Cambridge University Press
0521817277 - Practical Magnetotellurics
Fiona Simpson and Karsten Bahr
Excerpt
More information

Chapter 1
Introduction

Magnetotellurics (MT) is a passive exploration technique that utilises a broad spectrum of naturally occurring geomagnetic variations as a power source for electromagnetic induction in the Earth. As such, MT is distinct from active geoelectric techniques, in which a current source is injected into the ground as a power source for conduction. In fact, MT and geoelectrics have little in common other than the physical parameter (electrical conductivity) imaged. MT is more closely related to geomagnetic depth sounding (GDS), which was developed in the late nineteenth century after the existence of magnetovariational fields arising from induction was demonstrated by Schuster (1889) and Lamb (see Schuster, 1889, pp. 513–518). They applied a mathematical technique, invented by Gauss (1839) for separating magnetovariational fields originating internal to the Earth from those of external origin, to geomagnetic observatory data and detected a significant internal component. In the 1950s, Tikhonov (1950, reprinted 1986) and Cagniard (1953) realised that if electric and magnetic field variations are measured simultaneously then complex ratios (impedances) can be derived that describe the penetration of electromagnetic fields into the Earth. The penetration depths of electromagnetic fields within the Earth depend on the electromagnetic sounding period, and on the Earth's conductivity structure. This is the basis of the MT technique.

1.1 Magnetotellurics as a passive electromagnetic exploration method and its relation to active electromagnetic and geoelectric methods

The magnetotelluric (MT) technique is a passive electromagnetic (EM) technique that involves measuring fluctuations in the natural electric, **E**, and magnetic, **B**, fields in orthogonal directions at the surface of the Earth as a means of determining the conductivity structure of the Earth at depths ranging from a few tens of metres to several hundreds of kilometres. The fundamental theory of exploration MT was first propounded by Tikhonov (1950, reprinted 1986) and, in more detail, by Cagniard (1953). Central to the theses of both authors was the realisation that electromagnetic responses from any depth could be obtained simply by extending the magnetotelluric sounding period. This principle is embodied in the *electromagnetic skin depth* relation, which describes the exponential decay of electromagnetic fields as they diffuse into a medium:

$$p(T) = (T/\pi\mu\overline{\sigma})^{1/2}, \tag{1.1}$$

where $p(T)$ is the electromagnetic skin depth in metres at a given period, T, $\overline{\sigma}$ is the average conductivity of the medium penetrated, and μ is magnetic permeability. At a depth, $p(T)$, electromagnetic fields are attenuated to e^{-1} of their amplitudes at the surface of the Earth. This exponential decay of electromagnetic fields with increasing depth renders them insensitive to conductivity structures lying deeper than $p(T)$. Hence, in MT studies, one electromagnetic skin depth is generally equated with the *penetration depth* of electromagnetic fields into the Earth. In studies of the Earth, μ is usually assigned the free-space value ($\mu_0 = 4\pi \times 10^{-7}\,\mathrm{H\,m^{-1}}$), and Equation (1.1) can be approximated as

$$p(T) \approx 500\sqrt{T\rho_a}, \tag{1.2}$$

where ρ_a is *apparent resistivity*, or the average resistivity of an equivalent uniform *half-space*.

From Equations (1.1) and (1.2), we can deduce that for a given sounding period, the depth achieved by passive EM sounding will be dictated by the average conductivity of the overlying sample of earth that is penetrated. Electromagnetic fields that are naturally induced in the Earth and are exploitable for MT studies have wave periods ranging from $\sim 10^{-3}$ to $\sim 10^{5}$ s. Therefore, if we assume an average resistivity of the Earth's crust and upper mantle of $100\,\Omega\,\mathrm{m}$ (we hasten to add that the conductivity structure of the Earth is much more interesting!), we can see how penetration depths in the range of ~ 160 m to >500 km might be possible. The broad span of depths that can be imaged using the MT technique is one advantage of the method compared with active EM methods for which the maximum depth that can be probed is always limited by the size of the available source, and realisable source–receiver configurations.

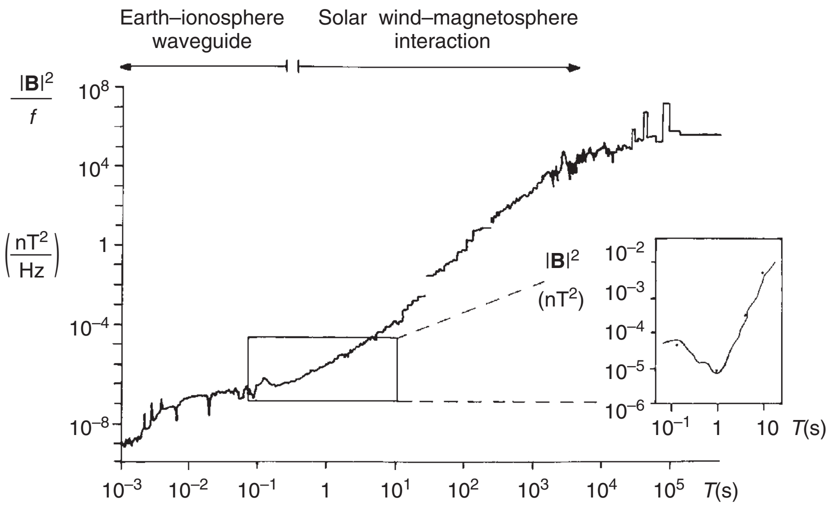

Figure 1.1 Power spectrum illustrating '1/*f* characteristics' of natural magnetic variations (modified from Junge, 1994). Short-period signals are generated by interactions in the Earth–ionosphere waveguide, whereas long-period signals are generated by solar wind–magnetosphere interactions. The spectral lines at periods of the order 10^5 s are harmonics of the solar quiet (Sq) daily variation. Inset illustrates the reduced signal power ($|\mathbf{B}|^2$) in the dead-band.

Whilst magnetohydrodynamic processes within the Earth's outer core generate the greater part of the Earth's magnetic field, it is the superimposed, more transient, lower-amplitude fluctuations of external origin, that MT sounding seeks to exploit. The power spectrum (Figure 1.1) of these fluctuations plummets in the 0.5–5 Hz frequency range, minimising at a frequency of ~1 Hz. This so-called *dead-band* of low-amplitude signals is attributable to the inductive source mechanisms, one effective above ~1 Hz, the other below ~1 Hz, and is frequently manifest in MT sounding curves by a reduction in data quality.

Electromagnetic fields with frequencies higher than 1 Hz (i.e., periods shorter than 1 s) have their origins in meteorological activity such as lightning discharges. The signals discharged by lightning are known as 'sferics' and encompass a broad range of electromagnetic frequencies. Local lightning discharges may saturate amplifiers, and it is not these, but rather the sferics from the highly disturbed equatorial regions, which propagate around the world within the waveguide bounded by the ionosphere and Earth's surface that are of most significance. Sferics propagate within the waveguide as transverse electric (TE), transverse magnetic (TM) or transverse electric and magnetic (TEM) waves, and are enhanced or attenuated depending on frequency. A description of these waveguide modes and discussion of the constructive and destructive interference that leads to certain frequencies being enhanced, whilst others are attenuated can be found in Dobbs (1985, Chapter 8). During the day, the waveguide is ~60 km wide, increasing to ~90 km at night-time. Sferics peak in the early afternoon. However, statistically a part of the

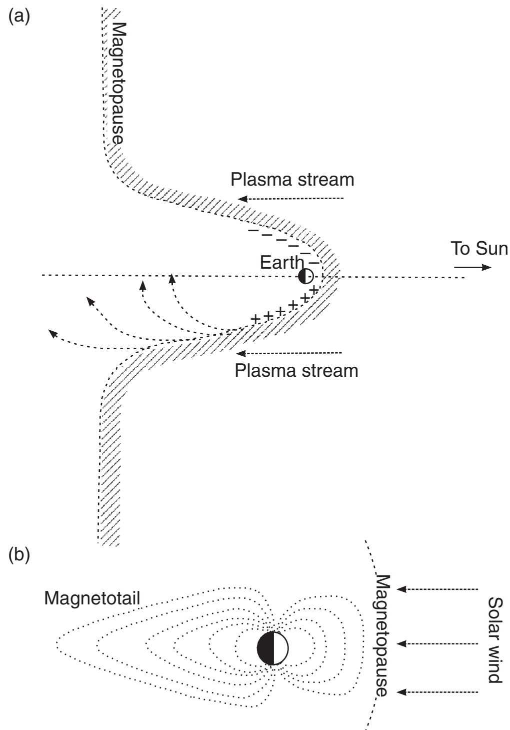

Figure 1.2 (a) Distortion of advancing front of solar plasma by Earth's magnetic field (after Chapman and Ferraro, 1931). The solar wind blows continually, but with varying intensity. Sudden increases in solar-wind pressure push the *magnetopause* closer to the Earth giving rise to magnetic storms. (b) Magnetic field lines showing the form of the Earth's magnetosphere, which extends further from the Earth than originally envisaged by Chapman and Ferraro (1931). The magnetosphere typically extends to approximately 10 Earth radii (i.e., ~64 000 km) on the dayward side of the Earth, whilst a long magnetic tail (the magnetotail) extends more than 300 000 km away from the nightward side of the Earth.

world may witness thunderstorm activity at any given universal time (UT).

Interactions between the solar wind (Parker, 1958) and the Earth's *magnetosphere* and ionosphere generate electromagnetic fluctuations with frequencies lower than 1 Hz (i.e., periods longer than 1 s). Briefly, the solar wind is a continual stream of plasma, radiating mainly protons and electrons from the Sun. On encountering the terrestrial magnetic field (at the *magnetopause*), these protons and electrons are deflected in opposite directions, thereby establishing an electric field (Figure 1.2). Variations in density, velocity and magnetic field intensity of the solar wind produce rapidly varying distortions of the Earth's *magnetosphere*. For example, increases in solar-wind pressure cause rapid compression of the magnetosphere, and therefore compaction of magnetic field lines, affecting an increase in the horizontal geomagnetic field. Oscillations of the magnetosphere generate small, almost sinusoidal variations of the geomagnetic field, called geomagnetic pulsations. Inductive and magnetohydrodynamic interactions between the magnetosphere and ionosphere complexly modify these fluctuating fields before they reach the Earth's surface. A more detailed discussion of these processes can be found in Campbell (1997).

The largest geomagnetic field variations (up to the order of a few hundred nT) occur during *magnetic storms*, which occur owing to sporadic increases in the rate at which plasma is ejected from the Sun. Magnetic storms last for several days, and in polar regions can lead to magnificent displays of light known as *aurora borealis* and *aurora australis*, or *northern* and *southern lights*, respectively.

Active EM induction techniques, in which the EM source field is man-made (e.g., the output from a generator), probe a more limited range of penetration depths than passive EM induction techniques: although, in theory, an artificial EM field with a period of several days can be generated, such a field will not be as powerful as a magnetic storm. Hence, MT sounding is better suited to probing depths of several hundred kilometres than are active induction techniques. However, active EM induction techniques can be used to improve the data quality in the period range of the dead-band.

An active exploration technique that is commonly used for probing the Earth's electrical conductivity structure is geoelectrics. Here, the physical process involved is conduction – which is not time-dependent – rather than induction. The spatial distribution of the geoelectrical voltage is described by the Laplace equation (rather than by the time-dependent *diffusion equation* that governs induction, which we explore in detail in Chapter 2): i.e., geoelectrics is a potential method. Being a potential method, geoelectrics suffers from the limitations which all potential methods have: in particular, a very limited depth resolution. In contrast, MT is not a potential method, and if EM fields spanning a wide period range are evaluated, the depth resolution of the MT method can be very high.

1.2 Problems for which EM studies are useful: a first overview of conduction mechanisms

Ascertaining subterranean electrical resistivity structure is rendered trivial unless electrical resistivity can be linked to other physical parameters or processes. Temperature, pressure, physical and chemical state, *porosity* and *permeability*, as well as the frequency at which measurements are made can all play a crucial role in determining the electrical resistivity exhibited by rocks and minerals.

The transmission of electrical currents by free charge carriers is referred to as *conduction*. Conduction occurs in rocks and minerals via three principal electrical charge propagation mechanisms: electronic, semi-conduction and electrolytic. Electronic conduction occurs in metallic ore minerals (e.g., magnetite, haematite), which contain free electrons that can transport charge. Electronic conduction is most significant in the Earth's core. Conduction in graphite is of particular relevance in EM studies. A single-crystal of graphite is an electronic conductor, due to the availability of free electrons. The conductivity of the amorphous graphite that occurs in the Earth's crust is lower than the conductivity of metallic ores, but higher than the conductivity of natural semi-conductors or electrolytes. Semi-conduction occurs in poor conductors containing few free charge carriers. Only a small proportion of the electrons present in a semi-conductor contribute to conduction, and a marked increase in conduction can be affected by the inclusion of minor amounts of weakly-bonded, electron-donating impurities. Semi-conduction is expected to dominate in mantle minerals such as olivine. Electrolytic conduction occurs in a solution containing free ions. Saline water is an electrolyte of particular relevance to crustal EM studies. Its free ions are derived from the dissolution of the constituent ions (e.g., Na^+, Cl^-) of the solid salt on entering solution. Only 'free' (i.e., not chemically bound) water has a significant effect on observed conductivities. In active tectonic regions, any partial melt generated by enhanced temperatures, adiabatic decompression or asthenospheric upwelling will also act as an electrolyte (e.g., Shankland and Waff, 1977).

EM sounding methods are highly sensitive to variations in abundance and distribution of minor mineral constituents and fluids within the rock matrix. A small fraction of conductive mineral (or fluid) will increase the *bulk conductivity* of a rock formation considerably if distributed so that it forms an interconnected network, whereas a larger fraction of the same conductive phase could have negligible effect if present as isolated crystals (or in pores). The interdependence of quantity and distribution of a conductive phase in determining the conductivity of a multi-phase medium is described by mixing laws (see Section 8.3). Conductive minerals that occur within the Earth in the required abundance and geometries to significantly affect the electrical resistivities include graphite, sulphides, magnetite, haematite, pyrite and pyrrhotite.

As we shall see later (Chapter 8), bulk electrical conductivity is relatively insensitive to the electrical conductivity of the host rock. Therefore, a direct link between electrical conductivity and lithology is generally not possible. However, porosity and permeability vary significantly according to rock type and formation, and variations in porosity and permeability influence electrical conductivity substantially. Sedimentary rocks typically contain significant interstitial fluid content, and are generally more porous than igneous or metamorphic rocks, unless the latter are extensively fractured, fissured or jointed. Unconsolidated sedimentary rocks can support porosities of up to 40%, but older, more deeply buried (and therefore more densely compacted) sediments might have typical porosities of less than 5%. In limestones and volcanic rocks, spheroidal pores (known as vugs) are common, but fluids contained in vugs will have less effect on bulk rock conductivity than fluids distributed in thin, interconnected microcracks or as grain-boundary films. Whereas the conductivity of a saturated rock decreases significantly with pressure owing to closure of fluid-bearing cracks, that of a partially saturated rock may be enhanced for small pressure increments, as a result of new fluid pathways being generated by microcracking. In addition to conduction via pores and cracks, adsorption of electrically attracted ions at grain interfaces results in surface conduction. Surface conduction is of particular significance in clays, which ionise readily.

The resistivity of a two-phase medium is governed primarily by the connectivity and resistivity of the conductive phase. Fluid resistivities are sensitive to salinity, pressure and temperature. With increasing temperature, the viscosity of a solution is decreased, promoting heightened ionic mobility and therefore tending to enhance conductivity. However, there are also conflicting tendencies owing to decreasing density and decreasing dielectric constant with increasing temperature

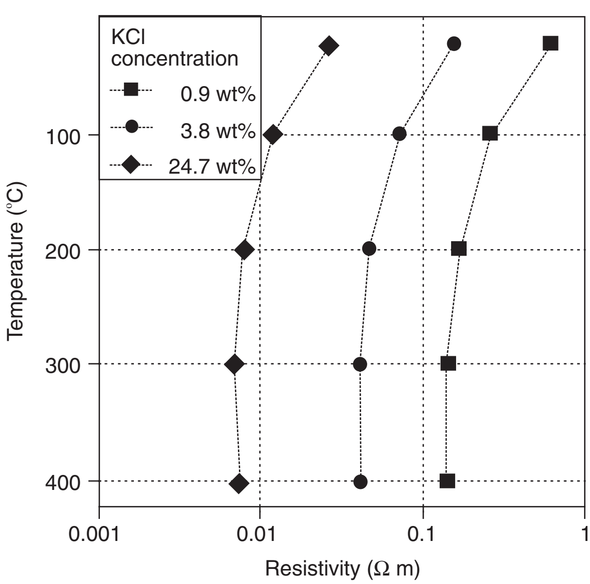

Figure 1.3 Electrical resistivities of varying concentrations of aqueous KCl solution as a function of temperature. (Redrawn from Nesbitt, 1993.)

that act to reduce conductivity. Figure 1.3 shows resistivity as a function of temperature for high-salinity KCl brines. It is known from boreholes that highly saline fluids exist to at least mid-crustal depths. For example, fluids pumped from depths down to 8.9 km from a deep borehole (KTB) in Bavaria, S. Germany have NaCl salinities of 1.7 M (Huenges *et al*., 1997), which translates to electrical resistivities of $\sim$0.1 Ω m at 20 °C, or $\sim$0.025 Ω m at 260 °C – the temperature at the 8.9 km depth from which they were pumped. Seawater has electrical resistivities of the order 0.2–0.25 Ω m.

A major contender to fluids as the cause of enhanced crustal conductivities is carbon – either mantle-derived as a product of CO_2 degassing (Newton *et al*., 1980) or biogenic (e.g., Rumble and Hoering, 1986; Jödicke, 1992) – in the form of graphite and/or black shales. It has been suggested that, under strongly reducing conditions, graphite might form as a grain-boundary precipitate from CO_2-rich or hydrocarbon-bearing metamorphic fluids (Frost, 1979; Glassley, 1982; Rumble and Hoering, 1986 and references therein). CO_2 inclusions in metasedimentary granulites (which may be representative of lower-crustal rock type) from Scandanavia, Tanzania and India are consistent with mantle-derived, CO_2-rich fluids and carbonate-rich melts emplaced by deep-seated intrusives (Touret, 1986). However, C_{13} isotope studies support a biogenic source of graphite in outcrops that have been sampled (e.g., Rumble and Hoering, 1986, Large *et al*., 1994). In contrast to artificial manufacture of graphite via carbon vapour diffusion for which formation temperatures of between 600 °C (with nickel as a catalyst) and greater than 2000 °C are required, graphite has been identified in low-grade metamorphic rocks with inferred formation temperatures of less than 450 °C (e.g., Léger *et al*., 1996). The low formation temperatures may be explained by the action of shear stresses, which promote graphitisation in the 400–500 °C temperature range (Ross and Bustin, 1990).

Graphitic horizons have been interpreted as markers of past tectonic events. For example, Stanley (1989) documented extensive outcrops of graphitic shales apparently associated with active and fossilised subduction zones, and exposures of black shales have also been mapped along the Grenville Front (Mareschal *et al*., 1991). Graphitic films have been observed in anorthosite rocks derived from the mid crust in Wyoming (Frost *et al*., 1989), whilst core samples from the KTB borehole and proximate surface exposures contain graphite both as disseminated flakes and concentrated along shear zones (Haak *et al*., 1991; ELEKTB, 1997). Graphite has also been detected in mafic xenoliths of recent crustal origin

(Padovani and Carter, 1977). During uplift, loss of connectivity of the graphite may occur (Katsube and Mareschal, 1993; Mathez *et al*., 1995), but Duba *et al.* (1994) and Shankland *et al*. (1997) have demonstrated that reconnection of graphite occurs when rock samples are pressurised. This pressure-induced reconnection of graphite suggests a rock mechanical model by which rocks, which are resistive at the Earth's surface may be conductive at depth. An alternative explanation involves loss of fluids (Shankland and Ander, 1983). Arguments for and against saline fluids and/or graphite as prime candidates for generating enhanced conductivities in the deep continental crust have been reviewed by Yardley and Valley (1997, 2000), Simpson (1999) and Wannamaker (2000).

Long-period MT studies (e.g., Lizzaralde *et al*., 1995, Heinson, 1999, Simpson, 2001b) indicate that average mantle conductivities are higher than the conductivity of dry olivine (Constable *et al*., 1992), which constitutes 60–70% of the upper mantle. Enhanced mantle conductivities may be generated by graphite, partial melts or diffusion of free ions (e.g., H^+, OH^-).

Partial melting may account for deep-crustal and upper-mantle conductivity anomalies that are imaged coincident with enhanced heat flow and seismic low-velocity zones (e.g., Wei *et al*., 2001). The conductivity of a partial melt is governed by its composition and its temperature (Figure 1.4), as well as by melt fraction and distribution (see Chapter 8). Sometimes, however, modelled electrical conductances imply implausibly high mantle temperatures or melt fractions for them to be generated solely by partial melting.

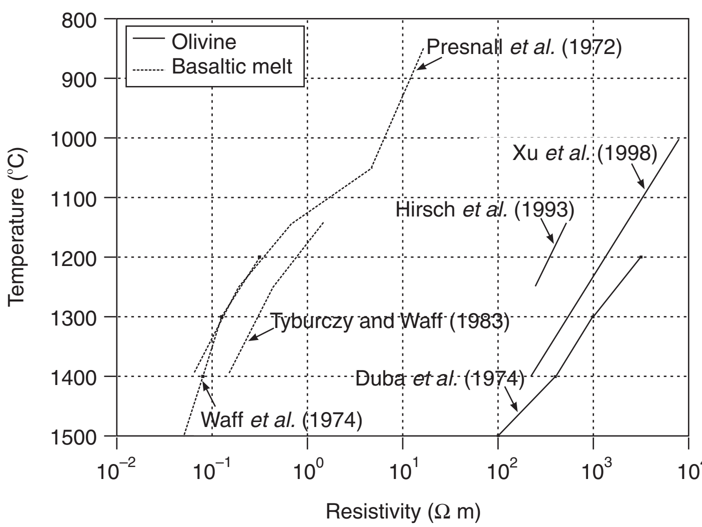

Figure 1.4 Electrical resistivities of dry olivine and basaltic melt as a function of temperature. Basaltic melt is approximately three orders of magnitude less resistive than dry olivine.

Trace amounts (p.p.m.) of water can significantly enhance the electrical conductivity of olivine at a given temperature by supplying mobile charge carriers in the form of hydrogen (H^+) ions (Karato, 1990). Hydrogen diffusivities are anisotropic in olivine (Mackwell and Kohlstedt, 1990), and the implications of this in terms of the potential contributions that MT studies can make to geodynamic models of mantle flow will be discussed in Chapter 9 (Section 9.3).

The maximum penetration depths that can be routinely achieved using MT sounding are of the order 400–600 km. At these depths, phase transitions from olivine to wadsleyite (~410 km) to ringwoodite (~520 km) to perovskite + magnesiowüstite (~660 km) are believed to occur. These phase transitions are expected to contribute to significant conductivity increases with increasing penetration into the *transition zones* (Xu *et al.*, 1998). Conductivity–depth models for the mantle derived from long-period MT data and from laboratory conductivity measurements on representative mineral assemblages are compared in more detail in Chapter 8 (Sections 8.1 and 8.2). Seismological studies also image transition zones at mid-mantle depths of ~410 km, ~520 km and ~660 km.

Overall, the electrical resistivities of rocks and other common Earth materials span 14 orders of magnitude (Figure 1.5). Dry crystalline rocks can have resistivities exceeding $10^6\ \Omega\,\mathrm{m}$, whilst graphite-bearing shear zones might have resistivities of less than $0.01\ \Omega\,\mathrm{m}$. Such a wide variance provides a potential for producing well-constrained models of the Earth's electrical conductivity structure.

1.3 An historical perspective

In 1889, Schuster (1889) applied the *Gauss separation* technique to the *magnetic daily (diurnal) variation*, and the subject of electromagnetic induction in the Earth was born. The Gauss separation is a mathematical formulation used by Gauss (1839) to deduce that the origin of the geomagnetic main field is internal. Suppose we have competition between two magnetic field models – a dipole within the Earth and a large ring current outside the Earth (Figure 1.6). In both cases, the magnetic field vector is the gradient of a potential:

$$\mathbf{B} = -\nabla U \tag{1.3}$$

and in both cases the components B_r, B_ϑ (in spherical co-ordinates, at the Earth's surface) are the derivatives of the potential U with respect to the co-ordinates r,ϑ, at the location $r = r_E$, where r_E is the radius of the Earth. For the dipole model