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Electrical Resistivity of Layer 3 in the Icelandic Crust

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Abstract.

The Electrical resistivity of oceanic layer 3 in the Icelandic crust has been measured in situ with electrical and magnetotelluric methods.
Outside the active zone of spreading and volcanism the resistivity in the upper part of layer 3 is 150-1000 Ωm. It is concluded that the layer there must be saturated with water. Beneath the neovolcanic zone much lower resistivity, about 10-20 Ωm, is caused by higher temperatures and probably somewhat higher porosity.

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Reykjavík, Iceland, 1.-7. July 1974

1. Introduction.

An active zone of rifting and recent volcanism the neovolcanic zone, crosses Iceland from south-west where it connects with the Reykjanes ridge, to the north-east where it connects with the Iceland - Jan Mayen ridge along a fracture zone (Pálmason and Saemundsson, 1974). It has a north-south direction in NE-Iceland and splits up in two separate zones in Southern Iceland having NE-SW direction. In Northern Iceland the active zone is expected to have been situated in the Langjökull-Skagi region until about 4 million years ago when it shifted ca 150 km eastwards to the present position (Saemundsson, 1974). On both sides of the neovolcanic zone there are strips of Quaternary flood basalts followed by Tertiary flood basalts on Western and Eastern Iceland.

A detailed seismic refraction survey of the Icelandic crust has been performed by Pálmason (1971). According to his interpretation there is a surface layer with P-wave velocities 2.0 - 3.3 km/s in the active zone of rifting and volcanism. Beneath it there are two layers of Tertiary flood basalts: layer 1 with P-velocity of 4.1 km/s and layer 2 with 5.2 km/s. These three layers are underlain everywhere in the island by a layer with P-wave velocity of 6.5 km/s, layer 3, which is considered to correspond to the oceanic layer 3. This layer is nowhere seen on the surface but the depth to its upper boundary has been mapped in detail by Pálmason (1971).

The temperature gradient in the Icelandic crust has been measured in boreholes by Pálmason (1971, 1973). He finds values ranging from 37°C/km on the eastern flood basalt zone up to 165°C/km in the Quaternary area in SW-Iceland. Within the neovolcanic zone the gradient is close to zero in the uppermost few hundred meters. This is caused by high fissure and crack porosity and hence a flow of cold ground water in the surface layers of the active zone.

Using the gradient and seismic measurement the temperature at the 2-3 layers boundary has been calculated (Pálmason, 1971). He concludes that the oceanic layer may be formed by a conversion of basaltic crust to amphibolite. Another possibility to explain the higher velocity in layer 3 is that it is largely composed of intrusives (Bödvarsson and Walker, 1964), which are denser than the overlying flood basalts. A combination of both these possibilities i.e. intrusives and some temperature-dependant effects, such as metamorphism or filling of pore spaces (Pálmason and Saemundsson, 1974) is possible. Beneath layer 3 there is a layer in the upper mantle with P-velocity of 7.2 km/s (layer 4). This anomalous low velocity is explained by partial melting. The temperature in this layer should therefore be around 1000°C, what is in agreement with values obtained from linear extrapolation of the temperature gradient (Pálmason, 1973) and from magnetotelluric measurements (Hermance and Grillot, 1974).

It seemed to be promising to measure in situ the electrical resistivity in the Icelandic crust, especially in layer 3, which cannot be studied at the surface. From comparing the in situ electrical resistivity with laboratory measurements on different types of rocks under various conditions, conclusions can be drawn about the moisture content, if electrolyte concentration and temperature in the deep crustal layers are known. The moisture content or porosity may play a role in explaining some geophysical phenomena like seismic activity or thermal anomalies, and may also be of an economical importance in exploitation of geothermal energy. On the other hand if porosity and electrolyte concentration are known in a certain area, then information about the temperature can be achieved in measuring the electrical resistivity.

The National Energy Authority (N.E.A.) has made numerous electrical soundings in Iceland in the last two decades, using in most cases the Schlumberger configuration with spacings up to 1500 m. These measurements have mainly been concentrated on geothermal areas, which are local anomalies in the regional behaviour of the electrical resistivity. In order to measure the resistivity of layer 3 it was necessary to extend these measurements to areas outside thermal fields and to greater depths. In 1972 the N.E.A. obtained on loan from Brown University in Providence, resistivity instruments capable of dipole-soundings with center spacing up to 10 km. Several dipole profiles were then measured in SW-Iceland. In 1973 a number of dipole profiles and magnetotelluric measurements were made on a profile across Northern Iceland. This was a part of a joint research program of the N.E.A. and Brown University, to determine the regional variance of the electrical properties of the Icelandic crust and upper mantle.

These measurements have shown a great scattering in the resistivity of the surface layers. Typical values for the neovolcanic zone (postglacial cold lava fields) are between several 1000 Ωm and several 10000 Ωm but 100-1000 Ωm in the Quaternary and Tertiary flood basalt areas. The resistivity usually decreases with depth and is often around 100 Ωm at 1-2 km depth in geologically different regions.

The main purpose of the present paper is to determine the resistivity and its lateral variation in the upper part of layer 3. In the geologically older flood basalt zones the resistivity lies between 150 and 1000 Ωm, but in the active zone it is around 10-20 Ωm.

The lower resistivity in the older regions is probably caused by much lower temperature and by lower porosity caused by gradually filling of pore spaces.

2. Location of field sites. Field measurements.

It is important to have in mind the regional geological structure and the variation of depth to layer 3 in selecting field sites. Their location is shown in Figure 1. At every site several Schlumberger soundings with 1.5 km spacing were carried out to evaluate the local homogenity and the resistivity of the surface layers. Then a equatorial dipole-dipole sounding with center spacing up to 12 km was performed.
At some of the sites magnetotelluric measurements (3-10.000 s) were made to find the resistivity in the deeper crust and upper mantle. Several other dipole soundings and a telluric survey were made during the field periods, especially around high temperature areas and other expected local anomalies. This data is being analysed and the results will be presented elsewhere.

The sites used for the present analysis are marked with the same letters as the dipole curves on the following figures. Two of them are located in the neovolcanic zone, DE-5 in Thingvellir and MVD-1 in Mývatnsöraefi east of Mývatn in Northern Iceland. The depth to layer 3 is 3.0-3.5 km at both these sites. Two sites are within eroded central volcanoes where layer 3 occurs very close to the surface HUD-3 in the Tertiary Vatnsdalur central volcano (Annells, 1968) and DE-4 in the Quaternary Kjalarnes volcano (Fridleifsson, 1973); the depths to layer 3 are 1 km and 2 km respectively. VED-1 and HED-1 are in the Tertiary basalt zones in Eastern and Western Iceland. Depth to layer 3 is around 3.5 km at the eastern site and probably around 4-6 km on the Vestfirdir peninsula. Site DDE-2 is on Mosfellsheidi in the Quaternary zone with depth to layer 3 around 2 km and finally the sites HUD-1 and HUD-2 are in the Audkúluheidi region, where the active zone is expected to have been situated 4 million years ago.

3. Presentation of the results.

Both the electrical sounding curves, obtained near Tertiary and Quaternary central volcanos in Vatnsdalur and on Kjalarnes where layer 3 is close to the surface, are shown on Figure 2. The resistivity structure with depth is obtained by comparing the field-curves with master-curves from Orellana and Mooney (1966). At both places, which are around 150 km apart, the resistivity increases in layer 3 to the same value of about 300 Ωm. At greater depth the data is scattered, probably because of lateral inhomogenites in the central volcanoes, but 300 Ωm seems to be a fair estimate of the mean resistivity in layer 3. In Vatnsdalur there is a thin low-resistivity layer at the top of the 300 Ωm layer.

Figure 3 shows dipole and Schlumberger curves from Thorskafjardarheidi on the Vestfirdir peninsula and from Mosfellsheidi in the Quaternary flood basalt zone in South-western Iceland. The resistivity of the surface layers is different in both regions but at greater depth the resistivities are similar. Also here we see an increase in resistivity to about 150-250 Ωm at a depth corresponding to the depth to layer 3. Cagniard estimates for the apparent resistivity at Thorskafjardarheidi are scattered around 200 Ωm for 10-20 s, which corresponds to ca. 20 km depth. At both sites the lowest resistivity observed is in a layer just above layer 3.

Schlumberger- and dipole-soundings and magnetotelluric measurements from the Audkúluheidi region in the old rifting zone (HUD-1 and HUD-2 in Figure 1) give a resistivity layering similar to that of Thorskafjardarheidi (Björnsson, unpublished data). At the surface there is a 270 m thick layer with 800 Ωm, then a 20 Ωm layer extending down to 1.5-2.0 km and then layer 3 with around 300 Ωm down to about 14 km depth.

Figure 4 shows Schlumberger- and dipole measurements from both the field sites lying in the neovolcanic zone. Thingvellir in South-western Iceland and Mývatnsöraefi east of Mývatn in Northern Iceland. The curves are strikingly similar although the sites are 250 km apart, but different from the curves for the sites outside the volcanic zone. The resistivity at shallow depth is somewhat higher and decreases with depth to a much lower value than in the flood basalt zones. To obtain information about the resistivity at greater depth in these areas it is necessary to extend the dipole results using the magnetotelluric measurements, which are shown in Figure 5.

The upper part of Figure 5 contains the Cagniard estimates of the apparent resistivities from Mývatns-öraefi and two model-curves, which are compatiable with the electrical structure obtained from the dipole curves. According to these models the 20 Ωm layer, which begins at 1200 m depth, reaches down to 20-30 km depth, where it is replaced by a layer with 70-100 Ωm. The lower part of Figure 5 shows the apparent resistivity at Thingvellir as a function of period, from Grillot (1973). Based on his results Hermance and Grillot (1974) have calculated models showing resistivities between 1 and 20 Ωm at depths from 2 to 10 km. This is in good agreement with the results from Mývatnsöraefi.

Finally dipole and Schlumberger measurements from Hérad in the Tertiary flood basalt region of Eastern Iceland, are shown in Figure 6. The resistivity here is much higher than on the older regions west of the active zone. It is of the order of 1000 Ωm at the depth corresponding to layer 3 and there is a minimum, around 300 Ωm, at about 1000 m depth. A magnetotelluric measurement at this site shows a strong tensor separation of the resistivities for periods greater than 50 s, but for 10 s the Cagniard estimates are scattered around 1000 Ωm in good agreement with the dipole results.

In summary the resistivity in layer 3 is highly variable in the Icelandic crust. At 5 sites in the geologically older flood basalt zone west of the volcanic zone the values are at least in the upper part of layer 3 accumulated between 150 and 300 Ωm, but in the active zone it is around 10-20 Ωm. At one site in the eastern flood basalt zone we find a 1000 Ωm resistivity in layer 3. The results are summarized in Table 1. The approximate depth to layer 3 from Pálmason (1971) and the assumed temperature gradient, inferred from Pálmason (1971) are also shown.

Table 1.

Place	dipole name see Fig.1	resistivity in layer 3 Ωm	depth to layer 3 km	gradient °C/km
Vatnsdalur	HUD 3	300	1(2)	70
Kjalarnes	DE 4	300	2	150
Mosfellsheidi	DDE 2	150	2	150
Þorskafj.heidi	VED 1	200	4-6	70
Hérad	HED 1	1000	3.5	40
Auðkúluheiði	HUD 1	300	3	70
Thingvellir	DE 5	10-20	3-3.5	180
Mývatnsöraefi	MVD 1	10-20	2-3.5	180

4. Discussion.

The electrical conductivity of water-saturated rocks depends on several parameters, in particular their content of solution (porosity), concentration of the pore fluid, temperature and pressure. Experiments on the dependence of conductivity of rocks as a function of the above mentioned factors are scarce. Several compilations on the resistivity of dry rocks as a function of pressure and temperature, and on water saturated rocks at room temperature are available (Brace and Orange, 1968). Brace (1971) and Hermance (1973) have calculated resistivities of crustal rocks at higher temperatures from the conductivity of the fluid (from Quist and Marshall, 1968) and the porosity of the rocks using Archie´s law. But they have also found discrepancies between measured and predicted resistivity at room temperature which may be caused by surface conduction along pore walls.

An experiment in which both pressure and temperature are varied for different porosity and salinity seems only to have been carried out by Parkhomenko et al. (1972). They have found a linear relationship between log resistivity and log porosity and also between log resistivity and the reciprocal temperature. Their results are summarised in Figure 7 for rocks saturated with 0,5 g/l NaCl solution, which is equivalent to concentration typical for Icelandic high temperature water. The data scatter somewhat for different samples and for different types of rocks but nevertheless it should be possible to obtain from such curves an estimate of porosity for given temperature and resistivity and for known salinity. In addition to the free moving water, rocks in sity may contain considerable amount of stationary water bound to the surfaces of the minerals and in closed pores. This water does not take part in the flow of ground water and may have much higher ionic strength than the free moving ground water and hence considerable lower resistivity (Bödvarsson, 1950; Keller and Frischknecht, 1966).

The rock samples measured by Parkhomenko et al. (1972) were first washed and dehydrated and then filled with solution of known ionic strength and hence the intercristalline water, which may have been present in situ, not considered. No measurements on resistivity seem to have been made with rock samples taken direct from the field. From this we see that porosity values obtained from Figure 7, based on Parkhomenkos data, are maximum values.

Figure 7 also shows resistivity in dry basalts after Bondarenko et al. (1972) and for comparison in 0.01 molal NaCl solution after Quist and Marshall (1968). Watanabe (1970) has measured the resistivity in basalt containing some but unknown amount of water at temperatures up to 1500°C. He finds resistivities between 3 and 30 Ωm in the temperature range 700-800°C, which is in good agreement with values expected from Figure 7.

Water from the Icelandic high temperature areas has ionic strength in the narrow range of 10-20 meq , corresponding to about 0.5 - 1.0 g/l NaCl (S.Arnórsson, personal communication). Some of the low temperature springs have similar salinity, but others are containing water with somewhat lower ionic strength. Only the high temperature areas close to the shore on the Reykjanes peninsula contain much more saline water similar to sea water in salinity.

If we compare the curves in Figure 7 with the results obtained in the previous section, it becomes obvious that the resistivity (10-1000 Ωm) of rocks in the upper part of layer 3 cannot be explained by conduction in solid dry rocks alone. The temperature of dry basalt must be at least 500°C to show such a low resistivity. Hence it is essential that these rocks are containing some amount of water.

If we assume 150-300 Ωm for resistivity in the upper part of layer 3 west of the neovolcanic zone, 70 Ωm for the overlying layer and 150-300°C for the temperature at the 2-3 layer boundary, then the porosity must be around 0.01-0.02 in layer 3 and about 0.04 in layer 2 according to the results of Parkhomenko et al. (1972), shown in Figure 7.

The anomalous high resistivity or 1000 Ωm at the site Hérad in Eastern Iceland can be satisfactorily explained with lower temperature. In this area there is one temperature gradient hole giving the lowest observed thermal gradient in Iceland 37°C/km (Pálmason, 1973). Assuming porosity around 0.01, this low temperature gradient can explain the observed resistivity of 1000 Ωm down to 5 km. Beneath about 5 km we have to conclude still lower porosity. This is in very good agreement with laboratory measurements on porosity of basaltic rocks from boreholes and from surface samples made by Pálsson (1972). He finds porosity around 0.01-0.03 for dolerite and many samples of fine grained or porefilled basalts.

In the active zone of spreading and volcanism the resistivity of layer 3 is much lower or 10-20 Ωm. This does not mean that layer 3, which is defined with seismic measurements as a layer with 6.5 km/s P-velocity, does not exist in this zone. The electrical resistivity depends much more on temperature than does the P-velocity. Pálmason (1971) has shown that a variation of 200°C at the 2-3 layers boundary would give a negligible variation (about 0.065 km/s) in P-velocity. But such a variation in temperature could cause a change in resistivity by a factor of 3-5.

According to Pálmason the temperature gradient may be of the order of 180°C/km in the western part of the neovolcanic zone in SW Iceland where the site Thingvellir (DE5) is located. He finds much lower gradient on a regional scale near the eastern brance of the active zone and also in Northern Iceland. But we believe that, at least in a narrow zone in the middle of the northern active zone, similar values are valid for the gradient as in SW-Iceland. The high hydrothermal and volcanic activity in this area support this opinion. Using a 180°C/km temperature gradient we find 500-600°C at the 2-3 layer boundaries beneath Thingvellir and Mývatnsöraefi. With this value and the observed resistivity of 10-20 Ωm we find, using Fig. 7, a porosity around 0.03 in the upper part of layer 3 beneath the volcanic zones. If the temperature is much lower i.e. about 300°C at 2-3 km depth, then the observed resistivity has to be caused by higher porosity or about 0.04-0.05 in the volcanic zone.

The low resistivity layer (10-20 Ωm) starting at about 2 km depth reaching down to the top of layer 3 must have similar porosity.

It should be emphasized that all above mentioned values for porosity are obtained from the laboratory experiments of Parkhomenko et al. (1972), where conductivity in more saline water, eventually absorbed to the surface of pores, has been neglected. These porosity numbers are therefore maximum values and the actual porosity may be considerable lower.

Nevertheless we believe that the obtained difference in conductivity beneath the neovolcanic and the flood basalt zones must be explained by difference in porosity, probably caused by gradually filling of pore spaces approaching the geologically older areas.

The difference in conductivity could also be explained by about ten times more saline ground water beneath the neovolcanic zone, but this is rather unlikely regarding the small variation in salinity of Icelandic high temperature water.

Acknowledgements.

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Fig.1

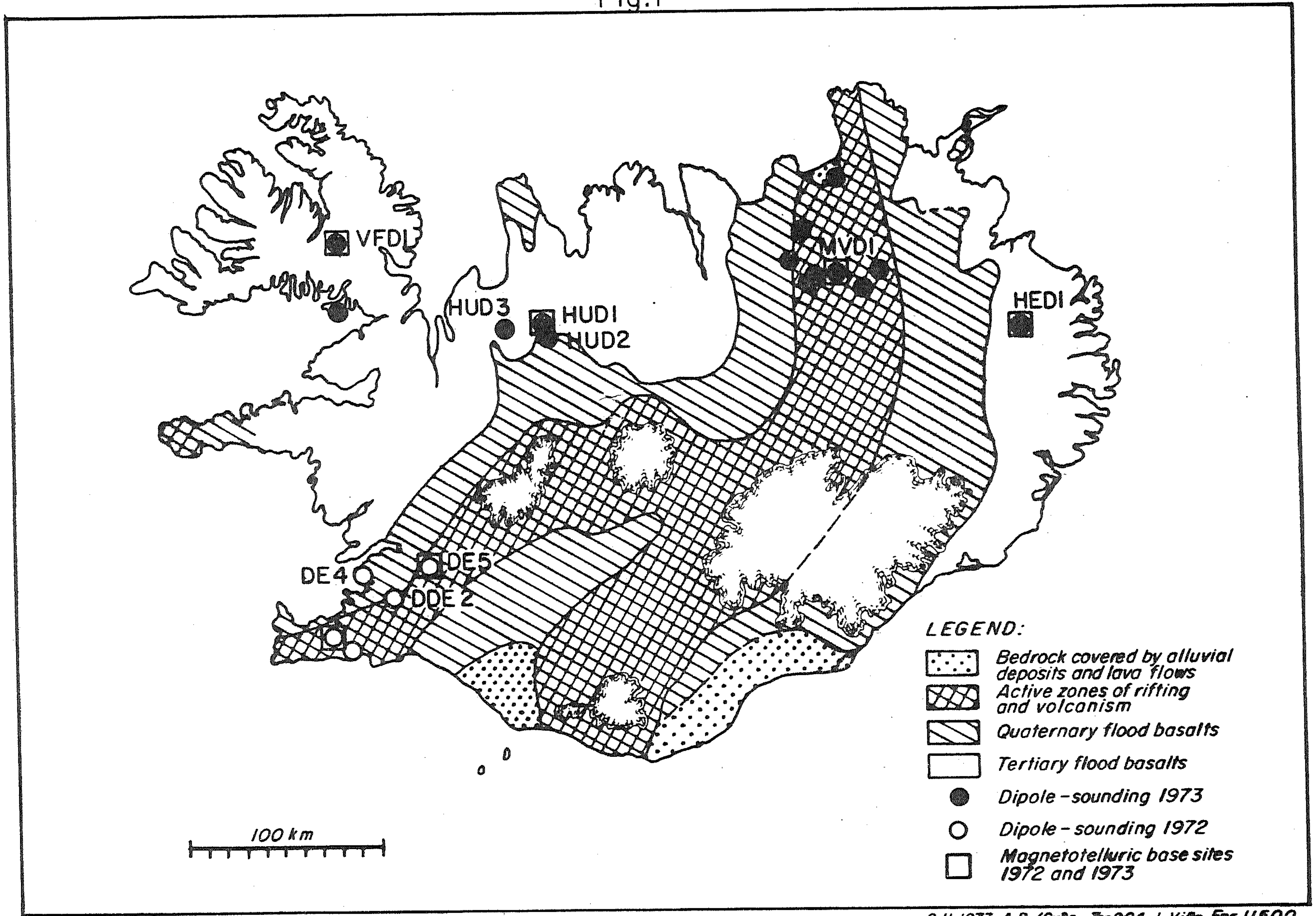

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Fig. 2

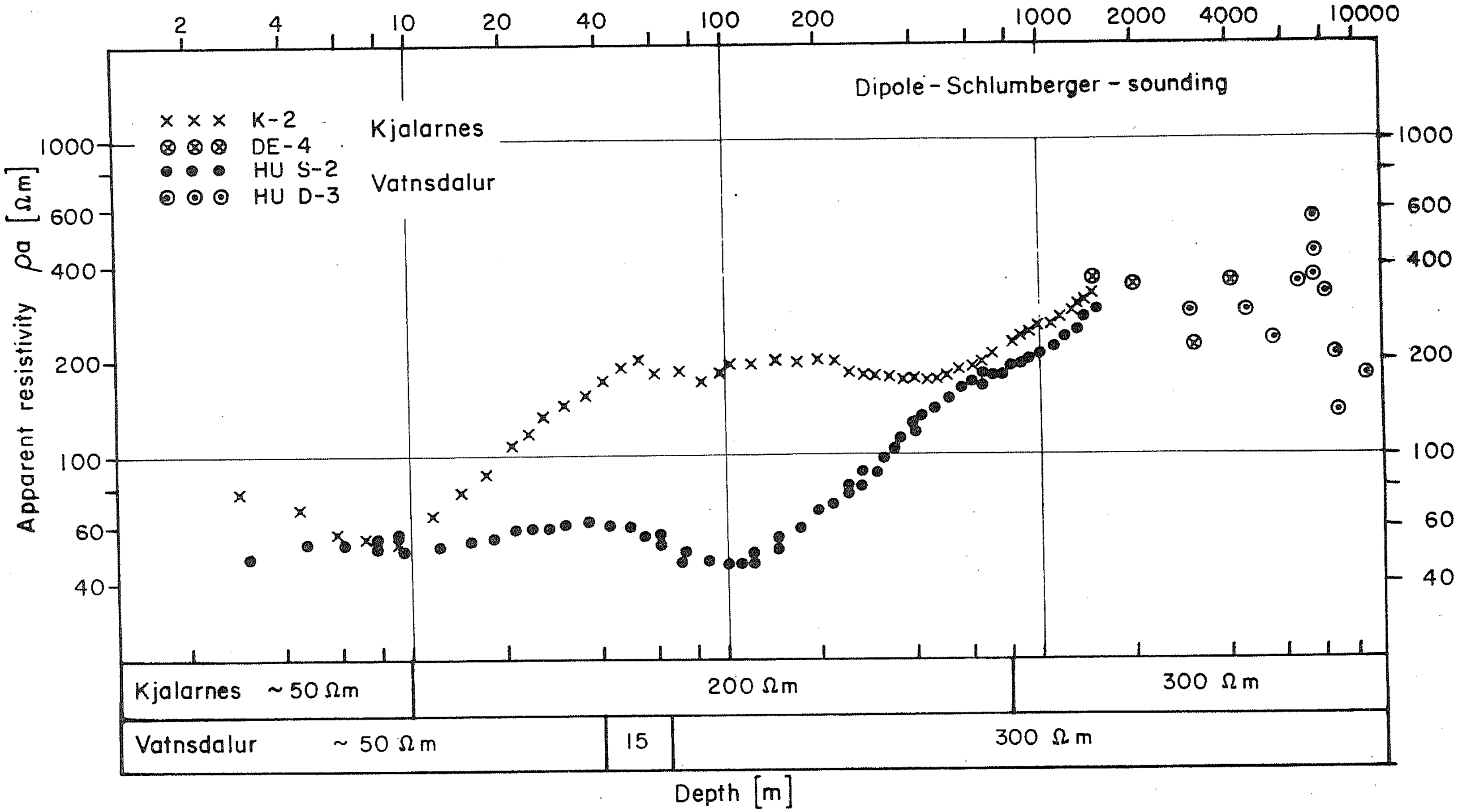

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Fig. 3

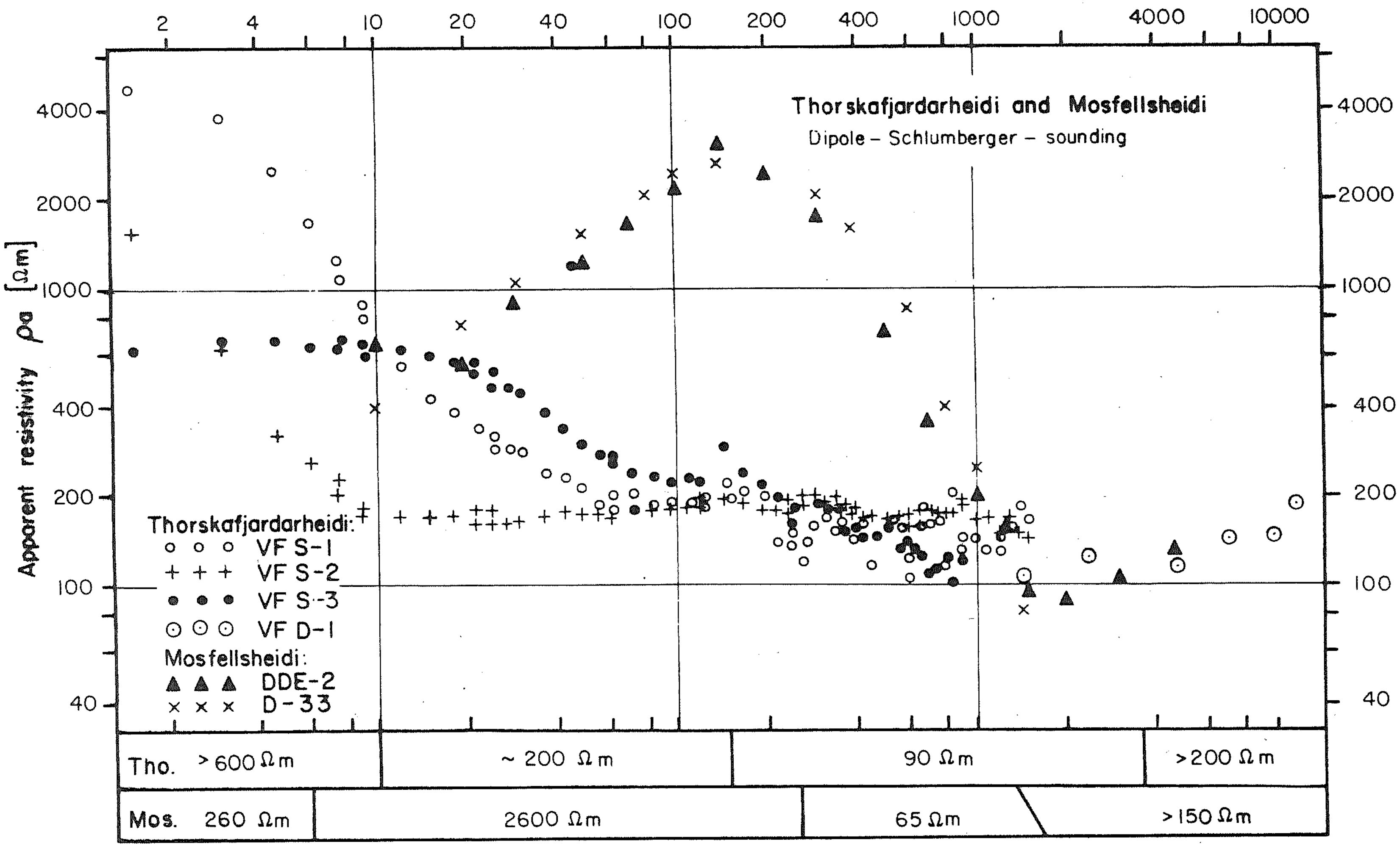

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Fig. 4

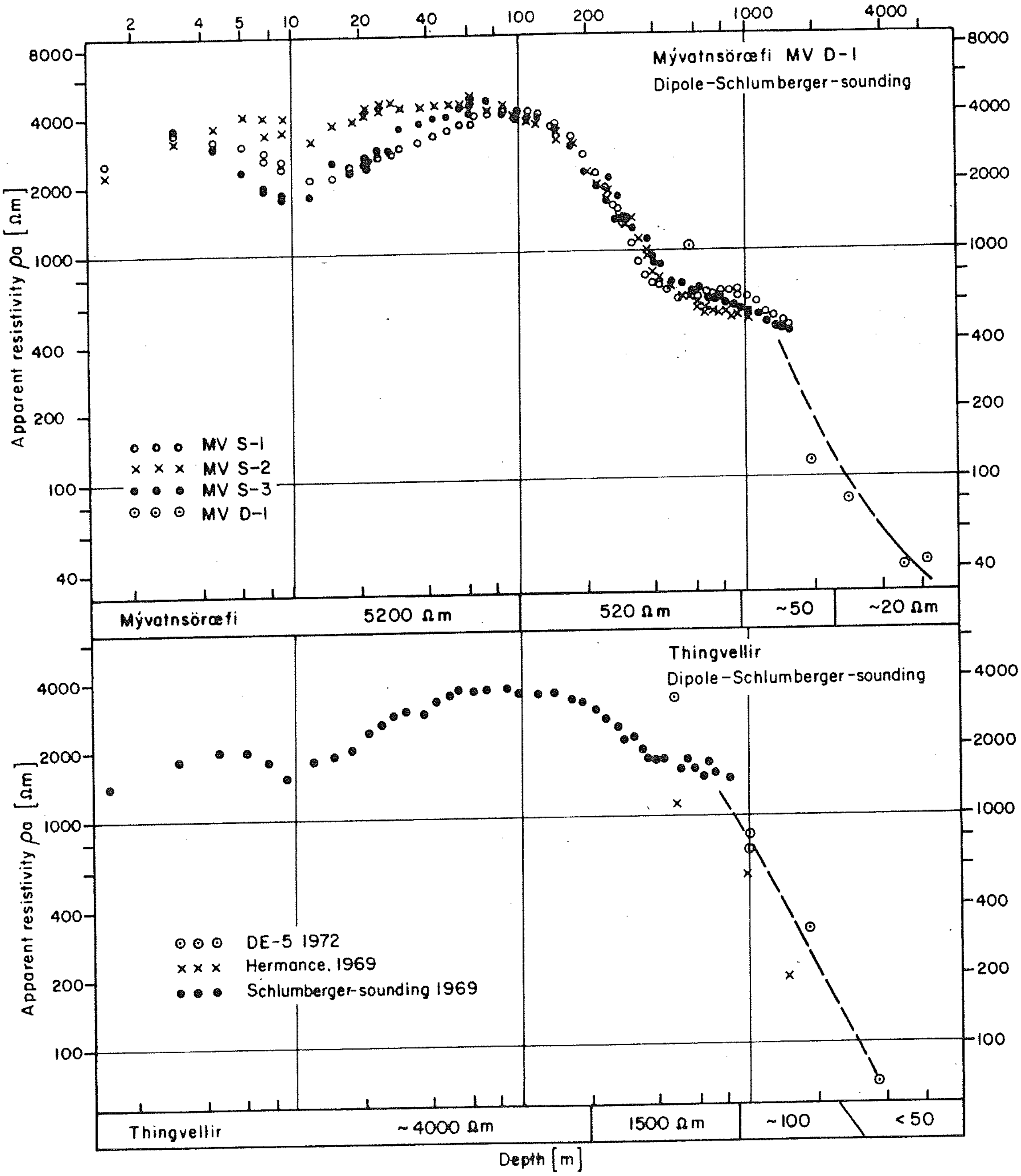

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Fig. 5

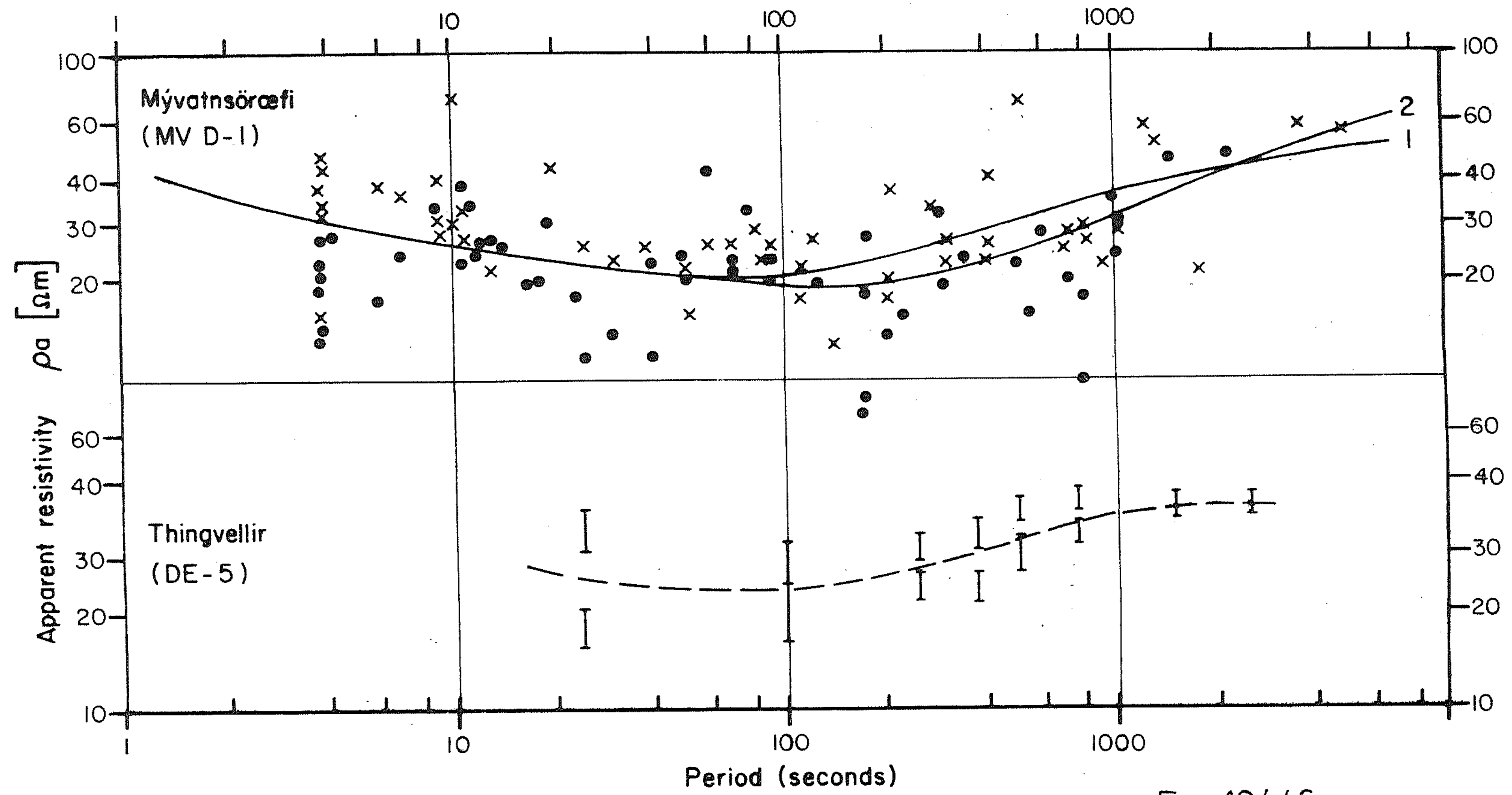

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Fig. 6

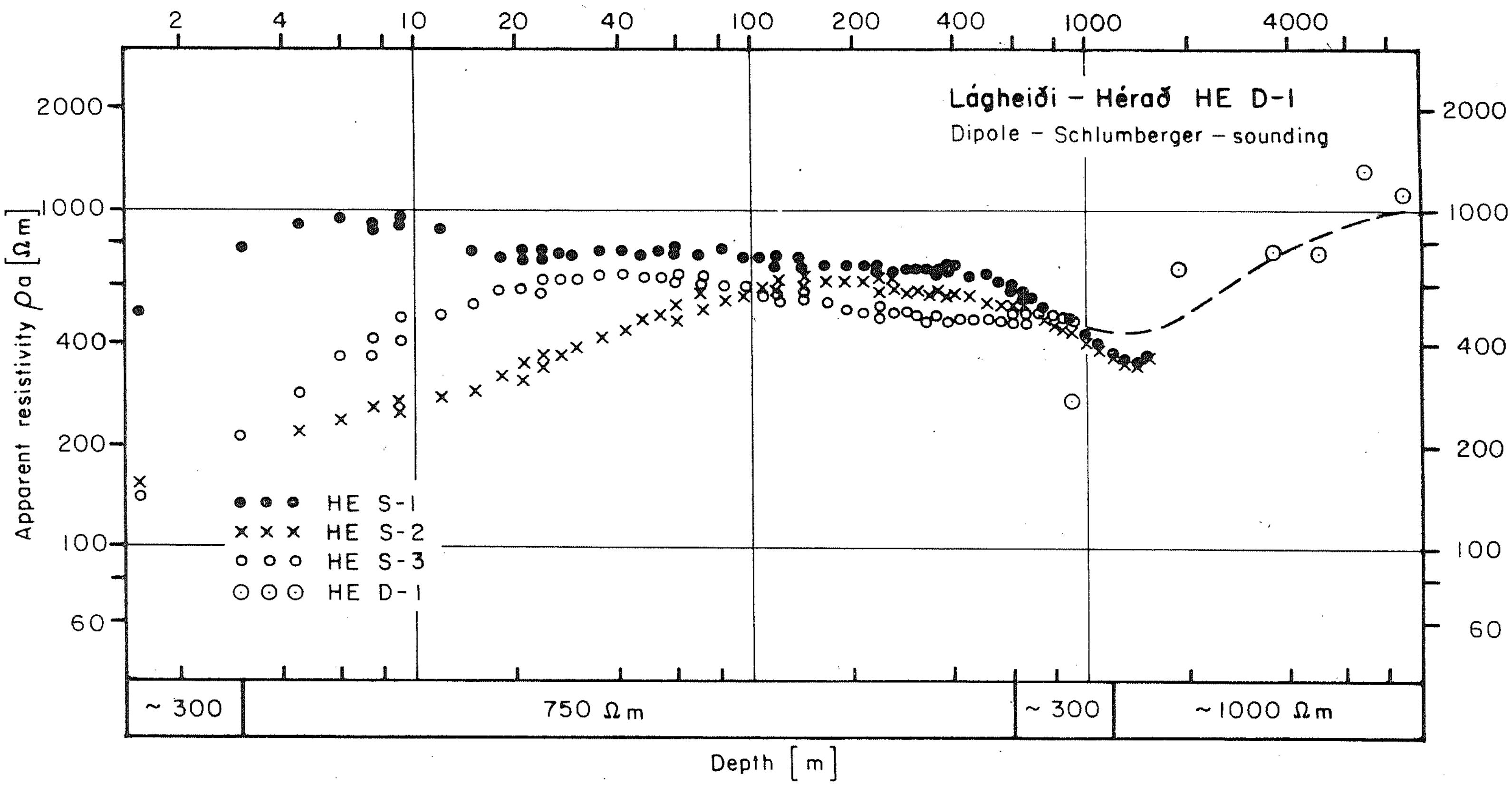

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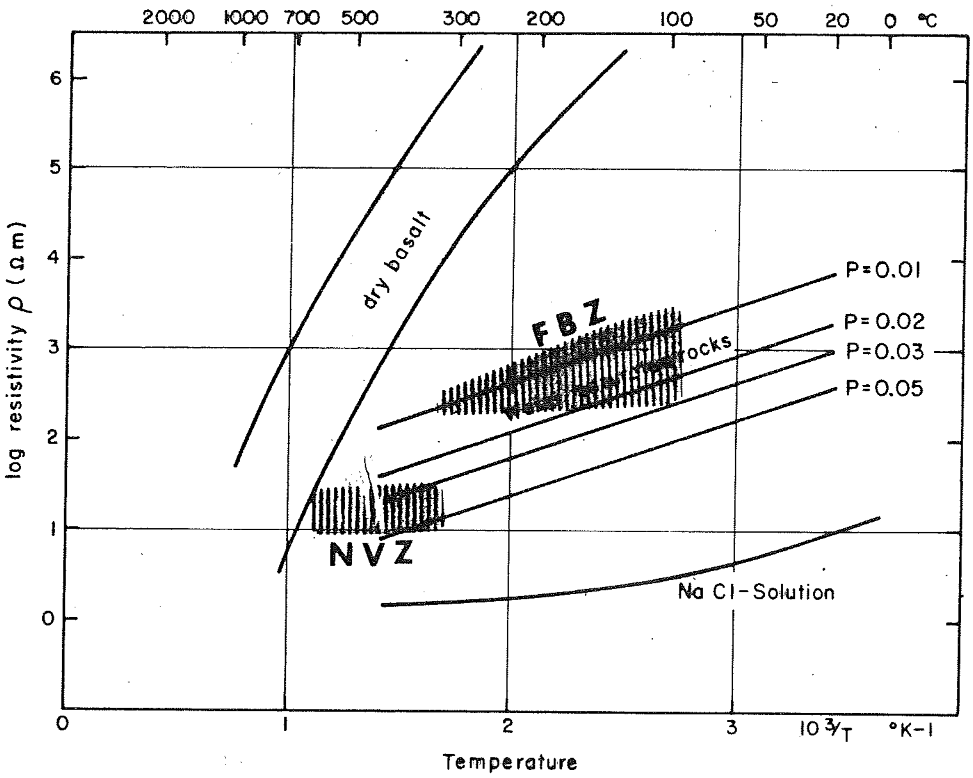
Fig. 7



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