces the melting temperature of the rocks significantly. Thus, the input of large amounts of water also may cause large scale melting in the lower crust. However, in wet rocks there can only as much melt be formed at temperatures below the solidus of the dry rock as water is available to saturate the melt. Further melting occurs at considerably higher temperatures (Partzsch, 1998).

Partial melting usually starts at grain corners where the maximum number of reactant phases is involved. In grain corners four different grains can meet, while in grain edges still three grains come together (Philpotts, 1990). The combination of different phases, i.e. minerals, lowers the solidus temperature of the local system (Partzsch, 1998). The melt first forms little melt pockets at the grain corners and grain edges. Several experiments (e.g. Partzsch et al., 2000) demonstrated that shortly after onset of melting a
small portion of the melt eventually escapes from the pockets to slowly form very thin films along part of the grain faces. The melt appears to be quite mobile and was observed at all types of grain boundaries, even at boundaries between minerals of the same type that could not have formed melt yet. In experiments at hydrostatic conditions van der Molen & Paterson (1979) found that the melt films were located at randomly oriented grain faces and cracks in minerals. With increasing amount of melt the spots of melting are less restricted (Philpotts, 1990).

As the fusion continues, the melt pockets spread out, and at a certain melt fraction the melt starts to interconnect along the grain edges and grain faces forming melt tubules and films (Sawyer, 1994). Preferably grain boundaries of low tortuosity and irregularity become channels, into which other more irregular or with lower melt volume passing through them drain melt (Sawyer, 2001). Further melting widens and lengthens favorably the channels that are oriented suitably with respect to the local tectonic stresses or the host fabric. This state of melting requires a minimum melt fraction between 8 vol% (Vigneresse et al., 1996) and approximately 15 vol% (dry melts; Schilling & Partzsch, 2001). Vigneresse et al. (1996) labeled this liquid percolation threshold (LPT). The melt can only be transported over very short distances due to deformation that produces dilatant sinks (Vigneresse et al., 1996), due to pressure gradients caused by the volume increases of melting (e.g. due to muscovite dehydration), or by adhesion effects relative to the respective minerals, which cause the fissures and cleavage cracks to become filled with melt (Büsch et al., 1974; Holyoke & Rushmer, 2002). As the melt fraction increases the rock may disaggregate. The pressure will then be distributed more evenly throughout the rock. That means that local pressure gradients probably do not act as force causing melt segregation and buoyancy plays a more important role (Holyoke & Rushmer, 2002).

Melt tends to take a shape that minimizes the interfacial free energy with the neighboring grains (textural equilibration) in the moment when it is formed in a grain corner. In ideal sources, that is in equilibrated systems with no anisotropic surface free energy, grains form faces of constant mean curvature (Laporte & Watson, 1995). In the beginning of the melting process areas of high curvature (grain edges) and areas of low curvature (grain faces) exist (Bulau et al., 1979). The grains adjust their crystal-liquid interfaces and angles of contact with the liquid by dissolving melt from high-curvature
(high-activity) edges and precipitating on low-curvature (low-energy) faces until the
system has reached a constant curvature and no sharp corners on the crystalline grains in
contact with the liquid exist (Bulau et al., 1979; Philpotts, 1990). Ideally this process
should squeeze the melt out of flat intergranular faces.

The result of the described arrangement of the melt is a characteristic dihedral angle
(wetting angle), $\Theta$, which is the contact angle between melt and two adjoining crystalline
grains (Waff & Faul, 1992). The wetting angle depends on the wetting behavior of both,
liquid and solid, and is a measure of the surface tension. Wetting is the ability of a liquid
to form a boundary surface with solids. Small wetting angles are caused by a large
wetting tendency (i.e. small surface tension). The wetting behavior of a liquid depends on
its chemical composition and on the chemical and structural properties of the wetted
mineral. The wetting angle is determined by the ratio of solid-solid ($\gamma_{ss}$) and solid-liquid
($\gamma_{sl}$) interfacial energy on each of the different interfaces (Schilling et al., 1997) (Fig,
3.3):

$$2 \cos \frac{\Theta}{2} = \frac{\gamma_{ss}}{\gamma_{sl}}$$  \hspace{1cm} (3.3)

Fig 3.3 Schematic sketch of a triple junction of three solid grains including melt. See
text for explanation. Modified from Cooper & Kohlstedt (1986)

$\gamma_{ss}$ depends on the degree of mismatches of the two bordering lattices along the grain
boundary and on the intrinsic anisotropy of $\gamma_{ss}$ associated with the crystalline structure of
the silicates (Laporte & Watson, 1995). The degree of anisotropy in $\gamma_{ss}$ depends on the
symmetry and complexity of the crystalline structure. It is low for quartz and feldspar,
but relatively high for clinopyroxene, hornblende, biotite and others. $\gamma_{sl}$ results from boundary mismatches along the crystal-melt interface. The crucial factor is here the orientation of the interface relative to the crystallographic lattice: $\gamma_{sl}$ is low at faces, which are parallel to the reticular planes of dense atomic packing (low (hkl) indices) (Laporte & Watson, 1995).

The wetting angle determines how the melt is distributed in the partially molten rock and how easily the melt can segregate to form magma bodies. Melts with high wetting tendency develop an interconnected network with $\Theta \leq 60^\circ$. The melt preferably extends along grain edges forming channels and connects the neighboring melt filled grain corners with each other (Best & Christiansen, 2001). This way the melt forms a continuous phase throughout the partially molten rock already at low melt fractions (Bagdassarov et al., 2003). Because of this structure the melt can segregate already at very low melt fractions. As long as the melt does not form extended layers along the grain faces the grains remain in contact until a sufficient amount of melt has formed that disaggregates the rock. Theoretically the crystal-crystal interfaces could also be wetted if $\Theta$ is zero. For partial melting in the crust mainly low wetting angles of 10 to 45° are relevant (Laporte & Watson, 1995). At $\Theta \geq 60^\circ$ the melt forms isolated pockets at the grain corners. These globules only connect at high degrees of partial melting (Best & Christiansen, 2001) (Fig 3.4).

Consider for example a melt in contact with a plagioclase grain. A silica-rich melt is highly polymerized, having a structure similar to that of the plagioclase. The degree of mismatch and therefore the surface free energy is small. The meniscus of a melt droplet that forms in a grain corner should be concave (curved negatively). In this case $\Theta$ is smaller than 60°. An iron-rich mafic melt is less polymerized, so that its structure shows a higher degree of mismatch with the crystallographic structure of the plagioclase. Thus, the surface free energy is higher, and the wetted surface tends to be smaller. The melt droplet forms a convexly shaped meniscus (positive curvature), and $\Theta$ is larger than 60°.

With regard to the pattern of melt distribution there appear two extreme conditions for a system with a certain amount of fluid: first, there can theoretically exist a perfectly (100%) interconnected network of melt (Hashin-Shtrikman upper bound), and second there can exist a completely non-connected fluid (Hashin-Shtrikman lower bound). These
extreme settings have severe consequences for the physical properties of the partially molten rock as will be discussed in a later chapter.

Fig. 3.4 Melt equilibrium geometry depending on the wetting angle. Upper line: view along melt channels (three grain junctions), middle line: view at grain corners (four grain junctions), lower line: melt distribution in a partially molten system. Modified from Best & Christiansen (2001), Walte et. al. (2003), Toramaru & Fujii, 1986.

All these configurations relating Θ with melt distribution are based on several assumptions. The first assumption is that the system is equilibrated. Equilibrium means that a system has reached the condition of lowest energy content possible in a certain setting. Only if parameters that determine the setting change, the system would alter its own conditions to again reach the state of lowest energy content possible. Especially in geological systems the process of attaining the state of equilibrium can cover very long time spans. Thus, it is difficult to estimate how close to equilibrium natural systems are, since measurements monitoring a system only cover very short time frames compared to the system’s development. Further the scale observed may play a role because small flow
motions of melt along grain boundaries do not necessarily alter the state of a whole magma zone.

The second assumption is that the system is isotropic and does not show any preferred crystallographic orientation. In reality this is not the case. That is why experimental systems often show deviations from the $60^\circ$ pattern described above (Best & Christiansen, 2001). All mineral grains have a certain crystallographic anisotropy. The melt wets the diverse grain faces differently easily. For example a mafic melt can wet more easily the (011) face of a hornblende grain than the (010) face (Philpotts, 1990). Laporte & Watson (1995) and Daines & Kohlstedt (1997) among others observed that at hydrostatic stress conditions flat crystal faces in contact with melt pockets rather than smoothly curved boundaries develop. They explained the deviations from the predicted geometry of the isotropic model with the strong influence of crystalline anisotropy, which can significantly alter geometrical features of the melt network. Sufficiently high anisotropy of $\gamma_{sl}$ leads to planar rational type solid-melt interfaces. Obviously, curved interfaces are characteristic for rock-melt systems with low anisotropy of $\gamma_{sl}$. Anisotropic fabrics can increase the permeability of the rock considerably.

The wetting angle of a partially molten rock significantly depends on the amount of water included in the system. Marchildon & Brown (2002) report that junctions of orthopyroxene grains with basaltic melt usually have a dihedral angle of approximately $70^\circ$. When the melt is saturated with water the dihedral angle drops to $52^\circ$. This change is significant because the interconnection state of the melt can be altered drastically. Most of the crustal rock compositions forms water bearing silicate melt in contact with quartz or feldspar grains and should therefore lead to interconnected melt networks (Schilling et al., 1997). Information and assumptions about the polymineralic source including data of water content, the anisotropy of interfacial energies, the presence of crystallographically preferred orientations and small-scale heterogeneities of the source are necessary for adequate modeling of the distribution of melt in the crust (Laporte & Watson, 1995).

Partial melts must form a widely continuous, i.e. nearly completely interconnected network throughout the source rock, before it can start to move over large distances buoyantly or driven by pressure gradients (Vigneresse et al., 1996; Schilling & Partzsch,
This occurs as the melt fraction increases to values of 20 to 25 vol%, which Vigneresse et al. (1996) call melt escape threshold (MET).

The permeability of the source rock and the viscosities of the rising melt and of the surrounding rock strongly influence the velocity and the pathway of melt migration and thus determine the rate of accumulation of melt. The permeability depends on the wetting angles in the partially molten rock and on the anisotropy of the fabric.

At wetting angles greater than 60° there is effectively no permeability until a critical melt fraction is reached where the separated melt globules get in touch with neighboring melt globules. For small wetting angles of less than 60° the permeability of the partially molten rock is fairly high already for small amounts of melt because the melt forms an interconnected network. Already small increases in the melt fraction can cause large increases in the permeability of the system (Philpotts, 1990). Higher permeability leads to increased flow velocity of the melt through the channels. In partially molten rocks with moderate melt content (20 vol%), which are characterized by a preferred elongation direction of the solid matrix grains the melt will be localized preferentially along the grain boundaries sub-parallel to this orientation. Thus, the whole melt network will be polarized to this preferred orientation, and therefore increased flow permeability along this direction will occur (Marchildon & Brown, 2002). A pronounced anisotropy in the melt permeability can be the result of differential stress. The long axes of the melt pockets are oriented predominantly at angles of 20° from the axis of maximum principal stress $\sigma_1$. This characteristic distribution of melt is called Melt Preferred Orientation (MPO) (Zimmerman et al., 1999) and can be controlled by surface tension and strain as well.

If there is a driving force and a site where the melt can move (dilatational structures may serve as sinks) the melt moves through the pathways along the grain boundaries and, depending on the fusion-generated volume change, through self-generated intergrain and intragrain fractures to the location of accumulation (Best & Christiansen, 2001, Marchildon & Brown, 2002). The principal driving force is the buoyancy of the less dense melt. The magma body starts to rise buoyantly through the overlying crustal sequences as soon as it exceeds a certain critical mass. Melt can also be extracted from the source rock when there is a pressure gradient that could be caused by the volume
expansion of the melt or by deformation processes that affect the whole system. For example granitic magmas typically form by dehydration melting under volume increase. This expansion enhances the pressure in the pores so that the melt is pushed out of the source along melt channels between the grains or through locally formed fractures. The pressure gradient forces the melt to move at grain scale according to the local stress conditions. At layer scale the melt flows from the most to the least competent layers. This process is called deformation assisted melt segregation (Sawyer, 1994). According to Petford et al. (2000), melt of a wide range of viscosity can be moved from the source to a local site of dilation in fairly short time scales of $10^{-1}$ to $10^4$ years. Applied stresses may force the melt out of the source earlier than the melt would leave the source solely based on buoyancy.

Under non-hydrostatic conditions applied shear stress can cause migration of the melt in the system. Especially in rocks with relatively low permeability the melt segregates from initially homogeneous distribution into melt-rich layers when they are sheared. This can significantly increase the interconnectivity of the melt because previously isolated melt pockets become smeared out and interconnect. This process might be the reason, why melt observed in crustal environment tends to form interconnected networks already at less than 20 vol% melt fraction. Over large regions 20 vol% melt may be a good estimate for conducting interconnected networks. Instead of being distributed throughout the whole rock the shear deformation will be concentrated along weak, melt-filled shear bands (Rosenberg, 2001, Holtzman et al., 2003), which are oriented at approximately 20° to the shear plane and separate melt-depleted zones in between. As the shear continues the bands may form anastomosing networks with larger shear bands at high angles connected by smaller bands at lower angles (Fig. 3.5; Holtzman et al., 2003). shear bands are means of efficient extraction and transport of melt from the source (Holyoke & Rushmer, 2002). The melt moves according to the pressure gradients (i.e. gradient of the mean stress) from stronger to weaker layers within the deforming rocks. A certain amount of melt (roughly 3 %) remains along the grain edges to minimize the surface free energy of the system. This melt can not be removed from the rock (Philpotts, 1990).
Melt-based conductivity

The conductivity of silicate melt is approximately two to three orders of magnitude higher than the conductivity of rocks close to their solidus, whereas the conductivity of different silicate melts varies only half an order of magnitude (van Ngoc et al., 1986). 1 S/m is a reasonable estimate for the conductivity of the complete melt body as has turned out from MT measurements (see below).

The conductivity of a melt-bearing system is apart from temperature and pressure determined by melt composition, amount of structurally bound and free water, oxygen fugacity, grain size, presence of cracks, and occurring preferred orientation of grains and melt (Presnall et al., 1972; Li et al., 2003). In partially molten systems the electrical conductivity is controlled by both components the solid and the liquid. Therefore at least a two-component system has to be analyzed. The solid contributes only little to the bulk conductivity so that the conductivity of the liquid dominates the system (Gaillard, 2004). Dry rocks are effectively semiconductors, that is, as the temperature increases the conductivity of the material increases, too. Since it is an electron-based mechanism, the
number and mobility of the electrons and the electron holes control the conductivity (Schilling et al., 1997). The electronic conduction is thermally activated and described by the Arrhenius-type equation

\[ \sigma = \sigma_0 \cdot e^{-E_A / kT} \]  \hspace{1cm} (3.4)

where \( \sigma_0 \) is a constant that describes the theoretical conductivity for infinitely high temperatures (i.e. a reference conductivity), and \( E_A \) is the activation energy, i.e. the energy necessary to enable the charge carriers to overcome the energy barrier between two lattice positions (Partzsch, 1998). However, solid conduction in the bulk silicate phase due to increased temperature cannot be the conductive mechanism that produces the conductivity measured in the crust. Only at more than 1000°C the resistivity would reach values smaller than 10 Ωm (van Ngoc et al., 1986). This is definitely to high a temperature for solid rocks at a depth of approximately 20 km, since dry basaltic rocks melt already between 1000°C and 1200°C, and more silica-rich wet rocks even start to melt at 650°C to 800°C at presence of water. Besides this, cracks common in grains would considerably lower the conductivity of the system because they are barriers for the charge transport (Partzsch et al., 2000).

As the melting starts, the conduction mechanism changes to ion-based conduction, which is mainly caused by alkaline ions and their interactions. The electrical conductivity of silicate melts strongly depends on temperature, but only weakly on pressure (Partzsch et al., 2000). The conductivity is described by the Arrhenius-type equation for ionic conductivity (Schilling et al., 1997):

\[ \sigma_i = \frac{\sigma_{0i}}{T} \cdot e^{-E_{A_i} / kT} \]  \hspace{1cm} (3.5)

where \( \sigma_{0i} \) is a constant, and \( E_{A_i} \) is the activation energy for ionic transport. \( 1/T \) describes the temperature dependence of the ion mobility.

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3The conductivity of andesites is most pressure dependent, whereas the conductivity of tholeiites shows lowest dependency. More Si-rich melts show intermediate pressure dependence of electrical conductivity.
Silicic melts are formed by different types of ions: The Si$^{4+}$ and Al$^{3+}$ cations are network formers, which are linked by bridging oxygens and build the interconnected networks of non-periodic SiO$_4^-$ and AlO$_4^-$ tetrahedrons of the polymers. Fe$^{3+}$ ions are also strongly bound to bridging oxygens. The large anionic polymers made of these tetrahedrons are assumed not to play a significant role in the conduction process (Waff & Weill, 1975). The cations Ca$^{2+}$ and Mg$^{2+}$ as well as dissolved H$_2$O, F$^-$, and Fe$^{2+}$ are network modifying ions. They are more weakly bounded to non-bridging oxygens, which are located in non-tetrahedral bounding arrangements (Best & Christiansen, 2001). Alkaline (e.g. Na$^+$, K$^+$) and alkaline earth ions act as the charge carriers moving within the gaps of the network (Partzsch, 1998). A certain positive correlation between conductivity and alkaline ion content and a negative correlation between conductivity and silica content was observed (Waff & Weill, 1975). An increase in silica content increases the viscosity of the melt due to the higher degree of polymerization and therefore decreases the conductivity. Increased temperature and, to a lower degree, increased pressure decrease the degree of polymerization and therefore decrease the viscosity of the melt. The most mobile species of charge carriers especially in viscous acid and intermediate melts are the small sodium and lithium ions. Since lithium is very rare, it contributes only little to the bulk conductivity (Gaillard, 2004). Thus, the conductivity is dominated by sodium ions, which are ions of an incompatible element. They tend to go into the liquid quickly and remain there for a long time span. In basic melt additionally calcium ions act as important charge carriers. Rocks with lower Si-content mostly have somewhat higher conductivity due to the higher NaO$_2$-content. Generally univalent ions are more mobile than divalent ions, since they interact less intensely with their environment. Melts with a high content of divalent cations contributing to conductivity show significant deviation from the Arrhenius temperature dependence, whereas melt conductivities dominated by univalent cations prove a conventional Arrhenius temperature dependence (Waff & Weill, 1975). The reason is according to Waff & Weill (1975) that because of their higher ionic potential (charge/ion radius) divalent cations polarize the oxygens of the surrounding polymerized SiO$_4^-$ and AlO$_4^-$-tetrahedrons in the melt structure more intensely than univalent cations do. This produces extended clusters around the divalent cations that have only a restricted mobility through the melt network.
in an applied electrical field (they have a large effective mass). Especially at lower temperatures the melt is widely polymerized, so that the divalent cations including their cluster have to move in long tortuous paths through the polymers. Divalent ions are significantly more mobile in regions between the polymerized units. With increasing temperature the melt’s degree of polymerization decreases and between the polymers more free space is produced. Thus, the divalent cations can move more freely through the network and consequently the apparent activation energy decreases. The apparent activation energy of univalent ions should be considerably less temperature dependent, since the paths of the ions through the melt less depends on size and distribution of the polymers, and therefore the ions are more mobile. The oxidation state of ions may also play a role in conductivity, since polyvalent ions (e.g. Fe$^{3+}$) tend to a lesser degree to dissociate into melts due to their greater binding energy. Then there would a smaller fraction of charge carriers be mobilized in the melt (Waff & Weill, 1975).

The melt’s conductivity increases significantly in presence of water. Water dissolves in rock melts and decreases the viscosity of the melt and thus increases the mobility of the ions in the melt (Partzsch, 1998). Li et al. (2003) obtained 1-10 S/m for dry melts and 1-20 S/m for wet melts.

The electrical conductivity is very sensitive to melt distribution. Especially the degree of interconnectivity of the melt is a crucial factor similar to rock-fluid systems. Hashin-Shtrikman upper and lower bounds ($HS^+$, $HS^-$) give a good estimate of boundary values of interconnection-dependent conductivity. In the case of perfectly interconnected melt the conductivity of the system is maximal and mainly depends on the melt conductivity. This is described by the formula of the Hashin-Shtrikman upper bound (Hashin & Shtrikman, 1962):

$$
\sigma_{HS^+} = \sigma_m \left(1 - \frac{3(1 - x_m)(\sigma_m - \sigma_s)}{3\sigma_m - x_m(\sigma_m - \sigma_s)}\right)
$$

\( \sigma_m = \) conductivity of melt
\( \sigma_s = \) conductivity of solid

\( x_m = \) volume of melt fraction
In the case of completely isolated melt pockets the bulk conductivity is low because it is dominated by the lower conductivity of the solid component (Partzsch et al., 2000). This can be described with the formula of the Hashin-Shtrikman lower bound (Hashin & Shtrikman, 1962):

\[
\sigma_{HS-} = \sigma_s \left( 1 + \frac{3 \chi_m (\sigma_m - \sigma_s)}{3 \sigma_s + (1 - \chi_m)(\sigma_m - \sigma_s)} \right)
\]  

(3.7)

![Figure 3.6](image)

Fig. 3.6 Calculated electrical conductivity of a partially molten rock as a function of the melt fraction based on the concept of Hashin-Shtrikman upper (HS⁺) and lower (HS⁻) bounds. The red lines indicate the measured overall conductance in the HCZ of the Altiplano and the melt fractions necessary to maintain the conductivity with melt conductivities of 5 and 10 S/m (perfect interconnection). After Schilling & Partzsch (2001)

Based on these equations, Fig. 3.6 shows that for a perfectly interconnected melt of 10 S/m a melt fraction of 12-13 vol% is needed to produce the overall effective conductivity of 1 S/m. For a melt conductivity of 5 S/m roughly 25 vol% are necessary. Natural melt is very likely not perfectly interconnected so that the melt content really required may lie around 20-30 vol%. Experiments have shown that it is possible to quantify the melt distribution from measured conductivity values and the observed melt
The ratio of interconnected melt to non-connected melt is deduced from the comparison of measured and modeled conductivities ($\text{HS}^+$ and $\text{HS}^-$).

Glassy material shows at lower temperatures higher conductivity values than completely crystallized material. This difference decreases with increasing temperature within the temperature range above the solidus temperature (Presnall et al., 1972). However, to produce glassy material the melt needs to be quenched quickly. Since it is not likely that the temperature of the high conductivity zone considered would drop intensely and rapidly enough, only a sudden pressure release caused by movement in the rock or certain mineral reactions could produce glass in the middle to lower crust. This phenomenon would be spatially very restricted. Furthermore, glassy material also tends to crystallize with the time, which would disturb the conductive structure. Thus, glassy material does not play a role in the explanation of the conductivity anomaly of the middle crust.

3.5 Conductivity based on aqueous fluids

The electrical properties of water depend on temperature, pressure, state of aggregation, and the amount of dissolved salts. Water itself is quite a bad conductor ($5 \times 10^{-6}$ S/m) (Partzsch, 1998). Electrolytes dissolved in water provide ions, which act as charge carriers. The concentration of ions and their interactions determine the conductivity of the solution. The higher the salinity of the fluid, the higher is the conductivity. The interactions of the charge carriers with each other and with their environment are determined by their charge, their mobility, the density of the fluid, pressure and temperature (Berktold, 1982). The term “aqueous fluid” as used in the following defines water containing dissolved ions as gases that all together can form a considerable conductor.

In rocks that contain a conductive electrolyte in an interconnected pore space serving as pathway for conduction, the electrolyte’s conductivity will provide the main part of bulk conductivity, depending on fluid saturation. A second conduction mechanism is the
electrochemical interaction between the fluid and the matrix material along the surface of the pores. Even a small percentage of water can increase the conductivity of a cold and dry basalt by several orders of magnitude due to this boundary surface effect. The conductivity of the rock matrix is here basically negligible. This can be expressed by the expanded Archie equation (100% saturation assumed) (Rink & Schopper, 1974):

\[
\sigma_0 = \frac{1}{F} \cdot \sigma_w + \sigma_{q0}
\]

(3.8)

\(\sigma_0\) being the bulk conductivity, \(\sigma_w\) the conductivity of the electrolyte, and \(\sigma_{q0}\) the surface conductivity of the pore space resulting from the electrolyte double layer that can be neglected in cases of very high salinity of pore fluid. \(F\) is a rock specific material constant that describes the geometry of the pore space (formation factor). The relation between \(F\) and the porosity \(\Phi\) is defined as (Archie, 1942; Telford et al., 1990)

\[
F = \Phi^{-m}
\]

(Archie’s Law) (3.9)

\(m\) is the cementation exponent that describes the compactness of the rocks, i.e. the packing density \((1.3 \leq m \leq 2.5)\). The difference in conductivity of dry and wet rocks is greatest at low temperatures and decreases as the temperature increases. Around the temperature of onset of melting this difference virtually disappears (Partzsch, 1998).

A possibility to describe the movement of a fluid through a porous matrix is given by Darcy’s Law (Bear & Bachmat, 1984; Rath, 2000)

\[
q = \phi \cdot v = -\frac{K}{\mu} (\nabla P - \rho g)
\]

(3.10)

with \(q\) being the intermediate velocity of the pore fluid \((v)\) times the porosity \((\Phi)\), \(K\) is the tensor permeability of the rock element, \(\mu\) the dynamic viscosity of the fluid, \(P\) the pressure, \(\rho\) the density of the fluid and \(g\) the gravitational acceleration. The precondition for the application of Darcy’s Law is that the total size of the system considered has to be significantly larger than the size of the grains and the pore space. Furthermore, the flow within the individual conduit has to be laminar, which provides an upper boundary of the
dimensions of porosity and the flow rate of the fluid (Turcotte & Schubert, 2002). Even though the latter precondition is very likely to be fulfilled, it is questionable, however, if the lower crust is a porous medium in the sense of Darcy.

An important condition of conduction based on saline fluid is the interconnection of the fluid in the rock. Interconnection in aqueous fluids follows the same rules as in melts. Dihedral angles smaller than 60° enable the fluid to interconnect along grain edges, whereas at dihedral angles greater than 60° the fluid remains in isolated pockets at grain corners. Increasing salinity can lower the dihedral angle significantly so that it becomes smaller than 60° (Li et al., 2003). With small dihedral angles only relatively small amounts of fluid are necessary to create an interconnected conductive network.

There are basically two boundary conditions that can describe a particular conductivity situation. There could be a fluid with low conductivity (low salinity) that occupies a relatively large volume and requires high porosity. The other extreme is a highly conductive, ion-rich fluid that only needs a relatively small volume and therefore little porosity (but still with a high degree of interconnection). In the upper crust the fluid is interconnected via pores and fractures throughout a strong wall rock. If the pore space is connected to the surface, the fluid pressure equals the hydrostatic pressure \( P_f = P_h \) (Yardley & Valley, 1997). In the lower crust the grains of rather ductile rocks collapse onto fluid-filled pores until the fluid pressure (pore pressure) equals the lithostatic pressure. That is, the pressure conditions at the depth considered can allow for a certain porosity if the effective pressure in the system (the difference between lithostatic pressure and pore pressure) approaches zero. Until approximately 10 years ago it was suggested based on laboratory experiments that from a depth of 3 km the pressure theoretically closes cracks and pores. Analyses of the KTB (ELEKTB-Group, 1997) however, demonstrated that in depths of 3 to 4, even of more than 5 km highly saline fluids invaded the drill hole from discrete horizons. This means that in those depths widely connected networks of high porosity still can exist filled with large amounts of aqueous fluids. Nevertheless, the high conductivity zone in the middle crust of the Altiplano is located in a depth range of 20 to 60 km with much higher pressure, and besides this in a zone of high tectonic activity. A high-porosity low-conductivity fluid situation is not very likely for this setting, since high porosity would cause the fluid to escape from the system.
within very short time scales, especially if located in an active orogen (Li et al., 2003). A porosity of 0.1% is suggested for the middle and lower crust by several authors (e.g. van Ngoc et al., 1986; Nesbitt, 1993), which refers to high-conductivity/low-volume settings. However, still a large amount of fluid would be necessary to explain the partly very high conductivities with saline fluids. These fluids have to be stable over a wide range of temperature and pressure for an appropriate time range or have to be supplied permanently by processes in the crust to compensate the loss due to escape. The large density differences between the rock and a few percent of fluid make it unlikely that under these conditions an interconnected network of saline fluids could be maintained in a conductive zone of the dimension considered over an adequate time period in a tectonically active region (Li et al., 2003). The fluid should move into shallower crustal levels and/or be caught in mineral reactions like the formation of hydrous minerals (e.g. muscovite) as soon as it reaches cooler settings. Hydration reactions usually cause a drop in water pressure below $P_l$, which means that an interconnected fluid phase will be interrupted under normal conditions. Moreover, in depths greater than 20 km a large amount of water should be dissolved in melt, since the solidus of water saturated rocks is exceeded there (Partzsch, 1998), which would lower the total amount of free aqueous fluid in the system. Wet silica-rich rocks begin to melt at approximately 650°C. To compensate the loss of water from the system several sources may contribute. First the subducted plate introduces large amounts of water into lower crustal and upper mantle levels. Hydrated minerals (e.g. amphiboles, biotite, and chlorite) release a lot of water as they are heated and experience prograde metamorphism (Frost & Bucher, 1994). Water could also stem from crystallization of igneous melts that contain water. Furthermore, the content of water in melt is associated positively with pressure, i.e. with overburden. The ascent of magma continuously lowers the amount of overburden and with it the pressure, so that water can escape the melt. Water is also transported by diffusion and advection even in almost impermeable crust (Rath, 2000). Even though this is a slow process, it runs throughout a large volume and a large (geological) time span so that in total a huge volume of water is involved. Sealing effects based on mineral redistribution (e.g. quartz) may slow down the water transport through the crust, but they do not seem to be able to completely stop the water from moving (Rath, 2000). Thus, even though much water will
escape from the system or be caught in mineral reactions there might still be a huge amount of water left that could be responsible for the conduction anomaly in the middle and lower crust (Fig. 3.7).

Fig. 3.7 Processes that determine the distribution of water in the crust. Reactions that produce free water are to a certain degree counteracted by those that consume water. The whole “circle” summarizes the processes over a large time range.

At low temperatures (below 300-400°C) the conductivity of a saline fluid increases with increasing temperature. The reason for this is that with increasing temperature the physical parameters viscosity, density and dielectric constant of the fluid decrease (Berktold, 1982). The temperature dependence of the fluid in the given temperature range is mainly based on the viscosity. A decrease in the viscosity causes a higher mobility of the ions, which results in increased conductivity. At temperatures above 300-400°C decreases in density and dielectric constant lead to decrease of conductivity with increasing temperature (Berktold, 1982).

At low temperatures (<300-400°C) the conductivity of the fluid is almost independent of pressure. The conductivity rises only slightly with increasing pressure; low-concentration solutions may even show a minor decrease in conductivity. At higher temperatures (above 300-400°C) the conductivity stronger depends on pressure. With increasing pressure (up to 300 MPa, Berktold, 1982) the conductivity increases significantly. This effect also occurs in low-concentration solutions, but there the
conductivity again decreases somewhat at very high pressures (1200 MPa) (Berktold, 1982).

Another factor that should not be ignored is the fact that saline solutions do not increase their conductivity with increasing salt concentration infinitely. The conductivity of a fluid increases with addition of salt up to a certain degree, but at very high concentrations (e.g. roughly around 40 wt% for KCl) the conductivity decreases towards values of the molten salt (Nesbitt, 1993). Since crustal fluids have only a salt concentration of 2-20 wt%, there can only a restricted increase in conductivity of crustal structures be explained by the existence of saline fluids (Nesbitt, 1993). Also HCO$_3^-$ and CO$_3^{2-}$ ions in the fluid (depending on the pH and the amount of dissolved CO$_2$) as suggested by Nesbitt (1993) are not very likely to increase the conductivity to the values that would be required for low porosity settings mentioned above.

The temperature regime of a region could be thought to be the key to distinguish between aqueous fluids and partial melts. The heat flow density in the central Andes is relatively high (max. 140 mW/m²). Modeling has proven that the temperature in 20 km depth can reach the solidus temperature of wet crustal rocks. Therefore, partial melting in a thickened crust could occur, taking large amount of convective heat transport, high radiogenic heat production, and frictional heat generation during collision into account (Schilling & Partzsch, 2001). However, the heat flow density is the least constrained of all parameters obtained in the Andes (Brasse et al., 2002).

### 3.6 Combinations of mechanisms

The conductivity of the HCZ beneath the Altiplano is widely considered to be the result of melt or aqueous fluid conductivity. However, there are some concerns about this approach. The measured data show an enormous conductivity of around 1 S/m within the center of the anomaly. This result has been proven to be stable by Schwalenberg (2000) using sensitivity analyses. Also other authors report conductivities in the range of 0.5 to 2 S/m (among others Schwarz et al., 1994; Lezaeta et al., 2000; Schilling & Partzsch, 2001). The measured conductance of the Altiplano represents integrated values for a
large volume. Since the volume certainly does not consist of 100% melt or aqueous fluid, the value includes the conductivity of the rock matrix as well as the conductivity of the fluid within the volume. Assuming 20 to 30 vol% of fluid, being interconnected to a high degree, the measured conductivity depends mainly on the fluid and only to a low portion on the conductivity of the solid rock.

There are numerous models combining different conduction processes to explain the high conductivity zones in the Andes as well as in the Tibet Plateau and other regions. The problem with combining conduction processes is that one has to be very careful about the conduction mechanisms being associated. There are diverse conduction mechanisms in different phases of the crustal conductive system. Conduction can be based on ions as in melts and aqueous fluids, but also on free electrons as in graphite or on electron-electron hole systems as in semi-conducting minerals. The conduction in a multi-phase system is crucially determined by the behavior of the charge carriers at the phase boundaries. For example a path of conduction based on electrons as in graphite can not be continued with an ion-based mechanism such as saline fluid or a melt, since the electrons would not be able to enter the new medium. For them the new medium is a barrier. In contrast, the accumulation of electrons at the boundary surface of the graphite would lead to accumulation of positive ions in the part of the liquid being in contact with the graphite. A strong polarization effect could be measured. This would even increase the resistivity of the system in this zone, lowering the overall conductance rather significantly. Also grain boundaries within the rock provide heterogeneities that influence the transport of charges within the semi-conductors.

Summary
The conductivity of the crust is assumed to normally range between $10^{-2}$ and $10^{-3}$ S/m. Zones of higher or lower conductivity are considered as anomalies. The conductivity of a material depends on the existence of mobile charge carriers such as ions or electrons. Several basic types of crustal conductivity anomalies were examined to characterize the high conductivity zone beneath the Altiplano. A conductivity anomaly based on conductive minerals like ores could be ruled out because of the scale of the anomaly and
because of other geophysical constraints. An increase of conductivity could theoretically be explained by the existence of graphite that is concentrated on fossil shear zones or in layers that have formed by metamorphosis of black shale into graphite. The stability of a graphite layer depends strongly on the oxygen fugacity in the system. Also the presence of melt in a partially molten system can produce increased conductivity if it forms a widely interconnected network throughout the volume considered. Tectonic deformation can change the degree of interconnectivity and therefore the conductivity significantly. To maintain a conductive system throughout a long time period the melt has to be stabilized in the crust to prevent it from rising and a permanent heat supply is necessary to avoid freezing of the system. The Hashin-Shtrikman upper and lower bounds give an estimate of how conductive a system can be depending on pure-melt conductivity, interconnectivity and melt fraction. Finally, water that is released from the heated subducting plate and forms brines in the crust can cause strong increase in the conductivity of a system. The brines rise through the mantle and the crust and eventually collect in zones of higher concentration of fluid where they form a wide network. Aqueous fluids tend to rise through the crust and leave the system rapidly. Thus it appears very likely that the conductivity is mainly based on the existence of partial melt and to a lower degree on aqueous fluids.
4 Modeling of magnetotelluric responses of specific geological settings

4.1 Principle of magnetotelluric methods

Magnetotelluric (MT) methods are based on the induction of alternating natural electromagnetic fields into the conductive earth. The source of the long periodic spectra (>1 s) of these fields are the current systems of the ionosphere and the magnetosphere of the earth, which vary in the course of time mainly because of the solar wind (Müller, 1995). The high frequency parts (10-10^4 Hz) are covered by AMT (audio frequency MT), which use mainly sferics produced for example by distant lightning as energy sources (Reynolds, 1997) (see Fig. 4.1 for comparison of the frequency ranges of electromagnetic methods).

![Frequency range of the electromagnetic spectrum and methods of geophysical measurements](image)

Fig. 4.1 Frequency range of the electromagnetic spectrum and methods of geophysical measurements. (VLF = Very Low Frequency, VHF – Very High Frequency, EHF – Extremely High Frequency). The ranges utilized for the methods partly overlap. Modified from Jiracek et al. (1995).
The electric \((E_x, E_y)\) and the magnetic \((B_x, B_y, B_z)\) field components are measured at the earth’s surface over a certain period range (normally \(10^{-4} - 10^5\) s) (Schwalenberg, 2000; Brasse, 2003). The variations of the electrical field are measured in mV/km and those of the magnetic fields are measured in nT. The advantage of the use of MT methods is the great penetration depth of the measurement (up to 1800 km) (Müller, 1995). In MT the horizontal components of the fields are connected in a linear relation via the magnetotelluric tensor \(Z\) (Červ & Pek, 1990):

\[
\begin{pmatrix}
E_x \\
E_y
\end{pmatrix}
=
\begin{pmatrix}
Z_{xx} & Z_{xy} \\
Z_{yx} & Z_{yy}
\end{pmatrix}
\begin{pmatrix}
B_x \\
B_y
\end{pmatrix}
\] (4.1)

The variations of the components provide insight in the distribution of subsurfacial conductivity depending on depth and lateral position within the crust. The depth penetrated strongly depends on the period of the fields and is described by the skin depth (Schwalenberg, 2000):

\[
p(\omega) = \frac{2}{\omega \mu \sigma} \approx \frac{1}{2} \sqrt{\frac{\rho \tau}{\xi}} .
\] (4.2)

In a homogeneous half space the E and B fields are constant in direction and magnitude over planes perpendicular to the vertical (downward +z) direction of propagation (Jiracek et al., 1995). The surface impedance equals the characteristic impedance of the medium. The value of the impedance is usually expressed as apparent resistivity \(\rho_a\) and the phase \(\Phi\) rather than the fields themselves (Schwalenberg, 2000):

\[
\rho_a(\omega) = \frac{\mu_0}{\omega} |Z(\omega)|^2 \quad \text{with} \quad |Z(\omega)| = \sqrt{\frac{\omega}{\mu \cdot \sigma}}
\] (4.3)

\[
\Phi(\omega) = \arctan \frac{\text{Im} Z(\omega)}{\text{Re} Z(\omega)}
\] (4.4)
\( Z(\omega) \) is the frequency dependent MT-transfer function. For a homogeneous half space the apparent resistivity \( \rho_a \) equals the specific resistivity of the material of the subsurface; the phase \( \Phi \) is 45°. However, usually geologic settings of interest are not just homogeneous half spaces, but rather 2D or 3D structures of contrasting resistivity.

Magnetotelluric methods are only sensitive to the presence of conducting material (e.g. interconnected fluids) that is distributed with a pronounced horizontal extent. They are blind to the internal structure of the conductive zone and produce only an integrated conductivity averaged over a certain depth range. A thin layer of high melt fraction and an equally conductive thick layer of low melt fraction would give the same measuring results (Li et al., 2003). The fact that only the depth to the top and the conductance or the thickness-resistivity ratio of the layer are well defined complicates the attempt to differentiate the resistivity decrease in a high conductivity zone (Nesbitt, 1993). It is impossible to determine the thickness of a single conductive layer within a system of layers, and therefore information about porosity of the system has to be obtained from other sources (van Ngoc et al., 1986).

The electrical conductivity of rocks and rock-melt systems has been intensely studied in laboratory experiments by means of impedance spectrometry. Impedance spectra of silicates and liquids have been measured under varying temperature conditions. The electrical impedance is a function of electrical frequency in the range of 1 kHz to 2 MHz. The spectra provide information about the actual conduction mechanism (electronic and ionic conduction) and the conduction path (grain boundary or grains, or melt).

Conductivities measured in lab experiments are often lower than those measured in the field by several orders of magnitude. Labendz (1999) explains this with scaling problems and with uncontrolled changes of conditions of the rocks after sampling. For example pressures in samples from drill holes hardly can be maintained during sampling and transport so that paths of conduction vary. Furthermore, many experiments have been carried out at elevated temperatures and normal pressures. Thus, cracks that have formed while heating the specimen and the fact that the samples are very dry are partly the reasons for those low conductivity values (Wei et al., 2001). According to Roberts & Tyburczy (1999) the conductivity of partially molten rocks determined at frequencies larger 10 Hz is not directly applicable in the interpretation of geomagnetic results. The
difference between the apparent conductivities of field and lab experiments depends on the degree of frequency dispersion. Further, the size of the sample determines the frequency of measurement. In impedance spectra the different conduction elements are separated by capacity, whereas in MT induction is used to divide the components.

The application of laboratory results to field data is a difficult process. Serious simplifications have to be made to simulate natural processes in experiments. The most radical simplification lies in the time factor. Most features of geological systems develop over wide time ranges that cannot equally be applied in experiments. Consequently, certain conditions such as the establishment of oxygen fugacity, the redistribution of melt within large bodies or any type of large scale equilibration are difficult to record (Partzsch, 1998). Certain geologic processes such as large scale partial melting, inherent heat production of radioactive decay, melt motion and establishment of melt containing systems within the partially molten crust also cannot be approached by experiments due to the spatial restrictions of laboratory equipment.

4.2 Modeling procedure

Based on petrological and geological constraints several models of geologically varying concepts have been developed for 3D-Forward modeling. The program used for this study was WinGlink\(^1\), which is a software system for processing, modeling and interpretation of data from non-seismic geophysical surveys. It uses a finite differences algorithm with discretization and Tichonov-regularization. The aim was to examine how sensitive a theoretical MT setup reacts on different geological settings, i.e. how constrained interpretation and modeling based on “real” (measured) data are. The modeling process employed here contained two basic steps. In the first step the models were constructed starting from a standard model. These models were run with the 3D-Forward modeling routine (set of modeling parameters see App. A1). The data obtained were then run with a 2D inversion routine (set of parameters see App. A2). The intention

\(^1\) WinGLink© Software is provided by geosystem http://www.geosystem.net/
was to find models that closely reproduce the data gained from the “real” data from the
ANCORP profile (see Schwalenberg, 2000).

Fig. 4.2 The inversion model obtained from the measured data of the ANCORP
profile ($\tau = 100$, Laplacian). From Schwalenberg (2000).

In the second step the models were varied in terms of shape, size and conductivity
values of the conductive structure as well as in terms of conductivity contrast between
conductivity structure and surrounding host rock. The aim was to examine how these
variations influence the computed response. All results were compared with the results
obtained from 2D-inversion of “real data” as presented in Schwalenberg (2000) and
Brasse et al. (2002) among others (Fig. 4.2).

The standard box model used for the modeling consisted of a central volume of
highly resistive matrix of 1000 $\Omega$m and a marginal shell (Fig 4.3). The x,y-dimension of
the central part were chosen to roughly approximate the dimension of the Altiplano (600
by 900 km) in order to develop an idea of the size of the geologic features modeled and
also in order to directly compare the modeling results with “real” data. The cell size was
20 by 20 km, and each cell had a specific constant conductivity value. The depth extent
has been designed to a depth of 70 km with constant cell sizes except at the surface where
the cells are increasingly smaller due to the discretization demands. Only in this central
volume the structural models were emplaced. For the creation of numerical models it is
necessary to construct a frame of increasingly larger cells around the central volume
towards the outer boundaries of the model box. Therefore the actual modeling volume
was extended by a few layers of 1000-$\Omega$m cells with increasing size in the ±-x, ±-y, and
+-z directions. An outmost margin of the model was constructed like a bawl around the
1000-Ωm box containing three layers of significantly lower conductivity (100 Ωm) to assure that the fields ceased within the model and that there were no disturbing 3D effects outside of the central volume. Thus, these marginal layers are a purely technical feature for computing and do not have any geological meaning. This standard model has been varied for comparative models to find out how the background conductivity of the matrix and with it the conductivity contrast between the geological structure and the matrix would change the MT response. One standard model box with a matrix resistivity of 500 Ωm (Altiplano-Standard 2) and one with 100 Ωm (the latter one with a 50 Ωm marginal bowl) (Altiplano-Standard 3) were constructed additionally.

Fig 4.3 3D image of model AP1Standard. The center has a resistivity of 1000 Ωm, the margin 100 Ωm. The profile marked in yellow is profile C with the stations C1 to C14 that is used for the presentation of the modeling results.

For the modeling procedure four profiles oriented east-west and one profile oriented north-south have been created. Only the results of the east-west profile C are shown, since this profile was used as modeling profile representing the ANCORP profile. The
other profiles were employed to control the calculation and transformation processes accomplished by the WinGLink programs.

Having the geological structures included in the models, discretization of the transition between low and high conductivity zones led to bad data and non-convergence when it was set too fine. This probably occurred because the cells of the model gained too extreme shapes, for example they became very long but exceedingly thin. Especially in models with significant 3D structure the discretization produced strongly disturbed data as could be seen in the sounding curves. This should not happen with synthetic data, which usually produce very smooth curves. In models with less prominent 3D structure often no significant difference between models run with and those run without discretization was obtained. Models without any or with only very low discretization ran by far the best. For models with decent discretization the program also often failed to start the forward calculation because the model (number of cells) became too large. It was not possible with this modeling program to only decrease the cell size of the modeled structure and its vicinity and keep the cells large further away from this area. The cell size could only be decreased by introducing complete layers into the model, which led to an exorbitant increase of the cell number. For the same reason the models could not have a complex internal structure, since this would have made the cell number exploding due to discretization. Therefore discretization and detail were generally kept at a relatively low level.

The forward calculation is an iterative process that is supposed to converge at a final error (set at $10^{-7}$). For several models convergence was not achieved. These models were rejected.

The regularization parameter $\tau$ is the smoothing parameter that controls the trade-off between fitting the data and observing the model constraint. Large values of $\tau$ lead to smoother models at the expense of worse data fit. $\tau$-values of 50 and of 100 have been generally applied for the 2D inversions to make the results of the different model concepts comparable with each other and with the “real data” models obtained by Schwalenberg (2000) and Brasse et al. (2002).
4.3 Model descriptions and modeling results

4.3.1 Conductivity model of dikes

Several models of dikes were tested for the concept of dikes serving as conductors. The idea was that a large dike has a pronounced horizontal dimension along its strike and therefore might be detected by MT. The examples included single-dike models as well as models with multiple dikes. Generally the dike width had to be modeled fairly large (20 km), since the computer capacity did not allow for smaller cells because of the restriction to the total cell number. Thus swarms of dikes of approximately parallel orientation were united to one large “effective” model dike. The decision on the construction of the dikes was based on the following consideration. The orientation of the dikes is determined by the local and regional stress field. The stress field in the Altiplano is mainly influenced by the compression due to the collision of the Nazca and the South American plates that formed the Andean mountain belt. If dikes formed then they only could appear oriented within a range of angles somewhat inclined (30°) to the compression direction within the brittle part of the crust. For simplification it is assumed that the crust behaves brittle throughout the whole dike extent\(^2\). Besides this, paleomagnetic measurements have revealed a fairly consistent pattern of rotations having their center at the Bolivian orocline, which forms a distinct bend between 18 and 20°S (Rousse et al., 2003). North of the deflection the rotation occurred mainly counterclockwise and south of it clockwise. The Altiplano is located right at the hinge of

\(^2\) In reality this may not be the case because in the lower, hotter part the crust behaves rather ductilely. Besides this the load of the overburden may switch the \(\sigma_2-\sigma_3\) relations of the compressive system in that way that \(\sigma_3\) is oriented vertically and \(\sigma_2\) horizontally (NS) in the upper part of the crust, but due to the increase of overburden in deeper levels \(\sigma_3\) may be horizontal (NS) whereas \(\sigma_2\) is vertical. This would have consequences for the orientation and propagation of dikes.
this bend where only little rotation is observed (Kley, 1999). Extension fissures in the hinge would be oriented in a very similar way as the dikes due to compression. It is not very likely for the Altiplano region that dikes are the source of the high conductivity zone.

![Diagram of conductivity zones and model dike orientations](image)

**Fig. 4.4** Model D1a: a highly conductive dike surrounded by a gradient zone stands vertically in a highly resistive host rock. The profile view on the right side is taken from the position of profile C in the map view (left). The profile is placed on an edge of the dike. The thick black lines are a number of separate thin lines beyond the graphical resolution.

With this orientation the dikes were only slightly inclined to the strictly east-west striking profile C. Thus, the general conductive direction of the dikes roughly paralleled the profile. The position of the profile relative to the dikes varied in the models. In model D1a the profile cut the dike only at one edge, whereas in model D4a the profile sat centrally on top of the dike.

The dikes were constructed as bodies with a highly conductive center (1 Ωm) surrounded by two 20 km wide (at the sides) and 3,5 km thick (on top and at the bottom) gradient shells of 10 and 100 Ωm, respectively, that bordered to the 1000 Ωm matrix representing the surrounding rock. This was probably a gradient too smooth for a geological system. Thus, models with a steeper gradient of only one 100 Ωm shell were tried as well. The highly conductive center of the dike simulated melt and fluid just filling and eventually propagating the dike. The horizontal extent of the single dike model was
320 km. The discretization had to be kept on a fairly low level to be able to start the forward calculation.

Fig. 4.5  Model D4a: a highly conductive dike surrounded by a gradient zone stands vertically in a highly resistive host rock. The profile view on the right side is taken from the position of profile C in the map view (left). The profile is placed centrally of the dike.

The 2D inversion was run with the regularization of the strongly smoothing “Uniform grid Laplacian operator” because intense 3D-effects had to be compensated (see App. B, Fig. 8.1). The inversion result of the single dike models revealed that only in the model, in which the profile was located centrally (i.e. symmetrically) on top of the dike (model D4a) a strong conductivity structure found by e.g. Schwalenberg (2000) was reproduced by the inversion. In the model D1a, in which the profile did not sit symmetrically on top of the dike but rather shifted onto the edge, the compact conductivity structure reproduced by the inversion modeling was too small (Fig.4.6a). The conductive zone was restricted to the stations C7, C8, and C9, which were the stations immediately on top of the dike center or immediately next to it. A zone of low to intermediate resistivity (5 to 100 Ωm) surrounds the high conductivity zone. The non-symmetry of it is due to the non-symmetric setting. The resistivity of the center of the conductive structure was locally somewhat underestimated (1 Ωm and less), whereas the resistivity of the host rock was partly overestimated (up to 7000 Ωm). The depth position of the anomaly was reproduced fairly well in the inversion result.
Fig 4.6 2D inversion results of the single-dike models D1a and D4a showing the resistivity distribution over the depth along profile C. a: model D1a rms = 4.152, b: model D4a, rms = 4.443
Model D1b was run without any discretization. It displayed no significant deviation from the one run with discretization. The 2D inversion results were virtually alike. The rms value even improved for the inversion of the non-discretized model. Models with a sediment layer of 5 Ωm and 1 km thickness in the uppermost levels were constructed for both D1 and D4 to test how a conductive layer on top of the system would influence the response. They were dismissed because the forward computation did not converge. Experiments with steeper gradient and no gradient completely failed to reproduce any highly conductive region in the 2D inversion. Only a very diffuse zone of intermediate conductivity in a host rock of locally highly overestimated resistivity was produced.

The models with the profile being located centrally on top of the dike (D4a) did consistently reproduce a large scale anomaly. The discretization of the model (D4a) caused significant disturbances that were recognized in the sounding curves. Inversions based on these data produced a large conductivity anomaly but with a huge rms of 17, 65 and a conductivity anomaly of infinite depth extent. Repeated inversion performed with corrected soundings that did not include outlying data anymore produced a more restricted conductivity anomaly with a significantly improved rms (Fig. 4.6). Tests with a model without any discretization (D4b) produced very smooth synthetic sounding curves without any disturbance and inversion results that were virtually like the results from the corrected data of D4a. It is not clear at the moment why the discretization causes significant trouble with the synthetic data. As seen in Fig. 4.6, the region of strongly decreased resistivity in the new inversion result extended from the stations C3 to C8, which were the stations in the vicinity of the conductive center of the dike. The resistivity of the anomaly was somewhat underestimated with values of less than 1 Ωm (“cell value 0”). The conductive center was surrounded by a quite wide frame of low to intermediate resistivity (5 to 100 Ωm), which was interpreted as the region that was influenced by the parts of the dike that were located farther away from the profile. The resistivity of the surrounding host rock was locally somewhat overestimated (up to 5000 Ωm).

Experiments with steeper gradient and none at all did not reproduce any high conductivity zone similar to the one obtained before. Hardly any zone of increased conductivity was imaged in the 2D inversion. A spot of 30 Ωm in a depth of 70 km beneath the stations C5 and C6 was all that was reproduced from the highly conductive
dike. The host rock became somewhat heterogeneous with values between 400 and 3000 Ωm. In experiments with decreased resistivity contrast the structure was not reproduced any better. An irregularly shaped zone of increased conductivity appears, but no compact high conductivity anomaly similar to the one based on real data is obtained.

The multiple dike models were derived from the single dike model and simulated systems of extended dikes and of shorter (quasi interrupted) dikes, respectively (see App. B, Fig. 8.2). The extent in width and depth was of the same order as for the single dike model. The forward calculation of model D2 did not converge. Therefore no 2D inversion was produced from the forward data. The forward calculation for model D3 converged and a 2D inversion was produced. The result was also strongly affected by 3D effects. The inversion result again demonstrated that the anomaly that was found was of unclear structure and fairly weak in intensity when the profile sat between the dikes. The resistivity of the host rock was represented with rather absurd values of several 100,000 Ωm (see App. B, Fig. 8.3).

4.3.2 Conductivity models extended melt zones

For this chapter numerous models of pluton-like conductive bodies were investigated. The conductive bodies had the very simple shape of a rectangular parallel-epiped of high conductivity. Variations of the basic model included changes in the gradient and the degree of discretization as well as different conductivities of the conductive body, i.e. of the conductivity contrasts between conductive body and host rock. The forward calculation was run as described before. The 2D inversion reproduced the body’s size in terms of its lateral extent and its depth position very well. This means that the EW-extent of the structure had to be set up in the lateral extent of the conductive anomaly of the model obtained by Schwalenberg (2000) to reproduce this anomaly by modeling.

The conductivity bodies had an E-W-extent of 200 km and a N-S- extent of 400 km to approximate a large conductive volume similar to the one obtained by e.g. Schwalenberg (2000). The large NS extent was chosen to keep the structure simple in terms of
modeling. As mentioned above, there is no clear data about the real NS-extent and the internal shape of the conductive structure beneath the Altiplano. The depth extent of the

Fig.4.7 M1a: original model and inversion result. The inversion shows the resistivity distribution over the depth along profile C. $r_{ms} = 1.97$. The conductive volume is reproduced quite well; however, the inversion result is significantly more heterogeneous than the original model.
conductive body was set up to 50 km and 63 km. Some of the models that were investigated have an EW-extent of 100 km and a NS-extent of 140 km, and are located in a depth of 20 to 35 km. Model M1a (Fig. 4.7) had a highly conductive center of 1 Ωm and a relatively wide gradient zone (i.e. a flat gradient) that formed a transition to the 1000 Ωm matrix. The model represented a melt-containing system that had existed already for quite a long time so that the heat could influence a fairly wide volume of the surrounding wall rock. The gradient was formed by a shell of 10 Ωm and 20 km of width and another 20 km wide shell directly around the first one with a resistivity of 100 Ωm along the sides. Each of them had a thickness of 1.7 km on top and at the bottom of the body.

The 2D inversion was run with “Standard grid Laplacian” for regularization because the parameter used for the dike models led to unreasonable results in which the anomaly of overestimated conductivity was extended into infinite depth. In the 2D inversion (Fig. 4.7) result the lateral extent and the depth of the top are reproduced very well. The bottom of the diapir is less good resolved and its depth is widely overestimated (80 km vs. originally 50 km). In models with even smaller conductive structures, that reached only 35 km of depth the reproduced thickness of the anomaly in the inversion result also was approximately 70 to 80 km. The poor depth resolution of conductive structures is a very typical phenomenon in MT investigations.

This model demonstrated that the depth extent of inversion results had to be considered with a good portion of scepticism, since the thickness of a conductive body did not necessarily have to be as large as the inversion result indicated. The conductive structure showed a tendency to appear split into a larger and a smaller part. The resistivity of the anomaly was somewhat underestimated and reproduced with values less than 1 Ωm. The inversion result also introduced additional heterogeneity into the matrix volume that surrounded the conductive body. As already seen in the dike models the matrix showed some kind of “bow waves” of alternating high (up to 5000 Ωm) and low (as low as 500 Ωm) resistivity zones departing at the conductive body and located along the sides of the anomaly. The resistivity at the surface in larger distance from the pluton (stations C1, C2, C13 and C14) was fairly similar to the value given in the original model. The resistivity of the very uppermost layers above the conductivity structure was
quite close to the original value, within a range of 700 to 1200 Ωm. The resistivity of the host rock below the actual conductivity body was decreased to 70 Ωm and connected with a layer of intermediate resistivity in a depth of 250 km. This zone of relatively high conductivity (20-80 Ωm) in a depth range of 130-210 km was an artifact that was not included in this discussion. Elimination of the discretization did not cause significant changes in the inversion result. Only the tendency to isolate the edges of the conductive structure was slightly increased. Eliminating the gradient zone or both the gradient and the discretization improved the imaging of the original model. A compact structure of high conductivity was produced. The host rock besides the structure was heterogeneous again and the resistivity values were locally overestimated.

Generally it also turned out that for this model the depth extent of 50 km of the original model is too small to reproduce the conductivity anomaly obtained by Schwalenberg (2000). In the inversion result, the lower boundary of the conductive zone based on real data was not shown, but clearly the anomaly extends to greater depth than it was reproduced with the models analyzed here.

Steepening the gradient around the 1 Ωm center somewhat increased the depth extent of the anomaly that was reproduced in the inversion result. The model represented a system that has been emplaced quite recently so that the surrounding host rock was not heated intensely yet. The models had only a gradient shell of 10 Ωm, which was 20 km wide; the rest the model was set like in the previous model. Only for the steep-gradient model in which the discretization was eliminated the 3D-forward calculation produced proper results (model M2b, App. B, Fig. 8.4) and the inversion reproduced the conductive body reasonably well. The synthetic data for the model M2a run with discretization showed significant outliers in the data. Correcting the data by masking the outliers led to a quite good inversion result in terms of reproducing the conductivity anomaly (Fig. 4.8). The conductivity anomaly was compact in shape and looked very similar to the previous models and to model M2b. Only the edges were separated from the central part. The depth extent was quite large and reached a level of 100 km.
Fig. 4.8 Inversion result of model M2a. The inversion shows the resistivity distribution over the depth along profile C. $\text{rms} = 3,463$. The conductive volume is reproduced quite well; however, the inversion result is significantly more heterogeneous than the original model.

The resistivity of the anomaly remained underestimated, whereas the resistivity of the host rock was heterogeneous and exhibited zones of high and zones of low resistivity. Directly beneath the anomaly the host rock in the inversion result showed a region of moderate resistivity (20 to 100 $\Omega\text{m}$). In the inversion result the gradient zone between high conductivity zone and host rock did not show significant changes compared to the model with flatter gradient.

The increase of the depth extent of the conductive body down to 63 km in model M3a did not considerably increase the depth extent of the anomaly obtained in the 2D inversion procedure. The internal structure and the contour of the top of the conductive zone varied slightly but the depth extent of the zone of maximum conductivity did not exceed 100 km (see App. B, Fig. 8.5).

An additional set of models was calculated testing the influence of decreased conductivity of the conductive body. In model M4b (Fig. 4.9) the central conductivity of
the conductive body was set to be 5 $\Omega$m. The lateral extent of the structure was set as before, and the maximum depth of it was 50 km. The inversion result showed a split high-conductivity body with three highly conductive (overestimated, <1 $\Omega$m) nests of similar size, which lay directly beneath a thin layer of 3 to 5 $\Omega$m that approximated the top of the original conductive body. These nests were interconnected and surrounded by a narrow zone of 10 to 20 $\Omega$m. The top of the conductivity body was at the right depth. However, the average thickness of the anomaly and therefore its depth was strongly overestimated (100 km vs. 50 km). The lateral extent of the reproduced high-conductivity anomaly was very similar to the originally implemented body. The highly resistive matrix showed “bow wave” structures of alternating high (up to 8000 $\Omega$m) and low (500-700 $\Omega$m) resistivity. The intermediate conductivity layer at 250 km depth was an artifact that was not discussed any further.

In model M5a (Fig. 4.10) the resistivity of the melt containing body was reduced to 10 $\Omega$m. The gradient zone is only a 20 km thin shell with a resistivity of 100 $\Omega$m. Lateral extent and thickness were like in the model before (M4b). The inversion result did not give any compact image of the conductive body. Three small nests in a depth range between 50 and 80 km with a conductivity of less than 1 $\Omega$m were interconnected by a field of intermediate conductivity of 10 to 20 $\Omega$m. The top of the conductivity structure was represented as a somewhat diffuse transition from lower to higher conductivities. The depth level of the zone of increasing conductivity was located approximately at the right depth. Due to the separation of the conductivity nests the bottom of the whole structure ended up way too deep (120 km). The host rock’s resistivity was heterogeneous as in all other models before. “Bow wave” structures of alternating high (up to 10000 $\Omega$m) and low (as little as 500 $\Omega$m) resistivity appeared along the sides of the anomaly down to large depths and also directly beneath the high-conductivity zone a high (up to 6000 $\Omega$m) resistivity zone occurred.

Elimination of the gradient zone led to an anomaly of somewhat more solid shape. The inversion result of model M5c (App. B, Fig. 8.7) exhibited a zone of high conductivity (<1 $\Omega$m) in a depth range of 70 to 120 km. A 40 km thick region of intermediate
Fig. 4.9 M4b original model and inversion result. This model is not discretized. (For some unclear reason it was not possible to start the forward calculation of the discretized model. However, experiments have shown that modeling is stable and gives reasonable results without discretization.) The inversion shows the resistivity distribution over the depth along profile C. rms = 1.211. The conductive volume is reproduced quite well; however, the inversion result is significantly more heterogeneous than the original model.
resistivity (5 to 20 $\Omega$m) was located above this zone. A thin gradient zone surrounded the high conductivity zone along the sides and the bottom. The host rock resistivity was heterogeneous and locally overestimated.

The models with decreased resistivity contrast (M4b and M5a and c) did not reproduce any large scale high conductivity zone as obtained by measurement in the Altiplano (App. B Fig. 8.6). Obviously, the conductivity contrast between conductive body and host rock had to be very large to reproduce a compact high conductivity zone in the inversion result. Furthermore, only structures of large extent led to anomalies of appropriate size compared to the result from “real” data because the lateral extent of the conductive structure always was reproduced quite well.

Decrease of the host rock resistivity to values of 100 $\Omega$m and of the conductive body to values of 5 $\Omega$m led to strong destruction of the anomaly in the inversion result. The high-conductivity block did not appear at its original position but significantly lower (at 60 to 140 km). The internal resistivity was underestimated (1 $\Omega$m and less). At a depth of 20 km only a zone of intermediate resistivity (20 to 50 $\Omega$m) appeared, which included two diffuse spots of 2 to 5 $\Omega$m. From this zone of intermediate resistivity a pair of “bow waves” of 20 to 40 $\Omega$m occurred. Besides and below the high-conductivity spot a zone of 400 to 600 $\Omega$m appeared, which demonstrated again a strong overestimation of the host rock resistivity. Only close to the surface the resistivity was well reproduced with values around 100 $\Omega$m. Generally the 2D inversion produced a high degree of heterogeneity within the model.

The trial to model a sediment layer as seen in the inversion result by Schwalenberg (2000) led to complete vanishing of the underlying conductive body in the inversion result because the sediment swallows the current. Only unclear zones of decreased conductivity appear in large depth.
Fig. 4.10  M5a original model and inversion result. The inversion shows the resistivity distribution over the depth along profile C. rms = 6,406. The conductive volume is reproduced quite well; however, the inversion result is significantly more heterogeneous than the original model.
4.3.3 Conductivity models of shear planes

The concept of conduction based on the existence of large scale shear planes as part of an underthrust system, in which the Brazilian shield dives beneath the Andes, was tested using two basic sets of models. The first set (models SP1a-c; SP1a Fig. 4.11) consisted of one large overthrust zone of four km thickness that extended from 20 to 40 km of depth. This could represent a combination of a group of single shear planes too small to be modeled separately as well.

The second set of models (models SP2a-c; SP2a Fig. 4.12) comprised three separate overthrust segments of approximately two km of thickness piled up on top of each other. The whole conductivity zone increased in thickness compared with the first model set due to this separation. The highly conductive centers of the multiple shear plane system were separated by 4 and 6 km less conductive host rock (10 Ωm). The dip angle is 3.3° to the west. The whole complex of shear planes in both sets was surrounded by a gradient zone of 10 and 100 Ωm that formed the transition to the highly resistive (1000 Ωm) host rock. The EW-extent resembled the size of the high conductivity zone as it was obtained by inversion computation done by Schwalenberg (2000). The shear plane models were not discretized due to the restriction in the amount of cells manageable for the computer.

Fig.4.11. Model of SP1a. The conductivity is based on one large shear plane. The white line in the cross section on the right denotes the level of depth that is shown in the map view on the left.
models converged in the 3D forward calculation process. Since the shear planes were of fairly wide extent, significant anisotropy in the MT responses was not expected.

![Diagram](image)

Fig. 4.12. Model of SP2a. The white line in the cross section on the right denotes the level of depth that is shown in the map view on the left.

The 2D inversion of model SP1a (Fig. 4.13) produced a very large high conductivity zone that was as wide as the original model at the top and narrows with increasing depth. The position of the top was reconstructed very accurately, and also the dip of the zone was clearly reproduced. The conductive zone showed a fairly high degree of heterogeneity, especially along the top. In a compact zone of high conductivity (<1 \( \Omega \) m) spots of lower conductivity (3-5 \( \Omega \) m) were emplaced. The lower boundary of the high conductivity zone was only poorly resolved. The depth extent of the conductive structure was by far overestimated: in the original model the shear plane reached a depth of 45 km, whereas the reproduced zone of extreme conductivity ended in a depth of 170 km. The gradient zone was very thin along the top of the high conductivity zone. It widened along the edges with increasing depth and was very pronounced below the structure showing a wide depth range with a resistivity of 10 to 100 \( \Omega \) m. ”Bow wave” structures of very high resistivity (up to 14000 \( \Omega \) m) appeared besides the conductive structure, especially in a larger depth of more than 60 km. In shallower levels the resistivity of the crust was reproduced quite well with values between 800 and 1400 \( \Omega \) m.
Fig 4.13. 2D inversion result of model SP1a. The inversion shows the resistivity distribution over the depth along profile C. rms = 1.608

In model SP1d the width of the gradient zone was decreased, i.e. the gradient was steeper (App. B). The 2D inversion result showed a strong increase in lateral heterogeneity in the conductivity zone. At the edges conductive nests were separated from the central conductive zone and also the middle part had zones of decreased conductivity. The overall appearance of the anomaly was not as compact as in model SP1a. The high conductivity zone only reached a depth of 120 km, which was considerably less than in the inversion result of model SP1a. Additionally, the bottom area of overestimated conductivity of 10 to 100 $\Omega$m right below the high conductivity zone was less pronounced. “Bow wave” structures of very high resistivity (5000 to 17000 $\Omega$m) appeared close to the high conductivity zone. The resistivity of the uppermost levels was reproduced quite well, except a layer right above the left half of the conductivity anomaly that reached 2000 $\Omega$m.
Fig. 4.14. Model SP1c: original model and 2D inversion result. The white line in the cross section on the right denotes the level of depth that is shown in the map view on the left. The inversion shows the resistivity distribution over the depth along profile C. rms = 0.791.
Tests with models of reduced resistivity contrast between the conductive shear plane and the resistive host rock were performed as well. The host rock resistivity was decreased to 500 $\Omega$m (model SP1b) and 100 $\Omega$m (model SP1c). The resistivity of the shear plane was 1 $\Omega$m as before. In the inversion result of model SP1b, the whole conductivity structure fell apart and formed zones of very low resistivity (<1 $\Omega$m) separated by regions of low to intermediate resistivity (10 to 50 $\Omega$m). A clear depth extent could not be stated for the whole structure but rather for the single conductive spots. They reached depth between 70 (eastern part) and 170 km (central part). The top of the conductive zone was located at the right position but did not show a sharp transition to high conductivity as in the models seen before. The resistivity of the uppermost layers was reproduced very well (300 to 600 $\Omega$m).

An extreme destruction of the high conductivity zone in the 2D inversion result was obtained in model SP1c with a host rock of a resistivity of 100 $\Omega$m (Fig. 4.14). There was no compact high conductivity zone at all anymore. Thus, this model could not be used to characterize the high conductivity zone obtained from “real” data (Schwalenberg, 2000). However, it is interesting that the shear plane clearly appeared in the 2D inversion result as an elongate slightly west-dipping structure. The depth and the thickness were very similar (virtually identical) to the originally given model. Apparently, little resolution of the lower boundary of the conductive structure was especially a problem of systems with high resistivity contrast. The decrease in resistivity contrast improved the presentation of the original structure, especially considering the resolution of the bottom of the conductivity structure. The resistivity distribution was patchy throughout the conductive structure and somewhat overestimated. Two large high conductivity spots appeared beneath the west-dipping structure that weakly reminded of the great depth extent of the conductive bodies of the models of large conductivity contrast. They were separated by a field of very high resistivity (>1000 $\Omega$m, strongly overestimated). High resistivity fields (>500 $\Omega$m) also appeared besides the high conductivity spots below the edges of the shear plane in depth larger than 70 km. In the uppermost layers the resistivity was again reproduced very well.

The investigation of the set of models containing three shear planes piled up on top of each other led to very similar results in the 2D inversion. The high conductivity structure
of model SP2a is only slightly thinner than in SP1a and the edges show conductivity fields more separated from the center. The high-resistivity “bow waves” and the intermediate resistivity zone beneath the high conductivity zone are virtually alike. The behavior of the system in the setting with a resistivity contrast of 100 Ω m (host rock) vs. 1 Ω m (shear plane) is the same for model SP2c and model SP1c. These results demonstrate that MT methods only ascertain the existence of horizontally interconnected conduction paths. They do not resolve the internal structure of the conductive system; in particular vertical inter-layering of highly and less conductive layers is not differentiated. Generally, the models of highly conductive shear plane(s) embedded in a highly resistive host rock of 1000 Ω m reproduce the inversion model obtained by Schwalenberg (2000), Brasse et al. (2002) among others very well.

Fig. 4.15. 2D inversion result of model SP2a. The inversion shows the resistivity distribution over the depth along profile C. rms = 1.74
Summary

Several geologic structures that may form a high conductivity zone measurable with MT-methods were tested. The experiments included variations of size and position of the conductive bodies as well as variations of the conductivity contrast between host rock and conductive structure. These models were 3D forward computed and then run with a 2D inversion using WinGLink. The inversion results were compared with 2D inversion results obtained from measured data of the ANCORP profile to find out which principal concepts under which conditions could reproduce the high conductivity zone beneath the Altiplano. Generally, the 2D inversions had the tendency to underestimate the resistivity of the conductive structure and to compensate this with overestimation of the host rock. No model could be found that produced intermediate resistivity of a few 100 Ωm for the host rock and a high conductivity zone with roughly 1 Ωm. Dikes only reproduced the compact high conductivity zone when the inversion procedure was run with a high degree of smoothing (Uniform grid Laplacian operator). Extended melt zones much better reproduced the anomaly in the inversion result with a lesser degree of smoothing (Standard grid Laplacian operator). Because of the very good resolution of the lateral extent and of the depth position of the top of the conductive structure a good idea of the size of the model structure required to reproduce the high conductivity anomaly based on ANCORP-data could be developed. Only models with an extreme resistivity contrast between host rock and structure reproduced an adequate anomaly in the inversion. Models with increased resistivity of the structure did not reproduce a compact large-scale conductivity anomaly in the inversion. Shear planes also were a modeling concept that for large resistivity contrasts led to an adequate high conductivity zone in the inversion result. Decrease of the resistivity contrast caused strikingly good reproduction of the originally given structure, but there was no similarity to the ANCORP data. It did not make any difference if a single shear plane model or a multiple shear plane model were employed; the inversion results were alike. There was also conspicuous similarity between the inversion results of the models of melt zones and the ones of shear planes. Thus, in conclusion it is technically possible to reproduce the given anomaly with three models of fundamentally different geological concepts. The shape characteristics and the magnitude of the anomaly obtained from inversion strongly depend on the inversion.
parameters, e.g. the regularization algorithms set up for calculation. Data from other geo-
scientific methods are required to finally decide, which geological concept best explains
the data all together.
5 Application to the Andes: Discussion of petrologic and tectonic constraints

In the results of the modeling it was shown that two different geological concepts can lead to very similar results in the MT modeling procedure. Obviously, it is not possible to simply characterize the geologic structure from the pure inversion results. It is therefore very important to include results from other geophysical and geological analyses to determine, which conductive structure most likely causes the conductive anomaly. This is not surprising, since a geological structure with complex geological, physical and chemical characteristics that causes the anomaly has to be described as a complex feature of the crust. Simplistic views only uncover parts of an aspect. In this chapter the results of the modeling are applied to the specific geological setting of the Central Andes. It is discussed, which one(s) of the purely technical solutions of the problem “how to produce a high conductivity zone” can explain the high conductivity zone beneath the Altiplano. In the discussion of the possible conductive materials in chapter 4 it turned out that melts are most likely the material responsible for the conductive anomaly. Thus, the interpretation of the high conductivity zone in terms of geological structures will mainly concentrate on the presence of melt.

If one tries to interpret the high conductivity zone beneath the Altiplano based on the existence of large amounts of melt, one has to explain what has caused the fusion, how the melt has migrated there and how it is stabilized in a depth range of 20 to 60 km of depth. In this relation it is also necessary to clearly distinguish between “magma” and “melt”. Melt is defined here as solely the liquid part of a partially (or totally) molten system. Magma is melt that contains solid particles of a partially molten system. Therefore, magma may be very heterogeneous and its physical properties, especially density and viscosity, may differ considerably from those of the pure melt. The magma density for example easily could be greater than the theoretical melt density and probably keep the magma from rising even though the pure melt would theoretically be able to rise.

A few arguments favor that the melt beneath the Altiplano plateau has to be intermediate to felsic in composition. First of all a large volume of basaltic melt is not
able to rise buoyantly through crustal material up to the depth range considered in an active collision orogen because its density is too high compared to the surrounding crustal rock. Further, basaltic melts have a temperature of around 1200°C. A melt volume like the one proposed would rapidly melt the surrounding host rock in a wide range. Finally no evidence for a large basaltic body has been found from gravimetric measurements. A possible source of felsic to intermediate magma could be the boundary between the crust and the mantle. Mafic to ultramafic mantle magma rises buoyantly through the mantle to the crust-mantle boundary or intrudes the lowermost levels of the crust. Somewhere at this depth it stagnates, since the density of the olivine-dominated mantle magma is larger than the density of the more felsic feldspathic crustal rock and therefore further magma ascent is blocked. Thus, the mantle magma basically underplates the crust, spreads out horizontally and only the heat of the magma can advance upwards. Provided that the total amount of mantle magma underplating is large enough the heat advancing into the crustal base could cause fusion in the material there. Since the crustal melt is more felsic than the mantle melt, its low density allows it to ascend through the crust. Also differentiated magma from mantle material could continue this path (Stolper & Walker, 1980): The successive enrichment in SiO₂ and other light oxides in the melt due to early crystallization of Mg-olivines, pyroxenes, and Fe-oxides in a basaltic melt lowers the density of the residual melt, but also increases its viscosity. Small amounts of water in contrast decrease to some extent the density as well as the viscosity of the melt. The new formed melt ascends through the crust until it becomes stabilized at a certain depth. The stability of melt depends on the density contrast between melt and country rock, the volatile content of the melt, the melt viscosity, the volatile overpressure in the magma chamber, and the width of conduits through which the melt could pass (Stolper & Walker, 1980). A rough estimate based on Buntebarth (1980) (see App. C1) gives a total thickness of the underplating magma layer of at least 1-2 km to provide sufficient heat for long enough a time (5 Ma) to fuse the crust. Smaller batches would cool too rapidly as that they could maintain a melt-containing system for a long time. Further sources of heat in the orogen may include increased radiogenic heat production due to the increase of crustal thickness, shear heating based on internal active deformation, and intrusion and
cooling of arc andesites (Babeyko et al., 2002). These all together can cause wide spread fusion.

A temperature-depth setting in the crust was generated based on the considerations above (Tab. 5.1). The calculation was based on Turcotte & Schubert (2002), eqn. 4-31:

\[ T = T_0 + \frac{q_m \cdot z}{k} + \frac{(q_0 - q_m) \cdot h_x}{k} \cdot (1 - e^{\frac{-z}{h_x}}) \]  

(5.1)

\[ T_0 = 10°C \text{ (surface temperature)} \]

\[ q_m = 30 \text{ mW/m}^2 \text{ (heat flux from the mantle to the base of the lithosphere)} \]

\[ q_0 = 100 \text{ mW/m}^2 \text{ (surface heat flow, e.g. Brasse et al., 2002)} \]

\[ h_r = 10 \text{ km (length scale for the decrease in H)} \]

\[ H = H_0 e^{z/h_x}, H_0 = \text{surface radiogenic heat production rate per unit mass, } z = \text{depth} \]

\[ k = 2.5 \text{ W/mK (thermal conductivity)} \text{ (Stephan Sobolev, GFZ, pers. communication)} \]

\[ z = \text{depth (m)} \]

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Table 5.1 Temperature values as functions of crustal depth based on the equation (5.1).

The aim is now to develop a petrologically and tectonically reasonable concept of how the melt or magma respectively migrates through the crust and forms bodies that are detected as conductive regions by MT. The transport mechanism as well as the emplacement structure has to be macroscopic in scale, i.e. diapirs and dikes, because
migration in the mere scale of grains is not efficient enough to produce conductive volumes as being detected. To characterize the crustal melt transport and emplacement mechanisms the behavior of the melt in the crust has to be investigated. This is best done by examining the “end members” of a certain physical quantity, which requires in turn the “end member values” of the contributing physical quantities. For example, to estimate the maximum buoyant ascent velocity the maximum possible (i.e. reasonable) density contrast and the minimum possible crustal viscosity have to be taken for the calculation. These estimates give valuable indications about processes likely to occur because in cases in which the most optimum values already rule out a process the intermediate ones won’t do any better. I especially apply extreme values for the density contrast in the calculation based on the following considerations. First of all the melt I consider is only a partial melt, i.e. only a few of the components of the original rocks are fused. According to Lucassen et al. (2001) the lower crust between 40 and 70 km of depth is not mafic in the sense of basaltic or andesitic-basaltic composition. It rather may form a homogeneous transitional composition or a heterogeneous composition with discrete felsic and mafic components (Lucassen et al., 2001). Generally the mineral components that have crystallized last are the ones that melt first. Many of the minerals that fuse to form early partial melt have fairly low densities. The minerals of early crystallization mostly have a relatively high specific weight and remain solid longest. If a rock of a certain density experiences partial melting and preferably loses the lighter components, the residual solid gains a higher specific weight. Thus, locally the density contrast of the partial melt and the surrounding residual rock is larger than it would be in the case of complete fusion of a volume of rock being surrounded by still solid rock of identical composition. Furthermore, a certain material input from the heated subducted plate may occur as well. Subducted oceanic crust may experience partial melting when the volcanic crust is relatively young and hot (Best & Christiansen, 2001). The Nazca plate is quite small, and its subduction rate has been fairly large for the last millions of years. The slab may well be hot enough to partially melt in the depth and introduce felsic components into the mantle and lowermost crust that ascend in the course of time. Thus, a fairly large density contrast for estimation of extreme quantities based on relatively high residual rock
densities, which are not supposed to characterize the whole lower crust but rather local conditions, and low density values for partial melts are a reasonable approach.

5.1 Conductivity associated with dikes

One approach to describe the migration of melt within the crust is the motion along fractures not only microscopically but also macroscopically. Dike propagation is a complex process and is determined by quite a number of parameters: magma viscosity (depending among others on water content), magma temperature in the source, the source pressure that triggers diking, extensional stress due to the tectonic setting, and a fairly continuous supply of magma are important factors of dike propagation (Rubin, 1995).

Before a dike can start to grow the melt has to segregate and to collect in a reservoir with a high percentage of melt, from which the dike can start and then is fed continuously. The melts produced by fusion of crustal material tend to be felsic as is indicated by rhyolitic and dacitic ignimbrites found on the Altiplano and farther west in the Western Cordillera. Since even wet granitic melt has a fairly high viscosity and therefore a low flow velocity through the rock pores, the melt in dikes has to be extracted from pores and melt bodies just in the vicinity of the base of the propagating dike (Weinberg, 2004). Thus, for large dikes large melt reservoirs with high melt fraction and highly interconnected networks are necessary.

The velocity of vertical ascent depends significantly on the magma’s viscosity, which in turn determines the dike width (Wada, 1994). Weinberg (2004) estimated a dike width between 2 and 20 m to be necessary for an efficient transport of felsic magma. Granite dike with widths of meters to several hundreds of meters have been observed. Clemens (1998) provides a formula to calculate the ascent velocity:

\[ v = \frac{\Delta\rho g \omega^2}{12 \eta_m} \]  

(5.2)

with \( \omega \) being the dike width, \( \eta_m \) the magma viscosity, \( g \) the gravitational acceleration, and \( \Delta\rho \) the density difference between magma and wall rock. Especially magmas with
relatively low viscosity ascend through a dike rapidly as long as enough magma is supplied from the source. The influence of the density contrast seems less pronounced.

Taking

\[ \omega = 10 \text{m}, \]

\[ \eta_m = 10^5 \text{ Pas (wet granite, Hall, 1996)} \]

and

\[ \Delta \rho = (3300-2170) \text{ kg/m}^3=1130 \text{ kg/m}^3, \]

the dike would ascent with 0.94 m/s. A dike of 100 m width (all other parameters are the same as above) would ascend with 94.17 m/s, which is rather catastrophic, but does not seem impossible. A 100 m wide dike of dry granite ascends with a velocity of only \(9.2 \times 10^{-4}\) m/s (values for viscosity and density from Hall, 1996).

Very large source pressure is needed to start and maintain the propagation of a relatively wide dike filled with felsic magma (Wada, 1994). The higher the source pressure is, the wider the dike can become (Rubin, 1993). The pressure of the reservoir and in cases of rapid melt segregation also a steep temperature gradient within the host rock close to the reservoir may weaken the rock by the formation of numerous small and randomly shaped and oriented “pre-dikes” within a marginal shell forming around the zone of partial melting before the “real” diking starts (Clemens & Mawer, 1992). According to Clemens and Mawer (1992), the initial process of diking could just be a connection of a few of these small “pre-dikes” that are located close enough and oriented appropriately relative to the local stress field. As the rock splits the fracture filled with granitic magma becomes self-propagating because the density contrast between magma and rock produces a buoyant force and because of the fluid pressure present in the reservoir. Additionally, at the tip of the fracture decompression due to dike propagation causes hydrous magma to expand due to the increase in partial molar volume of water dissolved in the melt. This supports the fracture propagation even more (Clemens & Mawer, 1992).

However, for several reasons it is not very likely that dikes cause the high conductivity zone beneath the Altiplano. First of all diking depends on high intensity tectonic stresses, which are not expected to be present in the lower crust of the Altiplano in the recent setting. Structural geological analyzes have shown that most of the compressive stress applied to the Altiplano resulted in large thrust systems in the west
and central Altiplano (17 Ma) and later mainly in intense uplift with little internal deformation (around 10 Ma) (Elger, 2003). Thus, there is no reason to expect recent dike propagation. This would be necessary because to serve as a conductor a dike has to be filled with conductive material, e.g. with melt. Since dikes cool very rapidly, the process of dike propagation would need to have occurred very recently. A granite dike of 10 m width would cool from 1000°C to 650°C in only approximately 100 years if the temperature of the ambient rock were roughly 640°C (model of cooling in Buntebarth, 1980; see App. C1.). A dike of 100 m width would need roughly 10,000 years at the same conditions. These estimates do not include the latent solidification heat (L) the granite in the dike transmits into the host rock as it solidifies at 650°C (wet granite). L is of the order of 300 to 400 kJ/kg. Thus, the dike’s cooling time will increase by several tens to hundred years due to the heating of the surrounding rock. Considering this cooling rate a dike has to rise rapidly in order to develop a significant size. Furthermore, diking, i.e. fracture propagation is based on the ability of the crustal material to fracture. This means that in regions of increased ductility this type of magma transport is rather unlikely. Fracture propagation is therefore more typical for shallower crustal levels than for the ultimate crustal base. The conductivity anomaly from this point of view can be explained with dikes in a depth range from 20 to 30 km. Immediately beneath the dikes, a huge reservoir with a very high melt fraction had to be located to maintain the dike as a conductor. The bottom of the high conductivity zone could not be resolved clearly with MT modeling or measurements. Based on these data it is not possible to decide if a conductive melt reservoir appears beneath the dikes. The MT data only would produce the integral conductance of the volume considered, but no details. However, Wigger et al. (1994) did not find any indication of dikes but detected a fairly wide zone of increased attenuation of seismic waves in a depth of 20 km, which was interpreted to be the result of the presence of melt. This is not consistent with dikes in a depth range of 20 to 30 km and a melt reservoir beneath them. Finally, a dike is a structure of extreme shape and therefore causes conductance anisotropy. The intensity of the anisotropy depends on the dike width and the overall conductivity of the dike and the wall rock (i.e. on the conductivity contrast between dike and wall rock), and probably also on the existence of sedimentary layers. Being filled with melt a dike would provide a direction of great
conductance parallel to the strike and another direction of basically no conductance perpendicular to the strike. This would produce a distinct horizontal anisotropy in the detected signal of MT. Anisotropy of this type has not been observed in the Altiplano. Assuming the existence of dikes, isotropy of the MT response is only possible when numerous dikes are oriented in different directions, that is when they either form a radial system of dikes or when they are distributed randomly. The first type of setting is typical for extensional or stress fields, probably in connection with plume ascent as has been observed in Iceland. The latter is simply unlikely.

The modeling results demonstrate a strong restriction regarding the position of the profile on the dike. Only for a profile sitting approximately symmetrically in the middle of the 20 km wide dike with the strike of the profile happen to be roughly equal to the strike of the dike is was possible to reproduce a high conductivity zone of significant extent. However, dikes are usually not 20 km wide. It could not be investigated how strong the anomaly in the inversion result of a considerably smaller dike would be, but it is clear that it would be significantly weaker than the modeled one. Furthermore, conductive structures have been fund in other profiles in similar depth (20 km), e.g. on that along 22°S. The shapes of these structures deviate from the ones obtained for the dike models and they are smaller, but the conductivity is fairly high as it only could occur for profiles sitting centrally above the highest conductive zone. It is not very likely to accidentally place several profiles sufficiently close to the center of a dike in 20 km depth several times to explain the high conductivity zone over the whole area with a system of dikes. Thus, even though under very particular conditions a dike may purely technically spoken be a solution of the imaging of a high conductivity zone, it does not provide a meaningful explanation of the anomaly in terms of geology. It is therefore concluded that the concept of dikes has to be dismissed from the list of solutions.

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1 It has been a popular idea for a long time that MT does not detect vertical structures because the current flows horizontally within the subsurface. This is only true for conductive structures that have no large extension in any dimension (small plutons) or merely a large vertical extension and very small horizontal ones (pipes) that solely allow for vertical currents and therefore cannot be detected. As soon as the structure has an additional horizontal extension as dikes have the current can flow horizontally within them and thus they ARE detectible by MT.
5.2 **Conductivity associated with extended melt zones**

The models based on the pluton-like conductive bodies reproduced the high conductivity zone obtained by Schwalenberg (2000) very well when the conductivity contrast between the conductive body and the surrounding host rock was very high (1 vs. 1000 $\Omega$ m). Therefore a conductivity zone of compact shape can technically be a solution of the problem what structure my cause the conductivity anomaly beneath the Altiplano. In chapter 4.3.2 conductive bodies of fairly large extent and compact shape were examined. The geological concepts of these different structures are very particular. However, the modeling procedure is for all of these models the same. In the following the different concepts are investigated in terms of geological significance and considering the interpretation of the conductivity anomaly.

**Melt transport via diapir**

Stoke’s law describes the buoyant rise of bodies through a matrix. Having a Reynold’s number ($Re$) significantly smaller than 1 indicating laminar flow the Stoke’s equation is (Turcotte & Schubert, 2002, eqn. 6-229):

\[
v = \frac{2 \cdot g \cdot r^2 (\rho_m - \rho_b)}{9 \cdot \eta}, \quad \text{with} \quad Re = \frac{2 \cdot \rho_m \cdot v \cdot r}{\eta}
\]

(5.3)

where $v$ is the velocity of the migration, $\rho_b$ the density of the rising body, $\rho_m$ the density of the matrix, $g$ the gravitational acceleration, $r$ the radius of the ascending body, and $\eta$ the viscosity of the matrix. Based on this model (see App. D) an andesitic melt drop with a radius of 10 cm would theoretically need $1.6 \times 10^{13}$ years to ascend ten kilometers. A granitic melt drop of 10 cm radius would need $1.2 \times 10^{13}$ years for ten kilometers. For spheres of a radius of 100 m the ascent time are $1.6 \times 10^7$ and $1.2 \times 10^7$ years, respectively. This is still way too slow. A melt body needs a radius of 500 meters to reach an ascent velocity in the order of 1.5 cm/year (andesite) or 2 cm/year (granite).

As long as the melt sphere moves through depth levels, at which the temperature is high enough to keep the magma liquid due to the geothermal gradient the cooling magma
can move upward slowly and also act as a conductor. The cooling process is fairly rapid: a sphere of a radius of 100 m would theoretically cool from 1000°C to 650°C in roughly 520 years when the host rock has a temperature of approximately 640°C (based on the temperature distribution given above this is equivalent to a depth of 30 km) (estimated from Buntebarth, 1980, see App. C2, not including the effects of latent heat). At this temperature even wet granite melt crystallizes and hardly moves any further. Dry granites and melt of somewhat more mafic composition would crystallize at much higher temperature and get stuck in larger depths. Since the cooling process is much more rapid than the ascent the melt sphere rather will continuously adjust its temperature to the temperature of the host rock during its ascent than cool stepwise. However, a diapir in 30 km depth does not explain a conductivity anomaly at a depth of 20 km. Assuming larger diapirs with radii of 1000 or (rather unusual) 10000m does not solve the problem. A diapir with a radius of 1000 m would still rise too slow compared to its cooling time. A diapir with a radius of 10000 m on the other hand cools very slowly, but theoretically rises too fast. The body would either need to have started to rise quite recently to be at the depth level of the HCZ, which is not supported by the recent state of basal heat flow, or it would have reached the surface already and erupt as volcanism, which is not observed on the Altiplano at the moment. A melt body of this size also should cause large scale high-degree-fusion in the crust. The heat flow measurements performed in this area did not prove this.

Considering the fact that compressible melts react on pressure release and on temperature loss due to ascent does not change the situation. Pressure release leads to volume expansion and thus to a decrease of density. Temperature decrease acts in the opposite way causing volume decrease and therefore density increase. The compressibility of melt ($\beta = 1/K$, K being the bulk modulus) is of the order of magnitude of $10^{-10}$ Pa$^{-1}$. The coefficient of thermal expansion ($\alpha$) is between 1.6 *10$^{-5}$ (Gabbro) to 2.4*10$^{-5}$ K$^{-1}$ (Granite; Turcotte & Schubert, 2002). Even though $\alpha$ is significantly greater than $\beta$, pressure has the larger influence on the depth-dependent volume/density behavior of the melt. The reason is that the temperature changes by a few hundreds to thousands of degrees over a depth range of around seventy kilometers, whereas the pressure variations in this depth range are of the order of hundreds to thousands of megapascals (see App. E).
Thus, a theoretical ascending melt body of approximately spherical shape or as often suggested upturned tear shape would start to rise from the depth in a significantly compressed mode and expand due to pressure release while ascending. That means that the density difference at the base of the crust is even smaller than estimated above, which slows down the calculated Stokes buoyant ascent of the magma even more. Therefore the times of migration become way too long for proving Stokes ascent being a realistic mechanism of magma transport over distances of several tens of kilometers. Besides this, diapiric magma ascent is not only a matter of density contrast but also (and probably even much stronger) of viscosity contrast. A diapir only can rise when the viscosity contrast of diapir and matrix does not exceed a certain maximum value (probably a ratio of melt viscosity to matrix viscosity of 1/100-1/10000; Podladchikov & Connolly, 2004, pers. comm.), i.e. when a diapir or magma pulse moves within a partially molten, i.e. highly ductile system. Analogue and numerical models resulting in mushroom shaped magma ascent simulate a magma pulse intruding into an existing magma chamber rather than a diapir ascending through solid crust. Structures observed in the field very likely developed this way. The explanation is that the matrix in this case is able to flow around the rising body due to its relatively low viscosity. In that way the matrix material is able to get out of the way of the moving body and replace the volume the body has left behind it. This is the case when a system contains significantly more than 50% melt and the crystals sink or rise through the magma. For the lower crust ($10^{18}-10^{23}$Pas) a much higher viscosity contrast has to be expected ($10^9-10^{15}$Pas). In this case the magma is rather transported via dikes or through a narrow irregular path than via diapir (Ramberg, 1981).

Marsh (1982) among others developed a model of “hot-Stokes”, in which a hot diapir weakens the surrounding cool crustal rock within a narrow marginal zone\(^2\). The diapir rises buoyantly through this soft low-viscosity shell by a distance of approximately the shell thickness and then again softens the crustal rock to form a new low viscosity margin to repeat the cycle. However, this is a very inefficient process that can explain the ascent of diapirs over a distance of a few tens to hundred meters, but surely not the ascent of

\(^2\) In this narrow zone the crustal material can be characterized as Newtonian fluid, and Stoke’s Law can be applied in this restricted volume of the weakened shell.
tens of kilometers because also in this concept the main problem is that the diapir would lose way too much heat to ascend a long distance.

The situation evaluated above makes it difficult to develop a model that explains purely diapiric magma ascent. Since regular-sized diapirs rise slowly and cool continuously they crystallize after short distances. Weinberg & Podladchikov (1994) developed a model, in which because of high buoyancy stress of a diapir the crustal viscosity becomes low and the diapir ascends fast. In this model they consider the crust as power law fluid, which is a fluid that requires less stress for deformation with increasing shear rate, i.e. its viscosity decreases with increasing shear rate. As a diapir ascends, crustal material has to descent to replace the volume, which establishes convective flow within the crust. Weinberg (2004) describes two consequences this would have. Firstly the upward flow may induce decompression melting of the hot aureole of the diapir, which would increase the amount of magma and therefore buoyancy of the diapir. Secondly, the downward flowing rocks replacing the diapir would bring new material into the source, which probably melts and rises forming a new diapir. However, no evidence has been found for plutons ascending as diapirs. There are no tracks of passing, blind diapiric aureoles or zones of deformation attributed to the passage of a diapir such as pipe-like high-grade shear zones, in which the cores have moved upward (Bateman, 1984; Clemens, 1998).

Summarizing we see that the ascent of magma through the crust cannot be described with the Stokes approach when transport of great amounts of magma over large distances have to be explained.

**Concept of ballooning diapirs**

Some authors (e.g. Dietl & Koyi, 2002) suggested, for example in the concept of “ballooning plutons”, that felsic plutons might be fed by more or less extended dikes, through which repeatedly magma pulses pass. The pluton would not ascend as a whole huge diapir, but form stepwise in approximately the position it finally crystallizes, even though it still might rise a small distance as a whole body by stoping. In this case the conductive structure is purely a feature of emplacement rather than a feature of magma ascent. The ascending magma is assumed to spread out horizontally in certain depths due
to density effects when the magma filled fracture reaches a crustal level where the magma density is the same or higher than the weighted mean density of the country rock (Clemens & Mawer, 1992). Magma ascent can also be stopped by emplacement-prone crustal structures, such as highly ductile zones. Subsequent magma pulses intrude the not yet crystallized magma body and cause it to thicken vertically and to grow laterally. In this way the setting may develop large lateral and vertical dimensions in the course of time. The heated wall rock of the pluton is thought to be pushed away and partly may be consumed by the expanding magma chamber. To solve the problem of providing space for the arriving magma concepts such as hydraulic roof lifting and floor-lowering of the growing magma chamber have been developed (Petford et al., 2000). In the concept of ballooning diapirs it should be possible that magma intruding an already existing magma chamber widens the conduit that has been produced by the first portion via diking because of weakening the host rock by dike formation and by heat conduction. This means even if the dike did not remain open during a possible gap in magma transport the crust is weakened enough to effectively provide a pathway for later magma transport. The fracture could provide a pathway for relatively slim, elongate “diapirs” to ascend relatively fast into the magma chamber, since the crust immediately around the dike could behave ductilely due to partial melting. Note, that the ductile crust in this case would be spatially very restricted to the adjacent zone of the conduit and therefore would not be a feature of the whole crust in the depth range considered. Having subsequent intrusions passing the same conduit toward the same magma chamber also could explain why only very few feeder dikes to granite plutons have been recognized. Estimates of volumetric filling rates between 0.01 and 100 m³/s yields emplacement times of a few days to roughly a million years for plutons less than 1 km to more than 100 km across (Petford et al., 2000). The lower values for volumetric filling rates are probably the more realistic ones for felsic magmas because of the magmas’ high viscosity values. The high viscosity of felsic magma also could be the reason, why granitic dikes are so rare compared to more mafic ones. The relatively low rising velocity of viscous magmas probably makes them more sensitive to crustal settings that together with decreased buoyant force stop the ascent. Subsequently arriving magma supplies additional heat and buoyancy, but this may not cause considerable further ascent but rather slow down the
differentiation and crystallization process, and increase the pluton size. Mafic magmas are less viscous and may therefore easier pass those crustal settings due to their ascent velocity and crystallize as dikes without forming large pluton-like structures. All these diapir models have one interesting thing in common. Magnetotelluric measurements only can detect the source in the lower crust and the recent emplacement position in the middle crust. The pathway of the magma is not detectable.

A pluton-like body of partial melt is in terms of conductance chiefly isotropic. The difficulty with plutons in MT measurements is that they hardly can be detected structurally if they are too small. They only may slightly increase the integral conductance of the observed volume. Plutons have to be very large (in the order of 10 km) to be detected by MT as separate conductive structure at a depth of 20 km. Their horizontal dimensions are the crucial factor for magnetotelluric recording, since plutons do not form extended large-scale networks, which would increase the effective volume of conductive, interconnected partial melt.

As seen in chapter 4 the MT modeling reproduces the lateral extent and the depth of the top of a conductive structure very well. This means that a magma chamber would have to be of adequate size to produce a high conductivity anomaly as found from ANCORP data. Considering this amount of melt caught in a pluton of this size the problem of stabilization of the system arises. Already an amount of 20 vol% melt is commonly thought to cause the melt to move out of the system and rise into higher levels. A body of significantly higher melt concentration as typically assumed for plutons would not remain stable within the crust if it had this size. Due to buoyant forces it would ascend and finally erupt at the surface. Eruptions are not observed in the Altiplano at time. Thus, a pluton does not provide a meaningful solution of the problem how to interpret the high conductivity zone beneath the Altiplano.

Concept of partially molten system based on migmatization

As already discussed in chapter 2 the Andes are the product of the collision of the continental South American and the oceanic Nazca plates. The Nazca plate is subducted under the South American plate. The result of this collision is the Andean mountain belt.
The subducting plate transports a tremendous amount of water and other fluids into the mantle wedge. In the course of the subduction the Nazca plate becomes heated intensely and the fluids are released. The fluids themselves are heated on their ascent through the wedge and the crust and cause melting in the crust along their way. In common sense the melt rises to the surface and erupts feeding a chain of volcanoes in a certain distance from the trench. The Western Cordillera at approximately 68.5°W forms the recent active volcanic belt of the Andes. However, the high conductivity zone beneath the Altiplano is located far behind the volcanic chain (67-68°W at 21°S) and no recent volcanism is related to it. Large-scale crustal partial melting initialized at a depth of 30-40 km may be a possible explanation for this phenomenon: hot fluids from the slab ascend through the crust and cause partial melting (Schilling et al., 2005, in prep). Note, that the amount of heat required here is significantly less than in the example of the diapir mobilization. Also the amount of melt is fairly low, since in this example no distinct melt body is formed. Conductivity values were modeled with highly conductive silicate melts of five to ten S/m using the Hashin-Shtrikman model. These values are very reasonable for wet granitic melt. According to Schilling & Partzsch (2001) a minimum melt content of 14 vol% is necessary to produce the given conductivity (see also chapter 3.4, Fig. 3.6). This value is based on the assumption that the melt is interconnected perfectly. Thus, the real amount of melt must be somewhat higher due to the fact that parts of the melt very likely are located in pockets and dead ends. The real amount of melt may be close to 20 vol% (and probably up to 25 vol%). At an amount of 20 to 30 vol% of melt it is commonly assumed that the melt starts to buoyantly move out of the partially molten system and slowly rises into higher levels. This mainly occurs by migration along grain boundary channels and grain-size fractures. The process could within certain time ranges lower the melt content enough to affect the conductivity of the system. However, this process is not very pronounced at this melt fraction but rather plays a role at higher melt fractions. Thus the melt is widely stabilized within the crust. The stability of the melt may additionally be explained with the fact that here a felsic melt is made of a relatively felsic crust as has to be expected at the given depth. In this case the buoyant ascent of the melt is fairly low because the density difference between melt and rock matrix is very small. Additionally, crystallization decreases the melt content of the system significantly. To preserve the
melt content, newly formed melt has to permanently replace the amount of melt that has escaped and crystallized. So, the system of partially molten crust may have established an equilibrium between melting and crystallization, and also between fluid input from the slab and fluid loss into the upper crust due to diffusion and melt-internal flow. In the course of time the upper boundary of the partially molten zone may have moved up to a depth of 20 km, and also the lower boundary can have moved away from the initial migmatite center due to the continuous heat inflow (Schilling et al., 2005, in prep.). This would be a classical example of large-scale migmatization. Migmatites are commonly interpreted to be partially molten rock, in which the granitic partial melt has not segregated from the complementary residue. They form layers, lenses, pods and irregularly shaped masses of granite that are mingled with mafic metamorphic rock (Best & Christiansen, 2001). Migmatites do not intrude anywhere and only deform by high viscous flow due to p,T-conditions. Partial melting (i.e. migmatization) is a very common process in the crust and therefore is an important factor in the process of crustal differentiation (Kriegsman, 2001).

Fig. 5.1 Melt distribution in a migmatite. The melt forms a system of horizontal layers that are interconnected locally by vertical bridges. The melt-containing crust keeps its solid framework.

A migmatite body should mainly consist of horizontal layers, which are sporadically interconnected by vertical bridges (Fig. 5.1). This would produce a strong isotropic signal in MT measurements, since the horizontally oriented currents parallel the conductive dimension. The overall conductance value is an integral of the highly conductive melt
layers and the little conductive rock matrix. Thus, a homogeneous conductive structure in the crust would be detected with MT.

The system will be maintained at this stage as long as constant amounts of fluids and heat are released from the slab to compensate the heat and fluid loss of the system, since melt and other fluids tend to rise through the crust. This is probably the status quo that has been observed by MT measurements. The concept of low-degree fusion of the crust also matches the seismic data that had obtained attenuation and high $v_p/v_s$ ratios in the region of the conductive zone that are interpreted to be related to the presence of fluids, e.g. melt. The existence of $v_s$ data from this volume indicates that the crust in the observed depth is not molten to a high degree and that the crust is not disaggregated there.

The presence of melt appears to be indicated by other geophysical results as well. Heat flow and gravity measurements support the interpretation of the high conductivity anomaly as being related to melt. The Altiplano is characterized by increased surface heat flow density of approximately 100 mW/m$^2$ (Springer & Förster, 1998). A system that contains melt introduces a certain amount of heat into the surrounding host rock, which is finally measured as heat flow anomaly. Since melt has a lower density than the rock from which it originates, the total density of a partially molten rock is lower than the density of the pure solid rock. This was confirmed by gravimetry measurements that obtained a gravity low for the area of the expected partially molten system. One more aspect makes the migmatite being the best interpretation. Taking together all MT data that had been measured on several profiles on the Altiplano-Puna plateau it turns out that the distribution of the conductive zones is not uniform but rather patchy. Large areas are not conspicuously conductive. Spots scattered throughout the plateau exhibit conductivity anomalies but of apparently smaller extent and intensity. It is impossible to describe the pattern in more detail because the number of profiles and therefore the overall measurement density does not provide enough data. However, the irregularity of the distribution of conductive spots is obvious.

Migmatization only occurs when several parameters such as heat input, fluid abundance, chemical/mineralogical composition of the rock all together happen to support the process of migmatization (Frank Schilling, GFZ pers. comm.). As soon as
one parameter does not meet the requirements of migmatization, no migmatite develops. Thus, in closer view migmatization is a rather localized phenomenon: At places where the combination of the parameters sustains migmatization melting occurs, but there are areas in between where no migmatization takes place. Therefore, a migmatite of crustal scale is heterogeneous. Areas of intense migmatization would neighbor zones of less or no migmatization.

A migmatite is a highly dynamic structure, in which the pattern of high and low degree migmatization changes in the course of time. In the case of increased heat input due to changes (flattening) of the subduction angle and/or increase of the subduction rate of the Nazca plate the rate of melt production increases. In this situation the migmatite may become unstable and the melt may ascend to the surface erupting in a wide-spread ignimbrite volcanism. This may have occurred 10-2 Ma ago when due to peak activity of heat flow from the mantle and the other heat producing sources around 10 Ma ago the Altiplano-Puna-Volcanic-Complex (APVC) was formed with ignimbrites, which are considered to be crustal melts (Scheuber, 1994; Babeyko et al., 2002). The distribution of the ignimbrites of the APVC is perceptibly patchy. The present state of the migmatite may be a state of ripening, probably towards a new episode of ignimbrite volcanism in a few million years.

Based on the evidence it appears most likely that the high conductivity anomaly is caused by a large scale melt system. The migmatite model is very attractive insofar as the melt acting as conductor forms right in the position where it is found. No transport of great melt volumes over large distances is necessary. Small-scale melt flow within the migmatite does not influence the overall appearance of the migmatite. The stability of the melt is explained even though a large volume is considered. The conductivity is basically due to the melt in the migmatite, since water in the approximately 900°C hot melt does not increase the conductivity of the system significantly (Robert Trumbull, GFZ, pers. comm.).
5.3 **Conductivity associated with shear planes**

The models of shear planes with conductive centers produced a very convincing conductivity anomaly that was in shape and structure similar to the one obtained by Schwalenberg (2000). There was virtually no difference between the models of one and of several separate shear planes that are piled up on top of each other. The inversion result always showed a large compact anomaly similar to that obtained for the models of extended melt zones because MT is insensitive to the internal structure of a conductor. Therefore, also a large-scale thrust system or a conductive network of shear planes technically can explain the high conductivity zone beneath the Altiplano. There is information from structural geology and tectonics as well as from other geophysical methods needed to decide if shear planes can provide a meaningful interpretation.

The Andes have formed by successive subduction of oceanic plates under a continental plate as described in chapter 2. This collision caused considerable crustal shortening and thickening which is mainly based on tectonic processes. Haschke & Günther (2003) report a ratio of tectonic to magmatic thickening of 2:1 for the Andes since the Eocene. The Altiplano region was subject to intense shortening as well. Since the plateau does not show much deformation at the surface, such as folding, a significant part of the shortening must have occurred in the subsurface (among others Elger, 2003). Besides the deformation and shortening caused by the subduction of the oceanic plates in the west, a deformation component from the east exists as well. The Brazilian shield is pushed beneath the Andes due to the westward motion of the whole South American Plate. This deformation must be concentrated along flat lying to moderately dipping shear zones besides the fault systems observed at the margins of the Altiplano. Isacks (1988) modeled the uplift of the Altiplano region based on intense shortening and crustal thickening, and on successive thrusting of the upper crust of the Andes over the foreland along an eastwardly vergent thrust belt. During the overthrusting the deformation of the brittle upper crust was according to Isacks (1988) concentrated along the eastern thrust belt. However, the underthrusting foreland continued to compress the ductile lower crust west of the thrust belt, which caused thickening of the lower crust and plateau-like uplift west of the thrust belt and also the E-W asymmetry of the plateau.
Schwarz & Krüger (1997) developed a model of an extended widely flat shear zone. Victor (2000) modeled the deformation of the western part of the Altiplano and found indications for a 30-35 km deep reaching transcrustal shear zone with a ramp-flat-ramp geometry. Beneath the western Altiplano another steep ramp is formed, which might cross the brittle-ductile transition zone down to the lower crust of 30-35 km of depth, probably connecting with the West Fissure fault zone. Steep ramps might be important features of crustal shortening because they can assimilate large amounts of shortening down to great depths (Victor, 2000).

In the case that the conductivity of the shear system is based on the presence of melts and/or aqueous fluids, the fluids have to concentrate along the shear bands that form a network throughout the shear zone as described in chapter 3.1. In the case of melt-based conductance the shear bands would need to be fairly thick to collect sufficient amounts of melt to maintain the overall conductance. Further a fairly stable heat flow from shear heating or from the mantle is necessary to avoid rapid freezing of melt. Shear heating also causes frictional weakening, which supports the continuation of shearing.

The conductivity of a shear zone is not necessarily restricted to melt or aqueous fluid, but also could be based on graphite. The process of underthrusting the Brazilian shield under the Andes from the east could transport significant amounts of carbon into the middle and lower crust. Oil and gas fields for example in the western Santa Cruz-Tarija Basin and in the Oran-Olmedo Basin demonstrate that large amounts of carbon are available in the region east of the Andes and therefore could be swallowed by tectonic processes. Since graphite can be a very efficient lubricant, it even would support fault movement along such a plane.

However, a large-scale thrust system should provide a more homogeneous distribution of the conductivity anomalies throughout the Altiplano-Puna plateau than the MT measurements revealed. Also the information from seismic experiments did not support a conductive thrust system. The melt-containing structure found with seismics was of rather large thickness. A shear plane should be detected as a thinner structure. A shear plane with graphite-based conductivity would not have produced a seismic signal like the obtained one at all. So, even though a shear system is geologically not completely unlikely the anomaly beneath the Altiplano is probably not caused by shear planes.
Summary

Investigation of geologic, tectonic and other constraints restricted the number of possible interpretations of the high conductivity zone beneath the Altiplano. The analysis was focused on the presence of melt causing the conductive anomaly. The concept of dikes as conductive structures was dismissed from the list of interpretations. Even though diking is a very efficient mechanism of magma transport, it is tectonically not likely to find dikes in the middle and lower crust of the Altiplano. Furthermore the results of other geophysical methods (especially seismics) argue against dikes as well. The models of extended melt zones turned out to be the best interpretation of the conductive anomaly when the melt fraction of the structure was assumed not to be exceedingly high. The model of a migmatite, i.e. of a zone of moderate melt fraction, in which the melt forms a network with a high degree of interconnection, was favored. Any other type of melt chamber of the size indicated by the anomaly with a high melt fraction typical for plutons would not remain stable within the crust but rise due to buoyant forces. The interpretation that the high conductivity zone is caused by a voluminous migmatite was also supported by other geological and geophysical examinations. The concept of shear planes was not confirmed by other MT measurements and also other geological and geophysical analyses. So even though the inversion result of the MT models very closely reproduce the anomaly found in the ANCORP data, it did not appear to be a consistent interpretation of the high conductivity zone beneath the Altiplano.
6 Conclusions

Magnetotellurics is a method to investigate the conductivity features of the subsurface of the earth. Zones of unusual high or low conductivity are located and the conductivity values are obtained. 2D inversion of the measured data finally images the distribution of the conductivity values throughout the depth of the profile. The problem is then to interpret the anomaly in terms of geological and tectonic constraints.

In the study presented here the opposite way was employed. Models for different geologic settings that consisted of conductive structures were constructed. From those synthetic measuring data were produced with 3D forward calculation. These data were run with a 2D inversion routine and the resulting images of the conductivity anomalies were investigated and also compared with inversion data that had been obtained from measurement data (ANCORP profile; Schwalenberg, 2000). It turned out that it is possible to produce very similar inversion results with very different models. Besides this the inversion results of the model data mostly reproduced the anomaly obtained from “real” data very well. Just based on the MT data it was not possible to unequivocally determine, which of the “candidates” was the most likely interpretation of the conductivity anomaly in the inversion results. Therefore, interpretation of 2D inversion results can only lead to reasonable conclusion when information of other experiments and investigations is included.

Similar investigations were performed in other regions such as Iceland, and the Tibet Plateau. All the high conductivity systems have different maximum conductivities due to the varying conductive systems and probably due to the differences in conducting material. These high conductivity zones are controlled by very different tectonic settings: The Tibet Plateau developed in connection with the collision of the Asian plate with the Indian plate, i.e. in a compressive regime comparable to the Altiplano. A high conductive layer of large extent was obtained with MT measurements of the INDEPTH project (Wei et al., 2001). The conductivity is interpreted to be mainly based on the existence of melt. Li et al. (2003) suggest for Tibet a thin layer of water located above the partial melt produced by cooling and crystallization of the melt. In their model free water is supposed to separate from the saturated melt and to collect above it through buoyancy effects. This
layer is supposed to form an independent strong conductive system besides the melt and to increase the integrated total conduction of the high conductivity zone. However, water tends to leave the crust rapidly under these conditions or is bound in newly forming minerals. This model has the difficulty to explain where the large amount of water comes from, since this collisional system has no subducting oceanic plate anymore that could transport significant amounts of water into the mantle.

Iceland on the other hand is located right on top of the Mid Atlantic Ridge beneath which a mantle plume is located. Kreutzmann et al. (2004) performed modeling of electrical conductivity distribution of this system based on temperature and melt fraction values. They figured out that for models with very low melt fraction MT signals can detect a mantle plume only in large depth (50 to 120 km). Within a ridge setting the plume models only slightly decreased the mantle resistivity. In field data this effect may be too small to be detected. In models with significantly higher melt fraction the plume stronger affected the data and could be detected within the ridge signal. Obviously the amount of melt being stable in the plume and with it the degree of interconnection of the melt are very important quantities here as well.

Quite a few aspects that could not be followed here are left for further investigation. It may be informative to run the inversion results with a 3D inversion instead of a 2D inversion. This surely would reduce the amount of artifacts that are introduced into the model by the inversion. 2D inversion produces always extreme conductivity values for the model because it is applied to 3D data. Especially for model structures of pronounced three dimensionality absurd structures and unrealistic conductivity values often were obtained that were due to the impact of 3D effects. Experiments with synthetic data performed by Siripunvaraporn et al. (2005) demonstrated that the application of 3D inversion routines to 2D profile data is able to reduce the generation of artifacts in the inversion. Another aspect worth to be tried may be to run the data with block inversion instead of smooth inversion. This may better constrain the shape, size and position of the target.

Babeyko et al. (2002) developed a model for the Altiplano plateau, in which intra-crustal convection occurs based on enhanced mantle heat flow. Hot crustal material is supposed to rise through a few narrow channels to a depth of about 20 km where it
spreads out laterally. According to their models large parts of the lower and middle crust are partially molten to a degree of mainly less than 10 vol%. Only sporadically zones with a melt fraction of 15 to 20 vol% occur in this model. It might be interesting to test this setting in an MT model to examine the conductivity distribution for it.

Concluding it should be mentioned that this approach tried here for interpreting MT data was probably a bit time consuming and sometimes complicated. However it turned out to strongly sensitize the mind for possible interpretations that are not close at hand. It demonstrated the necessity to include a large amount of additional data and information besides the actual measurement into the analysis that on one hand open new opportunities for interpretation but on the other hand restrict the number of possibilities very efficiently by leading approaches ad absurdum. Only if the interpretation of geophysical data describes a geologically, tectonically and geochemically tenable model we might be close to the truth.
7 References


http://www.amaps.com/spsamer.htm (April/7/2005)

https://zulu.ssc.nasa.gov/mrsid (April/7/2005)
Appendix

App. A Parameter sets used for numerical modeling

1. Parameter set for 3D forward modeling (WinGLink)

- min Error: 0.00000100
- Relaxations: 150
- Air Layers: 10
- Convergence Factor: 8
- Longest period: 10000,000
- # Decades: 4
- # Periods per Decade: 4

2. Parameter set for 2D inversion (WinGLink)

Main Parameters
- Invert TM mode rho and phase data
- Invert TE mode rho and phase data

Data select
- Min Frequency
  - TE data: 0.00010
  - TM data: 0.00010
- TE # Decades: 4,0
- TM # Decades: 4,0
- use station data

Smooth inversion
- Solve for smoothest model
- Uniform grid Laplacian operator (D-models); standard grid Laplacian operator (M-, SP-models)
- minimize integral of |Laplacian (m)|²
- tau for smoothing operator 100 (50)
- no static shift
- no fixed parameters
**Data Errors**

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<th>TM</th>
<th>rho</th>
</tr>
</thead>
<tbody>
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</tbody>
</table>

use data errors if existing

**Error floor**

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<th>rho</th>
<th>TM</th>
<th>rho</th>
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</thead>
<tbody>
<tr>
<td>phase</td>
<td>5,00</td>
<td>phase</td>
<td>5,00</td>
</tr>
</tbody>
</table>
Fig. 8.1 Inversion result of model D4, run with the regularization “Standard grid Laplacian operator”. rms 2.696. The very strong 3D effects of this model cause a spurious heterogeneity in the inversion. The inversions of this model and the other dike models were therefore run with a higher degree of smoothing. See text.

Fig. 8.2 Model of the multi-dike system D2. The forward calculation did not converge, so that the model was dismissed from further analysis.
Fig. 8.3  Model D3 model and 2D inversion result. rms 3.811. The inversion was run with the regularization “Standard grid Laplacian operator”. The very strong 3D effects of this model cause considerable distortion of the inversion result compared with the original model.
Fig. 8.4 Model M2b, inversion result showing resistivity distribution along Profile C. The original model was derived from model M2a, but here without discretization, rms 1.455

Fig. 8.5 Model M3a, inversion result showing resistivity distribution along Profile C. The original model was similar to M1a, only the depth extent was increased to 63 km, rms 2.079
Fig. 8.6 Model M5c, inversion result showing resistivity distribution along Profile C. The model had a fairly low resistivity contrast between the conductive structure ($10 \, \Omega \text{m}$) and the host rock ($1000 \, \Omega \text{m}$). No gradient zone was implemented. rms 1.74
Fig. 8.7 Model SP1d model and 2D inversion result. The white line in the cross section on the right denotes the level of depth that is shown in the map view on the left.
App.C Estimations of the cooling time of a dike and a sphere according to Buntebarth (1980)

C1 Dike

Fig. 8.8 Cross section of a magma-filled dike, explanation of the abbreviations see below

Fig. 8.9 Relative temperature distribution in a dike and the surrounding host rock depending on the relation d/D. Modified from Buntebarth (1980)

Time for cooling of a dike: \[ t = \frac{a \cdot D^2}{\kappa} \]

\( \kappa = \) thermal diffusivity, \( 0.8 \times 10^{-6} \) (granite), \( 7.9 \times 10^{-7} \) (basalt, gabbro)

\( a = \kappa t \cdot D^2 \), curve parameter, obtained from diagram

\( D = \) Dike width (given)
d = position within the dike
$T_1$ = Magma temperature
$T_2$ = Host rock temperature at a certain depth
$T$ = temperature of the magma ($d/D<1$) or the host rock ($d/D>1$) reached due to heat transfer from the dike into the host rock

This model is also used to estimate the cooling time of an underplating sill. For details see Buntebarth (1980).
C2 Sphere

Fig. 8.10 Cross section of a magma-filled sphere, explanation of the abbreviations see below.

Fig. 8.11 Relative temperature distribution in a sphere and the surrounding host rock depending on the relation \( r/R \). Modified from Buntebarth (1980).

Time for cooling of a sphere: \[ t = \frac{a \cdot R^2}{\kappa} \]

\( \kappa \) = thermal diffusivity, \( 0.8 \times 10^{-6} \) (granite), \( 7.9 \times 10^{-7} \) (basalt, gabbro)
\( a = \kappa \cdot t/D^2 \), curve parameter, obtained from diagram
\( R \) = Radius of the sphere (given)
\( r \) = position within the sphere
\( T_1 \) = Magma temperature
\( T_2 \) = Host rock temperature at a certain depth
\( T \) = temperature of the magma (\( d/D < 1 \)) or the host rock (\( d/D > 1 \)) reached due to heat transfer from the dike into the host rock
For details see Buntebarth (1980).
**App. D**

**Table 8.1** Ascent velocities calculated according to Stokes, depending on the radius of the ascending melt spheres.

### Alkali –Olivine-Basalt (1250°C) in lower crust (Hall, 1996)

<table>
<thead>
<tr>
<th>r (m)</th>
<th>$\rho_m$ (kg/m³)</th>
<th>$\rho_b$ (kg/m³)</th>
<th>$v$ (m/s) $(10^{18}$ Pas)</th>
<th>$v$ (cm/yr) $(10^{18}$ Pas)</th>
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<tbody>
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### Tholeiite (1400°C) in lower crust (Murase & Mc Birney, 1973)

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<th>$\rho_b$ (kg/m³)</th>
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### Andesite (1250°C) in lower crust (Hall, 1996)

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Granite (1250°C) in lower crust  (Hall, 1996)

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<th>( v ) (m/s) ( (10^{18} \text{ Pas}) )</th>
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App. E: Estimate of the depth-dependent behavior of a melt sphere related to pressure and temperature

The melt is a granitic melt sphere with an uncompressed radius of 1m (4.19m³) and a density of 2170 kg/m³; α=2.4*10⁻⁵ K⁻¹ and β=10⁻¹⁰ Pa⁻¹.

The geothermal gradient is calculated using an updated version of equation 4-31 in Turcotte and Schubert (2002) based on the estimate of underplating the crust by mantle magma:

\[
T = T_0 + \frac{q_m \cdot z}{k} + \frac{(q_0 - q_m) \cdot h_r}{k} \cdot (1 - e^{-z/h_r})
\]

\(T_0\) = surface temperature (10°C)
\(z\) = depth (m)
\(q_m\) = heat flux from mantle to base of the lithosphere (assuming magmatic underplating) (30 mW/m²)
\(q_0\) = surface heat flow (100 mW/m²)
\(h_r\) = length scale for the decrease of the heat relating the radiogenic heat production rate per unit mass to depth (10000m)
\(k\) = thermal conductivity (2.5W/mK)

The pressure estimation for the depth levels is based on a simple model of the crust: first layer 2600 kg/m³, 0-30 km of depth, second layer 3000 kg/m³, 30-55 km, third layer 3300 kg/m³, 55-70 km.

\[
P = \sum_{i=1}^{n} g\rho_i z_i
\]

<table>
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<th>depth (km)</th>
<th>Temperature (°C)</th>
<th>Pressure (MPa)</th>
<th>depth-dependent volume (m³)</th>
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Table 8.2 Dependency of volume and density of a melt-filled sphere on depth (temperature and pressure)