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GEOPHYSICS, VOL. 38, NO. 1 (FEBRUARY 1973) P. 3–13, 13 FIGS., 1 TABLE

AN ELECTRICAL MODEL FOR THE SUB-ICELANDIC CRUST†

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Active resistivity and magnetotelluric experiments in southwest Iceland suggest typical resistivities for the crust of 10 to 100 ohm-m. These low crustal resistivities are compared with calculations on the expected resistivity of fluid-saturated crustal rocks for plausible ranges of temperature, pore-pressure, and water chemistry. The comparison of the synthetic models with actual field data suggests: (1) The suppression of resistivity at shallow depth is caused by regional hydrothermal activity. (2) Appreciable effects from water are obtained from depths to 8 or 10 km. (3) Below 10 km the effects from conduction along electrolytic paths are probably dominated by conduction in the solid rock itself.

INTRODUCTION

Iceland intercepts the Mid-Atlantic Ridge and has served as a platform for a number of geophysical experiments studying regional tectonic problems. For several field seasons, our group at Brown University has been involved in research directed toward understanding thermal processes beneath the island through the interpretation of in-situ resistivity data using magnetotelluric, geomagnetic variation, and active resistivity measurements. It is apparent that an important aspect of the large regional heat flux is the generation of hydrothermal activity which is evident not only in surficial steam fields and hot springs but in suppressed resistivities at shallow depths in the crust.

The exact mechanism for coupling heat from deep-seated tectonic processes to hydrothermal activity in the crust is obscure at the present time. However, since of all geophysical properties (density, seismic velocity, magnetization, and electrical resistivity) resistivity is the most sensitive to the presence, composition, and temperature of water in rock pores, interpretations involving this parameter will no doubt be instrumental in placing constraints on the nature of the coupling process. The purpose of the following discussion is to cast a framework for interpretation of electromagnetic measurements made in Iceland by synthesizing an electrical model for the crust which is compatible with broad features of both field observations and laboratory measurements. In other words, we are proposing a model for the average electrical properties of the sub-Icelandic crust. We will see that this primitive model, although grossly oversimplified, is useful in estimating the electrical properties of water-saturated rocks in the deep crust. Moreover, it provides a backdrop against which future experiments can be planned.

FIELD RESULTS

The first indication that hydrothermal processes were important throughout the sub-Icelandic crust came to our attention during the field season of 1969 while performing magnetotelluric and active dipole-dipole resistivity measurements in southwest Iceland. Schlumberger, dipole-dipole, and magnetotelluric soundings were performed in the immediate vicinity of the field site shown in Figure 1. Moreover, rock samples collected in the area were brought back and electrical properties of water-saturated specimens were analyzed in the laboratory (Hermance et al, 1972).

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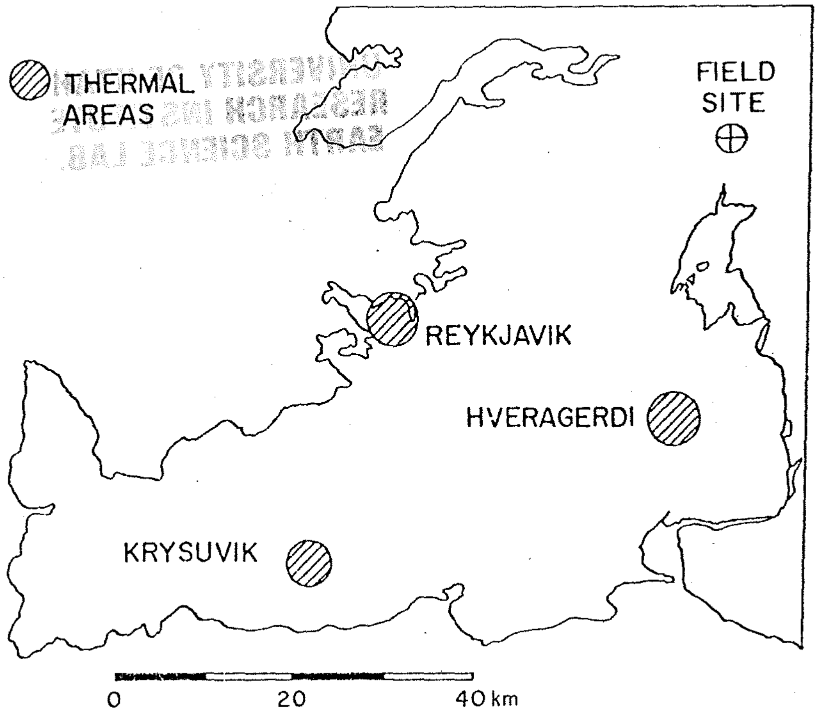

FIG. 1. Location of field site and hydrothermal areas in southwest Iceland.

From previous long-period magnetotelluric and geomagnetic variation measurements (100 sec and longer) we had anticipated suppressed resistivities at depths of 10 km and greater (Hermance and Garland, 1968). However, with a new field system in 1969 specifically designed to record short-period fluctuations, we were able to hand-pick amplitude ratios directly from the strip-chart records while still in the field and were surprised to see resistivities of 50 ohm-m at 3 to 5 sec periods as shown in Figure 2. Using the method of asymptotes (Keller and Frischknect, 1966), we estimated a *maximum* depth of 3 or 4 km to this zone of low resistivity. Subsequent digital analysis supports this model and shows remarkably small coupling between nonorthogonal electric and magnetic components. The diagonal elements of the impedance tensor are at least a factor of four smaller than the nondiagonal elements for all rotation angles. This suggests a very low degree of contamination of our data from the effects of lateral inhomogeneity.

On the other hand, surface Schlumberger soundings by the National Energy Authority of Iceland showed resistivities on the order of 1000 ohm-m (Figure 3) and no indication of lower resistivities at array spacings of up to 1 km. Only when we performed dipole-dipole measurements (Figure 4) with center spacings of 2 or 3 km was there evidence of low resistivity material at depth. The dipole-dipole interpretation suggests a *minimum* depth of 700 m.

The idea that these low resistivities are caused by *solid conduction* in *dry* materials is somewhat unreasonable as temperatures of 700–1000°C are needed at depths of only a few kilometers. It is far more reasonable to assume that hot electrolytic pore fluids are the predominant conductivity mechanism for near-surface crustal rocks.

In order to test this idea we have synthesized a number of electrical models of the sub-Icelandic crust using broad, though plausible, limits on temperature and pore-fluid resistivity, and in the following discussion we compare these models with our field observations. Our approach can be summarized by the following steps:

1) Composition: Assume, as a first approximation to seismic data, a basaltic crust 10 to 15 km thick underlain by an ultrabasic mantle.

2) Determine standard temperature and pres-

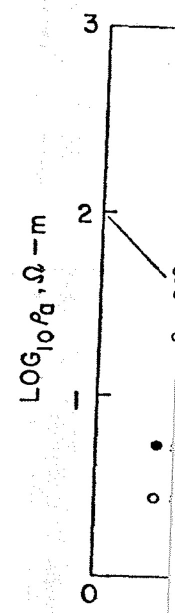

FIG. 2. Magnetotelluric ap
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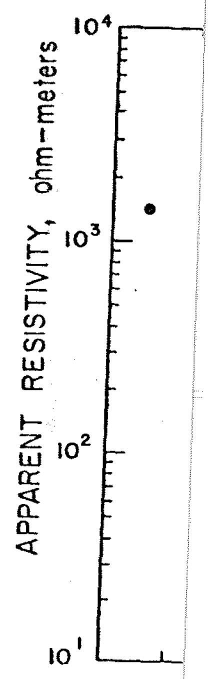

FIG. 3. Schlumberger resistiv

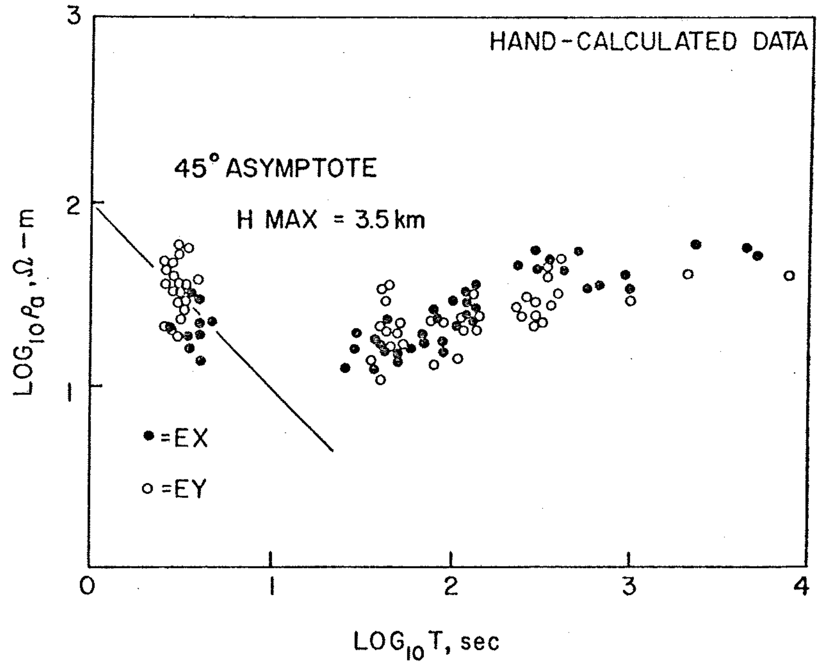

FIG. 2. Magnetotelluric apparent resistivity data. The straight line having a slope of 45 degrees suggests an asymptotic limit of 3 or 4 km as a maximum thickness for a relatively resistive zone at the surface.

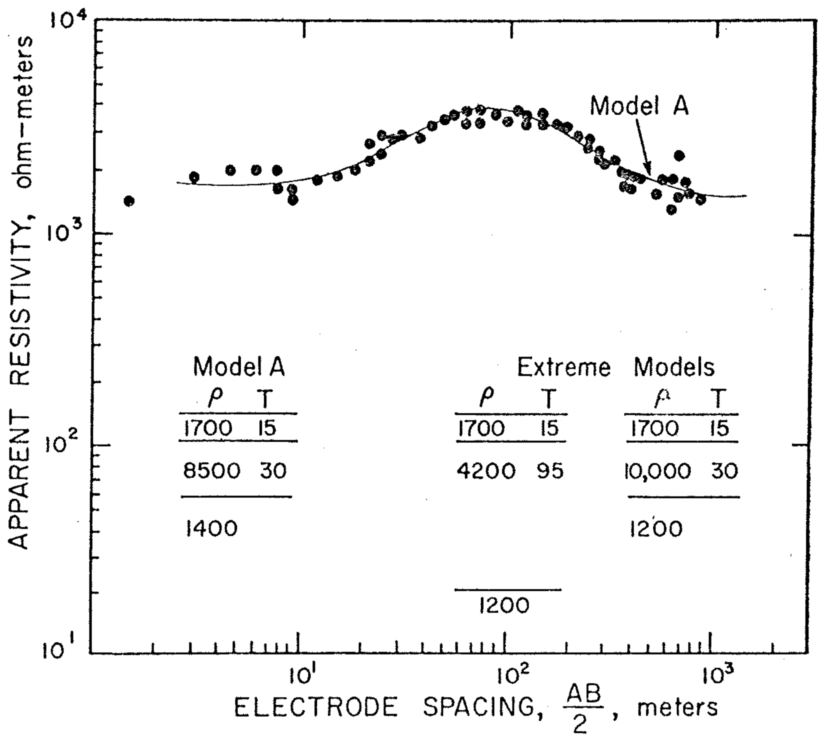

FIG. 3. Schlumberger resistivity data which indicate a 1000 ohm-m layer at greater than 50 to 100 m depth.

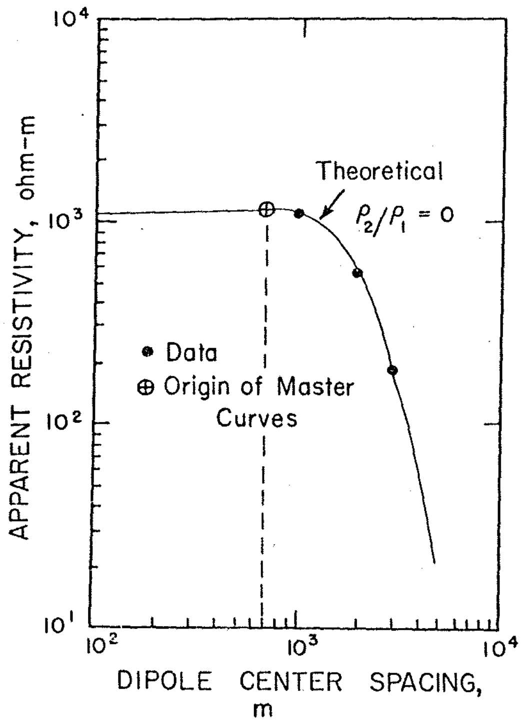

FIG. 4. Dipole-dipole resistivity data which indicate that the 1000-ohm-m surface layer (Figure 3) is underlain by a low-resistivity layer at 700-m depth.

sure conditions for a rock density of 2.8 gm/cm^3 and thermal gradients of 60°C/km and 120°C/km.

3) Estimate the contribution from solid conduction mechanisms in dry basalt.

4) Estimate the contribution from electrolytic conduction in pore fluids.

5) Determine a plausible range of total resistivity as a function of depth, calculate magnetotelluric response curves for extreme limits of this range, and compare with field magnetotelluric data from Iceland.

In principle, these calculations follow the pattern set by Brace (1971), who critically discusses the underlying assumptions. In practice, however, since our calculations apply to a specific tectonic province, we extend the application of this method in three important ways:

First, the geotherms used by Brace were based on theoretical calculations of Roy et al (1968) and were essentially extrapolations of surface gradients through a crust of high, though uncertain, heat productivity. The temperatures in our calculations, although extrapolated from surface gradients, are also tied to the temperature estimates of Hermance and Grillot (1970) at depths of 10 to 15 km based on magnetotelluric data. The assumptions of these workers that the predominant conductivity mechanism at 10-km depth beneath Iceland is solid conduction in the bulk material itself, and not electrolytic conduction through pore fluids, appears justified from the following discussion. Therefore, with a reasonably well-known surface geothermal gradient that, according to magnetotelluric interpretations, can be linearly extrapolated to depths of 10 or 15 km (Hermance and Grillot, 1970), we feel we have good temperature control throughout our crustal section.

Second, reasonably good values for the conductivity of pore fluids in near-surface rocks are obtained from extensive drilling operations by the National Energy Authority of Iceland in the area of our measurements.

Third, we compare our synthesized electrical models to actual magnetotelluric data from Iceland. The diagonal elements of the impedance tensor calculated for the data discussed below are generally small; hence, we are insensitive to lateral inhomogeneities. Therefore, magnetotelluric fields in Iceland are essentially downward-looking and provide good estimates of in-situ resistivities to which we can compare our synthesized calculations. In a sense, therefore, we are testing this primitive approach against real-earth data.

TEMPERATURE AND PRESSURE

The seismic refraction measurements of Båth (1960) and more recently by Pálmason (1970), suggest a basic crust on the order of 10 to 15 km thick. On the basis of this interpretation, we assume a specific rock density of 2.8 gm/cm^3 to calculate lithostatic pressure as a function of depth. From borehole temperatures in Iceland, temperature gradients probably lie within the range of 60°C/km to 120°C/km (Bodvarsson, 1961; Pálmason, 1970). For a thermal conductivity of .005 cgs units, these thermal gradients imply a surface heat flow of 3.0 to 6.0 microcal/cm^2/sec which is reasonable for such an active portion of the mid-ocean ridge system.

Because of low radioactive heat production in a basic crust, these gradients can very likely be *linearly* extrapolated to depths of 15 km with uncertainties of less than 10 percent; and temperatures estimated from these gradients probably bracket true temperatures in the Icelandic crust.

Data from standard tables of water density as a function of temperature and pressure (Burnham et al, 1969) can be numerically integrated to obtain hydrostatic pressure for these temperature gradients as shown in Figure 5. Also shown in this figure are curves of effective pressure for the case that pore pressure is hydrostatic. On the other hand, it is entirely plausible that pore pressure is lithostatic, in which case effective pore pressure would be zero.

CONDUCTION MECHANISMS

We assume that the total electrical conductivity of crustal material in place can be modeled by the parallel effects of conduction in the solid rock and conduction along pores and cracks containing electrolytic solutions.

A number of workers have measured the conductivity of dry basalt (Bondarenko, 1968; Coster, 1948; Khitarov and Slutskiy, 1965; Parkhomenko, 1967). Data points from these various

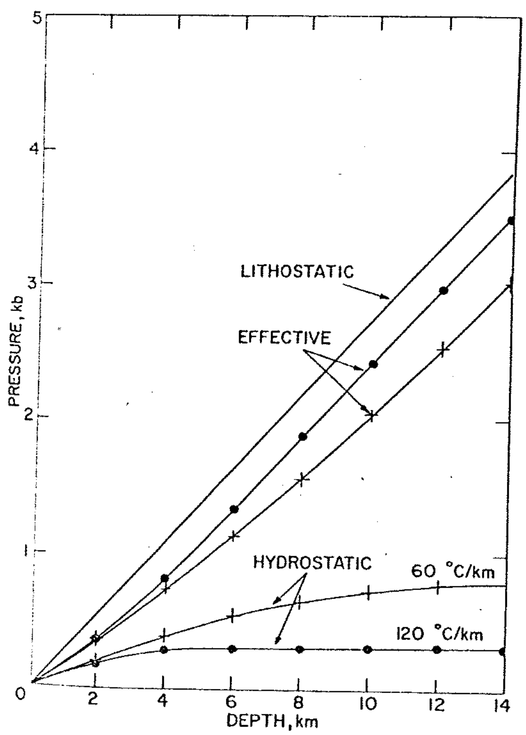

FIG. 5. Various pressure relationships in the sub-Icelandic crust for the geothermal gradients 60°C/km (crosses) and 120°C/km (dots). Shown are lithostatic pressure, hydrostatic pressures, and effective pressures for the case when the pore pressure is hydrostatic.

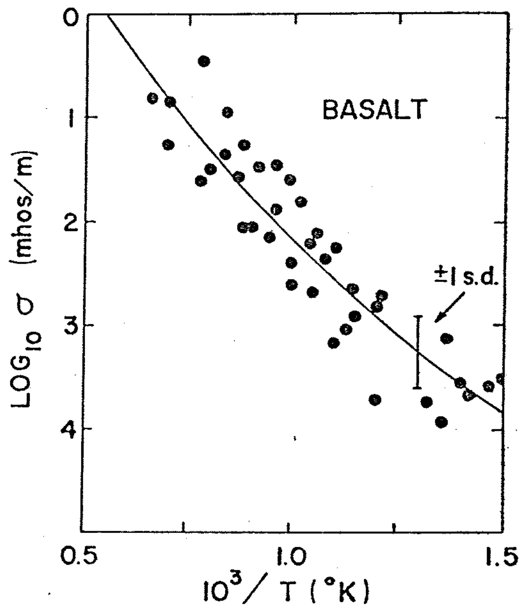

FIG. 6. Laboratory data on the electrical conductivity of dry basalt (black dots). The smooth curve is a second-degree polynomial fit to the laboratory data. The bracket spans ± one standard deviation.

experiments at essentially atmospheric pressure are shown in Figure 6, along with a curve representing a second-degree polynomial fit to the data in the least-squares sense. The second-degree fit was not a significant improvement over a first-degree fit.

The continuous curve in Figure 6 is used as an empirical relation between temperature and the contribution of electrical conduction through the solid basalt itself and allows us to estimate the component of resistivity in the sub-Icelandic crust due to solid conduction (Figure 7) for the two thermal gradients mentioned earlier.

From these curves it is quite evident that water must be playing a significant role in lowering resistivities in the upper crust, since at depths of 2 to 4 km we would predict that resistivities for *dry* basalt are at least 1000 ohm-m, whereas our magnetotelluric and dipole-dipole interpretations suggest *true* resistivities of less than 100 ohm-m.

The contribution to the bulk rock conductivity from electrolytic paths through rock pores can be estimated from Archie's Law in simple form (Brace et al, 1965; Brace and Orange, 1968) and is simply the conductivity of the fluid times the porosity squared. The porosity, in turn, depends

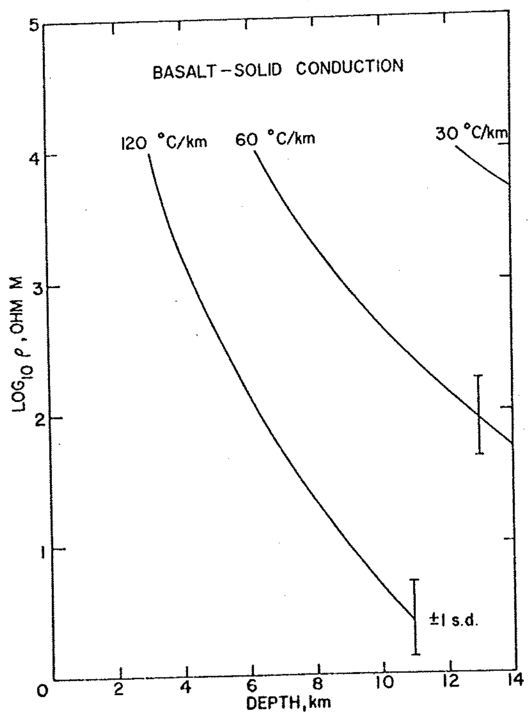

FIG. 7. The resistivity of dry basalt as a function of depth for the geothermal gradients: 30°C/km, 60°C/km, 120°C/km.

on effective pressure. The conductivity of the electrolyte itself can be treated as a sum over the contributions from various ionic members of the solution. However, using the relation of Dunlap and Hawthorne (1951), we can consider to a very good approximation the properties of the total electrolyte in terms of an equivalent concentration of NaCl. Therefore, the sum over the various species becomes an effective concentration.

In this discussion, a concentration is selected on the basis of well data, and the empirical laboratory data of Quist and Marshall (1968) is then used for determining conductivity as a function of temperature and pressure. Water from wells in the geothermal areas of Krisuvik, Hveragerdi, and Reykjavik, analyzed by the National Energy Authority of Iceland and the United States Geological Survey, has been found to have the approximate concentrations and resistivities shown in Table 1 (Bodvarsson, 1961). Also shown are data for a well on the Vestmann Islands off the southwest coast as well as data for sea water.

The salinity of sea water is not unreasonably large and represents a value between the concentrations found in the Icelandic hydrothermal fields and the much higher concentrations found for the Salton Sea and the Red Sea brines. It is not at all impossible that salinities in the deep crust might greatly exceed that of sea water, perhaps by an order of magnitude. However, we shall not consider this latter possibility in detail as there is apparently little surface evidence in Iceland to support the theory on a regional scale.

Having specified the possible range of ionic concentrations for pore fluids in rocks, in Figure 8 we show data from Quist and Marshall (1968) on the conductivity of a 0.1 molal solution of NaCl in terms of depth under conditions of hydrostatic and lithostatic pressure for the two geothermal gradients 60 and 120°C/km. At shallow depths the electrical conductivity *increases* with temperature because the viscosity is *decreasing*. The conductivity peaks at shallower depths for the larger gradient because the earth is hotter. The association of oppositely charged ions increases with temperature, which acts to inhibit the conductivity of the fluid. After the peak in conductivity is achieved, ionic association tends to decrease conductivity at greater depths.

This picture provides a pattern for the conductivity of the *pore fluid* as a function of depth. On the other hand, the bulk conductivity of the saturated *rock* depends on porosity as well as on conductivity of the pore fluid; and the porosity in turn depends on effective pressure, as shown by Brace et al (1965).

PRESSURE EFFECTS ON POROSITY

From the work of Brace et al (1965), we are fairly confident in assuming that the low-temperature resistivity of a rock containing conducting pore fluid will be given to a good approximation

Table 1. Analysis of representative hydrothermal waters, Iceland

Location	Approximate equivalent concentration of NaCl (ppm)	Approximate equivalent concentration of NaCl (molals)	Resistivity ohm-m, 25°C
Hengill well	430	0.0073	11.0
Krisuvik well	1300	0.02	3.6
Vestmann Islands	10,000	0.17	0.6*
Reykjavik	150	0.0025	35.0
Sea water	30,000	0.5	0.2

* Determined from chemical composition using method of Dunlap and Hawthorne (1951).

by Archie's Law in simple form: The rock resistivity is equal to the fluid resistivity divided by the porosity squared.

Figure 9 shows resistivity measurements reported by Hermance et al (1972) on samples from our survey area saturated with 90 ohm-m water. The porosity of the samples was about 5 percent, and the shape of the curves is typical. A point made by the above paper is that in-situ resistivities determined from dipole-dipole and Schlumberger measurements in the same area agree quite well with the laboratory measurements at zero pressure as shown in Figure 9.

Increasing effective pressure tends to close pore spaces and raise the bulk rock resistivity. Therefore, these laboratory measurements allow us to estimate the effects of pressure on the pore geometry. If the pore pressure is lithostatic, then $P_{eff}=0$, which implies that the bulk rock resistivity is simply $1/(\text{porosity}=.05)^2$ or 400 times the water resistivity at all depths.

In Figure 10 we show the saturated rock resistivity for lithostatic pore pressure and for electrolytic concentrations of 0.01, 0.1, and 0.5 molal (the latter is extrapolated from Quist and Marshall's data). Results for a geothermal gradient of 120°C/km are shown as solid lines, and results for a gradient of 60°C/km are shown as dashed lines. Since the pore pressure is lithostatic and the effective pressure is zero, these data simply reflect the behavior of the pore fluid multiplied by a factor of 400. Also indicated is the contribution from solid conduction in the rock itself. From the interception of the solid and electrolytic conduction curves, one can estimate at what depth solid conduction predominates.

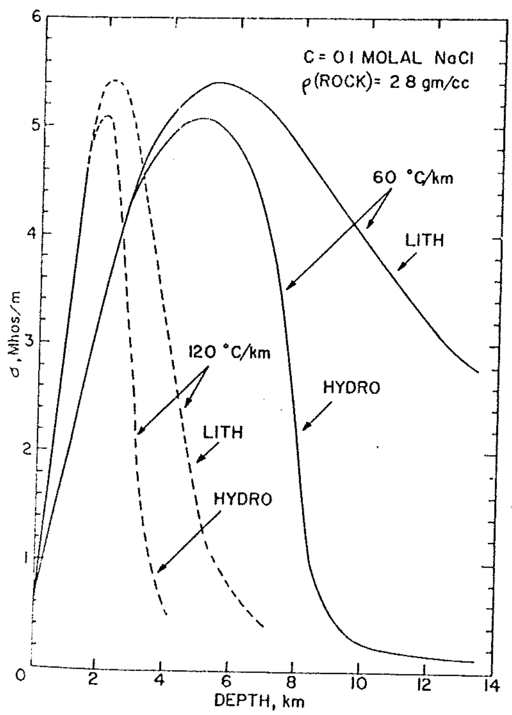

FIG. 8. Pore-fluid conductivity as a function of depth for lithostatic and hydrostatic pore pressures and the two geothermal gradients 60°C/km and 120°C/km. The electrolyte has an equivalent NaCl concentration of 0.1 molal.

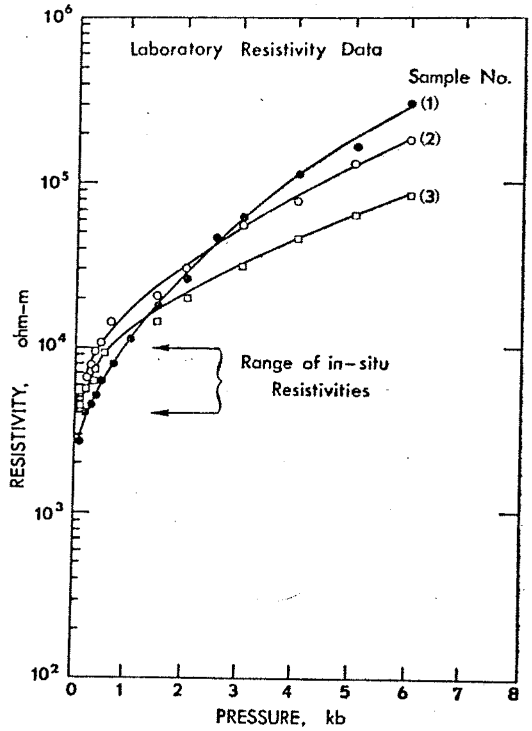

FIG. 9. Laboratory data on the low-temperature resistivity of water-saturated (90 ohm-m) specimen from the vicinity of the field site (after Hermance et al, 1972). Also shown is the range of in-situ resistivities for surface rocks estimated from Schlumberger measurements.

In Figure 11 the pore pressure is hydrostatic. The effective pressure, therefore, increases with depth, resulting in the closing of pores. Note how sharply these curves are bent upward with increasing depth as compared with the curves of Figure 10. This is simply the result of effective pressure closing up pore spaces with depth. Clearly, the actual pore pressure will be quite im-

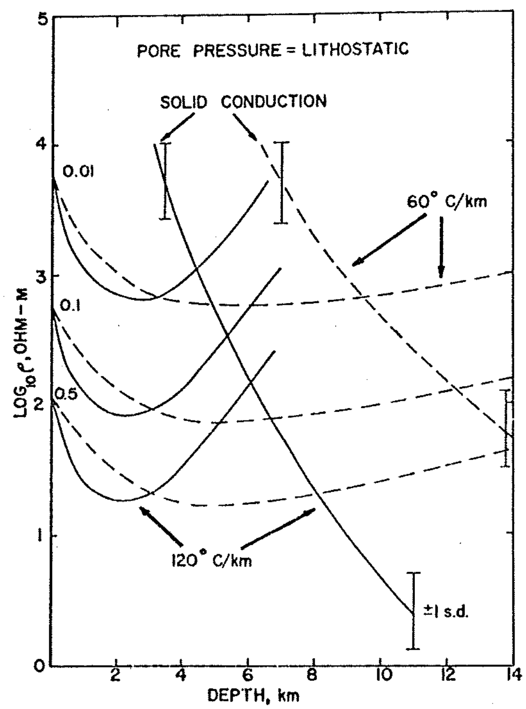

FIG. 10. Rock resistivity as a function of depth for the pore path conduction component only. The pore fluid is at lithostatic pressure and results for a range of NaCl concentration (0.01, 0.1, and 0.5 molal) are shown. Two geothermal gradients are considered: 60°C/km (dashed lines) and 120°C/km (solid lines). For comparison, the component due to solid conduction is reproduced from Figure 7.

portant in determining the electrical characteristics of rocks in the crust.

THE SYNTHESIZED MODEL

The total rock resistivity is determined by adding the contributions from solid conduction and electrolytic path conduction as parallel conductances. As a first approximation to conditions that might exist in the crust we assume that ionic concentrations may lie within the range 0.5 molal (sea water) to 0.05 molal (a concentration slightly greater than that found in the Hveragerdi and Krisuvik wells), and that the range of uncertainty in the solid conduction component is probably ± one standard deviation (defined in Figure 6). We then interpolate between the curves of Figure 10 to obtain the range of results shown in Figure 12 for a thermal gradient of approximately 100°C/km with the pore pressure equal to lithostatic pressure.

Several features of these curves bear comment. First, suppressed resistivities are seen at depths of 2 km or less. This agrees with our field evidence from magnetotellurics. Second, low resistivities characterize the entire crustal section. Finally, the transition from electrolytic conduction mechanism to a solid conduction mechanism appears to take place at a depth of 8 to 10 km.

COMPARISON OF SYNTHESIZED MODEL WITH FIELD DATA

Magnetotelluric apparent resistivity data from the field site is shown in Figure 13. In contrast to the simple analysis of hand-picked events shown in Figure 2, the data in Figure 13 are from machine calculations on digitized records using a number of techniques for estimating spectral amplitudes. The hand and machine calculations are essentially in agreement for the periods analyzed. We've determined (manuscript in preparation) that diagonal elements of the impedance tensor are less than 25 percent of the off-diagonal elements for all polarizations of the electric field. This would imply, as originally suggested by Hermance and Grillot (1970), that lateral in-

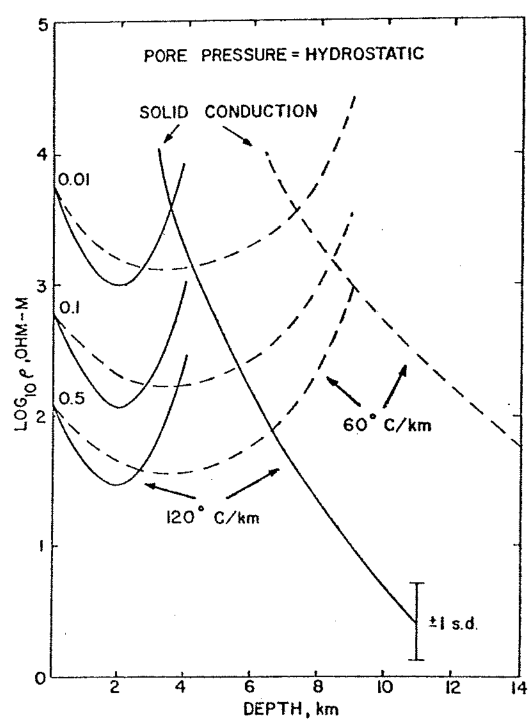

FIG. 11. Curves identical to those in Figure 10, except that pore pressures are hydrostatic.

homogeneities have second-order effects on the data, which may be neglected in the context of the present discussion. Therefore, we use simple Cagniard estimates for determining the apparent resistivity data in Figure 13 and do not discriminate between orthogonal electric field polarizations.

To compare the field data with the synthesized model of the Icelandic crust, we represent each of the continuous curves in Figure 12 by a number of discrete layers as indicated by the dashed lines in the figure. The smooth curves are for a homogeneous composition of basalt. In order to account for the compositional change from basic to ultrabasic material in crossing the crust-mantle interface, we introduce a resistivity contrast as suggested by Hermance and Grillot (1970). For curve (1), the crust is represented by 6 layers with an upper mantle having a resistivity of 60 ohm-m introduced at 10-km depth. For curve (2), a 6-layer crust extends to a depth of 15 km, where the resistivity in the mantle increases to 60 ohm-m. Note that the resistivity increase in curve (2) is off the diagram to the right in Figure 12. We should comment that the value of 60 ohm-m is probably a lower limit, and actual values could be as high as 200 ohm-m or possibly even greater. The upper mantle itself is simulated by an additional 10 discrete layers in which a resistivity of 60 ohm-m at a depth of 10 or 15 km grades smoothly to a value of 40 ohm-m at 100-km depth, such that the logarithm of resistivity is a linear function of depth. The latter functional dependence of resistivity on depth is symptomatic of a small geothermal gradient in the upper mantle. Theoretical magnetotelluric response curves calculated for the two 16-layer models are shown in Figure 13. The circled numbers in Figure 13 refer to the respective curves in Figure 12 with the introduction of the mantle described above. It is striking that models based on primitive assumptions regarding conductivity mechanisms in the crust succeed in bracketing the field data as well as they do.

One must be cautious however in drawing definite conclusions regarding ionic concentrations in pore fluids at depth from the models. It must be kept in mind that field experiments measure the *total* rock conductivity and interpretations of the *pore-fluid* conductivity are ambiguous by a factor of the porosity squared. That is to say that porosities in the deep crust may, in fact, be much less than the few percent implied above, providing

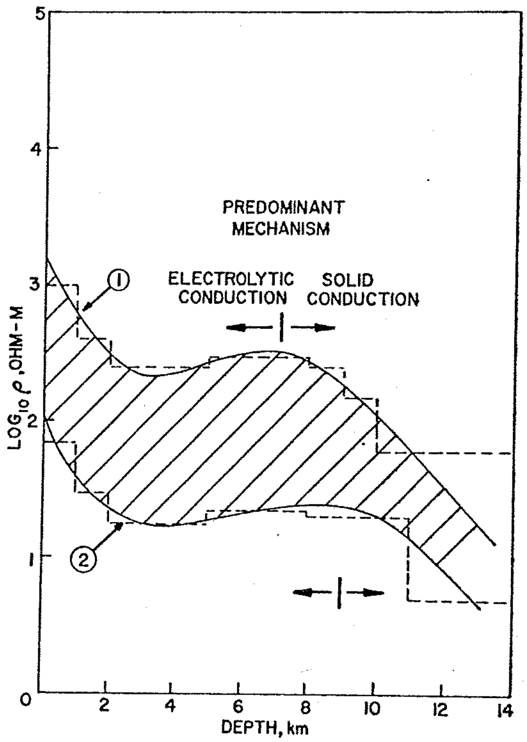

FIG. 12. An electrical model for the sub-Icelandic crust showing a plausible range of bulk resistivity as a function of depth. The depths are indicated at which electrolytic pore path conduction is dominated by solid conduction. The dashed lines are the discrete layers, approximating the smooth curves, used to calculate the magnetotelluric response shown in Figure 13.

ionic concentrations are considerably larger than those determined from well samples at the surface.

On the other hand, even though a number of resistivity models would fit the magnetotelluric data equally well, certain observations can still be drawn from these calculations. Apparently, at depths of 8 or 10 km, conduction along electrolytic paths gives way to conduction in the solid rock itself. This is not to say that water is absent at these depths, but rather that ionic association is increasing to such an extent that the electrical effects of the pore fluid are minimized. Thus, at depths of 12 to 15 km, laboratory measurements on *dry* basalt or ultrabasic materials may indeed be representative of real-earth conditions.

Moreover, at the present stage of our interpretation we are led to believe that appreciable water is present in the sub-Icelandic crust to depths of 8 or 10 km. This, however, is not conclusive since

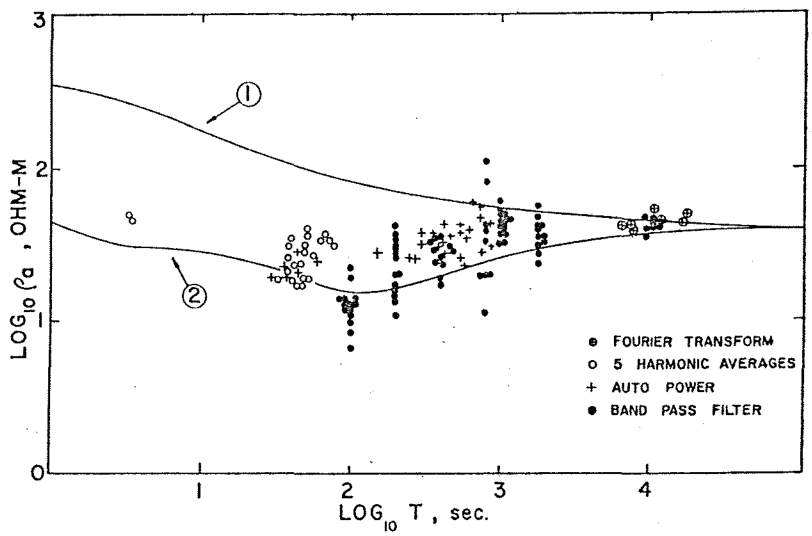

FIG. 13. Comparison of actual field data from Iceland with theoretical magnetotelluric response curves calculated from curves (1) and (2) in Figure 12. The real data have been reduced using a number of spectral analysis techniques.

an impermeable or resistive zone 2 to 4 km thick is difficult to resolve at depths of 5 to 10 km, using the magnetotelluric method alone.

Certainly questions regarding the presence of water at depths of 5 to 10 km are important to tectonophysics and petrology. The model calculations above provide a framework within which future experiments can be planned to study this problem and against which future interpretations can be contrasted.

ACKNOWLEDGMENTS

I would like to acknowledge the great benefit I have received from discussions with Bill Brace and Ted Madden at M. I. T. These workers, as well as George Keller at the Colorado School of Mines, have pioneered research toward providing quantitative estimates of the electrical properties and physical state of rocks in the deep crust. I am sure it is apparent that many of their ideas have found their way into the present discussion.

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