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Magnetotelluric Investigation of the Lower Crust and Upper Mantle Beneath Iceland

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Abstract. Magnetotelluric measurements were made at 12 sites on a 210 km long profile in northern and eastern Iceland. The profile crosses areas of different geological age ranging from Tertiary flood basalt to the presently active zone of rifting and volcanism.

Beneath the investigated area, at a depth of a few km down to about 20 km there exists a 5 km thick layer with low resistivity amounting to 15 Ωm which is imbedded between layers of higher resistivities. The depth to the low resistivity layer increases with increasing distance from the spreading axis.

The low resistivity layer is presumably caused by partial melting at the base of the crust and at the top of the anomalous mantle beneath Iceland. Comparison with laboratory measurements confirms a basaltic composition and a temperature of 1,000–1,100° C of the good conductor, and probably partially molten peridotite in the upper mantle beneath.

Mean temperature gradients in the crust calculated from the magnetotelluric data are in good agreement to surface temperature gradients measured in drill holes.

Key words: Magnetotellurics – Iceland – Electrical structure – Low resistivity layer – Temperature gradient – Partial melting.

1. Introduction

During the summer 1977 a magnetotelluric (MT) field program was performed in north-eastern Iceland.

The purpose of this experiment was to investigate vertical and lateral variations of the electrical resistivity in the lower crust and upper mantle beneath Iceland.

Among the physical properties of the earth's interior, which can be measured on the surface, the electrical resistivity is the parameter most sensitive to

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porosity and fluid content in the upper crust. It also correlates well with temperature and chemical composition in the lower crust and mantle. Hence, in measuring in some detail the resistivity distribution, it is possible to investigate crustal thickness and crustal structure. Magnetotelluric measurements could further give valuable informations on the temperature, composition and dimension of the mantle beneath Iceland. Therefore some of the questions concerning processes on the oceanic ridge system like plate accretion may be answered.

2. Geological Setting

Iceland is one of the few land areas, and certainly the largest one, situated on the Mid-Atlantic-Ridge system. The Mid-Atlantic Ridge, which is considered an

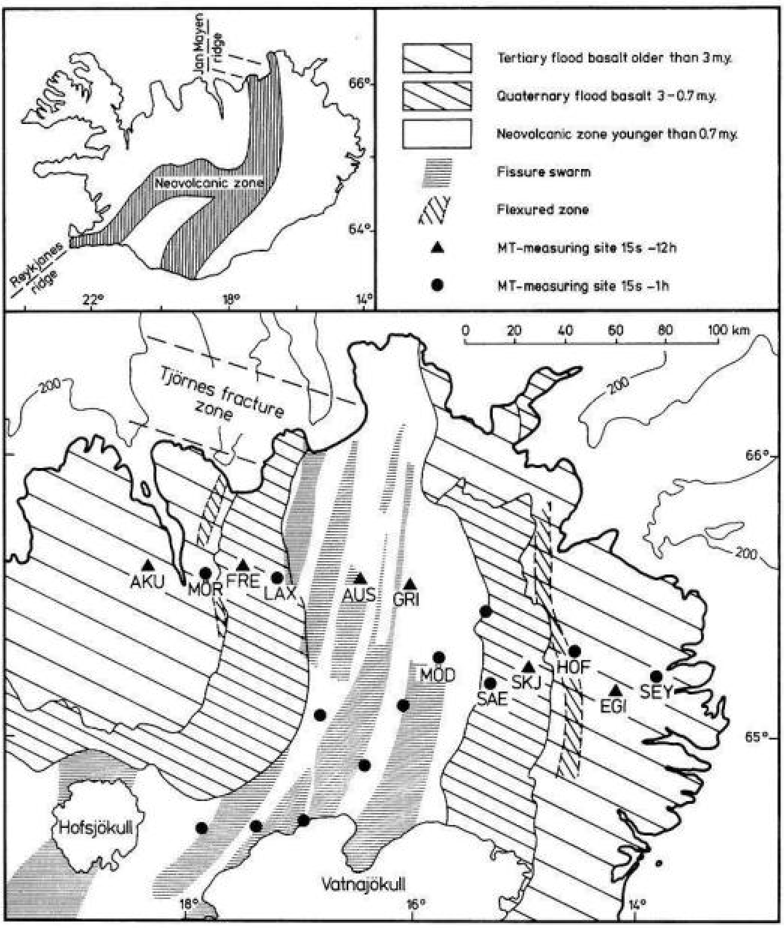

Fig. 1. A simplified geological map of north-eastern Iceland as redrawn from Saemundsson (1974). Triangles are locations of magnetotelluric base sites (period range 15 s–12 h). Filled circles show short period (15 s–1 h) magnetotelluric sites. Only the stations provided with initial letters of site-names are discussed in the present paper

accreting plate boundary of the European and American plates crosses Iceland from south-west to north-east. The surface manifestation of the ridge on land is a tectonically active zone of recent rifting and volcanism, the so called neovolcanic zone. On both sides of the neovolcanic zone there are strips of Quaternary flood basalts followed by Tertiary flood basalts in western and eastern Iceland (see Fig. 1). K/Ar age determinations indicate increasing ages with increasing distance from the active zone (Pálmason and Saemundsson, 1974).

The main features of the geology of north-eastern Iceland have been mapped by Saemundsson (1974). According to his interpretation the neovolcanic zone is characterized by large swarms of north-south trending faults and fissures. The individual swarms are only 5–10 km wide, but several tens of kilometers long. Most of them pass through central volcanos where volcanic activity and high-temperature geothermal fields are concentrated (see Fig. 1).

To the north the neovolcanic zone runs into a complex zone of fractures, the so called Tjörnes fracture zone, which has an east-west direction and joins the ridge system north of Iceland.

Rifting and volcanic activity within the neovolcanic zone in Iceland is mostly confined to the fissure swarms and the associated central volcanos. Major rifting activity is episodic rather than continuous occuring in episodes of a few years with intervals of 100–150 years of quiescence (Björnsson et al., 1977).

3. Previous Geophysical Research in Iceland

A seismic refraction survey of the Icelandic crust was performed by Pálmason (1971). According to his interpretation the upper crust of Iceland consists of a surface layer and two layers of Tertiary flood basalt. These layers are underlain everywhere in the island by a layer with a mean P-wave velocity of 6.5 km/s (called layer 3) which is considered to correspond to the oceanic layer 3. This layer is nowhere exposed at the surface and the depth to its upper boundary is usually 3–5 km. In the upper mantle underneath layer 3 there is a layer with a P-wave velocity of 7.2 km/s (layer 4). The depth to layer 4 is 8–10 km in the south-west and around 14 km in south-eastern Iceland. No values are available for northern and north-eastern Iceland. If layer 4 is interpreted as the upper mantle it has an anomalously low velocity. This has been explained by a small amount of partial melting. The temperature in this layer should therefore be around 1,000° C, which is in good agreement with values obtained from linear extrapolation of temperature gradients in drill holes which range from 40° C/km in the eastern flood basalt zone to 160° C/km in south-western Iceland (Pálmason, 1973).

The measurements of the time-varying magnetic field were initiated in Iceland by Garland and Ward (1965) and by Hermance and Garland (1968a). They found low resistivities of the order of 10 to 20 Ωm at several sites below a depth of 30 km, which they suggested might be caused by partial melting. They further assumed that the low resitivity layer underlays the entire island rather than just the neovolcanic zone. Magnetotelluric measurements made by Hermance and Garland (1968b) in the northern part of the neovolcanic zone gave similar results for that area.

In further work Hermance and Grillot (1970), Hermance et al. (1972) and Hermance (1973) performed magnetotelluric measurements in order to investigate the electrical structure both of the lower crust and the upper mantle. All these measurements were confined to the neovolcanic zone with special emphasis to its south-western part. Between approximately 2 and 15 km depth they found a conductive layer with a resistivity between 10 and 20 Ωm. Below 10–15 km, at the top of the mantle, the resistivity increases to 40–100 Ωm and shows little variation down to at least 100 km. Hermance (1973) believes that ground water has a considerable effect on the resistivity at depths down to 8–10 km, but below 10 km electronic conduction becomes dominant. Correlating the MT-results with the conductivity of basalt (lower crust) and peridotite (upper mantle) at high temperatures Hermance and Grillot (1974) estimated the temperature at the crust-mantle boundary to be 800–1,200° C. They further estimated the temperature gradient to be about 1° C/km between 15 and 100 km depth.

In further magnetotelluric and geoelectric works in Iceland Hermance et al. (1975), Thayer (1975), and Björnsson (1976) have extended the previous work from within the neovolcanic zone to the older Quaternary and Tertiary flood basalt areas. The aim was to investigate lateral differences in the electrical resistivity of the crust and upper mantle and also to obtain more detailed information on the vertical resistivity distribution, especially around thermal areas. On a regional scale the most significant result of this research is that the resistivity in the upper part of layer 3 at 3–5 km depth is much lower (10–20 Ωm) within the neovolcanic zone than beneath the Quaternary and Tertiary areas (300–1,000 Ωm).

The resistivity below a few tens of km in the mantle beneath Iceland seems to be uniform as reported in previous work. Among these findings there are also indications of differences in the resistivity of the lower crust inside and outside the neovolcanic zone but no detailed information was achieved on that matter.

4. Field Program, Measuring Sites

There were several reasons for performing the magnetotelluric experiment in northern and eastern Iceland:

(1) We intended to determine the temperature in the crust and upper mantle, especially its variations across and along the neovolcanic zone. This can be achieved by the magnetotelluric method since there exists a correlation between temperature and resistivity.

(2) In the part of the country under consideration there is only one neovolcanic zone (running north-south) and the older geological formations seem to be similar and symmetric on both sides of this zone. This means that one could expect a two-dimensional distribution of the electrical conductivity under the area which makes the analysis of MT-measurements on an east-west profile perpendicular to the geological strike much more simple than in the three-dimensional case.

(3) Further more, we planned to investigate possible resistivity changes along the neovolcanic zone from north to south in this area, caused by a proposed mantle plume under the zone in central Iceland (Saedmundsson, 1974).

(4) Long period MT-measurements have never before been made simultaneously at different sites in that part of the country.

(5) Some of the MT-sites were identical with measuring sites of an extensive seismic project (Reykjanes-Ridge-Iceland Seismic Project (RRISP)). The seismic measurements have been carried out in cooperation of Icelandic, German and Soviet geophysical institutions in July 1977. A joint interpretation of the experimental results probably will reduce the number of possible models of the Icelandic crust and upper mantle.

(6) In order to be able to quantitatively consider the influence of the non-uniform inducing magnetic field during night time the profile was orientated approximately perpendicular to the main direction of flow of the polar electrojet.

The major part of the magnetotelluric experiment was done along a 210 km long east-west profile of 12 MT-stations ranging from the Tertiary flood basalt area in the west across the neovolcanic zone to the Tertiary flood basalt area in eastern Iceland. At 6 base-sites on this profile the natural variations of the magnetic and electric field were recorded simultaneously and permanently for 8 weeks (triangles in Fig. 1). The instruments we used were Askania magnetographs for recording the time-variations of the magnetic field, and electrographs (Beblo, 1972) for recording the electric field. The observed period range was 300 s to 12 h.

At all the base-sites, at 6 additional sites on the east-west profile, and at 7 sites on a north-south profile along the neovolcanic zone, MT-measurements were made in the period range 15 s–1 h (filled dots in Fig. 1). A mobile MT-equipment was used consisting of electrograph, fluxgate magnetometer and tape recorder.

The measuring sites, especially the base-sites were installed in geologically homogeneous areas, in order to avoid local distortions of the electric field caused by inhomogeneities of surface resistivities (Beblo, 1974).

In the present paper first results of the magnetotelluric short period measurements (15 s–1 h) on the east-west profile are being presented.

5. The Magnetotelluric Method, Principles of Data Analysis

The electrical resistivity and its distribution within the earth's interior can be determined by observing time-varying magnetic ($\bar{B}$)- and electric ($\bar{E}$) fields at the surface. The depth penetration of the electromagnetic field into the earth depends on the resistivity distribution and the period (T) of the variation. According to the skin-effect, variations of shorter periods ($T < 1{,}000$ s) penetrate only into the uppermost layers (crustal depth). Longer period variations yield informations on the resistivity at greater depth (upper mantle).

5.1. The Impedance Tensor

For the determination of the resistivity-depth-distribution the observed time functions $E(t)$ and $B(t)$ are Fourier-transformed into functions of frequency. The horizontal and orthogonal components of the electric (E_i, E_j)- and magnetic (B_i, B_j) fields, which are measured at the surface, are combined by a complex transfer function, the impendance tensor $z=(z_{ij})$. The elements of z depend on the resistivity distribution and the orientation of the measuring coordinate system.

$$\begin{pmatrix} E_1 \\ E_2 \end{pmatrix} = \begin{pmatrix} z_{11} & z_{12} \\ z_{21} & z_{22} \end{pmatrix} \cdot \begin{pmatrix} B_1 \\ B_2 \end{pmatrix}$$

SI-units: $\bar{E}$ in V/m, $\bar{B}$ in nT, T in s.
Two parameters, the apparent resistivity $\rho_a(T)$ and the phase difference between the electric and magnetic field $\varphi(T)$ are calculated as functions of period from the impedance tensor.

Interpretation of MT-data in the general three-dimensional case is mathematically complicated and not practicable at the present stage.

In a two-dimensional resistivity distribution the diagonal elements of the impedance tensor disappear if the measured coordinates are parallel or perpendicular to the strike direction of the resistivity anomaly, and there remain the two elements

$$z_{\parallel} = \frac{E_{\parallel}}{B_{\perp}}$$

$$z_{\perp} = \frac{E_{\perp}}{B_{\parallel}}.$$

From these complex functions two different values are obtained for the apparent resistivity ρ_a and two for the phase difference φ:

$$\left.\begin{aligned} \rho_{a_{\parallel}} &= \frac{\mu_0 T}{2\pi} |z_{\parallel}|^2 \\ \varphi_{\parallel} &= \arg(z_{\parallel}) \end{aligned}\right\} \quad E\text{-polarization}$$

$$\left.\begin{aligned} \rho_{a_{\perp}} &= \frac{\mu_0 T}{2\pi} |z_{\perp}|^2 \\ \varphi_{\perp} &= -\arg(z_{\perp}) \end{aligned}\right\} \quad B\text{-polarization}$$

ρ in Ωm, φ in degrees.
In the case of an one-dimensional resistivity distribution, i.e., if the resistivity varies only with depth, the two values for the apparent resistivity and the two values for the phase become identical.

As Iceland is regularly situated below the polar electrojet the inducing magnetic field may be inhomogeneous. In that case ρ_a and φ depend both on the period T as well as on the spatial dimension, often called wavelength L, of the inducing magnetic field ($\rho_a(L, T)$, $\varphi(L, T)$).

Haak (1978) has made some model calculations to estimate the influence of inhomogeneous source fields on MT-measurements. He found that deviation from the homogeneous case can be neglected for a penetration depth shallower than the wavelength L. The half-width of the polar electrojet was estimated by Hermance and Garland (1968a) to be about 480 km. This means that the inhomogeneity of the source field in Iceland can be neglected for a penetration depth of less than approximately 100 km. Our present results are calculated from variations with periods shorter than 1 h. Such variations penetrate up to 100 km into the earth. Therefore we neglect effects from inhomogeneous inducing magnetic fields.

Near strong lateral changes in surface resistivity the electromagnetic field is changed considerably. This is for example the case near the ocean shore (ocean effect). It is well known, that for horizontal distances exceeding one or two skin-depths from the boundary the ocean effect can be neglected. Referred to the 200 m depth contour all our MT-stations are further inland by more than this critical distance which justifies our neglection of the ocean effect. Analysing MT-data from south-western Iceland Thayer (1975) arrived at a similar conclusion.

5.2. Polarization and Preference Direction of the Electric Field

The endpoints of the horizontal field-vectors $\bar{E}$ and $\bar{B}$ approximately move along an ellipse during one oscillation period T. The direction of the major axis of the ellipse is called the polarization direction, and the ratio of the major axis to the minor defines the intensity of polarization.

Direction and intensity of polarization of the magnetic source field usually change with time as well as with the period. Direction and intensity of polarization of the induced electric field depend on the magnetic source field but they are also a function of lateral variations of the electric resistivity in the underground near to the measuring site.

If the magnetic source field is randomly polarized and if there exists a preference direction of the polarization of the electric field, the preference direction depends mainly on the resistivity distribution near to the measuring site. If the resistivity distribution in the underground is two-dimensional it is possible to determine the preference direction of the electric field directly from the impedance tensor. For a constant magnetic field polarization direction α_B the tensor-element z_{21} shows a maximum for a certain electric field polarization direction α_E. The angle α_{E_0} where this tensor-element and therefore the electric field direction depends least on the direction of the magnetic field is identical to the preference direction (Kiessling, 1970, published by Haak (1978)). Determining the angle α_{E_0} of the preference direction simultaneously the absolute values of the diagonal elements of the impedance tensor were calculated. The ratio of these elements $|z_{21}/z_{12}|$ is a measure of the inhomogeneity of the electrical resistivity at the corresponding measuring site.

5.3. Model-Calculations

It is well known (Berdichevsky and Dmitriev, 1976) that the B-polarization depends much more on near-surface lateral resistivity variations than on resistivity anomalies at greater depth. In constrast to that case, the E-polarization ($\rho_{a_{\parallel}}$ and $\varphi_{\parallel}$) is far less affected by local near-surface anomalies and contains more information on the resistivity distribution at greater depth.

Because of the continuity of the electric field component parallel to a lateral resistivity contrast, the $\rho_{a_{\parallel}}$-values for the E-polarization are similar to the ρ_a-values for a plane layered earth. The difference between the E-polarization for a two-dimensional resistivity distribution and a one-dimensional case is small if the lateral resistivity gradient is small. From Fig. 2 it is evident that this is the case on our MT-profile in north Iceland. Therefore we have used the $\rho_{a_{\parallel}}$- and $\varphi_{\parallel}$-values of the E-polarization (not the B-polarization) to calculate one-dimensional models of the resistivity distribution with depth at each of the MT-stations.

6. The Magnetotelluric Results

6.1. Preference Direction of the Electric Field

At all the MT-sites along the east-west profile we have calculated the preference direction of the electrical field (α_{E_0}) using the method of Kiessling (1970). An investigation of the inducing magnetic field showed that no dominating preference direction of its polarization exists in the observed period range.

Figure 2 shows the distribution of the preference direction (α_{E_0}) along the profile for the period range of 15 to 300 s. The length of the lines is proportional to the ratio of the tensor elements $|z_{21}/z_{12}|-1$ and is a measure of the intensity of polarization (The lines should be reduced to points for a plane layered or homogeneous earth).

Within the Quaternary flood basalt zone and the neovolcanic zone the short lines indicate small horizontal variations of the resistivity. Strong horizontal variations of resistivity exist at the measuring sites within the Tertiary flood basalt zones. This means that the surface resistivity within the Tertiary flood basalts must be an order of magnitude greater than within the Quaternary and neovolcanic zones. This result is in good agreement with some resistivity (dipole-dipole) measurements made in the area, which show resistivities of the order of hundreds of Ωm at 1–4 km depth in the Tertiary zones, but resistivities well below 100 Ωm in the Quaternary and neovolcanic zones.

The preference direction within the Tertiary zones may be caused by the resistivity distribution of the underground but also by the topography near to the measuring sites.

6.2. Apparent Resistivities and Phase Differences

The impedance tensor has been calculated for all 12 sites on the EW-profil as a function of the angle of rotation. Herefrom we may infer that within the neovolcanic zone and the Quaternary flood basalt zone the tensor has the shape as for a one-dimensional resistivity distribution, whereas within the Tertiary

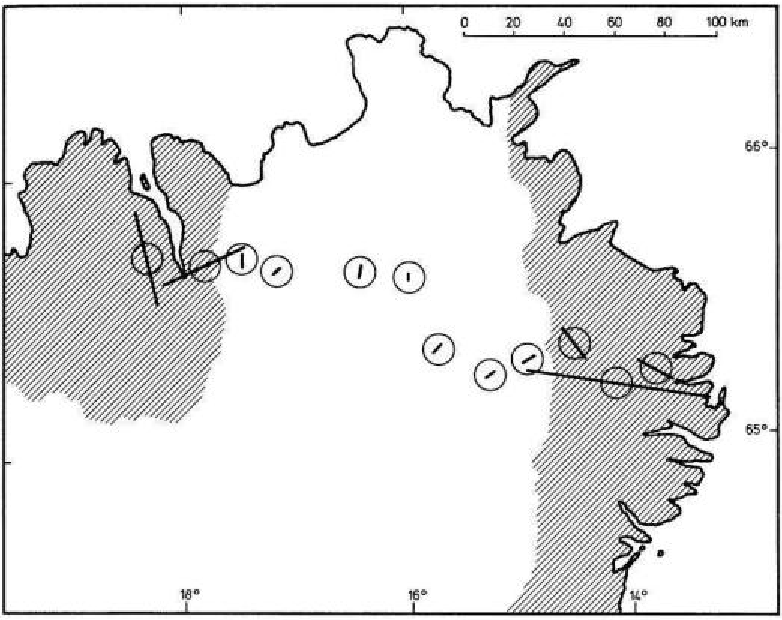

Fig. 2. Preference direction (lines) at each magnetotelluric site of the induced electrical field for the period range 15–300 s. The length of the lines indicates the intensity of polarization and is a measure of the inhomogeneity of the surface resistivity distribution at the sites

flood basalts we see a clear influence from lateral resistivity variations on the impedance tensor. Also from the distribution of the preference direction we may infer an approximately two-dimensional structure of the resistivity, coinciding obviously with the main geological features. Therefore we can discern the fundamental cases of *E*- and *B*-polarization.

Figure 3 shows the calculated values for the apparent resistivities and the phase differences for all the 12 sites, separately for *E*- and *B*-polarization. The phase differences are shown for the *E*-polarization only. The *B*-polarization (that means the component of the electric field perpendicular to the strike direction of the neovolcanic zone) demonstrates clearly strong horizontal resistivity variations. The apparent resistivity changes abruptly at the boundary of the Tertiary zones. The lower resistivity in the center regions can be seen as a trough bordered by higher values.

The *E*-polarization shows the component parallel to the boundary of the neovolcanic zone. The apparent resistivities for this case change continuously at the boundary.

The $\rho_{a_{\parallel}}$-values at the site Hofteigur (HOF) are elevated by a factor 2 or 3 compared to the sites on both sides. It is known (Berktold et al., 1976) that the ρ_a-values may be shifted by local near-surface resistivity anomalies to higher or lower values. The measuring site HOF is situated directly in a highly flexured zone near the border of the Tertiary and Quaternary zones (see Fig. 1) which might cause such an effect. A similar but not as significant shift is observed at

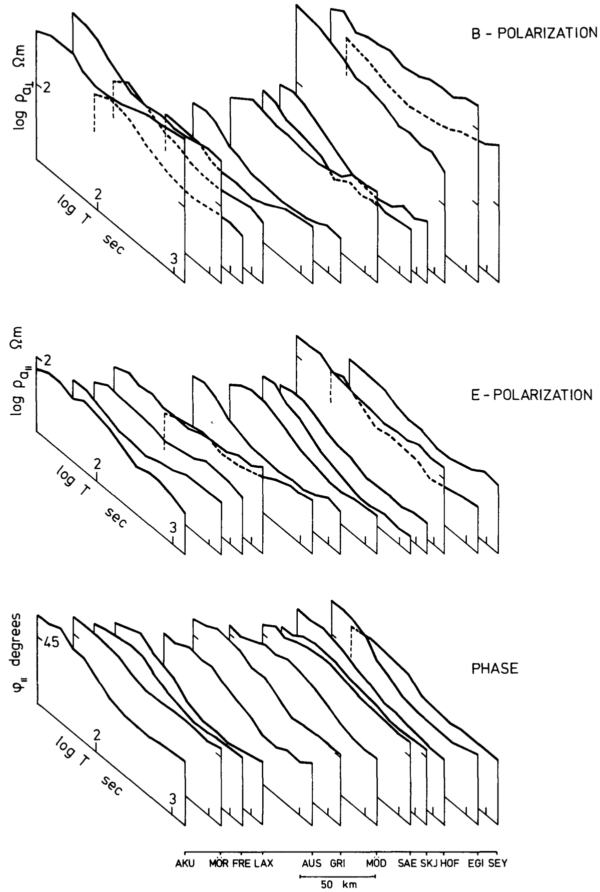

Fig. 3. Apparent resistivities for E-polarization $\rho_{a_\parallel}$, B-polarization $\rho_{a_\perp}$, and phase difference for E-polarization $\varphi_\parallel$ along the whole east-west MT-profile. The values for $\rho_{a_\perp}$ are deplated in the center part of the profile whereas the values for $\rho_{a_\parallel}$ and $\varphi_\parallel$ show similar behaviour across the profile

the sites Laxardalsheidi (LAX) and Saenautavatn (SAE), with ρ_a-values too high, or too low respectively, compared to adjacent sites.

In order to discern the fundamental cases of E- and B-polarization, we have additionally to consider the continuous or discontinuous behaviour of the apparent resistivities as a major discriminating factor. In AKU these directions seem to be exchanged, possibly caused by local topographic influence.

6.3. *Model-Calculations*

Because of the weak horizontal variations of the resistivity distribution within the Quaternary and neovolcanic zones we conclude that it is permitted to calculate one-dimensional models (plane layered earth) for the resistivity distribution at the MT-sites.

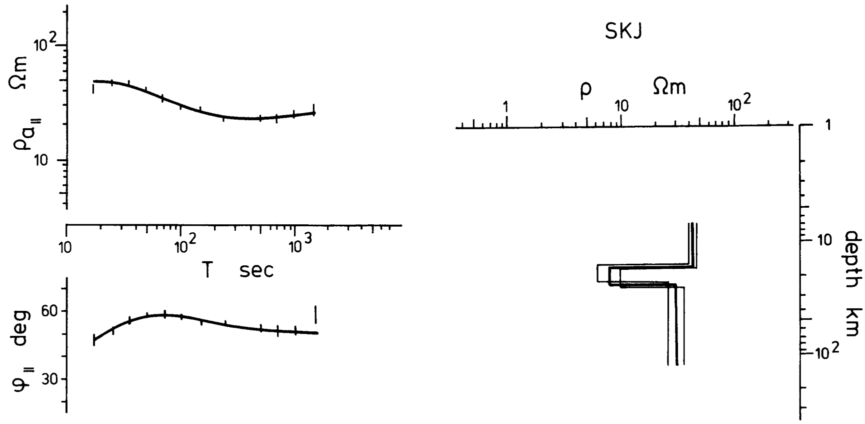

Fig. 4. To the left measured apparent resistivities ρ_a and phases $\varphi_{\parallel}$ as a function of period T at the site SKJ (Skjöldolfsstadir). The vertical lines are mean square deviations. To the right: a three-layer model with confidence levels computed by the mean square deviation of the measured values. The heavy lines through the measured values are theoretical curves calculated from the model

For calculating the models we used an inversion method given by Schmucker (1974). In all cases a good fit between the measured and calculated resistivities and phases was obtained for three-layer models. Models consisting of more than three layers did not show considerable improvement in the analyzed period range.

Figure 4 shows the measured values of resistivities and phases at the site Skjöldolfsstadir (SKJ), compared to the calculated curves and the corresponding three-layer model.

The absolute resistivity values in each layer obtained by the model calculations are different from site to site along the profile (see Fig. 5). This is probably caused by near-surface local anomalies which either tend to increase or decrease the resistivity values as a whole in all the measured period range. To a lesser extent this effect also affects the depth determinations. In spite of this uncertainty caused by resistivity variations at the surface the computed models are very similar at all MT-sites, showing only three distinct layers except the site AUS.

At approximately 8–15 km depth the mean resistivity is rather high, 50–100 Ωm. The second layer is only about 5 km thick and has a low resisitivity of 10–20 Ωm. The third layer has a resisitivity of 50–100 Ωm down to at least 100 km depth.

The most significant result of this model calculations is the good conducting layer, which appears beneath all the sites on the profile. The depth to the upper boundary of this layer increases with increasing distance from the neovolcanic zone. The depth is 13–14 km near the boundaries of the neovolcanic zone but 18–20 km in the Tertiary region. At the site Austaribrekka (AUS), the only MT-station within the center part of the active zone, the well conducting layer approaches the surface even by 1–2 km. This is clearly shown in Fig. 5.

The greatest deviation of computed phase values from measured values is observed for the longest periods (see Fig. 4). The measured phases increase with

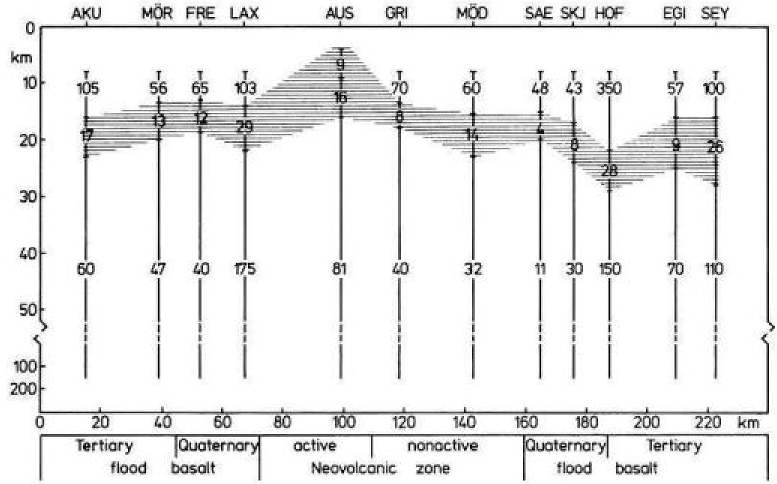

Fig. 5. Three-layer models along the east-west MT-profile calculated from apparent resistivities and phases of the *E*-polarization case. The hatched area is a low resistivity layer near the crust-mantle interface approaching the surface inside the neovolcanic zone. The numbers indicate resistivities in Ωm

increasing period. This behaviour indicates the existence of an additional low resistivity layer at great depth. From model calculations with four layers we have estimated the depth to this layer to be in the range of 200–300 km. This deep low resistivity layer can not be seen in the apparent resistivity. For that purpose the largest analysed periods of 1 h are too short. We hope to be able to delineate this deep low resistivity layer in a later stage having analysed the MT-data with periods up to approximately 12 h.

7. Discussion

In the MT-results the Quaternary and neovolcanic zones are clearly seen as an anomaly with lower electric resistivity than the Tertiary areas on both sides. This is evident both from the preference direction of the induced electric field as well as from the lowered apparent resistivity curves for the *B*-polarization in the central part of the measured profile. This behaviour must result from a zone of low resistivity underlaying the Quaternary and neovolcanic zones at shallow depth in the upper crust, probably 2–5 km. As the shortest measured periods in the MT-experiment were about 15 s the minimal skin-depth is by far too large to obtain detailed informations on this near-surface low resistivity zone. The MT-results show this low resistivity zone only as a marginal effect caused by its influence on the electric field. This result is supported by dipole-dipole measurements. At depths between 2–5 km the bulk rock resistivity is mainly controlled by ionic conduction in the pore fluid and hence by the porosity of the rocks. According to Saemundsson (1974) the Tertiary lava pile in north-eastern Iceland may be up to 4 million years older than the adjoining Quaternary and neovolcanic zones. Saemundsson has further found (personal communication), that the Quaternary lava pile contents much more sedimentary layers and is less altered than the Tertiary lava pile. This could cause an abrupt change in porosity and hence in resistivity at the boundaries of the Tertiary and Quaternary flood basalt zones in this part of the country. An increase in porosity from 1 to 3% would decrease the resistivity by one order of magnitude (Björnsson, 1976). It is noticable that no significant lateral change in resistivity is observed at the boundaries of the Quaternary and neovolcanic zones which indicates that no major change in porosity or temperature occurs at this geological contact.

The most significant result of the present MT-experiment is the existence of the low resistivity layer which can be traced along the whole east-west profile at a depth of 10–22 km, increasing in depth with increasing distance from the center of the neovolcanic zone. Only at the site AUS, which is located within the active part of the neovolcanic zone this layer approaches the surface with a minimum depth of 1–2 km.

The low resistivity layer is very likely somehow correlated to the crust-mantle boundary. The increasing depth with increasing distance from the spreading axis can be explained by a thickening of the crust with increasing age. Beneath Iceland Zverev et al. (1978) have found that focal depths of earthquakes increase from a few km down to 20–30 km on a 80 km long profile starting near our MT-site AUS running to the north-west. The depth to the base of the crust i.e., the interface of the seismic layers 3 and 4 is not known in northern and eastern Iceland. In south-eastern Iceland this interface is at a depth of around 14 km, and in the south-west it is at about 10 km (Pálmason, 1971). These depths are comparable with the depth determined by the MT-method, so it is very likely that the low resistivity layer characterises the crust-mantle interface beneath Iceland. If this conclusion is true, the magnetotelluric method can be used in a simple and time saving way to map in detail the lower boundary of the crust in Iceland.

The mean resistivity of the good conductor is about 15 Ωm. At shallower depth in the crust ionic conduction in pore fluids is dominating. Therefore, the good conducting layer within the active part of the neovolcanic zone at the site AUS, probably consists of several zones of different conductivity mechanism at different depths. The measured period range of 15 s–1 h is not adequate to obtain any detailed informations on the resistivity distribution at 0–5 km depth beneath the site AUS and hence the nature of this part of the low resistivity layer will not be discussed further in this paper.

At a depth below 10 km ionic conduction in pore fluids very probably plays a minor role, compared to electronic conduction in the solid rock itself or to conduction in a partial melt. It should therefore be possible to put some constraints on the temperature conditions and material composition at this depth by comparing the measured resistivity values to laboratory measurements of conductivity of different rocks at high temperatures and pressures.

A great number of workers have made laboratory measurements on electrical conductivity as a function of temperature and pressure for different crustal and upper mantle materials (see, for example, Parkhomenko (1967), Khitarov et

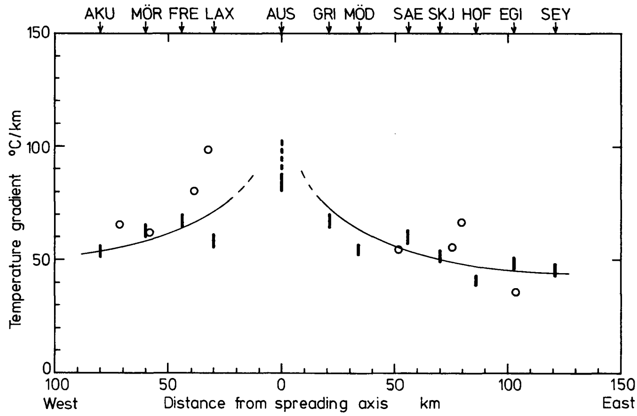

Fig. 6. Temperature gradient as a function of distance from the spreading axis in north-eastern Iceland. The vertical bars show the mean temperature gradient obtained from magnetotelluric measurements at the different sites assuming a temperature range of 1,000–1,100° C in the low resistivity layer beneath the MT-profile. Open circles are surface temperature gradients measured in shallow drill holes

al. (1970), Bondarenko (1972), Presnall et al. (1972)). All the results scatter considerably and seem to depend strongly on laboratory conditions and the measuring technique. Nevertheless the laboratory investigations on dry basalt samples yield temperatures in the range of 800–1,200° C for a resistivity of 15 Ωm, with a resonably good mean value around 1,000–1,100° C. Hence, assuming a basaltic composition of the low resistivity layer at 12–20 km depth the temperature in that layer must be about 1,000–1,100° C. This is in good agreement with the results of Hermance and Grillot (1970) obtained in south-western Iceland. According to some of the laboratory measurements partial melt starts occuring in this temperature range and then the resistivity can be lowered one or two orders of magnitude. The anomalous mantle beneath Iceland very probably consists of partially molten peridotite. If partial melting has reached a sufficient volume fraction the fluid phase and the volatile components will tend to rise to the top of the mantle and form a layer of high melt fraction at the base of the crust. It seems therefore plausible to explain the existence of the low resistivity layer as a thin zone of partial melt at the lower boundary of the crust. No exact information on the volume fraction of the melt phase can be obtained, because the resistivity strongly depends on the shape and liquid bridging of the melt pockets (Waff, 1974), but according to Shankland and Waff (1977) the melt fraction has to be at least 10% to obtain 15 Ωm in the temperature range of 1,000–1,100° C. The increased resistivity in the upper mantle below the good conductor is in good agreement with most of the laboratory measurements, which show that the resistivity of peridotite is aproximately an order of magnitude higher than that of basalt at the same temperature.

Assuming a constant temperature of 1,000–1,100° C in the center part of the low resistivity layer we have calculated the mean temperature gradient in the Icelandic crust below the whole MT-profile by dividing the temperature by the depth of the layer which is shown in Fig. 5. The results are shown in Fig. 6. The mean temperature gradient decreases with increasing distance from the spreading zone. The obtained value for the center station AUS is uncertain because of uncertain depth determination, but may be around 100° C/km. In east Iceland the calculated gradient is around 50° C/km and in the central part of north Iceland about 60° C/km. Figure 6 also shows surface-temperature gradients measured in shallow drillholes in north and east Iceland from Palmason (1973), and some new data from the files of the National Energy Authority of Iceland.

A relatively good correlation seems to exist between the surface gradients and the mean crustal temperature gradients calculated from the MT-measurements. This demonstrates the possibility to estimate temperatures in the deep Icelandic crust in an economic and time saving way from MT-soundings, especially in places where the surface gradient is disturbed by circulating near-surface ground water. In south Iceland Palmason (1973) has found a similar distribution of the surface temperature gradient but with much higher values across the western part of the active zone. On the other hand the eastern branch of the zone does not show any anomaly in the surface temperature gradient. He explains his observations as a consequence of different ages of the west and east branches of the neovolcanic zone (Saemundsson, 1974).

According to the present data it can be assumed that the low resistivity layer exists beneath the whole of Iceland at different depths. A wide spread partially molten layer at the crust-mantle interface can have an important tectonic implication as pointed out by Anderson and Sammis (1970). Especially beneath Iceland where this layer is at relatively shallow depth it can explain the mobility of the crust in the active spreading zones and the instability of these zones within the crust.

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References

Anderson, D.L., Sammis, C.: Partial melting in the upper mantle. Phys. Earth Planet. Inter. **3**, 41–50, 1970

Beblo, M.: A transportable battery-operated electrograph for permanent recording of the induced electric field of the earth. J. Geophys. **38**, 133–143, 1972

Beblo, M.: Die elektrische Leitfähigkeit unter den Ostalpen, abgeleitet aus magnetotellurischen Messungen längs eines Profiles vom Alpennordrand bis zu den Hohen Tauern. Diss. Fak. Geowiss. Univ. München 1974

Berdichevsky, M.N., Dimitriev, V.J.: Basic principles of interpreation of magnetotelluric sounding curves. In: Geoelectric and geothermal studies, A. Adam, ed.: pp. 165–221. Budapest: Akadémiai Kiadó 1976

Berktold, A., Beblo, M., Kemmerle, K.: On the distribution of the electrical conductivity below the Eastern Alps. Geol. Rundschau **65**, 715–732, 1976

Björnsson, A.: Electrical resistivity of layer 3 in the Icelandic crust. In: Greinar V. pp. 7–23, Reykjavik: Societas Scientiarum Islandica 1976

Björnsson, A., Saemundsson, K., Einarsson, P., Tryggvason, E., Grönvold, K.: Current rifting episode in north Iceland. Nature **266**, 318–323, 1977

Bondarenko, A.T., Galdin, N.Y.: On the physicomechanical and electrical properties of basalts at high pressures and temperatures. Izv. Earth Phys. **5**, 28–40, 1972

Garland, G.D., Ward, J.: Magnetic variation measurements in Iceland. Nature **205**, 269–270, 1965

Haak, V.: Interpretations-Verfahren für die Magnetotellurik unter besonderer Berücksichtigung lateral variierender elektrischer Leitfähigkeit im Erdinnern und eines räumlich inhomogenen induzierenden Magnetfelds. München: Bayerische Akademie der Wissenschaften 1978

Hermance, J.F.: An electrical model for the sub-Icelandic crust. Geophysics **38**, 3–13, 1973

Hermance, J.F., Garland, G.D.: Deep electrical structure under Iceland. J. Geophys. Res. **73**, 3797–3800, 1968a

Hermance, J.F., Garland, G.D.: Magnetotelluric deepsounding experiments in Iceland. Earth Planet. Sci. Lett. **4**, 469–474, 1968b

Hermance, J.F., Grillot, L.R.: Correlation of magnetotelluric, seismic and temperature data from southwest Iceland. J. Geophys. Res. **75**, 6582–6591, 1970

Hermance, J.F., Grillot, L.R.: Constraints on temperatures beneath Iceland from magnetotelluric data. Phys. Earth Planet. Inter. **8**, 1–12, 1974

Hermance, J.F., Nur, A., Björnsson, S.: Electrical properties of basalt: relation of laboratory to in situ measurements. J. Geophys. Res. **77**, 1424–1429, 1972

Hermance, J.F., Thayer, R.E., Björnsson, A.: The telluric-magnetotelluric method in the regional assessment of geothermal potential. In: Proceedings, Second United Nations Symp. on the development and use of geothermal resources, San Francisco, pp. 1037–1048. Berkeley-San Francisco: Lawrence Berkeley Lab., Univ. of California 1975

Kiessling, W.: Untersuchung der Variationen des erdelektrischen und erdmagnetischen Feldes im Oberpfälzer Wald zur Bestimmung der elektrischen Leitfähigkeit in größeren Tiefen mit der Methode der Magnetotellurik. Dipl.-Arbeit, Inst. Angew. Geophys. Univ. München 1970

Kitharov, N.I., Slutskiy, A.B., Pugin, A.V.: Electrical conductivity of basalts at high T-P and phase transitions under upper mantle conditions. Phys. Earth Planet. Inter. **3**, 334–342, 1970

Pálmason, G.: Crustal structure of Iceland from explosion seismology. Rit XL, Reykjavik: Societas Scientiarum Islandica, pp. 187, 1971

Pálmason, G.: Kinematics and heat flow in a volcanic rift zone with application to Iceland. Geophys. J. **33**, 451–481, 1973

Pálmason, G., Saemundsson, K.: Iceland in the relation to the Mid-Atlantic Ridge. Ann. Rev. Earth Planet. Sci. **2**, 25–50, 1974

Parkhomenko, E.I.: Electrical properties of rocks. New York: Plenum 1967

Presnall, D.C., Simmons, C.L., Porath, H.: Changes in electrical conductivity of a synthetic basalt during melting. J. Geophys. Res. **77**, 5665–5672, 1972

Saemundsson, K.: Evolution of the axial rifting zone in northern Iceland and the Tjörnes fracture zone. Bull. Geol. Soc. Am. **85**, 495–504, 1974

Schmucker, U.: Erdmagnetische Tiefensondierung mit langperiodischen Variationen. In: Protokoll DFG-Kolloquium Erdmagn. Tiefensond, A. Berktold, ed.: pp. 313–342, Grafrath-München: 1974

Shankland, T.J., Waff, H.S.: Partial melting and electrical conductivity anomalies in the upper mantle. J. Geophys. Res. **82**, 5409–5417, 1977

Thayer, R.E.: Telluric-magnetotelluric investigations of regional geothermal processes in Iceland. Ph. D. thesis, Providence: Dept. Geological Sciences, Brown University 1975

Waff, S.H.: Theoretical considerations of electrical conductivity in a partially molten mantle and implications for geothermometry. J. Geophys. Res. **79**, 4003–4010, 1974

Zverev, S.M., Boldyrev, S.A., Bourmin, V.Yu, Mironova, V.I.: Weak earthquakes in the northern part of the rift zone of Iceland. J. Geophys. **44**, 283–296, 1978

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