GEOMAGNETIC SOUNDING OF CONDUCTIVITY ANOMALIES IN THE LOWER CRUST AND UPPERMOST MANTLE

B.P. SINGH

Indian Institute of Geomagnetism, Colaba, Bombay 400 005, India

Abstract. Knowledge of the structure of the lower crust and uppermost mantle is of special importance in understanding plate tectonics. Electrical conductivity of this region has been measured recently in various parts of the world. Transfer functions are still the most widely used quantity in data analysis and model fitting. Anomalies in the horizontal magnetic components in combination with anomalies in the vertical component have been found very useful in locating conductivity contrasts. With these, when the cause of the anomaly is a concentrated line current, both the position of the line current and its depth can be directly located. The method of hypothetical event analysis is another new technique and this is highly suited for areas having complex subsurface geology or areas under non-uniform source fields or both. The anomaly is more suitable for modelling geophysical structures when it is separated into regional and local components. Model calculations still are not very satisfactory and the importance of one dimensional calculations must be emphasized for they give direct information on the variation of conductivity with depth, which is the purpose of GDS. We need more results, especially from tectonically active areas, before the underlying physical processes can be completely understood.

Introduction

In this paper, we will consider electrical conductivity anomalies in the lower crust and upper most mantle. For the basic theory of geomagnetic deep sounding (GDS) Rikitake (1966), Price (1967) and Schmucker (1970) with Ádám (1976a) constitute a good set of references. Arrays of magnetometers have been used in various parts of the world and results have shown considerable departures from spherical symmetry in the electrical conductivity of the earth. Areas having conductivities higher than 0.1 Sm^{-1} have been found at depths less than 200–300 km. Shankland and Waff (1977) have listed such anomalous areas. They are, the Rio Grande rift (Schmucker, 1964), the Wasatch fault belt and the Southern Rocky Mountains (Porath, 1971; Porath and Gough, 1971), the East African rift (Banks and Ottey, 1974), the Rhine Graben (Winter, 1973), Western Canada (Caner, 1971), Hawaii (Larsen, 1975), Iceland (Hermance and Grillot, 1970) and Yellowstone (Leary and Phinney, 1974). Some of the more recent studies discussing anomalous zones are: Kenya (Beamish, 1977; Rooney and Hutton, 1977), Carpathians (Rokityanski *et al.*, 1975; Jankowsky *et al.*, 1977 a and b), Scotland (Hutton *et al.*, 1977), Southern Africa (de Beer *et al.*, 1975; de Beer *et al.* 1976), Bermuda (Vanyan *et al.*, 1978); Northern Pyrenees (Vasseur *et al.*, 1977; Babour 1977), northeast of Quebec city (Honkura *et al.*, 1977), Southern half of Hokkaido, Japan (Nishida, 1976); and peninsular India (Rajaram *et al.*, 1978). These studies have identified many new areas having highly conductive bodies in the lower crust or upper mantle. Salient features of these studies will be covered in later sections. It is well known that both the structure of

Geophysical Surveys 4 (1980)71-87. 0046-5763/80/0041-0071$02.55.

the source field and the goelogical features of the medium in which currents are induced, vary from region to region and hence it is natural to expect that the mechanism of electromagnetic induction in the Earth – and hence in the planets – will only be fully understood when we have covered all possible cases.

Analyses of Data

The systematic features of transient variations of the Earth's magnetic field are used in GDS in studying subsurface geology. From the skin depth relationship, we know that variations of interest for studying anomalies in the lower crust and upper mantle cover the period range from pulsations to substorms. Their periods range from 10 sec to about 60 min, sometimes supplemented by data from quiet daily (Sq) magnetic variations. Data from a single station in general are not enough for detailed studies and usually one runs an array of magnetometers such as those designed by Gough and Reitzel (1967) or fluxgate magnetometers.

Another important development in instrumentation has been the use of highly sensitive directional magnetometers (Babour and Mosnier, 1977). The stability of this instrument is 1/20 nT for periods between 5 sec and 10 min, 1/5 nT for periods in the range of hours and 1/2 nT for periods of the order of several days. They found the instruments very useful in finding anomalies in the horizontal component. A telemetric link made possible the instantaneous comparison of variations of components at several stations.

The techniques for processing geomagnetic variation data to delineate conductivity contrasts have been reviewed recently by Banks (1973) and Frazer (1974). The contributions due to the source fields should ideally be removed from the observed variations. However, most commonly the size of the array and the number of magnetometers recording simultaneously are too small to permit a separation of the field into parts of internal and external origin. Even where it has been tried (Camfield 1973; Gough 1973) the results do not establish its superiority over other methods. However, Frazer (1974) emphasized that before this technique is abandoned, separation of array data collected over a complicated structure involving curved or multiple conductors should be attempted. In a recent study Jankowski *et al.* (1977 a and b) have separated the field into its constituent internal and external parts and their results are very interesting. These will be discussed later.

Several alternative methods of analysis have been adopted in GDS studies. The method of induction vectors developed by Wiese (1962) and Parkinson (1959) independently has been a very simple and elegant method to identify conductivity contrasts and it still stands as the most widely used method. Later attempts to make the analysis more quantitative and amenable to model calculations resulted in the introduction of transfer functions by Schmucker. Methods for estimating transfer functions $A(f)$ and $B(f)$ are described in detail by Schmucker (1964) and by Everett and Hyndman (1967). The technique as it stands today requires estimation of the transfer functions A and B as a function of frequency and then their fitting to a model structure which reproduces both the period and spatial variations of the A's and B's over the area of interest. The Dallas and Alberta groups (Gough 1973) introduced the use of maps of Fourier transform amplitudes and phases in GDS studies using data from magnetic arrays. These maps are very useful in qualitative studies. They are normally used to identify the anomalous regions and the area of interest may then be modelled directly or studied through transfer function techniques.

Some recent developments have considerably improved the utility of the transfer function technique of Schmucker (1970). One such has been the hypothetical event analysis (HEA) developed by Bailey *et al.* (1974) in which contour maps of the anomalous vertical component Z are prepared for a hypothetical uniform horizontal inducing field of unit amplitude. Maps are prepared for source fields having different directions of polarization. By using appropriate values of the X and Y components the polarization of the source field can be varied and the corresponding value of Z can be synthesised from the knowledge of $A(f)$ and $B(f)$ through the relation:

$$Z = A(f)\, X + B(f)\, Y\,.$$

The transfer functions are usually estimated from several events. An advantage of the HEA technique is that an observed source field of possibly complicated geometry and polarization may be replaced by a uniform linearly polarized source field. This aspect makes HEA suited for equatorial and auroral regions. Another advantage is that all stations of the array need not be run simultaneously. These points have been discussed by Hutton *et al.* (1977). Anomalies near a curved conductor will be strongly dependent on polarization and the HEA technique will be of special significance in such cases. The method has been used with success by Hutton *et al.* (1977), Beamish (1977), and Rooney and Hutton (1977).

A second important development in data analysis has been the use of the method of complex demodulates (Banks, 1975). Here the variation of amplitude and phase with time of selected frequency components are first separated. The method is suited for short data lengths, and even a record of one good magnetic storm may provide enough data for the estimation of transfer functions. This approach has been adopted by Agarwal *et al.* (1979) and they have found that from the complex demodulates, of only one magnetic storm, they could estimate the $A(f)$'s and $B(f)$'s as accurately as those determined by a large number of events in the conventional methods. Beamish (1977) has also used this technique and has found the method useful.

Earlier conductivity studies used only anomalies in the vertical component, Z, in the analysis but many recent studies have considered explicitly the anomalies in all three components. Anomalies in the horizontal components are more pronounced when the cause of the anomaly is channelled currents, which are induced in regions away from the observing sites. There is an obvious advantage in using the anomaly in the horizontal component (H) to locate conductors, since over the centre of conductors H_a has a maximum and falls rapidly to zero as we move away from it. By contrast Z_a is zero immediately above a conductor attaining maximal values at the edges of the conductor. A line current at a depth d from the point of observation, will produce an H field of width $2d$ at the half-maximum and a Z field with extremes $2d$ apart. The location of the current cannot be deeper than the equivalent current that fits the anomaly width data (Gough, 1973). If the conductor is not a thin conductor, but an extended one, the edge of the conductor can be determined from the Z-field and the H-field can be used to measure its lateral extent. This method has been used in locating conductors in many of the recent studies e.g. de Beer *et al.* (1976), Beamish (1977) and Jankowski *et al.* (1977a). As a word of caution, it is worth pointing out that while using the horizontal anomaly for determination of depth, it should be borne in mind that the field has two components and for reliable results both of these components should be considered.

The normal approach of estimating transfer functions through the relation

$$Z = AH + BD$$

has always emphasized that the method should be used for midlatitudes only, where normal $Z(Z_n) \not\approx 0$. Lilley and Sloane (1976) have considered also the formulae for horizontal layering given by Schmucker (1970) and suggested that the total field Z considering both its normal and anomalous part should perhaps be

$$Z = AH + BD + C\left(\frac{\partial H}{\partial x} + \frac{\partial D}{\partial y}\right)$$

The parameter C contains information on the depth and conductivity of the conductor. The real part of C gives the depth and the imaginary (C_i) gives the conductivity, through

$$\sigma = (0.8\,\pi\,\omega\,C_i^2)^{-1}$$

where σ is in Sm^{-1} and C_i is in km. Lilley and Sloan (1976) find the gradient method suitable for the study of conductivity of regions under non-uniform source fields. For bays and substorms at higher latitudes and for daily variations at mid-latitudes, the gradient method is good. Schmucker (1978) has recently given a method for calculating Z_n and it may be important to consider the complete equation in computations of A and B.

Two other important results of relevance to data analysis are those from Honkura *et al.* (1977) and from Babour (1977). Honkura *et al.* have demonstrated the presence of induction effects associated with Z-variations. Their results are based on observations in a seismically active region northeast of Quebec city. In this area the source field will have a non-zero Z-component. Perhaps, the induction by Z may be important in other areas too where the source fields are non-uniform. Babour (1977) shows that differential components are better suited for delineating conductivity anomalies than the total vertical component. His results are based on records collected in the Rhinegraben.

The above mentioned methods are some of the new techniques of data analysis and the results obtained using these are discussed below.

Results

ESKDALEMUIR ANOMALY

The Eskdalemuir anomaly, a region of high electrical conductivity in southern Scotland, is one of the most extensively studied anomalies. Recently, Hutton *et al.* (1977) have again surveyed this area with an array of 20 three-component recording magnetometers of Gough–Reitzel type. Hutton (1976) has summarized the results from magnetotelluric and magneto-variational surveys of this area. The apparent resistivity for Eskdalemuir in the stratified model as obtained from a magnetotelluric study consists of an uppermost layer of conductivity about 3×10^{-4} Sm^{-1} and thickness about 12 km, overlying a layer of higher than average crustal conductivity of about 2×10^{-2} Sm^{-1} to a depth of 30 km. The layers below this depth have conductivity of about 4×10^{-4} Sm^{-1}. Edwards *et al.* (1971) have suggested that currents induced in the Atlantic Ocean branch into the shallow seas surrounding the British Isles and create a potential difference that drives the current through the anomaly in the crust. The anomaly appears to have a width of approximately 80 km striking NE-SW across South Scotland.

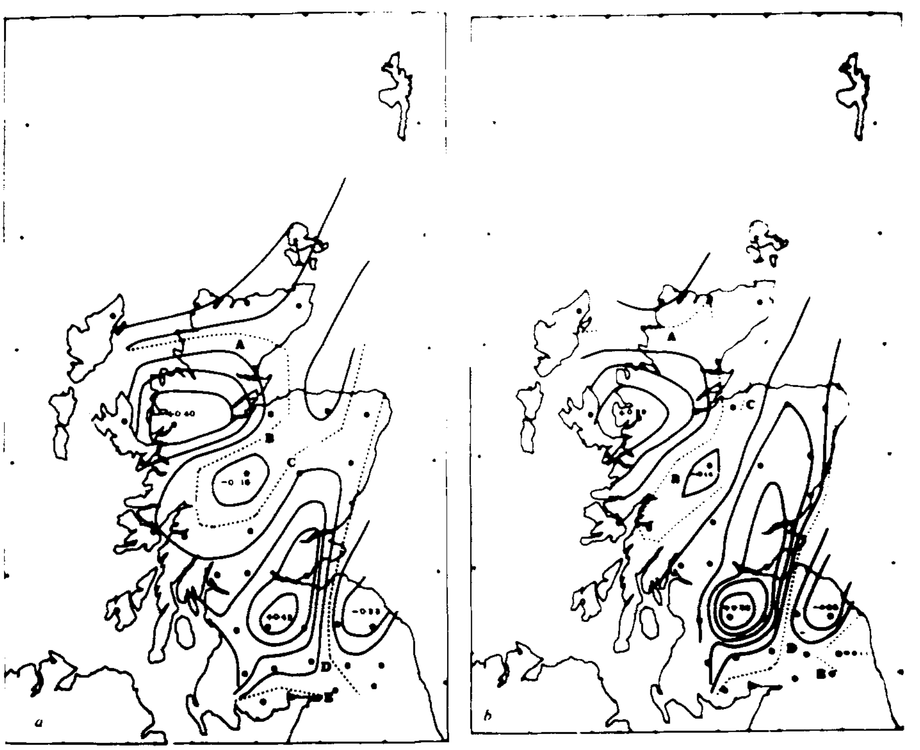

Fig. 1. (a) Z-contours for a horizontal magnetic field of unit amplitude, directed westwards, and having a period of 40 m. (b) same as (a) for a field directed northwards. (From Hutton *et al.*, 1977).

Hutton *et al.* (1977) have drawn Z-contours for hypothetical events in this area (Figure 1). Their technique is particularly suited for this area as the source field itself may cause non-zero Z-variation. The zero contour may represent the region where there is no large gradient in the conductivity. When the current concentration is complex, the above characteristics of zero contours may not be valid. In such cases the Z field gradients are more useful in the analysis. From the width of the Z anomalies, Hutton *et al.* set maximum limits of about 35 km on the depth of the conductor, situated in the crust. An interesting correlation has also been reported between the three regions of zero Z (Figure 1) and two Moinian and one Dalradian ophiolite suites regarded as marks of former subduction zones. The high conductivity could be due to the appearance of hydrated minerals (Hyndman and Hyndman, 1968).

ANOMALIES IN THE SOUTHERN HALF OF HOKKAIDO, JAPAN

Some of the foremost developments in GDS have their origin in the analysis of data collected in Japan as reviewed by Rikitake (1966). Here, we discuss recent work by

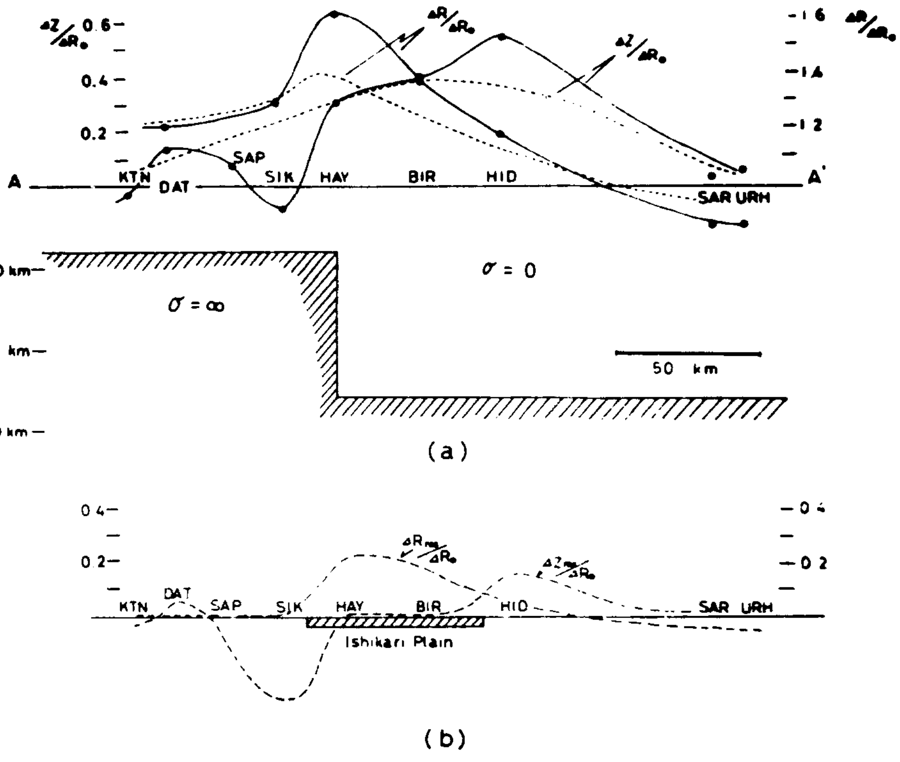

Fig. 2. (a) Observed $\Delta Z/\Delta R_0$ and $\Delta R/\Delta R_0$ for a period of 60 m (solid curves and circles) and that deduced from a step type super conductor (dashed curves). ($\Delta R_0 = \sqrt{(\Delta H^2 + \Delta D^2)}$). Results from a study in Hokkaido Japan (Nishida, 1976). (b) Differences between observed and calculated $\Delta Z/\Delta R_0$ and $\Delta R/\Delta R_0$.

Nishida (1976), of interest because of the model calculations. In the calculation the author has taken into account the scale length of the anomalies. The curves for $\Delta Z/\Delta R_o$ and $\Delta R/\Delta R_o$ ($\Delta R = \sqrt{\Delta H^2 + \Delta D^2}$ and $\Delta R_o = \Delta R$ at a normal station) were separated into two components of shorter and longer spatial wavelengths (Figure 2). The anomaly of shorter wavelength has its origin in current channelling in the Ishikari plain and the anomaly of longer wavelength is due to a sharp step in the surface of the upper mantle under the Ishikari plain. In the western region the depth of the conducting layers is considered to be 40 km, changing to 130 km in the eastern region. The layer is considered to lie in the mantle. The upper surface of the conductor on the western side is concluded to be the isotherm of about 1200°C whereas the temperature at the same depth on the eastern side is even below 500°C. Anomalously low P_n velocities (7.5 km/sec) have been found by explosion seismology under the western part of the Ishikari Plain. Most rocks except dry peridotite become wholly or partially melted at 1200°C at a pressure corresponding to the depth of 40 km. The increased conductivity on the western side may be due to the melting effect.

QUEBEC, CANADA

Honkura *et al.* (1977) from GDS studies northeast of Quebec city have shown that in areas where current channelling is important, the use of difference transfer functions is effective in reducing the scatter in computed transfer functions, due also partly to the geometry of the source field. They further show that in regions of high geomagnetic latitude, induction by the vertical component (Z) may also be important. It can be assumed that the effect of the source field will be approximately the same at stations separated by distances small compared to the source field. The difference transfer function, thus, is very useful in separating local anomalies from regional anomalies. The latter may be either of internal or external origin.

The authors also give a method for determining transfer functions when induction by all three components X, Y and Z is considered. They find the anomalous magnetic fields to arise both from surface currents and currents with different phases flowing at greater depths in the crust.

EUROPE

Europe has been extensively studied for geophysical features through GDS and MT. Some of the papers reviewed above have shown the presence of regional features, but these have used only local features in conductivity studies. The European data provides a good set for correlating regional features with geological structures. The regional components will indicate large-scale features. Analysis in this manner was started by Ádám *et al.* (1972) and has recently been completed in great detail by Pec *et al.* (1977). The authors have used 531 estimates of moduli and azimuths of Wiese-Parkinson vectors. The regional part was estimated through fitting a smooth surface $S(\theta, \lambda)$ in terms of spherical harmonics:

$$S(\theta, \lambda) = \sum_{n=0}^{N} \sum_{m=0}^{n} (g_n^m \cos m\lambda + h_n^m \sin m\lambda) \; P_n^m (\cos \theta)$$

where θ is the geographical colatitude and λ represents the geographical longitude. $P_n^m(\cos \theta)$ is the Legendre polynomial. The isoline contour map of the regional variation with terms up to $N = 1$, separates approximately the area affected by Alpine orogeny from the geologically older north-western Europe. The isolines from the fifth order approximation (terms up to $N = 5$) exhibit the North German-Polish conductivity anomaly merging with the Carpathian anomaly in the Western Ukraine and Roumania. The second-order approximation ($N = 2$) contours show an oval-like pattern in the western Ukraine close to the Czechoslovak-Hungarian-Roumanian border. The rough coincidence of this oval-like anomaly with the positive anomaly of geoid heights associates both anomalies with some uneven distribution of masses in the Earth's crust and upper mantle. Maps of apparent specific resistivities for Europe using the results of 250 MT sounding curves have been studied for regional anomalies by Porstendorfer and Gothe (1977).

CARPATHIANS

In the last few years, conductivity anomalies in the Carpathians have attracted considerable attention. We will discuss here some of the more recent results on this anomaly (Rokityanski *et al.*, 1975; Jankowski *et al.*, (1977a and b). On all the three traverses a reversal in the direction of Wiese–Parkinson vector was noticed close to the so called Piening Klippen belt marking the boundary between the Outer and Central Carpathians. The transfer functions on the two sides of the reversal zone show an asymmetry, connected with diverse surficial structures. The anomalies in geomagnetic variations are associated with an extended current concentration at crustal depths under the boundary between the Outer and Central Carpathians.

Jankowski *et al.* separated the internal and external parts of both horizontal and vertical components using Hilbert transforms. To study the frequency dependence of induced currents, Hilbert transforms were also obtained in narrow frequency bands. The depth of the internal current was calculated from the real part of a complex angle defined by the ratio of internal vertical and horizontal fields. The normals to these directions intersect at the common anomaly source (Figure 3). The results obtained by these authors exhibit the superiority of the method. The depth for various frequency ranges scatters between 22 and 45 km. The width of the anomalous zone is estimated to be 16–18 km centred at the reversal zone.

The induction vectors, transfer functions and field separation were used to model the geological structure along the profile No. V of their paper. The modelling technique was based on a finite difference approximation developed by Jones and Pascoe (1971). A model of a conducting body 16 km wide at depth between 22 and 45 km needs a conductivity of 0.02 Sm^{-1} to give a reasonable fit, though a better agreement is obtained when the conductor is considered to be present not only in the sediment but also in the rocks in the subduction zone of the crust.

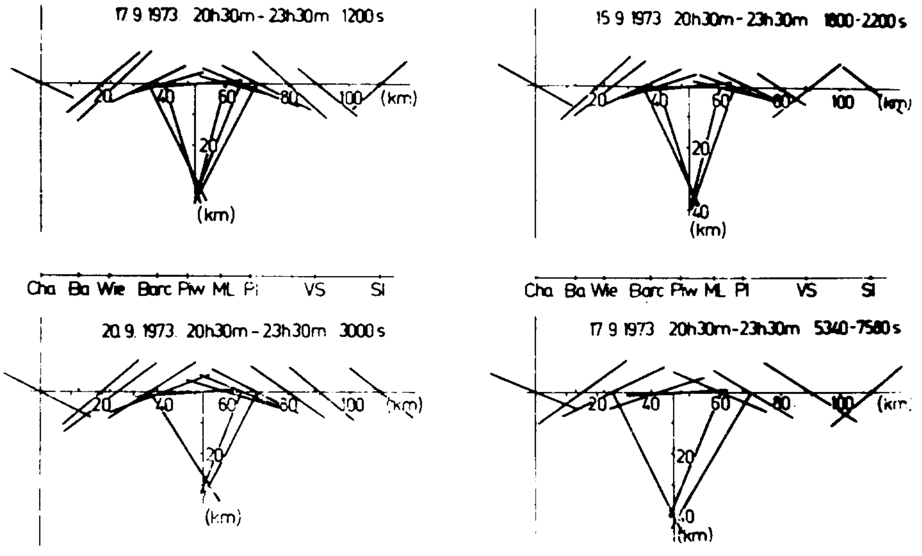

Fig. 3. Estimation of internal source depth from the real part of a complex angle defined by the ratio of internal vertical and horizontal fields. The results are for profile no. V in the Carpathians for various frequencies (From Jankowski *et al.*, 1977a).

Rokityanski *et al.* (1975) estimate the maximum depth of the upper edge of the anomalous body as 20 km in the western Carpathians and 40 km in the southern Carpathians. The difference could be due either to the depth of the conducting body increasing in the southern direction or to the extension of its width. They further suggest that the Carpathian anomaly connects larger conductors – the North German electric conductivity anomaly in the north and the Black Sea in the south. It is also suggested that the Carpathian electric conductivity anomaly is associated with metamorphic changes of the sedimentary rocks under the influence of temperature and pressure at depths of 10 to 25 km.

U.S.S.R.

Rokityanski *et al.* (1977) report results from the first magnetometer array study in U.S.S.R. covering the central part of the Russian platform. They detected an anomaly of complicated structure near Moscow. This is an interesting anomaly for studying the geological features of the Pre-Cambrian basement located at a relatively shallow depth.

KENYA

Discussions of results of recent geophysical studies in East Africa raise a controversy of whether or not the rifting in Kenya can be regarded as the first stage of continental break up. A recent review by Skinner (1977) describes in detail the various geophysical studies undertaken in the area. Results from GDS and MT studies have been reported by Banks and Ottey (1974), Beamish (1977) and Rooney and Hutton (1977). Beamish collected data using 22 magnetometers of the Gough–Reitzal type installed at sites in and around the Gregory Rift Valley in Kenya. Demodulated magnetograms were used to identify the anomalies, following the technique of Banks (1975). They showed an absence of anomalies in horizontal fields at the western stations, these were taken as normal stations. In the next phase of analysis single station transfer functions ($A(f)$ and $B(f)$) were computed and the maximum/minimum response functions of Banks and Ottey (1974) were calculated. For the period range 16–4 min, the in-phase parts of the maximum response functions (G_p) point towards the Rift axis, showing the presence of a shallow conductor in the graben. In the next step contours of Z-amplitude were drawn for hypothetical events. The phase maps of the hypothetical event analysis showed current concentrations quite clearly. From the extrema in the Z-variations, it was found that the current cannot be deeper than 30 km for the period band 8–4 min. For longer period (T = 128–16 min), the depth is estimated to be of the order of 50 km. In the south-east region a concentration of current was noticed, which caused a large gradient in the vertical field. These currents seem to have a crustal source.

Rooney and Hutton (1977) have studied the same area with both GDS and MT techniques. They point out the difficulty in MT interpretation from a single station near a conductivity contrast. However, if the boundary of the contrast can be defined through GDS, MT soundings across the anomalous structure at a series of sites can give reliable results (Reitmayr, 1975; Kurtz and Garland, 1976). Resistivities were calculated for two-dimensional models (Jones and Thomson, 1974) and for one-dimensional models (Schmucker 1970). The authors stress that not only should the calculated and observed resistivities be compared but a search should also be made to identify relationships between two-dimensional and one-dimensional curves. The latter provides information solely on the distribution of conductivity with depth which is the primary interest of study. The data could not resolve between the presence of a resistive zone 5 km thick near the surface or one 10–15 km thick at a depth below 5–10 km. A highly conductive zone (0.1 Sm^{-1}) is predicted to be present in the upper crust at a depth of 0–15 km. Another conductor is predicted at depths greater than 35 km. The high conductivity at upper crustal depths has been associated with high temperatures and water saturation of the crust under the rift valley. The possible existence of large quantities of water in the crust below the Gregory Rift and in the crust below Ethiopia and Iceland suggests a relationship between high water concentration, high temperature and tectonic activity.

SOUTHERN AFRICA

South of Kenya, de Beer *et al.* (1976) operated a 25 three-component magnetometer array in South-West Africa, Botswana and North-Western Rhodesia. Transfer functions showed the presence of a conductor that trends east-west in the western part of the array and curves north-eastward in the east (Figure 4). Four stations situated away from the conductor, were designated as normal stations and the mean of the X amplitudes from these was used to define the inducing field over the array. Z_a/X_n and X_a/X_n were then plotted across the conductor to estimate its depth; X_a/X_n gave 125 km and Z_a/X_n gave

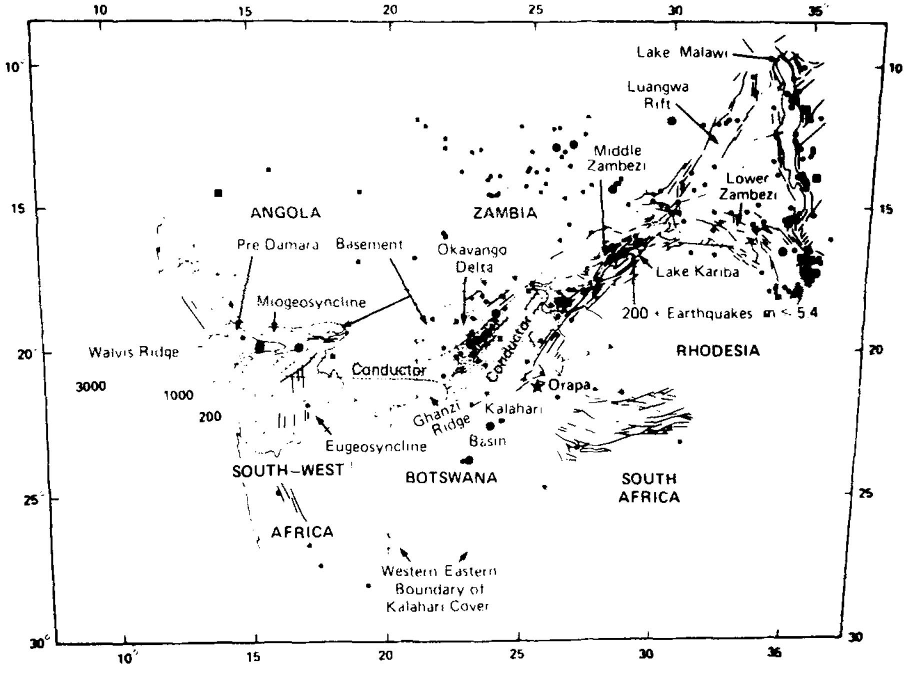

Fig. 4. The conductor in relation to seismicity and tectonics in southern central Africa (from de Beer *et al.*, 1975).

50 km. The authors favour the estimate of 50 km. The separation of the in-phase and quadrature induction vectors supports the location of the conductor at crustal depths. De Beer *et al.* (1975) using geological and geophysical data have concluded that this conductor marks an extension of the African Rift System along old weak zones in the lithosphere.

Pyrenees and North Morocco

Babour *et al.* (1976) and Vasseur *et al.* (1977) noticed a very large anomaly in Z and H in the period range from a few minutes to a few hours in the Northern Pyrenees. The anomaly corresponds to that of a current concentration. The geometry of the channelled current was found to remain invariant with time. Le Borgne and Le Mouel (1977) found a similar phenomenon in North Morocco. Babour (1977) further reports a strong channelling from the Rhinegraben. The time variations of the anomalous horizontal and the vertical component were found to be proportional from one station to another and from one component to another. The total anomalous field B_a (p, t) defined by the differences ΔH, ΔD, ΔZ (Figure 5) is linearly polarized in space and can be written as the product of spatial and time components, viz.

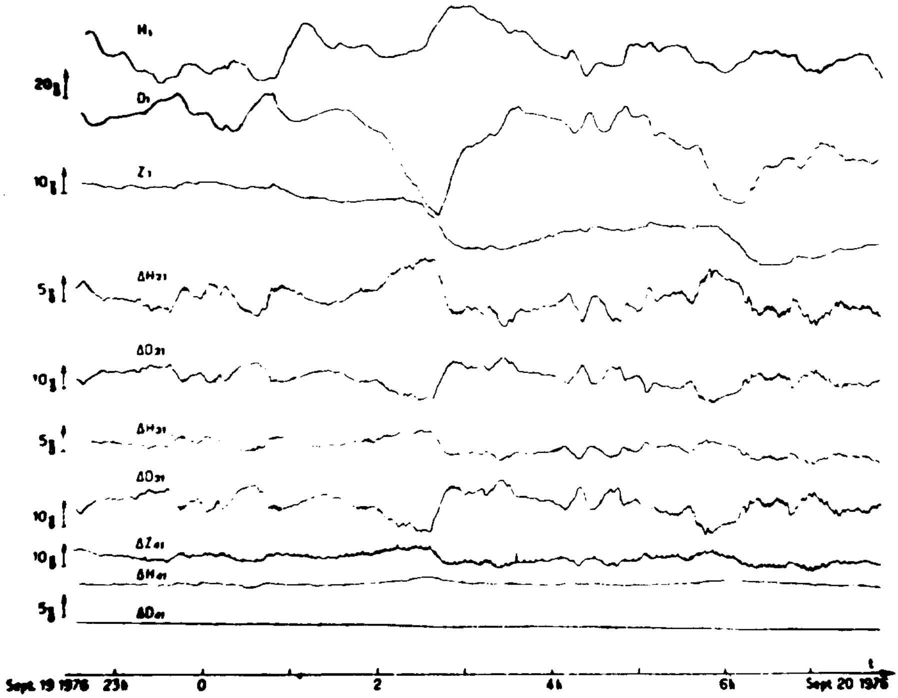

Fig. 5. (a) Variations of H, D and Z at the reference station in the Rhinegraben. (b) Simultaneous records of the components of the anomalous field at three stations in the anomaly (From Babour, 1977).

$$B_a(p, t) = b_a(p) R(t) .$$

Vasseur and Weidelt (1977) have developed a bimodal algorithm in terms of integral equations over the anomalous domain to solve problems of this class. Vasseur *et al.* (1977) used transfer functions to compute the structure of the area. They emphasize that the structure so calculated would give information on the area where currents are induced; in this case it would be the ocean. They deduce that there is a conducting layer at a depth of 50 km.

Babour (1977) finds (Figure 6) that the transfer function calculated from Z_a is compatible with the induction vector calculated from H_a, whereas the transfer function obtained by taking $Z_a = Z_{tot}$ gives a different direction. The result is of special relevance for those areas where the presence of channelled currents is very pronounced.

Eastern U.S.A.

New results on crustal electrical conductivity in the Eastern U.S.A. have been reported by Bailey and Edwards (1978). They calculated transfer functions and magnetotelluric resistivities at a number of sites in Pennsylvania, New York, New Jersey and Ontario (Canada). The period covered ranges from 2 seconds to 2 hours. Their results show that

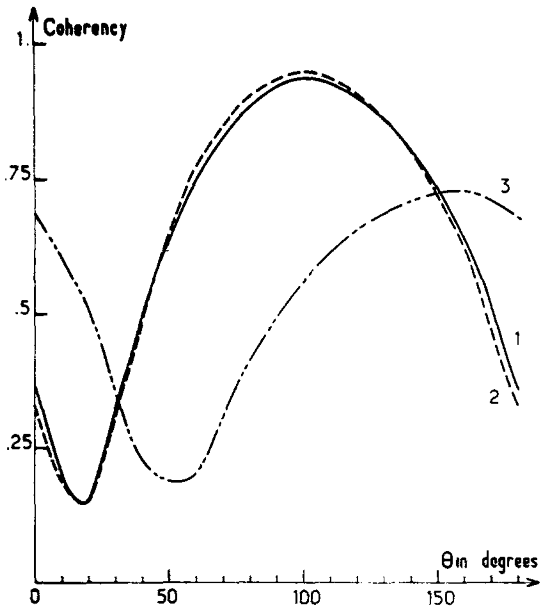

Fig. 6. Variation with azimuth of the coherency between the anomalous field (X_a) and the component of the horizontal field along the azimuth. Curves 1, 2, 3 correspond respectively to $X_a = \Delta H$, ΔZ, Z (From Babour, 1977).

geomagnetic features of the Appalachian system extend further west. The anomalies have been associated with conductivities in the crust.

INDIA

From the anomalies in *H*, *D* and *Z* recorded at the three permanent observatories near the southern tip of the Indian Peninsula, Rajaram *et al.* (1978) have shown the presence of a conducting channel in the crust or upper mantle between India and Sri Lanka Island. Their results also establish the presence of induced currents associated with the equatorial electrojet.

NORTH-WEST ATLANTIC OCEAN

Vanyan *et al.* (1978) analysed the resistivities (ρ_a) determined by Poehls and Van Herzen (1976) from the ratio of the horizontal component of the magnetic field measured at the surface (H_0) to the sea-floor value (H_d) in the northwest Atlantic. ρ_a values for the period range 1 h to 24 h were used in the analysis. The best model fitting both the amplitude and phase is found to be one having a lithosphere of thickness 90 km and an asthenosphere of conductance 1.5×10^4 S (100 km thick and 7 Ωm resistivity). This integrated value for the asthenosphere is greater than the conductance of the continent of (10^3 S) but about one third of the value found in the Pacific near California. The conductance of the sea bottom in the present work (2.5×10^3 S) is considerably higher. Such a high value can hardly be explained by the conductivity of sea-bottom sediments. Poehls and Van Herzen suggest that the crust on young sea floor may contain a substantial amount of sea-water, which will lower the resistivity. The present study further confirms the large difference in resistivity structure between continental lithosphere and oceanic lithosphere.

Conclusion

The technique of transfer functions though first introduced as an approximation by Schmucker (1964) to parameterize the induction effects seems still to be the most convenient quantity in modelling conductivity distributions. New studies reported in the last section show that if the normal component of the field (H_n, D_n and Z_n) can be accurately defined, the estimate of transfer functions becomes quite reliable. Estimation of H_n, D_n and Z_n is not a problem when an array of magnetometers is used. The HEA developed by Bailey *et al.* (1974) has made the transfer function analysis even more powerful. The contours of Z in the HEA make the method powerful for studying those regions where the naturally occurring source fields have small spatial wavelengths or for those areas where the conductor is curved or has a three-dimensional structure.

Jankowski *et al.* (1977 a and b) have shown that separation of fields into internal and external parts does not locate the current concentration accurately. My own experience with such separations is that the definition of Z_n is a very critical input in such computations. If array data contain anomalies of Z of scale lengths either smaller than the array spacing or larger that the array size, the results can be misleading. To estimate normal $Z(Z_n)$ the method developed recently by Schmucker (1978) can be highly useful.

Lilley and Sloane (1976) use the field gradient method to estimate electrical conductivity. Their method is especially suited for those areas which underlie non-uniform source fields. Their results are important since the conventional method of transfer functions is believed to be of rather limited use in regions of non-uniform source fields.

A major aim of all conductivity studies is to associate them with geothermal data. A clearer separation is necessary between electric anomalies in the crust and upper mantle, as the two arise from different physical and chemical processes. Ádám (1976b) has found that the average depth of conductive layers clusters around three different curves, when plotted against regional heat flow (Figure 7). One curve is for crustal conductors, which he calls FCL (= first conducting layer), another for upper mantle conductors called ICL (= intermediate conducting layer) and the third, again in the mantle, called UCL (= ultimate conducting layer). All three show decreasing depth with increasing heat flow. The three curves can be represented by:

$$h = h_0 q^{-a} \, ,$$

where h is the depth of the conducting layer, q is the heat flow, h_0 and a are constants.

Feldman (1976) supposes the conducting layer in the crust (FCL) to be located in the

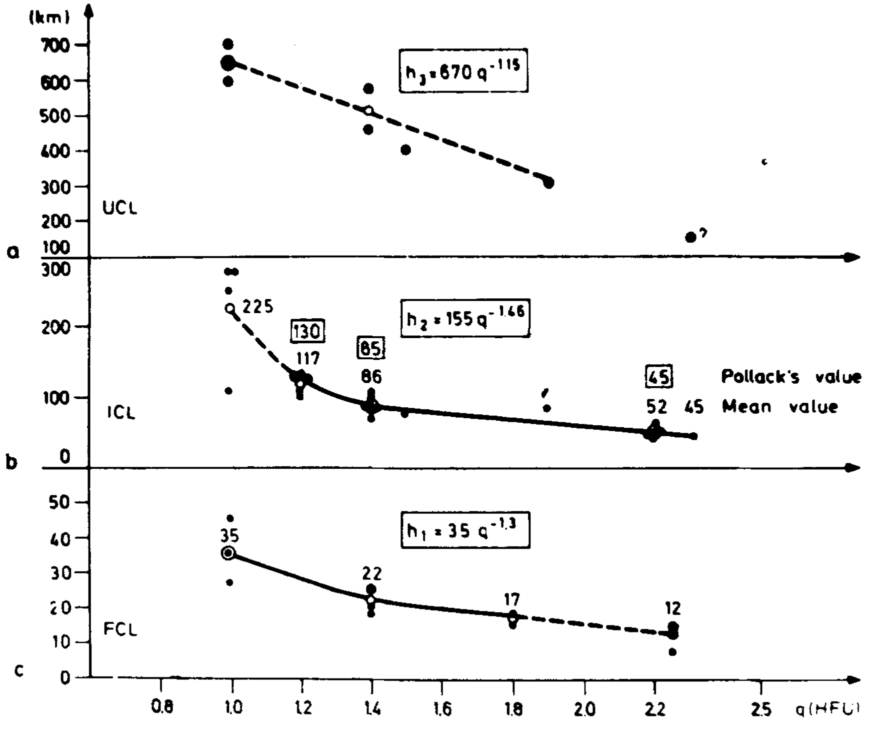

Fig. 7. Connection between regional heat flow and the depth of the conducting layers (From Ádám, 1976).

transition zone of the amphibolites and granulitic facia. Water plays an important part in these processes. This has also been discussed by Hyndman and Hyndman (1968). Ádám (1976b) suggests that the process of stabilization of the conductivity in ancient geologic formations can be explained by assuming that crustal anomalies are coupled with deep fractures. Examples of such anomalies are: the Ukranian crystalline shield at Kirovograd, the Baikal rift zone, the Rhinegraben and the Carpathians. The conductive layer in the upper mantle (ICL) is due to partial melting of basic rocks (basalts). The connection between the depth of the ICL and heat flow indicates the important role of the upper mantle in tectonic processes.

Acknowledgements

The author has greatly benefited from the correspondences with Professor U. Schmucker and Professor A. Ádám and wishes to thank them for clarifying many of his doubts. He also wishes to express his gratitude to Dr F.E.M. Lilley for critically reading the manuscript. Many useful discussions with Professor B.N. Bhargava and Mrs. N. Nityananda are gratefully acknowledged.

References

Ádám, A: 1976a, (Editor-in-Chief), *Geoelectric and Geothermal Studies (East Central Europe and Soviet Asia)*, KAPG Geophysical Monograph, Adakémiai Kiadó, Budapest.

Ádám, A: 1976b, *Acta Geodaet. Geophys. et Montanist. Acad, Sci. Hung.* **11** (3-4), 503-509.

Ádám, A., Verő, J., and Wallner, Á.: 1972, *Acta Geodaet. Geophys. et Montanist Acad. Sci. Hung.* **7** (3-4), 251-287.

Agarwal, A.K., Singh, B.P., and Nityananda, N.: 1979, *Phys. Earth Planet. Inter.* **18**, 232-237.

Babour, K.: 1977, *Geophys. J.R. Astr. Soc.* **31**, 265-269.

Babour, K., Mosnier, J., Diagniere, M., Vasseur, G., Le Mouel, J.L., and Rossignol, J.C.: 1976, *Geophys. J.R. Astr. Soc.* **45**, 583-600.

Babour, K., and Mosnier, J.: 1977, *Geophys.* **42**, 66-76.

Bailey, R.C., Edwards, R.N., Garland, G.D., Kurtz, R., and Pitcher, D.: 1974, *J. Geomag. Geoelec.* **26**, 125-146.

Bailey, R.C., and Edwards, R.N.: 1978, Paper presented in Canadian Geophysical Union Meeting on May 15-17.

Banks, R.J.: 1973., *Phys. Earth Planet. Inter.* **7**, 339-348.

Banks, R.J.: 1975, *Geophys. J.R. Astr. Soc.* **43**, 87-101.

Banks, R.J., and Ottey, P.: 1974, *Geophys. J.R. Astr. Soc.* **36**, 321-335.

Beamish, D.: 1977, *Geophys. J.R. Astr. Soc.* **50**, 311-332.

Camfield, P.A.: 1973, 'Studies with a two-dimensional magnetometer array in the western United States and south-western Canada', Ph.D. thesis, University of Alberta, Edmonton, Canada.

Caner, B.: 1971, *J. Geophys. Res.* **76**, 7202-7216.

de Beer, J.H., Gough, D.I., and van Zijl J.S.V.: 1975, *Nature* **225**, 678-680.

de Beer, J.H., van Zijl, J.S.V., Huyssen, R.M.J., Hugo, P.L.V., Joubert, S.J., and Meyer, R.: 1976, *Geophys. J.R. Astr. Soc.* **45**, 1-17.

Edwards, R.N., Law, L.K., and White, A.: 1971, *Phil. Trans. Roy. Soc.* **270A**, 289-323.

Everett, J.E. and Hyndman, R.D.: 1967, *Phys. Earth Planet. Int.* **1**, 24-34.

Feldman, I.S.: 1976, 'On the nature of conductive layers in the earth's crust and upper mantle'. In A. Ádám (ed.), *Geoelectric and Geothermal Studies*, pp. 721-730, KAPG Geophysical Monograph, Akadémiai Kiadó, Budapest.

Fraser, M.C.: 1974, *Rev. Geophys. Space Phys.* **12**, 401-420.

Gough, D.I.: 1973, *Geophys. J.R. Astr. Soc.* **35**, 83-98.

Gough, D.I. and Reitzel, J.S.: 1967, *J. Geomag. Geoelec.* **19**, 203-215.

Hermance, J.F., and Grillot, L.R.: 1970, *J. Geophys. Res.* **75**, 6582-6591.

Honkura, Y., Kurtz, R.D., and Niblett, E.R.: 1977, *Can. J. Earth Sci.* **14**, 256-267.

Hutton, V.R.S.: 1976, *Rep. Prog. Phys.* **39**, 487-572.

Hutton, V.R.S., Sik, J.M., and Gough, D.I.: 1977, *Nature* **266**, 617-620.

Hyndman, R.D., and Hyndman, D.W.: 1968, *Earth Planet. Sci. Lett.* **4**, 427-432.

Jankowski, J., Szymanski, A., Pěč, K., Pěčová, J., Petr, V., and Praus, O.: 1977a, *Acta Geodaet. Geophys. et Montanist. Acad. Sci. Hung.* **12** (1-3), 99-109.

Jankowski, J., Szymanski, A., Pěč, K., Červ, V., Petr, V., Pěčová, J., and Praus, O.: 1977b, *Studia Geophys. et Geod.* **21**, 35-57.

Jones, F.W., and Pascoe, L.J.: 1971, *Geophys. J.R. Astr. Soc.* **24**, 3-30.

Jones, F.W., and Thomson, D.J.: 1974, *Geophys. J.R. Astr. Soc.* **37**, 537-544.

Kurtz, R.D., and Garland, G.D.: 1976, *Geophys. J.R. Astr. Soc.* **45**, 321-347.

Larsen, J.C.: 1975, *Geophys. J.R. Astr. Soc.* **43**, 17-46.

Le Borgne, E., and Le Mouel, J.L.: 1977, *Ann. Geophys.* **33**, 237-242.

Leary, P., and Phinney, R.A.: 1974, *Geophys. Res. Lett.* **1**, 265-268.

Lilley, F.E.M., and Sloane, M.N.: 1976, *J. Geomag. Geoelec.* **28**, 321-328.

Nishida, Y.: 1976, *J. Geomag. Geoelec.* **28**, 375-394.

Parkinson, W.D.: 1959, *Geophys. J.R. Astr. Soc.* **2**, 1-14.

Pěč, K., Pěčová, J., and Praus, O.: 1977, *Studia geoph. et geod.* **21**, 383-393.

Poehls, K.A., and Von Herzen, R.P.: 1976, *Geophys. J.R. Astr. Soc.* **47**, 331-346.

Porath. H.: 1971, *J. Geophys. Res.* **76**, 2643-2648.

Porath, H, and Gough, D.I.: 1971, *Geophys. J.R. Astr. Soc.* **22**, 261-275.

Porstendorfer, G., and Göthe, W.: 1977, *Acta Geodaet. Geophys. et Montanist, Acad. Sci. Hung.* **12** (1-3) 117-122.

Price, A.T.: 1967, 'Electromagnetic induction within the earth'. In S. Matsushita and W.H. Campbell (ed.), *Physics of Geomagnetic Phenomena,* Vol. 1, pp. 235-298, Academic Press, New York and London.

Rajaram, M., Singh, B.P., Nityananda, N., and Agarwal, A.K.: 1979, *Geophys. J.R. Astr. Soc.* **56**, 127-138.

Reitmayr, G.: 1975, *J. Geophys.* **41**, 651-658.

Rikitake, T.: 1966, *Electromagnetism and the Earth's Interior,* Elsevier, New York.

Rokityanski, I.I., Kulik, S.N., Logvinov, L.M., and Schuman, V.N.: 1975, *Acta Geodaet. Geophys. et Montanist, Acad. Sci. Hung.* **10**, 277-286.

Rokityanski, I.I., Logvinov, I.M., Rokityanskaya, D.A., and Shuman, M.T.: 1977, *Acta Geodaet. Geophys. et Montanist. Acad. Sci. Hung.* **12**, (1-3), 131-137.

Rooney, D., and Hutton, V.R.S.: 1977, *Geophys. J.R. Astr. Soc.* **51**, 91-119.

Schmucker, U.: 1964, *J. Geomag. Geoelec.* **15**, 193-221.

Schmucker, U.: 1970, *Bull. Scripps Inst. Oceanogr.* **13**, 1-65.

Schmucker, U.: 1978, Personal communication.

Shankland, T., and Waff, H.S.: 1977, *J. Geophys. Res.* **82**, 5409-5417.

Skinner, N.J.: 1977, *Contemp. Phys.* **18**, 455-470.

Vanyan, L.T., Kharin, E.P., and Shilovski, P.E.: 1978, *Phys. Earth Planet. Int.* **16**, 7-9.

Vasseur, G., Babour, K., Menvielle, M., and Rossignol, J.C.: 1977, *Geophys. J.R. Astr. Soc.* **51**, 669-690.

Vasseur, G., and Weidelt, P.: 1977, *Geophys. J.R. Astr. Soc.* **51**, 669-690.

Wiese, H.: 1962, *Geofis. Pura.* **52**, 83-103.

Winter, R.: 1973, 'Der Oberrheingraben ad anomalie der elektrischen Leitfahigkeit, Untersucht mit Methoden der erdmagnetischem Tiefensondierung', Ph.D. dissertation, Univ. Gottingen, West Germany.