Combined Transient Electromagnetic and Magnetotelluric

Study of the Southern Kenya Rift Valley

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ABSTRACT

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The transient electromagnetic (TEM) method and the magnetotelluric (MT) technique have been applied to determine the electrical resistivity structure across the southern Kenya Rift Valley. The main profile extends from the shores of Lake Victoria, west of the Rift Valley, to the north of the Chyulu Hills volcanic chain, 150 km SE of the rift. A second profile runs parallel to the Chyulu Hills volcanic trend. Data from 19 stations along the two profiles have been processed using classical techniques and in the case of MT, analysed with modern tensor decomposition methods.

The TEM data have facilitated the removal of static shift effects from the MT data and recovery of the near-surface (<300 m) geoelectric structure. One-dimensional joint inversion of TEM and MT data yielded an approximate geoelectric structure for the region. Subsequent two-dimensional modelling has revealed a more realistic resistivity distribution for the complex environment of the Kenya Rift.

A resistive (>2000 $\Omega$ m) Archaean crust 30 km thick, with a 10-12 km mid-crustal conductive (=100 $\Omega$ m) zone, resting on a moderately resistive (=100 $\Omega$ m) mantle appears at the west end of the main profile. A conductive fault-like zone extending to mantle depths in the area of the Oloololo Escarpment coincides with the exposed boundary between the Archaean Nyanza Craton and the Proterozoic Mozambique Belt. A poorly constrained highly resistive (>10000 $\Omega$ m) (Proterozoic ?) crust is found at the western flank of the rift. Low resistivities (<50 $\Omega$ m) are found down to the base of the crust in the rift zone and are possibly due to the presence of sedimentary fill deposits at shallow depths, and the presence of magmatism and partial melt at deeper levels. East of the rift a less sharply defined geoelectric margin, offset from the accepted topographic and geologic boundary of the rift, marks the transition to a more resistive (=1000 $\Omega$ m) Proterozoic crust. Significantly enhanced conductivities (<100 $\Omega$ m) are implied in the complex 3-D region of the Chyulu Hills.

The 2-D model is shown to be consistent with previous geophysical models for the area but offers new insights into the structure and tectonics of the Kenya Rift Valley.
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Chapter 1

Introduction

An important objective in geophysical exploration is the determination of the earth's structure and physical property variations using surface measurements. One such geophysical method is electromagnetic (EM) induction soundings, which indirectly measure the electrical conductivity (the reciprocal of resistivity) of the earth.

The electrical resistivity of a rock is a measure of its ability to impede the passage of electrical currents, and is affected by rock composition, porosity, permeability, presence and salinity of fluids. A relation between bulk rock resistivity, porosity and fluid resistivity is given in Archie (1942). The electrical resistivity of materials within the earth can vary by over several orders of magnitude ($10^{-1}-10^{6}$ $\Omega$ m, Haak and Hutton, 1986; their Figure 1), especially when there are temperature variations or in the presence of specific minerals and volatiles (in connected pore), which is the widest range of any of the physical parameters that can be remotely sensed from the surface of the earth (Jones, 1993). Major changes in the resistivity of the rocks can be measured at the earth's surface by a number of different methods. The geophysical techniques for each resistivity determination (and depth probed) fall into the groups summarized in Table 1.1. They all have wide applications. Of particular importance is the determination of structural and compositional information about the earth from lateral resistivity variations. This particular study is mainly concerned with the application of a natural source technique known as the Magnetotelluric method (MT) to the study of the Kenya Rift, but additionally requires an artificial source technique known as the Transient Electromagnetic (TEM) method to constrain the shallow resistivity structure.

In this chapter, a brief historical background of the magnetotelluric method and the resistivity structure of the earth's crust and upper mantle is given, as well as an overview of the published relevant geological and geophysical studies of the region of interest. The thesis outline and the research objectives of the study are also presented.
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Introduction

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1.1 The Magnetotelluric method as an exploration tool

Magnetotellurics is a geophysical exploration technique employing simultaneous measurements of natural transient electric and magnetic fields to infer the electrical conductivity distribution within the earth beneath the site of the surface fields. The MT method is a relatively new geophysical technique. It was first proposed by Cagniard in 1953, although Tikhonov (1950), Kato and Kikuchi (1950) and Rikitake (1950) had published on the subject of deep crustal electrical conductivity investigations without applying the results to practical geophysical exploration. Even so, the fundamentals of the method are a direct consequence of the classical electromagnetic theory established by the work of Ampere, Faraday and Maxwell. A useful collection of MT papers describing the method up to the early 1980s can be found in Vozoff (1986).

The method differs from other EM geophysical techniques that require artificial sources of EM energy to probe the earth. The naturally occurring interminable variations in the earth’s magnetic field induce eddy currents, called telluric currents, in the conductive crust of the earth that are detectable as electric field variations at the surface (and within the earth). If the variation in the magnetic field is assumed to be derived from a plane EM wave propagating vertically into the earth, the EM impedance (i.e. the ratio of the horizontal electrical field in the
ground to the orthogonal horizontal magnetic field), measured at a number of frequencies, 
gives earth resistivities as a function of frequency or period resulting in a form of depth 
sounding (Cagniard, 1953).

A major advantage of the MT method is that its depth range of investigation is not restricted 
by the length of the measuring cables or the size of the power source. The electrical resistivity 
variation at deeper levels within the earth is simply measured by using increasingly lower EM 
wave frequencies. Also, MT surveys can locate vertical boundaries (Jones, 1987) and can 
investigate areas with deep sedimentary basins (Vozoff, 1972) or areas hampered by surface 
volcanics.

The MT method, like all geophysical techniques, does have its problems and limitations. To 
summarize, some of the more important ones are as follows:

- Its resolving power at depth is inherently limited by the attenuation of high frequency 
  components of the EM field by the presence of overlying rocks.

- MT soundings cannot resolve sharp boundaries or thin layers except with highly precise 
  response functions (Jones, 1987); the diffusive nature of the energy propagation “smears 
  out” the real earth structure.

- Models based on the estimated impedance as a function of frequency invariably produce 
  non-unique information on the resistivity variation with depth. Modelling is further hindered 
  by the presence of anisotropy, near surface distortions, frequency independent (static) shift 
  of the apparent resistivity sounding curves and topography.

- The minimum and the maximum depths of investigations are variable as they depend upon 
  the frequency range of the observations collected and the geoelectric structure of the site.

- The resistivity cannot be used to give an unambiguous identification of lithology as it 
  depends upon several different rock properties.

Some of these problems will be addressed in this study.

There are a number of case studies which describe the application of the MT method for 
the exploration of different geological targets (see for example the reviews by Hermance, 1983, 
and Chave and Booker, 1987). Since magnetotellurics can readily acquire information about 
the earth’s deep interior, a number of global induction studies have been undertaken (e.g. 
Madden and Swift 1969; Rooney and Hutton, 1977; Park et al., 1991; Wannamaker et al., 
1997). On a more localised scale, studies have been performed on granite intrusions (e.g. 
Beamish, 1990a; Jones P., 1992) and major thrust and fault zones (e.g. Hill, 1987; Meju, 1988); 
also MT has been used in mineral exploration (e.g. Strangway et al., 1973; Strangway, 1983); 
in basin evaluation for petroleum exploration (e.g. Tikhonov and Berdichevsky, 1966; Vozoff,
1972; Christopherson, 1991); and in civil engineering, ground water and archaeological problems (Guineau, 1975). The strong dependence of resistivity on both temperature and the presence of water has meant that MT has frequently been used for geothermal exploration (e.g. Hutton et al., 1984; Galanopoulos, 1989). With the advent of broad band MT recording systems, particularly at higher frequencies, increasingly more studies have been concerned with investigation of the shallower structure of the earth.

1.2 Electrical resistivity of the crust and upper mantle

Over the last few years, numerous reviews of the resistivity structure of the continental crust have been published (e.g. Shankland and Ander, 1983; Haak and Hutton, 1986; Gough, 1989; Jones, 1992). In the early 1970's it was generally believed that there was a gradual increase in resistivity with depth in the crust, due to the high resistivity of the silicate rock matrix and a gradual decreasing percentage of fluid, as pore space decreased with depth (Brace, 1971). The increased number of MT studies showed that the middle to lower continental crust is generally characterized by a sharp decrease in resistivity (Shakland and Ander, 1983; Haak and Hutton, 1986; Hyndman et al., 1993). Hyndman and Shearer (1989) and Hyndman et al. (1993) treated as “lower crust” those parts of the crust which showed significantly low resistivity relative to the resistivity of dry igneous rocks at the same temperature but at atmospheric pressure. This corresponds to rocks at temperatures of at least 350°-400°C and minimum depths of between 15 and 35 km.

In most regions, the upper crust (<13 to 20 km) is relatively resistive ($10^7$ to $10^5$ $\Omega$.m). The values vary with the type of the crust, from more resistive ($10^3$ to $10^5$ $\Omega$.m) in Precambrian shields to less resistive ($10^2$ to $10^3$ $\Omega$.m) in regions which have been tectonically active in the Phanerozoic (Gough, 1986; Hyndman et al., 1993). Gough (1986) proposed that the electrical state of the crust depends on the different stress regimes, and the differences between shields and tectonically active crust depends on the density of fractures more than the resistivities of the fluids.

Limitations of the magnetotelluric method make it difficult to uniquely characterize the resistivity decrease in the crust because only the depth to the top and the conductance (the conductivity-thickness ratio) of the layer are well defined (Hyndman and Shearer, 1989). In Phanerozoic continental margins, the upper resistive crust extends to depths of 13 to 20 km (e.g. Jones, 1992). In active areas, zones of low resistivity rise to depths of 10 km or less (Jones, 1992). In Precambrian shields, an extensive zone of low resistivity is not generally encountered above 25 km (Nesbitt, 1993). Jiracek et al. (1979) from studies of four rifts (Rio
Grande, East Africa, Baikal, and the Rhinegraben) found low resistivity zones (<50 $\Omega$.m) at depths of 10-30 km. The variations in depths to the top of the low resistivity zones appear to be related to variations in regional heat flow and it has been proposed that they correlate with crustal temperatures of about 400±50°C (Shankland and Ander, 1983; Gough, 1986; Hyndman and Shearer, 1989).

Gough (1986), Jones (1987, 1992), and Hyndman and Shearer (1989) have emphasized that the zones of low resistivity in the crust often coincides with zones of relatively high seismic reflectivity. In addition, the top of the reflective and conductive middle crust is often coincident with the bottom of the zone of earthquake activity (Gough, 1986). This suggests that the transition from the upper resistive to the middle conductive crust correlates in general with the brittle to ductile transition in rock rheology (Bailey, 1990), and this is also believed to occur around 400°C (Schwarz, 1990).

Many causes of the observed enhanced conductivities have been proposed during the last two decades, the following are the four most likely candidates (Shankland and Ander, 1983, Haak and Hutton, 1986; Gough, 1986; Wannamaker, 1986; Frost et al., 1989; Bailey et al., 1989; Bailey, 1990; Newton, 1990; Jones, 1992; Hyndman et al., 1993; Duba et al., 1994):

- water - aqueous fluids (saline fluids)
- carbon grain - boundary films
- conducting minerals, and
- partial melts.

Temperatures in the lower crust (e.g. Hyndman et al., 1993) are above the critical temperature (374°C for pure H$_2$O) so the term aqueous fluid is preferred instead of water since liquid and gaseous H$_2$O are indistinguishable (Jiracek, 1995). Early interpretations suggest that low resistivity zones, especially in tectonically active areas such as rifts, were caused solely by magma (e.g. Banks and Beamish, 1979). Recent studies (e.g. Jodicke, 1992; Gough, 1992; Nesbitt, 1993; Katsube and Mareschal, 1993; Hyndman et al., 1993) proposed that aqueous fluids or conductive minerals, mainly graphite are more likely to cause the high conductivities in the lower crust. However, petrological evidence has strongly suggested that the lower crust is very dry, and no free connected fluid phase can be present in deep stable crust (Yardley, 1986). Moreover, they question the proposal that low crustal conductivity is due to thin grain boundary films of graphite, since films of sufficient thickness would be readily visible on broken surfaces in hand specimens, and this is not the case in most of the investigated rock samples (Yardley and Valley, 1997).
For H\textsubscript{2}O to accumulate in the crust, there must be both a source and a capping mechanism. Sources include deep circulating meteoric water, fluids released from metamorphic dehydration reactions in progressively buried or heated rocks, and fluids released from mantle derived magmas crystallizing higher up in the lithosphere (e.g. Hyndman and Shearer, 1989; Marquis and Hyndman, 1992; Jones, 1992). Additionally, mantle devolatilization processes are a poorly defined but potentially large source of water for the lower crust (e.g. Kay and Kay, 1986). In active regions (rift structures and continental hot spots) localized zones of mantle upwelling and higher than normal mantle temperature where water is carried in partial melts are the prime candidates for water sources (Hyndman and Shearer, 1989).

Bailey's (1990) calculations proved that aqueous fluids have residence times that are very short in geological terms, so their existence in deep stable continental crust is precluded. To explain the fluid retention Etheridge et al. (1983) postulated the existence of an impervious cap that is sealed by precipitation in pores and fractures. However, this model has not been generally accepted (e.g. Yardley, 1986). Frost and Bucher (1994) argue that this model is not valid since there is no geologic evidence for the presence of this impermeable cap that will entrap free aqueous fluids for a long periods of geological time. A trapping mechanism for aqueous fluids in cracks at the brittle ductile transition, a theoretical model was presented by Bailey (1990). According to him, the vertical fluid transport below the brittle-ductile transition zone is rapid, but in the cooler crust, above the transition zone, it slows abruptly. The resulting overpressure induces horizontal hydrofractures in which water will accumulate. However, Jiracek et al. (1995) pointed out that Bailey's (1990) mechanism would operate most effectively where principal regional stress is horizontal (compressional) rather than vertical as it is in extensional rift environments.

Graphite, the reduced form of carbon, is highly conductive\(^{\dagger}\) with values of \(\approx 10^{-4} - 10^{-5} \ \Omega \cdot m\) at 500°C (e.g. Hyndman et al., 1993), so only small amounts are required to generate the typical bulk resistivities of the lower crust (Katsube and Mareschal, 1993). The source of graphite is usually thought to be from the reduction of CO\textsubscript{2}-rich fluids during cooling (Frost et al., 1989). Even though the amount of graphite needed to produce a conductive zone is small, there must be a connected film on regional scales (Katsube and Mareschal, 1993). Two of the main origins of deep graphite are (Schwarz, 1990; Jodicke, 1992; Hyndman et al., 1993): (i) organic material of sedimentary origin, carried to great depth and metamorphosed, and (ii) igneous and metamorphic processes, i.e. from reduction of CO\textsubscript{2} or some other deep source in igneous

\(^{\dagger}\) Graphite is more conductive (\(10^4 \ \Omega \cdot m\); Duba and Shankland, 1982) than highly saline water (\(10^2 \ \Omega \cdot m\); Nesbitt, 1993) which is more conductive than magma (0.5 \(\Omega \cdot m\); Hermance, 1979).
rocks. In a conductive rift environment the presence of graphite is not usually invoked as a source of low resistivities, since a continuous supply of deep fluids is expected and also there appears to be a strong correlation with temperature which is not expected for graphite (Jiracek et al., 1995).

It is important to notice that there are other possible causes of enhanced conductivity apart from fluids and graphite. Mareschal et al. (1992) found evidence of sulfur and iron on grain boundaries, indicating the possibility of other grain boundary phases, such as pyrrhotite, pyrite, or magnetite, being present as a grain boundary conductor (Frost and Bucher, 1994). However, these minerals are rarely dominant, so they are implausible explanations for the world-wide occurrence of lower crustal conductive layers (Merzer and Klemperer, 1992).

Finally, the occurrence of rock-melts in the crust is a potential cause of conductivity enhancements (connected melt along grain boundaries would lead to lowering of resistivities; Waff, 1974; Sato and Ida, 1984). Small percentage of partial melt is connected with high temperatures (above 700°C), and may only be found in tectonically active zones at crustal depth (Schwarz, 1990). Melts must be connected over large distances to explain the decrease of the resistivity on a large scale (Shankland and Waff, 1977).

The question that emerges why is the upper crust is resistive? Gough (1986) suggested that saline fluid permeating the whole crust is under compressive stress in the upper crust, and thus is in separate cavities, whereas in the lower crust the fluid forms interconnected films on the crystal surface. Sanders (1991), proposed that the resistant upper crust results from the absence of interconnected brine-filled porosity. Hyndman et al. (1993) proposed that at lower temperatures in the brittle upper crust the confining pressure will close the thinner interconnected grain boundary. Nesbitt (1993), explained that when H₂O-CO₂ solution cools to the low temperatures of the upper crust, a CO₂-rich phase separates from the aqueous solution. The bubbles of the CO₂-rich phase rise through fractures and pores and tend to create pockets of conductive brines which are electrically separated by resistive zones. Consequently the upper brittle crust is largely saturated with aqueous brine and is anomalously resistive due to insulating properties of discrete zones. For the graphite mechanism, the contrast between upper and lower crust resistivities requires that either the amount of graphite increases downward (Hyndman et al., 1993), or that the continuous interconnected film in the lower crust is broken by cooling when the rocks are uplifted to the upper crust (Frost et al., 1989).

Nevertheless, it has to be emphasized that there is no single cause for the crustal enhanced conductivity zones since different effects operate in different tectonic environments and at different depths (Jones, 1992).
From the information summarized above, there is an emerging picture of the continental crust being composed of a brittle, resistive, < 400°C upper crust with a relative rapid transition to a ductile, conductive, >400°C middle crust (Nesbitt, 1993). There is also evidence of a second transition around 700° to 750°C to a more resistive lower crust (Hyndman and Shearer, 1989; Marquis and Hyndman, 1992). These transitions can also be roughly correlated with metamorphic facies of upper greenschist at the first transition and upper amphibolite to granulite at the second (Nesbitt, 1993; granulite facies are presumed to have no free water and to be dry and resistive, the estimated temperature is about 730°C; e.g. Wannamaker, 1986).

It is assumed that the upper mantle in stable regions is fairly resistive (>100 Ω.m) (e.g. Jones, 1987; Hyndman et al., 1993; Banks et al., 1996). However, in tectonically active areas the presence of partial melt in the upper mantle may result in the decrease of bulk resistivity (e.g. Sato, 1986; Hjelt and Korja, 1993), although alternative models have also been proposed (e.g. Karato, 1990; Bai and Kohlstedt, 1992). In the latter case, the enhanced electrical conductivity is attributed to solid state conductivity of olivine if hydrogen is present (olivines and pyroxenesis are the most prominent and abundant upper mantle minerals).

1.3 The geology and geophysics of the Kenya Rift

This thesis deals with the magnetotelluric study of the southern part of the Kenya Rift Valley. This requires understanding of the continental rifting process and the geology of the Kenya Rift. This section therefore describes the mechanisms and process of continental rifting, and finally it proceeds to describe the Kenya Rift and the results of previous geophysical studies in the area.

1.3.1 Continental rifts

Rifting of the continents is one of the prominent intraplate phenomena which can ultimately lead to the break up of continents. The study of the process of rifting is useful for the understanding of the physical and rheological properties of the continental lithosphere, the forces which drive, deform and break the plates, and also for the evolution of sedimentary basins (Fuchs et al., 1994). Continental rifts are distinct from oceanic rifts in that they represent the genesis, rather than the mature stage of lithospheric extension (Morgan, 1997). Hydrocarbon, mineral and geothermal reserves associated with continental extension and basin formation are some of the geological and economic considerations in the study of the process and evolution of continental rifts.

Olsen and Morgan (1995) define a continental rift to be “an elongate tectonic depression associated with which the entire lithosphere has been modified in extension”. A rift system is
defined as a tectonically interconnected series of rifts. There are several examples of continental rift zones with a wide range of extensional styles. These include (Ruppel, 1995): passive margins (North Atlantic, Western Australia), discrete intercontinental rift zones (East African Rift, Rio Grande Rift, Baikal Rift), diffuse rifts (Basin and Range Province) strike-slip dominated rifts (Dead Sea Rift) and rifts in zones of regional compression (Tibetan grabens). The Kenya Rift is regarded as the typical active continental rift, implying that the extensional process is presently in action. Although all the rifts have unique characteristics, continental rifts can generally be characterized by some common features (Ruppel, 1995): (i) normal faulting with subsidiary strike-slip faulting, (ii) lithospheric thinning which outpaces crustal stretching, (iii) varying amounts of alkaline magmatism, (iv) heat flow locally elevated near faults and magmatic centers, and (v) crust that has experienced magmatic underplating and some amount of magmatic intrusion.

1.3.1.1 Rifting mechanisms

The main categories of mechanical rifting processes are pure shear, simple shear, combination mechanisms and lower crustal flow (Ruppel, 1995). Figure 1.1 shows the models of strain geometry in rifts. Each mechanism is based on specific assumptions about lithospheric rheology and has particular consequences for the thermal evolution of the rift zones, basin formation, subsidence and uplift history, and sedimentation patterns (Ruppel, 1995).

In the pure shear model, the crust and upper mantle are uniformly attenuated along any given vertical reference line (McKenzie, 1978). There are two main stages: (i) rapid stretching which produces thinning, block faulting and subsidence and (ii) thickening due to the heat conduction to the surface followed by slow subsidence. It should be noted that there is a close relationship between heat flux and subsidence and that the asthenosphere plays a passive role during the lithospheric thinning. The pure shear thinning formulation represents a simplification of extensional process and does not account for the complicated brittle-over-ductile rheology of the crust and upper mantle.

The simple shear mechanism emphasizes the role of brittle faulting, localized strain (along fault zones), and predominantly crustal deformation in the evolution of rifted terrains. The relative extension of crust and mantle lithosphere along any given vertical line is non-uniform. The main stages are: (i) a low angle (<30°) normal fault penetrates most of the crust (and possibly the mantle) (Wernicke, 1985), (ii) the upper-plate may extend relative to the surrounding region, and (iii) isostatic rebound occurs in the unloaded terrains. In the simple shear model thinning of the crust occurs offset from the area of maximum lithospheric thinning. Magmatism and topographic uplift generated from the upwelling asthenosphere may be located beneath the zone of thinned lithosphere, far removed from the region of crustal deformation.
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Figure 1.1: Models of strain geometry in rifts. (a) The *uniform pure shear* model (McKenzie, 1978). Rapid uniform stretching of the entire continental lithosphere produces thinning and passive upwelling of hot asthenosphere. The lithosphere consists of a crust of thickness $h$ and mantle lithosphere of thickness $l$. Extension is assumed to be confined to a zone with original width $x_e$, and is defined by $\beta$, where $\beta$ is the ratio of the initial to the final thickness ($\beta>1$ for extension). Instantaneous lithospheric stretching raises the geothermal gradient, and asthenosphere passively upwells beneath the rift. (b) Wernicke’s (1985) *simple shear* faulting mechanism. A low-angle large scale detachment penetrates the whole crust, and terminates at a ductile shear zone within the mantle lithosphere. The maximum thinning of the lithosphere occurs offset from the most extended crustal rocks. (c) Combined pure shear and simple shear model (after Kuszmir and Ziegler, 1992). Extension takes place on planar faults in the upper crust, but by distributed plastic deformation (pure shear) in the lower crust and mantle.
In the combined pure shear and simple shear model faulting is dominant in the upper crust, and distributed plastic deformation occurs in the lower crust and mantle. This model includes (Kusznir et al., 1991): (i) crustal thinning during extension by simple shear in the upper crust and pure shear in the lower crust and mantle, (ii) the lithosphere temperature field is perturbed during extension and re-equilibrated after, and (iii) a flexural-isostatic response of the lithosphere to crustal thinning and thermal loads both syn- and post- rifting event. The flexural cantilever model has been successfully applied to parts of the western (Kusznir and Ziegler, 1992) and eastern (Hendrie et al., 1994) branches of the East African Rift System.

The lower crustal flow model suggests the ductile lower crust may have sufficiently low viscosity to flow laterally from unextended to extended regions in response to horizontal pressure gradients (Ruppel, 1995) induced by localized thinning of the upper crust (Artyushkov, 1973). Lack of constraints on the range of values of lower crustal viscosity limits the understanding and modelling of this process.

The driving force involved in the continental rifting process has lead to the classification of the rifts into two categories (see Ruppel, 1995; Bott, 1995 and references therein): passive rifts (driven by distant plate boundary forces) and active rifts (driven by sub-lithospheric mantle dynamics). Basically, both passive and active rifting occur in response to extension in the lithosphere, but the source of the extensional driving force differs (Figure 1.2).

The passive rifting mechanism results from plate interaction moving under the influence of large scale convective flows in the mantle, i.e. the rifting is due to convection at a distance (Ruppel, 1995). This associates rifting with extensional stresses in the lithosphere at zones of weakness; in the passive case the driving force is the plate motion and the lithosphere is thinned only in response to extension (Olsen and Morgan, 1995).

Active rifting is defined as rifting in response to a thermal upwelling of the asthenosphere as result of lithospheric thinning (e.g. McKenzie and Bickle, 1988; White and McKenzie, 1989). The causative stresses of rifting are associated with lateral thermal density variations in the lithosphere and the underlying asthenosphere (Olsen and Morgan, 1995). In active rifting the lithosphere is thermally thinned by heating and absorption into the asthenosphere, in addition to necking in response to extension, and hence the volume of asthenosphere rising into the lithosphere exceeds the volume of lithosphere displaced laterally by extension.
1.3.2 The Kenya Rift

The Kenya Rift is an active continental rift that has developed since the Oligocene - Lower Miocene (Morley et al., 1992); it is part of the eastern branch of the East African Rift System, itself part of the Afro-Arabian Rift System which extends for 6000 km from Syria in the north,
traversing through the Jordan Valley, Dead Sea, Red Sea, Gulf of Aden, Afar, East Africa and terminating in a large number of splayed faults in southern Africa (Prodehl et al, 1997).

The East African Rift System (Figure 1.3) is an assemblage of connected rift segments, extending for 3200 km from Afar triple junction in the north to the Zambesi river in southern Africa (Achauer et al., 1992). There is a distinct bifurcation north of the Nyanza Craton in Uganda, splitting the rift system into a western branch (Albert Rift, Tanganyika Rift) and an eastern branch (Kenya Rift). The active rift systems of East Africa have developed above two broad plateaux: the Afar Rift and main Ethiopian Rift sectors transect the 1000 km wide Ethiopian plateau, while the Eastern and Western Rifts cut the 1300 km wide East Africa plateau (Ebinger et al., 1997).

The Afar region is the most complicated part of the East African Rift system. The Red Sea, Gulf of Aden and East African Rift systems intersect in a 300 km wide zone of deformation within the Afar depression (Ebinger et al., 1997) which has been interpreted as an assemblage of rotated micro-plates (Acton et al., 1991). The main Ethiopian Rift begins in the southern Afar and extends southwards to the Kenya border. The Ethiopian Rift crosses the Ethiopian plateau, a broad area of uplift, covered by a thick sequence of Eocene to Recent eruptive volcanics and is thought to overlie a mantle plume (Mohr, 1987; Smith, 1994). South of the Ethiopian Rift, the East African Rift System splits into two branches, the Western and Eastern (Kenya) Rifts, which have no clear surface connection. These rifts skirt the elevated East African Plateau underlain by the strong Nyanza Craton. The presence of a mantle plume (Ebinger et al., 1989), may have caused the uplift and resulted in the bifurcation of the rift into the Western and Eastern branches in the surrounding weaker Proterozoic lithosphere. The Western branch of the rift is characterized by deep asymmetric basins bounded by 50-100 km long normal border fault systems, with estimated extension ~10 km in an east-west direction (Ebinger et al., 1989). Lake Tanganyika and Lake Malawi are located in the rift segments. The southern part of the spoon-shaped Western rift loses its original N-S trend via a number NW-SE trending segments and interbasinal accommodation zones, providing a component of strike-slip faulting (Ebinger, 1989). The Usangu Rift, in southern Tanzania, branches north-east towards the southern end of the Eastern branch of the East African Rift System. South of this region lies the N-S orientated Malawi Rift, confining the 400 km Lake Malawi. South and west of Lake Malawi the structure and the seismicity suggest the early stages of continuing rift propagation in this direction (Girdler, 1991).
The Kenya Rift (see Figure 1.4) is one of the most well studied magmatic continental rifts. The southern part of the rift is known also as Gregory Rift, since the first detailed description of the rift came from Gregory (1921) in his book "The Rift Valleys and Geology of East Africa". The Kenya Rift traverses Kenya from Lake Turkana in the north to Lake Magadi in the south and dies out in northern Tanzania. The Turkana region of northern Kenya lies in the topographic depression between the Ethiopian and East African plateaux (Morley et al., 1992). This area was rifted in the Cretaceous - Palaeocene (> 56.5 Ma) forming the Anza graben and was probably part of the Karoo rift system across central Africa related to the break up of Gondwana (Hendrie et al., 1994). The Kenya Rift cuts across this structure with a N-S orientation, as a
wide zone of faulting and volcanism. Lake Turkana covers most of this area and the rift floor is about 400 m above sea level.

The central part of the Kenya Rift bisects the Kenya Dome, a topographic high rising from 300 m in Turkana to more than 1900 m between Nakuru and Naivasha (Figure 1.4) and decreasing in altitude to 500 m around Lake Magadi. Smith (1994) suggests the presence of limited (<1 km) crustal uplift which formed prior to the initiation of the volcanism in the Kenya Rift. In this central region the rift consists of a complex series of grabens 50-70 km wide bounded by major normal faults arranged en echelon, with escarpments of up to 1600 m high. The graben floor is characterized by younger, dense, sub-parallel faults, and is dotted with volcanic centres, hydrothermal systems and a number of both saline and freshwater rift valley lakes (Birt, 1996).

South of the Kenya Dome the altitude of the rift floor decreases to 600 m towards lake Magadi and the rift zone widens to 70-100 km. The sharp Nguruman escarpment (Figure 1.4) forms the western boundary of the rift in this area, the eastern margin is made up of a series of gentle step-faults. South from lake Magadi in northern Tanzania the morphology changes into a wide region of diffuse faulting with less volcanic activity and many diverging rift segments (Birt, 1996).

1.3.2.1 Geology of the Kenya Rift

The study area is located in the southern part of the Kenya Rift. The profile extends from Lake Victoria in the west to the southern part of the Chyulu Hills in the south east. Figure 1.4 shows the main geological units of the Kenya Rift. The area includes outcrops ranging in age from the Precambrian (>570 Ma) to Recent. The basement is a complex of metamorphic and igneous rocks of mainly Precambrian age. The Archaean Nyanzian Craton at the western end of the line is overthrust by the Mozambique Orogenic Belt to the east (Figure 1.5) (Smith and Mosley, 1993). The volcanic activity related to the rifting process started in the Middle Miocene (16-20 Ma) (Morley, 1994). Sediments are exposed in the western end of the line (Lake Victoria) and in the basins of the rift floor. The Chyulu Hills volcanic field is situated about 150 km east of the Gregory Rift, near the border of Tanzania and just 40 km north of Mt. Kilimanjaro. The area around Chyulu Hills is a peneplain with an average elevation of 1000 m that contains metamorphic inselbergs and the narrow Quaternary chain of volcanoes. Another major geological feature in this area is the 300 km long Yatta Plateau (Figure 1.4) which is formed by a remnant of a Miocene phonolite flow (Smith, 1994).
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**Precambrian basement**

The development of the Kenya Rift zone has been strongly influenced by two factors (Smith and Mosley, 1993): first its location above a major lithospheric heterogeneity, and second, a framework of crustal-scale ductile/brittle shear zones in the Precambrian basement.

The rift has developed close to the boundary between the Archaean Nyanza Craton in the west and the Proterozoic Mozambique Belt in the east (Figure 1.5). The Nyanza Craton is composed of low-grade volcano-sedimentary ophiolitic terranes (2.9-2.5 Ga) (Smith and Mosley, 1993); lithologies range from basic to intermediate and acid and mainly consist of granulites, greenstone associations and granitoids including narrow metasedimentary fold belts. The Mozambique Belt trends N-S through Tanzania and Kenya, with a well defined western front, a zone of mylonites and thrusts separating it from the Archaean rocks of the Tanzania Craton (Shackleton, 1993). The Mozambique Belt is suggested to include Archaean rocks (2.5 Ga) and younger nappes (900 Ma). It exposes metasediments which include conspicuous shelf-facies quartzites, marbles, migmatites and graphitic schists and gneisses (Shackleton, 1986). It has been affected by a complex sequence of metamorphic episodes as a consequence of the collision of the Nyanza Craton in the west and the Kibaran craton in the east. This is interpreted as a continent-continent late Proterozoic collision (Shackleton, 1986; Smith and Mosley, 1993).

The collisional event can be divided into three main phases (Smith and Mosley, 1993). An early tectonothermal event (850-720 Ma) was associated with large scale horizontal tectonism and overthrusting of the craton margin, a period (635-550 Ma) of sinistral deformation along the craton margin, and finally end-collisional events (530-430 Ma) that developed NW-SE trending shear-zones cutting obliquely across the boundary. The Nandi fault and more generally the Aswa Shear zone mark the outcropping western boundary of the Mozambique Orogenic Belt on the western flank of the rift in central Kenya and the Oloololo escarpment in the south (Figure 1.5) (Smith and Mosley, 1993). Note, however, that Smith and Mosley (1993) and Smith (1994) argue that the true position of the craton margin in the southern Kenya is obscured by gravitationally collapsed nappe structures in the region of the Mara-Loita Hills to the west of the rift, and the buried Archaean basement extends eastwards from the outcropping boundary at the Oloololo Escarpment influencing the volcanism and the rift development (Figure 1.5).
Figure 1.4: Geological map of Kenya (after Birt, 1996). Cenozoic volcanics and recent sediments cover large parts of central and eastern Kenya, overlying Precambrian basement. Top right: Location map of the East African Rift System showing the major rift faults. The Archaean Nyanza Craton is shown shaded.
**Chapter 1**

**Introduction**

- Olivine rich nephelinitic volcanic centres
- Mixed nephelinitic volcanic centres
- Olivine poor nephelinitic volcanic centres
- Zone of overlap of magmatic provinces
- Exposed Archaean basement

**Figure 1.5:** Tectonic settings of the Kenya Rift. Rift segments and faults shown in relation to principal late Proterozoic structural elements in the Precambrian basement (after Smith and Mosley, 1993). Inset shows distribution of nephelinitic and carbonatitic volcanism along the margin of the Nyanza Craton (from Smith (1994), modified from Le Bas (1987)). Smith (1994) suggests that the overlapping area of magmatic provinces (where there are both: olivine-poor nephelinitic and carbonatitic volcanism, and olivine-rich nephelinitic volcanism) marks the true position of the suture between the Precambrian units.
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Introduction

**Rift volcanics and tectonics**

The Kenya Rift was initiated on the western side of Lake Turkana in the Late Oligocene - Early Miocene (e.g. Baker and Wohlenberg, 1971; Morley, 1989), when extensive volcanism in the Lotikipi Plain (located across the Uganda border in northern Kenya) preceded the rifting. The observation that volcanism preceded faulting indicates that the melts were generated in response to a mantle thermal anomaly, rather than decompressional melting in response to crustal extension (White and McKenzie, 1989). The 37-25 Ma volcanism in the northern Kenya Rift may not represent Kenya Rift volcanism, and it may related to a thermal source originating from beneath the Afar Dome in Ethiopia (Morley, 1994; Mechie et al., 1997). This source drove the early volcanism in the north, and may be responsible for 25-30 km out of the total 40 km of surface extension proposed for this region (Hendrie et al., 1994).

In contrast, the central and southern parts of the rift are not related to this early phase of rift development. Smith (1994) reviewed the geomorphic, stratigraphic, and fission-track evidence concerning uplift of the Kenya Dome, and concluded that an uplift of <1 km occurred prior to volcanism. Geologic evidence constrains the timing to be post-Cretaceous in age. A pre-rift depression formed in the early Miocene but clear development of half grabens did not occur until about 12 Ma. From the Pleistocene to Recent, there has been a tendency for the zone of faulting and volcanism to migrate towards the centre of the rift valley. The main phase of tectonism was Plio-Pleistocene (<5 Ma) in age.

To the south, the age of the rift initiation becomes younger with major faulting beginning in the central part of the rift around or more likely prior to 16 Ma; and the surface volcanism (Middle Miocene) predates rifting (rifting followed at about 10 Ma) (Morley, 1994). Thus there is not a smooth propagation of volcanism ahead of rifting to the south with time (Morley, 1994). The volcanism in the south may probably be associated with the ascent of a separate thermal anomaly from beneath the Archaean Nyanza Craton eastwards to the site of the present day rift (Mechie et al., 1994a). In southern Kenya, the Upper Miocene-Recent extension along surface faults is about 10 km (Mechie et al., 1997). Beginning in the Quaternary, a series of young volcanoes developed along the rift valley axis, but there are also impressive off-axis volcanoes (Mounts Elgon, Kenya and Kilimanjaro, and the Chyulu Hills). The NW-SE striking Chyulu Hills are located directly on a major shear zone of the Mozambique Belt basement (see Figure 1.5) (Smith and Mosley, 1993). This shear zone can be regarded as a preexisting structure that could have given way to the magmas during the final part of their ascent. The volcanism took place during the Late Pleistocene and Holocene (<1.6 Ma) and the youngest flows are of historical age (Novak et al., 1997a; and references therein).
1.3.3 Previous geophysical investigation

Geophysical data acquired in Kenya during the last 30 years have helped to answer questions concerning the development and the structure of the rift, and gain some insight into the forces driving continental rifting.

Heat flow (e.g. Morgan, 1982; Nyblade and Pollack, 1993; Wheildon et al., 1994) and seismicity (e.g. Maguire et al., 1988; Tongue et al., 1992; Tongue et al., 1994) measurements demonstrate that the Kenya Rift is an active feature. Gravity and seismic data (e.g. Birt, 1996; and references therein) are consistent with an upwelling of the asthenosphere beneath the rift axis. Teleseismic results (e.g. Achauer and the KRISP teleseismic working group, 1994) suggest a degree of melting, and magnetotelluric and geomagnetic deep sounding data (e.g. Rooney and Hutton, 1977; Beamish, 1977) indicate anomalously high conductivity in the upper mantle possibly suggesting partial melting. Swain et al., (1994) reviewed the geophysical work prior to 1989. Some of the most recent geophysical work from the Kenya Rift can be found in Prodehl et al. (1994a) and Fuchs et al. (1997). Geological, geochemical and petrological perspectives are discussed by Latin et al., (1993), Macdonald et al., (1994), Hay et al., (1995a), Hay et al., (1995b), Wendlandt et al., (1995). Only some of the most important results will be briefly outlined here.

There have been many gravity studies within the Kenya Rift (e.g. Khan and Mansfield, 1971; Swain and Khan, 1977; see also Swain et al., 1994;), which provide density information for the rift sediments and volcanics as well as Proterozoic and Archaean basement. Figure 1.6 shows the Bouguer anomaly map for the whole of Kenya.

The regional field shows that the Kenya Dome and the East African Plateau are characterized by negative Bouguer anomaly values, suggesting a low density body in the base of the lithosphere, rising into the lower crust beneath the rift (Searle, 1970; Khan and Mansfield, 1971). Low density rift volcanics at the surface are added to account for the negative anomaly. The axial gravity high was modelled as a dense intrusion, 16-20 km wide, of mantle material into the upper crust coming to within 2-3 km from the surface (Searle, 1970; Baker and Wohlenberg, 1971). A recent model by Swain (1992) suggests that the central trough of the rift is underlain by basement gneiss invaded by a swarm of basic dykes, rather than a single massive intrusion. Nyblade and Pollack (1992) emphasizing the importance of the boundary between the Archaean Craton and the Mozambique Belt proposed that the Bouguer anomaly over the Kenya Rift and surrounding region may be a combination of a "suture" anomaly and a "rift" anomaly. The "rift" anomaly derives from crustal structures and a low density zone about 100 km wide in the mantle lithosphere beneath the rift valley. The "suture" anomaly would arise from a low density crustal root along the boundary between the Archaean Nyanza Craton and
the Mozambique Belt and higher density crust in the mobile belt (producing a negative anomaly over the boundary and a positive anomaly over the younger tectonic block). Their model places the suture between the crustal units in southern Kenya about 95 km west of the rift.

**Figure 1.6:** Bouguer anomaly map of Kenya (after Birt, 1996). The low gravity values are associated with the Kenya Dome and the East African Plateau. Contour interval is 10 mGal. Gravity station locations are marked by black dots.

Data from recent gravity studies (Hay et al., 1995b; Simiyu and Keller, 1997; Tesha et al., 1997) across the whole East African Rift System demonstrate lithospheric thinning beneath the rifts and suggest the existence of a deep mantle anomaly centred beneath the East African Plateau and the Nyanza Craton. The mantle anomaly is widely believed to be an upwelling mantle plume around the sides of the Nyanza Craton, and the diameter of its head is about 600 km (Simiyu and Keller, 1997). Low-density material is suggested to rise up beneath the
plateau and to spill around the sides of the cratonic lithosphere to shallow depths. At the equator the low-density upwelling material reaches the base of the crust in the Kenya Dome, but is deeper beneath the western branch. South of the Kenya Dome, the low Bouguer anomaly decreases in amplitude and the upwelling material is considered to be contained at deeper mantle levels. In order to explain the Bouguer anomalies derived from gravity observations in the southern seismic profile of KRISP 95, Birt (1996) proposed a low-density, westward dipping zone in the upper mantle to the west of the rift, consistent with the presence of an upwelling mantle plume, similar to the model of Hay et al. (1995b) and Simiyu and Keller (1997).

Seismic refraction - wide angle reflection experiments (Figure 1.7) carried out between 1985 and 1994 in the Kenya Rift by the Kenya Rift International Seismic Project (KRISP '85, KRISP '90 and KRISP '94) show major crustal thickness variations along and across the rift (Figure 1.8) (Mechie et al., 1997). Along the rift axis crustal thickness varies from 35 km in the south beneath and to the south of the Kenya Dome to 20 km in the north beneath the Turkana region (Prodehl et al., 1994b). This variation is accomplished to a large extent by the thinning of the 6.8 km/s basal crustal layer. This may be a normal feature of Proterozoic crust (Durrhein and Mooney, 1991), or may be due to underplating caused by the rifting process (Mechie et al., 1994a). The decrease of crustal thickness northwards from the Kenya Dome can be correlated with changes in surface topography (northwards decrease), rift width (northwards increase), surface estimates of extension (5-10 km in the south and 35-40 km in the north) and Bouguer gravity (the regional northwards increase of which can be explained entirely by the change in crustal thickness; Mechie et al., 1997). From the differences in the thickness of the crust along the rift it is apparent that rifting in the south is not as advanced as in the central or northern parts of the rift.

The upper mantle velocity ($P_n$) is reduced from normal value of 8.1 km/s beneath the rift flanks to 7.5-7.7 km/s beneath the thin crust of the rift axis. This is considered to be due to the presence of a small degree (3-5%) of partial melt (Mechie et al., 1994b). The profile in the southern Kenya at the latitude of Lake Magadi also shows a low $P_n$ velocity of 7.8 km/s confined to beneath the surface expression of the rift (Birt, 1996). However, crustal thinning beneath the rift along this profile is much subdued and to the west towards Lake Victoria the crust remains thin with a total thickness of around 34 km. In contrast, east of the rift, in the vicinity of the Chyulu Hills volcanic field some of the thickest crust (40-44 km) beneath Kenya is revealed over a distance of about 250 km (Novak et al., 1997a). Further south-east the crust thins to 20-25 km towards the Indian Ocean. Beneath the Chyulu Hills the $P_n$ velocity is around 7.9 km/s. A zone of reduced velocity detected in the base of the crust in this region suggests the presence of a zone of partial melt, but the existence of small magma chambers, rather than
a zone of asthenospheric upwelling is the preferred explanation in this region (Ritter and Kaspar, 1997).

**Figure 1.7:** Location map of the KRISP 90 and KRISP 94 seismic refraction-wide-angle reflection surveys (after Birt, 1996). In KRISP 90 three profiles were acquired across and along the rift, targeting the north and central part, an axial line from Lake Turkana in the north, to Lake Magadi in the south (Mechie et al., 1994a; Keller et al., 1994); a cross-rift line from Lake Victoria in the west to Chandlers Falls in the east (Maguire et al., 1994; Masotti et al., 1997); and an eastern flank line, from Chandlers Falls to Lake Turkana (Prodehl et al., 1994b). In KRISP 94 two profiles were acquired in the southern part of the rift, a cross-rift line running from Lake Victoria, crossing the rift at Lake Magadi and terminating north of the Chyulu Hills (Birt, 1996; Birt et al., 1997; Prodehl et al., 1997), and a profile ran from the eastern flank of the rift near Nairobi, to the Indian Ocean at Mombasa, crossing the Chyulu Hills (Novak et al., 1997a; Novak et al., 1997b).
The magnetotelluric profile of this study was selected to lie along the geophysical profile of KRISP 94 (Birt, 1996; Birt et al., 1997) and the most relevant points from the previous experiment will be noted here:

- Asymmetric rift basin, 4 km deep adjacent to the western bounding fault.
- Thin low velocity layer 150 km west of the rift (deformed Proterozoic basement).
- Higher velocities in the Proterozoic (east) than the Archaean (west), indicating differences in crustal structure between the two units.
- High velocity (7.0-7.1 km/s) lower crustal layer thickens to the east. Beneath the rift, its origin may be magmatic underplating that has minimised the crustal thinning by the addition of new material.
- Small amount of axial crustal thinning (1-2 km) superimposed on transition from thin Archaean (33 km) to thick Proterozoic crust (40 km).
• Poorly resolved upper mantle velocity reduction beneath rift axis, thought to be due to 3-5\% partial melt. Outside the rifted region, the upper mantle has higher velocities in the Archaean (8.3 km/s) than the Proterozoic (8.1-8.2 km/s).

The transition in basement velocity, crustal thickness and the pinching out of the mid-crustal layer are interpreted as differences between Archaean and Proterozoic crustal units (Birt 1996). In southern Kenya the rift seems to have developed above this ancient crustal weakness. The surface expression of the suture is obscured by a thin sliver of deformed Proterozoic basement that was thrust over the Archaean margin at the time of collision (Smith, 1994).

Teleseismic data from the KRISP '85 and '90 experiments show in the uppermost mantle a steep-sided low velocity body confined to the rift that indicates the present of partial melt (Green et al., 1991; Achauer and the KRISP teleseismic working group, 1994). Strong variations in velocity structure were found both parallel and perpendicular to the rift axis (Green and Meyer, 1992) and may be correlated with pre-existing basement structures (Achauer and the KRISP teleseismic working group, 1994). Slack et al., (1994) interpreted the same data set, and proposed a plume impacting on the base of the lithosphere causing it to erode, with the centre of the plume located near to the southern part of the Kenya Rift.

The Kenya Rift Valley is an area of concentrated seismicity of low magnitude. Micro-earthquake studies show good correlation of the seismicity with surface faults, hot springs and geysers, and more generally with geothermal and volcanic regions and the large scale structural features of the main rift (Maguire et al., 1988). The most active regions are Lake Bogoria, the E-W trending Nyanza Trough and north of the Kenya Dome towards Lake Turkana. Recent studies showed that the micro-earthquakes were confined to the brittle upper part of the crust, with the seismicity occurring between 12-16 km depth (Young et al., 1991; Tongue et al., 1994). Aseismic creep may occur at relatively shallow depths where strain is accommodated along low angle faults, or by the intrusion of magma bodies into the upper crust (Tongue et al., 1994).

Heat flow data have been presented by Morgan (1982), Nyblade and Pollack (1993), and Wheildon et al., (1994). Their data define generally low heat flow on the flanks of the Kenya Rift (40-60 mWm\(^2\)), with high, but variable heat flow on the rift floor (50-100 mWm\(^2\)) spatially associated with fluid circulation (hydrothermal activity within the rift), Quaternary volcanism and faulting. Wheildon et al., (1994) suggest that any deep thermal anomaly associated with the Kenya Rift has not yet been conducted to the surface, and the high heat flow values result from the emplacement of magmas in the crust and by hydrothermal convection. These interpretations support a model of a relatively young evolution of the asthenospheric anomaly.
beneath the rift zone, with the age of heating of the mantle at the Moho to have occurred recently, within the last 10 Ma (Wheildon et al., 1994). On a regional scale the Archaean Craton has lower heat flow values than the Proterozoic Belt, with values increasing away from the centre of the craton (Nyblade and Pollack, 1993).

Whilst the present study was the first magnetotelluric venture undertaken in Kenya under the auspices of KRISP, earlier Magnetotelluric and Geomagnetic investigations have been performed (Figure 1.9).

Banks and Ottey (1973) obtained geomagnetic depth sounding records from six stations along a 300 km near-equatorial profile. Their model includes two zones of anomalous conductivity, a low resistivity (10 $\Omega$ m) zone at 20 km depth beneath the rift axis, and another of more moderate conductivity (200 $\Omega$ m) 100 km east of the rift at a depth of about 50 km. However, the 10 $\Omega$ m conductor beneath the rift floor was poorly constrained, such that a 5 km thick, 5 $\Omega$ m near surface conductor could also satisfy the data. Beamish (1977) and Banks and Beamish (1979) made use of a more extensive data set from 22 sites to support the idea of conductivity anomaly beneath the rift, and confirmed the existence of enhanced conductivities to the east of the rift, modelling low resistivity zones in both the crust and mantle. These regions of high conductivity were interpreted as zones of partial melting in the upper crust and mantle coinciding with regions of Quaternary volcanism. To the west of the rift, no anomalies were detected.

Rooney and Hutton (1977) acquired 12 magnetotelluric soundings along a similar near-equatorial profile, but the resolution of the data was poor, and could only define a low resistivity body (of the order 2-20 $\Omega$ m) beneath the rift axis extending to depths of at least 40 km below the rift floor. They interpreted the low resistivities as result of partial melting caused by enhanced temperatures under the rift, and they estimate that less than 5% melt in a mantle composed predominantly of olivine would be sufficient to produce the observed resistivity values.

In the central part of the rift valley, in the geothermal plant of Olkaria, more MT measurements have been made (Galanopoulos, 1989; Hutton et al., 1989) and still continue (Onacha, pers. commun., 1995), for the Kenya Power and Lighting Company investigating the resources of the geothermal field.
A model of a plume upwelling beneath the thinned lithosphere in the Gregory Rift is, also, supported by geochemical studies, based on the volume and composition of basaltic melt generated beneath the rift. Latin et al., (1993) inverted rare earth element concentrations of the
most magnesian basalts to study the depth range and degree of partial melting beneath the central rift and the rift flanks. For a mantle temperature of about 1500 °C the top of the melting region predicted from the inversions is at 70 km beneath the rift axis and 80 km beneath the rift flanks. Information on degrees of partial melting suggest that the amount of melt has varied from <1% to a maximum of 3% (Latin et al., 1993), but the estimates are highly model-dependent (Macdonald, 1994). To explain the estimated basaltic magma (924,000 km³) generated beneath the rift zone in the past 30 Ma Latin et al., (1993) infer that the asthenospheric mantle must continually upwell through the melting region (extending from 70 to 150 km in depth) with a vertical velocity of between 40 and 140 mm/yr. Further geochemical studies (high-pressure / high-temperature experiments on phonolites; Hay et al., 1995a) suggested that alkali basaltic magmas may have injected and/or underplated the lower crust in central Kenya prior to the rift-related basaltic volcanism at the surface, between 23-14 Ma. Hay et al. (1995b) integrated geophysical data (seismic refraction and gravity) and petrological results to model an extensive lens of anomalously high density / high velocity material in the lower crust beneath the rift axis representing material of basaltic composition; the model supports the idea that the subrift crust was extensively modified by underplating and/or basaltic intrusion and subsequent partial melting. Later the underplating material was the source for the extensive flood phonolites which erupted in southern Kenya between 14 and 11 Ma (Hay and Wendlandt, 1995; Hay et al., 1995a).

Detailed interpretations, evolutionary and present-day structural summary derived from the KRISP seismic refraction-wide-angle reflection experiments, teleseismic studies, petrology and surface geology can be found in Keller et al., (1995) and Mechie et al., (1997). The active uprising of anomalously hot mantle material (a mantle plume) combined with pre-existing lithospheric structures is believed to have generated and controlled the evolution of the rift. Smith (1994) concluded that a small mantle plume rose through the lithosphere to the base of crust. He suggested that initially the axis of the plume was located beneath the central part of the (Gregory) rift, where the E-W trending Nyanza Rift forms a third arm to a tri-radial junction. Summerfield (1996) and Burke (1996) have presented maps with hypothesized mantle plumes and hot spot tracks for the African Plate indicating Lake Victoria and Afar regions as two of them. Recently, Ebinger and Sleep (1998) suggested a single large plume impinging beneath the Ethiopian plateau and affecting a much broader region, extending down the East African Rift, offshore and south to the Comoros islands near Madagascar, west to the Darfur uplift, and possibly much further to the Adamawa plateau and the Atlantic coast (their Figures 1 and 2). Lateral flow and ponding of plume material was facilitated by pre-existing zones of lithospheric thinning. George et al. (1998) argue that there are actually two plumes in the area of East Africa, the Afar plume in the north and an older plume now located under Lake Victoria that has
migrated southwards and caused eruptions on the southern Ethiopian plateau as early as 45 million years ago.

1.4 Aims of this study

Under the auspices of the KRISP project, a series of both active and passive seismic experiments have investigated the seismicity and deep structure of the rift. The results of these studies have generated feasible models for the structure of both crust and upper mantle along and perpendicular to the rift. The advances in geophysical knowledge have been matched by a greater understanding of the evolution of the rift in relation to tectonic settings. In particular, the importance of structures in ancient basement in controlling the evolution of rift faulting and magmatism is more widely appreciated (Smith and Mosley, 1993). In parallel with this progress, great strides have been made in the acquisition, processing and interpretation of electromagnetic measurements. It was therefore deemed worthwhile to carry out a detailed magnetotelluric study of the Kenya Rift Valley, as it probably can provide some key answers to the deep geology of this region.

The south of Kenya is a transition region, to the north the rift faults and extensional structures are well developed, in the southern part the well-defined narrow graben of central Kenya is replaced by a wider asymmetric rift valley and the influence of the uplifted area of the Kenya Dome becomes more distant. It is now recognised that the oldest volcanism and structures associated with the rifting event, in the Kenya Rift, young southwards, and the rift appears to be attempting to cut into the Archaean Nyanza Craton into northern Tanzania. In order to examine this region of the rift, long-range seismic refraction and gravity measurements were performed in 1994 as part of the Kenya Rift International Seismic Project. The same area was selected for electromagnetic studies the following year. In early 1995, high-quality broad band magnetotelluric data were collected at 19 sites, that lie along the pre-existing seismic and gravity profiles. Transient electromagnetic soundings were also performed in order to facilitate the removal of static shift effects from the MT data (Stenberg et al., 1988; Meju, 1996) and recover the near surface structure. The line was about 500 km long, starting from Lake Victoria in the West, crossing the rift at Lake Magadi, crossing the Chyulu Hills and terminating on the southern flanks of the Taita Hills in south-east Kenya. The primary objectives of this study were:

- The acquisition of accurate and unbiased MT response functions
- The determination of the optimum resistivity distribution for the southern Kenya Rift Valley
The analysis and interpretation of the results for both near surface and deeper structure and the correlation of any observed resistivity structures with previous geophysical and geological data from the area.

The previous electromagnetic studies in the Kenya Rift suffered from low spatial density (about 50 km separation), static shift problems, and narrow period range (10-2000 sec), absence of high frequency data which restricted the resolution of the shallow structure and lack of low frequencies which limited the penetration depth when the surface layers are conductive, especially beneath the rift. Given these limitations, the picture conveyed by the past studies was simple and uncertain. Rooney and Hutton (1977) carried out limited 1-D modelling of impedances for the rift stations and tested the validity of 2-D models in a more quantitative way, but static shift problems and poor quality data seriously questioned the results. The present study is aimed to produce a more quantitative and realistic geoelectric model for the Kenya Rift Valley. The higher density station spacing (~20 km; although still widely spaced compared with the best modern practice of station separation of less than 5 km), the broad period range of the MT soundings (10^2 to 10^4 sec; providing information from shallow, ~200 m, to deeper layers, 40 km), and the use of transient electromagnetic technique (to recover the shallow resistivity distribution (<200 m) and remedy the "static shift" problems from the MT impedance tensor) are expected to help to accomplish the targeted aims. Additionally, the use of improved instrumentation and modern processing and modelling techniques will provide better resolution and more reliable results.

The effect of localized surface bodies of low or high resistivity material which shift the whole apparent resistivity curve up or down, known as "static shift" effect, is a major problem for MT measurements. The use of TEM soundings to recover the true level of the MT apparent resistivity is a new approach towards the solution of the problem. This survey is one of the first regional studies that applied TEM soundings to every MT location (see also Meju et al., 1999), therefore, it was an aim of the project to evaluate the use of the method in the complex environment of Kenya Rift.

Comparison with the results of the previous seismic-gravity profiles would assist in obtaining an improved image of the subsurface structure of the area. The profile crosses the boundary between the Archaean and Proterozoic crustal units, and would therefore allow an investigation of the suture zones, its relation with the rift and the influence on the resistivity distribution of the area, and may help define the position of the craton margin. Since magmatism is an important factor in the rift mechanism, and partial melt has been shown in laboratory studies to increase the conductivity of rock by orders of magnitude (e.g. Waff, 1974; Sato and Ida, 1984) magnetotelluric measurements on the rift floor could provide information about the upwelling thermal anomaly that has been detected and the magmatic underplating or
intrusion in the lower crust (Mechie et al., 1997). Finally, the proposed model for the region of the Chyulu Hills can be examined. The model endorsed the idea that the volcanism in the Chyulu Hills was started by only one or a small number of melt blobs originating from a larger plume below the Kenya Rift and not from a major upwelling of the asthenosphere in the region of the hills. It was proposed by Karson and Curtis (1989) and supported by Macdonald (1994) and Novak et al., (1997a).

1.5 Thesis outline

This introduction has presented an overview of the geoelectric structure of the continental crust, the process of continental rifting, and a review from previous studies of the Kenya Rift. It draws on both geological and geophysical observations to provide some background information concerning the known evolution, structure and understanding of the region. It also tries to lend additional support to the interpretation placed on geophysical models from this study. The rest of this thesis is as follows.

Chapter 2 starts by describing the source of the EM field variations and continues by deriving some of the fundamental MT relationships from basic EM wave theory. Other relevant topics, including the processing methods for the data and short presentation of the current distortion correction techniques, are also reviewed. Since TEM observations were also collected, some of the fundamental applied theory is developed. Finally, this chapter proceeds to give a short review of 1-D MT response function modelling methods and their limitations, and concludes presenting the 2-D modelling technique emphasizing the theory behind the algorithm that was used in this study.

The equipment, the field methods and strategy employed for the acquisition of MT and TEM data are given in Chapter 3. This includes a discussion of MT field techniques site selection criteria and a brief discussion of the KRISP 95 MT experiment and the sites which were occupied in southern Kenya.

The processing of the field records and their conversion to a form convenient for interpretation follows in Chapter 4. The Groom and Bailey (1989) and Bahr (1991) analysis of the impedance tensor and the reasons for the selection of the regional electric strike direction, are also discussed. The remainder of this chapter is concerned with the correction of the static shift effects from the MT apparent resistivity using the TEM data.

One dimensional and two dimensional models for the profiles are presented in Chapter 5, with discussion about their uniqueness and resolution. The interpretation of the 2-D models,
their integration and comparison with previous geophysical and geological data from the area, and a robust model for the present structure of the Kenya Rift are discussed in Chapter 6.

Chapter 7 summarizes some of the more significant conclusions and results of this study before providing some recommendations for further research work.
Chapter 2

The Magnetotelluric and Transient Electromagnetic Methods

Two geophysical measuring tools were used in this study to observe the electrical conductivity changes with depth in the earth. This chapter reviews the theoretical and practical background to the Magnetotelluric and Transient electromagnetic methods. Then it briefly recounts the theoretical background to the forward and inverse 1-D and 2-D modelling schemes used. Finally, the chapter proceeds to describe the overall properties of the measured earth response functions that influenced the selection of particular modelling methodology in this study.

2.1 Magnetotelluric fields: source mechanism

The magnetotelluric field can be defined generally as the time-varying portion of the earth's magnetic field which induces current flow in the earth. These fluctuations may occur over periods ranging from milliseconds to centuries. There are various types, sources and frequencies of the variations. Figure 2.1 is an approximation and schematic representation of origins and frequencies. The *Secular variations* have internal origin, thought to be caused by motion in the electrically conducting fluid core of the Earth. The *Diurnal (daily) variations* are thermal effects of the sun and attraction effects of the moon, which cause the ionosphere to oscillate. *Micropulsations* are variations of exospheric origin, and these are the most significant for exploration. *Vibrations* is a general name for variations whose frequencies are higher than 5 Hz, and the lightning is the major source (Yungul, 1996).

<table>
<thead>
<tr>
<th>Period (time per cycle)</th>
<th>Frequency (cycles per second = Hz)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 year (internal)</td>
<td>1 year</td>
</tr>
<tr>
<td>6 h (ionospheric)</td>
<td>1 Hz (external)</td>
</tr>
<tr>
<td>600 sec (exospheric)</td>
<td>5 Hz (atmospheric)</td>
</tr>
<tr>
<td></td>
<td>30 MHz (stellar)</td>
</tr>
</tbody>
</table>

![Figure 2.1: Approximate and schematic frequencies and origins of the natural electromagnetic fields (after Yungul, 1982)](image-url)
Within the period range which is used in MT surveys ($10^2 - 10^4$ sec), noise is contributed primarily by three types of sources: atmospheric electrical discharges, industrial noise from power distribution systems and micropulsations in the magnetic field (Keller & Frischknecht, 1966). Magnetotelluric signals must be generated at a sufficient distance from the survey location so they can be considered as being virtually vertically incident "plane waves". The source of most MT signals are therefore external to the earth. The assumption of a vertically incident plane wave, and the fact that the resistivity contrast between the air and the earth is large enough, means that waves at all angles of incidence are transmitted into the earth nearly vertically.

![Figure 2.2: Typical spectrum of amplitudes of electromagnetic noise in the extremely low frequency (ELF) range (after Keller and Frischknecht, 1966).](image)

Magnetotelluric source fields can be classified as originating either from lightning or from solar activity. A typical amplitude vs. frequency spectrum for magnetic variations is shown in Figure 2.2. Between 10 and 0.1 Hz there is a minimum in the amplitude of the magnetic variations and a maximum in the natural noise (due to microseismic activity, mainly wind noise but also ocean effects) and cultural noise spectra, with resultant low signal-to-noise ratios (Jones, 1992). Therefore this frequency (1 Hz) distinguishes two kinds of activity with relative sources above 1 Hz and below 1 Hz. Within this so called "dead band", to get good quality MT data can take a great deal of patience because of the persistently low level of the signals and the normally high level of noise (Boehl et al., 1977).

**Sources above 1 Hz**

At frequencies above one cycle per second, the magnetotelluric field is generated almost entirely from meteorological activity. The meteorological component consists of fields from lightning associated with thunderstorms. The signals created from lightning strikes are known as "sferics". The "sferics" attain their peaks in the early afternoon, local time. They propagate
around the world trapped in the waveguide formed between the ionosphere and the earth’s surface. In the day time the waveguide width is 60 km and increases to 90 km during the night time. The waveguide absorbs energy at certain frequencies (e.g. >1 kHz) and at others - the Schumann cavity resonance frequencies (8, 14, 20, 26 and 32 Hz)- the energy is enhanced (Jiracek et al., 1995).

There are three major storm centers in the equatorial regions which have an average of 100 storm days per year, with small areas within these centers which average more than 200 stormy days per year. These centers are located in Brazil, central Africa and Malaysia, distributed in such a way that during any hour of the day, there is probably a storm in progress in one of these centers.

**Sources below 1 Hz**

Sferics contribute very little to the naturally existing electromagnetic fields at frequencies less than 1 Hz. At these low frequencies, the natural electromagnetic field consists of periodic or nearly periodic variations in the earth’s magnetic field. These variations cover the range in frequencies from 1 cycle per second to one cycle per eleven years, though variations with periods longer than a day are rarely used to determine earth resistivity. These variations represent a variety of phenomena, but the complex interaction between the solar wind (charged particles ejected from the sun) and the earth’s atmosphere and magnetosphere is the main cause of the natural electromagnetic fields below 1 Hz.

When ionized particles moving outward from the sun encounter the earth’s magnetic field, they cause rapid distortions to the earth’s magnetosphere. The ionized particles are protons and electrons, and when they encounter the terrestrial magnetic field, they are deflected in opposite directions and set up current systems with their own secondary magnetic field which cancels the effect of the main field at a boundary called the “magnetopause” which is the outer boundary of the earth’s magnetosphere. This boundary is not symmetrical but is more compressed in the direction of the solar wind and moves erratically with variations in the solar wind.

When sunspots occur the solar wind is enhanced and magnetic storms are observed. The solar wind is also enhanced by smaller masses of plasma which activate substorm phenomena. At altitudes of 80-160 km the earth’s atmosphere is strongly ionized. The motion of charged particles in this zone which is known as “ionosphere” is associated with the induction of the current. The interaction of magnetic and inertial forces gives rise to “Magnetohydrodynamic” waves. Magnetic effects due to these waves are the micropulsations, which are the main source of the natural electromagnetic waves below 1 Hz.
Pulsations are classified according to their continuity and period (Parkinson, 1983). Those that continue with either steady or regularly fluctuating amplitude are called "Pc" (pulsations continuous). There is also a second type of pulsation that resembles a damped oscillation, each group containing between 5 and 20 pulsation cycles. These are known as "Pi" (pulsation irregular). The amplitude of the Pc is dependent upon the time of the day but they tend to diminish during the night when Pi become more dominant.

2.2 Maxwell's equations

The theory behind induction studies is based upon Maxwell's equations. Electric and magnetic fields are connected by Maxwell's field equations:

\[ \nabla \times E = -\frac{\partial B}{\partial t} \] (2.1)

\[ \nabla \times H = J + \frac{\partial D}{\partial t} \] (2.2)

\[ \nabla \cdot B = 0 \] (2.3)

\[ \nabla \cdot D = \rho^* \] (2.4)

where \( E \) is the electric field vector (V/m), \( H \) is the magnetic field vector (A/m), \( B \) is the induction field vector (W/m² or T), \( D \) is the displacement vector (C/m²), \( J \) is the current density and \( \rho^* \) is the charge density (C/m³). Since any charge distribution decays very rapidly within a conductive material, it can be assumed that no charges accumulate and consequently equation 2.4 can be simplified by setting \( \rho^* \) to be zero.

To derive relationships between Maxwell's equations, the experimentally determined constitutive equations must be used:

\[ D = \varepsilon \cdot E \] (2.5)

\[ B = \mu \cdot H \] (2.6)

\[ J = \sigma \cdot E \] (2.7)

\[ \sigma = 1/\rho \] (2.8)

where \( \varepsilon \) is the electrical permittivity (\( \varepsilon_0 = 8.9 \times 10^{-12} \text{F/m} \) in free space), \( \mu \) is the magnetic permeability (\( \mu_0 = 4\pi \times 10^{-7} \text{H/m} \) in free space), \( \sigma \) and \( \rho \) are the electrical conductivity and resistivity.
of the medium in S.m$^{-1}$ and ohm.m$^{-1}$ respectively. The variation of $\varepsilon$ and $\mu$ is small compared with the vast range of resistivities within the earth, therefore by assuming that all media are linear, isotropic, homogeneous, and possesses electrical properties which are independent of time, temperature or pressure, we may simplify the analysis and the dielectric permittivity and the magnetic permeability is assumed to be that of free space, $\varepsilon = \varepsilon_0$ and $\mu = \mu_0$. With these hypothesis the constitutive equations allow a simplification of Maxwell's equations by eliminating the $B$, $J$ and $D$ terms (Ward and Hohmann, 1987).

### 2.2.1 The magnetotelluric diffusion equation

Taking the curl of equation (2.1) and then substituting equation (2.6) into the right hand side:

$$\nabla \times (\nabla \times E) = -\partial / \partial t (\nabla \times B) = -\mu \cdot \partial / \partial t (\nabla \times H)$$

(2.9)

Expanding the left hand side of equation (2.9) and then substituting equation (2.2) and using equation (2.7) to eliminate $H$:

$$\nabla \cdot (\nabla \cdot E) - \nabla^2 E = -\mu \cdot \sigma \cdot \partial E / \partial t$$

(2.10)

Since equation (2.4) states that $\nabla \cdot E = (\nabla \cdot D / \varepsilon_0) = 0$, then the first term in equation (2.10) disappears. As MT measurements consist of time domain observations of the EM field variations ($E(t)$ and $H(t)$) a Fourier transformation is applied to determine their amplitude as a function of angular frequency ($\omega = 2\pi f$). Assuming the primary field has a harmonic time variation of $e^{-i\omega t} (E(t) = E_0 e^{-i\omega t})$, taking the derivative of $E(t)$ in equation (2.10) gives:

$$\nabla^2 E = -(\mu \cdot \varepsilon \cdot \omega^2 - i\omega \mu \cdot \sigma) \cdot E = k^2 \cdot E$$

(2.11)

in which $k = (\mu \varepsilon \omega^2 - i\omega \mu \sigma)^{1/2}$ can formally be regarded as the complex wave number of a medium. A similar equation can also be derived for the propagation of the magnetic field in a homogeneous medium:

$$\nabla^2 H = -(\mu \cdot \varepsilon \cdot \omega^2 - i\omega \mu \cdot \sigma) \cdot H = k^2 \cdot H$$

(2.12)

Equations (2.11) and (2.12) are the wave equations in the frequency domain, and they are known as Helmholtz equations in $E$ and $H$. At the frequencies used by the magnetotelluric method, $\mu \varepsilon \omega^2 \ll \mu \sigma \omega$ for earth materials; displacement currents are much smaller than conduction currents. Thus equations (2.11) and (2.12) may be rewritten as:

$$\nabla^2 E = -i\omega \mu \cdot \sigma \cdot E = k^2 \cdot E$$

(2.13)
\[ \nabla^2 \mathbf{H} = -i \omega \mu \sigma \mathbf{H} = k^2 \mathbf{H} \]  

(2.14)

under this circumstance, the propagation constant is given by: 
\[ k = (-i \omega \mu \sigma)^{1/2}. \]

In general form equations (2.13) and (2.14) can be written as:

\[ \nabla^2 \mathbf{F} + k^2 \mathbf{F} = 0 \]  

(2.15)

\( \mathbf{F} \) denotes either the magnetic field \( \mathbf{H} \) or the electrical field \( \mathbf{E} \). Equations (2.13) and (2.14) represent diffusion equations, therefore the electromagnetic induction in the earth is a diffusion process.

To simplify the solution of the diffusion equation, more assumptions have to be made. As was mention before (Section 2.1) the basic assumption is the existence of plane electromagnetic waves incident on a plane earth surface (Cagniard, 1953). In the early days, some researchers (Wait, 1954; Price, 1962) argued about this plane wave assumption of the source field. Madden and Nelson (1964) came to the defence of the Cagniard model by showing that when the source's horizontal wavelength is much greater than the skin depth (skin depth will be defined later), as it is for the earth, the model remains valid. Dmitriev and Berdichevsky (1979) further proved that the horizontal magnetic field components need to be uniform, but can vary linearly over a layered earth.

### 2.3 Induction in 1-dimensional structures

A one dimensional structure is one in which conductivity is a function of depth only, i.e. \( \sigma = \sigma(z) \). This is the simplest model structure consisting of a homogeneous and isotropic half space of uniform conductivity; \( \sigma \) occupies the half space \( z \geq 0 \), and \( z < 0 \) is free space.

In a Cartesian coordinate system (\( x: \) north, \( y: \) east, \( z: \) vertically downward) it is possible to select that only one component of the electric field is non-zero. Unless there are lateral changes in resistivity, the current sheets induced in the earth by a varying magnetic field must be parallel to the surface of the earth, since no appreciable amount of current can be forced to flow across the air-earth boundary. This restriction permits only vertically travelling waves. Orientating the coordinate system so that the \( x \) direction is along the current flow, there is no electric field in the \( y \) or \( z \) directions (\( E_y = E_z = 0 \)). With this choice of coordinates, equation (2.13) simplifies to:

\[ \frac{\partial^2 E_x}{\partial z^2} = -i \omega \mu \sigma E_x \]  

(2.16)
This equation is a second order linear equation; the elementary solution of this equation is:

\[ E_x = Ae^{kh} + Be^{-kh} \]  \hspace{1cm} (2.17)

The magnetotelluric field is known to be quasi-periodic in the frequency range commonly used in electrical prospecting and for a harmonically time-varying field there is a general solution:

\[ E_x = Ae^{(i\omega \cdot k)z} + Be^{(-i\omega \cdot k)z} \]  \hspace{1cm} (2.18)

where \( k \) is the wave propagation constant, and \( A \) and \( B \) are arbitrary constants which are evaluated by applying the boundary conditions, i.e., the tangential component of \( E \) and \( H \) is continuous across the boundary. Since in the earth half space \((z>0)\), the intensity of \( E \) decreases with increasing \( z \), due to the transformation of electromagnetic energy to heat, it is important to impose the boundary condition that \( A = 0 \). Taking the derivative of equation (2.18) with respect to \( z \) \( \left( \frac{dE_x}{dz} = -kBe^{-kh} \right) \), and equating this with equation (2.1) \( \left( \frac{dE_x}{dz} = -i\omega \mu H_y \right) \), at the earth's surface \((z=0)\):

\[ B = (i\omega \mu / \sigma)^{1/2} \cdot H_y \]  \hspace{1cm} (2.19)

This coefficient contains information about the conductivity distribution. The ratio of the electric field strength to magnetic field strength can be defined as the wave impedance of a medium. It can be shown that for plane wave propagation, this ratio depends on the electrical properties of the medium and on the frequency, and that the dimensionality of the ratio is expressed in ohms. In this particular problem, it is:

\[ \frac{E_x}{H_y} = Z = \sqrt{\frac{i\omega \mu}{\sigma}} \]  \hspace{1cm} (2.20)

where \( Z \) is the wave impedance of the homogeneous ground defined by the ratio of orthogonal components of \( E \) and \( H \). From the equation (2.20) an apparent resistivity may be defined, the so called Cagniard apparent resistivity (Cagniard, 1953):

\[ \rho = \frac{1}{\omega \cdot \mu} |Z|^2 \]  \hspace{1cm} (2.21)

with a constant 45° phase difference between the oscillations in the magnetic and electric field intensities. In more general cases, the phase relation between the two fields varies, and may be used to determine resistivity distributions, just as variations in the magnitude of the wave impedance. The phase difference between the \( E \) and \( H \) fields of \( Z \) is:
\[ \Phi = \arg \left( \frac{E_x}{H_y} \right) = \arg(Z) \]  
\hspace{1cm} (2.22)

The Cagniard impedance can be evaluated from any other orthogonal pair of horizontal electric and magnetic field observations. Employing more practical field units of mV.km\(^{-1}\) for the measurement of the electric field and nT for the magnetic field variations, equation (2.22) reduces to:

\[ \rho = 0.2 \ T|Z|^2 \]  
\hspace{1cm} (2.23)

where \( T \) is period in seconds (\( T = 1/f \)).

### 2.3.1 Skin depth \( \delta \)

The propagation constant can be separated into real and imaginary parts:

\[ k = (\omega \mu / 2 \rho)^{1/2} + i(\omega \mu / 2 \rho)^{1/2} \]  
\hspace{1cm} (2.24)

(indicating an exponential decay of amplitude along the travel path and an oscillatory nature of the wave amplitude), and can be used to define the skin depth, the depth at which the amplitude of the field has been attenuated by 1/e of its value at the surface of the medium. For a uniform earth it is given by:

\[ \delta = \frac{1}{\text{Re}(k)} = \sqrt{\frac{2}{\omega \mu \cdot \sigma}} \]  
\hspace{1cm} (2.25)

A useful approximation is given by:

\[ \delta = 500 \cdot \sqrt{\rho / f} \text{ meters} \]  
\hspace{1cm} (2.26)

This equation contains the important information that the skin depth of a magnetotelluric sounding depends on both the period of the signals and the conductivity of the medium. The longer the period measured or the smaller the conductivity of the medium, the deeper the electromagnetic fields penetrate. Therefore, the MT impedance is estimated at a broad range of frequencies in order to obtain a response function which can provide information concerning the conductivity distribution over a range of depths. Curves showing the relationship between wave number, resistivity and frequency are given in Figure 2.3.
2.3.2 Electromagnetic induction in an N-layered half space

In an n-layered earth, continuity conditions which must hold at each boundary permit us to express the wave impedance observed at the surface in terms of the wave impedances in each of the lower layers. Both the electric field and the magnetic field, must be continuous across the boundaries between layers. The wave impedance at the bottom of the first layer must equal the wave impedance at the top of the second layer, and so on. If we consider a model in which the earth is represented by n horizontal layers where the conductivities of the layers are \( \sigma_1, \sigma_2, \ldots, \sigma_n \) respectively and the thickness of the top n-1 layers are \( h_1, h_2, \ldots, h_{n-1} \), the MT impedance at the surface, can be evaluated by first computing the impedance for each individual layer in the stack. This can be done by solving the equation (2.16) for each individual layer and imposing continuity conditions for \( E_x \) and \( H_y \) across boundaries. The following relation can be obtained for the impedance at the surface (Ward et al., 1973; Kaufman and Keller, 1981):

\[
Z_n(\omega) = \frac{i \mu \omega}{k_i} \coth(k_i h_i + \coth^{-1}\left(\frac{1}{k_i} \coth(k_i h_i + \coth^{-1}\left(\ldots \coth^{-1}\left(\frac{k_{n-1}}{k_n} \coth\left(k_n h_{n-1} + \coth^{-1}\left(\frac{k_{n-1}}{k_n}\right)\right)\right)\ldots\right)\right))
\]

(2.27)

where \( k_i \) is the propagation constant for the i-th layer. Equation (2.27) is used in the 1-D MT forward problem to calculate the apparent resistivity and phase of a layered earth structure at any given frequency.
2.4 Induction in 2-dimensional structures

A homogeneous plane layered earth model is clearly inadequate for the interpretation of most induction data. The structure of many geological bodies such as dykes, faults and rift valleys, ideally, remains virtually unchanged in one direction (the strike direction), over large distances. If the earth's structure is constant in the strike direction for distances over a few skin depths, a 3-D feature can be reasonably well approximated by a two-dimensional model (Patra and Mallick, 1980).

To solve this problem it is necessary to make the same basic assumptions as for the 1-D case, but assuming further that there is no conductivity variation in the x direction (\(\partial \rho / \partial x = 0\)). Expanding Maxwell's equations into Cartesian component form will give rise to 6 equations. When the geoelectric structure is truly two dimensional, the EM fields de-couple into 2 distinct polarizations (Figure 2.4). The first of these, the TE-mode (transverse electric) describes the field components (\(H_y, H_z, E_x\)) observed when the currents are flowing along (parallel to) the structure, and the TM-mode (transverse magnetic) relates the field components (\(H_x, E_y, E_z\)) when the current are crossing (perpendicular to) the structure. The TE-mode is also often called E-polarization or E-parallel, and the TM-mode B-polarization or E-perpendicular. The two sets of equations are given below:

\[
\begin{align*}
\frac{\partial H_y}{\partial y} - \frac{\partial H_z}{\partial z} &= \sigma E_x, \\
\frac{\partial E_x}{\partial z} &= i\omega \mu H_z, \\
\frac{\partial E_y}{\partial y} &= -i\omega \mu H_z \quad (2.28.a)
\end{align*}
\]

\[
\begin{align*}
\frac{\partial E_y}{\partial y} - \frac{\partial E_z}{\partial z} &= i\omega \mu H_x, \\
\frac{\partial H_x}{\partial z} &= \sigma E_y, \\
\frac{\partial E_z}{\partial y} &= -\sigma E_z \quad (2.28.b)
\end{align*}
\]

Figure 2.4: Magnetotelluric polarization modes.
The most important characteristics of 2-D MT apparent resistivity sounding curves are realized by understanding the variations in TE and TM mode results on either side of a simple geoelectric contact. Figure 2.5 depicts such a contact with $\rho_1$, representing the more conductive quarter-space to the left of the resistive quarter-space $\rho_2$. The physics of the response of this 2-D model is profoundly different for the two polarizations.

![Diagram showing variation of TE and TM apparent resistivities (top) and phase (middle) curves versus distance for the fault model illustrated (bottom) at a single period (1 sec). The TE-mode responses (solid curves) varies smoothly across the structure, whereas the TM-mode (dashed curves) apparent resistivity curve is discontinuous, with values far outside the range of $\rho_1$ and $\rho_2$.](image)

All three EM field components in TE mode ($E_y, H_x, H_z$) are continuous across all boundaries so the fields vary smoothly across geologic contacts (solid lines). Therefore, the apparent resistivity, which is proportional to $E_y / H_x$ is smoothly varying across a contact. In TM polarization, $H_y$ and $E_z$ are continuous across a contact but $E_x$ is discontinuous (dashed lines). The discontinuity in $E_x$ follows from the continuity of the normal component of the current density $J$. The electric field discontinuity at a contact is due to electric charge accumulation on the contact, which is the underlying cause of the galvanic effect (Berdichevsky and Dmitriev, 1976). The galvanic effect produces the discontinuous behaviour of the $\rho_{TM}$ results. At locations sufficiently away from the contact, neither $\rho_{TE}$ or $\rho_{TM}$ responses sense
the contact. This horizontal distance is about one skin depth, so that the distance is greater at low frequencies than high frequencies (Vozoff, 1991). Two other major characteristics of the $\rho_{TE}$ and $\rho_{TM}$ are (Vozoff, 1991; Jiracek et al., 1995): (a) both curves converge to the true resistivities at the highest frequencies, and (b) $\rho_{TE}$ is above $\rho_{TM}$ on the conductive side of a contact, and vice versa on the resistive side.

The TE and TM modes offer a different kind of information about the resistivity structure. The TM impedance is more sensitive to near-surface resistive structures, while the TE impedance is more sensitive to deep conductive structures (Berdichevsky et al., 1998), the TM mode is more robust to plausible 3-D effects than the TE impedance (Wannamaker et al., 1989). A number of 2-D studies exist that used only the TM mode to interpret the investigated areas. The $\rho_{TM}$ and $\phi_{TM}$ are relatively insensitive to finite strike extent (Wannamaker et al., 1984), and if sampled adequately laterally, will return model resistivity estimates whose long-wavelength averages presumably are unbiased by shallow, smaller scale structures (Torres-Verdin and Bostick, 1992). This fact allows valid resistivity cross-sections to be derived from careful modelling of observed TM mode data across orientated geological terranes (Wannamaker and Hohmann, 1991). The approach has been a key component of several successful MT transect interpretations in recent years (e.g. Mackie et al., 1988; Wannamaker et al., 1989; Park et al., 1991; Wannamaker et al., 1997). Nevertheless, Banks et al., (1996) argue that models which fit the TM mode are not resolving large-scale structures evinced only by the TE mode. Berdichevski et al. (1998) proposed that the gaps left by one mode can be filled by the other mode, and the use of both modes can lead to more comprehensive and reliable information for the resistivity distribution of the earth's interior.

2.5 Earth response functions

Any function that can provide valuable information about the earth is an Earth response function (Rokityanski, 1982). Such a function characterises the conductivity structure of the earth and could include the impedance tensor, the apparent resistivity or the phase of the impedance.

2.5.1 The Impedance tensor

The relationships among the field components at a single site are systematically contained in the impedance. It is the quantity from which conductivity structure is interpreted. In general, $H_x$ has an associated $E_y$ and some $E_z$, both of which are proportional to $H_x$. Likewise, $H_y$ causes an
$E_x$ and some $E_y$, so that at each frequency we would expect a linear system to behave as (Cantwell, 1960; Rokityanski, 1961):

$$E_x = Z_{xx} H_x + Z_{xy} H_y$$
$$E_y = Z_{yx} H_y + Z_{yy} H_x$$

where each term is frequency dependent. This is commonly written:

$$\begin{pmatrix}
  E_x \\
  E_y 
\end{pmatrix} = \begin{pmatrix}
  Z_{xx} & Z_{xy} \\
  Z_{yx} & Z_{yy}
\end{pmatrix} \begin{pmatrix}
  H_x \\
  H_y
\end{pmatrix}$$

(2.29)

or

$$E(f) = Z(f) \cdot H(f)$$

(2.30)

where $Z_{xy}$, $Z_{yx}$ are the principal impedances and $Z_{xx}$, $Z_{yy}$ are the subsidiary (or additional) impedances due to contributions from parallel components of the fields.

In a simple 1-D environment the off-diagonal impedance elements relate the orthogonal components of the electromagnetic field, and their corresponding apparent resistivity and phase estimates are given by:

$$\rho_{xy} = 0.2 \left| \frac{Z_{xy}}{} \right|^2 / f \quad \& \quad \Phi_{xy} = \tan^{-1} \frac{\text{Im}(Z_{xy})}{\text{Re}(Z_{xy})}$$

(2.31)

$$\rho_{yx} = 0.2 \left| \frac{Z_{yx}}{} \right|^2 / f \quad \& \quad \Phi_{yx} = \tan^{-1} \frac{\text{Im}(Z_{yx})}{\text{Re}(Z_{yx})}$$

A similar relationship, with the impedance tensor, is also assumed to exist between the horizontal and vertical components of the magnetic field:

$$H_z(f) = \left( T_x(f) \cdot T_y(f) \right) \begin{bmatrix}
  H_x(f) \\
  H_y(f)
\end{bmatrix}$$

(2.32)

where $T_x(f)$ and $T_y(f)$ are known as the tipper elements.

### 2.5.2 Rotation of the impedance tensor

In a uniform or horizontally layered earth, $Z_{xx}$ and $Z_{yy}$ are zero, $Z_{yx} = -Z_{xy}$, and the equations (2.29) reduce to:
\[ E_x = Z_{x,y} H_y \]  
\[ E_y = Z_{y,x} H_x = -Z_{x,y} H_x \]  
(2.33)

In a 2-D case, if \( x \) or \( y \) axis is along strike then:

\[ Z_{xx} = Z_{yy} = 0 \quad \text{but} \quad Z_{xy} \neq -Z_{yx} \]  
(2.34)

If neither axis is along strike then:

\[ Z_{xx} = -Z_{yy} \neq 0 \]  
(2.35)

The strike direction is seldom known very precisely during an MT survey. Suppose our measurement axes \( x \) and \( y \), form an angle \( \theta \) with the true strike, we want to determine our field components in the (preferred) principal anisotropy axes \( x' \) and \( y' \). For a Cartesian rotation, when the new axes are rotated \( \theta \) degrees clockwise as in Figure 2.6, the transformed field components are:

\[ E' = R \cdot E \quad \text{and} \quad H' = R \cdot H \]  
(2.36)

where

\[ R = \begin{pmatrix} \cos \theta & \sin \theta \\ -\sin \theta & \cos \theta \end{pmatrix} \]  
(2.37)

\[ \theta \]

\[ x' \]

\[ x \]

\[ y' \]

\[ y \]

**Figure 2.6: Axis rotation**

To transform the \( Z \) tensor, such that:

\[ E' = Z' \cdot H' \]  
(2.38)

then \( Z' \) must satisfy:

\[ Z' = R \cdot Z \cdot R^T \]  
(2.39)

where \( R^T \) is the transpose of \( R \).
As the diagonal elements of the rotated impedance tensor rarely reach zero, several different ways have been used to find the rotation angle $\theta_0$ between measurement direction and strike. One way for example, is to rotate the $Z_{ij}$ in steps, plot them on a polar diagram, and pick an optimum angle from the plots (Vozoff, 1991). Alternatively it is possible to use one of Swift's (1967) solutions, in which the expressions for $Z_{xy}(\theta)$ and $Z_{yx}(\theta)$ are differentiated with respect to $\theta$ to give an angle $\theta_0$ which minimizes:

$$\left|Z'_{xy}(\theta_0)\right|^2 + \left|Z'_{yx}(\theta_0)\right|^2$$

at each frequency, and results in an expression for $\theta_0$:

$$\theta_0 = \frac{1}{4} \arctan \frac{(Z_{xx} - Z_{yy})(Z_{xy} + Z_{yx})^* + (Z_{xx} - Z_{yy})^*(Z_{xy} + Z_{yx})}{\left|Z_{xx} - Z_{yy}\right|^2 - \left|Z_{xy} + Z_{yx}\right|^2}$$

where * denotes complex conjugate. This solution also maximizes $|Z_{xy}|$ and minimizes:

$$\left|Z_{xx}\right|^2 + \left|Z_{yy}\right|^2$$

In a ideal 2-D situation $Z'_{xx}$ and $Z'_{yy}$ should be zero at this angle $\theta_0$, but in practice they seldom reduce to zero. The angle $\theta_0$ is estimated with a $\pm 90^\circ$ ambiguity, which can be resolved if additional magnetic or geological information is available.

On rotation through 180 degrees, two minima for the diagonal elements are obtained and the corresponding axes are termed the principal conductivity axes. The apparent resistivities and phases in these directions are called major and minor apparent resistivities and phases. These are expressed as:

$$\rho_{maj} = 0.2 \cdot T \left|Z'_{xy}\right|^2 \quad \& \quad \Phi_{maj} = \arg(Z'_{xy})$$

$$\rho_{min} = 0.2 \cdot T \left|Z'_{yx}\right|^2 \quad \& \quad \Phi_{min} = \arg(Z'_{yx})$$

In order to minimize two-dimensional and three-dimensional effects it is common to define an effective response function for a medium which is rotationally invariant (Tikhonov and Berdichevsky, 1966; Ranganayaki, 1984). The effective impedance, which is the square root of the determinant of the impedance tensor and related to the determinant of the system equations (2.29) is given by:

$$Z_{eff} = (Z_{xx}Z_{yy} - Z_{xy}Z_{yx})^{1/2}$$
Chapter 2 MT and TEM methods

It has the physical sense of mean impedance for the medium. The corresponding apparent resistivity and phase are given by:

$$\rho_{\text{eff}} = \frac{1}{\mu \omega} |Z_{xx}Z_{yy} - Z_{xy}Z_{yx}| \quad \& \quad \Phi_{\text{eff}} = \text{arg}(Z_{xx}Z_{yy} - Z_{xy}Z_{yx})$$ (2.45)

These two quantities ($\rho_{\text{eff}}$, $\Phi_{\text{eff}}$) are the geometric mean of parallel and perpendicular resistivities and phases.

2.5.3 Impedance skew

To determine if two-dimensional interpretation is possible, a “two-dimensionality” measure can be constructed from the elements of the impedance tensor. An indicator of the “dimensionality” of the conductivity structure is the magnitude of the complex ratio of the off-diagonal elements of the impedance tensor, this ratio is called the Impedance Skew, and is expressed as (Swift, 1967):

$$\text{Skew} = \frac{|Z_{xx} + Z_{yy}|}{|Z_{xy} - Z_{yx}|}$$ (2.46)

It is a measure of the EM coupling between the measured $E$ and $H$ fields along coincident directions (Vozoff, 1972). For ideal 1-D and 2-D conductivity distributions, skew obviously reduces to zero. However, for real data this is seldom the case and it will only approximate to zero for a 1-D earth or a 2-D earth measured in strike angle. For the general 3-D structures skew will not be zero and an upper limit of 0.4 is usually placed on its value for a 2-D interpretation to be valid. Skew, does not change with rotation of coordinates.

2.5.4 Estimation of the impedance elements

The objective of the magnetotelluric data processing is to transform the electromagnetic field recorded as time functions into a frequency dependent response function ($Z$, $\rho$) used for interpretation. The processing thus includes two operations (Rokityansky, 1982): (a) spectral analysis of the time series and (b) estimation of elements of the impedance tensor.

The equations that must be solved are [equations (2.29)]:

$$E_x(f) = Z_{xx}(f)H_x(f) + Z_{xy}(f)H_y(f)$$

$$E_y(f) = Z_{yx}(f)H_x(f) + Z_{yy}(f)H_y(f)$$

i.e., two complex equations in four complex unknowns, the $Z_j(f)$. The equations are complex because all the quantities have both magnitude and phase. Obtaining two independent
estimates of the $E_i(f)$ and $H_j(f)$ ($i$ and $j$ can be $x$ or $y$), the equations can be solved. Since any measurement of $E$ and $H$ contains errors, it is desirable to have more than two independent records to allow a certain averaging for reducing the noise effect and estimating the error in the result. Sims and Bostick (1969) and Sims et. al. (1971) have discussed a least squares spectral analysis procedure for optimizing the estimates of the impedance elements from a large number of independent record sets. Suppose there are $N$ independent measurements of $E_x$, $H_x$, $H_y$ at a given frequency. To obtain the mean-square estimates of $Z_{xx}$ and $Z_{xy}$ we use the function (Sims et. al., 1971):

$$F = \sum_{i=1}^{N} (E_{xi} - Z_{xx}H_{xi} - Z_{xy}H_{yi}) \cdot (E_{xi}^* - Z_{xx}H_{xi}^* - Z_{xy}H_{yi}^*)$$  \hspace{1cm} (2.47)

where $E_{xi}^*$ is the complex conjugate of $E_{xi}$, etc., and find the $Z_{xx}$, $Z_{xy}$ minimizing $F$. For this purpose we take derivatives of the real and imaginary parts of (2.47), with respect to $Z_{xx}$ and, similarly, to $Z_{xy}$, equate them with zero and obtain:

$$\sum_{i=1}^{N} E_{xi}H_{yi} = Z_{xx} \sum_{i=1}^{N} H_{xi}H_{yi} + Z_{xy} \sum_{i=1}^{N} H_{yi}H_{yi}^*$$ \hspace{1cm} (2.48)

The summations are crosspower and autopower estimates of the field components. Equations (2.48) may be solved simultaneously for $Z_{xx}$ and $Z_{xy}$, and the solution will minimize the error caused by the noise on $E_x$. It is clear from the (2.48) that these expressions can be derived from equations (2.29) by multiplying by $H_{xi}^*$ and $H_{yi}^*$ and then summing, i.e. averaging all over the realizations. Multiplying both the equations (2.29) in turn by $E_{xi}^*$, $E_{yi}^*$, $H_{xi}^*$, $H_{yi}^*$ we obtain (Kao and Rankin, 1977; Goubau et. al., 1978):

$$|E_x|^2 = Z_{xx}b^* + Z_{xy}c^*$$ \hspace{1cm} (2.49)

$$a = Z_{xx}d^* + Z_{xy}e^*$$ \hspace{1cm} (2.50)

$$b = Z_{xx}|H_x|^2 + Z_{xy}f^*$$ \hspace{1cm} (2.51)

$$c = Z_{xx}f + Z_{xy}|H_y|^2$$ \hspace{1cm} (2.52)

$$|E_y|^2 = Z_{xx}d^* + Z_{xy}e^*$$ \hspace{1cm} (2.53)
\[ d = Z_{xx} |H_x|^2 + Z_{xy} f^* \]  \hfill (2.54)

\[ e = Z_{xx} f + Z_{xy} |H_y|^2 \]  \hfill (2.55)

\[ a^* = Z_{xx} b^* + Z_{xy} c^* \]  \hfill (2.56)

Here \( a, b, c, d, e \) and \( f \) are crosspowers defined by:

\[ a = \overline{E_x E_x^*}, \quad b = \overline{E_x H_x^*}, \quad c = \overline{E_x H_y^*} \]

\[ d = \overline{E_y H_y^*}, \quad e = \overline{E_y H_x^*}, \quad f = \overline{H_x H_y^*} \]  \hfill (2.57)

and the bar denotes the averaging. In principle, one can use any pair of equations (2.49) through (2.52) to obtain values for \( Z_{xx} \) and \( Z_{xy} \), and any pair of equations (2.53) through (2.56) to obtain values for \( Z_{yx} \) and \( Z_{yy} \). So for determining two impedance elements, e.g. \( Z_{xx} \) and \( Z_{xy} \), we have four equations [(2.49)-(2.52)] to form six pairs from which the two elements \( Z_{xx} \) and \( Z_{xy} \) can be calculated by six ways. But only four of them give stable results (Sims et al., 1971).

For example, from equations (2.49) and (2.58) one obtains:

\[ Z_{xy} = \frac{\overline{|E_x|^2} d^* - ab^*}{d^* c^* - e^* b^*} \]  \hfill (2.58)

while from equations (2.59) and (2.52) one obtains:

\[ Z_{xy} = \frac{c \overline{|H_x|^2} - bf}{\overline{|H_x|^2} |H_x|^2 - |f|^2} \]  \hfill (2.59)

In the absence of noise, all methods of estimating the impedance are equivalent. However, in the presence of noise, each estimate is different. The different values arise from the combination of random errors in the crosspower and autopower densities, and systematic bias errors produced by noise power in the autopowers. The estimate \( Z_{xy} \) in equation (2.58) is, evidently, biased upward by the noise in \( |E_x|^2 \), while in (2.59) \( Z_{xy} \) is biased downward by the noise in \( |H_x|^2 \). In this study a weighted average of the upward and downward averages was used (see section 4.1.2, equation 4.7). The influence of random and bias errors on impedance estimates has been discussed by several authors (Swift, 1967; Sims et al., 1971; Kao and Rankin, 1977; Goubau et al, 1978).
2.5.5 Coherence

The coherence (Swift, 1967) is a frequently used MT data quality test statistic calculated for two components (e.g. A and B) of the Fourier transformed EM fields:

\[
Coh(A, B) = \frac{[A^*B]}{([A^*A][B^*B])^{1/2}}
\]  

(2.60)

where \([A^*B]\) is the cross spectral power averaged over a range of frequencies. The coherence is real valued and lies between 0 and 1; the unity implying that the two signals are perfectly linearly related, and the zero that they are completely uncorrelated.

After the impedance elements have been determined one can substitute the observed magnetic field data in the equation (2.29) and calculate the electric field conventionally called the predicted field \(E_{ip}\). By comparing \(E_{ip}\) with the observed electric field one can judge the quality of \(Z_{ik}\) estimates. The result of the comparison is characterized by predicted coherency (Swift, 1967):

\[
Coh(E, E_{ip}) = \frac{E \cdot E_{ip}^*}{\left\{E^2, E_{ip}^2\right\}^{1/2}}
\]

(2.61)

\[
= \frac{E \cdot E_{ip}^*}{\left\{E^2 \left[Z_{1x}^2|H_s|^2 + Z_{0y}^2|H_y|^2 + 2 \text{Re}(Z_{1x} Z_{0y}^* H_s H_y^*)\right]\right\}^{1/2}}
\]

In the MT work the predicted coherence is preferred to the ordinary coherence.

In conclusion, we note that the presence of correlated noise in the components of the magnetotelluric field means that the estimates of the impedance tensor elements are generally biased.

2.6 Distortions in Electromagnetic Induction

The final objective of electromagnetic investigation is to obtain original and accurate information about the area studied and to estimate the likelihood of the quantitative models produced. Small-scale (relative to the background induction scale) or near-surface conductivity inhomogeneities can severely affect the magnetotelluric impedance tensor, reducing the quality and resolving power of MT-soundings and creating erroneous interpretations. We can define three different types of distortion results (Groom and Bahr, 1992): (1) The Static Shifts of sounding curves. (2) The mix of the regional impedances in areas of underlying two-
dimensional structure causing distortion to the level and shape of each sounding curve and phase. (3) The generation of anomalous magnetic fields that in turn alter the background phases, at sufficiently high frequencies.

Earlier reviews of the EM distortion problem are presented in Jones (1983a), Menvielle (1988), Jiracek (1990), Cerv and Pek (1990), Groom and Bahr (1992), and Chave and Smith (1994).

2.6.1 Characteristics of EM distortions

The physical principles governing the electromagnetic distortions have been understood and investigated for several years. A plane EM wave interacting with a laterally inhomogenous, 3-D earth induces anomalous excess currents and charge concentrations (Berdichevsky and Dmitriev, 1976; Jones, 1983; Berdichevsky et al., 1989). These produce respectively, inductive and galvanic distortion of the electric field measurements and consequently the impedance estimates.

Inductive distortion results from currents flowing preferentially in the more conductive region producing a frequency dependent distortion of both the apparent resistivity and phase response functions. At the highest frequencies and in strictly 2-D situations this will not be a problem. The inductive effect follows Faraday's law, whereby the time derivative of the primary magnetic field induces excessive currents. These currents flowing in closed loops produce secondary magnetic fields which add vectorially to the primary magnetic fields (Jiracek, 1990).

Galvanic distortion has been the subject of much recent interest, particularly with regard to near surface inhomogeneities. The galvanic effect is caused by the primary electric field producing electric charges where variations in conductivity occur. These charges result in a secondary electric field which add vectorially to the primary electric field. The underlying cause of galvanic electric field redistributions is that charges are built up on the boundaries of inhomogeneities when there is a component of the electric field \( E \) in the direction of a conductivity gradient (Jiracek, 1990).

An understanding of the galvanic electric field distortions alone enables the estimation of the surface effects on three-dimensional MT sounding curves even though the apparent resistivity calculation includes both the electric and magnetic fields. This is feasible because the 3-D galvanic magnetic field distortion is determined by a volume integral to the current distortion, which is small for near-surface bodies of small volume. In these cases, the apparent resistivity results are dominated by electric field variations.

Applying Ohm's law and the equation of current continuity, it is easy to show (e.g., Kaufman, 1985) that the volume charge density \( \rho \), in a conducting medium is given by:

\[
\rho = -\frac{j}{\sigma} \frac{\partial E}{\partial t} \tag{2.20}
\]
\[ \rho_v = -\frac{\varepsilon_0}{\sigma + i \omega \varepsilon_0} \mathbf{E} \cdot \nabla \sigma \]  
\hspace{1cm} (2.62)

where, \( \omega \) is the angular frequency, and the dielectric permittivity \( \varepsilon_0 \) has been taken as a constant. For most EM methods the quasistatic approximation is appropriate, therefore \( \sigma \gg i \omega \varepsilon_0 \) and

\[ \rho_v = -\frac{\varepsilon_0}{\sigma} \mathbf{E} \cdot \nabla \sigma \]  
\hspace{1cm} (2.63)

The secondary electric field set up by these boundary charges can be written in the form (LeMouel and Menvielle, 1982):

\[ \mathbf{E}_s = -\frac{\partial \mathbf{A}_s}{\partial t} - \nabla \phi \]  
\hspace{1cm} (2.64)

where \( \mathbf{A}_s \) is the secondary magnetic vector potential resulting from the boundary charges and \( \phi \) is the electrostatic potential. Following the hypotheses that the thickness of the inhomogenous upper layer is small with respect to the penetration depth of the transient field (LeMouel and Menvielle, 1982) the first term in equation (2.64) is negligible compared to the second so the secondary electric field is independent of frequency. The secondary electric fields \( \mathbf{E}_s \) can be quite large since, in the low frequency limit, by Coulomb's law:

\[ \mathbf{E}_s = -\nabla \phi = -\frac{1}{4\pi \varepsilon_0} \int \frac{\rho_s}{|r|^2} \mathbf{r}_0 ds \]  
\hspace{1cm} (2.65)

where \( \mathbf{r} \) is a vector between a differential surface element \( ds \) and the observation point, and \( \mathbf{r}_0 \) is a unit \( \mathbf{r} \) vector. From equations (2.63) and (2.65) it can be seen that the secondary galvanic electric field is proportional to the primary field, and that the two are in phase. Thus the effect of near surface inhomogeneities is to scale the electric field by a multiplicative factor. The magnetic field for a small size inhomogeneity and low frequencies remains almost unaffected by the build-up of boundary charges (Wannamaker et al., 1984a), however, the galvanic magnetic field distortion could be significant in situations of large galvanic distortion (Groom and Bailey, 1989).

Because only the electric field is affected, the MT log-log apparent resistivity is shifted by a time independent multiplicative scale factor and the phase lead of the electric field over the magnetic field is unchanged. MT sounding curves are shifted upward when measuring over surficial resistive bodies, and they are depressed over conductive patches, and this is because the effects of boundary charge build-up are a reduction or an enhancement of the total electric fields at different locations. Where the shifting up or down of the apparent resistivity curves is
frequency independent, the constant shift is referred to as a Static Shift. At very high frequencies, when the skin depth is less than the scale size of the inhomogeneity the effect is no longer frequency independent (Chave and Smith, 1994), since the first term in equation (2.64) is not negligible, and the agreement of the undistorted impedance phase is not valid.

2.6.2 Distortion correction techniques

Several proposals have appeared to partially address the 3-D galvanic distortion problem. The proposed techniques can be classified under six categories (Jiracek, 1990): (i) use of invariant response parameters (Berdichevsky and Dmitriev, 1976), (ii) curve shifting, (iii) statistical averaging (Berdichevsky et al., 1980), (iv) spatial filtering (Berdichevsky et al., 1989; Bostick, 1986), (v) use of distortion tensors, and (vi) computer modelling (Wannamaker et al., 1984b; Park, 1985). The details of these are beyond the scope of this discussion. However, since curve shifting and distortion tensor were applied in this study, the main theme of these two techniques is briefly presented here.

2.6.2.1 Static Shift correction

Accurate interpretation of the MT curves requires the removal of the static shift effects. The distorted curves have to be shifted to the true, correct, level they would have assumed in the absence of the anomalous small-size 3-D bodies that caused the distortion.

In absence of any complimentary information, Jones (1988) suggested as a solution to the problem, shifting the curves into agreement with the modal resistivity value of a known layer within a sedimentary basin. This technique is possible if many observations are available and the layer resistivity can be treated as a fixed parameter along the profile. Beamish and Travassos (1992) have considered three methods which attempt to remove static offsets: (i) individual curve shifting, (ii) statistical/spatial averaging, and (iii) the application of parameter constraints. In this proposal one or more independent constraints are required at each measurement location for the removal of static offsets. In the absence of such control the three methods are statistical and each must supply a constraint that has least conflict with the data characteristics.

It seems the most promising approach to a static shift correction is based on utilisation of independent surface electromagnetic data, specifically by use of transient or time domain EM soundings (TEM) (Sternberg et al., 1988). Unlike the telluric field the magnetic field is much less affected by surface inhomogeneities. TEM curves become insensitive to the surface inhomogeneity, thus the joint interpretation of MT and TEM curves is a way to eliminate the static shift. Transient Electromagnetic soundings were used by Andrieux and Wightman (1984), Sternberg et al., (1988), Pellerin and Hohmann (1990), Meju and Fontes (1993), Meju (1996)
and Meju et al., (1999). Sternberg et al., (1988) jointly invert the TEM and MT data with a 1-D computer code that includes a vertical MT static shift parameter but does not address any multi-dimensional modelling. They also demonstrated the accuracy of the corrected MT curves by comparison with well log information. Pellerin and Hohmann (1990) highlighted the importance of TEM based corrections in multi-dimensional environments. Meju (1996) jointly invert MT-phase and TEM apparent resistivity, emphasising that the impedance phase is not affected by static shift.

Joint interpretation of MT and TEM curves requires that the maximum depth of TEM sounding is greater than the minimum MT skin depth, and the medium at this depth interval is laterally uniform. Using joint inversion of MT phase and TEM apparent resistivity Meju (1996) showed that an overlap between the TEM and MT data sets is not always essential for retrieving a reliable resistivity structure. In the 1-D case, the TEM and MT curves will be parallel in the overlap area, and the scheme will accurately correct the data. If the curves are not parallel within the specified frequency range, 2-D or 3-D effects are present (Pellerin and Hohmann, 1990). The static shift correction procedure based on auxiliary TEM soundings seems to be very promising, but it does require another survey. However, relatively inexpensive field portable TEM equipment is now available for shallow-depth probing.

2.6.2.2 Distortion tensor technique

One dimensional and two dimensional modelling assume that the diagonal and rotated diagonal elements of the impedance tensor are zero. Any values present in these elements of a measured impedance tensor are classed as noise. However, experimentally determined magnetotelluric impedance tensors rarely conforms to the ideal 2-D impedance tensor. Therefore, a number of tensor decompositions have been proposed that utilize all four complex impedance elements in the calculation of their parameters.

Two tensor analysis methods more general than that of Swift (1967) have been developed. They are eigenanalysis and explicit tensor decomposition. The first is a mathematically based decomposition (Eggers, 1982; LaTorraca et al., 1986; Spitz, 1985; Yee and Paulson, 1987). This method is not used in this study, and will not be discussed further here. The second is presented by Bahr, (1988) and Groom and Bailey (1989, 1991), and is based on physical models.

**Physically-based decompositions**

A physical approach that yields more obviously physically meaningful parameters (e.g. regional strike and regional impedances) rather than mathematical parameters (e.g. eigenvalues) was introduced by Larsen (1977), and generalised from Bahr (1988) and Groom and Bailey (1989,
In summary the purpose of Groom and Bailey (1989) decomposition is to separate local and regional parameters as much as possible under the assumption that the regional structure is at most 2-D and the local structure causes only galvanic scattering of the electric fields, and to do so in the form of a product factorization.

The physical model used in the decompositions of Bahr and Groom and Bailey is of an inductively small local body of varying conductance overlying a regional two dimensional structure, the local body galvanically distorting only the telluric field produced by the regional structure. The model makes two simplifying assumptions: (1) The regional electromagnetic field is produced by only one dimensional or two dimensional bodies. Thus the decomposition model neglects any three dimensional inductive effects. (2) The model neglects distortion of the magnetic field by inductively weak bodies.

Writing $E = CE_r$, where $E$ is the "measured" electric field, $E_r$ is the "regional" field and $C$ is the scattering tensor operating on the regional field, $E$ can be written $E = Z_m H$, where $Z_m$ is the measured impedance. Then if $Z_2$ is the background or 2-D off-diagonal impedance tensor expressed in the strike co-ordinate system (regional impedance), $Z_m$ can be decomposed as:

$$Z_m = RCZ_2R^T$$

(2.66)

where $R$ is the rotation matrix from the measurement axes to regional coordinates. Groom and Bailey (1989) showed that the real 2x2 electric scattering or channelling tensor $C$ can be written as a product factorization of three basic suboperators: $C = g_T S A$. The tensors are responsible for "Twist", "Shear", and "Anisotropy", effects respectively, and the factor $g$ is a scalar "site gain" and is necessary for normalization. The suboperators have the form:

$$T = g_T \begin{bmatrix} 1 & -t \\ t & 1 \end{bmatrix}, \quad S = g_S \begin{bmatrix} 1 & e \\ e & 1 \end{bmatrix}, \quad A = g_A \begin{bmatrix} 1+s & 0 \\ 0 & 1-s \end{bmatrix}$$

where $t$, $e$ and $s$ are real numbers. The "anisotropy" tensor multiplied by the regional impedance tensor, simply adds to the anisotropy already existing in the regional induction impedance tensor $Z_2$. The "shear" tensor develops anisotropy along axes dissecting the principal directions of the regional impedance. The effects of tensors $A$ and $S$ on a family of unit vectors is shown in Figure 2.7. The right-hand side in Figure 2.7, shows the multiplication by tensor $S$ where $e$ is positive. A vector on the x axis is deflected clockwise through an angle $\tan^{-1}e$, and a vector along the y axis counter anti clockwise by the same angle. Therefore the
shear \( \epsilon \) is defined as an angle given by: \( \phi_\epsilon = \tan^{-1} \epsilon \). The effect of the "twist" tensor is simply to rotate the electric field vectors through a clockwise angle defined as twist angle: \( \phi_t = \tan^{-1} t \).

![Figure 2.7: A family of unit vectors (a) before and (b) after the application of the splitting tensor \( A \) on the left-hand side, and the Shear tensor \( S \) on the right-hand side. The \( x \) axis is up, the \( y \) axis to the right (After Groom and Bailey, 1989).](image)

Applying the factorization to equation (2.66) we have:

\[
Z_m = gRTSAZ_2 R^T
\]

Groom and Bailey concluded that is not possible to tell whether \( gA \) is different from the identity matrix \( I \) on the basis of a simple local measurement, so the terms \( gAZ_2 \) are lumped together.

So if \( gA \) is absorbed into \( Z_2 \) to give \( Z_2' \) then we have:

\[
Z_m = RTSZ_2' R^T
\]

(2.67)

The representation of \( Z_m \) in equation (2.67) constitutes the desired decomposition, and contains seven real parameters, three angles (azimuth, shear angle, and twist angle) and two complex principal impedances (real and imaginary part of the major and minor). They can be calculated from least squares or other fitting procedure to the data, giving a misfit \( \varepsilon \) that can be used as the eighth parameter, and detect the deviation from the ideal distortion model (Groom and Bailey, 1989). The seven real parameters that Groom and Bailey (1989) suggest,
compared to five in the conventional 2-D analysis looks more reliable, and do provide more physical insight than the full eight parameter impedance decompositions (e.g. Eggers, 1982; LaTorraca et al., 1986).

Groom and Bailey (1989) and Groom and Bahr (1992) show that although the first order effects of electric field scattering can be the most severe effect, magnetic effects caused either by the current channelling or induction can have significant effects on the estimated phases.

Finally, Bahr (1991) demonstrates how data can be categorized into seven grades of distortion and suggest more reliable undistorted skew and strike parameters, similar to Groom and Bailey (1989). Following the notation of Bahr (1991), we define:

\[ S_1 = Z_{xx} + Z_{yy} \]
\[ S_2 = Z_{xy} + Z_{yx} \]
\[ D_1 = Z_{xx} - Z_{yy} \]
\[ D_2 = Z_{xy} - Z_{yx} \]

In the following equations, the phase differences between two complex numbers and the corresponding amplitude products are abbreviated by the commutators:

\[ [C_1, C_2] = \text{Im}\{C_2C_1^*\} = \text{Re}\{C_1\} \text{Im}\{C_2\} - \text{Re}\{C_2\} \text{Im}\{C_1\} \]

The angle \( \alpha \) of the direction of the regional strike is:

\[ \alpha = \frac{1}{2} \arctan \left( \frac{[S_1, S_2] - [D_1, D_2]}{[S_1, D_1] + [S_2, D_2]} \right) \quad (2.68) \]

we define also, the phase-sensitive skew \( n \) :

\[ n = \frac{\sqrt{[D_1, S_2]-[S_1, D_2]}}{|D_2|} \quad (2.69) \]

a rotationally invariant measure of phase differences in the impedance tensor:

\[ \mu = \frac{\sqrt{[D_1, S_2]+[S_1, D_2]}}{|D_2|} \]

and a rotationally invariant measure of the two-dimensionality :

\[ \Sigma = \frac{(D_1^2 + S_2^2)}{D_2^2} \]

Using the above parameters \( (n, \mu, \Sigma) \) and the skew \( k \) as defined by Swift \( (k = |S_1|/|D_2|) \) Bahr (1991) provides a "cookbook" for the evaluation of measured impedance tensors. The
seven classes defined below emphasise the separation of effects due to galvanic distortion and those due to regional induction.

- **Class 1**: The simple 2-D anomaly characterised by $\Sigma > 0.1$ and $k < 0.1$. The tensor impedance is either undistorted (Cagniard's (1953) model of a layered half space would be appropriate) or it is described by Swift's (1967) model.

All “higher” classes deal with cases in which the skew $k$ does not vanish, e.g. $k > 0.1$, and the method of Groom and Bailey (1989) needs to be applied.

- **Class 2**: The purely local 3-D anomaly superimposed upon a layered earth characterised by $n < 0.05$

- **Class 3**: A regional 2-D anomaly with weak local distortion, characterised by either $\phi_x - \phi_y < 5.0$ and $\phi_x + \phi_y < 20.0$ or $\phi_x - \phi_y < 20.0$ and $\phi_x + \phi_y < 5.0$.

- **Class 4**: A regional 2-D anomaly in rotated coordinates, characterised by $\phi_x = 0$.

- **Class 5**: A regional 2-D anomaly with strong local distortion characterised by $n < 0.3$ and large twist and shear angles.

- **Class 6**: A regional 2-D anomaly with strong local channelling characterised by $\phi_x \approx 45.0$.

- **Class 7**: A regional 3-D anomaly characterised by $n > 0.3$, which cannot be interpreted with two-dimensional models.

In conclusion, the current decompositions can be extremely useful, but they are also fraught with many problems. Strong current channelling can lead to a poorly determine inverse problem. Large noise coupled with a weak 2-D inductive response results in a poorly resolved strike direction. 3-D phase effects from an anomalous magnetic field adds to these problems (Groom and Bahr, 1992). Nevertheless, mathematical and galvanic 3-D decomposition methods adds valuable information about the scattering processes governing the data. And as Groom and Bailey (1989) have concluded, there is no internal solution to the statics problem from the MT data at a single location, and external data help to solve the problems of statics and strike direction.
2.7 The Transient Electromagnetic Method (TEM)

In the Transient Electromagnetic method (TEM) a strong direct current (dc) is passed through an ungrounded loop. The flow of the current in the loop creates a primary magnetic field. The rapid termination of the current generates a system of induced currents in subsurface conductors, called eddy currents, which flow in such a way as to preserve the magnetic field that existed before the current was switched off. The eddy currents and the secondary magnetic field which they produce decay with time and induces a voltage in a receiver coil.

It can be shown (Nabighian, 1979) that, on or above the surface of the earth at any given time after switch off, the combined effect of all induced currents in the ground can be approximated by the effect of a single current filament of the same shape as the transmitter loop, that moves outwards and downwards with time (Figure 2.8). This is known as the "smoke ring" concept and is a useful way of visualising the complex process occurring in the earth.

![Figure 2.8: The use of equivalent current filament concept in understanding the behaviour of TEM fields over a conducting half space (after Nabighian and Macnae, 1991)](image)

The amplitude of the current filament decreases with time as does its velocity in a homogeneous earth. For a homogeneous earth of conductivity $\sigma$, the wave velocity $V_z$ with which the ring expands away from the transmitter, at a time $t$ is given by the equation (Nabighian, 1979):

$$V_z = \frac{2}{\sqrt{\pi \sigma \mu}}$$
The diffusion depth, \( d \), of the wave at time \( t \) is given by:

\[
d = 2\pi \sqrt{\frac{2t}{\mu \sigma}}
\]

The current filament moves down away from the transmitter loop at an angle of approximately 47 degrees with the surface (Nabighian, 1979).

The equation solving the TEM forward problem for the layered earth problem with a step source is given in Knight and Raiche (1982) and Raiche (1984). This gives the mutual impedance, \( Z \), between the transmitter and receiver loops, located on the ground surface, as a function of measurement time \( t \). This equation is:

\[
Z(t) = -\pi \mu a b \int_{0}^{\infty} L_{-1}^{-1} \left\{ \left( I(p) p A_{0}(m, p, \lambda) \right) J_{1}(\lambda a) J_{1}(\lambda b) d\lambda \right\}
\]  \hspace{1cm} (2.70)

where \( a \) and \( b \) are respectively the radius of the transmitter and receiver loops, \( A_{0}(m, p, \lambda) \) is the layered earth impedance function, \( m \) represents the layer thickness and conductivities, \( p \) is the Laplace transform variable corresponding to \(-i\omega\) where \( \omega \) is the angular frequency, \( \lambda \) is the integration variable for the inverse Hankel transform and is equal to \( -p^{-1} \) for a step function turn-off, \( J_{1} \) is the Bessel function of order one, and \( L_{-1}^{-1} \) is the inverse Laplace transform operator with respect to \( p \). Equation (2.70) is evaluated using the Gaver-Stehfest method of Knight and Raiche (1982) and linear digital filters for calculation of the inverse Hankel transforms.

The layered earth impedance function \( A_{0} \) is found by a back-substitution process (Knight and Raiche, 1982; Raiche, 1984). For a structure with \( n \) layers above a uniform half-space of conductivity \( \sigma_{n+1} \), the reflection coefficient at the interface between layer \( j \), of conductivity \( \sigma_{j} \), and layer \( j+1 \) of conductivity \( \sigma_{j+1} \) is:

\[
R_{j} = \frac{S_{j} - S_{j+1}}{S_{j} + S_{j+1}}
\]

The exponential factor \( E_{j} = \exp(-2S_{j}h_{j}) \) is defined where \( h_{j} \) is the thickness of layer \( j \) and \( S_{j} = (\lambda^{2} - i\omega\mu\sigma_{j})^{1/2} \). Working from the bottom to the top for the sequence of the layers, we calculate

\[
F_{n+1} = 0
\]
for the basal half-space,

\[ F_N = R_N E_N \]

for the overlying layer \( n \), and

\[ F_j = E_j \frac{(R_j + F_{j+1})}{1 + R_j F_{j+1}} \]

for any layer \( j \) \((0<j<n)\) on top of layer \( n \). Using this recursive relation, the function \( A_0 \) is simply:

\[ A_0 = \frac{R_0 + F_1}{1 + R_0 F_1} \]

2.7.1 TEM measurements

In TEM surveys the transmitter is connected to a loop of wire laid on the ground. Most of the TEM transmitters generate a bipolar waveform with a ramp turn-off time as illustrated in Figure 2.9. The receiver measures the decaying voltage when the transmitter is turned off; this allows the very small voltages to be measured without interference from the large primary field. The receiver samples the amplitude of the transient decay using a number of gates.

![Schematic diagram showing transmitter current linear ramp turn-off time and received signal (after Swift, 1990).](image)

**Figure 2.9:** Schematic diagram showing transmitter current linear ramp turn-off time and received signal (after Swift, 1990).
The calculation of the transient electromagnetic response is usually based on the assumption that the current turn-off waveform is a step function. In reality electronic limitations mean that this is not valid and the current switch off takes a finite time called turn-off time or ramp time. Many transient systems are designed to turn off the current in a linear ramp whose width is measured by the transmitter. The advantage of this is that the finite turn-off time can be accounted for during data processing.

The depth of investigation is an important aspect for both the design and the interpretation of TEM surveys. The depth of exploration depends primarily on the duration of the measuring time, \( t \), on instrument specifications (e.g. dynamic range and resolution) and on ambient noise; it also depends much on the resistivity profile of the subsurface. So many independent parameters render the depth of exploration difficult to quantify. Spies (1989) demonstrated that in the near-zone (transmitter-receiver separation is less than the depth of investigation) TEM soundings, where the induced voltage is measured with an induction coil, the depth of investigation is proportional to the \( 1/5 \) power of the source moment and ground resistivity \((I\Lambda\rho)^{1/5}\). If the receiver is an \( H_z \) magnetometer, the depth of investigation is controlled by the source loop moment instead of ground resistivity (see Spies, 1989; equation 12, Figure 7). In far-zone (loop size is much greater than the depth of investigation) TEM soundings, the depth of investigation for an induced voltage receiver is proportional to the \( 1/4 \) power of source moment and resistivity and is inversely proportional to the source-receiver separation (Spies, 1989). This analysis gives the maximum depth of investigation based on detection of a buried layer. The minimum depth, at which individual conductivity variations can be resolved is determined by the earliest sample time.

### 2.7.1.1 Loop configurations

To produce magnetic fields in the earth to generate eddy currents some form of inductive transmitter (Tx) is usually used together with an inductive receiver (Rx) to measure the secondary magnetic field. The most common TEM survey configurations are the **coincident** and the **central loop** arrangements (Figure 2.10).

The **coincident** or **single-loop** configuration is the most simple, and perhaps the most commonly used TEM array. The same wire loop is used for both transmitter and receiver, and the configuration is particularly suited to fast efficient reconnaissance surveys with typical loop sides of 25-200m. Major advantages of the coincident loop include good signal to noise ratios and a uniform response as a function of transmitter location. This ensures a degree of coupling with any target at any orientation, and a minimisation of geological noise. Main problems with coincident loop surveys are IP and superparamagnetic effects (description is given in section
2.7.1.2). Thermal transients left from the transmission, since the same wire is used as a Tx and a Rx, are also a problem, especially when using high currents.

![Figure 2.10: Common TEM loop configurations.](image)

The central loop or in-loop configuration uses a multi turn receiver (usually a multiturn coil of 1m diameter) centred within a square transmitter loop. An operational requirement is that the receiver coil needs to be accurately located on the centre of the transmitter loop. This significantly increases the time needed for each sounding in comparison with single-loop method, but the signal amplitude is largest at the centre of the Tx loop. There are, however, several advantages to the central loop configuration. Several different receivers can be used consecutively to record the transient over a large dynamic range with better resolution. This is also ideal for 3-D environments, where horizontal components of the secondary magnetic field, $B_x$ and $B_y$, can be recorded by placing the receiver vertically at the loop centre at two orthogonal orientations. Shallow anomalies near the centre of the loop will have a large effect on the central receiver. Generally the central loop arrays give better resolution for conductors, and have better S/N ratio, than the coincident loop.

### 2.7.1.2 Noise sources

In the frequency range (1 to $10^5$ Hz) that TEM utilises there are three main types of noise sources that can distort a TEM sounding: a) cultural or man made noise, b) natural electromagnetic noise and c) geological noise.

Instrument noise arising from electronic components on circuit boards and wind noise, are also a source of errors. Instrument noise is significantly reduced by stacking whereby a series
of readings is taken and the results averaged to give the final figure. The wind noise can be reduced to zero if care is taken to place the entire loop on the ground.

**Cultural or man made noise**

Currents induced in metallic conductors such as power and telephone lines, pipelines, fences etc. can produce anomalous TEM responses. Man made noise comes mainly from the electric power distribution grid (50 or 60 Hz) while VLF radio stations create higher frequency noise (10 to 25 kHz). Most of the TEM instruments have internal filters (notch) which are set to the local mains frequency reducing the effect of the noise. Harmonics of the steady power-line noise can be rejected by tapered stacking. In areas with very high power line noise it is not possible to carry out TEM soundings as the instruments become saturated with the noise.

**Natural Electromagnetic noise.**

There are several sources of EM noise that effect TEM measurements. Sferics are one of the most serious sources of noise. Sferics are electromagnetic signals produced by lightning strikes (section 2.1.1). The waves travel directly to the receiver if the source is near, or by multiple reflections from the earth's ionosphere if the source is distant. This sferics noise can be of such high strength to become the main limitation in obtain sufficient S/N ratio. Below about 6 Hz, the natural EM noise field is primarily of geomagnetic and ionospheric origin with relatively little noise present in the range 1 to 6 Hz (McCracken et al., 1984). Additional details on EM noise can be found in McCracken et al., (1984), and Spies and Frischknecht (1991).

**Geological noise**

In terms of TEM soundings interpreted with 1-D algorithms, geological noise can be considered as any feature which will distort the fields from those expected from a homogeneous horizontally layered earth. The noise sources can be divided into several groups (Spies and Parker, 1984; Swift, 1990):

- Lateral and vertical resistivity variations within layers which are caused by differences in pore fluid content, alteration, weathering and sediment deposition.
- Depth variations of the interfaces between layers.
- 3-D bodies within the survey area which cause current channelling and gathering.
- Discontinuities such as faults and dykes which will interrupt the current flows especially where they form the interfaces between two lithologies.

McCracken et al., (1986) discuss the effects of lateral variations in the overburden usually produced by weathering. Newman et al., (1987) assess the bias of fitting constrained layered-earth models to TEM data obtained over 3-D structures. They concluded that layered-earth
interpretations can accurately locate the depth of the burial of these structures but not the depth extent and resistivities. Geological noise also occur when the rock properties depart from the simple resistive and magnetic properties found in most rocks. Variations in dielectric constant $\varepsilon$ cause displacement currents which are not normally considered in the calculation of TEM responses. This happens when very high frequencies or large spacings are employed, or the resistivity is very high, as in a granite. Superparamagnetic rocks cause additional errors. Buselli (1982) reports anomalous transient recordings in Australia, characterized by $1/t$ time dependence, leading to incorrect apparent resistivity determinations with time. The main cause of this anomalous transient behaviour was shown to be the presence of superparamagnetic material in the lateritic soil cover having a frequency-dependent magnetic permeability (Swift, 1990).

Finally, another source of noise in TEM surveys is the induced polarisation (IP) effect. Where variation in conductivity with frequency exists, the conductivity tends to increase with increasing frequency, and such materials are termed polarizable. The effect of IP currents on a TEM decay curve is to increase the voltage at early times and to decrease it at late times. In extreme cases this may lead to the transient decaying through zero and becoming negative before asymptoting to zero. More information can be found in Spies (1980), Lee (1981), Flis et al. (1989), and Hatzichristodulu, (1998).

2.7.2 TEM data processing

TEM data from a field instrument consist of an approximately exponential voltage decay curve. The way to interpret this data is to convert the field data to apparent resistivities. The apparent resistivity is a form of data normalisation and is the resistivity of a homogeneous half-space which will produce the same response as that measured over the real earth with the same acquisition parameters. The raw Sirotem field data are in nV/A and have been normalised for the transmitter (Tx) current and the receiver (Rx) gain. In addition, the voltages have to be normalised for the receiver area before being converted to apparent resistivities.

The turn-off time has to be accounted for before conversion of the voltages to apparent resistivities. Several ways of dealing with the turn-off time have been reported in the literature (Raiche, 1984; Asten and Price, 1985; Fitterman and Anderson, 1987). The approach that has been followed in this project is based on Raiche (1984). The advantage of this method is that the field data give a representation of the actual resistivity variations. Once the voltages have been converted to apparent resistivities then the next stage is the modelling of the data, and the ultimate goal is to deduce the subsurface resistivity distribution (resistivity-depth information) from these sounding curves.
2.8 Numerical Modelling of Electromagnetic Data

The first objective of interpretation in any geophysical study is to derive a model compatible with the observed data. Then inferences may be made from the model as to the nature and extent of the particular structure under investigation.

The response functions acquired by MT and TEM soundings usually consist of estimates of the apparent resistivity and phase over a range of frequencies or times. The conductivity structure is modelled to fit the observed variation of these responses.

A preliminary study of the results may afford a qualitative interpretation but any quantitative analysis requires modelling of the data. Modelling of EM data may proceed either via (i) the forward solution, in which a conductivity model is constructed, its inductive response computed and compared with the measured data, or (ii) the inverse solution, the direct retrieval of the electrical properties (i.e. the conductivity distribution) of the earth directly from the data. Both methods have been used in this study.

2.8.1 1-D Electromagnetic Modelling

The main task of mathematical modelling in electromagnetic induction is to predict the observed responses over a given distribution of the earth's electrical parameters. One dimensional modelling is the simplest and the most widely used class of EM modelling methods which applies to TEM and MT. Strictly, it can only be applied in regions which have no lateral variation in resistivity, but in practice it is frequently used even in complex geoelectric environments to obtain starting models for more sophisticated, higher dimensional procedures.


2.8.1.1 Approximate models

The usual way to present magnetotelluric data is to plot the apparent resistivity and the phase against frequency, possibly in a rotated coordinate system. Alternatively, there are two widely used schemes which give an approximate model from smoothly varying estimates of the apparent resistivity. The Schmucker (Schmucker, 1970; 1987) and Niblett-Bostick (Niblett and Sayn-Wittgenstain, 1960; Bostick, 1977; Jones, 1983b) transformations give a continuously varying resistivity – depth profile. They are ideal for use in the field owing to their very small computational cost, or as in this study, the Niblett-Bostick transformation was used as the starting point for 1-D modelling or preliminary interpretation.
In Schmucker transformation the data can be presented in $p^* - z^*$ diagrams (Schmucker, 1987):

$$p^*(\omega) = p_a(\omega) \cdot \begin{cases} \frac{1}{(2\sin^2 \phi(\omega))}; & \text{for } \phi \leq 45^\circ \\ \frac{2\cos^2 \phi(\omega)}{\mu_0\omega}; & \text{for } \phi > 45^\circ \end{cases}$$

$$z^*(\omega) = \frac{1}{\mu_0\omega} \text{Im}\{Z(\omega)\}$$

The $p^* - z^*$ transformation is based on plane 2-layer models with two adjustable parameters: an $h$-model for a very poorly conducting top layer ($\phi > 45^\circ$) above a uniform halfspace and a $\tau$-model ($\phi \leq 45^\circ$) for which the top layer is a very good conductor (Schmucker, 1987).

The Niblett-Bostick transformation is a simple mapping of the MT data into resistivity-depth information using an algorithm based upon the asymptotic response of the impedance data in a horizontally layered model with infinitely conducting or resistive substrata. A continuous resistivity-depth curve may be obtained for the resistivity and phase independently. The relevant expressions are:

$$h = \sqrt{\frac{\rho_a(T)T}{2\pi\mu_0}}$$

and

$$\rho^*(h) = \rho_a(T) \left(\frac{\pi}{2\phi} - 1\right)$$

where $h$ is the penetration depth in a half space medium of resistivity equal to the observed value $\rho_a$ at the particular period and $\rho^*$ is an alternative expression that incorporates phase ($\phi$, in radians) information for the Niblett-Bostick resistivity at depth $h$ (Weidelt et al., 1980).

In the TEM case, a comparable data transformation scheme recently developed by Meju (1995, 1996) has been applied. This data transformation scheme closely resembles the Bostick (1977) scheme for magnetotelluric data transformation. It was determined empirically from numerical modelling studies and skin depth considerations. According to this transformation scheme, the effective depth of electromagnetic imaging is given by:

$$\delta_{eff} = \frac{\delta_r}{2.3} \quad (2.71)$$
where $\delta_t$ is the analogous quantity to the skin-depth for the TEM situation, and is referred as the "diffusion depth" given by:

$$\delta_t = \left(\frac{2t\rho_a}{\mu_0}\right)^{1/2}$$

(2.72)

where $\rho_a$ is the observed apparent resistivity at time $t$ and $\mu_0$ is the magnetic permeability of free space. For the same transformation scheme the effective resistivity is given by:

$$\rho_{eff} = k\rho_a e^{-(1-a)}$$

(2.73)

where the variable parameter $a$ in the exponent takes a value between 0.15 and 0.2 for field data, and the constant $k$ may be needed to improve the fit between the field and computed model response.

### 2.8.1.2 Forward modelling

The forward solution to the 1-D induction problem represents the simplest approach to EM interpretation. The usual strategy in curve matching methods (Cagniard, 1953; Yungul, 1961; Wait, 1962), is one of trial-and-error in an attempt to find a close fit between experimental data and the computed responses for a known and specified earth model. These curves are generated by application of equation (2.27) in the MT case and equation (2.70) for TEM assuming a horizontally homogeneous model with 1-D resistivity distribution. If there is a compatibility between measured and theoretical data then the model can be considered a possible approximation to the earth's actual conductivity structure. By varying different parameters of the model and recomputing the inductive response, the sensitivity of the data to certain parts of the model can be investigated and a best fit to the data can be obtained.

Forward solution offers an intuitive feeling for the most critical parameters. However, the method cannot quantify the non-uniqueness of the solution, since only one conductivity distribution is derived that satisfies the data. The introduction of high speed digital computers allowed the EM response function to be computed more quickly by direct solution of the recurrence relation for any specified multi-parameter model.

### 2.8.1.3 Inverse modelling

The inverse problem attempts to retrieve the conductivity structure of the earth directly from the data. The problem is described as: given some information on the values of some measured quantities (field data) a theoretical relationship is used to derive the values of the set of parameters that explains the field observations. The inversion of the MT data for the
conductivity distribution is a non-linear and a non-unique process (e.g. Oldenburg, 1979; Parker and Whaler, 1981; Smith and Booker, 1988).

2.8.1.3.1 Aspects of the inverse problem

Inversion is the procedure of estimating physical earth parameters from a set of observed data. In electromagnetic methods, the model responses are non-linear functions of the appropriate model parameters. Linearized or non-linearized inversion schemes provide two methods for dealing with non-linear inversion problems. The main types of these solution methods are least squares (e.g. Marquardt-Levenberg and Gauss Newton methods) and gradient (e.g. steepest descent and conjugate gradient) methods (Marquardt, 1963; Levenberg, 1944).

It is worth discussing some aspects of the inverse problem for the geoelectromagnetic induction model. As stated previously, the inverse problem of EM is non-linear but the solution method effectively relies on the linear inversion theory. The outlines given here are based on some books (e.g. Menke, 1984; Meju, 1994). The non-linear electromagnetic induction model can be expressed in the form:

\[ d = F(m) + e \] (2.74)

where \( d \) is a vector of the \((n \times 1)\) observed data, \( m \) is the vector of the unknown \((m \times 1)\) model parameters (here, layer resistivities and thicknesses), \( F \) is the set of nonlinear functions of the model parameters that facilitates the theoretical predictions known as the forward model, and \( e \) is the \((n \times 1)\) column vector of additive noise (errors).

The mathematically robust least-squares formalism is generally adopted and nonlinearity is usually addressed via iterative procedures. In inversion the goal is to find the model parameters that minimise the differences between the field data and that predicted by the forward theory. This is achieved by minimising the sum of the squares of the residuals \( e \) which is given by:

\[ q = e^T e = [d - F(m)]^T [d - F(m)] \] (2.75)

where \( T \) indicates transpose. Differentiation yields a system of nonlinear equations that are difficult to solve. The general procedure is to start with a first approximation to the actual model parameters (initial model) and then successively improve the values of the estimates until \( q \) is minimized. It can be assumed that the model response function is linear about the initial guess \( m^0 \) and it is standard practice to perform the first order Taylor series expansion, an approximation that is valid only if the series converges:

\[ d \approx F(m) = F(m^0) + \frac{\partial F(m^0)}{\partial m}(m - m^0) \]
so that,

$$d - F(m^0) = \frac{\partial F(m^0)}{\partial m}(m - m^0). \quad (2.76)$$

Defining an \((n \times p)\) matrix of partial derivatives of \(F\) with respect to each of the model parameters, \(A\), called the Jacobian matrix:

$$A = \left[ \frac{\partial F(m^0)}{\partial m} \right] \quad (2.77)$$

defining an \((n \times 1)\) discrepancy vector, \(y\), containing the differences between the initial model response \(F(m^0)\) and the observed data:

$$y = d - F(m^0), \quad (2.78)$$

and a \((p \times 1)\) parameter change vector \(x\):

$$x = (m - m^0) \quad (2.79)$$

equation (2.76) can be re-expressed as: \(y = Ax\), and the problem \((d = F(m))\) is fully linearized. So equation (2.75) becomes:

$$q = e^T e = (y - A x)^T (y - A x). \quad (2.80)$$

To minimize the above equation, differentiation is performed with respect to each model parameter \(x_i\) and set the result equal to zero. Using the least squares method it follows that:

$$\frac{\partial q}{\partial x} = \frac{\partial}{\partial x} (y^T y - x^T A^T y - y^T A x + x^T A^T A x) = 0$$

or

$$A^T A x + x^T A^T A - y^T A - A^T y = 0 \quad (2.81)$$

which is equivalent to the vector equation:

$$A^T A x - A^T y = 0 \quad \text{or} \quad (2.82)$$

Assuming that \((A^T A)\) exists, the above equation leads to the least squares solution for the parameter perturbations:

$$x = (A^T A)^{-1} A^T y$$
which is the parameter correction to be applied to the initial model \( m^0 \). The quantity 
\((A^T A)^{-1} A^T\) is known as the pseudo inverse or Lanczos (1961) inverse and is usually referred
to as the generalized inverse of the rectangular matrix. This is the Gauss-Newton solution and
can be applied successively to improve the initial model until an optimal model \( m^* \) is obtained.
The iterative formula \( m^k = m^{k-1} + x^k \) is used to solve the nonlinear problem. When \( x \)
becomes negligible then the \( m^k \) is a good approximation of the true solution \( m \). The
drawbacks of this technique are that a good approximation to the actual model is required for
the procedure to converge and that the matrix \( A^T A \) may be singular or near singular (Meju,
1994).

The \((n \times p)\) matrix, \( A \), can be factored (Lanczos, 1961) into a product of three other
matrices:

\[
A = U \Lambda V^T
\]

where for \( n \) data and \( p \) model parameters, \( U(n \times p) \) and \( V(p \times p) \) are the data space and
parameter space eigenvectors respectively and \( \Lambda \) is a \((p \times p)\) diagonal matrix containing at
most \( r \) non-zero eigenvalues of \( A \), with \( r \leq p \). The diagonal entries in \( \Lambda(\lambda_1, \lambda_2, ..., \lambda_p) \) are
called the singular values of \( A \). The eigenvalues of \( A^T A \) are explicitly related to those of \( A \)
and \( A^T A \) is said to be ill-conditioned if the eigenvalues are very small or near zero. This is
called Singular Value Decomposition (SVD). The SVD method is popular in geophysical
analysis because it is robust mathematically and stable numerically, providing vital information
on the state of the model and data, enabling model resolution and covariance studies.

To prevent unbounded solution growth when \( A^T A \) is ill-conditioned, the step length of the
solution is constrained by placing a bound on the parameter changes (Levenberg, 1944). To
damp the absolute values of the increments during the successive applications of Taylor
approximations, arbitrary positive weights are added to the main diagonal of \( A^T A \). The
direction of the residual sum of squares has a minimum when the weights are equal
(Levenberg, 1944). This technique is known as the Marquardt-Levenberg or Ridge regression
method (Levenberg, 1944; Marquardt, 1970).

In Ridge regression it is not only the residual \( e \) that is minimized but also the solution
length \( x \). The minimization problem becomes:

\[
\phi = q_1 + \beta q_2 = e^T e + \beta(x^T x - L_0^2)
\]

\[(2.83)\]
where $L_p^2$ is the bound on the energy of the parameter increments and $\beta$ is the damping factor, an undetermined Langrange multiplier which determines the relative importance that is given to $q_1$ and $q_2$. From minimization, is derived the Ridge estimator for the parameter perturbation:

$$x = (A^TA + \beta I)^{-1} A^T y .$$  \hspace{1cm} (2.84)

Adding the damping factor to the matrix $A^TA$ increases the size of the eigenvalues, thus leading to a stable inversion of the matrix $(A^TA + \beta I)$. There are many different ways of obtaining the damping factor $\beta$ (Jupp and Vozoff, 1975; Johansen, 1977; Meju, 1992).

Using the SVD representation of $A$ ($A^T = V \Lambda U^T$) the least squares solution given by equation (2.84) can be written as:

$$x = (A^TA)^{-1} A^T y = (V \Lambda^{-2} V^T) U \Lambda U^T y = V \Lambda^{-1} U^T y$$  \hspace{1cm} (2.85)

where it is assumed that $r = p$. In ridge regression, the element $1/\lambda_i$ in the $\Lambda^{-1}$ matrix can be replaced by the element:

$$\lambda^{-1}_{nj} = \frac{\lambda_j}{(\lambda_j^2 + \beta)}$$

so that the inverse formula of equation (2.84) becomes:

$$x = V \left( \frac{\Lambda}{\Lambda^2 + \beta I} \right) U^T y .$$

If observational errors are available, they can be incorporated directly in the problem formulation to obtain a more acceptable weighted solution (Meju, 1994). If the $n$ associated errors $\sigma_i$ are Gaussian with zero mean and statistically independent, then a diagonal weighting matrix $W$ can be defined as: $W = diag\{1/\sigma_1, 1/\sigma_2, ..., 1/\sigma_n\}$. This is used to scale the observed data to prevent undue importance being given to poorly estimated data (e.g. Meju, 1992). So the ridge regression estimate (2.84) becomes:

$$x = [(WA)^T W A + \beta I]^{-1} (WA)^T W y .$$

In addition to the model parameters, the inversion procedure provides a plethora of related information that can be used to assess the solution to the inverse problem. The root mean square (rms) error, the parameter resolution matrix, and the covariance matrix are three
parameters that gauge the goodness of the solution. The root mean square error evaluates the fit between the observed and calculated data, it is defined as (Meju, 1994):

\[
\text{rms} = \frac{1}{n} \sum_{i=1}^{n} \frac{(d - F(m))^2}{\sigma^2}
\]

and for a weighted solution:

\[
\text{rms} = \frac{1}{n} \sum_{i=1}^{n} ||Wd - WF(m)||^2
\]

Another measure of goodness of a solution is the chi-square misfit given by the formula (e.g. Meju, 1988):

\[
q = \sum_{i} |y - Ax|^2
\]

The parameter resolution matrix describes how well the predictions match the data (Menke, 1984), given by (Jackson, 1972):

\[
R = HA = (VA^{-1}U^T)(UAV^{-1}) = VV^T
\]

where \( U, V \) have been previously defined and the matrix \( H \) is the inverse used. \( R \) is of dimension \( p \times r \) where \( r \), the number of non-zero eigenvalues, is the degree of freedom of the problem. If \( R=I \), then each model parameter is uniquely determined. The deviation of the rows of \( R \) from those of the identity matrix \( I \), measures the lack of resolution for the corresponding parameters. The covariance matrix \( C \) is a \((p \times p)\) matrix that depends on the covariance of the experimental errors and the way in which error is mapped from data to model parameters (Menke, 1984). If the data are uncorrelated and of equal variance \( \delta^2 \), then:

\[
C = \delta^2 (A^T A)^{-1}.
\]

The off-diagonal elements, the covariances, indicate the correlation between the model parameters. The square roots of the diagonal elements of the matrix are generally referred to as standard deviations of the least squares parameter estimates and may be used to estimate the bounds of the model parameters (Meju, 1994). Large off-diagonal elements \( C_{ij} \) mean the \( i^{th} \) and the \( j^{th} \) model parameters are highly correlated.
2.8.1.3.2 Inversion of EM data

The non-linear nature of the MT problem, the limited bandwidth of measurements, coupled with observational errors, biasing effects both due to cultural noise or lateral effects, along with the "smearing out" of sharp conductivity contrasts or thin layers by the diffusing EM energy, ensure that an infinite set of conductivity structures will satisfy a finite set of EM data (Meju and Hutton, 1992). This property of non-uniqueness forms the major problem in the interpretation of resistivity depth models derived from EM data, and it becomes apparent that additional independent geophysical or geological information concerning one or more of the parameters should be incorporated into the procedure to reduce the range of non-uniqueness. Apart from finding the model which best fits the observational data, some representation of the uniqueness of the model should be provided.

There are two main approaches to the construction of solutions to the 1-D inverse problem. They may be categorized as parametric (discontinuous) methods or continuous. The commonest technique is the first method in which the model earth is parameterized in terms of a small number of layers (less than the number of observations). Each layer has a homogeneous resistivity within a specified depth range and the model is terminated at its base by a half space (e.g. Wu, 1968; Jupp and Vozoff, 1975; Larsen, 1981; Fischer and Le Quang, 1982; Meju, 1992). The objective is to find the simplest model that fits the data best. The continuous conductivity models derived in the second category of solutions are inversion schemes that produce resistivity structure that is isotropic and a continuous function of depth (e.g. Niblett and Sayn-Wittgenstain, 1960; Parker, 1970, 1977; Oldenburg, 1979; Hobbs, 1982).

The non unique nature of the EM data implies that there is more than one model which satisfy the recorded data in the absence of a priori information. Random search or Monte Carlo methods can provide a partial solution to the problem and can explore the range of uniqueness of the solutions. The Monte Carlo method (e.g. Jones and Hutton, 1979b) generates a large number of random models around an initial set of specified parameter values. It then evaluates the forward model response for each and retains those which most closely fit the observations. Subsequently, the starting model is revised on the basis of these best fitting models and the whole procedure repeated. The method is too consumptive of computing time and such a random search can never be exhaustive (Meju and Hutton, 1992). Linearized inversion methods are now widely used in EM data analysis. The uniqueness of the final solution can be investigated by computing the resolution matrix, or by finding the most extreme parameter values which will produce a response function which is consistent with the observed. The application of such method to MT data has been described by Meju (1988) and Meju and Hutton (1992). There are other techniques for addressing the non-uniqueness inherent in the inversion of MT data (e.g. Oldenburg, 1983; Oldenburg at al., 1984; Constable et al., 1987).
Recognizing the limitations of discontinuous model parametrizations "Occam" type modelling have been proposed (Constable et al., 1987; Meju, 1988; Meju and Hutton, 1992). The aim of these methods is to generate smooth models which are devoid of the sharp discontinuities that typify conventional least squares inversion.

The resulting best fitting parametric models from both the linearized and Monte Carlo methods are strongly dependent upon the starting models. Extraneous thin layers can greatly alter the interpretation of the geoelectric structure as they can be readily added to a model with little significance in the misfit. The non-uniqueness of the inverse problem is greatly reduced with the use of a correctly specified starting model (Oldenburg, 1990). Problems in the inverse modelling are also caused by the sensitivity of the MT models to the shape of the response functions. It only takes a very small change (e.g. due to observational noise) to result in a very large effect on the estimated model parameter values (e.g. Parr, 1991). Another problem for the geoelectric interpretation of the models results from the principle of suppression of thin layers (e.g. Patra and Mallick 1980; Raiche et al., 1985; Jones, 1992). In certain situations the EM response functions cannot be used to determine the model layer thickness and resistivity separately and the only parameter that may be resolved is the layer conductance (conductivity - thickness product: $S = \sigma h$). To distinguish properly a thin conductive layer from a thicker more resistive layer, constraining information concerning the layer thickness or resistivity must be applied. Some inversion algorithms favour thin zones ($D^*$: Parker, 1980), whereas others tend towards thicker zones (e.g. Constable et al., 1987).

The non-uniqueness of a data set can be assessed using a damped most-squares inversion scheme (Meju, 1988; Meju and Hutton, 1992). The scheme uses a compensating relationship between the model parameters during extremal inversion and provides the confidence limits of the optimal least squares model. The main advantage of this method is that the errors in the data and the non-unique nature of the solution are accounted for using the class of solutions which are maximally consistent with the observations (Meju and Hutton, 1992).

2.8.1.3.3 Joint modelling of geoelectromagnetic data

Different geophysical techniques aim to define the resistivity structure with depth (e.g. MT, TEM, CSAMT, D.C. resistivity). None of these methods applied separately can unambiguously resolve the subsurface structure throughout the entire depth range for reasons mentioned already. Joint modelling of D.C. resistivity and MT (e.g. Vozoff and Jupp, 1975), D.C. resistivity and TEM (Raiche et al., 1985) or combined D.C resistivity, TEM and AMT (Meju et al., 1999) data lead to increased resolution of the model and provide a more complete description of the underlying structure.
In this study joint inversion of TEM apparent resistivity and MT impedance phase (Meju, 1996) provided the tool to correct the MT apparent resistivity curves for static shift effects (section 4.5). For the joint inversion problem, the inversion formula expands to (Meju, 1994):

\[ x = (A_t^T A_t + \beta^2 I)^{-1} (A_t^T y_t - \beta^2 m^0) \]

where

\[ y_t = \begin{bmatrix} y_{t1} \\ y_{t2} \end{bmatrix} = \begin{bmatrix} d_t - F_t(m^0) \\ d_\phi - F_\phi(m^0) \end{bmatrix} \quad \text{(dimension: } [n_t + n_\phi] \times 1) \]

and

\[ A_t = \begin{bmatrix} A_{t1} \\ A_{t2} \end{bmatrix} = \begin{bmatrix} \partial F_t(m^0) / \partial m^0 \\ \partial F_\phi(m^0) / \partial m^0 \end{bmatrix} \quad \text{(dimension: } [n_t + n_\phi] \times n_p) \]

for the \( n_t \geq 1 \) TEM apparent resistivity data, \( n_\phi \) MT phase measurements, and the \( n_p \) model parameters. The subscripts \( t \) and \( \phi \) denote the contributions to the system of equations from TEM and MT phase data, respectively. Scaling the observations by their associated uncertainties prevents the poorly estimated data from influencing the solution significantly. The weighted solution is (Meju, 1994, 1996):

\[ x = [(WA)^T(WA)_t + \beta^2 I]^{-1} \{ (WA)^T(WA)_t y_t - \beta^2 m^0 \} \]

where the diagonal weighting matrix \( W \) contains the reciprocals of the observational errors.

The above parameter corrections \( x \) are applied to \( m^0 \) in successive applications to recover the desired model parameters, \( m \). Normalising the data to some common-scale before assembling them in \( y \) and \( A_t \) overcomes the problem of different magnitudes between the two data sets (Meju, 1996). In this inversion scheme the logarithms of the TEM data are considered with the actual phase data (Meju, 1994):

\[ y_t = \{ \ln d_t - \ln F_t(m^0) \} \]

\[ A_t = \partial \{ \ln F_t(m^0) \} / \partial m^0 \]

The components of \( m \) are taken to be the logarithms of the resistivities and thickness of the subsurface model sought. An additional normalization has to be applied to the contributions from the phase data to the system of equations, since phase data vary from 0 to 90° while the transformed apparent resistivities would typically assume a range of values that is less than one-tenth of the phase data (Meju, 1996). For this study a standard MT parametric computer
modelling program MTINV (Meju 1992) has been used for the joint inversion, adapted to take account of the additional TEM observations and then to invert the TEM apparent resistivity and MT phase.

2.8.2 2-D modelling

Various 2-D modelling procedures can be applied to MT observations. Assuming invariance in one horizontal direction the induction response of a finite 2-D anomaly can be studied by solving Maxwell's equations separated into two field polarizations (TE and TM modes described in section 2.4) and subject to certain boundary conditions. The TM mode is related to the $B$-polarized wave (telluric current flows across the structures), and the TE mode is related to the $E$-polarized wave (telluric currents flows along the structures). Because of the difficulties encountered in applying boundary conditions to general 2-D problems, analytical solutions have been obtained only for simple discontinuities like faults and dikes (e.g. d'Erceville and Kunetz, 1962; Rankin, 1962; Weaver, 1963).

To study more general 2-D models, Maxwell's equations and the appropriate boundary conditions are usually discretized prior to a numerical solution. There are four main methods for the solution of 2-D induction problems: the finite difference (e.g. Patrick and Bostick, 1969; Jones and Price, 1970; Brewitt-Taylor and Weaver, 1976), the finite element (e.g. Reddy and Rankin, 1973; Wannamaker et al., 1986, 1987), the integral equation (e.g. Patra and Mallick, 1980; Hohman, 1983), and the transmission line analogy (e.g. Madden and Thompson, 1965; Madden and Swift, 1969; Swift, 1971). In all these methods the region to be modelled is divided into a mesh of elements at which field values are evaluated subject to boundary conditions. The outer boundaries of the mesh must be sufficiently distant from any vertical discontinuities that any perturbations in the field are insignificant at these points, so that the boundary conditions are satisfied. For a given spacing the finite element methods are usually the most accurate, and the finite difference methods the quickest and simplest (Madden and Mackie, 1989; Mackie et al., 1993). In this study a computer program based on the finite difference equations has been used.

2.8.2.1 Computer modelling technique

The 2-D scheme used in this study is an inversion modelling algorithm which finds regularized solutions to the 2-D inverse problem for MT data using Tikhonov (1963) regularization. The forward model simulations are computed using difference equations generated by network analogs to Maxwell's equations.
2.8.2.1.1 Finite difference forward solution

The 2-D forward program used to calculate the MT forward problem is in effect a finite difference scheme, but is based on a network formulation of Maxwell's equations (Swift, 1971; Madden, 1972). The curl operators in Maxwell's equations for a 2-D earth can be rearranged into divergence and gradient operators by making the proper analogy between current and voltage variables with electric and magnetic field components. The resulting equations (transmission surface equations) are the same as the equations for a rectangular network and can be efficiently solved for the given earth model and frequency of the inducing field.

In a transmission surface, voltage ($V$) and current ($I$) along the line are related to each other by the equations (Swift, 1971):

\[
\frac{\partial I_x}{\partial y} + \frac{\partial I_y}{\partial z} = -YV
\]

\[
\frac{\partial V}{\partial z} = -ZI_z
\]

\[
\frac{\partial V}{\partial y} = -ZI_y
\]

where $Z$ and $Y$ are the distributed impedance and admittance parameters per unit length. The analogy is clear if equations (2.86) are compared with Maxwell's equations in two dimensions (see section 2.4 reproduced here as equations (2.87) and (2.88)):

\[
\frac{\partial H_z}{\partial y} - \frac{\partial H_x}{\partial z} = \sigma E_x
\]

\[
\frac{\partial E_z}{\partial y} - \frac{\partial E_x}{\partial z} = i\omega\mu H_z
\]

\[
\frac{\partial E_z}{\partial z} = i\omega\mu H_y
\]

\[
\frac{\partial H_z}{\partial z} = \sigma E_y
\]

\[
\frac{\partial E_z}{\partial y} = -i\omega\mu H_z
\]

\[
\frac{\partial E_x}{\partial y} = -\sigma E_z
\]

Over the transmission surface, the scalar voltage is cross-coupled in equation (2.86) with two components of the current. In each of the two MT polarizations, the one field component linearly polarized in the strike direction is cross-coupled in equations (2.87) and (2.88) with two components of the other field component. $V$ plays the same role in equation (2.86) as $E_x$ and $H_x$ play in equations (2.87) and (2.88). By associating $V$ with $E_x$ (or $H_x$) and the
components of $I$ with the components of $H$ (or $E$), it can be given the distributed circuit parameters of the equivalent transmission surface can be given in terms of the appropriate earth parameters.

If the principles of energy conservation are maintained, the associations for the TE-mode case are:

$$E_x \leftrightarrow V \quad \text{with} \quad Z = -i\omega \mu, \quad Y = \sigma$$

$$H_y \leftrightarrow I_z$$

$$H_z \leftrightarrow -I_y$$

and for the TM-mode case:

$$H_x \leftrightarrow V \quad \text{with} \quad Z = \sigma, \quad Y = -i\omega \mu$$

$$E_y \leftrightarrow -I_z$$

$$E_z \leftrightarrow I_y$$

The analogy is maintained in the boundary conditions. A constant vertical current is associated with $H_y$ at the top of the air layer, which is constant in the TE-mode. A constant voltage at the surface in the TM-mode, is equivalent to $H_x$ being constant at $z=0$ and at the sides a one dimensional transmission line is solved to find the boundary values of $V$. The model is covered with a mesh of cells each electrically homogeneous and the cells are linked by Kirchoff’s laws of electrical continuity, thus ensuring a set of equations at each node point.

### 2.8.2.1.2 2-D Inversion

The 2-D inversion method used in this study is based on the maximum likelihood procedure described by Mackie et al. (1988).

The derivation of the maximum likelihood inverse can be found in Menke (1984) and Tarantola (1987). The maximum likelihood inverse gives the solution which minimizes a weighted sum of the variance of the difference between the data and the model predictions and the difference between the model parameters and the a priori information. It can be assumed that the errors between the observed data and the model predictions have Gaussian statistics, and are therefore proportional to (Mackie et al., 1988):

$$\exp\left[-\left(\mathbf{d} - \mathbf{F}(\mathbf{m})\right)^T \mathbf{R}_{dd}^{-1} \left(\mathbf{d} - \mathbf{F}(\mathbf{m})\right)\right]$$

where $\mathbf{d}$ are the observed data vector, $\mathbf{F}$ is the forward model operator which maps the model space to the data space, $\mathbf{R}_{dd}$ is the data covariance matrix and is a measure of the
experimental uncertainties in the data, and the superscript \( H \) stands for Hermitian, or complex conjugate transpose. Assuming that the errors between the model parameters and the \textit{a priori} information have Gaussian statistics, the above expression is equivalent to:

\[
\exp\left(-[(m - m^0)^H R_m^{-1} (m - m^0)]\right)
\]

where \( m \) is the model, \( m^0 \) is the \textit{a priori} model, and \( R_m \) is the model covariance matrix, and describes the uncertainties in the \textit{a priori} model. Both the model and the data covariance matrices are assumed to be real. Assuming the data errors and the model errors are uncorrelated, their joint probability function is proportional to:

\[
\exp\left(-[(d - F(m))^H R_d^{-1} (d - F(m)) + (m - m^0)^H R_m^{-1} (m - m^0)]\right)
\]

The maximum likelihood solution is that solution which maximizes the above probability function, or equivalently, minimizes the exponent.

As in the 1-D case, the nonlinear problem can be solved by linearizing the forward problem around some model \( m^0 \):

\[
F(m) = F(m^0) + A\Delta m
\]  

(2.89)

where \( m = m^0 + \Delta m \). \( A \) is the sensitivity matrix, and \( \Delta m \) are the model changes which must be determined. The terms of the sensitivity matrix describe the perturbations in the data (or sensitivity of the data) due to perturbations in the model parameters, and are of the form \( A_{ij} = \frac{\partial d_i}{\partial m_j} \). Since the problem is nonlinear, \( \Delta m \) describes only local changes, so the inversion must be iterated, each time updating the model. The model at the \( k + 1 \) step is given by: \( m_{k+1} = m_k + \Delta m_k \). The maximum likelihood solution for the nonlinear problem is obtained from (Mackie et al., 1988):

\[
(A^H R_d^{-1} A + R_m^{-1})^{-1} \Delta m_k = A^H R_d^{-1} [d - F(m_k)] + R_m^{-1} (m^0 - m_k).
\]

The interpretation of the above equation is straightforward. The model changes calculated at each step represent a compromise between fitting the data and adhering to an \textit{a priori} model (Mackie et al., 1988). The compromise is weighted by the inverse of the data and model covariance matrices.

Because of the large variations in electrical properties in the earth, it is useful to use the logarithm of the conductivity or resistivity as parameters. This also guarantees the positiveness of the parameters and allows larger changes in the model parameters as the inversion is iterated. For the electrical problem, where the resistivity can vary by several orders of
magnitude, this reduces the total number of iterations needed to reach an acceptable solution (Madden and Mackie, 1989).

A particular approach used to solve the maximum likelihood equations is the use of conjugate gradient relaxation technique (Madden and Mackie, 1989; Mackie et al., 1993; Mackie and Madden, 1993). At each iteration of the inversion, conjugate gradient relaxation is used to obtain an approximate solution for \( \Delta \mathbf{m} \), so that there is no need to do a large matrix inversion at each iteration of the inversion. Conjugate gradient relaxation is an iterative method for finding the solution to real, symmetric and positive-definite systems of equations \( \mathbf{A} \mathbf{x} = \mathbf{b} \), such as would arise in numerically solving partial differential equations. Conjugate gradient methods can be accelerated if a non-singular matrix \( \mathbf{M} \) with known inverse can be found to approximate \( \mathbf{A} \). Thus, instead of solving \( \mathbf{A} \mathbf{x} = \mathbf{b} \), an alternative set is solved \( \mathbf{M}^{-1} \mathbf{A} \mathbf{x} = \mathbf{M}^{-1} \mathbf{b} = \mathbf{c} \), which should converge much faster since \( \mathbf{A} \mathbf{M} = \mathbf{I} \). This is called "preconditioning". The simplest preconditioner is the matrix \( \mathbf{M}^{-1} = \text{diag}(1/\alpha_{11}, 1/\alpha_{22}, \ldots, 1/\alpha_{nn}) \) where \( \alpha_{ii} \) refers to the diagonal elements of the \( \mathbf{A} \) matrix.

The specific inversion program used in this study was written by Mackie (1996). In this program, conjugate gradient relaxation methods are used to solve the maximum likelihood system of equations for both the forward and inversion solution. The inversion process incorporates the Tikhonov’s method which defines a regularized solution of the inverse problem to be a model \( \mathbf{m} \) that minimizes the objective function:

\[
S(\mathbf{m}) = (\mathbf{d} - \mathbf{F}(\mathbf{m}))^T \mathbf{R}_{dd}^{-1} (\mathbf{d} - \mathbf{F}(\mathbf{m})) + \tau \| L(\mathbf{m} - \mathbf{m}^0) \|^2 
\]  

(2.90)

where \( \mathbf{d}, \mathbf{F}, \mathbf{R}_{dd}, \mathbf{m}^0 \) have been defined above, \( L \) is a linear operator and \( \tau \) is a regularization parameter. Each datum \( d_i \) is a log amplitude or phase of TE or TM complex apparent resistivity at a particular station and frequency. The model vector is a log resistivity as a function of position: \( \mathbf{m}(x) = \log \rho(x) \). The linear operator that is used is: \( L = \Delta \) (Laplacian operator), yielding:

\[
\| L(\mathbf{m} - \mathbf{m}^0) \|^2 = \int (\Delta(\mathbf{m}(x) - \mathbf{m}^0(x)))^2 \, dx \, .
\]  

(2.91)

The program uses the nonlinear conjugate gradient relaxation method to directly minimize the objective function \( S \). The model sequence is given by:

\[
\mathbf{m}_{j+1} = \mathbf{m}_j + \alpha_{j+1} \mathbf{h}_{j+1}
\]
where $h_{j+1}$ is a given search direction. The nonlinear conjugate gradient technique computes the search direction as:

$$h_{j+1} = C_j g_j + \beta_j h_j$$

where $g_j$ is the gradient of $S$ evaluated at $m_j$, and the operator $C_j$ is known as the preconditioner. In the nonlinear conjugate gradient algorithm, the preconditioner has a big impact on efficiency. The two considerations in choosing the $C_j$ are (1) how well it steers the $g_j$ in the direction of the true solution and (2) the amount of computation involved in applying the operator. The preconditioner used by the algorithm is:

$$C_j = (\gamma I + \alpha L^T L)^{-1}$$

where $\gamma$ is proportional to the matrix norm of $R_{dd}^{-1/2} A_{j}$. The $\alpha_{j+1}$ is computed to minimize the exact $S$. This involves an iterative line minimization procedure. The line minimization automatically defaults to one step computation of $\alpha_{j+1}$ when $F$ can be well approximated by its linear expansion about the previous model. Operations with $A_j$ and $A_j^T$ are computed efficiently using reciprocity of the forward problem. They require solving only two extra forward problems per frequency.
Chapter 3

The Electromagnetic Induction Survey

The measurements performed in this study, were of three types: (1) Long-Period magnetotelluric (LMT), (2) Audio-magnetotelluric (AMT), and (3) Transient Electromagnetic (TEM). The use of the two MT systems, preserve a reasonable depth of penetration (10-40 km) at most locations, with supplementary geoelectric information about structures above the minimum depth of MT investigation (100-200 m), was gained by the TEM measurements.

A description of the equipment and methods used in the acquisition of the geoelectric field observations from the Southern Kenya Rift Valley are now given, followed by information about the “K.R.I.S.P. 95 - MT experiment”.

3.1 Magnetotelluric field data acquisition

The K.R.I.S.P. 95 - MT experiment was designed to probe the crust and upper mantle in the southern part of the Kenya Rift Valley, where the well-defined narrow graben (50-70km wide) of central Kenya is replaced by a wider (100 km) asymmetric rift valley. Previous magnetovariational magnetotelluric investigations across the central part of the Kenya Rift (Banks and Ottey, 1973; Beamish, 1977; Rooney and Hutton, 1977) demonstrate low resistivities (of the order of 2-20 Ω.m) extending to depths of at least 40 km below the rift floor, as well as enhanced conductivities to the east of the rift in both the crust and upper mantle, when no anomalies were detected to the west of the rift. Using the above information, the potential maximum and minimum measurement frequencies were in range of 100 Hz~5000 sec (estimated through the skin depth equation (2.25)).

3.1.1 Audiomagnetotelluric field measurements

Audiomagnetotelluric measurements were made using the S.P.A.M. (Short Period Automatic Magnetotelluric) MKIIb system developed at the University of Edinburgh by Graham Dawes (1984). The basic S.P.A.M. field equipment is summarized in Figure 3.1. The system can simultaneously record up to a maximum of seven different sets of EM field observations. To allow the full tensor impedance to be evaluated, two orthogonal components of both the horizontal magnetic and electric fields need to be recorded. Variations in the vertical magnetic
field amplitude can be collected on a fifth channel, whilst the remaining two channels are available for locally referenced (electric and/or magnetic measurements).

The basic MT instrumentation consists of the signal detecting system (magnetic and electric sensors) and the data recording equipment.

![Diagram of S.P.A.M. magnetotelluric field equipment](image)

**Figure 3.1:** A schematic drawing depicting the S.P.A.M. magnetotelluric field equipment deployed in the field.

### 3.1.1.1 Field sensors

Magnetic fields can be measured directly by SQUID (Superconducting Quantum Interference Device) magnetometers, but more recently, the development of highly sensitive, low noise induction coils has largely replaced SQUID measurements. Magnetic field variations, in this study, are measured in the N-S, E-W and vertical magnetic components by induction coils. The induction coils in this study (Societe E.C.A. - France, CM11E) consist of a large number of fine copper wires wound around a high permeability core, and a preamplifier with low-noise connections and components to minimise signal losses prior to recording, all enclosed in a waterproof casing. The coils had low electronic noise, relatively high sensitivity (50 mV/γ), and a flat response between 0.012 and 100 Hz.
The electric ("telluric") field fluctuations were detected by measuring the potential differences induced between two grounded electrodes positioned approximately 100 m apart. Two pairs of electrodes were deployed in orthogonal directions (N-S, E-W), to measure the $E_x$ and $E_y$ field components. Numerous types of electrodes are used in geophysics, which can be divided into three groups: (a) metallic electrodes (brass and steel stakes), (b) graphite electrodes and (c) non-polarizing electrodes (copper, cadmium, silver, mercury and lead). Petieau and Dupis (1980) compared the performance of several types of electrodes and illustrated the overall superiority of non-polarizable electrodes as regards potential difference of polarization, noise and time stability, and temperatures coefficients, for low MT frequency (<10 Hz). In this study, most of the AMT measurements were performed using copper-copper-sulphate non-polarizable electrodes, while in some sites silver-silver-chloride non-polarizable electrodes were used. The Cu-CuSO$_4$ electrodes consisted of copper rods immersed in a concentrated copper sulphate and gelatine suspension contained in a ceramic porous pot. This was sealed with a perspex cap. Electrical contact with the ground was achieved through the fairly large surface area of the ceramic pot.

### 3.1.1.2 Data recording equipment

The SPAM MKIIb system is a portable battery operated, real-time data acquisition and analysis system. A detailed description of the SPAM MKII system can be found in Hutton et al., (1984) and Meju (1988). A block diagram of the complete in-field system is given in Figure 3.2.

![Block Diagram of complete in-field MT system](after Dawes, 1984)
The SPAM recording system records data digitally on a cartridge tape over the frequency range 0.012-128 Hz. The amplitudes of the EM fields vary substantially with direction, time and especially frequency. Furthermore, as high frequency EM data can be recorded much more quickly than low frequency data, the signals were not recorded simultaneously over the entire frequency band, but were instead measured in one of four overlapping bands (Band 0: 128-13 Hz, Band 1: 16-1.625 Hz, Band 2: 2-0.203 Hz, Band 3: 0.250-0.012 Hz). The gain can therefore be optimised. The main units of the system are: (i) a microcomputer (LSII 11 with a 64 kilobyte RAM), a programmable amplifier/filter bank and a power regulator; (ii) twin cartridge (programs and data) storage deck; (iii) alphanumeric pocket VDU and Keyboard; (iv) 40 column miniature electrosensitive printer for listing and plotting results; (v) sensor distribution box and (vi) a power distribution box.

The field programs are read off the program tape by the computer and any useful results written onto the other tape. The main program facilities are: software band selection, instrument response corrections, automatic gain adjustment and signal selection, with manual override, alteration of signal selection criteria, and continuous or switchable display of results, which can be listed on the printer or stored on tape.

The sensor distribution box contains the telluric preamplifiers and connections to the induction coils. The signal from each electrode pair is connected to a preamplifier with a gain of 40. The signals from the induction coils are brought to the sensor distribution box by 3 m long connecting cables. A 50-100 m individually shielded multi-conductor cable powers the "telluric" box and takes the electric and magnetic signals back to the amplifier/filter bank. The maximum allowable input range of the analogue to digital converter (ADC) is ±5V and all incoming signals are tested to see that they do not exceed this range. All the 7 channels have notch filters to eliminate frequencies of 50 and 150 Hz, the mains electricity fundamental and its first odd harmonic. Each amplifier/filter unit has automatic (or manual) gain adjustment with hardware band selection. From the amplifier/filter bank the signals are sent to the multiplexer - ADC unit where the channels are sequentially sampled and digitised. The sampling frequency is different for each frequency band. Each data set is called an "event" or "window" and consists of 256 samples for each of the 7 channels. The data window is transferred in real-time into computer memory for infield analysis and is stored if it satisfies certain present conditions. The computer then initiates the acquisition of the next window. Depending on the quality of the data, about 25-30 data windows are considered adequate for a particular frequency band. A simplified infield flowchart is shown in Figure 3.3.

To reduce the sheer volume of data recorded at period bands with high sampling rates and ensure good quality data for infield modelling, several optional acceptance criteria were set,
and windows not meeting these criteria were rejected. These criteria include a stated number of averaged frequencies within a window having a signal power in the magnetic components, as well as the multiple predicted coherency of impedance tensors, above stated levels, the phase of $Z_{xy}$ and $Z_{yx}$ being in opposite quadrants and the existence of an unpolarised source field. Acceptance criteria were normally set at 5 frequencies with a predicted coherency of at least 0.8 and a magnetic power level above 0.3 nT for the two MT impedance sets.

Infield analysis enables a very quick appraisal of the data being collected, and involves converting the data into apparent resistivities, phases and other vital information as a function of frequency.

![Simplified flowchart of the in-field system (after Dawes, 1984).](image)

Figure 3.3: Simplified flowchart of the in-field system (after Dawes, 1984).
3.1.1.3 Field procedures

Careful field procedure is probably more crucial to successful results in MT than in any other EM technique. Site selection and sensor installation are the two major factors governing data quality.

Ideally MT soundings are conducted in open flat-lying areas, clear for 60-100 metres square, away from features which might affect the fields. Powerlines, pipelines, electric fences, radio transmitters, vehicular traffic, and even pedestrian traffic can act as local field sources for which the impedance differs from the plane wave value. Metal structures, (passive) metal fences, and metal strands on wooden fenceposts can locally distort the magnetic field, or cause sensor motion when the wind blows. Natural features might affect the fields, especially the electric field. These objects include abrupt topographic features such as sinkholes and pinnacles, small salt pools, outcrop in areas that are otherwise soil covered (Vozoff, 1991).

As a general statement, the better the installation, the better the data and the less time required at the site. The actual field procedure followed in this survey in setting up an AMT site is described below.

First the sensor box was located in the centre of the sensor system. The telluric cables and the position of the electrodes was decided with a use of a tape and prismatic compass. The electrodes pairs were always placed in a (magnetic) North-South, East-West, cross-shaped array, thus ensuring common direction of measurement at all the sites. The electrodes were buried in shallow holes at the end of the telluric lines, for temperature stability and protection from any other interference, and they were kept moist at all times creating mud inside the holes. In general, an increase in separation produces an increase in potential difference between the electrode pairs, but also produces an increase in noise due to wind moving the electrode cables, therefore a compromise distance of between 70-100 m (depending on the area) was chosen. To minimize the effects of wind induced disturbances, the telluric cables were kept flush against the ground particularly if the cable traversed a bush or fence. The contact resistance between an electrode pair (N-S, or E-W) was checked before recording began, a typical good resistance being about 0.4-3 kΩ. Values greater than 10 kΩ did not give good results. The electrodes were checked to see that they were firmly planted in the ground and that the cables were properly connected.

The two induction coils, used to measure the magnetic north and east components of the horizontal magnetic field, were placed in shallow trenches, and a deeper hole (80-100 cm) was dug for the vertical coil. The horizontal coils were precisely orientated (to within 1 degree), so that the measured signals did not include a significant component of the magnetic field from the other direction. Additionally, the horizontal coils were levelled with a spirit level, so as not to
measure a component of the vertical magnetic field. The coils were connected to the sensor box with 3 m long cables, and they were always buried, to reduce vibration noise. As a precaution, the same coil was used for the same magnetic element throughout the field campaign.

The sensor box was placed 80-100 m away from the S.P.A.M. system, to avoid magnetic interference, and they were connected to each other through a multi-conductor cable. Finally, the whole system was earthed, and then tested (initial system calibration). When operation of the system appeared satisfactory, digital recording was started.

3.1.2 Long Period field measurements

Long Period data were digitally recorded to static ram using RAP 5.3 geologgers (Erich Steveling, Gottingen). The approximate period range was 40-10000 sec. Particular features of the RAP dataloggers are:

- Modular construction
- 8 Analogue channels with ± 5 V inputs
- A 16 Bit separate ADC for each channel
- Precise clock
- Data reliably saved to 8 MB SRAM
- User definable 1-60 second digitisation period
- 0.15 A current intake from external 12 V battery
- Large LCD display for operational management
- Keyboard

The magnetic sensors consisted of three fluxgate magnetometers (Magson), fixed at right angles to each other in a sealed drum. Parkinson (1983) describes the physical principles used in the design of a fluxgate magnetometer. The magnetometer drum was fixed to a spike base driven into the ground at the bottom of a hole (1m deep), and the hole was covered with a wooden board, and buried. Before covering the magnetometer, it was accurately levelled and aligned, so that the fluxgate magnetometers pointed north, east, and vertically down. Power was sent to the magnetometers along a 20 m cable and the signals sent back along the same cable, which was also placed below the surface. The fluxgate drum was placed as far away as possible (=30 m) to eliminate any magnetic interference with the rest of the equipment. All the equipment was inside a metallic box, half-buried to reduce noise from vibration and try to ensure stable temperature environment.

The electric field fluctuations were measured via 100 m, perpendicular (N-S and E-W) orientated dipoles having silver-silver-chloride electrodes buried at either end. These
electrodes are protected from damage and pollution by a cylindrical plastic chamber, with a porous base, filled with saturated potassium chloride solution. To ensure good contacts, the electrodes were totally buried in a sheath of wet clay. The cables connecting the electrodes were buried, to minimize the effects of the wind disturbances.

After amplification and compensation for background fields, the measurements were digitally recorded on static ram, with datalogger being powered by 12 V batteries recharged by solar panels. Simultaneous recordings from several sites were synchronised by the use of GPS clocks (Simpson et al, 1997).

### 3.2 Transient Electromagnetic (TEM) measurements

In order to be able to compensate for the effects of static shift, and to constrain the apparent resistivity of the uppermost kilometre, transient electromagnetic (TEM) measurements were performed. A TEM system must satisfy two conditions. It must produce a high amplitude time varying current for the transmitter, and secondly, it must be capable of detecting and storing small voltages induced in a receiver coil.

The SIROTEM MKIISE system, developed in Australia by CSIRO (Buselli and O'Neil, 1977) was used in this project, with central and coincident loop configurations. Sirotrem is a portable multichannel instrument, transmitter and receiver components are housed in the same unit and powered by two 12 V batteries. External transmitter synchronisation and dual time base facilities are also available with the system powered by a SATX-1 high power transmitter for certain configurations.

The Sirotem transmitter wave-form is a rectangular bipolar current pulse, with equal on and off times that are multiples of 10 ms for 50 Hz interference rejection and range from 30 to 180 ms. When the transmitter pulses into the loop, the full wave form shows an exponential current rise of time constant $\tau$ producing a constant current pulse, followed by linear turn off ramp $\tau_0$. Magnitudes of $\tau$ and $\tau_0$ are a function of the size and resistance of the transmitter loop. For a 100 m loop of 3.2 $\Omega$ resistance the turn-off time is $\approx 140$ $\mu$s. The maximum transmitter current is 10 Amps, and it is a function of the loop wire resistance. The mean current amplitude is used to normalise recorded voltage. The gain of the receiver can be set to 0.1, 1 or 10 depending on the voltage amplitude.

The receiver can sample the transient signal at specified times across two overlapping time-bases, which can provide information from 0.049 ms to 161.4 ms. The time-bases, called “Early Time” and “Standard Time”, are each comprised of 32 continuous channels, with
bandwidths of 10 kHz and 1.2 kHz respectively. The early time (ET) records data from 0.049-20.1 ms with closely spaced gates for shallow work. The standard time (ST) records data from 0.49-161 ms. The time channels' widths vary between 400 μs and 25.6 ms and are contiguous, so no signal is lost. Table 3.1 gives the recording times for each channel.

An internal microprocessor stacks voltage decays and normalises the transmitter current. By stacking n records the signal to noise ratio improves according to $1/n^{1/2}$. Sirotem has a high degree of ambient noise rejection capability due to its ability to stack up to $2^{11}$ (i.e., 2048) separate readings to obtain the output voltage. Thumbwheel switches allow the user to set the number of stacks between $2^8$ (i.e., 256) and $2^{11}$ (i.e., 2048) and to select the number of channels to be recorded.

<table>
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<tr>
<th>Channel</th>
<th>Early Time (msec)</th>
<th>Standard Time (msec)</th>
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<tr>
<td>2</td>
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<td>0.879</td>
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<tr>
<td>3</td>
<td>0.147</td>
<td>1.271</td>
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<tr>
<td>4</td>
<td>0.196</td>
<td>1.663</td>
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<tr>
<td>5</td>
<td>0.245</td>
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<td>6</td>
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<tr>
<td>10</td>
<td>0.711</td>
<td>5.779</td>
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<tr>
<td>11</td>
<td>0.858</td>
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<td>1.054</td>
<td>8.523</td>
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<td>32</td>
<td>20.164</td>
<td>161.403</td>
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</table>

Table 3.1: Recording times for Sirotem MKIISE, in the middle of the recording windows.

Output from the Sirotem, in the form of stacked, and normalised voltage decay values, is recorded to magnetic tape. In addition to the tape, the unit has a printer for immediate control, and also to give an error statistic for each voltage or the number of decays rejected on stack formation. This helps decide which of the transients will be discarded and whether the sounding should be repeated.

Deep exploration with transmitter currents of 1-20 Amps can be facilitated by synchronising a SATX-1 transmitter to the main Sirotem unit using highly stable crystal oscillators. This unit is used to record the turn-off time, output voltage and transmitter current, for later processing. A separate multitud receiver can be used with the system for central loop surveys. It has a fixed moment of 10000 m². It can be used to measure both the vertical and horizontal fields.
The magnetotelluric method allows a unique glimpse of the crust and upper mantle beneath rift zones via the derivation of a geoelectric section. This description of the earth is sensitive in different ways to the prevailing conditions compared to models imposed by seismic, gravity or heat flow measurements. By virtue of the distinct earth properties they yield, seismic and electromagnetic techniques offer the potential to powerfully complement each other when data are jointly interpreted.

Since 1985, a series of seismic experiments have been carried out by the Kenya Rift International Seismic Project (KRISP Working Group, 1987, 1995; Henry et al., 1990; KRISP Working Party, 1991; Prodehl et al., 1994) along and across the axis of the Kenya Rift, with the objectives of elucidating the state and structure of the crust and upper mantle in the vicinity of the rift, and gaining some insight into the forces driving continental rifting. The results are summarised by Keller et al., (1994) and Mechie et al., (1997), and illustrate the structural complexity associated with the rifting process. Whilst the seismic lines of 1985 and 1990 provide valuable information on the central and northern portions of the rift, the southern part has been left unexplored. The KRISP 94 experiment was designed to investigate the upper lithosphere across the rift in the southern Kenya, using long-range seismic and gravity measurements (Birt, 1996; Birt et al., 1997). Two profiles were selected, targeting different aspects of the rift and its relationship to the surrounding geology. The first profile was 400 km long, running from the eastern flank of the rift near Nairobi, to the Indian Ocean at Mombasa, crossing the Chyulu Hills (a Quaternary volcanic range north-east of Mt. Kilimanjaro). The second profile was 440 km long, starting from Lake Victoria in the west, crossing the rift, and terminating on the northern flanks of the Chyulu Hills.

KRISP 95-MT experiment was designed to complement the KRISP 94 seismic lines. The actual field work took place during January and February 1995. Electromagnetic data were collected at 20 sites across southern Kenya, the traverse extended from Lake Victoria in southwestern Kenya crosses the rift at the alkaline soda lake at Magadi, just north of the border with Tanzania, and terminates south of the Chyulu Hills, in the east of the rift. Sites were located to lie along the pre-existing geophysical profiles of KRISP 94, to enable a comparison with the seismic section.

At each of the 20 sites marked in Figure 3.4 and Table 3.1 all the three types of data were acquired (TEM, AMT, LMT). The 580 km long profile crosses both Archaean and Proterozoic basement, and recent volcanics and sediments associated with the rift Valley. A brief description of the geology in each station is given in Table 3.3.
Figure 3.4: Geological map of the southern Kenya, showing the locations of KRISP 95-MT observational sites.

- Holocene sediments
- Cenozoic rift volcanics
- Miocene and Pliocene sediments
- Paleozoic and Mesozoic sediments
- Proterozoic Mozambique Belt
- Archaean Nyanza Craton
- Volcano
- TEM - AMT - LMT stations

- L. Victoria
- Mt. Kenya
- Mombasa
- Indian Ocean
- TANZANIA

Temperate sediments
Cenozoic rift volcanics
Miocene and Pliocene sediments
Paleozoic and Mesozoic sediments
Proterozoic Mozambique Belt
Archaean Nyanza Craton
Volcano
TEM-AMT-LMT stations

Chapter 3: The Electromagnetic Induction Survey
Chapter 3

The Electromagnetic Induction Survey

<table>
<thead>
<tr>
<th>Station Name</th>
<th>Short Name</th>
<th>Latitude (°)</th>
<th>Longitude (°)</th>
<th>Elevation (m)</th>
<th>Geomagnetic Declination (°)</th>
</tr>
</thead>
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<td>34.2054</td>
<td>1190</td>
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<td>Keh</td>
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<td>Keek</td>
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<td>Sin</td>
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<td>1600</td>
<td>1.85</td>
</tr>
</tbody>
</table>

Table 3.2: Station locations with site names, short names, elevation and geomagnetic declination.

3.3.1 Geology along the profile

At the western end, close to Lake Victoria, the profile (as detailed by Birt, 1996) crosses outcropping Archaean intrusives and meta-sediments. These rocks are part of the Nyanza Craton, which extends west of the rift and makes up much of the elevated East African Plateau. Then the profile crosses the Migori Gold Belt, where the craton is disturbed by a series of intrusions orientated parallel to the profile, and reaches the Oloololo escarpment. The 400 m Oloololo escarpment is a striking linear fault-scarp, thought to be a surficial expression of the suture zone associated with the proterozoic collision of the Nyanza craton and the Mozambique belt (Shackleton, 1986). Smith and Mosley (1993) and Smith (1994) argue that this feature does not mark the true boundary between the units, but the Proterozoic rocks to the east have been extensively thrust over the craton margin, for as much as 100 km, so the real crustal suture lies further to the east.
### Station Name | Geology
--- | ---
Nyamanga | Precambrian - Granites.
Migori | Precambrian - Basalts.
Kehancha | Precambrian - Shales, Slates and slaty tuffs, Greywackes.
Lolkoroin | Precambrian - Metabasalts and Pillow lavas.
Oloololo | Upper Miocene - Phonolites and olivine basalts.
Keekorok | Archaean - Biotite gneisses.
Leganishu | Quaternary - Reddish brown soils, and black soils overlying Precambrian basement system rocks.
Morijo | Archaean - Granitized biotite gneisses.
Irkiba | Archaean - Feldspathic quartzites and white flaggy quartzites.
Nguruman | Tertiary - Pyroxene- and olivine- basalts.
Magadi | Lower Pleistocene - Alkali trachyte.
Singleraine | Pliocene to Miocene - basalts.
Selengei | Quaternary - Olivine basalts, with volcanic ash, breccia and agglomerate.
Chyulu North | Pleistocene - Olivine basalts, with ash, agglomerate and alluvial volcanic detritus
Chyulu West | Recent - Olivine basalts
Chyulu 2 (Range) | Holocene - Basalt flows, pyroclastics, volcanic soils
Chyulu South | Archaean - gneisses.
Mwatate | Quaternary - Soil overlying undifferentiated Precambrian basement rocks.
Lukenya | Upper Miocene - Phonolites, trachytes and olivine basalts.

**Table 3.3: Generalised geology for each station.**

The profile then drops down onto the flat plains of the Masai Mara. A blanket of Miocene phonolite lavas partially obscure the basement rocks. In the folds and thrusts of the Loita Hills, east of the Mara plains, Proterozoic rocks of the Mozambique Belt are exposed. These structures thought to record the Pan-African age collisional event between the Precambrian units (Shackleton, 1986; Smith and Mosley, 1993). Continuing eastward the Nguruman escarpment, a 2000 m major fault escarpment, marks the western boundary of the rift in southern Kenya (Bosworth, 1987; Smith, 1994). From there the profile cuts across the strike of the rift, traversing the approximately 60 km expanse of the rift floor. The elevation of the rift floor drops to 650 m above the sea level, and it is covered by a sequence of trachyte and basalt.
lavas, which are extensively faulted. The profile crosses Lake Magadi, an alkaline soda lake, to reach the more gently faulted eastern margins of the rift, rising in elevation to reach Proterozoic basement. Then the profile traverses an area of patchy sediments and reaches the Chyulu Hills range, a Quaternary volcanic field, situated about 150 km east of the Kenya Rift. The profile finishes south of the Chyulu Hills in Archaean basement next to Mwatate.

3.3.2 Field work

The precise location of the stations was constrained by the need to follow driveable roads. The average spacing of the stations was =25-30 km outside the Rift Valley, and reduced to =20 km inside the rift. Nine stations were surveyed west of the Rift Valley, three stations inside and eight stations in the east (four of them in the area of the Chyulu Hills). The elevation varied from 1100 m in the west to 2000 m on the top of the Nguruman escarpment, drops to 620 m in Magadi area, and increases from 1100 m, east of the rift, to 1500 m in the Chyulu Hills before it starts decreasing again, in Mwatate area, east of the Taita Hills (Figure 3.5). The stations were separated in two different lines. The main line (Figure 3.4), from station Nyamanga to Chyulu North, was 420 km long orientated =68 degrees west of the geographic North, and was coincident with the seismic line G. The line was targeting the structure of the Kenya Rift Valley. The second line, 150 km long, has been designed to coincide with the seismic line F, and was along the Chyulu Hills, orientated =36 degrees west of geographic north, and included the stations in the Chyulu Hills area, and the Mwatate station.

The need to choosing safe places to install the LMT stations for a period of one week resulted in some of the stations being installed next to farm houses or fenced areas so as to be protected. This resulted in some of the telluric lines being reduced to less than 100 m long, and in cultural noise being induced to the data (e.g. Nyamanga, Lukenya). Nevertheless, lack of powerlines in most of the countryside, assisted noise free data to be acquired in the majority of the sites. The main problem during the campaign was the bad condition of the roads, lack of accommodation close to the sites, and high temperatures during the day. Inside the Rift Valley high temperatures often caused malfunction in the operation of the S.P.A.M. system. A lot of stations had to located inside National Parks, so the measurements had to be made with the presence of a ranger, during the day-time. At some of the stations, the crew could camp next to the instrument so measurements during the night were facilitated, particular with the AMT readings.
Figure 3.5: a) Digital elevation model along the profile. b) Topographic profile along the line (elevation in metres above sea level, and distance in km).
The LMT systems were deployed in three phases, progressing eastwards across the rift, each system was laid in the field for a period of about 5-7 days. A crew of 3-4 people needed about 3 hours to install each station, since all the cables and the box with the instrument had to be buried. Each station was serviced once during the recording time, downloading the data to a PC, checking the solar panels, and the condition of the batteries and electrodes. For logistical reasons Lukenya was chosen as a base station. Instrumentation was deployed there for the entire campaign period.

For the AMT measurements, a crew of 3 people was needed to install the S.P.A.M. system, and a typical set-up time was 40-60 minutes. Measurements were made in a day depending on the signal activity. Where possible, the same telluric lines and electrodes were used for LMT and AMT readings for better data compatibility match. As the electromagnetic signal and noise strengths over the frequency spectrum can be strongly dependent upon the time of the day, it was common practice to alternate the frequency bands in which measurements were made. For band 3 with the lowest sampling rate the better time was during early morning or late afternoon, also in some stations measurements were acquired during the night. Normally 60 windows for the first three period bands were recorded. For band 3, a number of 30 windows were considered adequate, although in some sites 100 windows were recorded during the night.

Most of the surveys employed TEM loops of dimension 100x100 m, and where it was not possible (limited space) the loop size was 50x50 m, trying always to locate the centre of the loop in the centre of the MT sensor system to ensure that both methods were sampling the same geology. In all the sites the Sirotem system was used in two field configurations (central and coincident loop). Time limitations in some sites, forced the crews to shift the TEM loop 50-60 m away from the centre of the MT system, so the same time that TEM readings were taken another crew could install the MT station. Two different cables were used for the transmitter loop, one with low resistance for the 100 m loop, and a high resistance cable for the 50 m loop. A typical turn-off time for the 100 m loop was 150 μsec, and 60 μsec for the 50 m loop, the average current was about 6.2 and 5.2 Amps, respectively. The TEM readings were performed in about 40 minutes while the MT instruments were not recording.

Of the 20 sites occupied, 19 produced reliable data, in the Kajiado area the data were contaminated by cultural noise. LMT data were not recorded in Lolkoroin station, and in Leganishu only coincident loop TEM data were acquired. A major problem with the S.P.A.M. sensor box (broken capacitor and resistor burn-out) forced the team to abandon a site west of Mwatate (Kedai station).
Chapter 4

Data Processing and Analysis

The purpose of data analysis is to extract reliable values of impedances, apparent resistivities, and the other earth response functions from the field records, and present them in a form convenient for interpretation.

This chapter describes the procedure adopted in processing the AMT field records to yield the response functions needed for qualitative and quantitative studies. It presents the long period data and the determined induction arrows. It then proceeds to evaluate the use of the Groom and Bailey (1989) decomposition method for this set of MT data and the selection of the regional strike direction. Finally, the TEM data are presented, and their use to remove the static shift effects from the AMT data is illustrated.

4.1 AMT Data processing

The data processing of the magnetotelluric soundings consisted of classical tensorial estimation. The raw field data were originally stored on floppy disks and then transferred to UNIX workstations (SUN systems) for processing. The time series data were recorded in 4 overlapping frequency bands (Table 4.1). In Figure 4.1 are shown the time series for a window in each band. The telluric channels are scaled to mV/km and the magnetic to nT. A good correlation between channels $H_x, E_y$ and $H_y, E_x$ can be seen (e.g. Band 1, 500 - 800 msec). The vertical magnetic field information appears noisy (Band 3) and scanty (Band 0, 1) and was abandoned from the interpretation process.

<table>
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<th>sampling rate</th>
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<td>512 Hz</td>
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<tr>
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<td>2 Hz</td>
<td>64 Hz</td>
</tr>
<tr>
<td>Band 2:</td>
<td>2 Hz</td>
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<td>8 Hz</td>
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<tr>
<td>Band 3:</td>
<td>4 sec</td>
<td>32 sec</td>
<td>1 sec</td>
</tr>
</tbody>
</table>

Table 4.1: The four frequency bands of S.P.A.M. Mk1lb.

The analysis was done using the data analysis program AMTANAL originally written by Rooney (1977), and improved for computational efficiency by Dawes (1984). The tensor elements were determined for the upward and downward biased case (Sims et al., 1971) and a weighted
average of these two was adopted for the final impedance tensor estimation (see section 4.1.2, equation 4.7).

Figure 4.1: Time series of band 0 (128-16 Hz), 1 (16-2 Hz), 2 (2 Hz-4 sec) and 3 (4 sec-32 sec), in site Oloololo. The telluric fields are scaled to (mV/km), the magnetic fields to (nT).
4.1.1 General window analysis

The data were reprocessed in the frequency domain; the standard window analysis was as follows:

(i). Time series were scaled by using the recording gains. Instrument correction tables were then set up.

(ii). Signals were detrended and tapered by using a least squares method (Bendat and Piersol, 1971), and a cosine bell window (Harris, 1978) respectively.

(iii). The data were Fast Fourier transformed (FFT).

(iv). The Fourier transform coefficients were corrected for the instrument's (telluric preamplifier and CM11E induction coils) responses.

(v). The auto- and cross-power spectral estimates were calculated and band averaged over 8 neighbouring frequencies to reduce the variance of the estimates.

(vi). Multiple predicted coherencies were computed for overcoming the bias of responses by noise on the electric components. Minimum multiple predicted coherency was set at 0.8.

These procedures were executed for each recorded data window in turn.

In the first step of the time series analysis, after the initial checks and amplifier gain corrections, there were a series of filter and window rejection options. Details of all the rejection procedures are given in Fontes et al. (1988). Notch and delay line filters were built into the MT data analysis program. These filters are routinely used in geophysical data analysis, and the reader is referred to Kanasewich (1981, pp. 247-252) for a description of notch filtering. The delay line filter acts upon a fundamental frequency and all the higher harmonics up to the Nyquist frequency. The impulse response of the filter is given by (Fontes et al., 1988):

\[ W(z) = 1 - z^n \]  \hspace{1cm} (4.1)

where

\[ z = e^{i2\pi/n} \hspace{1cm} j = 0,1,2,\ldots,(n-1) \]  \hspace{1cm} (4.2)

\( n \) is the number of digital time units by which an input signal is delayed and is given by:

\[ n = \frac{\text{sampling frequency}}{\text{perturbation frequency}} \]

\( z_0 \) corresponds to the fundamental frequency to be filtered, and \( z_1,\ldots,z_n \) are the higher harmonics. The delay line filter had an obvious application in suppressing 50 Hz noise and its higher harmonics produced by power lines. Although most of the stations were away from power lines, for the few stations that were deployed close to urban areas the implementation of the delay line filter resulted in a remarkable improvement in the quality of the data. In Figure 4.2 are shown the time series for a
window of Band 0 (high frequency band) before and after the application of the delay line filter in Lukenya station. In Figure 4.3 are shown the results of analysis of Lukenya site after the application of a simple subtractive delay line filter.

**Figure 4.2:** Time series for a data window in the high frequency band (Band 0) in site Lukenya, before (left-hand site) and after (right hand site) the application of a delay-line filter. The site located close to a fence and a power line, the time series are dominated by strong sinusoidal 100 Hz oscillations. After the application of the delay-line filter for the 50 and 100 Hz frequencies the signal has dramatically improved, and the expected correlation between channels $H_{x}, E_{y}$ and $H_{y}, E_{x}$ can be seen.
Figure 4.3: Results of data processing in site Lukenya, before and after application of delay-line filtering (50-100 Hz). The apparent resistivity curves have been improved, while the impedance phase has not been affected.

4.1.2 Calculation of Average Response Functions

In the frequency domain analysis procedure implemented in program AMANAL, the impedance tensor elements were determined for the upward and downward case (Sims et al., 1971; section 2.5.4). In the first case noise was assumed to be present only in the two magnetic channels while in the second case noise was assumed to be present only in the telluric
channels. An example of the estimation of the off-diagonal tensor element $Z_{xy}$ is given for the two cases of biasing:

**Electric component noise free - Upward bias**

$$Z_{sy} = \frac{(H_x E_x^*) (H_x E_x^*) - (H_x H_x^*) (E_x E_x^*)}{(H_x E_x^*) (H_x E_x^*) - (H_x H_x^*) (H_x E_x^*)}$$  \hspace{1cm} (4.3)

$$Z_{sy} = \frac{(H_y E_y^*) (H_y E_y^*) - (H_y H_y^*) (E_y E_y^*)}{(H_y E_y^*) (H_y E_y^*) - (H_y H_y^*) (H_y E_y^*)}$$  \hspace{1cm} (4.4)

**Magnetic component noise free - Downward bias**

$$Z_{sy} = \frac{(H_x E_x^*) (H_x E_x^*) - (H_x H_x^*) (E_x E_x^*)}{(H_x E_x^*) (H_x E_x^*) - (H_x H_x^*) (H_x E_x^*)}$$  \hspace{1cm} (4.5)

$$Z_{sy} = \frac{(H_y E_y^*) (H_y E_y^*) - (H_y H_y^*) (E_y E_y^*)}{(H_y E_y^*) (H_y E_y^*) - (H_y H_y^*) (H_y E_y^*)}$$  \hspace{1cm} (4.6)

where $*$ denotes conjugates and $(A^*_j)$ represents auto-power spectra if $A = B$ and $i = j$ or cross-power spectra if $A \neq B$ and $i = j$ or $i \neq j$. The effect of noise on the data was to elevate or decrease the apparent resistivity estimates. Equations (4.3) and (4.4) would yield upward shifted apparent resistivity curves while equations (4.5) and (4.6) produce downward shifted curves (Sims et al., 1971). The bias range was taken to characterise the bias in the data. Less than half a decade bias range was found in all the stations across the profile, as shown by the example from one site in Figure 4.4.

The impedance tensor elements $Z_{xx}, Z_{xy}, Z_{yx}, Z_{yy}$ for each frequency for each kind of biasing were computed by using similar expressions to (4.3) - (4.6), and were averaged for the whole number of accepted windows assuming a log-normal distribution (Bentley, 1973). Accepted windows were required to satisfy the minimum power, minimum coherence and minimum number of frequencies criteria. Two other conditions were imposed, concerning the sign and outlier rejection criteria. It was required for each frequency set to possess a preferred sign of the off diagonal elements of the impedance tensor. All estimates lying outside ±2.2 standard deviations from the mean value were rejected. A new mean and a new standard deviation were calculated for each biased case.

A weighted average of the upward and downward averages was computed using the equation (Jones, 1977):
Chapter 4

Data Processing and Analysis

\[ Z_u = \frac{Z_u}{(e_u)^2} + \frac{Z_d}{(e_d)^2} \]

where \( Z_u, Z_u, Z_d \) are impedance tensor elements with average up and down biases, respectively, and \( e_u, e_d \) are the errors on the impedance tensor elements with up and down biases.

**Station: Leganishu**

![Graphs showing response function estimates for E and H noise-free channels at Leganishu station.](image)

**Figure 4.4:** Example of bias in response function estimates assuming noise-free E or H channels, in Leganishu station. The bias appears to increase in Band 3, which was the noisiest band for all the stations, compared to high frequency Band 0.
Figure 4.5: MT response curves for a station west of the Rift Valley. Figures up left and centre are the apparent resistivity and phase curves of the off diagonal elements of the unrotated impedance tensor. The curves plotted are those of the average of the up and down biased estimates. The error bars represent one standard deviation. Figures down left and centre are the apparent resistivity and phase calculated after rotation to the principal direction (major and minor apparent resistivity and phase) using the methods of Swift (1967). In the right hand site is plotted the regional azimuth of the major impedance (measured positive clockwise from magnetic north), the impedance skew (Swift, 1967), predicted coherence, and the number of estimates averaged.
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Unrotated XY

Unrotated YX

Station: Morijo

Frequency (Hz)  Frequency (Hz)  Frequency (Hz)


Phase (%)  Phase (%)  Phase (%)

Major

Minor

Num of Estimates  Coherency  Slow  Azimuth (°)

0.5  0.0  1.0  0.5  0.0  1.0

0.0  0.5  0.0  0.0  45  90

-90  -10  -90  -10  0  0

Frequency (Hz)  Frequency (Hz)  Frequency (Hz)
Figure 4.6: MT response curves for Nguruman site, a site located on the western margin of the Rift Valley. The data types and symbols are the same as for Figure 4.5.
Several parameters were then computed as a function of frequency. They include the apparent resistivities and phases (unrotated, major and minor - see section 2.5.2), the azimuth of the principal impedance axis, and the impedance skew values. Typical examples of the magnetotelluric data for stations Morijo (west of the Rift Valley) and Nguruman (inside the Rift) are given in Figure 4.5 and Figure 4.6 respectively. They illustrate the frequency variation of the apparent resistivity and phase responses along the measuring and principal directions, the coherence, the number of estimates, the impedance skew and the azimuth of the major apparent resistivities. For the apparent resistivities and phases, the curves plotted are those of the average of the up and down biased estimates calculated using equation (4.7). The error bars assigned to the parameter estimates are ±1 standard deviation about the mean assuming a normal distribution (lognormal for those on log scale). The phase of $Z_{xy}$ has been reflected from the third quadrant to the first quadrant. The regional azimuth is measured positive clockwise from magnetic north. The coherency is a multiple predicted coherency and is the average of all estimates at a particular frequency. The average includes those estimates rejected because they had a coherence below the minimum value, hence the sometimes very low values. The mean tensor response functions from all the sites are presented in Appendix A. The apparent resistivity and phase plotted in Figure 4.7 is derived from the effective impedance, which is the rotationally invariant average given by Berdichevsky and Dmitriev (1976), (see section 2.5.2, equations (2.44)-(2.45)).

![Figure 4.7: Invariant apparent resistivity and phase data, for stations Morijo and Nguruman.](image)
4.2 Long Period Data

The Long Period data have been processed (Simpson et al., 1997) and transfer functions have been calculated in the approximate period range 20 sec to 10000 sec. The sampling interval was 10 sec. The author was provided with the impedance tensor elements and the magnetic response functions as part of the collaborative KRISP 95 agreement.

The AMT and LMT data overlap at ~30 sec. Examples of combined AMT and LMT data for two stations are presented in Figures 4.8 and 4.9. They show the apparent resistivity and phase responses along the measuring axes, the coherence, the impedance skew, the azimuth (Swift, 1967), and the number of estimates. The figures suggest that the apparent resistivity curves and phases are slightly shifted relative to each other. The gap is not always well defined though, as the AMT data are noisy towards the longer periods. The quality of the LMT data at the short period end (20 sec - 30 sec) is also doubtful. The data were gathered using different types of instruments (Section 3.1.2) and data processing procedures. The shifts in the apparent resistivity curves are most likely to be caused by miscalibration between the two measurement MT systems, but high noise levels in the AMT data, may also cause some differences in the resistivity estimates. Low coherencies over the instrumentation boundaries indicate that the data quality is reduced over the frequency range from 0.1 Hz to 0.01 Hz in the AMT data.

<table>
<thead>
<tr>
<th>convention</th>
<th>Real arrow</th>
<th>Imaginary arrow</th>
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</thead>
<tbody>
<tr>
<td></td>
<td>Magnitude</td>
<td>Azimuth</td>
</tr>
<tr>
<td>Parkinson</td>
<td>$\sqrt{A_R^2 + B_R^2}$</td>
<td>[\arctan\left(-\frac{B_R}{A_R}\right)]</td>
</tr>
<tr>
<td>Wiese</td>
<td>$\arctan\left(\frac{B_R}{A_R}\right)$</td>
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</tr>
</tbody>
</table>

Table 4.2: The definition of induction arrows in Parkinson (1959) and Wiese (1962) conventions.

4.2.1 Computation of Induction arrows

Measurements of the vertical magnetic field variations enabled computations of the induction arrows. Following the standard conventions of (Hobbs, 1992) the vertical magnetic field is related to the horizontal magnetic field components in the frequency domain and at a single station by:

$$Z(\omega) = A(\omega)X(\omega) + B(\omega)Y(\omega)$$ (4.8)
where $X$, $Y$, and $Z$ are the Fourier transforms of magnetic field variations in the north, east and downward directions respectively. $A$ and $B$ are complex:

$$A = A_r + iA_i$$
$$B = B_r + iB_i$$

The combination $(A, B)$ forms a two dimensional vector called the magnetic response function. The real and imaginary parts can be combined to form two induction arrows. Two definitions of induction arrows are common. In the Parkinson convention (Parkinson, 1959) the arrows tend to point towards areas of higher conductivity\(^1\) (increased current density), whereas the Wiese arrows (Wiese, 1962) tend to point away from areas of higher conductivity. Table 4.2 gives the definitions of the induction vectors.

The visual display of the induction arrows in maps is a comprehensive presentation of changes in vertical magnetic field anomalies, both as a function of frequency and location. For two dimensional conductivity structures, real and imaginary arrows point either parallel or anti-parallel to each other. If the induction anomaly is not in the range of the skin depth, both induction arrows are zero. The imaginary arrows point in opposite directions to the real arrows if the induction anomaly is within the range of the skin depth. An anomaly at the surface, which is small in comparison to the skin depth, causes the imaginary arrows to point in the direction of the real arrows. The largest real induction arrows are observed close to a conductivity contrast; they become smaller with increasing distance from the boundary.

Summarizing, the imaginary components are generally more variable, being more sensitive to shallow induced features, whilst the real components are sensitive to more regional features. Although vertical magnetic field response functions do not contain direct information about the underlying conductivity structure ($\sigma = \sigma(z)$), they detect the lateral extension of conductivity anomalies, and they are indicative of lateral conductivity contrasts.

\(^1\)When measurements are taken near the sea and if the conductivity of the earth is much smaller than that of the seawater, then considerable distortion of the Parkinson induction arrows is expected (Coast effect).
Figure 4.8: Sample of combined AMT and LMT apparent resistivity and phases curves at Irkiba, at the western boundary of the Rift Valley. The Swift rotation angle, impedance skew, coherency and the number of estimates are also presented.
Figure 4.9: Combined AMT and LMT data from Nguruman in the west margin of the Rift Valley. The annotation is similar to that of Figure 4.8.

Induction arrows calculated from the AMT data appear noisy and scattered especially in the low frequencies while induction arrows computed from the long period data appear more consistent and stable. Figure 4.10 illustrates the induction vectors computed from the AMT frequency range at two stations with high quality vertical magnetic component data. For completeness Figure 4.11 portrays induction arrows for the same stations, computed from the long period data, collected using a fluxgate magnetometer. The good match in the overlapping frequency range is noticeable. The induction arrows computed take the Parkinson convention of \((-A_x, -B_x)\). Figure 4.12 shows the induction arrow maps for the periods 40 and 1000 sec. For the 40 sec period, the induction arrows at the stations close to the western boundary of the Rift Valley (Nguruman Escarpment) and inside the Rift point in an almost perpendicular sense to
the N-S trending rift. For the 1000 sec period, the perpendicular conductivity boundary lies in an almost NW-SE direction, rather than parallel to the N-S trending rift axis. In the Chyulu Hills area, the induction arrows are relatively small and unstable, indicating a possible underlying conductivity anomaly. The induction arrows from all the stations are presented in Appendix B.

Figure 4.10: Parkinson Induction arrows computed from the AMT data recorded at Irkiba and Nguruman.
Figure 4.11: Parkinson Induction arrows computed from the long period data at Irkiba and Nguruman. Notice the good match between the AMT (Figure 4.10) and the LMT induction arrows in the overlapping frequency range.
Figure 4.12: Real component Parkinson induction arrows for (a) 40 sec, and (b) 1000 sec, superimposed on the regional geology.
4.3 Application of Tensor Decomposition techniques

The real earth is complex and so channelled currents, near surface conductivity variations, topography, and higher dimensional structures can affect MT response functions.

The presence of galvanic distortion (section 2.6) in magnetotelluric measurements, and its effects can possibly lead to a serious misinterpretation of the measured impedance tensor, so it is vital to separate the galvanic effects from the inductive response within the tensor. As described previously (section 2.6.2.2), Bahr (1988) and Groom and Bailey (1989), have developed similar methods based on decomposing the impedance tensor under a physically realisable model and attempt to separate regional and local information. The main assumptions for their model is that any inductive response is restricted to one- or two-dimensional bodies, with an overlying inductively small body, and that the distortion of the telluric and the magnetic fields by inductively weak bodies is negligible. This technique has two serious limitations (Bahr, 1991): (i) the conductivity structure might be less complex than assumed in the general model and therefore irrelevant model parameters are derived; (ii) the regional conductivity structure may be more complicated than indicated by a two dimensional model. For these reasons Bahr (1991) suggest that this decomposition method can only be reliably employed if the data satisfy certain conditions. These are distributed between seven classes of general model of increasing complexity.

In order to evaluate the necessity for a Groom and Bailey (1989) tensor decomposition of the Kenya data sets the data collected at each station were analysed using the methods described by Bahr (1991) (see section 2.6.2.2, page 2.27). From this analysis, data at each frequency were grouped into one of the seven sub-classes. Table 4.3 gives a summary of this analysis. A number of points relating to the nature of the induction problem arise from this analysis. The data recorded at many sites show characteristics of weak local distortion (class 3), many sites fall into class 1 in the higher frequency bands, and the tensor can be described with the conventional method. The data measured at a number of sites and particularly in the long periods, fall into class 7, with skew values $n > 0.3$ (equation 2.69) suggesting that $Z$ describes a regional 3-D anomaly, for which the superimposition model is not appropriate. Nevertheless, if the decomposition method yields a regional strike that is coincident with the regional strike of neighbouring sites, then the regional structure can be considered to be approximately 2-D (Groom and Bailey, 1989; Bahr, 1991).

In order to demonstrate the decomposition technique based on the physical model of Groom-Bailey, two sounding data from Leganishu and Irkiba stations are shown in Figure 4.13 and Figure 4.14 respectively. The indication that the $Z$ has been distorted (class 3) and that the stations are located close to the investigating structure (Rift Valley), were the reasons for
their selection. The data are presented in the form of apparent resistivity and phase impedance. The observed data have been plotted together with the predicted data, and the major and minor principal response elements. The lower part of the figure illustrates the results of unconstrained decomposition, while in the upper diagram the regional strike direction has been constrained to 0° (more detailed explanation of the selection of the strike angle will be presented later). To simplify the figures, error bars have not been plotted.

<table>
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<th>Stations</th>
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<th>LMT data</th>
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<td>Band 1 16-2 Hz</td>
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<td>Singleraine</td>
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<td>7</td>
</tr>
<tr>
<td>Chyulu West</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Chyulu 2 (Range)</td>
<td>1</td>
<td>-</td>
</tr>
<tr>
<td>Chyulu South</td>
<td>7</td>
<td>1</td>
</tr>
<tr>
<td>Mwaiate</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>Lukenya</td>
<td>3</td>
<td>1</td>
</tr>
</tbody>
</table>

Table 4.3: Summary of results of analysis of the Kenya data set using the method of Bahr (1991) for seven classes of telluric distortion.

An important aspect of applying the Groom-Bailey physical model to real data is the computation of some measure of misfit to appraise quantitatively the appropriateness of the model. A useful statistical test of how well the ideal distortion model is significantly in error is the $\chi^2$ - like misfit variable. This is a chi-squared residual error of fit normalised using the
square of the variances ($\sigma^2_{ij}$) of the four elements of the measured impedance tensor. It is given by (Groom et al., 1993):

\[ \gamma^2 = \frac{1}{4} \sum_{ij} \sum_{ij} \frac{(\hat{Z}_{ij} - Z_{ij})^2}{\sigma^2_{ij}} \]  

(4.10)

where $\hat{Z}_{ij}$ and $Z_{ij}$ are the modelled and measured tensor elements, respectively. If a model is appropriate it is reasonable to presume that it will fit the data to within three standard deviations, hence $\gamma^2$ will be between 0-9. If none of the simplified earth models fit within these levels this could imply that not all physical effects have been included (i.e. 3-D induction). This error parameter has been plotted in the figures.

At Leganishu station differences between estimated and observed data, as well as between principal response elements and observed data, are negligible for the apparent resistivity and YX impedance phase. The differences become more significant, but still small, and inside the error limits only for the XY phase. In addition, no obvious improvement in data quality is observed with the decomposition technique. The reason for this may be that the method requires certain conditions such as strong 2-D host and weak 3-D scatter source, as well as that the parameters recovered by the decomposition can be unstable in the presence of noise (Groom and Bailey, 1991). At Irkiba station, the differences are still small for the apparent resistivity curves and the YX impedance phase, but the decomposed XY phase changed noticeably. To further weigh the requirement of Groom and Bailey decomposition, the root mean square relative error of fit $\epsilon$ has been calculated (Groom and Bailey, 1989):

\[ \epsilon^2 = \frac{\sum_{ij} \sum_{ij} (\hat{Z}_{ij} - Z_{ij})^2}{\sum_{ij} \sum_{ij} |Z_{ij}|^2} \]  

(4.11)

where $\hat{Z}_{ij}$ are the modelled data, and $Z_{ij}$ are the measured data. The rms error of fit from the conventional method and the Groom-Bailey decomposition are compared in Figure 4.15 for

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2 There are three simplified earth models (Groom et al., 1993): (i) The one dimensional earth, in which the conductivity varies only with depth, (ii) the conventional two dimensional model, in which the conductivity varies with depth and in one lateral direction, and (iii) a model in which a small, inductively weak, 3-D anomaly lies above a 2-D earth.
unconstrained strike direction. Both stations display relatively small error differences between the two methods.

Figure 4.13: MT parameters obtained by using the procedure of Groom and Bailey (1989) for removing the near-surface distortion effect, in site Leganishu. Apparent resistivities and phases of the measured impedance (circles) have been plotted together with the major and minor principal response elements (x) and the predicted data by the decomposition (+) for unconstrained regional strike direction, and constrained regional direction to 0°. The regional strike from the unconstrained decomposition (circles) has been presented with the conventional strike direction (stars). The distortion parameters twist (circles) and shear (stars) have been presented for the unconstrained model. The chi-squared normalised residual error of fit is given for unconstrained (circles) and constrained (triangles) decomposition.
Station: Leganishu

- Measured data
- : Major and minor principal response elements
+ : Predicted data by the decomposition

Distortion Parameters
- Twist
- Shear

Azimuth
- Regional (Swift)
- Regional (Groom-Bailey)
Figure 4.14: Application of Groom-Bailey decomposition method in site Irkiba. The regional strike is constrained to 0°. The symbols are similar to Figure 4.13. Strong differences are observed between the unrotated decomposed and measured impedance phase of XY direction. Complex geology, and topographic effects making the decomposition questionable.
The above analysis and controls was performed with the measurements recorded at each of the 20 stations occupied. For all the stations (except Irkiba site), the differences between observed and decomposed data were insignificant and limited. In the light of these results, it has been decided that the decomposition process was not necessary for this study. Another reason not to apply Groom-Bailey decomposition was that, a significant number of frequencies at many stations described a regional 3-D anomaly (class 7), and the superimposition model was no longer valid. For the complex environment of station Irkiba, it was decided not to apply the Groom-Bailey decomposition. The presence of topographic effects (section 4.6), static shift (the physical decomposition does not solve the problem of static shift), and a possible 3-D regional structure, make the physical model of Groom-Bailey unwarranted, and could lead to erroneous results. Moreover, a priori TEM information can help to recover a more reliable geoelectric structure.

![Graph](image)

**Figure 4.15:** Frequency dependency of the RMS error of fit, as defined in equation (4.11), for sites Leganishu and Irkiba. The triangles represent the conventional technique (Swift, 1967) and the circles the unconstrained Groom and Bailey (1989) decomposition. In station Leganishu Groom and Bailey model appears slightly better than the conventional, while in Irkiba the conventional technique gives better results.

### 4.4 Strike Direction

Defining regional trend is an important objective for all magnetotelluric surveys. Incorrect determination of the strike azimuth may result in estimates of $Z_{2D}$ which are not rotated to the correct axes with possible mode mixing of the TE and TM responses. There have been many approaches to solve the problem. Swift's (1967) algorithm is the most common one, but local distortions and noise can cause meaningless angles to be resolved, and new methods have
been developed to obtain the correct strike angle of the 2-D regional structure in the presence of local distortion (Bahr, 1988; Groom and Bailey, 1989).

The MT profile in Kenya was designed to follow the seismic line of KRISP. The profile crosses the Oloololo escarpment and the Rift Valley at an angle of ≈N70°W, and further east becomes approximately parallel to the NW-SE striking chain of the Chyulu Hills. It is clear that the number of lateral and vertical conductivity contrasts that the profile crosses in this 600 km will affect the data set. The effects of these contrasts not only combine with each other but also with the effects associated with the conductivity depth profile and any conductivity anisotropy that may exist. Thus, these soundings are likely to yield a complex pattern of regional azimuths. Moreover, the estimation becomes more enigmatic when three dimensionality enters, and local 3-D distortions and noise are present.

Figures 4.16 and 4.17 shows the regional azimuths that were obtained using the conventional Swift method, and the unconstrained Groom-Bailey decomposition, at each station, respectively. Characteristic frequencies for each AMT frequency band and four different periods from the long period data have been selected to represent distinct depths of penetration. At the high frequencies, the azimuth depends on the topography or any electrical anisotropy caused by a local body within the top few kilometres. At lower frequencies, when the skin depth is more than three times the scale length of any local body (so that the body becomes inductively weak), the azimuth represents the dominant regional structure at depth. By definition, the conventional strike direction has a ±90° uncertainty. When the site is on the resistive side of a two-dimensional contact, the regional azimuth of the major apparent resistivity is perpendicular to this contact. The difference in angle between the regional azimuth obtained by the two methods can be regarded as an indicator of the degree of the galvanic distortion. In order of increasing distance from the west end of the profile, four zones of contrasting azimuths can be identified. These are described next.

**Lake Victoria to Oloololo Escarpment**

High frequencies (34-7 Hz) indicate a general strike trend of N50°W. Differences of about ±90° between Swift and Groom-Bailey are explained by the ambiguity of the conventional method. These rotation angles correlate with the intrusions in the Nyanza Craton, which run parallel to the profile (≈N60°W) and form the host rocks of the Migori Gold Belt. Just west of the Oloololo Escarpment, at Lolkoroin station, the Swift’s azimuth trace the Miocene phonolite lava, presumed to have flowed Southwest away from the topographic uplift of the central Kenya Dome (Smith, 1994). In the same site, the Groom-Bailey angle appears less affected by this local feature. At lower frequencies Swift’s angles become unsteady compared to Groom-Bailey, with approximately N-S direction (±15°), suggesting an underlying linear regional grain.
**Oloololo Escarpment to Nguruman Escarpment**

The influence of the N-S trending Rift Valley axis on the geoelectric structure of the area becomes more intense in this part of the profile. The regional strike direction rotates =N-S close to the Nguruman Escarpment which marks the west boundary of the Rift.

At Oloololo station, the conventional method marks a strike direction approximately N-S (uncertainty of 90° rotates the Swift azimuth to a N80°W direction). Groom-Bailey regional strike defines a N30°W direction for the lower frequencies. Keekorok and Leganishu sites show a N40°W strike. Approaching the Nguruman Escarpment, the azimuth swings to an almost N-S direction (±10°) across the whole frequency range. For the Swift determination, the expected 90° rotation is present at Irkiba site, indicating the passing from a resistive to a conductive area. Groom-Bailey regional strike appears noisier; in Morijo the strike shows a scatter of around ±20° from the North, while at Irkiba, it rotates steadily from N30°W to N-S at frequencies less than 0.1 Hz.

**Rift Valley and sites to the East**

High conductivities (=5-20 Ω.m) on the floor of the Rift reduce the penetration depth at each frequency, in contrast with what obtains at stations west of the Rift which have a more resistive property (=1000 Ω.m).

The site located on the foot of the Nguruman Escarpment defines a N-S strike direction with very small variations (in both methods), but the Groom-Bailey technique appears more irregular at the high frequencies. For the other two stations inside the Rift (Magadi and Singleraine), the azimuth at higher frequencies appears noisy and inconsistent. It looks like it tracks the direction of the faulted trachyte and basalt lavas that the rift floor is covered with, whilst apparent resistivity and phases curves suggest that one-dimensional structure could explain the instability. At lower frequencies the azimuth angle rotates to a more steady N-S direction (±10°).

East of the Rift, at the Lukenya site the strike direction swing from N30°W to N-S at 1 Hz frequency, and to N40°E at lower frequencies; additionally, the values of the skew rise from 0.2 to 0.6 indicating a strong three dimensional effect. The Selengei site appears more erratic at high frequencies but at lower frequencies an azimuth of N30°W is indicated by Groom-Bailey, while the Swift angle rotates from N60°W in the high frequencies to ≈N15°E for the lower frequencies.

**Chyulu Hills area and sites to the South**

In the Chyulu Hills area the regional azimuths generally follow a direction parallel or perpendicular to the NW-SE striking direction of the basaltic cones, rotating to a ≈N30°E direction at lower frequencies. The strike angle seems to trace the presence of a major shear
zone within the Pan-African Mozambique belt (Smith and Mosley, 1993) running almost parallel to Chyulu Hills (see Figure 1.5). One-dimensional shallow conditions (i.e. layered volcanics) and underlying three-dimensional structure affect the consistency of the azimuth.

It is obvious from Figures 4.16 and 4.17 that for most of the sites along the profile there is a marked change in the regional azimuth. This change occurs at periods where the skin depth is such that the presence of the Rift is affecting the measurements, even for sites located more than 200 km away from it. Long period data west of the Rift may indicate a drift to the geoelectric strike direction from N-S at latitude 2° south (close to Nguruman Escarpment) to NW-SE further north. Approaching the rift and inside it, the regional azimuths at all sites swing to the north, but by different angles ensuring that all azimuths align to within 5° about magnetic north. East of the Rift, azimuth varies from site to site, being mainly dependent on the distance from the measurement site to the Rift edge and the Athi Ikutha shear zone (to the east). Principally a N40°E direction appears relatively constant at all the sites from Lukenya in the north to Mwatate at the southeast end of the line.

In order to perform two-dimensional modelling the principal impedances need to be obtained by rotating the measurement impedance by an angle defined here as the regional azimuth. The objective of this study was the geoelectric structure of the southern part of the Kenya Rift Valley. A regional N-S azimuth has been chosen for the following reasons.

(i). This is the regional azimuth at sites close to and inside the Rift Valley over a broad frequency range provided by both methods of strike determination, with small variations (±10°) from this orientation.

(ii). This is the azimuth obtained from the long period data, which are less affected by shallow distortions and topographic effects.

(iii). The regional geology/tectonic structure indicates a N-S trending axis, which correlates with the geoelectric strike.

Due to the above reasons, and since in all the soundings the telluric lines and the magnetic sensors have been orientated N-S and E-W (in geomagnetic coordinates), no rotation has been applied. Thus the XY polarization involving only $Z_{xy}$ is defined as the TE mode, while the YX polarization involving the $Z_{yx}$ is selected as the TM mode.

In the area of Chyulu Hills the orientation of the profile swings from N70°W to about N40°W and runs parallel to the regional strike. It was decided that these sites will be interpreted differently, using a simplistic 1-D approach since the setting of this line is parallel to the strike. However, a 2-D modelling, has also performed, rotating the impedance tensor to 40° west of north (for the TE mode).
Recent sediments
Cenozoic rift volcanics
Mesozoic sediments
Proterozoic Mozambique Belt
Archean Tanzania Craton
Rift faults

Figure 4.16: Regional azimuths defined following the conventional method of Swift (1967). Maps (a) to (d) present azimuth directions obtained from four characteristic AMT frequencies, while (e) to (h) are representing Long Period data. Lines in the direction of the regional azimuth are the same length for all sites.
Azimuth (Swift)

- T = 40 sec
- T = 1000 sec
- T = 500 sec
- T = 250 sec
- T = 1000 sec
Figure 4.17: Regional azimuths defined from the unconstrained Groom and Bailey (1989) decomposition technique. Maps (a) to (d) present azimuth directions obtained from four characteristic AMT frequencies, while (e) to (h) are representing Long Period data. The geological units are the same as for Figure 4.16. Lines in the direction of the regional azimuth are the same length for all sites.
Azimuth (Groom-Bailey)

T = 40 sec

T = 500 sec

T = 1000 sec
4.5 TEM data analysis

An important objective of the TEM data analysis is to produce a picture of the vertical resistivity distribution beneath the sounding point. TEM data processing involves converting the recorded voltages to apparent resistivities. In this project TEM data were collected at each of the MT sites, using Central and Coincident loop configurations, with either a 50 x 50 m or 100 x 100 m loop. The simple data transformation developed by Meju (1995, 1998) (see section 2.8.1.1.) has been applied to each data set in order to gain information about the shallow (<300 m) geoelectric structure. Figure 4.18 illustrates this scheme for the TEM data recorded at station Oloololo located west of the Rift Valley. The late-time central (x) and coincident (+) loop data appears the same, but they differ in the early-time channels; this may be due to a possible 2-D or even 3-D very shallow structure. The different geometry of the receiver may explain the dissimilarity between the two curves. Swift (1990) describes how the different geometry of the two loop configurations affects the responses in a complex environment. If a conductor channels the eddy currents away from the loop the effect on the centre of the loop may be quite small. Where a larger volume is being sampled the loop may contain a magnetic field of the opposite polarity reducing the effective magnetic field. A simpler explanation is that the central loop receiver is incorrectly calibrated or installed.

![Station: Oloololo](image)

**Figure 4.18:** Comparisons of TEM field data and transformation results for two different configurations, central (x) and coincident (+) loop. The observed apparent resistivity and voltage decay curves are shown in the left-hand panels. The corresponding results of simple resistivity versus depth transformation are shown in the right hand plot.
In ideal stratified ground, the central loop and the coincident loop curves should match each other but in areas of strong lateral heterogeneity they may differ (Meju, 1996). The Kenya Rift Valley is an area of complex geology, inducing different responses to each configuration loop. Figure 4.19 illustrate the discrepancy, caused by multi-dimensional geology, to the TEM data. TEM depth sounding at Magadi station suggest shallow 1-D characteristics, and the curves from both configurations match each other. At Singleraine, the differences between the two data sets possibly indicate the presence of a 2-D subsurface structure.

**Figure 4.19:** Comparison of central loop (x) and coincident loop (+) TEM depth soundings in a station with 1-D shallow depth conditions (Magadi) and a station (Singleraine) located in a 2-D environment. 1-D conditions produced curves that are identical for both configurations while 2-D structure differentiates the sounding curves.

Strong disparity between the two loops is also present at the Mwatate station located southeast of the Chyulu Hills (Figure 4.20). The overburden changes at this station with residual soil overlying undifferentiated Precambrian crystalline basement. Possible strong 2-D or 3-D effects may explain the big contrast between the two curves. Nevertheless, this data set appears to be similar to the data that Buselli (Buselli, 1982; Figure 2) used to illustrate the superparamagnetism effect in TEM data. Laboratory studies have shown the effect to be a feature of intergranular magnetism, inversely proportional to grain size and temperature and concentrated at the surface (Buselli, 1982). At the surface, arid weathering products of lateritic soil (hydrated iron oxides) have superparamagnetism properties that reduce surface resistivities. High frequency magnetotelluric data indicating shallow 1-D conditions, support the idea that the TEM curves have been affected by the presence of superparamagnetic material in the soil.
4.6 Static shift removal from the MT data using TEM data

One of the most serious effects that small-scale or near-surface conductivity inhomogeneities can generate is the static shift of the magnetotelluric apparent resistivity data to arbitrarily low frequencies (section 2.6.1). Static shifts are manifested in the data as vertical, parallel shifts of log-log apparent resistivity sounding curves, the impedance phase being unaffected. Small-scale features are likely to be three-dimensional; therefore, both apparent resistivity curves are shifted, usually by different amounts. This is so since both curves have contributions from electric fields normal to boundaries in the 3-D case, i.e. there is no true TE-TM mode separation. A recent promising approach to the remedy of the problem is based on the utilization of auxiliary TEM field measurements.

In this study, the TEM data have been jointly interpreted with the MT data to estimate the correct level of the MT apparent resistivity (Figure 4.21). TEM and MT sounding centres were located (for the most of the stations) in the same position to ensure that both methods are sampling the same geology. In sites where the TEM data were noisy, only the central loop data have been used since the strength of the induced signal is highest in the centre of the Tx loop. For the comparative analysis of TEM and MT soundings, MT wave frequencies \( f \) are related to transient times \( t \) with the empirically determined relation (Meju, 1995, 1998):
where \( t \) is in seconds and \( f \) is in hertz (cf. Sternberg et al., 1988).

**Station: Oloololo**

- TEM data
- MT data

![Graphs of TEM and AMT data](image)

**Figure 4.21:** A comparative analysis of TEM and AMT from a station with 1-D apparent shallow depth conditions. The MT phases have been plotted together with the TEM voltage decay curves. The TEM data are shown by triangles, and the circles represent the AMT polarizations. The AMT apparent resistivity curves have been slightly affected by different amounts of static shift. The central loop (CN) data have been correlated with the TE mode, while coincident loop (CC) data have been used to correct the TM mode.
For the sites in which the AMT sounding curves for the two polarizations illustrate shallow 1-D conditions, but are affected by static shift, it was found that either central or coincident loop data sets could be used to shift the MT apparent resistivity to the TEM level. Figure 4.21 shows a sample of MT data with static shift problem. The MT data show 1-D features at frequencies above 1 Hz, the similarity of the MT and TEM data for both loop modes and the MT polarizations in the overlapping segments is a sign that the MT data have been slightly affected by static shift. MT apparent resistivity and the error bars are eventually multiplied by a constant factor to bring the MT curve to match the TEM curve at the highest frequencies.

Comparative analysis of the TEM data and the short period MT data (128 Hz – 32 sec) helped to correct equipment related shift in the AMT data, and suggests that the long period data (40–10000 sec) are unaffected by this kind of shift. A representative data set is presented in Figure 4.22. The AMT TE mode was shifted upwards by a factor 1.6 (using TEM control), while the TM mode appears unaffected (compare with Figure 4.9). The recovered AMT apparent resistivities match the observed LMT apparent resistivities. As has been suggested above, miscalibration between the MT systems may explain this behaviour of the apparent resistivity curves.

**Station: Nguruman**

- TEM data
- AMT-LMT data

![Figure 4.22: Comparative analysis of TEM, short period MT, and long period MT data in station Nguruman. The TE mode for the AMT data has been shifted upwards, after joint interpretation with the TEM data. The long period data appeared unaffected, confirming that the proposed level of shift is the correct one. This is not the case for other stations that both, short and long period, apparent resistivity curves appeared shifted by different amounts.](image-url)
Pellerin and Hohmann (1990) in their interpretation scheme had set two criteria for the interpretation to be accurate: (i) The minimum MT penetration depth has to overlap with the maximum TEM penetration depth, and (ii) the MT sounding data should display a 1-D response in the overlapping area. In sites of complex geology MT sounding curves are dissimilar; the conventional method of shifting the MT apparent resistivity to the correct level may not be appropriate since the overlapping area may not be parallel. In these sites, it has been noted, that each of the MT polarizations can be adequately correlated with one of the loop configurations A coherence between one of the TEM loop and an AMT mode has been identified; central loop data are related to the TE mode, while the TM mode matches with the single loop (see also Meju et al., 1999). In these environments joint inversion of MT phase and TEM apparent resistivity (Meju, 1996) appears to be more efficient for static shift correction, since the requirement for an overlap region between the MT and TEM data and flat high-frequency ends of the MT curves are not essential. This correction technique is based on the assumption that the impedance phase remains unaffected by the static shift. Another useful feature of this dual inversion is that limited TEM data would suffice for recovering a reliable resistivity structure in areas of simple geology. Data from station Chyulu West in Figure 4.23 have been used to illustrate the above technique. TEM apparent resistivity and MT phase for the TE mode have been jointly inverted and a likely level for the MT apparent resistivity has been defined. Comparison of the TEM apparent resistivity with neighbouring stations indicates similar shape and magnitudes, sustaining the belief that the shifted MT apparent resistivity represents a true, or at least a reliable, subsurface resistivity structure. The results of applying static shift correction in AMT and LMT data using TEM data are presented in Appendix C for all the sounding sites.

Finally, special reference has to be made to Irkiba site. AMT polarizations display 2-D, or even 3-D, characteristics (Figure 4.24), with splitting phases and apparent resistivity curves. 2-D effects are imaged, also, in the TEM data. Comparative analysis of the TEM and AMT data in this site illustrates strong discrepancy between the TEM and AMT apparent resistivity curves, indicating distortion of the AMT data. Potential explanation for the distortion is the surface topography and surface inhomogeneities. The site is located in the north-east part of the Loita Hills, 7 km west of the 2000 m high Nguruman Escarpment. It is well established that strong topographic features (like the Nguruman Escarpment) can severely distort the impedance tensor causing shift problems (Jiracek, 1990). The apparent resistivity values due to topographic effects are highest in valley troughs and lowest on topographic peaks (Jiracek, 1990). Several techniques have been proposed in the literature for the theoretical treatment of topographic effects (Wannamaker et al., 1986; Chouteau and Bouchard, 1988; Jiracek et al., 1989). Although static corrections based on topography have proven valuable, particularly in
mountainous terrain, they do not solve static shift problems due to subsurface bodies. The geological complexity of the area suggests that the source of the distortion in this site, besides the topography, can be the existence of subsurface irregularities. Shifting the MT apparent resistivity curves in the proposed from the TEM data level, as it appears in Figure 4.24, effectively removes the static shift problem, so that the numerical two-dimensional modelling can be reliably performed to produce interpretable 2-D subsurface models.

![Station: Chyulu West](image)

**Figure 4.23:** Result of joint inversion of TEM apparent resistivities (central loop configuration) and MT phase data, for the TE mode in site Chyulu West, a 2-D environment. The calculated MT apparent resistivity, from the inversion process (left-hand panels), pointed the level that the observed MT data had to be shifted (right-hand panel).

To conclude, it has to be mentioned that the use of the TEM data in this study for the remedy of static shift problems led to a reliable estimation of the undistorted AMT resistivity curves. For the shallow 1-D cases (where the TEM and AMT curves are parallel) the scheme accurately corrected the data. For sites at which 2-D and 3-D effects were present, implementation of this technique appeared to be very effective in providing a more accurate picture of the resistivity structure, and increased the interpreter’s insight by ascertaining if the area is approximately 2-D. Nevertheless, it has to be pointed out that so far none of static shift removal techniques have been proven to work well in every geological environment (Groom and Bahr, 1992).
Figure 4.24: Illustration of static shift correction using the TEM data at the Irkiba site before (left-hand panels), and after (right-hand panels). Topographic effects and surficial inhomogeneities can be blamed for the strong distortion. The AMT TE and TM mode apparent resistivity curves were respectively corrected using central- and coincident-loop data.
4.7 Qualitative interpretation of the data

A qualitative interpretation of the data before the modelling can provide prior information about the geoelectric structure of the study area and lead to a more realistic view of the subsurface structure. In Appendix A (pages A2-A15) are shown the sounding results for all the sites across the Rift Valley from Lake Victoria in the west to north of the Chyulu Hills in the east. The quality of the data based upon scatter error bars and data point continuity is generally good at most sites in the high frequency bands (127 - 0.01 Hz). Lower frequency data appeared more scattered and noisy. In many stations the apparent resistivity curves appear to have been shifted along the vertical axis suggesting influence of near-surface inhomogeneities.

Apparent resistivities at the sites located west of the Rift Valley (sites Nyamanga to Irkiba) are moderately anisotropic, with the values in the principal directions differing by less than one decade for high frequencies but in the lower frequencies the anisotropy increases. Phase information in most of these stations suggest a 2- or 3-D environment. At most of the sites, the skew values appear to increase from 0.2 at high frequencies to 0.4 at frequencies below 1 Hz. Strong lateral variations occur between stations Lolkoroin and Oloololo in the west. The data from Keekorok appear very noisy and scattered with big error bars and high skew factors. Approaching the flank of the Rift Valley, the telluric field appears polarized in the N-S direction and the apparent resistivity decreases indicating a more conductive lower crust. At Irkiba the data are at least 2-D in character; the apparent resistivity curves are strongly anisotropic, the phase data are very different for the two principal directions and the skew values are 0.2-0.4.

For the stations located on the rift floor (Nguruman, Magadi, Singleraine) and the stations east of it (Selengei, Chyulu North) the azimuthal values are scattered; this feature together with the somewhat isotropic resistivities for the high frequencies (>0.1 Hz) suggest that 1-D interpretation of the results could provide a reasonable indication of the actual resistivity distribution at these sites. In this area the resistivities appeared to be about two decades lower than those from west of the rift and the penetration depth is distinctively less. At the station north of the Chyulu Hills (Chyulu North) the geoelectric structure appeared 3-D with very high skew factor (0.6) and strong disparities between the two polarizations.

Figure 4.25 show pseudosections of the XY and YX polarizations responses for the stations of the main profile, and serves as a measure of the relative depth of penetration of the electromagnetic field. The pseudosections are valuable in showing how well the available impedance estimates constrain the resistivity distribution, both in terms of the spatial density of the sites and the frequency range of the observations (Banks et al., 1996). Figure 4.25 shows very clearly the importance of close station spacing in the vicinity of geological discontinuities such as the western boundary of the rift (between stations Irkiba and Nguruman). Broadly
speaking, the profile can be divided into three main geoelectric sections: a resistive section west of the Rift Valley (>1000 Ω.m), a low resistivity rift valley floor (=10-50 Ω.m), and a moderate resistivity sector east of the Rift Valley (>100 Ω.m).

**Figure 4.25:** Psuedosections of shifted apparent resistivity [(a) & (c)] and impedance phase [(b) & (d)] of the XY-pol [(a)-(b)] and YX-pol [(c)-(d)] responses for the stations of the main line across the southern Kenya Rift Valley. The resistivity values have been contoured on a logarithmic scale. Values shown correspond to log (apparent resistivity Ω.m) and degrees. The vertical line dots below the site locations show the data points used for contouring.
Chapter 5

Modelling of Kenyan Data

In this chapter 1-D and 2-D models of the conductivity structure across the southern part of the Kenya Rift Valley are presented along with a discussion of the possible errors and the extent of the non-uniqueness present in the models.

5.1 One-dimensional modelling of the Kenya data

One-dimensional (1-D) modelling (forward and inversion) remains an important tool for interpreting magnetotelluric data. While the Kenya Rift Valley is at least 2-D in character, it is known from 2-D and 3-D modelling studies that 1-D modelling can recover representative profiles in regions of complex structures (Ranganayaki, 1984). Since the data at most sites are approximately 1-D for frequencies above 10 Hz, 1-D modelling was undertaken as a first approach to the interpretation of the measured responses and a guide for a starting model for the 2-D modelling.

5.1.1 Niblett-Bostick contour section

Traditionally, pseudosections of the resistivity or phase variations with frequency provide useful information in 2-D environments (Ranganayaki, 1984). The pseudosections give an initial picture of the amount of structural detail that can be anticipated in the resistivity models to be derived for a region. A drawback in this technique is that features shown at common frequencies may in reality be generated by structures at different depths. A more realistic structural picture of the area can be obtained when resistivity data are mapped onto their appropriate depths via the Niblett-Bostick transformation (e.g. Meju, 1988). This approach is implemented for the Kenya data using the TE and TM modes of the AMT observations. The MT data have been corrected for static shift effects as previously described (section 4.6).

The Niblett-Bostick transformation was used to compute estimates of the resistivity-vs.-depth variation for sites across the Kenya Rift Valley. The transformed values were contoured on a logarithmic scale to obtain a pseudosection (Figure 5.1). The upper parts of the Figure 5.1 show the near-surface structure (down to 5 km), whilst the lower diagrams display the total crustal structure, derived from the AMT and LMT data.
Figure 5.1: The variation in the Niblett-Bostick determined resistivity of the TE and TM-mode with depth across the Rift Valley. AMT data have been used for the western section, while both, AMT and LMT data were contoured for the area of the rift and east of it. Parts (a) and (b) present the TE-mode, while (c) and (d) are representing the TM-mode. The resistivity values have been contoured on a logarithmic scale. The vertical line dots below the site locations show the data points used for contouring. West of the Rift high resistivity crust is found in the majority of the stations. Inside the Rift more moderately resistive rocks (\(\approx 2.0 \log_{10}\ \\text{resistivity; } 100 \ \Omega\cdot m\)) underlie the conductive sediments (\(\approx 1.0 \log_{10}\ \\text{resistivity; } 10 \ \Omega\cdot m\)). East of the Rift lack of observations reduce the efficiency of the quantitative geoelectric section. (a) and (c) shows the shallow structure (<5 km) whilst (b) and (d) give the observed deeper variation (<40 km) at the corresponding MT site locations annotated at the top.
Unequal spacing of the data in the vertical direction, compared with the larger separation between the stations (~20 km) counts against any detailed interpretation. It is obvious in Figure 5.1 that in the west part the crust is resistive (up to 1000 $\Omega \cdot m$) and becomes more conductive inside the rift and east of it (~10 $\Omega \cdot m$). The depth transformation of the TE and TM mode responses at all sites (Figure 5.1) provide a general guide about the penetration depth of the data. West of the rift where the structure is more resistive, the TE-mode penetrates to depths of about 50 km while the TM-mode provides information to about 45 - 50 km. Inside the Rift Valley
the depth of penetration is reduced to about 5-10 km for both modes for the AMT data and the additional long period data provide information to about 20-30 km. East of the rift the penetration depth is about 15-20 km.

The resistivity contours show strong lateral variations in structure especially at the west boundary of the Rift Valley, close to the Irkiba station. In the TE mode the west geoelectric boundary of the rift is located between stations Morijo and Irkiba and appears shifted west of its topographic expression. Additional indication of structural discontinuity in the upper crust appeared in the vicinity of station Lolkoroin which is close to the Oloololo Escarpment. East of the Rift Valley the underlying structure appears less deformed, but lack of observations between stations Singleraine and Selengei reduce the integrity of this inference. On the whole, inferences can be more confidently drawn from the west section because of the higher density of the measurements, and in the area of the Rift Valley where the long period data have increased the resolution in the lower crust.

5.1.2 1-D inversion results

The qualitative interpretation and the Niblett-Bostick pseudosection demonstrate the strong 2-D character of the Rift Valley and the use of the 2-D modelling technique is imperative. As has been noted, a priori information for the 2-D numerical model is necessary in order to obtain a reliable model and reduce the computing time. The "initial guess" for the 2-D interpretation is usually based on collated 1-D models and the use of independent data. In this study 1-D inversion of both polarization modes were used to build pseudosections along the profile which were then adopted as starting models for higher dimensional interpretation schemes.

The parametric inversion program that has been used in this study for the 1-D modelling was an algorithm developed by Meju (1992, 1996) for joint interpretation of TEM and MT data. The algorithm uses the least squares method to solve the non-linear problem. The algorithm handles constrained problems and the resulting system of equations are solved using the singular value decomposition (SVD) technique with resistivity and depth as the model parameters. The general formulation of the inverse problem has been described in section 2.8.1.3.

The code requires an initial model and the Niblett-Bostick depth transformation provided a preliminary guess to the geoelectric structure; some modification to the model through forward modelling improved the fit of the model to the observed data and finally inversion was applied to produce the best fit. Both resistivity and depth were allowed to vary. Examination of the model resolution matrix, parameter covariance matrix and error misfit provided an indication of
the quality of the inversion result. Non-linearity and noise prevent an exact solution being found (Parker, 1984).

Joint interpretation of TEM and AMT-LMT data would allow improved model resolution and especially unravel the shallow structure. The Niblett-Bostick transformations of the MT data and the TEM resistivity-depth transformation (equations 2.72, 2.73) are used to derive initial models and the two data sets were jointly inverted to enhance resolution of the resistivity model. Several other starting models were also inverted; the aim was to identify the resulting models for which the inverted parameter values produced the best fit, and from them to select the one to represent the site in a collated geoelectric section. The retention of those models which produced similarly small misfits helped to show the range of equally acceptable models and to identify the "simplest model" (i.e. the one with the fewest layers) which was consistent with the acquired observations and to assess its overall uniqueness. In most of the stations, a relatively small number of models satisfied the best fit criteria out of which one was ultimately selected as representative of the true geoelectric structure of the site.

This exploratory modelling was routinely applied to all the TEM and MT soundings within the study area, and was particularly useful for determining the 'least information' (minimum number of layers) which can be gleaned from the response functions. To estimate the range of possible parameter values a damped most-squares inversion scheme (Meju, 1988; Meju and Hutton, 1992) was applied. The TEM sampling times were transformed to their equivalent frequencies (equation 4.12) and joint inversion of TEM and MT apparent resistivity was effected. Sites with scattered and noisy data have resulted in a wider range of acceptable parameter values. Probably the least well-determined parameters were the depth position and resistivity of the terminating layer. It was noted that a wider range of parameter values are found for the resistive layers reflecting the ability of the EM method to detect preferentially more conductive sequences of rock.

Figure 5.2 illustrates the joint modelling of TEM and MT data at one station (Singleraine) using the most squares technique. In this figure, the responses of blocked 1-D structures containing the main features of the transformed resistivity profile are superimposed on the field data in the left hand panels. On the right hand plots are shown the blocked inversion models corresponding to the simple TEM data transformation and the Niblett-Bostick transformation for the MT data.

Although the available MT bandwidth was fairly broad (127 Hz ~ 5000 sec) a main assumption about the resistivity distribution with depth was that the proposed model had the minimum structure possible for some tolerable level of misfit to the data (Fischer and Le Quang, 1982; Smith and Booker, 1988).
Station: Singleraine

- TEM data
- MT data

**CN TEM - TE mode**

- Resistivity (Ω.m)
- Depth (m)

**CC TEM - TM mode**

- Resistivity (Ω.m)
- Depth (m)

**Figure 5.2:** Results of joint inversion of TEM and MT data for the two polarization modes using most squares inversion scheme. The TEM data are shown by triangles and have been plotted in the left-hand panels at their equivalent MT frequencies (1/T) for comparison. The MT apparent resistivity and phase field curves and the corresponding Niblett-Bostick transforms (shown in the right hand plots) are represented by circles with their associated errors. The solid response curves in the left-hand panels are those computed for the blocked models shown in the right-hand plots. In TM mode the deep crustal resistors vary from 700 to 3000 Ω.m, while the overlying conductor appears particularly well resolved as revealed by the very small range in the bounds of the acceptable models for this parameter.
The most-squares models clearly cover a range of possible models and thus provides a measure of the non-uniqueness. Only the apparent resistivity data have been inverted. The figure clearly portrays the extra information that the TEM data have added to the 1-D models generated from the MT data providing a more complete image of the near-surface resistivity features for each station.

5.1.3 Interpretative geoelectric section

The 1-D results for the TE and TM mode were collated into a geoelectric structure that may be interpreted as characterising the resistivity distribution of the area and by implication its geotectonic structure. The interpretative geoelectric sections for the region of the Rift Valley are shown in Figure 5.3. The error bars represent variability in the depth determination of each layer as determined using the most squares method.

West of the rift, the TE mode geoelectric model shows a highly resistive (2000-5000 \( \Omega \cdot m \)) crystalline upper unit that is interrupted in the area of the Lolkoroin and Ooololo stations characterized by less resistive (100-600 \( \Omega \cdot m \)) material and possibly marking the Ooololo Escarpment. A lower resistivity unit (200-1000 \( \Omega \cdot m \)) underlies the resistive upper crust from stations Migori to Leganishu. The lower crust and upper mantle are very conductive (about 10-50 \( \Omega \cdot m \)) but have zones of higher resistivity (400-4000 \( \Omega \cdot m \)) underneath stations Nyamanga, Kehancha and Keekorok. The western section is bordered to the east by a highly resistive zone underneath station Morijo (=10000 \( \Omega \cdot m \)). The TM mode for this area exhibits similar structure for the upper crust with changes in the depth limits and resistivity range. In the TM model the resistor extends to about 10-15 km depth and appears more uniform; it is interrupted only at station Lolkoroin. The underlying layers are less resistive (150-100 \( \Omega \cdot m \)) in the vicinity of station Kehancha, and their resistivity increases (1500-5000 \( \Omega \cdot m \)) at stations Ooololo and Keekorok. The upper mantle appears highly conductive (= 50 \( \Omega \cdot m \)). In comparison with the TE mode, in the TM mode the resistivity distribution in the lower crust and upper mantle close to the western flank of the rift (underneath station Morijo) appears more uniform without strong lateral discontinuity, as it appears in the TE mode. Lack of long period data for the Lolkoroin station limits the resolution of the structure in the area for both modes.
Figure 5.3: Interpretative geoelectric sections for the southern Kenya Rift Valley, derived from the 1-D interpretation of the TE and TM modes. The vertical bars represent the most squares limits of the 1-D models. The values of the resistivity units are in $\Omega \cdot m$. 

5.8
At station Irkiba, a sequence of thin layers marks the geoelectric boundary of the rift and defies any correlation with the west sites. Both TE and TM modes inside the rift recover conductive surface and basement (4-40 $\Omega\cdot$m). The conductive structure appears only to be interrupted from a resistive block (500-1000 $\Omega\cdot$m) underneath station Magadi. The conductive surface detects valley sediments, while the conductive basement may be interpreted as magma that permeates the whole crust. High conductivities limited the penetration of the MT signals and the thickness of the layers is not well determined. The TE mode in this section of the profile appears more laterally uniform as expected from MT theory, compared to the western section.

East of the rift, the structure remains conductive and uncertain. Insufficient data coverage for this section (80 km spacing between stations Singleraine and Selengei), poor quality long period data at Selengei station and influences from neighbouring 3-D structures (Chyulu Hills) suggests that the 1-D geoelectric sections cannot provide any reliable information about this part. The erratic limits of the basement layers illustrate the problem of non-uniqueness in the 1-D modelling.

5.2 Two-dimensional modelling study of the Kenya Rift

The structural complexity of the Kenya Rift Valley requires the use of 2- or 3-dimensional modelling procedures for realistic quantitative interpretations. Considering that the Rift Valley, to a fair approximation, is a linear structural element with a dominant N-S trend, two-dimensional interpretation can be considered sufficient for this regional study.

The 2-D inversion modelling program used in this study was written by Randal Mackie (1996). An initial guess of the electrical structure of the subsurface (starting model) is required. This is done by assigning resistivity values to represent portions of the subsurface materials underlying the survey line. The inversion program uses the predicted impedances from the forward problem to modify the model parameters such that, over a number of iterations, the inversion will find a 'better' set of model parameters that minimize the error between the model response and the observed data. The formulation of the inverse problem is similar to that used by Jupp and Vozoff (1977) in that the block geometry is fixed. Only the resistivities are allowed to vary during the inversion.

The region to be modelled was divided by a number of horizontal and vertical lines into a mesh of rectangular cells of variable sizes. The intersections of the horizontal and vertical lines form the nodes of the grid. There are three factors that must be considered in determining the grid spacing (Madden and Mackie, 1989). First, the spacing has to adequately describe the important electrical parameter of the area of interest. Secondly, the errors in the approximation
resulting from the discretization have to be at an acceptable level. Thirdly, limits have to be imposed on the computational effort involved in obtaining the solution.

Proper treatment of the model discretization is important in calculating the magnetotelluric response of realistic 2-D models. Because of the damped nature of electromagnetic propagation in conducting media, it is common to accept that the area responsible for a magnetotelluric response is confined to the order of a skin depth, and this applies to the horizontal as well to the vertical distances. This is correct in ordinary circumstances, but very different situations can occur when dealing with anisotropic media (Madden and Mackie, 1989). The earth is essentially very anisotropic electrically simply because it has layers with strong electrical contrasts. If an anisotropic conductivity zone lies between a measurement point and a region of different structure, the influence of the distant structure on the measurement may be underestimated. Another problem is that the spacing requirements are frequency dependent. Both of these factors can be dealt with by using graded nets. In a graded net the spacing varies with position. By grading the 2-D model in the horizontal direction, it is possible to extend the side boundaries to large distances without creating unmanageably large systems, and discretizing the model on a finer scale near regions of strong resistivity gradients allows one to obtain more accurate solutions. It is possible also to grade the model in the vertical direction (Mackie et al., 1993) since: (a) the diffusive nature of electromagnetic waves in conducting media smears the information content with depth and (b) this allows the use of the same model for a wide range of frequencies without changing the vertical spacing, at least up to some high frequency limit that is determined by the thickness and conductivity of the first layer. To satisfy the boundary conditions several graded air layers are added on top of the earth model to account for perturbations in the H fields from lateral current gradients. All the H field perturbations are required to be damped out at the top of the air layers. At the bottom of the earth model, a 1-D impedance for a layered earth is used to relate the E field to the H field. Madden and Mackie (1989) present an analytical discussion about the spacing problem and the boundary conditions.

The input parameters for the inversion program are the observed data, the expected \( \text{rms} \) error, the smoothing factor TAU, the noise floor, and the fixed parameters. Any combination of TE and/or TM mode data may be input to the inversion algorithm. The program uses period data, so all the data had to be converted from Hz data to period. The \( \text{rms} \) error is the error at which the program terminates if reached before the maximum number of inversion iterations (usually less than 60 are expected). The \( \text{rms} \) error is defined:

\[
\text{rms} = \sqrt{\frac{(d - F(m))^T R^{-1}_{dd} (d - F(m))}{ndpts}}
\]
where ndpts is the number of data points, and the rest of the symbols have been defined above. Additionally, noise in the data and non-2D data may prohibit the program from reaching the target rms error values. TAU is the regularization parameter that controls the trade-off between fitting the data and adhering to the model constraint. Larger values cause a smoother model at the expense of a worse data fit. The noise floor defines the value that the input errors that are below this value will be reset to this value. Finally, a file with the model parameters to remain fixed through the inversion can be added.

5.2.1 Prior information and initial models

A realistic initial model has to be constructed to minimize the computing time and effort involved in the 2-D modelling studies. In order to construct a realistic starting model for the 2-D inversion, independent geological data were combined with the 1-D results, the information from previous geoelectric studies in the area, and information from the evaluation of the structural characteristics of the $\rho_{TE}$ and $\rho_{TM}$ responses across the profile. Previous studies of MT and GDS data from the Kenya Rift (Banks and Ottey, 1973; Beamish, 1977; Rooney and Hutton; 1977; Banks and Beamish, 1979) suggest that much of the crust within the rift is highly conductive (10 Ω.m) from the surface down to >35 km depth. Seismic studies in the region propose a crustal thickness of ~33 km beneath the Archean craton in the western part and 40 km beneath the Mozambique belt in the east (Birt, 1996).

The behaviour of the $\rho_{TE}$ and $\rho_{TM}$ responses across a geological steep contact (e.g. fault) can also be used in initial model construction. The observed apparent resistivities at 4 selected frequencies for the AMT recordings and two frequencies from the LMT data for the TE and TM polarization cases are plotted against the station distance in Figures 5.4 and 5.5 respectively. The figures illustrate two anomalous zones; one west of the rift in the area of Oloololo Escarpment (stations Lolkoroin and Oloololo) and the Rift Valley itself.

West of the rift in Oloololo area (60-80 km from the west end of the line), the AMT data for high frequencies (>0.5 Hz) identify a zone ~50 km wide of "dyke-like" appearance. An oblique conductor appears to exist near Lolkoroin extending to shallow depths close to station Oloololo. The resistivity (~100 Ω.m) is an order of magnitude less than that of the surrounding crust. The pattern of the $\rho_{a}$ in the vicinity appears quite similar for the two polarizations, indicating that the two modes are not well defined, suggesting a smaller scale local structure with different orientation from the Rift Valley. This is supported by the azimuth of the data in the area (see Chapter 4). This structure is not well defined in the LMT data, noting that long period observations are not available for Lolkoroin.
Figure 5.4: Variations of TE and TM apparent resistivities versus distance across the Kenya Rift Valley at four selected AMT frequencies. The dash lines border the two anomalous zones across the profile, the Oloololo area (60-90 km from the beginning of the line), and the Rift Valley. The expected greater discontinuity in the apparent resistivity values for the TM than for the TE mode data in the west contact area of the rift is apparent only in the lower frequencies.
From theory, a greater discontinuity in the apparent resistivity values for the TM mode than for the TE mode data is expected across the rift zone. Notice that the AMT data show almost the same $\rho_a$ pattern in both TE and TM modes across the profile for frequencies above 0.5 Hz. For the stations immediately to the west of the rift, this may be due to the long distance of the stations (>1 skin depth) from the contact zone (area between stations Irkiba and Nguruman) which is therefore not imaged by the data, and the 1-D character of the shallow structure. Inside and east of the rift, the 1-D characteristics of the shallow sedimentary basin prevent any clear separation between the two modes. However, at the lower frequency of 0.04 Hz and the long period data the $\rho_{TE}$ varies smoothly across the rift, while the $\rho_{TM}$ indicates a greater discontinuity. This provides a justification for the 2-D interpretation of the field data and a first quantitative indication of the basic structure of the rift. The spatial resistivity plots for the traverse also show that the resistivity (1-10 $\Omega$.m) of the rift is two to three orders of magnitude less than the crust west of the rift (1000 $\Omega$.m) and an order less than the crust east of the rift.
The western margin of the rift valley is located in the vicinity of Irkiba station (175 km) and the eastern margin in the area of Selengei station (330 km). Another area of resistivity contrast inside the rift is proposed between stations Nguruman (205 km) and Magadi (231 km), with the Nguruman station to be located at the conductive part of the contact. Finally, to the east of the rift a resistivity contrast is apparent in the lower frequencies (<0.5 Hz) between stations Selengei (330 km) and Chyulu North (376 km) with Selengei being on the resistive part ($\rho_{TE}$ is above $\rho_{TM}$).

The 1-D geoelectric sections for the TE and TM modes (Figure 5.3) illustrate features that are not entirely common with the above observations across the whole profile. In station Oloololo the 1-D interpretation modelled a conductor at depths of 35 km for the TE mode and ~50 km for the TM mode that has not appeared in the spatial resistivity plots in the relating frequencies. Additionally, the 1-D shallow structure of the rift floor proposed from the spatial resistivity plots is not supported from the differences that appeared between the two pseudosections in the area. Combination of the two geoelectric sections, resulted in the construction of a realistic starting 2-D model for the profile.

### 5.2.2 Modelling procedure

The main traverse is trending ≈68 degrees west of the geographic North. Since the geoelectric strike of the Rift Valley was defined to be N-S (chapter 4) the stations have been projected onto a line orientated E-W, crossing the Rift Valley in the area of Magadi station. Projecting the stations onto an E-W line the two polarization modes (TE and TM) are accurately defined orthogonal (parallel and perpendicular, respectively) to the strike direction (N-S) for the 2-D modelling procedure. This may affect the model resolution since different distances from the Rift Valley are now involved. Because of the large separation between the sites and the differences in geoelectrical characteristics between the area west and the section inside and east of the rift, this author decided to model the profile in two overlapping parts with different grid size and discretization; it was considered capable of providing better resolution as well as reducing the computation time. The vertical and horizontal discretization of the models were dictated somewhat by the network formulation (Mackie et al., 1988); that is, the error introduced by the discretization should be as small as possible, the model response needs to be calculated over a range of frequencies, and the model should attain 1-D values at its edges. In the 2-D finite difference algorithm the response functions in the centre of each grid element are calculated (see Madden, 1972).

The western segment was modelled using a model with grid size of 77 x 56 elements and extended from site Nyamanga at the west end of the line to site Magadi inside the rift. The area
of interest in this section is mainly the area west of the Nguruman Escarpment. The grid was extended down to 35 km depth (excluding the last 4 rows that were used as the basal half space) and was more detailed between the depth interval 3 and 30 km. The two stations inside the rift (Nguruman and Magadi) that were included in the model provided control for the boundary conditions at the eastern edge of the model. In the other model, the Rift Valley is of the main interest. In this eastern section, the rift and the area east of it were modelled using a mesh design of 64 x 43 elements. The section extended from station Leganishu west of the Nguruman escarpment to the end of the line at Chyulu North. Analogous with the western segment, the sites to the west of the rift were used to provide the appropriate boundary conditions for the western border of this rift model. The grid for this model emphasized the shallow structure and was finer for the top 10 km, extending to 25 km depth. Tables 5.1 and 5.2 illustrate the gridding of the western and the rift sections, respectively. Care was taken to add more grid elements in areas of certain interest (e.g. close to Oloololo and Nguruman Escarpments). Since the topography changes significantly along the profile (1500 m above sea level west of the rift, 600 m inside the rift and 1000 m east of it) topography was included in both models.

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| Z - Block thicknesses (m)               |
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| 80.0000 130.000 130.000 80.0000 40.0000 40.0000 60.0000 |
| 100.000 100.000 200.000 500.000 1000.000 1000.000 1000.000 |
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| 2000.00 2000.00 1000.000 500.000 500.000 500.000 500.000 |
| 1000.000 2000.00 3000.00 1500.00 5000.00 20000.0 800000. |

Table 5.1: Grid size used in the 2-D modelling of the western section. In this section the sites that were modelled were: Nyamanga, Migori, Kehancha, Lolkoroin, Oloololo, Keekorok, Leganishu, Morijo, Irkiba, Nguruman and Magadi. The gridding for the last 2 sites is not well discretized since the section focuses on the stations west of the rift and reducing the size of the mesh reduces the computation time.
Table 5.2: Grid size used in the 2-D modelling of the eastern segment. The sites that were modelled were: Leganishu, Morijo, Irkiba, Nguruman, Singleraine, Selengei and Chyulu North. Similar to the western section, the grid elements for the first station (Leganishu) are quite big in an effort to reduce the size of the grid; more attention was taken for the sites inside the rift and close to the Nguruman escarpment.

Initially, a few inversion runs were performed to determine the best TAU for the problem. According to the program manual, the value of the TAU should optimally be chosen such that the $\text{rms}$ error for the inversion is between 1.0 and 1.5. A TAU factor of 5 and 10 for the western and eastern sections, respectively, has been decided upon and used. Additionally, the noise floor in the data was arranged to be 5% for both sections. Furthermore, the first few iterations helped to identify the most likely zones where the true geoelectric boundaries may be located and subsequent modifications have been performed to both of the models to improve the fit to the data. This procedure of trial and error and steps of 2-5 iterations have been performed several times, other features were added into the basic framework with necessary adjustments until a reasonable fit for both polarizations and an acceptable structure was found. Furthermore, these first inversion runs identified the most suitable discretization of the problem and crucial changes to the cell sizes helped to accomplish a better fit. It has to be noted that during the whole 2-D inversion both modes (TE and TM) were inverted with equal weighting.

Once the best possible starting model has been created and the more appropriate input parameters have been determined, the desired $\text{rms}$ error is set to 1.0, and the maximum number of iterations to achieve these criteria is set to 60. Note that noise in the data and non 2-D data prevent the program from reaching the target $\text{rms}$ error value. Nevertheless, the
program automatically stops when the minimization converged and could not reduce the error further. In this study 60 distinct frequencies were used, the number of frequencies at each individual site was 40-53 depending on whether LMT data were included or not. The program in the end produced 4 files containing: (i) the calculated responses of the TE and TM mode, (ii) the resistivity map corresponding to the same grid elements used to specify the starting model, (iii) statistics about each iteration, and (iv) the CPU elapsed time versus iteration. The model statistics for the eastern section model are presented in Table 5.3. In this table “tau” is the smoothing factor, “chi” is desired chi square that is actually the number of the data points used in the inversion, “s1” is the chi square, “s2” is the model roughness, and “s2/t” is the model roughness divided by the smoothing factor, “s3” is the closeness measured if fixed parameters have been specified (in this case 0), and finally “s” is the objective function that the program tries to minimize (s=s1+s2+s3). For the western section model, 32 iterations produced a good misfit with rms error of 1.17655 (i.e. chi square of 2574.73 for 1860 data values). For the eastern section model the program stopped when it completed 37 iterations with rms error 1.74272 (chi square misfit of 4336.96 for 1428 data values).

The modelling of the profile in two sections, was done to reduce the computation time by decreasing the grid size, and to facilitate the optimization of the mesh design. An overlap zone of the 5 stations, was considered sufficient and this served as the conductive rift in the western section model and as the resistive western flank to the eastern section model. For completeness, after the general resistivity structure has been determined from the two separate segments, the two sections were combined into one inversion model for the total length of the profile (station Nyamanga to station Chyulu North). The grid size was 107 x 56. For this model 31 inversion iterations were performed and a more satisfactory model with the best possible fit between the calculated and observed data has been obtained. A satisfactory model here means that the model response curves lie within the zone defined by the error bars of the E-pol and H-pol data for most of the sites (Figure 5.7). The rms error for this model was 1.91423 (chi square 10487.19 for 2862 data points). The results from the inversion of the whole profile in a single model have not changed the features that appeared in the two separate sections.

During the inversion problems have been noted in station Selengei. The bad quality of the long period data (especially the TM impedance phase) in this station and the steep nature of the apparent resistivity curve in frequencies $10^1$ - $10^2$ Hz combined with the small error bars of the LMT data, forced the program to model the long period data causing a large misfit to the AMT data. Excluding the LMT data from the inversion resulted in a satisfactory fit of the AMT data. Subsequent modifications to the deep structure of the area and further 2-5 iterations
including the Selengei LMT data helped to obtain a sensible fit for both AMT and LMT observations without affecting the misfit for the rest of the stations.

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Table 5.3: Statistics for the inversion of the eastern section model. The inversion converged after 37 iterations. The parameters and their meaning have been described in the text.
5.2.3 2-D geoelectric models

The interpretative 2-D model for the southern part of the Kenya Rift Valley is presented in Figure 5.6. The fit of the model curves to the observed TE and TM mode data for all the sites are shown in Figure 5.7. Notice the good fit in both amplitude and phase for the two polarizations in most of the sites. The main features of the final model are summarized below.

West of the Rift Valley the top 10 km is very resistive (=2000 Ω.m) but is broken by a 20 km wide zone of moderate resistivity (=100 Ω.m) at Lolkoroin. This resistive upper crustal unit is underlain by a mid-crustal zone of lower resistivity (100-600 Ω.m) which varies in thickness from 10-12 km between Migori and Lolkoroin to about 20 km between Oloololo and Keekorok. The mid crustal unit is separated into two blocks between Lolkoroin and Oloololo by a vertical conductive zone that transects the entire crustal thickness (this may be the boundary zone between the Archaean rocks to the west and the Proterozoic rocks to the east). The lower crust is highly resistive (>4000 Ω.m) in the westernmost block down to about 32 km where it rests on conductive (10-50 Ω.m) substratum. The highly resistive lower crustal unit is absent in the other block to the east where an anomalous conductive layer (5-30 Ω.m) is present at about 30 km depth. The absence of interpretable LMT data at Nyamanga and west of it and at Lolkoroin have limited the resolution of the conductive structures at =30 km depths. Nevertheless, the overall decrease in \( \rho_{TE} \) and \( \rho_{TM} \) and values of \( \phi_{TM} \) of 60° to 70° for the lower frequencies in the stations of the area clearly indicate the existence of mantle conductors. The conductive block in the Oloololo area is better constrained by the high quality LMT data at this station. The entire crustal thickness is resistive at the western flank of the Rift Valley (from Morijo to Leganishu).

The geoelectric boundary of the rift in the upper crust appears in the Irkiba area. The sharp boundary in the model resulted from the discrete approximation of the earth used in finite difference modelling; the high resistivities (>4000 Ω.m) underlying station Morijo may be an artefact of inversion. Below 7 km depth, moderate to high resistivities (=400 - 1000 Ω.m) compared with the resistivity distribution east of the area, suggest that the lower crust geoelectric boundary of the Rift Valley lies further to the east, coinciding with its surface expression (Nguruman Escarpment). Evaluation of the fit for the stations Morijo and Irkiba shows that the shallow structure is better resolved for both sites. At lower frequencies the fit appears satisfactory for the TM mode, but the TE mode is not well resolved. The model structure at station Irkiba simulates abrupt variation in \( \phi_{TE} \) and in the slope of the \( \rho_{TE} \) for the very low frequencies. A likely explanation for the discrepancy is the 3-dimensionality of the intensively faulted area indicated by the data characteristics and the strong topographic
influence of the escarpment. Three dimensional modelling would be appropriate in studying such an area.

Low to moderate model resistivities at the three sites located on the floor of the Rift Valley imply about 4-5 km of fairly conductive (<10 to >50 Ω.m) material in the rift zone. The high frequencies illustrate 1-D conditions at all the sites and especially Singleraine. The model response fits the observations satisfactorily across the whole frequency range. At the Magadi station the two polarizations exhibit a high peak in $\rho$ and an inflection in $\phi$ for frequencies 1-10 Hz incompatible with the rest of the curves and with the observations at the nearby stations. This abrupt variation in the curves is likely to be due to cultural noise induced in the data from the neighbouring industrial centre. Since this anomaly appears only at this site, these frequencies were excluded from the inversion.

A distinctive lower crustal feature in the eastern part of the Rift Valley is the conductive zone that rises to depths of ≈17-20 km underneath stations Magadi and Singleraine, and seems to be connected with highly conductive upper crustal horizon east of Singleraine. The ≈7 km thick shallower conductive horizon dominates the upper crust of the area and appears to thicken eastwards. The lack of observations in the east flank of the rift created a big gap in the model with station Selengei positioned ≈60 km east of the rift. Unlike the western flank where a clear geoelectric boundary was found to correspond with the surface and geological expression of the rift, there is no clear indication of a boundary in the east flank. A possible geoelectric margin may be present in the Selengei station.

At Selengei, a 25-30 km wide resistive (>600 Ω.m) intrusive-like feature is seen at depths of 15-20 km, flanked by moderately resistive (100-400 Ω.m) zones. This resistive block becomes evident near $10^{-2}$ Hz frequency where the phase values fall to around 45° and the apparent resistivity curve becomes flat for the TE mode. In an attempt to evaluate the resistivity distribution of this locality, a comparison was performed with the responses from station Lukenya (see Appendix C; page C.11), located about 30 km NNW of Selengei. This site has not been included in the model since it was away from the trend of the line. In Lukenya the TE mode for low frequencies (<$10^{-2}$ Hz) has an essentially flat apparent resistivity curve (=100 Ω.m) and the phase falls to 45° following a similar trend with the $\phi_{TE}$ of Selengei. Similarities between the two stations also showed in the TM mode confirming the consideration of the 2-dimensionality of the area. Finally, at station Chyulu North the discrepancy between observed and calculated data is likely to be caused from the 3-D structure of the Chyulu Hills chain that is lying farther to the east, and the assumption of the mesh design that in the boundaries of the model the structure is extending to infinity.
Figure 5.6: Cross section of the geoelectric structure derived from 2-D inversion modelling for the southern part of the Kenya Rift Valley.

A resistive (≥1000 Ω.m) upper crust overlies a conductive (=100 Ω.m) region in the west of the rift. High conductivities (<50 Ω.m) appeared at mantle depths in two belts in this section. One to the westernmost part, and another one underneath Oloololo escarpment. The electrical signature of the rift with low crust resistivities starts to appear beneath site Irkiba in the west and becomes clear below all the sites located on the rift floor. The eastern electrical margin is not well defined. A mid-crustal conductive horizon east of the rift is likely to be linked with the conductor beneath stations Magadi and Singleraine, which is bordered to the east by a resistive zone beneath Selengei.
Figure 5.7: Fit of the 2-D response curves to observed data for the Southern Kenya Rift Valley. The apparent resistivity and phase information are presented for the two modes (TE and TM) at each station. The error bars in the data are 1 standard deviation. The sites are presented in geographical order from the west (Nyamanga) to the east (Chyulu North) end of the line.
Chapter 5

Modelling of Kenyan Data

Station: Oloololo

TE-mode

TM-mode

Frequency (Hz)

Phase (°)

Station: Keekorok

TE-mode

TM-mode

Frequency (Hz)

Phase (°)
Chapter 5 Modelling of Kenyan Data

Station: Leganishu

TE-mode

TM-mode

Station: Morijo

TE-mode

TM-mode
Chapter 5 Modelling of Kenyan Data

Station: Irkiba

TE-mode

TM-mode

Station: Nguruman

TE-mode

TM-mode
Station: Selengei

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Station: Chyulu North

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Chapter 5 Modelling of Kenyan Data

5.2.4 Qualitative assessment of uniqueness of the 2-D inversion results

An important adjunct to numerical inversion is the study of the resolution properties of the final model parameters (Jupp and Vozoff, 1975; 1977). This section will concentrate on estimating the uniqueness and sensitivity of the 2-D inversion results to the regional scale features that the models revealed.

It has not been advocated that the model produced by the inversion is unique, but rather that some certain features in the model are important for fitting the observed data. In order to evaluate the significance of the structures derived from the inversion, three different qualitative controls have been applied to the final model. These controls were targeting: (i) the range of the resistivity distribution of different units across the profile, (ii) the location of the major lateral boundaries, and (iii) the depth of the main horizontal discontinuities. These controls have been performed separately in the two previously inverted sections (western and eastern segment).

**Resistivity variations**

In order to estimate the resistivity range of the main geoelectrical units, each model was divided into several blocks (42 for the western and 34 for the eastern section) of specified resistivities. A resistivity value representative of the area was assigned to each block. After changing the resistivity of a single block, a forward calculation was performed and the $rms$ error of the model was stored, while the rest of the blocks were kept in their initial values. Each block was allowed to vary 6-12 times (equally spaced in logarithmic scale) in a resistivity range of approximately 4 decades around its optimal value (e.g. 1-5000 $\Omega$.m for a block that originally was set to 100 $\Omega$.m). Not all of the blocks were examined with this method. The near surface blocks were kept constant. Finally, 25 blocks were adjusted in the western and 24 in the eastern section. A simple algorithm was developed that allowed the whole procedure to be automated. The apparent resistivity error in the data is about 8 - 20%. A change of $\pm$ 5% in the total $rms$ error of the initial models was considered sufficient to describe the resistivity range of each individual block. Since some of the zones that were tested made little contribution to the total $rms$ error (e.g. small sized blocks, enhanced error in the data), the resulting range for a given block was found to be quite broad. In such cases, additional controls were applied; (inspection of the $rms$ error for the individual stations located in the area), and if these controls were not sufficient then the observed and calculated curves were plotted for visual inspection. Similar tests were also performed whenever the proposed range appeared to be quite distinctive from the calculated value from the 2-D inversion. The final results of this procedure appear in Figure 5.8.
Figure 5.8: Resistivity range of the main geoelectrical units for the profile of the southern Kenya Rift Valley.
Comparing the Figures 5.6 and 5.8, qualitative measures of the uniqueness of the inversion model can be derived. The general geoelectric structure remains unchanged. Some of the units that were not clearly imaged in the inversion procedure became more dominant, and others were minimized. West of the Rift Valley the main points are:

- The mantle low resistivity block below station Nyamanga has been replaced by a more resistive unit of 100 - 400 $\Omega$.m. The range was quite broad; resistivity values as high as 1000 $\Omega$.m caused insignificant changes in the $rms$ error, and they did not change the fit of the data for site Nyamanga. Nevertheless, increase of the resistivity in this area more than 400 $\Omega$.m generates large misfit in the LMT data for Migori and smaller changes to those of Kehancha and Oloololo. In addition, the mantle blocks situated underneath Migori and Kehancha at depths of over 32 km were kept at relatively similar levels compared to the initial inversion results.

- The existence of the highly conductive zone below station Oloololo was emphasized. The range of the resistivity remained the same as in the initial inversion. Attempts to reduce or to remove this zone caused large misfit for frequencies below $10^2$ Hz to the Oloololo and the surrounding stations.

- The presence of the conductive fault-like zone in the area between Lolkoroin and Oloololo site was confirmed with resistivity ranges 100-500 $\Omega$.m. The lower crust and upper mantle underneath station Keekorok are not well resolved from the model, the resistivity range is broad, varying from 400-1500 $\Omega$.m. The noise-degraded AMT data in Keekorok have contributed to this uncertainty. The lower crust - upper mantle in station Leganishu is clearly imaged with resistivity values of 200 - 600 $\Omega$.m. At station Morijo, the range of the resistivity remained high, with the higher limit set to 50000 $\Omega$.m for the entire crustal thickness. Nevertheless, the large misfit in the TE mode for this station limits the accuracy of the proposed resistivity range.

Inside and east of the rift, small deviation from the final inversion results occurred, probably due to the use of a simplified version of the model, and the large size of the resistivity blocks. In this part of the profile the main points can be summarized as follows:

- Underneath Irkiba site the resistivity of the upper crust remains at low values of 10-50 $\Omega$.m, and the mid-crustal unit increases to resistivities of average 2000 $\Omega$.m. Underneath stations Nguruman and Magadi the resistivity range stays similar to the optimal one (100-400 $\Omega$.m). High conductivities in the upper crust reduce the penetration depth and the lower crust (>20 km) is not well resolved. For the western part of the rift (stations Irkiba and Nguruman)
the resistivity range is very broad (100 - 2000 Ω.m), the aberrant behaviour (for the lower frequencies) of the calculated TE mode in Irkiba add to the uncertainty. In stations Magadi and Singleraine the lower crust remains conductive, but the data do not provide sufficient information for better resolution.

- At the eastern boundary of the rift near Singleraine site, the deep conductor appears not to extend to shallower depths or to be linked with the low resistivity horizon to the east. Lack of observations for the area between Singleraine and Selengei limits the uniqueness and the adequacy of the model in this zone.

- The resistive unit that breaks the lower crust below the Selengei site remains and it appears to extend eastwards beneath Chyulu North, where the resistivity appears higher (200 - 500 Ω.m) compared to the inversion results.

**Lateral variations**

It has been previously stated that a major problem during this survey and the interpretation of the results was the wide spacing between the stations. In an attempt to define the most likely position of the main vertical boundaries and assess the lateral variations of horizontal invariant areas, as constructed from the inversion model, a similar approach with the resistivity analysis (multiple forward calculations with chain alterations of the location for the proposed vertical boundaries) was performed. However, since the spacing between the recordings is about 20 km this analysis has not proved elucidating.

In the westernmost part of the profile the low resistivity mantle block is limited direct underneath station Nyamanga and west of it. The area between Migori and Lolkoroin appears almost laterally uniform. The vertical limits of the mantle conductor in Oloololo area are not well resolved, and can broaden without affecting the misfit of the model. The western geoelectric border of the rift, is likely to located 5-8 km west of the Nguruman station. The zone below the rift floor appears almost laterally uniform with small changes, mainly caused by shallow local structures. The deep conductor to the east of the rift (underneath site Singleraine) appears to extend westwards. Analysis of the area between Singleraine and Selengei has not resulted in a more accurate limit of the conductor to the west and the resistor to the east.

**Depth variations**

In the final 2-D model for the southern Kenya Rift Valley (presented in Figure 5.6) a vertical resistivity sequence has been identified rather than appropriate horizontal boundaries. However, for the qualitative assessment and to facilitate the description of the model the term
of "horizontal boundaries" has been adopted to describe the approximate depths where the transition to different resistivity sequences occurred.

Following similar operation with the analysis for lateral variations, the main horizontal boundaries of the model were examined. To the west of the rift three main horizontal electrical discontinuities are apparent in the model: (i) the \(=10\) km boundary between the resistive upper crust and the conductive mid-crustal zone, (ii) the \(=20\) km boundary between the mid-crust and resistive lower crust, and (iii) the base of the lower crust at about \(30\) km depth. Inside and east of the Rift Valley the main horizontal discontinuities appear around 5, 12 and 20 km.

Analysis of the west horizontal discontinuities indicates that the 10 km boundary may lie between 8 and 13 km, and the deeper boundary (20 km boundary) could extend to a maximum depth of 24 km. The base of the resistive lower crust varies from 28 to 33 km depth. In the eastern part of the profile the 5 km discontinuity may lie between 3 and 6 km, and the deep end of the conductive horizon could extend to 16 km. The 20 km discontinuity is not well resolved by the data, it can be placed at shallower depths of 15 km, and can be shifted down to 25 km without noticeable alterations to the total \(rms\) error.

The analysis was extended also to three other local horizontal discontinuities. The top of the deep conductive block at Oloololo could be between 27 and 35 km depth without increasing the misfit of the data. The top of the 20 km deep conductor beneath Singleraine could be shifted to 10 km depth and be linked with the upper crust horizon, but the misfit of the TM mode for the low frequencies has been increased at stations Magadi and Singleraine. Finally, the top of the resistor in Selengei could be shifted to 24 km depth without affecting the fit between the observed and calculated data.

### 5.3 Chyulu Hills modelling

Southeast of station Chyulu North the profile was orientated in a NW-SE direction, almost parallel to the strike of the Chyulu Hills chain. The Chyulu Hills volcanic field consists of several hundred volcanic cones and lava flows resting on a basement peneplain with an average elevation of about 1000 m. The unrotated apparent resistivity and phase curves (see Appendix A; pages A.15-A.19) indicate a one dimensional structure for the high frequencies but the character of the curves for lower frequencies (<0.1 Hz) suggest a 3-D structure with strong discrepancy between the two polarizations (XY and YX). The induction arrows are relatively small and unstable (see Appendix B, pages B.8-B.10), precluding their interpretation in terms of regional, two-dimensional, lateral conductivity contrasts alone.
Since the trend of the line was parallel to the strike direction and the area shows 3-D characteristics, it was considered that the use of the effective impedance was more convenient for a simple modelling interpretation. As for the main line across the Kenya Rift Valley, a pseudosection of the invariant apparent resistivity has been created. The data were mapped onto their approximate depths via the Niblett-Bostick transformation to give a more realistic structural pattern (Figure 5.9). Low resistivities in the upper and lower crust and the poor quality long period data at sites Chyulu West and Chyulu 2 reduced the penetration depth to about 15 km. The figure shows a resistive (1000 $\Omega$ m) upper crust underlain by a more conductive (10 $\Omega$ m) lower crust. At station Chyulu 2 the conductive layer is elevated to shallow depths. Large separation between Chyulu South and Mwatate sites limited model resolution.

![Figure 5.9: Chyulu Hills invariant apparent resistivity section. The resistivity values have been contoured on a logarithmic scale](image)

Additional to the contour pseudosection of the average invariant, 1-D inversion modelling was performed and the results were collated to form a geoelectric cross section (Figure 5.10). As shown in this figure a three layer structure describes the Chyulu Hills area. The resistive topmost layer (=2000 $\Omega$ m) is found to be interrupted at station Chyulu 2 by a low resistivity
zone that rises close to the surface (1 km). Underlying this layer is a less resistive (10-20 Ω.m) zone which is itself underlain by a highly conductive basal layer. The dominant feature across the whole section is the highly conductive zone (0.1 - 3 Ω.m) that extends from Chyulu West to Mwatate (suggesting a possible zone of partial melts), resting on a moderate resistive lower crust (100-200 Ω.m). However, the poor quality of the LMT data reduce the integrity of the assumption of a resistive lower crust.

Figure 5.10: One dimensional geoelectric cross-section of the Berdichevsky average invariant along the Chyulu Hills.

In an attempt to produce a more realistic geoelectric model for the area the impedance tensor has been rotated into a direction of 40° west of north (TE mode). This rotation angle is in accord with the strike direction of the Athi Ikutha shear zone that crosses the area (Shackleton, 1986; see also Figure 1.5), and the average electrical azimuth (see section 4.4). Similar tactics for the mesh design, as have been described earlier, were adopted. Two models are presented, the first one (Figure 5.11a) is the result of the 2-D inversion of the TE mode. This model served as the initial model to a second inversion. In this second step (Figure 5.11b) both TE and TM modes were inverted. The fit between the calculated and observed data was sufficient for both models (Figure 5.12 presents the results from the joint inversion of TE and TM modes).
The results are quite similar to the 1-D model, the main differences observed are in the mid-crustal layer where the resistivity distribution was increased to about 25 $\Omega$.m, and its depth to base was shifted down to 16 km. A main discontinuity arises at about 10 km depth where the base of the top conductor is located. The model of the joint inversion of the TE and TM mode revealed a resistive structure between the Chyulu South and Mwatate stations. The large separation between these two sites decreased the resolution of this structure, but it proved essential to fit the TM mode for these two sounding points. However, as has been mentioned, the character of the area is 3-D and in order to obtain a reliable model for the region a higher inversion scheme would have to be applied.

Figure 5.11: Two dimensional models for the area of the Chyulu Hills. The impedance tensor was rotated W40°N. (a) Only the TE mode was inverted, since the strike is parallel to the profile, (b) TE and TM modes were jointly inverted, the initial model for this inversion was the resulting model from the TE mode modelling.
Figure 5.12: Fit of the 2-D response to observed data for the Chyulu Hills MT profile. The results are from the joint inversion of the TE and TM mode. The error bars in the data are 1 standard deviation. The sites are presented in geographical order from NW to the SE.
Chapter 5 Modelling of Kenyan Data

Station: Chyulu North

TE-mode

TM-mode

Station: Chyulu West

TE-mode

TM-mode
Chapter 5 Modelling of Kenyan Data

Station: Chyulu 2 (Range)

**TE-mode**

**TM-mode**

Station: Chyulu South

**TE-mode**

**TM-mode**
Station: Mwatate

TE-mode

TM-mode

App. Res. (Ω.m)

Phase (°)

Frequency (Hz)
Chapter 6

Model Appraisal - Interpretation

The results of the magnetotelluric experiment in the southern Kenya Rift are appraised in the light of the available geological and geophysical data. The discussion aims to integrate the results of various geophysical data to unravel the deep structure of the study area and to produce a speculative model for the present state of the rift in southern Kenya. The possible causes of the observed crustal resistivity distribution are discussed based on the general present day knowledge of the electrical structure of the crust, and on the concept of an upwelling mantle plume beneath the Kenya Rift. As the inferences are drawn from models with insufficient data coverage and simplified models, any conclusions must be treated with some caution.

6.1 Comparison with previous geophysical observations

In many respects the simplified 2-D geoelectric model (represented in Figure 6.1) is in accord with the results obtained from previous geophysical studies in the region.

6.1.1 Comparison with geoelectric models

A comparison of the results of the current MT experiment with the previous electric studies near the equator can provide a comprehensive picture of how the geoelectric structure of the rift varies along the axis. This survey is about 250 km south of the location of the previous measurements. Rooney and Hutton (1977) and Banks and Beamish (1979) from magnetotelluric and magnetovariational studies showed that low conductivity zones exist within and east of the rift, but they did not identify any conductivity anomaly to the west of the rift (see Chapter 1). The current study found the low resistivity zones inside the rift and their extension eastward, with the west boundary being highly resistive. A very low resistivity near-surface (top 5 km) layer suggested from all the surveys possibly represents the rift infill of conductive sediments. Moderate resistivities (=50 Ω.m) in the middle crust and low resistivities (<10 Ω.m) in the lower crust-upper mantle below the rift to depths greater than 15-20 km, suggested by the previous and the current investigations stress the general 2-D character of the rift along its axis. However, a deep conductor to the east of the rift shoulders inferred by Beamish (1977)
and Banks and Beamish (1979) was not found in the present investigations which has poor coverage in the east. Additionally, the conductivity anomaly detected in this study underneath Ooololo Escarpment was not suggested by the previous studies.

Figure 6.1: Simplified 2-D geoelectric model for the southern Kenya Rift Valley. The traverse coincides with the seismic profile, the differences in the distance are due to projection of the stations on different lines. Only the main features are emphasized here.

6.1.2 Comparison with seismic models

The main reason for the selection of the location of the present MT profile was to coincide with previous seismic profiles in the region (KRISP 94, Birt, 1996; Novak et al., 1997a). The two techniques essentially sound different material properties of the crust (velocity-density in seismic and conductivity in MT) at widely differing spatial scales (Beamish and Smythe, 1986).
Therefore, independent images for the earth's interior can be provided from the two methods. If the results from the two techniques are combined, a more comprehensive image for the crust can be obtained (e.g. Jones, 1981; Beamish and Smythe, 1986; Gough, 1986; Jones, 1987; Merzer and Klemperer, 1992; Booker et al., 1996). Previous studies have shown that there is a general trend of decreasing velocity with decreasing resistivity (Marquis and Hyndman, 1992).

It is accepted that the low resistivity (<100 $\Omega$.m) zone in the middle-lower crust correlates with the low velocity (<7 km/s) layer (Hyndman and Klemperer, 1989), and reflective zones correspond with conductive layers (Hyndman and Shearer, 1989). Approximately, "normal" continental crystalline rock is interpreted as transparent / resistive and fluids in inter-connected pores as reflective/conductive (e.g. Gough, 1986; Jones, 1987). However, reflective / resistive zones, for which mylonites or layered intrusives may be acceptable explanations, have been identified, along with transparent / conductive layers (Jones, 1987).

The seismic models of KRISP 94 (Birt, 1996; Novak, 1997) are shown in Figure 6.2. The general parallelism between some of the features of the seismic and geoelectric model, for the main line, is visible. To the west of the rift the upper-crustal bottom boundary at about 10-12 km depth, an underlying layer with basal boundary at about 20 km in some zones and the lower crustal boundary at 30 km are apparent in both models. Inside and east of the rift valley, the depth of the rift basin and the mid-crustal discontinuity are not well correlated in both models. A more detailed comparison follows.

To the west of the rift, the cover layer in the seismic model has been modelled as a thin (<1 km) low velocity (~4 km/s) sequence at the surface with variations in basement velocity (~6 km/s); this layer becomes thicker (average 1 km) between the Oloololo Escarpment and the edge of the rift. The electrical model for the area did not identify an equivalent uniform structure. The resistivity distribution across the region deduced from TEM and high frequency AMT data describes a more complex near surface layered structure with moderately resistive (~1000 $\Omega$.m) and more conductive (~200 $\Omega$.m) thin layers resting on resistive (>1500 $\Omega$.m) crystalline basement. The rift basin itself has a P-wave (average) velocity of 4.55 km/s and is asymmetric, being deepest (4 km) in the western margin adjacent to the Nguruman Escarpment. The high conductivities (<10 $\Omega$.m) found on the rift floor coincide with the low velocities. However, the MT model did not identify a rift basin similar to the seismic section, and only to the west did the transition from a low to a higher resistivity zone at depth of about 5 km seem to correlate with the seismic results. To the east of the rift zone, the increase in velocity (5.7-6.0 km/s) approaching normal basement values corresponds to the changes of the resistivity from conductive sediments to normal resistive upper crust.
The seismic interface at about 12 km depth that is imaged from the KRISP 94 and from previous seismic refraction work further to the north (Maguire et al., 1994; see also Figure 1.7) is clearly imaged in the geoelectric model as the boundary between the upper resistive crust and the conductive middle-lower crust. The correspondence is more distinct in the region west of the Keekorok station. A second seismic reflector at 17 km depth, deeping eastwards and disappearing beneath the Rift Valley, follows the trend of the discontinuity between the conductive and the highly resistive lower crust zones of the MT model at a depth of about 20 km. The velocity beneath the first interface varies from 6.35 to 6.42 km/s, and increases beneath the second boundary, varying from 6.54 to 6.66 km/s. The argument for correspondence between decreasing velocity and decreasing resistivity seems to apply to this sector. However, the similarity between the models over this interface is limited in the area underneath and east of the rift where MT suffers from insufficient spatial coverage. Inside the rift, the top of the highly conductive body at 18 km depth may correspond to the point where the
second interface pinches out, but such a hypothesis cannot be positively established due to
uncertainties in both of the models. To the east of the rift, the top of the basal resistive block
beneath station Seiengei (at about 17 km depth) can be regarded as the mid-crustal reflector,
since the increase of the velocity coincide with increase of resistivity.

The lowest crustal boundary at 31 km in the west, shallowing to 22 km in the east,
indicating the transition from low (< 7 km/s) to higher velocities is in good agreement with a
continuous boundary in the geoelectric model stretching from the west end of the line to the
western flank of the rift where it also indicates the transition from the highly resistive upper part
of the lower crust to a more conductive basal sector. Limited depth penetration from MT inside
and east of the rift have led to poor model resolution at depth. Finally, the seismic Moho
marking the transition to the upper mantle is located at a depth of 33 km at the western edge,
reaching 35 km beneath the eastern shoulder of the rift, and deepening to 40 km at the eastern
end of the profile. The thinning of the crust and a localized low velocity of 7.8 km/s directly
beneath the rift axis has been interpreted as the seismic signature of the rift, but this anomaly is
not well constrained (Birt, 1996). No coincident electrical resistivity expression was identified at
the same depth but it may be considered that the shallower boundary at 30 km in the MT model
marks the top of the upper mantle.

From the above description it is evident that the two methods have recovered various
compatible features. The seismic model has described better the deeper structure, especially in
the area of the rift valley and east of it, and has delineated the eastern boundary of the rift
better than the electrical model. Nevertheless, the signature of the MT model is quite distinctive
to certain lateral variations. The resistivity anomaly in the area of the Oloololo Escarpment is
possibly an important geological feature and emphasizes the contribution of MT to the deep
structure of the area. Additionally, the western margin of the rift is clearly marked in the MT
model as a strong lateral discontinuity. Inadequate station spacing east of the rift margin have
limited the resolution of the geoelectric model but the MT method still succeeded in setting a
plausible lateral limit (beneath station Seiengei) to the rifting extension.

6.2 Interpretation and implications of the Kenya Rift model

The discussion in this section will draw upon the 2-D modelling results of the MT data from the
main traverse across the southern Kenya Rift Valley (Figure 6.1). A main feature of this model
in the west Nyanza Craton is the presence of a "fault-like" zone in the area of the Oloololo
Escarpmnt, and a deep mantle conductor in the same area. A sharp boundary in the west
margin of the rift valley, and low resistivities on the floor of the rift are notable features inside
the rift. East of the rift more resistive upper crust and a highly resistive segment in the middle to
lower crust are the main structural characteristics. The lower crust - upper mantle is imaged by the data only to the west of the rift.

6.2.1 Upper Crust (top 10 km)

A very thin veneer of conductive sediments (<50 Ω.m) in stations Nyamanga and Migori, east of Lake Victoria is suggested by TEM inversion down to depths of 100 m. The sediments overlie Archaean basement, part of the Nyanza craton, of high resistivities (>1000 Ω.m) typical in a stable upper crust (e.g. Haak and Hutton, 1986). The complex upper crustal resistivity structure between the lake shore and the Oloololo Escarpment, with oblique conductive (~100 Ω.m) fault-like zones is probably related to the complex geology of the Migori Gold Belt. The presence of minerals deposited in veins by mineralizing fluids, sulphide ore deposits, and graphite (Shackleton, 1946) may have resulted in the low resistivities. A series of intrusions orientated NW-SE can explain the ~35° electric strike direction (section 4.4) that the high frequency AMT data indicate for the area. The strong resistivity contrast between stations Lokkoroin and Oloololo, coincides with the surface expression of the Archaean - Proterozoic collision zone (Shackleton, 1986) at the Oloololo Escarpment. The low resistivities may be caused by meteoric fluid circulation (of higher salinity at depth) occurring in the upper crust by fracture - focused flow or topographic control, often interpreted to extend to great depths (10-20 km) within the continental crust (e.g. McCaig, 1988, 1989; Torgersen, 1990), or by graphite- and sulphide-bearing rocks, the protoliths of which were deposited in fractures between the colliding terrains (Hjelt and Korja, 1993).

The upper crust is highly resistive, in the area between Oloololo and Morijo stations. The seismic model (Birt, 1996) shows a low velocity surface layer in the area resting on a highly fractured basement, and an increase in the basement velocity from west to east - with the transition from Archaean Craton to Mozambique Belt occurring beneath the rift. Since the upper crust shows a small resistivity range globally and tectonically (Haak and Hutton, 1986; Jones, 1992), the MT results are unable to associate the upper crust in the region with the Archaean or the Proterozoic unit. The arguments that have been described in section 1.2 can be used to explain the high resistivities. Small resistivity decrease (~100 Ω.m) in zones in the upper crust can be attributed to the presence of relatively resistive (>100 Ω.m) meteoric fluids and cannot reduce the bulk resistivity further. A deformed, cracked and sheared basement caused by the collision of the two units has been proposed by Birt (1996) to explain the reduced seismic velocity (compared to normal values of 6.0-6.3 km/s); this model is in agreement with the explanations of Gough (1986) and Nesbitt (1993) for the high resistivities observed in the upper
crust and may justify the remarkably high resistivities (>20000 Ω.m) approaching the Nguruman Escarpment.

The greatest degree of static shift was revealed below Irkiba station on the top of Nguruman Escarpment and suggests the complexity of the area. Resting on the basement system rocks are Tertiary and Quaternary sediments, and volcanic rocks associated with the formation of the rift valley, and recent deposits (volcanic black cotton and red-brown sandy soils). The upper 5 km are unusually conductive (<10-50 Ω.m) including sequences of resistive (>2000 Ω.m) blocks, lying on a more resistive crystalline unit. Meteoric fluid circulation or graphite (formed from organic-rich sedimentary material) may be used to explain the high conductivities; the occurrence of the geothermal area of Narosura (Riaroh and Okoth, 1994; their Figure 1) in the region may also contribute to the low resistivities. In the northeast area, lavas are tilted gently as a result of rift faulting. Lavas and ignimbrites were extruded from a number of north-south vents and fractures, and these features may correlate with the resistive units that the model has reconstructed. The greater part of the area is underlain by metamorphosed sedimentary rocks of Precambrian age of high resistivities.

The floor of the rift valley is affected by a dense network of faults in a N-S trend and is covered by a sequence of trachyte and basalt lavas; Lake Magadi occupies an elongated depression at the eastern side of the rift basin. An obvious interpretation for the enhanced conductivity (<10 to 50 Ω.m) inside the rift valley is magma that permeates the whole of the crust beneath the rift and intruded into the sediments (Davies, 1998). However, the top 5 km of high conductivities can be possibly attributed to hot, saline fluids contained in the porous sedimentary and pyroclastic formations that may promote preferential current channelling along the rift axis. The western part of the rift floor close to the Nguruman fault, which marks the west boundary of the rift, is more conductive than the east part but, as mentioned before, fault zones are frequently cited as being conductive since they offer favourable conduits for graphitization, trapping fluids and even ascending magmas. The station density is very low to identify any structure that might be correlated with geothermal activity or volcanism at the surface. However, the low near-surface resistivities (<5 Ω.m) in area of Lake Magadi possibly relate to geothermal activity in the area which consists of hot springs (see Riaroh and Okoth, 1994). A distinctive feature in the western part of the rift floor is the resistive (200-600 Ω.m) structure that underlies the top conductive zone (at a depth of about 5 km) and appears to be linked to the resistive upper crust to the west. This resistive zone laterally shifts the onset of the rifting region further to the east than its surface expression. A smoother transition to higher resistivities beneath the east part of the rift may relate to the more gentle step-faulting on the eastern side of the valley compared with the sharp formation of the Nguruman Escarpment to the west.
Nevertheless, this may simply reflect the broader site spacing on the eastern margin. The zone beneath Selengei station is suggestive of the Proterozoic basement in the area with high resistivity values, typical of the upper crust. Low resistivities further to the east, below the Chyulu North station may correlate with the surface volcanism of Holocene to recent times in the Chyulu Hills area.

### 6.2.2 Middle- Lower Crust, Upper Mantle (>10 km)

The Archaean interior beneath the west end of the profile is a representative image of the global model of the continental crust (Figure 6.3). From Lake Victoria to Oloololo Escarpment a low resistivity (=100 $\Omega$.m) middle crust (10-22 km) lies on a highly resistive (>4000 $\Omega$.m) lower crust underlain at about 30-32 km depth by a more conductive (<1000 $\Omega$.m) deep crust-upper mantle. As previously stated the three boundaries at about 10, 20 and 30 km depth are in general agreement with seismic interfaces.

![Figure 6.3: Schematic diagram of the continental crust with a resistive upper crust underlain by a two-layer lower crust, proposed by Jones (1987). The top part of the (reflective ?) lower crust contains interconnected free water and thus is a well-conducting zone. It is divided from the upper crust by an impermeable zone, necessary to trap fluids. The temperature is about 300-400°C. A graphite film coating grain boundaries will effectively lower resistivity, and not require the impermeable zone. The crust-mantle boundary (Moho) is seldom exposed in field data by changing electrical resistivity.](image)

The transition from the upper to middle-lower crust is characterized by the change from brittle to ductile deformation style at temperatures of 300 to 350°C caused mainly by hydrolytic weakening of quartz (Jodicke, 1992). A fixed isotherm in the earth's interior (i.e. 400°C) represents the start of the conductivity increase in the lower crust (e.g. Hyndman et al., 1993). No electrical boundary coincident with the seismic Moho was observed in this study. It is
generally accepted that there is no dramatic change in conductivity across the Moho discontinuity (e.g. Schwarz, 1990; Beamish, 1990b; Jones, 1992). A general model for the continental crust describes the deeper crust as moderately resistive (=1000 Ω.m), with the upper mantle more conductive (=100 Ω.m) (Haak and Hutton, 1986; Jones, 1987).

Saline aqueous pore fluids and grain boundary graphite are the two most probable mechanism candidates for the low resistivity in the crust. The presence of fluids is supported by the model of Mooney and Christensen (1994). Mooney and Christensen (1994) propose a model for the composition of the crust in the Kenya Rift, based on a comparison of the KRISP 90 crustal velocity structure with laboratory measurements of compressional-wave velocities of rock samples. In their model, the crust comprises three primary layers whose composition changes with depth from felsic to mafic, and in metamorphic grade from greenschist to granulite facies (Figure 6.4). According to this model, on the flanks of the rift, the upper and middle crust appear to be comprised largely of Precambrian greenschist to amphibolite facies felsic-to-intermediate composition metamorphic rocks. The lower crust may consist of granulite facies mafic rocks. The upper and middle crustal layers are likely to be intruded by mafic dykes and sills beneath the rift axis. In ordinary rocks the most important hydration / dehydration reactions, in terms of water volume, take place at the boundary between the amphibolite facies and the greenschist facies (Sanders, 1991; Hyndman et al., 1993). In the Kenya environment this transition seems to correlate with the top of the conductor at about 10 to 12 km depth.

Figure 6.4: Composition of the crust beneath the Kenya Rift at the Equator from crustal seismic velocity structure and laboratory measurements of Kenyan rock samples (after Mooney and Christensen, 1994).
The presence of fluids can also be used to explain the low resistivities in the upper crust in the area. Fluids derived from lower crustal reservoirs move through fault-shear zones to the upper crust (McCaig, 1988). However, Yardley (1986) and Yardley and Valley (1997), argue that the granulite facies rocks having cooled to steady-state conditions well below their peak metamorphic temperatures, are water deficient. This would mean that the water activity is so low that they would spontaneously react with free water to form hydrous minerals, leaving a dry and therefore insulating environment (Sanders, 1991). Given the petrological arguments against fluids in stable regions (such as the Archaean unit), graphite accumulations seems to be a more likely explanation and is in accord with recent results from the KTB deep drill hole (Emmermann and Lauterjung, 1997; ELEKTB Group, 1997) which point to the presence of graphite as the more possible source for lower and upper crustal conductivities. The idea that the Archaean deep crust is especially resistive, supported by results from the Lewisian (Scotland) and Ukrainian terrains (Haak and Hutton, 1986) and from recent MT results from the Slave Craton (Jones and Ferguson, 1997), seems to apply to the deep Archaean crust in the region between Lake Victoria and Oloololo station with resistivities higher than 3000 Ω.m. The transition from the conductive middle crust to the resistive lower crust (at about 22 km depth) seems to coincide with the Mooney and Christensen (1994) transition from “wet” amphibolite to “dry” granulite facies (at about 27 km depth) and may thus be a potential explanation for the increase of the resistivity (see also Wannamaker, 1986). Dehydration by loss of fluid porosity or through successive thermal dehydration events, and breaking of graphite film interconnections with time may also provide another explanation for the high resistivities.

The resistivity structure of the middle-lower crust changes eastwards. The conductive zone in the Oloololo Escarpment may be a contact-fault zone and is consistent with the geological interpretation that the escarpment is spatially coincident with the suture zone between the Nyanza Craton and the Mozambique Belt, which collided during the Proterozoic (Shackleton, 1986). A similar interpretation has been put forward from a gravity model of the lithosphere by Nyblade and Pollack (1992). During and after such active tectonic events, fluids may be abundant in the crust and may be associated with such diverse phenomena as granulite-facies metamorphism, shear zone alkaline granites, and various kinds of metasomatism (Newton, 1990) that could have resulted in enhanced conductivities. Additionally, plate tectonic collisional processes of the Proterozoic could have increased input of carbon to the deep crust directly as organic-rich sedimentary distributions or from the mantle below via slab devolatilization (Wannamaker, 1997). However, as mentioned above, Smith and Mosley (1993) and Smith (1994), supported the idea that the area extending from the east of Oloololo Escarpment to the western flank of the rift is Archaean basement. The geoelectric model of the area cannot differentiate between the two units, although it has been recognised that there is a
significant difference between the average resistivity of the lower crust in areas of different age (e.g. Hyndman and Shearer, 1989; Jones, 1992). The differences are stronger between Proterozoic and Phanerozoic, but it has not been determined if there is a continuous variation with the age, or the appropriate age divisions (e.g. Precambrian, Archaean, etc.; Hyndman et al., 1993). According to Hjelt and Korja (1993), in old regions where post-orogenic events have deformed the lower crust (e.g. via mafic underplating), the conductive traces of collisions can be imaged only in the upper to middle crust, while in younger areas the collisional conductors may extend throughout the crust. It can be stated that the different resistivity distribution in the middle (\(\approx 2000 \Omega \cdot m\)) and lower crust (\(\approx 400 \Omega \cdot m\)) in the region between Keekorok and Leganishu stations compared to the conductive (<300 \(\Omega \cdot m\)) middle and resistive (>4000 \(\Omega \cdot m\)) lower Archaean crust to the west is evidence of different units. However, the noisy and scattered data at Keekorok station cannot justify the uniqueness of the model.

Between Oloololo and the rift flanks the structure is considerably more complex; a resistive (\(\approx 2000 \Omega \cdot m\)) middle crust and an ill-defined moderately resistive (\(\approx 400 \Omega \cdot m\)) lower crust, bordered by a wide (50 km), thick (>30 km), highly resistive (>5000 \(\Omega \cdot m\)) crustal unit directly adjacent to the rift. Underneath Morijo and Irkiba stations, the middle and lower crust are exceptionally resistive. The fit of the model for these two stations is not sufficient, especially for the low frequencies, and therefore any interpretation has to be treated with caution. Assuming the model of Smith and Mosley (1993) and Smith (1994) that the collision between Archaean and Proterozoic units coincides with the location of the rift, a possible interpretation of the high resistivities in the west margin of the rift is that the highly fractured and reworked rock matrix which resulted from the Proterozoic collision probably disrupted the continuity of graphite films (grain-boundary graphite acts as conductor only if the film is continuous; Frost et al., 1989), with the result that the conditions in the middle-lower crust become similar to the brittle, resistive upper crust. Nevertheless, it is apparent that the influence of the rifting processes extended beyond the limits of the rift; high temperatures and pressures may have altered the bulk composition and resistivity properties of the crust west of the rift. Rock porosity is progressively closed by the increase in effective pressure and the electrical resistivity increases (Marquis and Hyndman, 1992). The seismic data for the western flank of the rift have not recorded any dramatic change, but resistivity is the physical rock parameter that is more sensitive to various tectonic and composition factors (Haak and Hutton, 1986).

Durrheim and Mooney (1991; 1994) (Figure 6.5) describe the crust in Archaean provinces to be about 35 km thick, whereas Proterozoic crust has a significantly greater thickness of about 45 km and a substantially thicker high-velocity (>7.0 km/s) layer at the base, probably representing predominantly mafic rocks. In the western part of the profile a high velocity layer at
the base of the crust correlates with a low resistivity zone in the MT model at a depth of about 30 km. The presence of free aqueous fluids (Marquis and Hyndman, 1992) or mafic intrusions of partial melt derived from the upper mantle (Newton, 1990; Singh and McKenzie, 1993) may cause the reflectivity and the low resistivity. Birt (1996) explains the high velocity material at the base of the crust as magmatic additions, and its overall thickening towards the east as reflection of the change from Archaean to Proterozoic crust based on the model of Durrheim and Mooney (1991; 1994).

Figure 6.5: Model for Archaean and Proterozoic crustal evolution emphasizing the differences in the chemical properties of the uppermost mantle (from Durrheim and Mooney, 1991). Archaean crust developed above initially hotter mantle; magmatic underplating, if present, is ultramafic and is seismically indistinguishable from normal mantle. A cold lithospheric keel may also act as a thermal boundary against crustal underplating by basaltic melts from asthenosphere. Proterozoic thicker (45 km) crust develops above cooler, fertile mantle that is the source of basaltic underplating.

The presence of magmatic underplating can also explain the low resistivities. The mafic magmas underplating the crust may generate silicic magmas by secondary melting of overlying rocks; this heat dehydrates greenschist and amphibolite grade rocks producing more hydrous fluids that flux additional silicate melt (Brown, 1994) and consequently enhancing the conductivity. Moreover, Lee et al. (1983) measured conductivity of mafic rocks (amphibolites and mafic gneiss) in the laboratory and found remarkably high values; a conduction mechanism related to the presence of hydrous silicates such as chlorite was proposed to justify the high conductivities. Rising partially molten mantle diapirs can also explain the very low resistivity in the area of the Oloololo Escarpment. Small magma blobs generated from an upwelling thermal mantle anomaly beneath the rift axis may have exploited lithospheric weakness and fractures (such as the suture zone between the Precambrian units) and migrated upward in the Oloololo area. A hypothesis of a mantle plume circulating below the western flank of the rift (Ebinger,
1989) or an older plume located under Lake Victoria (George et al., 1998) generating the low resistivities cannot be supported by this model.

Inside and east of the Rift Valley, thick sedimentary sequences and high conductivities in the upper crust mask the features of the lower crust. The resolution of the model is limited to the top 20-25 km. To the west of the rift, the resistive (200-600 Ω.m) zone described in the upper crust appears to have deeper roots in the middle and lower crust. The rest of the rift crust is moderately conductive (<100 Ω.m) with sections of high conductivities (<10 Ω.m). Fluid exsolution from cooling mantle derived magmas or partial melts is the most likely explanation of these low resistivities. In the rifting environment of Kenya the fluids will be mostly thermally driven (Frost and Bucher, 1994). At deep crustal levels the fluids may ultimately be of mantle origin transferred to the crust via magmas (Frost and Frost, 1987). At shallower depths aqueous fluids of meteoric and magmatic origin may dominate. Rifting events may have created pathways for fluid flow, but because extensional faulting moves wetter (and cooler) upper crust over drier (and hotter) lower crust (Frost and Bucher, 1994), the driving force for the fluid flow is likely to be the strong thermal gradient found in the region (Figure 6.6).

**Figure 6.6:** Petrologically based model for the regime of fluid flow in a continental rift environment (from Frost and Bucher, 1994). In the continental rifts the fluid flow is thermally driven.

Partial melting as a cause of the high conductivities in the middle and upper crust in the region is unlikely since temperatures of over 700°C (e.g. Schwarz, 1990) would be required at depths of few km. According to Mechie et al. (1994b; 1997) the estimated temperatures at Moho depths beneath the rift are 630°C at 21 km depth beneath the northern part of the rift and 1000°C at 35 km depth in the southern part corresponding to heat flow value of 105±51 mW/m². Moreover, water is a more probable source of the low resistivities compared to the partial melts because it may be removed from a much larger mantle volume and over a much larger area than magma, since it is more mobile than the partial melts (Hyndman and Shearer, 1989). The
seldom conductive (<10 \(\Omega\).m) zone beneath the east-central part of the rift may be correlated with hot mafic dykes and sills that Mooney and Christensen (1994) described in their model.

To the east of the rift, the resistive (=1000 \(\Omega\).m) crustal diapir-like feature may be interpreted as a remnant of the Proterozoic crust. Whether this feature is just an isolated block detached from the resistive upper crust or a deep root zone of the lower crust, cannot be accurately determined from the present data. 1-D modelling of the Lukenya station further to the north recovered a similar resistive block (3000 \(\Omega\).m) extending to depth of about 23 km, overlying a more conductive (<100 \(\Omega\).m) zone. Assuming a 2-D character for the area it can be accepted that the block underneath Selengei station is also underlain by a more conductive zone, describing a general continental crust, with a resistive upper part and more conductive lower section. However, this assumption is debatable since the separation of the two stations is about 20 km and the assumption for the 2-D character in such a complex environment cannot be always accepted.

6.2.3 Chyulu Hills

The 1-D and the 2-D models for the area of the Chyulu Hills have revealed a top 3-5 km resistive upper crust (=1000 \(\Omega\).m) underlain by a highly conductive (<20 \(\Omega\).m) zone down to a depth of about 15-17 km resting on a moderately resistive (>100 \(\Omega\).m) basement. The Chyulu Hills are located on a major shear zone within the Pan-African Mozambique Belt (Smith and Mosley, 1993; see also Figure 1.5). This shear zone can be regarded as a pre-existing structure that could have facilitated magma ascent to the surface, and resulted in the observed low resistivities.

A comparison between the MT models (Figure 5.11) and the seismic line (Figure 6.2) (Novak, 1997) reveals similar features. The near-surface low resistivity region beneath station Chyulu 2 (Range) can be associated with the pyroclastic material in the area, and coincides with the low velocity feature from the seismic profile. The basement velocities of 6.0-6.2 km/s at depths of few km from the surface seem to harmonize with the resistive upper crustal structure which correlated with Proterozoic metamorphic rocks (Omenge and Okelo, 1992). The high conductivities have limited the MT penetration depth and mask the structure of the deeper layers. A low-velocity zone beneath the Chyulus was attributed to the presence of 2-5% of partial melt (Novak et al., 1997b) and elevated temperatures (Ritter and Kaspar, 1997) and interpreted as a sign for the active magmatic processes related to the rifting in East Africa; the MT data failed to penetrate these depths. Novak (1997) interpreted the lower crustal low-velocity anomaly as small magma bodies segregated from the asthenosphere-lithosphere boundary which ascended to basal crustal levels where they stagnated and assimilated lower
crustal material. The top of the resistive layer at about 16 km depth may correlate with the upper boundary of the lowermost crust at a depth of about 18 km in the seismic model. The conductive blocks in the upper 10 km can be associated with the active volcanism in the area (Shaitani Volcano) and the presence of a small volume of molten magma in active chambers, or active hot dyke systems that facilitate the uprising of "hot" material from the magma chambers (responsible for the low velocity zone) in the lower crust to the upper crust.

According to the 2-D MT model the area between the Chyulu South and Mwatate station is likely to be resistive. The area lies entirely within the Mozambique Orogenic Belt and consists of high grade crystalline rocks with a sedimentary cover (Pohl and Niedermayr, 1979). The resistive region may reflect the Proterozoic crust and clearly separates the Chyulu Hills from the south-east part of the Belt. This hypothesis in combination with the resistive zone beneath Selengei station limits the extension of the conductive region inside the topographic and geologic expression of the Holocene volcanic field, and suggests no direct connection to the Gregory Rift. This conclusion is compatible with teleseismic tomograms from the area (Ritter and Kaspar, 1997) that present evidence for small magma chambers in the lower crust and uppermost mantle strictly located beneath the Chyulu Range. The geoelectric model also seems to coincide with the flank volcanism model of Karson and Curtis (1989, their Figure 4) which is also supported by Macdonald (1994). In these models it was speculated that the heat for melt production in the Chyulu area originated from the upwelling asthenospheric body that causes the rifting (Achauer et al., 1994); small volumes of melt propagate through the lithosphere underneath the eastern shoulder of the rift and these small size blobs generated the volcanic field of the Chyulu Hills. However, the approach to relate the Chyulu Hills with the Kenya Rift and the upwelling mantle anomaly is not certain, since the spatial station separation cannot provide a clear link. Moreover, it is possible that the Chyulus are a feature which has to be seen in the context of the formation of Mount Kilimanjaro which is located only 40 km away.

6.2.4 Induction Arrows

The Parkinson induction arrows (see Appendix B; Figure 4.12) calculated from the AMT data appeared noisy and scattered, whilst the arrows produced from the long period data are more reliable. The real components of the arrows are generally more sensitive to regional features, thus the discussion will concentrate on their interpretation.

At the west end of the line, close to lake Victoria, the high frequency real induction arrows are probably influenced by the conductivity zones in the area of the Migori Gold Belt and their pattern is not well defined, small imaginary parts implies that the conductors are probably quite shallow. Approaching the rift valley, the real component arrows point in an almost perpendicular sense to the N-S trending rift. Close to the Nguruman Escarpment, real and
imaginary parts become anti-parallel emphasizing the 2-D character of the area. The Nguruman fault and the sedimentary fill within the rift, channelling currents in a N-S direction, strongly influence the orientation of the real induction arrows, especially at high frequencies.

At lower frequencies (< 10^{-2} Hz) the induction arrows to the west of the rift are rotated, pointing perpendicular to a conductivity boundary that lies in a NW-SE direction, and their magnitude increases towards the rift. The arrows from the stations around Oloololo Escarpment do not appear to be influenced by the conductivity anomaly beneath Oloololo station questioning its lateral extent, it is surprising that such a feature (i.e. a contact fault between different geological formations) does not have a stronger effect on the induction arrows. However, in the frequency range 0.1-10 Hz the amplitude and the azimuth of the induction vectors at station Oloololo indicate a conductivity boundary in a direction almost parallel to the escarpment. Inside and east of the rift the direction and magnitude of the induction arrows are preserved for the lower frequencies, and only in the area of the Chyulu Hills do they become relatively small and unstable at all frequencies. The pattern and the amplitude of the induction arrows in the Chyulu Hills indicate that they may be situated directly over a major conductivity anomaly. It is characteristic that the eastern margin of the rift does not seem to exert a strong influence on the induction vectors, verifying the assumption of a less acute eastern rift flank. However, it may be that the conductive NW-SE Proterozoic trend effectively masks the signature of the rift's eastern margin (see also Simpson et al., 1997).

Previous surveys by Banks and Ottey (1973), Rooney and Hutton (1977), Beamish (1977) and Hutton et al., (1989) have also produced SW-NE to SSW-NNE trending induction arrows. These observations support previous conclusions of some uniformity along the rift valley, in a grossly heterogeneous 3-D environment. Simpson et al. (1997) suggest as a possible explanation of the principal direction of all these arrows, a second principal strike direction controlled by the pre-rift NW-SE trending Proterozoic fault fabric. Smith and Mosley (1993) and Smith (1994) proposed that the rifting was initiated above a basement shear system marking the contact between the Archaean Craton and the Proterozoic Mobile Belt and was influenced by a frame-work of large NW-SE trending shear zones; these lineaments provided weak conduits for the ascent of magma. Furthermore, Achauer (1994) from 3-D tomographic studies described the development of the rift to be affected from two features: a narrow wedge of asthenosphere beneath the rift in the area of the Kenya Dome, and the Proterozoic NW-SE trending Aswa-Nandi shear zone that marks the transition from the Archaean Nyanza craton to the Mozambique belt; the upwarping of the mantle flow used the old suture zone. Smith, (1994, Figure 5) correlated the position of the suture between the Precambrian units with a NW-SE overlap zone (see Figure 1.5), 75-100 km wide, containing both olivine-rich and olivine-poor nephelinitic volcanism (Olivine-rich nephelinitic volcanism is associated with the thinner
lithosphere of the mobile belts, while olivine-poor nephelinitic volcanism is restricted in East Africa to thick Archaean lithosphere). In southern Kenya the margin is located directly beneath the rift valley.

6.3 A speculative tectonic model for southern Kenya Rift

An active model of rifting provides an effective framework for the evolution of the Gregory Rift (Smith, 1994). The emplacement of a thermal mantle perturbation or a small plume at the base of the lithosphere and adjacent to the northeastern margin of the Nyanza Craton (Smith and Mosley, 1993) combined with the presence of pre-existing lithospheric and crustal heterogeneities, such as shear zones and the suture between the Archaean Nyanza Craton and the Proterozoic Mozambique Belt, were the critical driving forces for the formation, location and the development of the Kenya Rift.

It can be advocated that many features of the rifting in the Kenya Rift are present in the geoelectric model of this study. An integrated model for the present-day structure of the rift in southern Kenya has been developed based on the work of Karson and Curtis (1989), Smith (1994), Mooney and Christensen (1994), Birt (1996) and Mechie et al., (1997) and the present magnetotelluric results. Figure 6.7 shows a schematic diagram illustrating the interpretation of the electromagnetic profile across the rift, the main features of which are illustrated below:

**Rift Valley:** An ascending low density mantle intrusion, containing 3-5 % partial melt accumulates below the base of the crust and generates partially molten diapirs that separate and rise into the crust where they form mid-crustal small magma chambers and sills, causing the low resistivities that have been observed. Dyke injection to shallow crustal levels (1-5 km) combined with heavily weathered sedimentary fill, may generate low resistivities in the upper crust. The well defined western margin confined the shape and extension of the anomalies to the area of the surface expression of the rift. This inferred interpretation is reinforced by the seismic model of Birt (1996) and the upper mantle model of Byrne (1997). Byrne (1997) developed a model with an upper mantle reflector at about 50 km depth beneath the western flank of the rift that shallows immediately beneath the rift to about 43 km depth. Additionally, the teleseismic tomographic studies from the central section of the rift constrained a low velocity zone to the western flank of the rift (Achauer, 1994).

**Western section:** Melt containing blobs from the top of the asthenospheric wedge may have been segregated and intruded into the lithosphere, rising towards the surface along pre-existing structural discontinuities. The suture zone between the Archaean and Proterozoic crust may have facilitated the intrusion and accumulation of magma beneath the region of the Oloololo
Escarplet, and possibly underneath the western end of the line around station Nyamanga. If the Oloololo Escarpment marks the structural boundary between the Archaean Craton and the Proterozoic Belt in the southern Kenya (supported from the gravity model of Nyblade and Pollack (1992), and the surface geology), then the geoelectric model has clearly imaged it as the conductive zone between the representative Archaean crust to the west and an ill-defined, deformed Proterozoic (?) crust to the east. However, this hypothesis leans on the results of a single station, and contradicts the results of Smith and Mosley (1993), Smith (1994) and Birt (1996), that argue that the suture zone in the southern Kenya lies directly under the rift axis. Correlation of the westward deepening conductive zone in the Oloololo area and the conductive layer at depth of about 30 km beneath sites Nyamanga to Lolkoroin with the presence of a plume (circulating hot material) beneath the Lake Victoria, and possibly feeding the rift is speculative.

Figure 6.7: Schematic cross-section along the southern Kenya Rift Valley, based on the geoelectric model of the area and deductions from past gravity, seismic and geological observations (Shackleton, 1986; Nyblade and Pollack, 1992; Smith, 1994; Birt, 1996).

Eastern section: Absence of a geoelectric boundary correlatable with the geologic and topographic expression of the rift combined with the eastward extension of the conductivity anomalies may reflect an asymmetric eastward extension of the rifting processes. The plume may have spread laterally and melt travelled away from the source exploiting crustal weakness and formed the off-axis volcanic centre of the Chyulu Hills (Figure 6.8) (resulting in the
observed low resistivities). It may be argued that the presence of a conductive body underneath the eastern flank of the rift at a depth of 100 km as modelled by Beamish (1977) may be the geoelectric signature of the plume. However, the suggested possible presence of a section of ‘normal’ crust between the rift and the Chyulu Hills may prevent any direct link between the two features at crustal depths, but does not preclude a deeper connection at asthenospheric depths, or even at the lower crust which the present MT model did not resolve.

Figure 6.8: Sketch showing the simplified features of the southern Kenya Rift. Modified from Smith (1994) and Birt (1996). Mafic magma from an upwelling mantle plume beneath the East African Plateau intruded and underplated the lower crust as magma chambers and sills. Some magma rises into the crust through pre-existing crustal weaknesses beneath the rift and the Chyulu Hills.
Deep magnetotelluric soundings together with transient electromagnetic measurements were performed at 19 sites in the southern part of the Kenya Rift. The sites were positioned along a profile running across the southern part of the Kenya Rift Valley and another one parallel to the Chyulu Hills volcanic chain. Data acquisition was hampered by numerous factors of which the hard terrain conditions together with instrumentation failures, financial and time constraints resulted in a wide station separation of 20-25 km which limited the model resolution.

The MT frequency range varied from 128 Hz to 5000 sec providing information down to 25-40 km. Upward biased, downward biased and average impedance tensors (Sims et al., 1971) were calculated from the recorded time series. Data quality varied from site to site, but from a number of soundings noise-free, smooth estimates of the apparent resistivity and phase curves were calculated. However, cultural noise contaminated some of the soundings (e.g. station Magadi) and one site to the east of the rift (Kajiado) produced no worthwhile data.

The measured impedance tensors were analysed using the decomposition methods of Groom and Bailey (1989) and Bahr (1991). The impedance tensors at a number of stations for frequencies lower than 0.1 Hz show the effects of three dimensional induction, while in many sounding sites near surface inhomogeneities caused static shift problems. Strong topographic effects from the Nguruman Escarpment deteriorate even further the impedance tensor of the station located on the top of it. The regional azimuth derived from the Groom and Bailey (1989) decomposition and the classical method of Swift (1967) indicated a N-S electrical azimuth for the area of the Rift Valley, coincident with the geological strike of the rift, and about N40°W for the Chyulu Hills area, almost parallel to the shear zone that crosses the area. The dual mode TEM method (Meju et al., 1999) used to correct the MT apparent resistivity from static shift effects allowed a more detailed examination of the near surface geoelectric structure (the top 300 m).

The frequency domain response functions were transformed into models of the resistivity with depth variation by a number of different routes. The approximate Bostick transformed apparent resistivity and phase information from each site allowed a rapid appraisal of the gross geoelectric variations. One dimensional inversion models of both polarization modes and the
invariant were collated to form pseudosections for the MT profiles and present realistic images of the area, and an a priori model for the final two dimensional inversion. The 2-D geoelectric model revealed features that the simple 1-D inversion failed to detect, and should be regarded as the best approximation to the more realistic 3-D earth structure. Nevertheless, wide station spacing, lack of long period data (stations Nyamanga, Lolkoroin), scattered and noisy data (stations Kehancha, Keekorok) limited the vertical and lateral resolution of the model and added to the non-uniqueness of the MT method. An exploratory error analysis of the final 2-D model helped to elucidate and emphasize some of the main features. The resulting earth models from two dimensional inversion combined with geological information and the results from previous geophysical studies were used to gain a better understanding of the geological structure and architecture of the crust beneath the Kenya Rift and the region around it.

7.1 Conclusions

Some of the key results of this study are catalogued here:

- The use of the multi-geometry transient electromagnetic technique to remedy the static shift problems from the magnetotelluric responses provided the right level for the MT apparent resistivity curves, and thus eliminated one of the major problems of the MT technique. In shallow 1-D conditions (e.g. station Singleraine) the procedure was straightforward; the MT apparent resistivity curves are shifted to the level of the TEM apparent resistivity. In complex environments (e.g. station Irkiba) joint inversion of $\rho_{TEM}$ with the unaffected MT impedance phase proved sufficient to define the right level for the $\rho_{MT}$. Both central and coincident TEM data have been used and correlated with the TE and TM modes respectively.

- The area of the southern Kenya Rift Valley is shown to be approximately two dimensional in geoelectric structure. The 2-D inversion model derived from this experiment is consistent with the known geology and previous geophysical studies of the area. It is thus the first unequivocal geoelectric model for the Kenya Rift. However, the wide station separation limited the lateral resolution of the geoelectric structure, and hence the uniqueness of some of the features present in the model.

- At the western end of the main profile the Archaean Nyanza Craton has been clearly imaged in the geoelectric model. A highly resistive upper crust overlies a conductive middle crust that rests on a highly resistive lower crust. The hypothesis of a rising mantle plume is not supported by this dataset, but cannot be ruled out as a possibility.

- The Ololololo Escarpment coincides with a conductive fault-like zone and a mantle depth conductor. Whether this feature marks the boundary between the Nyanza Craton to the west
and the Mozambique Belt in the east is not well defined, but clearly the electric model greatly enhances the importance and the influence of the escarpment in the deeper structure of the area than the previous seismic investigation which showed variations limited to the near surface layers.

- The crustal characteristics of the region between Ololoolo and Nguruman Escarpments are not well resolved. Gradually the resistivity increases from the west to the east reaching values of tens of thousands of $\Omega\cdot m$ at the rift flanks. Whether the crust to the east is Proterozoic or Archaean cannot be determined from this study.

- The western flank of the rift is apparently more sharply imaged than the eastern flank. Low resistivities in the upper crust inside the rift identify sediments and pyroclastics that fill the floor of the rift valley, and which are probably permeated by salty fluids. Moderate conductivities in the middle crust may arise from the presence of cooling intrusions and the circulation of hot brines to which they give rise. The deeper conductive diapir-like bodies are likely to be related to magmatism and partial melt. The surface conductors mask the deeper structure and the MT penetration depth and hence model resolution is limited to 20-25 km depth. The sparser station density to the east of the rift may be the reason for the absence of a clear geoelectric boundary in the area. However, it is possible that the geoelectric structure reflects the gentle step-faulting of the eastern rift margin, and an asymmetrical extension of the rifting towards the Proterozoic Mobile Belt to the east. The crustal resistor underneath station Selengei appears to set a boundary to the rift area and images the Proterozoic Mozambique Belt.

- The induction arrows are influenced by the N-S trending rift at AMT frequencies and become oblique to the rift (NW-SE) at lower frequencies (ca. LMT data). Pre-existing structures and shear zones can be linked to this behaviour of the induction arrows, since graphite and related mineralization probably exist at shear zones channelling current along them. Moreover, it has been suggested (Smith and Mosley, 1993) that these lineaments have strongly influenced the evolution and location of the rift providing paths to the upwelling magma.

- The invariant 1-D section provides a useful image for the area of the Chyulu Hills but the depths to interfaces were underestimated, while the resistivity range was overestimated in this complex environment. The impedance tensor was subsequently rotated to 40° west of north and the 2-D inversion model provided a more realistic image of the 3-D area. Low resistivities in the upper and middle crust coincide with areas of surface volcanism and the possible presence of small magma chambers. The crustal resistor underneath Selengei station combined with the possible existence of a resistor between Chyulu South and
Mwatate stations, suggests the presence of a small diapir under the Chyulus probably caused by small melt blobs that originated from a larger plume below the Kenya Rift. The induction arrows have not revealed a dominant inductive strike direction for the area, emphasizing the 3-D character of the volcanic field.

### 7.2 Suggestions for further research

The Kenya Rift Valley is a complex environment and it is clear that more geophysical and geochemical research is required to fully understand the structure and the evolution of the rifting processes in the area. The geoelectric model of this study provided new insights into the architecture of the crust and towards the understanding of the importance and influence of a possible mantle plume and pre-existing structures on the location and development of the Kenya Rift. However, a better understanding of the causes of the resistivity variations in the earth's crust in stable and active environments will help to improve the present image of the continental rift.

Suggestions for further work can be divided into two main groups: (i) further field measurements, and (ii) further mathematical modelling of the present and previous data from the Kenya Rift.

Electromagnetic induction soundings at various locations could be performed in order to confirm or eliminate interpretations drawn from the present data set. During the modelling procedure the high station separation caused problems and generated interpretational uncertainties. Closer station spacing in certain areas will be able to provide the desired lateral resolution in the model and minimize the degree of non-uniqueness in the MT model. The Oloololo conductivity anomaly, the eastern flank of the rift, the region around Selengei station and the Chyulu Hills are some of the places where the lack of high density observations generated uncertainties. Furthermore, a new 3-D magnetotelluric survey is required to image the deep electrical structure beneath the Kenya Dome and further north; these measurements are needed to provide a more complete image of the proposed mantle plume and will attribute to the better understanding of the rifting processes in a well developed rifting area.

A re-interpretation of this data set and previous results from the near-equator profiles using 3-D numerical modelling procedures may be beneficial to estimate the geoelectric character of the northward extension of the rift and other structural boundaries (such as the suture zone between the Nyanza Craton and the Mozambique Belt). Additionally, three-dimensional modelling of these data, and especially the data from the Chyulu Hills will add to the uniqueness of the models that are presented in this study.
Appendix A

The MT results obtained at 14 stations across the southern Kenya Rift Valley (A.2-A.15), 5 stations along the Chyulu Hills (A.15-A.19), and station Lukenya (A.20), 10 km SE of Nairobi. The Rift and Chyulu Hills stations are presented in geographical order from the west (Nyaragata) to the east (Mwatate).

For each site, the results shown are: the apparent resistivity and phase curves of the off diagonal elements of the unrotated impedance tensor, the major and minor apparent resistivity and phase, the number of estimates averaged, the coherency, skew and the regional azimuth of the major impedance (measured positive clockwise from magnetic north). The error bars are one standard deviation.
Station: Nyamanga
Station: Ooololo

Unrotated XY

Unrotated YX

App. Res. (Ω m)

Phase (°)

Frequency (Hz)

Major

Minor

Num of Estimates

Coherency

Skew

Azimuth (°)

Appendix A
Station: Nguruman
Station: Magadi
Station: Chyulu 2 (Range)

Unrotated XY

App. Res. (\Omega m)

Phase (°)

Major

Minor

Frequency (Hz)

Unrotated YX

App. Res. (\Omega m)

Phase (°)

Num of Estimate

Coherency

Skew

Azimuth (°)

Frequency (Hz)
Appendix B

Parkinson Induction Arrows computed from the AMT and LMT data recorded at 19 stations across the Southern Kenya Rift, and along the Chyulu Hills.
Station: Nyamanga

Station: Migori
Station: Ooololo

Real

Imaginary

Station: Keekorok

Real

Imaginary
Station: Leganishu

Real

Imaginary

Station: Morijo

Real

Imaginary
Appendix B

Station: Irkiba

Station: Nguruman
Station: Magadi

Real

Magnitude

Azimuth (°)

Frequency (Hz)

Station: Singleraine

Real

Magnitude

Azimuth (°)

Frequency (Hz)
Station: Selengei

Station: Chyulu North
Station: Chyulu West

Station: Chyulu 2 (Range)
Station: Chyulu South

Station: Mwatate
Station: Lukenya

Real

Imaginary

Magnitude

Azimuth (°)

Frequency (Hz)

Magnitude

Azimuth (°)

Frequency (Hz)
Appendix C

TEM, AMT and LMT data for the 19 stations of the *KRISP 95 - MT experiment*. The MT data have been corrected from static shift effects using the TEM data.

The MT phases have been plotted together with the TEM voltage decay curves. The TEM data are shown by triangles, and the circles represent the MT polarizations. The central loop (CN) data have been correlated with the TE mode, while coincident loop (CC) data have been used to correct the TM mode.

For the Chyulu Hills the TEM data have been plotted together with the unrotated MT polarizations.
Station: Nyamanga

- TEM data
- AMT-LMT data

**TEM CN - TE mode**

**TEM CN - TM mode**

Station: Migori

- TEM data
- AMT-LMT data

**TEM CN - TE mode**

**TEM CN - TM mode**
Station: Kehancha

- TEM data
- AMT-LMT data

**TEM CC - TE mode**

**TEM CN - TM mode**

Station: Lolkoroin

- TEM data
- AMT data

**TEM CN - TE mode**

**TEM CC - TM mode**
Appendix C

Station: Oloololo

- TEM data
- AMT-LMT data

TEM CC - TE mode

TEM CN - TM mode

Station: Keekorok

- TEM data
- AMT-LMT data

TEM CC - TE mode

TEM CN - TM mode
Station: Leganishu

- TEM data
- AMT-LMT data

TEM CN - TE mode

Station: Morijo

- TEM data
- AMT-LMT data

TEM CN - TE mode
Appendix C

Station: Irkiba

- TEM data
- AMT-LMT data

Station: Nguruman

- TEM data
- AMT-LMT data
Station: Magadi

- TEM data
- AMT-LMT data

**TEM CN - TE mode**

Frequency (Hz)

**TEM CC - TM mode**

Frequency (Hz)

Station: Singleraine

- TEM data
- AMT-LMT data

**TEM CN - TE mode**

Frequency (Hz)

**TEM CC - TM mode**

Frequency (Hz)
Station: Selengei

- TEM data
- AMT-LMT data

**TEM CN - TE mode**

**TEM CN - TM mode**

Station: Chyulu North

- TEM data
- AMT-LMT data

**TEM CN - TE mode**

**TEM CN - TM mode**
Station: Chyulu West

- TEM data
- AMT-LMT data

Station: Chyulu 2 (Range)

- TEM data
- AMT-LMT data
Appendix C

Station: Chyulu South

- TEM data
- AMT-LMT data

**TEM CN - XY**

**TEM CN - YX**

Station: Mwatate

- TEM data
- AMT-LMT data

**TEM CN - XY**

**TEM CN - YX**
Station: Lukenya

- TEM data
- AMT-LMT data

**TEM CN - XY**

**TEM CC - YX**
References


Beamish, D., 1990b, A granite window to the lower electrical crust and upper mantle, Terra Nova, 2, 314-319.


Bosworth, W., 1987, Off-axis volcanism in the Gregory rift, East Africa: Implications of the models of continental rifting, Geology, 15, 397-400.


Cagniard, L., 1953, Basic theory of the magnetotelluric method of geophysical prospecting, Geophysics, 18, 605-635.
References


References


Groom, R.W., and Bailey, R.C., 1991, Analytic investigations of the effects of the near surface three dimensional galvanic scatters on MT tensor decompositions, Geophysics, 56, 496-518.


KRISP Working Group, 1995, Group takes a fresh look at the lithosphere underneath Southern Kenya, *EOS* 76, 73, 81-82.


References


Meju, M.A., 1995, Simple effective resistivity-depth transformations for infiel or real-time data processing, Computer & Geosciences, 21, 985-992.


Morley, C.K., 1989, Extension detachments and sedimentation in continental rifts (with particular reference to East Africa), Tectonics, 8, 1175-1192.


References


Rankin, D., 1962, The magnetotelluric effect on a dike, Geophysics, 27, 666-676.


Strangway D.W., Swift, C.M.(Jr), and Holmer R.C., 1973, The application of audio-frequency magnetotellurics (AMT) to mineral exploration, *Geophysics*, 38 (6), 1159-1175.


Wannamaker, P.E., 1997, Comment on "The petrologic case for a dry lower crust" by Yardley, Internet site: MTNET "http://nazca.cg.emr.ca/mtnet/mtnet.html".


