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Tectonophysics 346 (2002) 169–185

TECTONOPHYSICS

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Magnetotelluric images of the crustal structure of Chyulu Hills volcanic field, Kenya

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Received 23 January 2001; accepted 26 November 2001

Abstract

Electromagnetic experiments were conducted in 1995 as part of a multidisciplinary research project to investigate the deep structure of the Chyulu Hills volcanic chain on the eastern flank of the Kenya Rift in East Africa. Transient electromagnetic (TEM) and broadband (120–0.0001 Hz) magnetotelluric (MT) soundings were made at eight stations along a seismic survey line and the data were processed using standard techniques. The TEM data provided effective correction for static shifts in MT data. The MT data were inverted for the structure in the upper 20 km of the crust using a 2-D inversion scheme and a variety of starting models. The resulting 2-D models show interesting features but the wide spacing between the MT stations limited model resolution to a large extent. These models suggest that there are significant differences in the physical state of the crust between the northern and southern parts of the Chyulu Hills volcanic field. North of the Chyulu Hills, the resistivity structure consists of a 10–12-km-thick resistive (up to 4000 Ω m) upper crustal layer, ca. 10-km-thick mid-crustal layer of moderate resistivity (~ 50 Ω m), and a conductive substratum. The resistive upper crustal unit is considerably thinner over the main ridge (where it is ca. 2 km thick) and further south (where it may be up to 5 km thick). Below this cover unit, steep zones of low resistivity (0.01–10 Ω m) occur underneath the main ridge and at its NW and SE margins (near survey positions 100 and 150–210 km on seismic line F of Novak et al. [Novak, O., Prodehl, C., Jacob, A.W.B., Okoth, W., 1997. Crustal structure of the southern flank of the Kenya Rift deduced from wide-angle P-wave data. In: Fuchs, K., Altherr, R., Muller, B., Prodehl, C. (Eds.), Structure and Dynamic Processes in the Lithosphere of the Afro-Arabian Rift System. Tectonophysics, vol. 278, 171–186]). These conductors appear to be best developed in upper crustal (1–8 km) and middle crustal (9–18 km) zones in the areas affected by volcanism. The low-resistivity anomalies are interpreted as possible magmatic features and may be related to the low-velocity zones recently detected at greater depth in the same geographic locations. The MT results, thus, provide a necessary upper crustal constraint on the anomalous zone in Chyulu Hills, and we suggest that MT is a logical compliment to seismics for the exploration of the deep crust in this volcanic-covered basement terrain. A detailed 3-D field study is recommended to gain a better understanding of the deep structure of the volcanic field.

Keywords: Kenya Rift; Off-axis volcanism; Electromagnetic depth sounding; Conductivity imaging

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0040-1951/02/$ - see front matter
PII: S0040-1951(01)00276-1

1. Introduction

The Kenya Rift is characterised by extensive volcanic activity (Fig. 1) and is one of the most intensively studied magmatic continental rifts (e.g., Gregory, 1921; Baker and Wohlenberg, 1971; Khan and Mansfield, 1971; Banks and Ottey, 1973; Beamish, 1977; Rooney and Hutton, 1977; Banks and Beamish, 1979; Shackleton, 1986; Karson and Curtis, 1989; Smith and Mosley, 1993; Macdonald et al., 1994; Wheildon et al., 1994; Birt et al., 1997; Novak et al., 1997a). Rift-related magmatic activities started in northern Kenya in the Oligocene (see Morley et al., 1992 and references therein), propagated southwards—being initially restricted to the rift proper and its western flank—and is still continuing (see Williams, 1978). Large basaltic fields started to develop on the eastern flank of the rift in Late Miocene times and Quaternary volcanism was restricted to the actual rift and the eastern flank (Baker, 1987; Karson and Curtis, 1989; Class et al., 1994; Macdonald, 1994). A belt of several hundreds of young volcanic fields formed 100–150 km off the axis of the rift and are thought to represent the largest volcanic features related to the rift structure (Bosworth, 1987).

The Chyulu Hills volcanic field in southern Kenya (Fig. 1) is one of such fields. It is an area of high magmatic activity and is situated 150 km east of the southern part of the rift, about 40 km northeast of Mt. Kilimanjaro. The volcanic chain is located directly on a preexisting NW–SE striking shear zone in the Proterozoic basement of the Mozambique orogenic belt that may have served for magma injection into the upper crust (Smith and Mosley, 1993). The age of volcanism ranges from 1.4 Ma in the northern part to the recent (Saggerson, 1963; Haug and Strecker, 1995), with active volcanism in the last century restricted to the southern edge of the main ridge. The hundreds of cones characterising the Chyulu field are aligned in a NW–SE direction. The metamorphic basement terrain on which the Chyulu Hills rests is a peneplain with an average elevation of 1000 m and the enclosed narrow Quaternary chain of volcanoes attains a maximum altitude of 2175 m. A smaller volcanic edifice, the Taita Hills, occurs further south of the Chyulus. The origin and evolution of the volcanic activities off the axis of the main rift in the region have attracted considerable attention (e.g., Bosworth, 1987; Karson and Curtis, 1989; Smith and Mosley, 1993; Smith, 1994) but are still not well understood.

1.1. Past work and motivation for current studies

The structure and composition of the deep crust and lithosphere across the Chyulu Hills volcanic field have been the main focus of recent intensive studies (e.g., Henjes-Kunst and Altherr, 1992; Novak et al., 1997a,b; Ritter and Kaspar, 1997). Geothermobarometric data on mantle xenoliths suggest an apparent lithospheric thickness of about 105 km (Henjes-Kunst and Altherr, 1992). The northern part of the Chyulu field especially east of Selengei (Sel in Fig. 1) contains remnants of the earliest stage (Early Pleistocene) magmatic activity in the Chyulus (Saggerson, 1963). There are widely spaced, partially eroded volcanic cones in this area and studies of xenoliths in the volcanic rocks show that magma transport was characterised by rapid ascent from about 105–110-km depth to the surface without stagnation in crustal chambers (Henjes-Kunst and Altherr, 1992). The southern part of the Chyulu Hills volcanic field contains a steep volcanic ridge (hereafter referred to as the main ridge) with closely spaced cones over a distance of 50 km and formed during Late Pleistocene to recent times (Saggerson, 1963; Omenge and Okele, 1992). Here, the magma is thought to have had significant residence time in the crust since the volcanic rocks contain abundant crustal-derived xenoliths but lack lower crustal and mantle-derived xenoliths (see Novak et al., 1997b).

Based on these geological and geothermobarometric considerations, it is to be expected that there will be differences in the physical state of the crust in the northern and southern parts of the Chyulu Hills volcanic field. The seismic refraction P-wave model of Novak et al. (1997a, Fig. 8) shows a 9–11-km-thick upper crust, a <10-km-thick middle crust, and up to 25-km-thick lower crust with no major lateral variations over a distance of ca. 275 km along the profile. The revised interpretative model supported by an integrated data set (Novak et al., 1997b, Fig. 6) suggests the presence of a low-velocity body in the lower crust (30–44 km deep) beneath the volcanic field. From a tomographic study of the Chyulu Hills, Ritter and Kaspar (1997) demonstrated the presence of lateral velocity contrasts of about 5% with a prominent low-velocity zone located directly beneath the volcanic

range down to 70-km depth (see also Novak et al., 1997b, Fig. 7); however, near-surface effects were not well resolved and there were no major laterally varying structures in the 3–23-km-depth range in the interpretative models. It is apparent from the foregoing discussions that further geophysical constraints on the upper crustal structure of the region are desirable, and may also help improve our understanding of the structure and evolution of the Chyulu Hills volcanic field. Since there was active volcanism in the southern edge of the main ridge in the last century (Saggerson, 1963; Omenge and Okele, 1992), vestiges of magma injection in the upper crust may be expected and could be detected using ideal geophysical depth soundings. The structure of the upper 25 km of the earth's crust in Chyulu Hills will be examined in this paper.

1.2. Adopted methodology and study focus

The magnetotelluric (MT) method is a proven wavefield electromagnetic technique for deep subsurface imaging in volcanic-covered and other crystalline terrains (e.g., Christopherson, 1991; Jones, 1993; Simpson et al., 1997; Simpson, 2000; Bai et al., 2001). The time-domain or transient electromagnetic (TEM) method is a well-established tool for geological mapping (e.g., Meju et al., 1999) and mineral exploration in complex terrains (e.g., Peters and de Angelis, 1987 and references therein). The MT method has capability for probing several tens of kilometers but the data may be affected by galvanic distortions—manifesting as frequency-independent static shifts of the apparent resistivity curves—when small-size surficial heterogeneities are present (Berdichevsky and Dmitriev, 1976), as can be expected in the weathered and volcanic-covered basement terrain of Chyulu Hills. The TEM technique provides a logical shallow-depth (<1 km) compliment to MT and also serves for correction of MT static shifts (e.g., Sternberg et al., 1988; Pellerin and Hohmann, 1990; Meju, 1996). The combined TEM–MT approach was therefore selected as the technique with optimum potential in the Chyulu Hills crystalline terrain and was incorporated into a multidisciplinary program aimed at understanding the process of continental rifting in Kenya (see Simpson et al., 1997; Sakkas, 1999).

Preliminary analysis of MT data from the region suggests the presence of significantly enhanced conductivities below the Chyulu Hills (Simpson et al., 1997); however, no quantitative modelling of the data was undertaken and the actual physical parameters of the suggested zone of enhanced conductivities were not known. An improved TEM-based technique for static shift correction in MT data from complex geological terrains was recently suggested (Meju et al., 1999) and it is highly desirable to develop a quantitative MT model for the upper crustal structure of the Chyulu Hills volcanic field to compliment the models of the deep crust and lithosphere from other geophysical investigations along the same transect (e.g., Novak et al., 1997b; Ritter and Kaspar, 1997). This is the main focus of our paper. However, it is pertinent to mention that the TEM–MT station spacings (dictated by the resources available for this aspect of the multidisciplinary project) are much too wide (>30 km) in comparison with the seismic refraction and gravity station spacings (ca. 2.5 km) on the Chyulu Hills line, and the present investigation may therefore be best regarded as a feasibility study of the applicability of the combined TEM–MT method in this particular volcanic field. We note, however, that there have been past electromagnetic induction studies north of this region (e.g., Banks and Ottey, 1973; Beamish, 1977; Rooney and Hutton, 1977; Banks and Beamish, 1979) but these have different research thrusts; and for those involving MT measurements, little attention was paid to the distortion of the MT sounding curves by small-size surficial heterogeneities. Surficial small-size bodies cause galvanic distortions of MT apparent resistivity sounding curves (Berdichevsky and Dmitriev, 1976) leading to erroneous subsurface models if unaccounted for (see Jones, 1988; Groom and Bahr, 1992 and references therein). TEM data have been shown to solve this problem in simple-layered geological terrains but only provide partial remedies in complex terrains (see e.g., Sternberg et al., 1988; Pellerin and Hohmann, 1990; Meju, 1996; Meju et al., 1999). It is hoped that joint TEM–MT measurements will lead to improved model identification.

2. Field experiments and data analysis

Eight sounding locations (along the seismic and gravity line F of Novak et al., 1997a,b) were occupied in early 1995 as part of the KRISP94-MT project

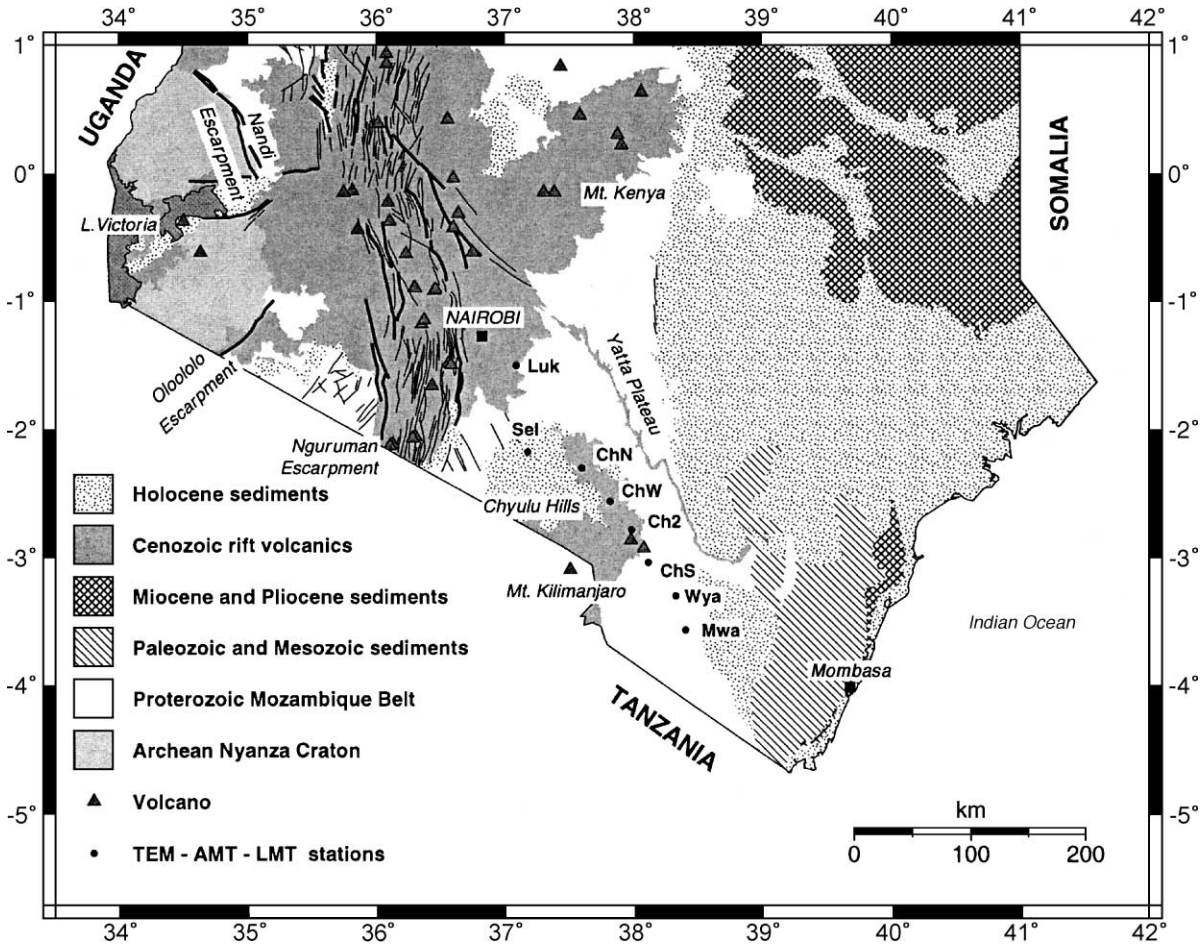
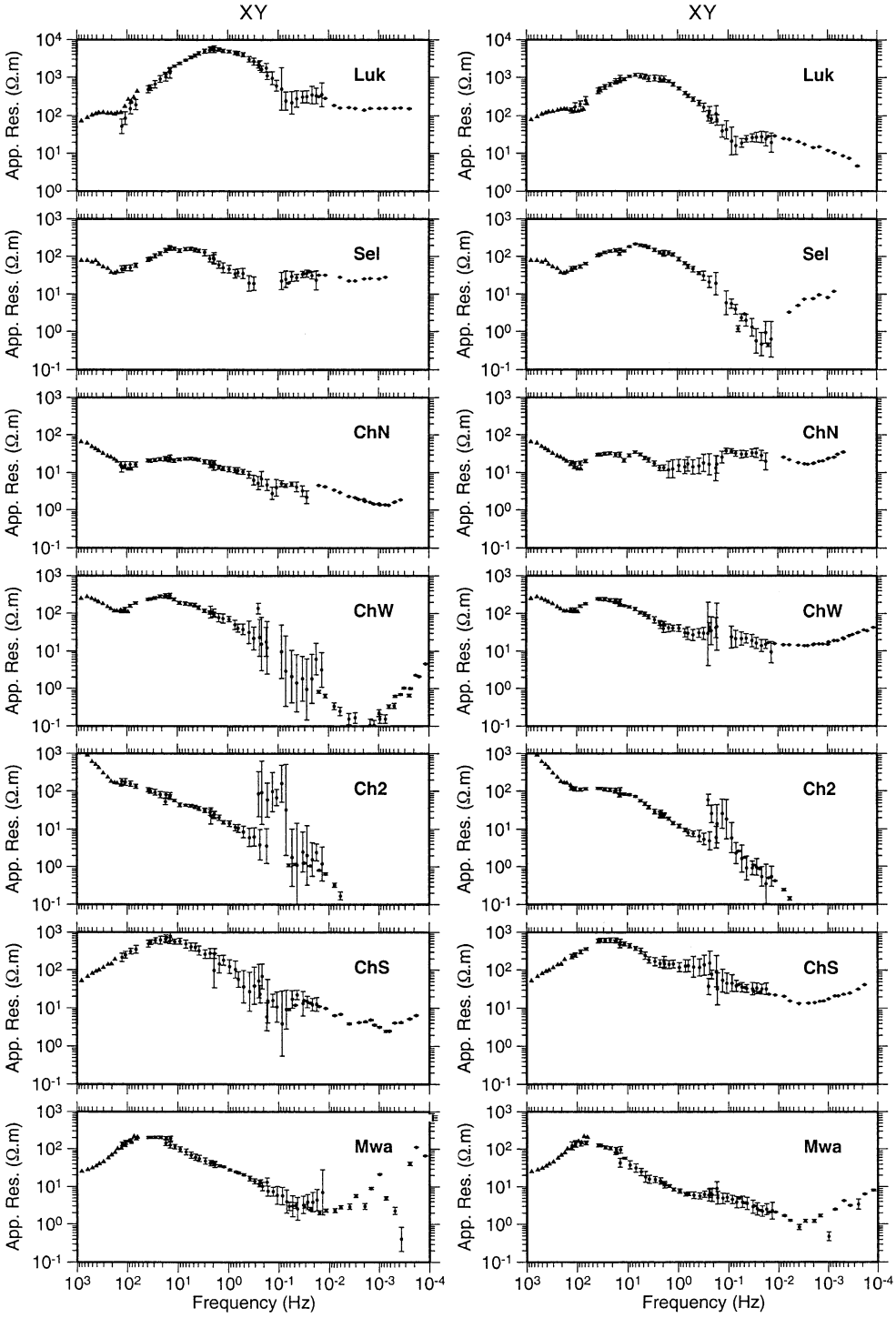


Fig. 1. Geology map of Kenya showing location of TEM–MT stations in the Chyulu Hills volcanic field. The stations are: Lukenya (Luk), Selengei (Sel), Chyulu North (ChN), Chyulu West (ChW), Chyulu 2 (Ch2), Chyulu South (ChS), Wyandani (Wya), and Mwatate (Mwa).

(Simpson et al., 1997). The field experiments have been described in detail by Simpson et al. (1997) and only the salient points will be recounted here. The TEM–MT transect extends from Lukenya (ca. 30 km south of Nairobi), past the Chyulu Hills, to Mwatate south of the Taita Hills about 160 km from Mombasa and the Indian Ocean to the southeast (see Fig. 1). The northernmost part of the profile is composed of metamorphic rocks of the Proterozoic Mozambique orogenic belt (see Saggerson, 1963). Most of the TEM–MT stations were situated close to the main seismic shot points of the KRISP94 experiments (Lukenya is near Athi, the northernmost seismic shot point ATH; Chyulu North and Chyulu South are seismic shot points at the NW and SE edges of the Chyulu Range. Mwatate station is 46 km north of the seismic shot point RUK of Novak et al., 1997a) and just south of Taita Hills. The average station spacing over the Chyulu Hills (the main zone of anomalous seismic signature) is 33 km, which is greater than the 25-km depth of exploration interest in this paper. The wide spacing between sounding stations means that

Fig. 2. Combined MT and TEM apparent resistivity data for Chyulu Hills stations. The MT *xy* and *yx* data (shown as round symbols with error bars in the left-hand and right-hand plots) have been corrected for static shift using the central-loop TEM data (triangular symbols) plotted at their equivalent frequencies based on Meju (1996). Note that the AMT data with larger error bars overlap with the higher quality long-period MT data (cf. Table 1).


XY
XY
Luk
Sel
ChN
ChW
Ch2
ChS
Mwa
App. Res. (Ω.m)
Frequency (Hz)

short wavelengths features will not be resolved in this pilot experiment.

TEM and high-frequency MT data were not recorded at station Wyandani due to a technical problem at the terminal stage of the field experiments. At every other station, TEM soundings were performed using the central-loop and single-loop configurations with a 100-m-sided square loop acting as the transmitter (Tx). The same loop serves as the receiver in the single-loop case since in the TEM method, we only measure the transient responses of the ground to inductive energization when the Tx is switched off. In the case of central-loop sounding, the receiver was a small multi-turn coil with an effective area of 10 000 m^2 and was placed at the centre of the Tx loop. The MT measurements employed induction coils and 100-m-long grounded electric dipoles. The recordings consisted of two orthogonal horizontal (magnetic north–south and east–west) components of both the electric and magnetic fields on the ground surface and the vertical magnetic field component. The MT electric dipoles were arranged in an L-shaped configuration and covered the same ground as the TEM Tx loop to ensure that both methods sampled the same geology, a practical necessity for accurate static shift correction (Meju, 1996). The MT data cover a frequency band width of 120–0.0001 Hz while the TEM soundings effectively cover a frequency band of 1000–10 Hz (based on Meju, 1996, Eq. (1)) as in the examples presented in Fig. 2. The MT surveys employed a short-period field system developed at Edinburgh University and long-period RAP systems developed by Erich Steveling at the University of Gottingen (see Simpson et al., 1997 for details), and data were recorded in six frequency sub-bands (Table 1).

The TEM transient voltage responses have been adjusted for transmitter turn-off effects (Raiche, 1984) and converted to apparent resistivities using a non-linear iterative scheme with initial values furnished by the late-time approximation (Kaufman and Keller, 1983). The MT data were processed using standard tensorial techniques (Swift, 1967; Egbert and Booker, 1986) with attention being paid also to data sets affected by power line noise (Fontes et al., 1988). The resulting impedance tensor elements were used to compute the apparent resistivities, phases, azimuths, induction arrows, and other geoelectrical indices (Parkinson, 1959; Swift, 1967; Bahr, 1988, 1991; Groom and Bailey, 1989, 1991; Groom and Bahr, 1992) necessary for subsurface structural interpretation. We obtained the apparent resistivities in the measurement directions (*xy* or north–south, and *yx* or east–west polarisations) for preliminary assessments and corrected them for static shift using central-loop TEM data since they look fairly similar in shape at most stations (see Fig. 2). Note the good overlap between the AMT (128 Hz to < 100 s) data and the higher-quality, long-period MT data (40–>1000 s). The gross features of these MT curves suggest a simple resistive–conductive–resistive sequence underneath variable overburden evinced by the TEM response.

2.1. Distortion analysis and regional geoelectrical strike

In order to determine whether Groom and Bailey (1989) decomposition was necessary for the MT data sets, the data have been analysed for telluric distortion and classified using the methods described in Bahr

Table 1
Classification of telluric distortions in the Chyulu Hills data subsets (six frequency bands) using the method of Bahr (1991)

Stations	128–16 Hz	16–2 Hz	2 Hz–4 s	4–32 s	40–1000 s	>1000 s
Lukenya	3	1	3	–	7	7
Selengei	1	1	7	7	7	7
Chyulu North	7	7	7	7	7	–
Chyulu West	–	–	2	7	3	–
Chyulu 2	1	–	7	7	–	–
Chyulu South	7	1	–	–	3	–
Mwatate	1	1	7	7	7	7

The first four frequency bands refer to short period (AMT) measurements while the last two refer to long period (LMT) measurements using different field equipment (cf. Fig. 2).

(1991). The resulting classifications (1–7) are given in Table 1. Some of the data bands show characteristics of weak local distortion (class 3) with the high-frequency bands showing classes 1 and 2 features, and can be interpreted using conventional methods. However, the data at most sites fall into class 7 (especially at long periods), with skew values (Swift, 1967) greater than 0.3 suggesting a regional 3-D structure, for which the superimposition model (Groom and Bailey, 1989) is inappropriate. Interestingly, the Groom–Bailey decomposition method yielded fairly consistent regional geoelectrical strike (NE–SW or NW–SE) directions for the neighbouring stations as in the examples presented in Fig. 3 for selected frequencies, and we may elect to interpret the MT profile data using a 2-D approximation.

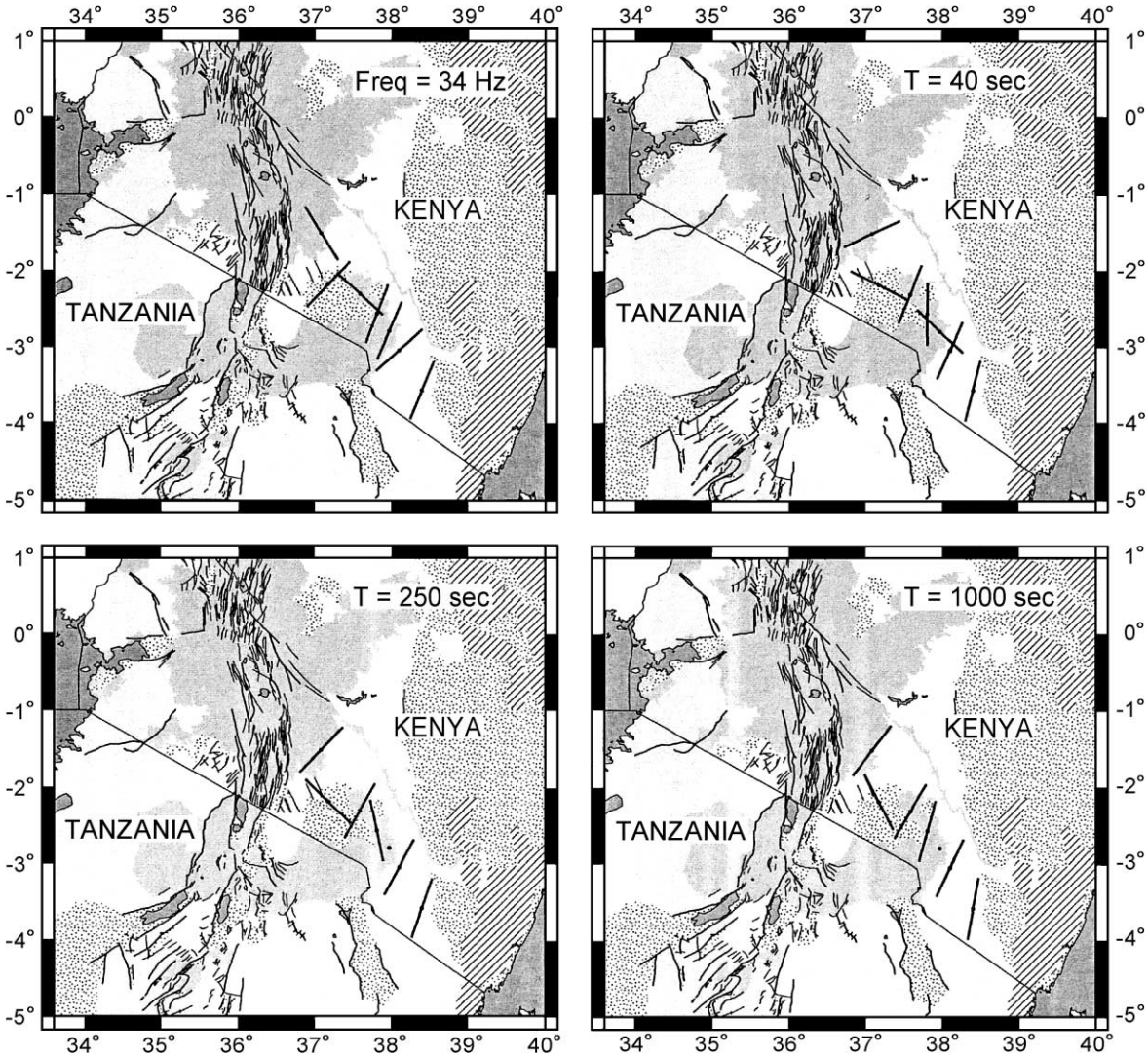


Fig. 3. Pictorial representation of regional geoelectric azimuths determined using the Groom–Bailey (1989) method for selected periods for the MT data. The azimuths are indicated by the site-centred dark bars and are superimposed on the regional fault fabric (re-drawn from Smith and Mosley, 1993).

The ratios of the vertical-to-horizontal magnetic fields from MT records are often displayed graphically as 'induction arrows' (Parkinson, 1959, 1983). These facilitate a qualitative areal assessment of the degree of conductivity variations in the subsurface. In an area with a dominant structural trend, the induction arrows will be orthogonal to and point towards anomalous current concentrations such as that which will be induced in a linear conductive body (Jones and Price, 1970). The Parkinson induction arrows for Chyulu Hills are of small magnitudes and unstable (i.e., do not define a dominant trend at all frequencies) for the seven stations shown in Fig. 4 but appear to point to the presence of linear conductors near stations ChN and

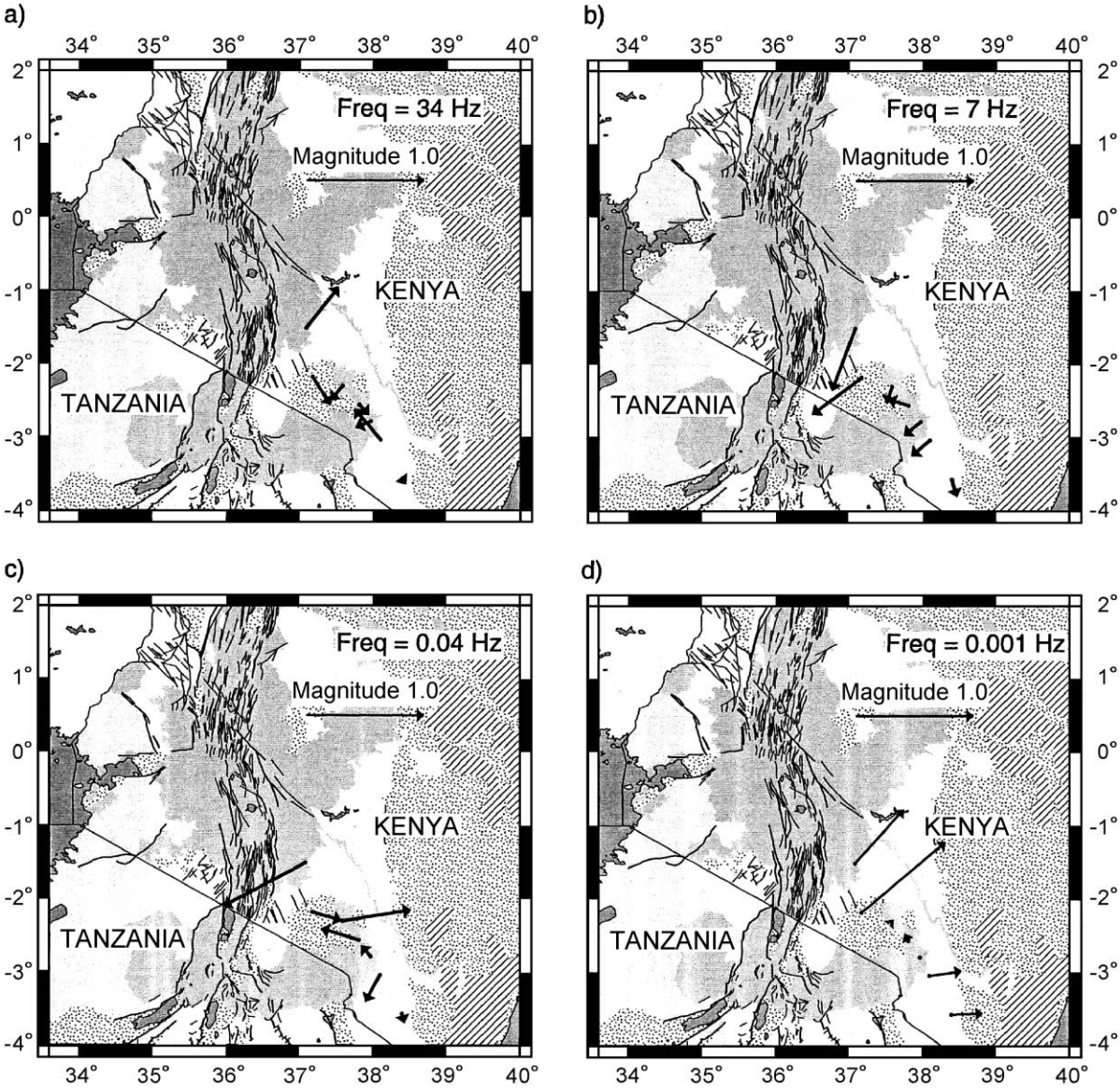


Fig. 4. Real Parkinson induction arrows for (a) 34 Hz, (b) 7 Hz, (c) 0.04 Hz, and (d) 0.001 Hz superimposed on the regional fault fabric. Station Wya is not shown since there are no high-frequency recordings at this site.

Ch2. It is noteworthy that away from the main ridge (stations ChN, ChW, and Ch2), the induction arrows at low frequencies (ca. 0.001 Hz) seem to suggest the presence of a NW–SE conductive axis to the east of the Chyulu Hills. Simpson et al. suggested that this may be the signature of the NW–SE basement structural lineations (Shackleton, 1986), some of which possibly controlled the localisation of the volcanic chain (Smith and Mosley, 1993). It would thus appear that the interpretation of the composite (high and low frequency) data set in terms of a regional 2-D model could be fraught with difficulties since the composite MT data seem to show 3-D characteristics over the main ridge. However, full-domain 3-D modelling (Druskin and Knizhnerman, 1988; Mackie et al., 1993; Smith, 1996a,b) is beyond the scope of the present pilot study and more importantly, it will be more appropriate for 3-D data arrays. Our single survey line with poor spatial coverage does not justify a rigorous 3-D modelling approach since the resulting model cannot be unequivocal. For consistency with the 2-D seismic refraction and gravity models for the region (Novak et al., 1997a,b), a 2-D approach will be adopted in this pilot study. For 2-D modelling, MT responses were obtained in the NW–SE (N40°W) direction and for the orthogonal direction (N50°E), which for notational simplicity will be respectively termed the transverse magnetic (TM) and transverse electric (TE) modes data here. The sounding curves for the TE and TM modes have been respectively corrected for static shift using central-loop and single-loop TEM data as suggested by Meju et al. (1999, Fig. 6).

3. 2-D MT data inversion for deep crustal structure

For 2-D MT data inversion, we have used a popular nonlinear conjugate-gradient, finite-difference-based inversion program (Mackie, 1996; Mackie et al., 1988; Mackie and Madden, 1993). The TE and TM apparent resistivities and phases (1496 data points) and their associated standard observational errors constitute the data for inversion; the threshold rms data misfit was set at 1.4. Two kinds of inverse modelling studies were undertaken—one involving the use of structured initial models derived with the aid of constraints furnished by 1-D inversion of TEM and MT data sets and available seismic models, and the other using only featureless (half-space or smooth) models. The undergirding philosophy is that any common features in the resulting 2-D models are warranted by the data and may be deemed geologically significant.

3.1. Inversion for maximum structure

We first inverted the MT data using a starting model derived by joint 1-D inversion of TEM and MT data (cf. Meju, 1996) and augmented with the middle–lower crustal-layered aspects furnished by the seismic model for this line (Novak et al., 1997b, Fig. 6). Note that there was no attempt to force the MT model into conformity with the seismic model; the iterative parameter update scheme was free to modify our initial model as appropriate to match the field data. The resulting optimal model is shown in Fig. 5. The fit between the computed model response and the observed apparent resistivity and phase data at all, but Wyandani station, are satisfactory as shown in Fig. 6. The resistivity model (Fig. 5) suggests the presence of strong lateral changes in resistivity in the top 20 km of the crust. The nonvolcanic northern end of the profile (positions 0–20 km), i.e., Lukenya area, is characterised by a 10–12-km-thick, highly resistive (100–4000 Ω m) upper crustal unit and an underlying relatively conductive (< 50 Ω m) middle crustal unit. It would appear that a highly conductive layer exists at about 20-km depth in the Lukenya area. The resistive upper crustal unit is much thinner (< 3 km) at profile positions of 125–210 km and appears to be overlain by conductive materials in the Selengei area. Major conductive (< 10 Ω m) steep zones are suggested at mid-crustal depths (8–23 km) in the Chyulu North area (profile positions 100–125 km) and immediately north of Chyulu South (profile positions 190–210 km). An anomalous, 4–5-km-thick, highly conductive (0.01–10 Ω m) body possibly exists in the upper crust at profile positions of 140–185 km (i.e., top 10 km of the main ridge). This particular body appears to dip gently (4°–5.5°) to the northwest and is bounded in the northwest and southeast by the steep zones mentioned above. The area around Wyandani appears to be highly resistive but the lack of high-frequency MT data would suggest limited model resolution. A conductive (< 10 Ω m) vertical zone extends from about 1–17-km depth around Mwatate.

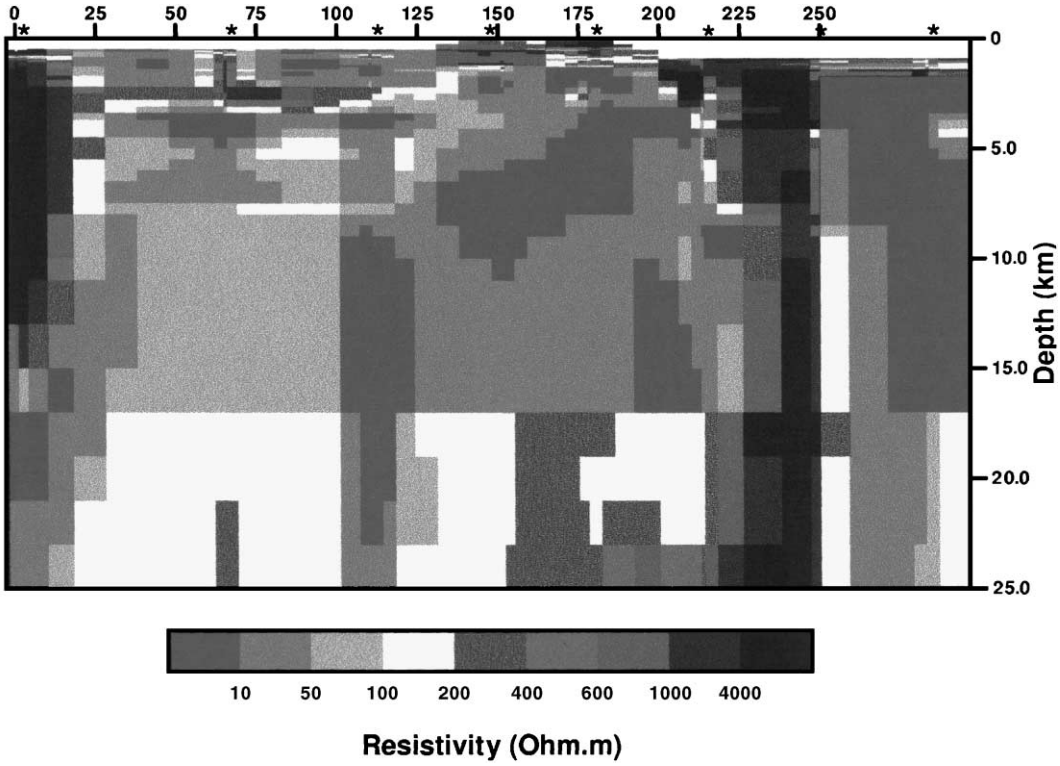


Fig. 5. A resistivity model derived by 2-D inversion with a structured initial model. The numbers at the top of the section are the distances in km along the coincidentally located seismic line of Novak et al. (1997a).

3.2. Inversion for minimum structure

An effective practical strategy when inverting a scanty set of noisy field data is to seek the smoothest model that can reproduce the main features of the data (e.g., Constable et al., 1987; Smith and Booker, 1988; deGroot-Hedlin and Constable, 1990; Meju and Hutton, 1992). For the Chyulu Hills 2-D data inversion problem, the subsurface was discretised into a large number of rectangular blocks of initially constant resistivity (i.e., a half-space initial model) and we sought an optimum model with minimized differences between the resistivities of adjacent blocks in both the vertical and horizontal directions using the Twomey–Tikhonov (Twomey, 1963; Tikhonov, 1963) derivative regularisation measures. In the ideal situation with a dense network of measurement stations, only those features that are essential for fitting the field data will be retained or recovered in the resulting optimal model. In the present case with widely spaced stations, only those features in the neighbourhood of the sounding stations may be accepted as justified by the data. The use of different half-space starting models (cf. Bai et al., 2001) provides a simple consistency check for model resolution.

The TE and TM modes data were simultaneously inverted using various half-space initial models and Twomey–Tikhonov smoothness constraints. For brevity, only the resulting models for the 100 and 500 Ω m half-spaces (Fig. 7a and b) will be discussed here. For these 2-D models, the fit between the calculated and observed apparent resistivity and phase data is marginally better than that shown in Fig. 6. The model generated from the 100-Ω m starting model (Fig. 7a) appears to suggest the presence of major conductivity anomalies in the areas of Late Pleistocene volcanism; note the conductive (<10 Ω m) zones in the upper and middle crusts with associated narrow steep zones underneath the main ridge and near its margins. The model obtained from the 500-Ω m half-space initial

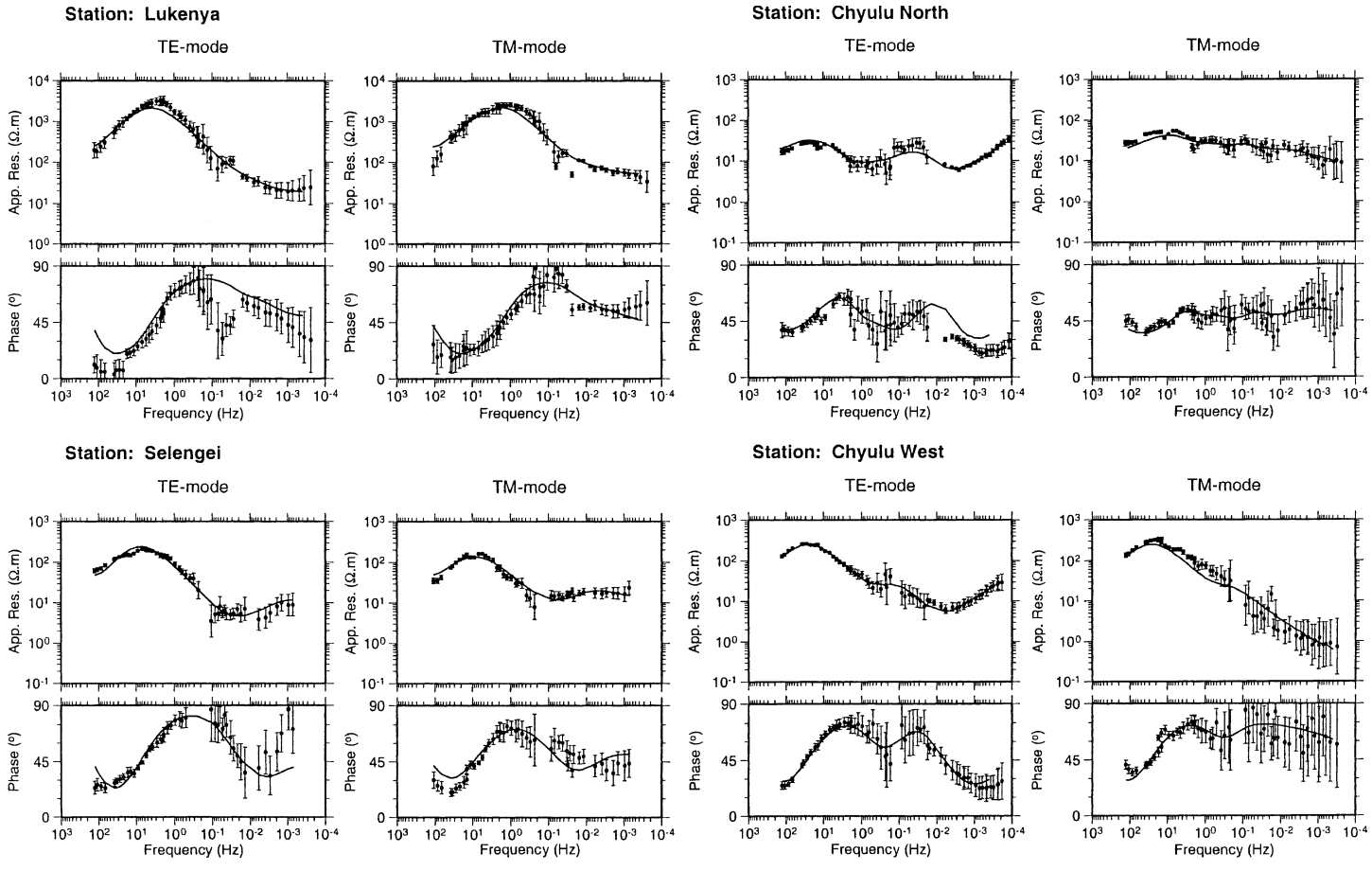


Fig. 6. A comparison of the actual field data and the computed response curves for the 2-D model of Fig. 5.

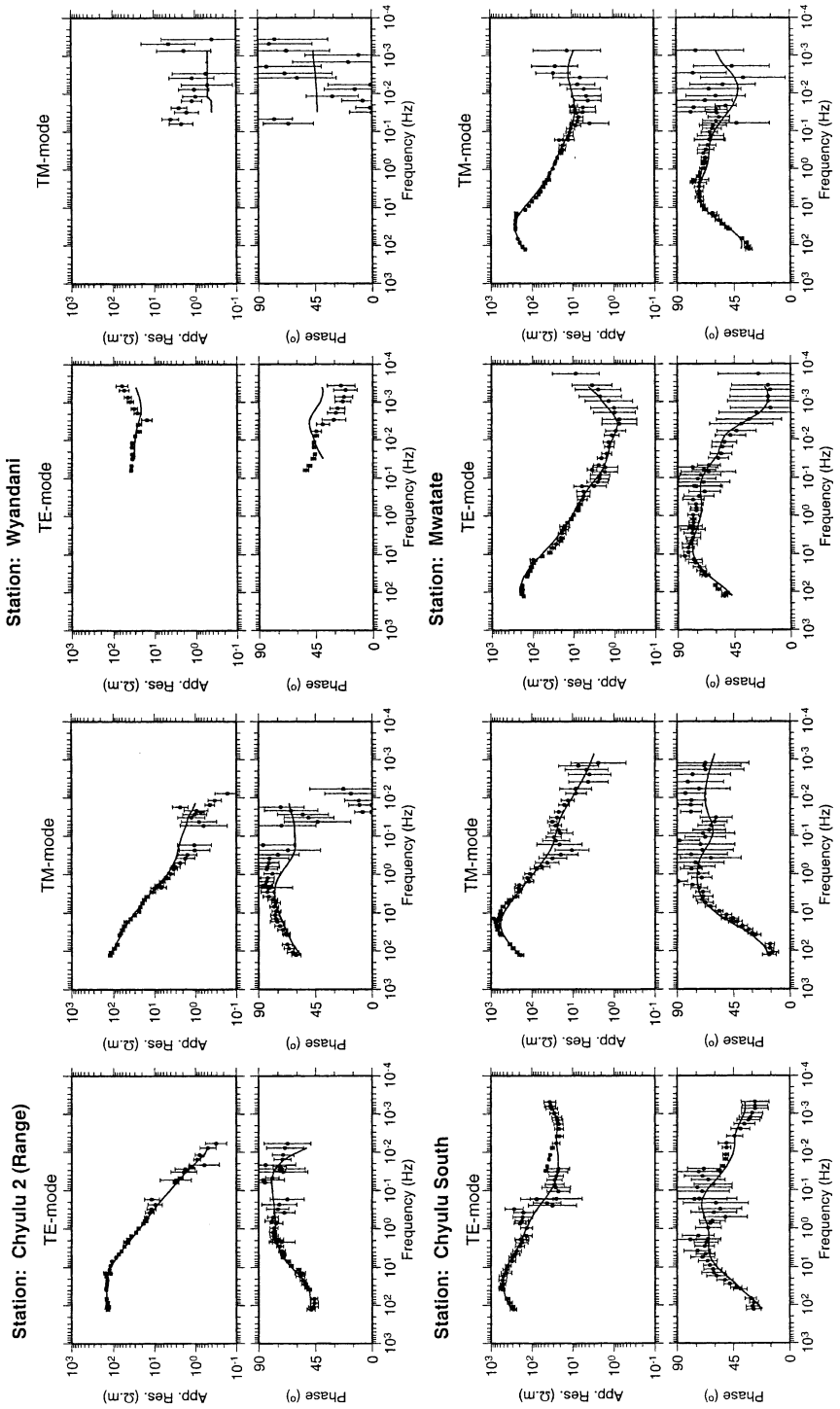


Fig. 6 (*continued*).

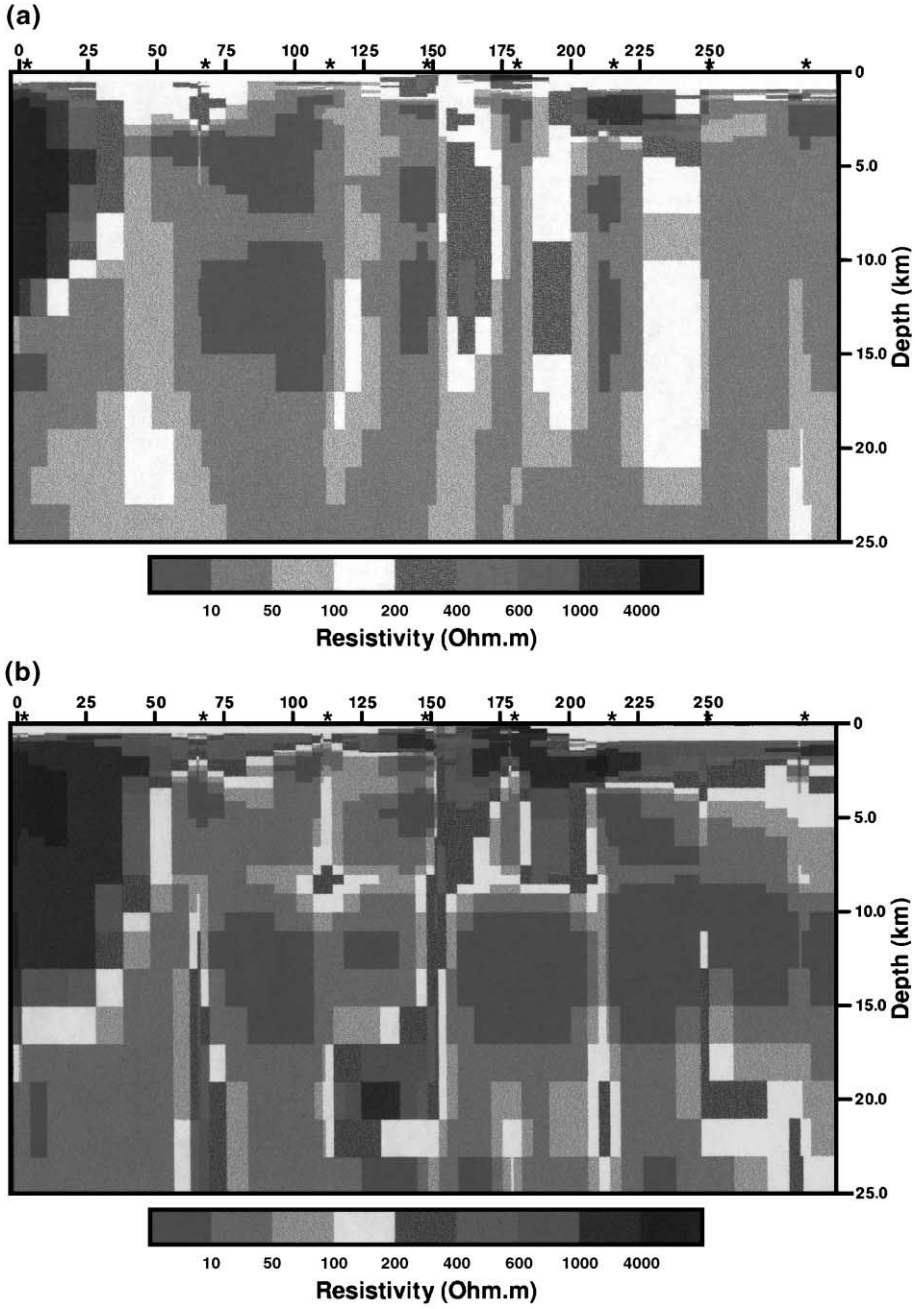


Fig. 7. Examples of 2-D models generated by regularised inversion using different smooth initial models. Shown are the models from (a) 100 Ω m and (b) 500 Ω m initial half-space models. These models are statistically equivalent and highlight the nonuniqueness of models constructed from scanty field observations. Only those features common to both models are constrained by the data.

model is shown in Fig. 7b and suggests a somewhat layered aspect with relative enhancements of the resistive features seen in the other models. The resistive upper crustal unit is clearly shown to vary laterally; it is more than 10 km thick in the Lukenya area, about 5–9 km thick over the volcanic field, less than 1 km around profile position 112 km (ChN station), and about 5 km thick near Mwatate. The thickness of the resistive cover unit is not well resolved underneath the main ridge. It would appear that the upper crust contains conductive bodies in its bottom part around profile positions 60 and 250 km. The middle crust (9–20-km depth) appears to be anomalously conductive ($<10\ \Omega$ m) southward from Selengei.

The main features common to these models (and to the model of Fig. 5) are therefore: (i) the resistive upper crust which is thinnest over profile positions 100–150 km, (ii) the horizontal conductive ($<10\ \Omega$ m) zones in the upper crust across the Chyulu Hills and north of Mwatate, and (iii) the steep anomalous deep-reaching conductive zones near profile positions 100–150 and 175–225 km (i.e., bordering the main ridge). These low-resistivity anomalies may be geologically significant and could be vestiges of the magmatic activities that took place in the region; the steep zones bordering the main ridge may be faults or magma conduits.

4. Discussion

The seismic refraction P-wave model of Novak et al. (1997a, Fig. 8) shows a simple-layered crustal structure with prominent refractors at 9–10-km depth and at 19–21-km depth above a 22–23-km-thick lower crust with no major lateral variations but their revised interpretative model (Novak et al., 1997b, Fig. 6) suggests the presence of a low-velocity body in the lower crust (30–44 km deep) beneath the volcanic field. There is thus a good agreement between the depth to the base of the resistive upper crustal and conductive mid-crustal geoelectric units in the electromagnetic transect and the independently obtained depths to the crustal refractors of Novak et al. (1997a). Ritter and Kaspar (1997) also demonstrated the presence of lateral velocity contrasts of about 5% with a prominent low-velocity zone located at profile positions of 100–150 km, directly beneath the volcanic range down to 70-km depth; Novak et al. (1997b, Fig. 7b) show two low-velocity zones around profile positions of 100 and 150–175 km. Despite the wide spacings between our TEM–MT stations which limited model resolution to a large extent, the constructed resistivity models strongly suggest that there are major laterally varying structures in the top 20 km of the crust and more significantly, that the area around profile positions 100–200 km contains major low-resistivity bodies at depths of about 4 km down to at least 19 km in the 2-D models.

We suggest that the low-resistivity features in the MT models are related to the low-velocity zones detected at greater depth in the same geographic location by the abovementioned investigators. If this contention is correct, then our MT results provide a necessary upper crustal constraint on the structure of the anomalous zone in Chyulu Hills. Note that Novak et al. (1997b) favoured an interpretation of the anomalous seismic signature in terms of partial melt (or magma chambers) in the lower crust or upper mantle. Since the volcanic rocks on the surface contain abundant crustal-derived xenoliths but lack lower crustal and mantle-derived xenoliths (see Novak et al., 1997b), we speculate that the conductors found below the main ridge especially around profile positions 100–200 km could be magma chambers or conduits. This geographical location is also associated with a low gravity signature (see Novak et al., 1997b, Fig. 8). The other conductive anomaly south of the Chyulu Hills may be a hallmark of the volcanic activity of the past century (Saggerson, 1963). Overall, there is good agreement between the seismic, gravity, and MT models for the Chyulu Hills and we may suggest that MT will be a logical compliment to seismics in the exploration of the deep crust in this terrain.

5. Conclusion and suggestions for further studies

The results of quantitative interpretation of the invariant and rotated MT apparent resistivity and phase sounding curves suggest that the top 25 km of the earth's crust in the Chyulu Hills volcanic field comprises three main geoelectric units. The topmost unit is highly resistive and shows a progressive southward thinning; it is thickest in the Lukenya area located 30 km south of Nairobi, and thinnest around Mwatate

located 160 km north of Mombasa. Directly beneath the Chyulu Hills, the uppermost and the underlying geoelectric units have been dissected by zones of low resistivities (< 10 Ω m) interpreted as possible vestiges of faulting or magmatic activities in the area. The southern and northern edges and the midsection (near Chyulu West) of the main range contain anomalous steep conductors in the depth range of 1–19 km which may be deep conduits for magma migration in Late Pleistocene to recent times. It is possible that the anomalous conductive (0.01–1 Ω m) body directly beneath Chyulu West and Chyulu Range stations may be an upper crustal magma chamber.

This study was severely hampered by the lack of dense spatial coverage of the field measurements, thus militating against any rigorous 3-D EM modelling. It will be unrealistic to undertake 3-D modelling of our current data sets acquired along a single profile. Therefore, it is recommended that a detailed 3-D field survey be undertaken in this region involving broadband (1000–0.00001 Hz) MT and TEM soundings with station spacings not exceeding 2–5 km (as for the seismic and gravity components of the KRISP94-MT collaborative project). Three-dimensional TEM–MT modelling will be appropriate for the resulting dense network of observations. Using an available 3-D modelling code (Mackie et al., 1993), Simpson (2000) recently produced an MT model for southern Kenya Rift emphasising the influence of the preexisting NW–SE shear zones (suggested by Smith and Mosley, 1993) on the MT responses recorded along another profile running east–west across the main rift. Although simplistic due to the scanty data set available, the model is a useful step towards understanding the deep structure of the region. It will be desirable to undertake a similar study for the Chyulu Hills line but with additional depth soundings and with the 3-D modelling interpretation adequately incorporating the known geological features of the area.

Acknowledgements

This study was conducted as part of the KRISP94-MT project and was funded by the European Union Human Capital and Mobility program under contract ERBCHRXCT930308. We thank P. Denton, M. Eisel, P. Denny, S. Onacha, B.P. Ouma, C. Birt, and S. Sundqvist for assistance with the fieldwork. We are grateful to the UK Natural Environment Research Council Equipment Pool for a loan of the S.P.A.M. Mk 2B field equipment. The field studies benefited from the use of the Sirotem Mk2 equipment belonging to Leicester University and RAP LMT data loggers belonging to GeoForschungsZentrum Potsdam. We thank the Kenyan Government for permission to undertake this study.

References

Bai, D.B., Meju, M.A., Liao, Z., 2001. Magnetotelluric images of deep structure of the Rehai geothermal field near Tengchong, southern China. Geophys. J. Int. 147, 677–687.

Bahr, K., 1988. Interpretation of the magnetotelluric impedance tensor: regional induction and local telluric distortion. J. Geophys. 62, 119–127.

Bahr, K., 1991. Geological noise in magnetotelluric data: a classification of distortion types. Phys. Earth Planet. Inter. 66, 24–38.

Baker, B.H., 1987. Outline of the petrology of the Kenya Rift alkaline province. In: Fitton, J.G., Upton, B.G.J. (Eds.), Alkaline Igneous Rocks. Geol. Soc. Publ., vol. 30, pp. 293–312.

Baker, B.H., Wohlenberg, J., 1971. Structure and evolution of the Kenya Rift Valley. Nature 229, 538–542.

Banks, R.J., Beamish, D., 1979. Melting in the crust and upper mantle beneath the Kenya Rift: evidence from Geomagnetic Deep Sounding experiments. J. Geol. Soc. (London) 136, 225–233.

Banks, R.J., Ottey, P., 1973. Geomagnetic deep sounding in and around the Kenya Rift Valley. Geophys. J. R. Astron. Soc. 36, 321–335.

Beamish, D., 1977. The mapping of induced currents around the Kenya Rift: a comparison of techniques. Geophys. J. R. Astron. Soc. 50, 311–332.

Berdichevsky, M.N, Dmitriev, V.I., 1976. Distortion of magnetic and electrical fields by near-surface inhomogeneities. Acta Geod., Geophys. Mont. Hung. 11, 447–483.

Birt, C.S., Maguire, P.K.H., Khan, M.A., Thybo, H., Keller, G.R., Patel, J., 1997. The influence of pre-existing structures on the evolution of the southern Kenya Rift Valley—evidence from seismic and gravity studies. In: Fuchs, K., Altherr, R., Muller, B., Prodehl, C. (Eds.), Structure and Dynamic Processes in the Lithosphere of the Afro-Arabian Rift System. Tectonophysics, vol. 278, pp. 211–242.

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

Christopherson, K.R., 1991. Application of magnetotellurics to petroleum exploration in Papua New Guinea: a model for frontier areas. Geophysics: The Leading Edge of Exploration 10, 21–27.

Class, C., Altherr, R., Volker, F., Eberz, G., McCulloch, M.T., 1994. Geochemistry of Pliocene to Quaternary alkali basalts from the Huri Hills, northern Kenya. Chem. Geol. 113, 1–22.

Constable, S.C., Parker, R.L., Constable, C.G., 1987. Occam's inversion: a practical algorithm for generating smooth models from electromagnetic sounding data. Geophysics 52, 289–300.

deGroot-Hedlin, C., Constable, S., 1990. Occam's inversion to generate smooth two-dimensional models from magnetotelluric data. Geophysics 55, 1613–1624.

Druskin, V.L., Knizhnerman, L.A., 1988. A spectral semi-discrete method for the numerical solution of 3-D nonstationary problems in electrical prospecting. Phys. Solid Earth 24, 63–74.

Egbert, G.D., Booker, J.R., 1986. Robust estimation of geomagnetic transfer functions. Geophys. J. R. Astron. Soc. 87, 173–194.

Fontes, S.L., Harinarayana, T., Dawes, G.J.K., Hutton, V.R.S., 1988. Processing of noisy magnetotelluric data using digital filters and additional data selection criteria. Phys. Earth Planet. Inter. 52, 30–40.

Gregory, J.W., 1921. The Rift Valleys and Geology of East Africa. Seely Service, London, 479 pp.

Groom, R.W., Bahr, K., 1992. Corrections for near-surface effects: decomposition of the magnetotelluric impedance tensor and scaling corrections for regional resistivities: a tutorial. Surv. Geophys. 13, 341–379.

Groom, R.W., Bailey, R.C., 1989. Decomposition of magnetotelluric impedance tensors in the presence of local three-dimensional galvanic distortion. J. Geophys. Res. 94, 1913–1925.

Groom, R.W., 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.

Haug, G.H., Strecker, M.R., 1995. Volcano-tectonic evolution of the Chyulu Hills and implications for the regional stress field in Kenya. Geology 23, 165–168.

Henjes-Kunst, F., Altherr, R., 1992. Metamorphic petrology of xenoliths from Kenya and northern Tanzania and implications for geotherms and lithospheric structures. J. Petrol. 33, 1125–1156.

Jones, A.G., 1988. Static shift of magnetotelluric data and its removal in a sedimentary basin environment. Geophysics 53, 967–978.

Jones, A.G., 1993. Electromagnetic images of ancient and modern subduction zones. Tectonophysics 219, 29–45.

Jones, F.W., Price, A.T., 1970. The perturbations of alternating geomagnetic fields by conductivity anomalies. Geophys. J. R. Astron. Soc. 20, 317–334.

Karson, J.A., Curtis, P.C., 1989. Tectonic and magmatic processes in the eastern branch of the East African rift and implications for magmatically active continental rifts. J. Afr. Earth Sci. 8, 431–453.

Kaufman, A.A., Keller, G.V., 1983. Frequency and Transient Soundings. Elsevier, Amsterdam.

Khan, M.A., Mansfield, J., 1971. Gravity measurements in the Gregory Rift. Nature 229, 72–75.

Macdonald, R., 1994. Petrological evidence regarding the evolution of the Kenya Rift Valley. In: Prodehl, C., Keller, G.R., Khan, M.A. (Eds.), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, vol. 236, pp. 373–390.

Macdonald, R., Williams, L.A.J., Gass, I.G., 1994. Tectonomagmatic evolution of the Kenya Rift Valley: some geological perspectives. J. Geol. Soc. (London) 151, 879–888.

Mackie, R.L., 1996. Two-dimensional inversion of magnetotelluric data: Software. Internet site: MTNET http://nazca.cg.emr.ca/mtnet/mtnet.html.

Mackie, R.L., Madden, T.R., 1993. Conjugate direction relaxation solutions for 3-D magnetotelluric modelling. Geophysics 58, 1052–1057.

Mackie, R.L., Bennett, B.R., Madden, T.R., 1988. Long period magnetotelluric measurements near the central California coast: a land-locked view of the conductivity structure under the Pacific Ocean. Geophys. J. R. Astron. Soc. 95, 181–194.

Mackie, R.L., Madden, T.R., Wannamaker, P.E., 1993. Three-dimensional MT modelling using difference equations—theory and comparison to integral equation solutions. Geophysics 58, 215–226.

Meju, M.A., 1996. Joint inversion of TEM and distorted MT soundings: some effective practical considerations. Geophysics 61, 56–65.

Meju, M.A., Hutton, V.R.S., 1992. Iterative most-squares inversion: application to magnetotelluric data. Geophys. J. Int. 108, 758–766.

Meju, M.A., Fontes, S.L., Oliveira, M.F.B., Lima, J.P.R., Ulugergerli, E.U., Carrasquilla, A.A., 1999. Regional aquifer mapping using combined VES–TEM–AMT/EMAP methods in the semi-arid eastern margin of Parnaiba Basin, Brazil. Geophysics 64, 337–356.

Morley, C.K., Wescott, W.A., Stone, D.M., Harper, R.M., Wigger, S.T., Karanja, F.M., 1992. Tectonic evolution of the northern Kenya Rift. J. Geol. Soc. 149, 333–348.

Novak, O., Prodehl, C., Jacob, A.W.B., Okoth, W., 1997a. Crustal structure of the southern flank of the Kenya Rift deduced from wide-angle P-wave data. In: Fuchs, K., Altherr, R., Muller, B., Prodehl, C. (Eds.), Structure and Dynamic Processes in the Lithosphere of the Afro-Arabian Rift System. Tectonophysics, vol. 278, pp. 171–186.

Novak, O., Ritter, J.R.R., Altherr, R., Garasic, V., Volker, F., Kluge, C., Kaspar, T., Byrne, G.F., Sobolev, S.V., Fuchs, K., 1997b. An integrated model for the deep structure of the Chyulu Hills volcanic field, Kenya. In: Fuchs, K., Altherr, R., Muller, B., Prodehl, C. (Eds.), Structure and Dynamic Processes in the Lithosphere of the Afro-Arabian Rift System. Tectonophysics, vol. 278, pp. 187–209.

Omenge, J.M., Okele, R.E., 1992. Geology of the Chyulu-Oloitokitok area. Kenya Mines and Geology Department, Nairobi, Geological Survey of Kenya, Report No. 112.

Parkinson, W.D., 1959. Directions of rapid geomagnetic fluctuations. Geophys. J. R. Astron. Soc. 2, 1–14.

Parkinson, W.D., 1983. Introduction to Geomagnetism Scottish Academic Press, Edinburgh, 433 pp.

Pellerin, L., Hohmann, G.W., 1990. Transient electromagnetic inversion: a remedy for magnetotelluric static shifts. Geophysics 55, 1242–1250.

Peters, W.S., de Angelis, M., 1987. The Radio Hill Ni–Cu massive sulphide deposit: a geophysical case history. Explor. Geophys. 18, 160–166.

Raiche, A.P., 1984. The effect of ramp function turnoff on the TEM response of a layered earth. Bull. Aust. Soc. Explor. Geophys. 15, 37–42.

Ritter, J.R.R., Kaspar, T., 1997. A tomography of the Chyulu Hills,

Kenya. In: Fuchs, K., Altherr, R., Muller, B., Prodehl, C. (Eds.), Structure and Dynamic Processes in the Lithosphere of the Afro-Arabian Rift System. Tectonophysics, vol. 278, pp. 149–169.

Rooney, D., Hutton, V.R.S., 1977. A magnetotelluric and magnetovariational study of the Gregory Rift Valley, Kenya. Geophys. J. R. Astron. Soc. 51, 91–119.

Saggerson, E.P., 1963. Geology of the Simba–Kibwezi area. Geol. Surv. Kenya, Rep. 58, 70 pp.

Sakkas, V.A., 1999. Combined transient electromagnetic and magnetotelluric study of the southern Kenya Rift Valley. PhD thesis, University of Leicester, UK, 250 pp.

Shackleton, R.M., 1986. Precambrian collision tectonics in Africa. In: Coward, M.P., Ries, A.C. (Eds.), Collision Tectonics. Geol. Soc. London, Spec. Publ., vol. 19, pp. 329–349.

Simpson, F., 2000. A three-dimensional electromagnetic model of the southern Kenya Rift: departure from two-dimensionality as a possible consequence of a rotating stress field. J. Geophys. Res. 105, 19321–19334.

Simpson, F.L., Haak, V., Khan, M.A., Sakkas, V., Meju, M.A., 1997. The KRISP-94 magnetotelluric survey of early 1995: first results. Tectonophysics 278, 261–271.

Smith, M., 1994. Stratigraphic and structural constraints on mechanisms of active rifting in the Gregory Rift, Kenya. In: Prodehl, C., Keller, G.R., Khan, M.A. (Eds.), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, vol. 236, pp. 3–23.

Smith, J.T., 1996a. Conservative modeling of 3-D electromagnetic fields: Part I. Properties and error analysis. Geophysics 61, 1308–1318.

Smith, J.T., 1996b. Conservative modeling of 3-D electromagnetic fields: Part II. Biconjugate gradient solution and an accelerator. Geophysics 61, 1319–1324.

Smith, J.T., Booker, J.R., 1988. Magnetotelluric inversion for minimum structure. Geophysics 53, 1565–1576.

Smith, M., Mosley, P., 1993. Crustal heterogeneity and basement influence on the development of the Kenya Rift, East Africa. Tectonics 12, 591–606.

Sternberg, B.K., Washburne, J.C., Pellerin, L., 1988. Correction for the static shift in the magnetotellurics using transient electromagnetic soundings. Geophysics 53, 1459–1468.

Swift, C.M., 1967. A magnetotelluric investigation of electrical conductivity anomaly in the southwestern United States. PhD Thesis, M.I.T. In: Vozoff, K. (Ed.), 1986, Magnetotelluric Methods, SEG, 1986.

Tikhonov, A.N., 1963. Regularization of ill-posed problems. Dokl. Akad. Nauk SSSR 153, 1–6.

Twomey, S., 1963. On the numerical solution of Fredholm integral equations of the first kind by the inversion of the linear system produced by quadrature. J. Assoc. Comput. Man. 10, 97–101.

Wheildon, P., Morgan, P., Williamson, K.H., Evans, T.R., Swanberg, C.A., 1994. Heat flow in the Kenya Rift zone. In: Prodehl, C., Keller, G.R., Khan, M.A. (Eds.), Crustal and Upper Mantle Structure of the Kenya Rift. Tectonophysics, vol. 236, pp. 131–149.

Williams, L.A.J., 1978. Character of Quaternary volcanism in the Gregory Rift Valley. In: Bishop, W.W. (Ed.), Geological Background to Fossil Man. Scottish Academic Press, Edinburgh, pp. 55–69.