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Introduction to the magnetotelluric method

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1.1 Introduction

Over the past three decades, and particularly over the intervals 1985–1995 and 2005–2010, magnetotellurics has undergone a revolution driven by four main factors: (1) the emergence of low-power, low-cost, 24-bit digital electromagnetic sensing and recording technologies, (2) dramatic improvements in the understanding of noise in electromagnetic measurements, with the concomitant evolution of data processing algorithms, (3) substantial advances in the ability to recognize and remove distortion by near-surface structure local to the measurement point that is the bane of practical magnetotellurics, and (4) the development of fast two- and three-dimensional (2D and 3D) modeling and inversion capabilities concurrent with the constantly increasing power of computers. In the 1970s, a typical magnetotelluric survey consisted of a handful of sites whose data were analyzed using ordinary least-squares methods, smoothed in the frequency domain to reduce data scatter, and interpreted using one-dimensional (1D) models "stitched" together into a 2D pseudo-section that may, or may not, be tested through 2D forward modeling. By the 1990s, surveys comprising several tens of sites along a single line were common, data were processed using robust methods, which produced substantially more reliable response estimates that were subsequently analyzed for galvanic distortion, and rapid 2D modeling and inversion were standard. By the 2010s, magnetotelluric surveys consisting of many hundreds of sites, with areal rather than linear coverage, are being carried out, data processing is semi-automatic, usually using bounded influence or multivariate approaches, multi-site distortion removal is being applied routinely, 2D interpretation, often including anisotropy, is routine, while 3D interpretation based on 3D inversion is becoming commonplace. The key purposes of this book are documenting this magnetotelluric revolution and providing an up-to-date, rigorous reference on the field that is much more than a practical guide, and that is useful to the novice, expert practitioner and interested non-magnetotelluric geoscientist alike.

As an illustration of the technological and scientific distance that magnetotellurics has come over the past three decades, two representative studies from the mid-1970s and mid-2000s will be compared and contrasted. The first is the thorough analysis and

The Magnetotelluric Method: Theory and Practice, ed. Alan D. Chave and Alan G. Jones. Published by Cambridge University Press © Cambridge University Press 2012.

geological interpretation of 16 magnetotelluric sites from the eastern Snake River Plain–Yellowstone region of the western USA presented by Stanley *et al*. (1977). The second is a 3D magnetotelluric study utilizing 100 areally distributed sites from the Betic Cordillera on the Iberian peninsula published by Rosell *et al*. (2011).

Stanley *et al*. (1977) is a (for the era) cutting-edge magnetotelluric data collection, analysis and interpretation effort that is also notable for its rigorous geological interpretation, a practice that was not widespread in the 1970s. The strong guidance of Francis Bostick, who published infrequently but was very influential in the 1970s and 1980s, is evident in the paper. Data were collected using the magnetotelluric–audiomagnetotelluric system of the University of Texas Geomagnetics Laboratory based on coil magnetometers and electric field lines recorded in analog form on frequency-modulated (FM) tape after band-pass filtering. These were supplemented by a digital cryogenic system developed by the US Geological Survey. Little is stated about processing the data into magnetotelluric response functions, but it can be assumed that a least-squares approach without remote referencing was applied. The resulting responses were represented in a 2D sense following Sims & Bostick (1969) and Word *et al*. (1970), with an adopted geoelectric strike angle consistent with surface geology. However, the magnetotelluric responses were not smooth functions of frequency, especially at low frequencies, probably due to a lack of robustness in response function estimation, and were often not interpretable due to the presence of galvanic distortion of the electric fields. As was sometimes the practice at the time, the response functions were smoothed after the fact, in this case using the novel approach of enforcing a Hilbert transform relationship between the apparent resistivities and phases (Figure 1.1). Each site was then inverted for a 1D model, and the results at adjacent sites were "stitched" together to produce the electrical cross-section shown in Figure 1.2. The key notable feature in the cross-section is the conductive zone at a depth ranging from 7 km beneath the Raft River thermal area to 18 km beneath the central part of the Snake River Plain to 5 km beneath Yellowstone. The resistivity at these depths ranges from 1 to 10 Ω m, and was interpreted to have a primarily thermal origin.

Rosell *et al*. (2011) utilized 100 broadband magnetotelluric sites (41 of which also included long-period data) distributed across a 400 km × 200 km area of southern Spain. The data were processed into response tensors using robust methods, and covered a period range of 0.001 to 20 000 s. The dimensionality of the dataset was assessed using the method of Martí *et al*. (2009) based on rotational invariants (see Chapter 6). The predominance of 3D behavior led to 3D inversion of the off-diagonal tensor elements using the WSINV3DMT code of Siripunvaraporn *et al*. (2005). Figure 1.3 depicts three vertical slices through the model, along with earthquake hypocenters, together with a geological map showing the locations of the magnetotelluric sites. The CB2 conductive body is the upward-oriented tongue near the center of each slice that divides the structure into three domains as follows: (1) a southwest domain D1 characterized by a resistive lithosphere and hypocenters with increasing depth toward the east and south, (2) a less resistive southeast domain D2 with hypocenter density concentrated at upper crustal levels, and (3) a northern domain D3

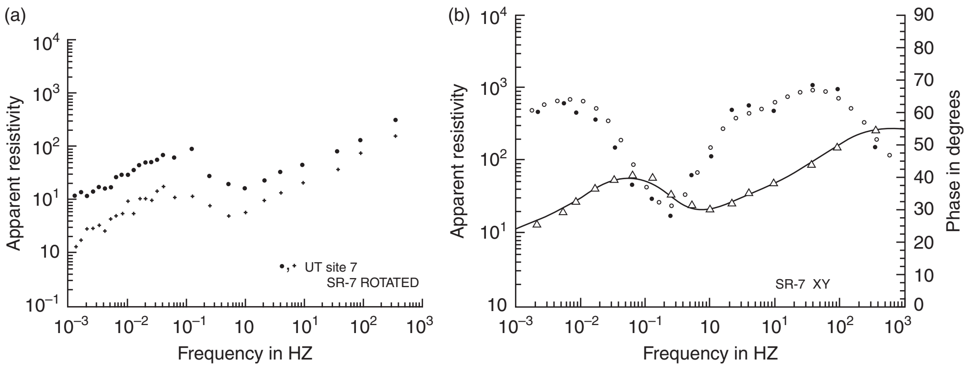

Figure 1.1. (a) The principal apparent resistivities for Site 7 from Stanley *et al.* (1977) after rotation to maximize the sum of the absolute values of the off-diagonal response tensor elements. (b) The transverse electric (TE) phase (solid circles) and the apparent resistivity computed from the phase through a Hilbert transform relationship (triangles). The open circles and solid line are the phase and apparent resistivity for a 1D model fitted to the transformed data. Taken from Stanley *et al.* (1977).

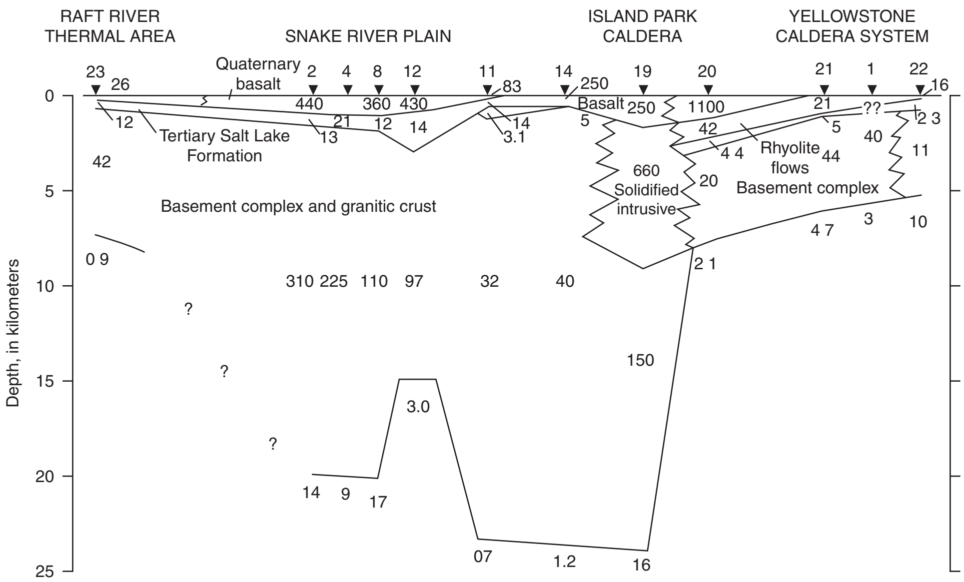

Figure 1.2. Southwest to northeast electrical cross-section extending from the Raft River thermal area to Yellowstone obtained by stitching together 1D inversions at each of the sites shown as triangles at the top. The numbers in the cross-section are the electrical resistivities obtained at that depth. The vertical exaggeration is 10:1. Taken from Stanley *et al.* (1977).

that is resistive to greater depth than in D1 and D2 and displays a paucity of hypocenters. Rosell *et al.* (2011) interpreted CB2 as the intrusion of asthenospheric material into the lithosphere caused by lateral lithospheric tearing and fracture of the eastward-directed subducting Ligurian slab beneath the Alboran region.

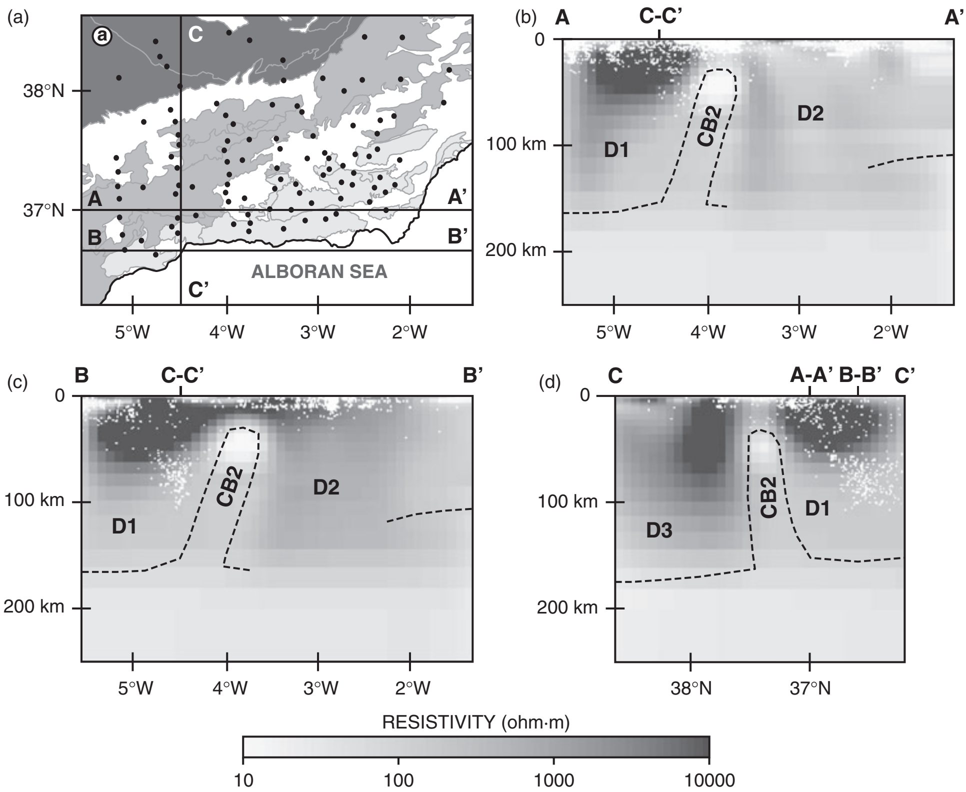

Figure 1.3. (a) Geological map of the Betic Cordillera showing the locations of the magnetotelluric sites (black dots) and (b)–(d) the slices A–A′, B–B′ and C–C′, respectively, that are vertical slices of the 3D resistivity model crossing the CB2 body shown as the vertical tongue near the center of each slice. Earthquake hypocenter locations within 8 km of each profile recorded since 1900 are shown by white dots. The dashed black line shows the lithosphere–asthenosphere boundary inferred from the resistivity distribution. D1, D2 and D3 are the main tectonic domains described in the text. Grayscale adaption of color figure from Rosell *et al*. (2011).

1.2 A quick tour of magnetotellurics

The propagation of electromagnetic waves through a medium with homogeneous physical properties is proportional to $e^{-\gamma x}$, where the complex propagation constant γ is

$$\gamma = \sqrt{\omega\mu(i\sigma - \omega\varepsilon)} \tag{1.1}$$

and ω is the angular frequency under the assumption of $e^{i\omega t}$ time dependence, μ is the magnetic permeability, σ is the electrical conductivity and ε is the electric permittivity. The constitutive relations, μ, σ and ε in (1.1) that also link the electric and magnetic fields in the Maxwell equations parameterize the physical properties of Earth that may be sensed in an electromagnetic experiment of any design (i.e. from megahertz ground-penetrating radar to sub-millihertz deep-mantle studies using the 11 yr solar cycle).

As shown in Chapter 2, at the frequencies used and for the targets of interest in magnetotellurics, the magnetic permeability may be taken as the free-space value μ_0 in nearly all Earth materials. In addition, $\omega\varepsilon \ll \sigma$, with the difference amounting to many orders of magnitude. The second term under the square root in (1.1) is due to displacement current in the Maxwell equations (the so-called Maxwellian term), and, once its effect is neglected, the electromagnetic fields are not governed by a wave equation. Instead, they are governed by a diffusion equation that also obtains in other fields of geophysics such as heat flow, with the important distinction that in electromagnetism they are vector fields rather than scalar in form. The diffusion equation limit of the Maxwell equations yields electromagnetic induction, and the magnetotelluric method is its primary geophysical application.

In a uniformly conducting medium, the length scale for electromagnetic induction is the familiar skin depth, which is the distance over which the electromagnetic field decays by $1/e \approx 0.37$, and is given by

$$\delta(\omega) = \sqrt{\frac{2}{\omega\mu\sigma}} \tag{1.2}$$

Using SI units, (1.2) may be simplified to

$$\delta = 503\sqrt{\rho T} \tag{1.3}$$

where δ has units of meters, T is the period in seconds and ρ is the inverse of the electrical conductivity, or the electrical resistivity in Ω m. At a zeroth-order conceptual level, the magnetotelluric method comprises measurement of the skin depth as a function of period to infer resistivity as a function of position in Earth. As shown in Chapter 3, the period range over which magnetotellurics can operate is about 10^{-4}–10^{5} s, and most bulk Earth materials have a resistivity range of 10^{-1}–10^{5} Ω m. More typical values in Earth's crust and uppermost mantle are 10–10 000 Ω m, usually decreasing with increasing depth (due to rising temperature), yielding a skin depth δ ranging over 50 m to 500 km and beyond, covering everything from the surface into the deep upper mantle. This broad range of penetration depths is one of the appeals of the magnetotelluric method, as it can be used in studies whose focus is near-surface to upper mantle without altering the underlying principles. Only the instrumentation needs to be changed.

The magnetotelluric method uses time variations of Earth's magnetic field caused at low frequencies (<10 Hz) by the interaction of the solar plasma with the ionosphere and magnetosphere as a source, and at high frequencies (>10 Hz) by global lightning activity (Chapter 3). The fluctuating magnetic field induces an electric current within Earth whose magnitude depends on electrical conductivity, and from Ampere's law, measurements of magnetic field fluctuations at Earth's surface determine the total electric current in the subsurface. The addition of an electric field measurement at Earth's surface yields the electrical conductivity at that point, and transformation of electric and magnetic field data into the frequency domain allows the geophysicist to map the electrical conductivity as a function of depth and position (and occasionally, with time, in continuous or repeat four-dimensional experiments). Since the electromagnetic fields are vector entities, it is possible to measure three components of the magnetic field and the two horizontal components of the electric field at Earth's surface, where the vertical electric field vanishes due to the presence of the insulating atmosphere. This limitation does not exist at the seafloor, but other, non-magnetospheric source types typically dominate the vertical electric field in the ocean, as shown in Chapter 2.

Owing to the vector form of electromagnetic fields, the fundamental datum in the magnetotelluric method is a tensor relationship $\overleftrightarrow{\mathbf{Z}}$ (called the magnetotelluric response function or tensor) between the surface vector horizontal electric and magnetic fields as a function of period, so that

$$\mathbf{E} = \overleftrightarrow{\mathbf{Z}} \bullet \mathbf{B} \tag{1.4}$$

where • denotes the inner product, along with an auxiliary equation relating the vertical and horizontal magnetic fields

$$\mathbf{B}_z = \mathbf{T} \bullet \mathbf{B}_h \tag{1.5}$$

where $\mathbf{T}$ is a vector called the Z/B transfer function, geomagnetic depth sounding transfer function or tipper function. Owing to the tensor form of the relation between the electric and magnetic fields, $\overleftrightarrow{\mathbf{Z}}$ and $\mathbf{T}$ from a single site contain information about the electrical conductivity as a function of coordinate direction and depth. Combining $\overleftrightarrow{\mathbf{Z}}$ and $\mathbf{T}$ from sites along a line or over an area can yield a 2D or 3D understanding of Earth structure. A formal physical and mathematical treatment of the properties of the magnetotelluric response $\overleftrightarrow{\mathbf{Z}}$ for 1D, 2D and 3D media is contained in Chapters 2 and 4. Its estimation from measurements is a difficult statistical problem whose solution is described in Chapter 5.

Once reliable estimates of the magnetotelluric response $\overleftrightarrow{\mathbf{Z}}$ and Z/B transfer function $\mathbf{T}$ are obtained from a set of data, the focus of the magnetotelluric practitioner shifts to their analysis, modeling and ultimately a geological interpretation. Distortion of regional electric fields by local structures is arguably the greatest bane of the magnetotelluric method, and one that has caused poor interpretations directly leading to low acceptance of magnetotellurics within the broad geoscience community. Consequently, the first step in understanding estimates of $\overleftrightarrow{\mathbf{Z}}$ and $\mathbf{T}$ is assessing the extent and nature of their distortion. Approaches and methods developed over the past two decades to recognize and remove such distortion, primarily through tensor decomposition, enabled significant advances in reliable imaging of the subsurface electrical conductivity structure using magnetotellurics, as discussed in Chapter 6.

For most geologically plausible structures, numerical methods are required to solve the forward problem of predicting the magnetotelluric responses that would be observed at arbitrary locations and frequencies given a hypothetical model of Earth's resistivity structure. This is straightforward for 1D structures, as described in Chapter 2, and requires the use of sophisticated numerical finite-difference, finite-element or integral equation methods in

2D and 3D, as presented in Chapter 7. Solving the forward problem is an essential element in the second step of magnetotelluric data interpretation, but is only a means to an end.

Simply producing a model that qualitatively fits the measurements does not allow the magnetotelluric practitioner to fully assess and interpret a given set of measurements. This requires consideration of the inverse problem of inferring Earth's conductivity structure on the basis of observed magnetotelluric responses at specific locations and frequencies. Solving the magnetotelluric inverse problem involves finding one or more models of conductivity whose predicted responses fit the observed responses. The notion of "fit" must take into consideration measurement errors in the observations, computational errors in theoretical predictions, and even the appropriateness of the model itself. This is exactly the same notion as used to assess whether a given model of distortion is appropriate for a dataset. In most circumstances, a wide range of conductivity models will provide acceptable fits to a set of magnetotelluric observations. In other words, solutions to the inverse problem are non-unique, and it becomes necessary to impose additional constraints on the model or to find a concise characterization of the range of acceptable models. The revolution in magnetotelluric inversion to deal with such non-uniqueness is covered in Chapter 8.

In magnetotelluric surveys, it is essential to acquire high-quality data over a broad frequency or period range. The components of a magnetotelluric recording system can be divided into three parts: electrometers for sensing electric fields (or, more correctly, potential differences), magnetometers for sensing the magnetic field (or, more correctly, the magnetic induction) and recording/timing units for controlling the timing, digitization, filtering, and recording of the data, all of which have improved dramatically in recent decades. However, modern instrumentation alone will not yield time series of high fidelity; this also necessitates thoughtful site selection, careful installation and thorough field documentation that comprise state-of-the-art magnetotelluric field procedures. The advances in magnetotelluric data acquisition are documented in Chapter 9.

Chapter 10 concludes the book with a review of a selected set of case histories, with the emphasis placed on illustrating many of the analysis and modeling methods described in Chapters 2–8, and using magnetotelluric data either alone or with other geophysical data to make geological inferences about Earth. The selected examples span near-surface to crustal to mantle studies on both land and the seafloor, and provide an up-to-date set of examples of what can be accomplished using the magnetotelluric method.

1.3 Historical perspective

Electromagnetic induction studies have their theoretical origin in the spherical Earth treatment of Lamb (1883). Subsequently, Schuster (1889) applied Lamb's theory to data from globally distributed geomagnetic observatories to show that Earth is a conducting body. Chapman (1919) and Chapman & Price (1930) produced spherically symmetric but disparate models for the electrical resistivity of Earth, consisting of a thin surface insulating shell overlying 28 or 2.3 Ω-m conductors, respectively. The difference was attributed to a decreasing resistivity with depth because the Chapman & Price study used longer-period data than the earlier Chapman analysis. In a sequence of now classic studies, Lahiri & Price (1939) and Rikitake (1950, 1951a,b,c,e) extended these results to produce more detailed, radially varying models for the resistivity of Earth. By the early 1980s, the ability to resolve large-scale lateral variability in Earth structure emerged, as reviewed by Tarits (1994). All of these studies used the *geomagnetic depth sounding* (GDS) method based on measurements of the geomagnetic (but not the geoelectric) fields at periods of a day to a week or more. Under the assumption that the driving external source is dominantly an axisymmetric ring current flowing many Earth radii above the geomagnetic equator, the fundamental GDS datum is the ratio of the vertical to the horizontal magnetic field as a function of period, which may be used to infer the electrical resistivity as a function of depth and, in some cases, location. It differs significantly from magnetotellurics in many important respects, not least of which is that classical GDS operates at substantially longer periods, hence is primarily sensitive in the mid-mantle transition zone (400–600 km), and will not be further covered in this book. However, from a historical perspective, it is important because the theory for GDS is similar to that for magnetotellurics, and the academic rigor from GDS studies provided much of the theoretical basis for magnetotellurics.

In parallel to the evolution of GDS, electrical geophysics emerged as a viable tool in the exploration for petroleum and conductive mineral ores beginning in the 1920s with the Schlumberger brothers. In contrast to GDS, exploration electrical geophysics typically utilizes an artificial source to inject or induce electric currents in the subsurface. As a result, the physical scale of structures that can be studied is much smaller than with GDS, and in fact is typically limited to, at most, the uppermost crust. A comprehensive survey of electromagnetic exploration methods appears in the two-volume compendium edited by Misac Nabighian (1987, 1991) that remains the standard reference today. Electrical exploration methods differ substantially and practically from magnetotellurics in many ways, but represent a field that was more rapidly evolving than academic GDS. The combination of the need to span the gap lying between the exploration and global scales, and the constant drive for innovation in the exploration world, contributed to the emergence of the magnetotelluric method by the middle of the twentieth century.

In the nineteenth century, descriptions of geomagnetically induced fluctuations on telegraph cables led to the simultaneous measurement of the terrestrial electric and magnetic fields at the Royal Observatory, Greenwich (Airy, 1868). While their relationship was not studied quantitatively, Airy did note that the electric field frequently led the magnetic field, and inferred a subsurface origin for the field variations. This observation led a German research team to make measurements of the time-varying magnetic and electric fields during their contribution to the 1887 First International Polar Year in the Arctic. These data, the first ever field campaign magnetotelluric data recorded, were analyzed in the mid-1980s by Jones & Garland (1986). Subsequently, in an attempt to explain the polarization and temporal behavior of short-period (<2 h) geomagnetic fluctuations such as bay disturbances and micropulsations, Terada (1917) calculated the phase difference between the vertical and horizontal components of geomagnetic variations assuming that it was caused by electric currents in Earth. The relationship between orthogonal electric and magnetic field components in (1.4) was introduced by Hirayama (1934), and was at one time known as Terada's relation in Japan (Hirayama, 1934; Rikitake, 1951d). Hirayama (1934), and later Rikitake (1948), showed that the *E*/*B* amplitude is proportional to the square root of period with a constant phase difference of 45° for a uniform half-space. However, Hatakeyama & Hirayama (1934) reported simultaneous measurements of time-varying surface electric and magnetic fields at Toyohara, Japan (now Yuzhno-Sakhalinsk, Russia), showing that their phase difference varies with period, being smaller than 45° at long periods. The recurring observation that the measured *E*/*B* amplitude and phase changes with period was rigorously explained by Kato & Kikuchi (1950a,b) and Rikitake (1951d) through a varying resistivity with depth under the assumption of a plane-wave source outside the atmosphere. The content of Rikitake (1951d) was first presented at a Japanese national meeting in October 1949.

Geophysical tradition holds that the theory of magnetotellurics was proposed simultaneously and independently by Tikhonov (1950) in the USSR and Cagniard (1953) in France. Cagniard notes in the Acknowledgements to his paper that: "The theoretical work reported in this paper was done some time ago and has been mentioned in applications for patents which have been made in several countries to protect the new prospecting method involved. Because of the potential practical applications I have had to postpone any publication related to magneto-telluric phenomena for many years." But he goes on to say that: "Meanwhile, the Russian scientist Tikhonov, and the Japanese scientists Kato, Kikuchi and Rikitake had also recognized the existence of such an effect. To my knowledge, they have not pointed out the possibility disclosed by my work of applying these results to practical geophysical exploration. They have, however, paid attention to their possible use for investigating the electrical conductivities of very deep regions in the Earth's crust." Thus, although Cagniard deserves the credit for appreciating that magnetotellurics could become a geophysical exploration tool for resources and gives practical examples of its use, the key concepts underlying magnetotellurics originated earlier in Japan than in Europe, and had achieved an advanced level of development and application by the early 1950s. Rikitake deserves as much of the credit for its development as the customary choices of Tikhonov and Cagniard, although magnetotellurics is really the cumulative result of research by a number of investigators over the first half of the twentieth century.

It should be noted that Tikhonov (1950), Rikitake (1951d) and Cagniard (1953) are all physically wrong because they are based on the concept of a plane-wave source that cannot be responsible for electromagnetic induction, as shown in Chapters 2 and 4. However, they remain mathematically correct, and constitute the historical basis for the developments in the subsequent 60 years, with primarily those of the past 25 years comprising the subject of this book.

However, controversy over the plane-wave source model continued into the 1960s. Wait (1954, 1962) cast aspersions on the validity of the assumption of plane-wave sources, since finite ionospheric sources do not give rise to normally incident plane waves. Further, magnetotelluric results inferring conductivity distributions well into the mantle (Migaux *et al*., 1960) did not agree with geomagnetic depth sounding analyses, and the difference was sometimes ascribed to finite source fields. This led Price (1962) to develop a general theory for the magnetotelluric method with a finite dimension for the source fields. Srivastava (1965) extended the plane-wave recursion algorithms of Wait (1954) to include the effect of finite source dimensions, and gave a curve-matching technique for estimating not only the conductivity layering parameters but also the scale of the source when both horizontal components of the magnetic field are measured.

Computer modeling studies undertaken by Madden & Nelson (1964) and Srivastava (1965) indicated that, for realistic Earth conductivity profiles, the plane-wave source field assumption is valid for periods up to 10^3 s. Swift (1967) analyzed magnetograms from two stations 1300 km apart (Dallas and Tucson) at mid-geomagnetic latitude, and concluded that, for the period band 10^3–10^5 s, the wavelength of the source field must be in excess of 10^4 km. This corresponds to a wavenumber ($k = 2\pi/\lambda$) of the order of 10^{-5} m^{-1}, and therefore from Srivastava's (1965) studies, the plane-wave mathematical assumption is valid at mid-geomagnetic latitudes for periods up to 10^5 s. This conclusion has stood the test of time, and it is only in unusual locations (e.g. beneath the equatorial electrojet or during auroral activity at high latitudes) that finite source fields seriously bias magnetotelluric data.

During the 1950s and 1960s, it also became apparent that the scalar nature of the magnetotelluric response from Tikhonov (1950), Rikitake (1951d) and Cagniard (1953) was inadequate for Earth studies. Neves (1957) appears to have been the first to recognize the tensor nature of the relationship between the electric and magnetic fields, and defined a finite-difference algorithm for solving the 2D magnetotelluric forward problem.

The authors have chosen to stop the historical perspective at this point out of deference to living colleagues who played a role after about 1960, and in the belief that the authors of the chapters in this book have provided their perspective about each specialty area that should take precedence.

1.4 Commercial use of magnetotellurics

Although almost all of the authors of the chapters in this book are academics, magnetotellurics is used in industry to a far larger extent than in academia. Large academic magnetotelluric groups have perhaps 10 or so broadband magnetotelluric (BBMT) systems, whereas there are over 500 BBMT systems from one manufacturer in continuous use in China alone for oil and gas exploration.

As far as the editors of this book are aware, the first commercial use of magnetotellurics was for geothermal exploration in the USA in the late 1950s and early 1960s. Magnetotellurics was in significant use in the Soviet Union during the 1970s, with over 100 crews actively recording data primarily for hydrocarbon exploration (Spies, 1983), although the majority used analog equipment and determined the apparent resistivity in