# SCHLUMBERGER VERTICAL SOUNDINGS: TECHNIQUES AND INTERPRETATIONS WITH EXAMPLES FROM KRÍSUVÍK AND GLERÁRDALUR, ICELAND AND OLKARIA, KENYA

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## ABSTRACT

Location of geothermal fields with strong surface expressions such as hot springs, mud pools and fumaroles is relatively easy. However not all geothermal systems have surface manifestations. The goal of electrical resistivity measurements in geothermal energy exploration is to delineate resistivity anomalies in the subsurface. Several methods have been applied in this type of prospecting; the Schlumberger vertical sounding and Headon profiling being the most common.

The Schlumberger sounding has its own inherent characteristics. Theoretical background, field procedure and treatment of data is discussed. To illustrate how this method has been applied within the framework of integrated geological, geochemical and geophysical exploration two case histories of Gleradalur, Iceland, and Olkaria, Kenya, are presented. Both Schlumberger sounding and Headon profiling have proven fairly successful at these geothermal fields. Data from Krísuvík in Reykjanes Peninsula, Iceland, shows high surface resistivities which rapidily decrease with depth to less than 10 Ωm in a couple of hundred metres.

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# 1 PREFACE

## 1.1 Scope of work

This report is as a result of the six-month course undertaken by the author at the UNU Geothermal Training Programme in Iceland in 1986. The course was partly sponsored by UNDP and partly by the Icelandic Government. The training programme is designed for professionals with some experience in geothermal work in their home countries. The participants attend an introductory lecture course lasting four to five weeks, which is followed by practical training in a specialised field. The lecture course is composed of topics on Geology, Geophysics, Geochemistry, Borehole geology, Reservoir engineering and Utilization of geothermal energy resources. Participants are required to sit for two written tests.

The author participated in geophysical exploration specialised training; several detailed lectures were given. However emphasis was placed on practical training on how to conduct geophysical surveys of geothermal fields and data interpretation. The author carried out resistivity measurements of Schlumberger soundings and Headon profiling at Krísuvík and Arbaer, respectively. The data was modelled in 1-D and 2-D using the computer facilities at the National Energy Authority (NEA). The report reviews and discusses some of the features and results from the Schlumberger sounding. Some data from Eburru, Kenya, was also interpreted; the results, which are in a separate report, shall be presented to the Kenya Government.

There was a two-week field excursion to the many high and low temperature geothermal areas of Iceland, including several district heating systems and factories using hot water.

## 1.2 Introduction

In the past decades resistivity sounding and profiling have proved to be valuable tools in prospecting for geothermal energy in many countries. Three principal variations of direct current methods have found use in geothermal energy exploration, though there have been controversies in the literature over the relative merits of these techniques. The best tested and most widely used method is the Schlumberger vertical sounding. It is mainly used in surveys which are carried out in order to map resistivity anomalies in the subsurface. The primary objective of this paper is to review some of the inherent characteristics of the Schlumberger sounding. Some of these features have advantages over other geophysical methods, while others act as limitations. In this paper the importance of identifying the optimum field working conditions, data treatment and interpretations shall be emphasized. Often it is hard to assess the degree to which models fit a given data. Several interpretative methods (inversion, curve matching, iteration) shall be discussed. Data from Krísuvík in Reykjanes Peninsula, Iceland is interpreted. A review of two case histories from Gleradalur (Iceland) and Olkaria (Kenya) is given in order to asses the success or failure in the application of this method.

## 2. FIELD PROCEDURE

### 2.1 Introduction

Regardless of the specific electrode spread employed, there are only two procedures in resistivity work. The particular method to be used depends on whether one is interested in resistivity variation with depth or with lateral extent. The first one is called Vertical Electric Sounding (VES) while the second is Electrical Profiling. In this paper we shall mainly concern ourselves with VES.

In Vertical Electric Sounding the fraction of total current which flows at depth varies with the current electrode separation. Thus in order to delineate resistivities downwards in the subsurface it is necessary to arrange electrodes in such a manner that change in current electrode separation gives information about resistivity structure at depth.

### 2.2 Electrode separation and instrumentation

Figure 1 shows the field arrangement of the Schlumberger VES. Four electrodes are arranged collinearly, the current electrodes A and B on the outside while the potential electodes M and N on the inside. It may be of interest to note that, in virtue of Helmholt's reciprocity theorem in the theory of electric circuits, the resistivity value will be unaltered if AB and MN are interchanged.

To change the depth range under investigation the distance AB is progressively moved outwards symmetrically, keeping MN fixed. However when the ratio AB/MN becomes too large the potential drop across MN becomes too small to be measured within reasonable accuracy. This necessitates increasing MN. It shall be discussed in the next chapter why MN should be kept at MN≤AB/5 in order to give a resistivity value correct to within a couple of per cent. At any AB/2, readings should be made at several MN/2 values in order to indicate the presence of lateral inhomogenity.

The current used in resistivity measurements is usually DC from a battery; or a low frequency square wave with a commutator to alternate the direction of current flow at time intervals ranging from a tenth of a second to tens of seconds. If AC is used, the current density, due to 'skin-effect' phenomenon, tends to
concentrate towards the surface (Koefoed, 1979).

Requirements for MN electrodes are different from those of current electrodes. First, contact resistance is not as important. However if metallic electrodes are used, polarization potentials will form across them. This is due to the fact that electrolytic solutions in the ground have different compositions in different places. To minimise this effect 'porous pots' are used as potential electrodes. Often when potential values are read they are disturbed by the voltmeter taking some current from the ground and by natural noises in the ground. These effects are minimised by using voltmeters of high impedance and by taking measurements at both directions of current flow in conjunction with injecting an adjustable voltage from the battery.

## 2.3 Initial treatment of data

It is desirable to make some initial judgement about the quality of the data collected as measurements are being carried out in the field. As data values are measured they are plotted on a double log paper. Any values that are obviously out of step should either be repeated or the source of error (such as current leakage) checked. With experience many subsurface structures can be inferred from just looking at the field graph. This will enable the field crew to know when an anomaly is being traversed and hence decide if more soundings are necessary.

The preliminary field interpretation is followed by more precise assessment back in the office using standard master curves or computer facilities. It will be discussed in chapter four how the data is analyzed to yield quantitative results which in turn provide information regarding subsurface geological parameters.

## 2.4 Summary of field precautions

(i) Surface inhomogenities introduce misleading results in the potential reading by distorting the current flow pattern. Therefore the centre of sounding should be sited in places where potential electrodes will encounter little surface inhomogenity.
(ii) Good contact should be ensured at the current electrodes. If need be, the area around the electrodes should be moistened.
(iii) Since currents being injected into the ground can be quite big, it is important that personnel take extra care in handling the cables.
(iv) Checks for current leakages both at the start and end of measurements will contribute to the quality of the data.
(v) Good conductors like wire fences, underground telephone lines and pipes tend to distort the current pattern. These can be avoided by placing the potential electrodes as far as possible from these objects.
(vi) Powerlines, especially those accompanied by earth wires, can be a nuisance when measurements are taken close by as they may introduce high voltages in the ground.
(vii) Poorly maintained equipment cannot be expected to perform a good job. Hence regular checkups are necessary including replacement of wornout accessories like batteries.
(viii) Finally, it should be noted that good book keeping of the field data and locations of measurement positions are essential.

# 3 THEORETICAL BACKGROUND

## 3.1 Potential distribution at surface of ideal earth

In resistivity measurements using VES method the measured parameters are: current injected into the ground and a voltage drop at a known electrode configuration. Often it is not easy to relate these measured quantities to resistivity stratification at the subsurface. To derive this relation many techniques have been employed, relying on several assumptions. It must be pointed out, however, that the determination of resistivity may solve an electric problem but not necessarily a geological one; the reason being that given geological formations are not associated with definite resistivities except in a broad and general manner (Parasnis,D.S. 1972).

The model which best describes many geological situations is that represented by an earth composed of a finite number of horizontal layers. Each layer is electrically homogeneous as well as isotropic. The field is generated by a point current source at the surface, the current being direct current. Koefoed (1979) and others have derived the expression for the potential at the surface; an outline of the derivation is presented below.
The basic equations for the derivation are

$E=\beta J$ Ohm's Law ..........................(1)

$div.J=0$ Divergence condition ...............(2)

where E=potential gradient
J=current density
$\beta$=resistivity of the medium.

Combining the two equations,

$$\nabla^2 V=0 \qquad \qquad (3)$$

where V=potential.

Equation (3) can be solved by use of polar or cylindrical coordinates. Using the relationship

$$I = \int_s J \cdot ds$$

it can be shown that for a single current source

$$V = \beta I/2\pi r \quad \ldots\ldots (4)$$

For several sources of current the potential at any point on the surface is found by adding all the contributions of the current sources, since potential is a scalar. That is, potential at a point M is

$$V_m = \beta/2\pi \; \Sigma I/r \quad \ldots\ldots (5)$$

and for potential at point N

$$V_n = \beta/2\pi \; \Sigma I/r \quad \ldots\ldots (6)$$

where r is the distance between the current and potential electrode.

However in practice there are two current electrodes A and B (Figure 2). One being a source (positive) and the other a sink (negative) i.e.

$$I_a = -I_b$$

From the Schlumberger configuration, shown in Figure 1 it follows that

$$V_m = \beta I/2\pi \; (1/AM - 1/MB) \ldots\ldots (7)$$

$$V_n = \beta I/2\pi \; (1/AN - 1/NB) \ldots\ldots (8)$$

The potential difference divV across MN is then

$$V_m - V_n = \beta I/2\pi \ (1/AM - 1/MB - 1/AN + 1/NB) \ \ldots\ldots(9)$$

Simplifying,

$$\text{divV} = \beta IK \ \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots(10)$$

where K is known as the geometric factor depending on the electrode configuration.

In the Schlumberger array AM=NB and, designating

$$AB/2 = s$$
$$MN/2 = p$$

it can be shown that

$$\beta = \text{divV}/I \ \ \pi/2 \ \ (s^2 - p^2)/p \ \ldots\ldots\ldots\ldots\ldots(11)$$

or

$$\beta = \text{divV}/I \ \ K_s$$

$K_s$ being the geometric factor for the Schlumberger array. Since the earth is not homogeneous $\beta$ is not the true resistivity but the apparent resistivity, $\beta_a$ which may be larger or smaller than, or in rare cases identical with, one of the resistivity values in the heterogeneous earth. Therefore,

$$\beta_a = \text{divV}/I \ \ K_s \qquad \ldots\ldots\ldots\ldots\ldots\ldots(12)$$

In the field, parameters are divV and I. However divV, the potential gradient, can not be determined exactly, since it is defined as the limit of the ratio of the voltage to spacing as the two potential electrodes, MN, move closer together. The error introduced into the resistivity by having a measurable MN can be evaluated. Keller et al, (1970) have shown that in order to keep the apparent resistivity within a given error limit, say e, the following relation must be satisfied.

$$p^2 = 4e/(1+e)\ s^2$$

For Schlumberger Sounding it is desirable to use e=5 per cent, which means that,

$$p \le 0.435s$$

## 3.2 Relationship between Apparent Resistivity Function and Resistivity Transform Function

In the previous section we discussed the relation between potential difference across two points on the surface and current distribution for a perfectly homogeneous earth. However in practical field problems it is realistic to use a model that has some resemblance to the really earth. There are a number of approaches in mathematics that are used in determining potential fields in a layered medium. The commonest approach has been discussed by Koefoed (1979) in which a special solution to Laplace equation

$$\mathrm{divgrad}V = 0$$

is satisfied and holds true for each layer of stratified earth. Here we shall only quote the final results of the derivation. The potential function through the layers solved from the above equation must meet the following two conditions, namely, the potential function, V, must be continuous across the boundaries of different layers otherwise $\mathrm{grad}V \rightarrow \infty$ which would imply $J \rightarrow \infty$ which is impossible. Secondly, the normal component of the current flow across the boundaries must also be continuous; this is from the requirement that the current charge must be conserved. The component of current along the plane boundary is in most cases discontinuous since resistivity of various layers are different.
The equation for the potential distribution due to a point current source on the surface is given by

$$V = \beta_i I/2\pi \ [1/r + 2\int_0^\infty K(\alpha)\ J_0(\alpha r)\ d\alpha] \quad \ldots.(13)$$

where $\beta_i$ = resistivity of a given layer

$I$ = current

$\alpha$ = variable of integration

$r$ = distance between current source and measuring point

$K(\alpha)$ = Kernel function controlled by resistivities and thicknesses of the layers

$J_0$ = Bessel function of zero order.

In resistivity measurements (in our case using Schlumberger array) the measured quantities which occur in the expression in the apparent resistivity are I, divV, s, and p. It can be shown from Equations 11 and 13 that

$$\beta_a = s^2 \int_0^\infty K(\alpha)\ J_1(2\alpha s)\ \alpha\ d\alpha \quad \ldots\ldots\ldots\ldots\ldots.(14)$$

In theory, it is possible to calculate the apparent resistivity curve from Equation 14. In practice, however, the reverse procedure is the case. The apparent resistivity curve is known and we wish to determine the Kernel function $K(\alpha)$ which contains the layer parameters. The work involved in computing $K(\alpha)$ is enormous and can best be done with the aid of a computer. Normally, the computing time is lessened by transforming $K(\alpha)$ to another function $T(\alpha)$ such that Equation 14 becomes

$$\beta_a = s^2 \int_0^\infty T_1(\alpha)\ J_1(2\alpha s)\ \alpha\ d\alpha \quad \ldots\ldots\ldots\ldots\ldots.(15)$$

where $T_1(\alpha)$ = resistivity transform function which contains the required parameters. This is possible because the relation between apparent resistivity and resistivity transform functions is linear; Koefoed (1979).

In moving from apparent resistivity function to resistivity transform function (or vice versa) digital linear filtering is used. Samples of one function are taken at a constant interval along the abscissa and expressed linearly in the sample values of the other. The coefficients in this relation are called Filter Coefficients. The number of filter coefficients chosen depend on the computer time and the degree of accuracy required. This is the method behind the inversion process discussed in the next chapter.

# 4 INTERPRETATION OF RESISTIVITY SOUNDINGS

## 4.1 One-dimensional interpretation

### 4.1.1 Introduction

In most circumstances in resistivity geosurvey work 1-dimensional interpretation is done. The basic assumption here is that the earth is electrically homogeneous being composed of different layers whose resistivities change with depth. The raw data is first transformed into apparent resistivities by use of Equation 11 and then plotted on a double log paper. The apparent resistivity curve obtained is interpreted employing either the Auxiliary method or the Inversion process.

### 4.1.2 Auxiliary point method

The method requires a pre-calculated catalogue of master curves for the field parameters observed. The master curves are prepared in dimensionless coordinates from Equation 15. Keller et al, (1970) have described how the curve matching is done. By using these master and auxiliary curves the parameters can be determined thus obtaining a model that best fits the observed data plus being geologically and geoelectrically sound. Nowadays, however, it is rare that curve matching is used as the only and final form of interpretation. This stems from the fact that innumerable numbers of theoretical curves are needed for different geological and geophysical situations. The accuracy of the method is also not high enough because interpretation can never be wholly objective. However the method is useful as it presents a quick picture of the subsurface and the models obtained can act as starting models for the more accurate methods.

### 4.1.3 Inversion method

With the advent of computing facilities the inversion process has become widely used in geophysical data interpretation.

Inversion is an iterative process that is controlled by the mathematics originating from Equation 15. A model is proposed from the field data, fed into a computer and an inversion programme is run; the result of which is a new model whose parameters are compared to the field data. These parameters are altered and again run in the computer. This is repeated until a model is found whose calculated data best matches the field data.

Given a set of observations ($O_j$) and a theoretical earth model the accuracy of the approximation to the really model is given as a square error of E

$$E = \Sigma(O_j - C_j)^2$$

where $C_j$ is the calculated values for the model. The iterative process stops when E is less than a pre-determined value. Inman et al (1973) have shown that the inverse method is quite a powerful scheme. Ideal models were chosen to test the inversion method. They found out that noise had little effect on the well resolved parameters but the poorly resolved ones were in error by as much as 15%. They further noted that the uniqueness of the final model cannot be guaranteed as it may depend on the model at which the inversion was initiated. 1-D models are a good starting point for 2-D interpretation. Chapter 6 discusses how 1-D and 2-D were employed to interpret data from Krísuvík.

### 4.1.4 Features of curves

By definition the apparent resistivity measured by the Schlumberger array, $\beta_s$, equals the apparent resistivity, $\beta_a$, when the potential electrode separation is very small. Often in the field it becomes necessary to increase p as s becomes large to allow for more accurate measurement. The measured resistivity curve is composed of various segments which don't tie, i.e,

$$\rho_a(s,p) \neq \rho_s(s,p)$$

and the apparent resistivity curve becomes discontinuous. These shifts tend to cause problems in interpretation. Figure 3 shows a converging shift while Figure 4 is that of a nonconverging one. Converging shifts are caused first by the change in the ratio of p to s, usually referred to as eccentricity, and, second, by large resistivity contrasts between the horizontal layers. Hence the type of shift gives information about the resistivity structure (Arnason, 1984).

The nonconverging shifts are caused by lateral resistivity variations near the surface. This source of error can be minimized by using 2-D interpretation. The Ellipse programme at the NEA in Iceland has been developed to automatically correct for the constant shifts by use of the linear filter method. Often most field curves are a combination of the two phenomena (Figure 5). It then becomes necessary to interpret resistivity curves using both 1-D and 2-D techniques.

Another feature of apparent resistivity curves is that they cannot have slopes exceeding 45° on their ascending sections, though they may have values bigger than 45° at the descending sections. Consider a two layer model (Figure 6) where the uppermost layer has a finite resistance, say $\rho_1$, underlain by a much more resistive layer. If current is sent from the surface there will be a potential drop divV which will depend on the value of $\rho_1$. Now $\rho_1$ can, theoretically, take any values between 0 to $\infty$. That means that the descending slope can have any value between 0° & 90°. When s is comparable to the overburden thickness much of the current will flow along the boundary and the voltage drop reading will be constant. When s exceeds the boundary thickness

$$\rho_a \approx \rho_a(s).$$

The relationship between log apparent resistivity and log electrode spacing is linear, with a slope of 1. In fact it can be shown that

$$\rho_a = s\ \rho_1/t$$

where $\rho_1$ = resistivity of first layer

$t$ = its thickness

for any given two layers with the deeper one more resistive. This is so because in nearly every exploration problem, the lowermost layer reached by the current is an insulator, so that the portion of the sounding curve obtained with large s should approach the behavior predicted by the above relation. Taking two points (a,b) on the ascending part of the curve the following relation hold

$$\rho_a = s_a\ \rho_1/t \quad \text{and}$$
$$\rho_b = s_b\ \rho_1/t$$

Combining the two equations,

$$\rho_b/\rho_a = s_b/s_a$$

from which

$$(\log \rho_b - \log \rho_a)/(\log s_b - \log s_a) = 1$$

which is a slope of 45°. If the ascending slope is bigger than 45° then there is possibly a lateral variation in the resistivity or there is something wrong with the measurement which should be checked.

## 4.2 2-D interpretation

In most cases, soundings made in the field produce curves which are characterised by sharp bends or none-converging shifts. Using 1-D interpretation on such curves can bring erroneous results. This is because the basic assumption that the layers are horizontal, each with uniform resistivity is not met. This results from lateral variations in layer parameters, presence of dykes, or dipping layers. In such circumstances 2-dimensional interpretation has to be used. Knowledge of when to use 1-D or 2-D comes with experience. As already noted in section 4.1.3, 1-D interpretation is used as a starting point, and in some cases different sections of the curve are interpreted using one type alone.

Two dimensional interpretation is done with the aid of a computer. At the NEA, Iceland, a programme called Twodim is in use. This is a finite difference element modelling programme, FDM. The modelling is done by first constructing a rectangular grid. At the centre of the sounding the grid is equally spaced becoming widely spaced as one moves away in both x- and z-directions.

Once the grid is made the 2-D earth model is divided into blocks according to the grid. The blocks should have resistivities and thicknesses as inferred from pseudo-sections of 1-D. A file for the output of the calculated data is created and the programme Interpr is run. The Schlumberger apparent resistivity at different electrode spacings is calculated and stored in the output file. A calculated pseudo-section is plotted and compared manually with the measured one. The block parameters are adjusted by trial and error until a pseudo-section is obtained which best approximates to the measured data.
The determination of apparent resistivity values depends on the number of filter coefficients. A big size and a large number of coefficients increases the accuracy, however one has to compromise between accuracy and computer time which could be anywhere between 10 to 40 minutes.

## 4.3 Effects of deviations from the ideal earth picture

In section 3.1 the basic assumptions used in deriving the potential distribution on the surface of an ideal earth were outlined. In this section, some of the major features of a real earth and their effects will be discussed.

### (i) Near surface inhomogenity
This phenomenon is fairly common in geothermal areas. The distance that is traversed by the potential electrodes can be

composed of different resistivities, due to, say, roads, ditches, or pipelines. They will distort the readings. It is especially serious if AB/2 is comparable to MN/2. This error can be minimised by choosing suitable locations.

(ii) Topography

Ketsela (1984) has pointed out that topographic effects can be erroneous to sounding readings, by increasing or lowering their values depending on whether the feature is a hill or valley. This is usually overcome by including these effects in the 2-D modelling.

(iii) Equivalence

The depth penetrated by the current is usually determined by the maximum distance of current arm, s, which in practice is finite. On account of this and the inhomogenity problem coupled with finite accuracy of measurement, widely different resistivity distributions may lead to curves which, although they are not identical, cannot be distinguished from each other in practice. This produces ambiguity in the interpretations. For example, a thin layer sandwiched between two layers with much higher resistivities will make much of the current to flow through it. If the resistivity and thickness of the thin layer are, say, increased by the same proportion the amount of current flowing through it will remain unchanged. Hence the two cases will be electrically equivalent. This situation can be solved by using relevant available geological information.

(iv) Conductive overburden

In some situations measurement may be carried out on ground where the resistivity of the near surface structure is much smaller than below. This causes much of the current to concentrate in the overburden and may not reach the base rock and resolution becomes virtually impossible. To overcome this problem measurement may be made with two transverses using different electrode separations and interpretation done 2-dimensionally.

(v) Anisotropy

In reality, geological formations are far from isotropic; this is especially true of shale and clay formations. In clay, for example, the resistivity is the same in all directions along a layer but has a different value in the direction perpendicular to the stratification. It can be shown that if resistivity is measured in the field with an array oriented parallel to the bedding planes the measured resistivity is higher than the longitudinal resistivity by a ratio, $\alpha$ (Keller et al, 1970).

(vi) Dipping beds and vertical contacts

In practice the subsurface structures are not horizontally stratified. Mwangi (1982) has quoted resistivity curves obtained from sounding over dipping contacts. The effects of these phenomenon will depend on the size, and location relative to the sounding centre. In general, the mathematical formulation of this feature is enormous and studies have been limited to simple models. The effect on the resistivity curves are steep rising slopes and a discontinuity in the slope where one of the current electrodes crosses the outcrop at the boundary plane.

(vii) Natural noise

Telluric, thunderstorms and other natural electromagnetic phenomenon can cause the ground to develop induced currents, thus affecting resistivity measurements. Thunderstorms, for example, release energy which is propagated in a waveguide bounded by the ionosphere and the earth's surface. At any point on the earth's surface the measured noise includes this waveguide propagated energy plus atmospheric discharges from nearby sources. Such a feature is not easy to deal with; however, equipment with systems which only accept data within a prescribed range of amplitude can be employed. Unfortunately, the author couldn't come across any literature that has a discussion on these effects and their possible impact on electrical resistivity vertical sounding; this would have been of interest as Kenya is prone to severe thunderstorms, being in the equatorial region.

## 4.4 Conclusion

As a consequence of the above inherent problems it is hard to place any description on the uniqueness of a model of interpretation from a given sounding data. However the Schlumberger sounding method, though it has several limitations of its own, including the relatively slow progress with which work has to be done in the field and insensitivity to some lateral discontinuities, has been applied successively in many countries in the prospecting of geothermal energy.

# 5 CASE HISTORIES

## 5.1 Glerárdalur area, Eyjafjördur, N-Iceland

### 5.1.1 introduction

The Gleradalur geothermal field lies close to the town of Akureyri in N-Iceland. Though geoscientific studies in this area started as early as 1920 serious work didn't start until 1980. Part of the explanation could be that the results from drillholes sunk in up to 1965 turned out to be negative and more emphasis was put into studies in other areas, notably Laugaland and Reykir in Fnjoskadalur. Attention was brought once again to Gleradalur when production from Laugaland and other areas supplying Akureryi with warm water was being faced with drawdown problems. Studies have shown that a large fault across cuts the dyke where hot springs occur at the surface; so it is not the dyke itself but this fault which is the aquifer (Einarsson, 1986).

### 5.1.2 Geological setting

Björnsson and Saemundsson (1975) have given an overview of the geological picture in this region. The strata consists of Tertiary subaerial basaltic lava flows about 10 m.y. old. There are many dykes which form about 6% of the total volcanic mass. Relative age of the dyke swarm is not known. Most of the dykes in the area have a strike N 15° E; thicknesses vary between 1-10 M and dip 88-80° W. Figure 7 is a simplified geological structure of Glerargil. The country rock is cut by faults with direction N 0° to 10° E and most have been displaced 10-20 M. The dip of the faults is unknown; only one fault is exposed in the gully of river Glera. Much of the country rock is covered by moraine formed probably during the last ice-age (Einarsson, 1986). Since there were very few outcrops, resistivity measurements were resorted to.

### 5.1.3 Resistivity measurements

Many Schlumberger soundings have been made in the gully of Gleradalur (Flovenz, 1981). Some of the results were interpreted by Mwangi (1982). Correlation was then tried to be made between the low and high resistivities obtained, with the known dykes and faults. This was no easy task as the picture was complicated by the many dykes and faults crossing one another. For the first time in Iceland Headon profiling was used, results from it aided in modifying the model. The barriers with high resistivities seemed to correlate with faults which must act as aquicludes. Several drillholes were sunk into the low resistivity anomaly and these provided more information about reservoir conditions. They showed a high thermal gradient which indicates that the hot water follows a resistivity layer there. A barrier of high resistivity, probably a fault, lies in between the low resistivity areas - that with holes 5,6,7 and that with holes 8,10,12; and so no connection should be expected between these two systems (Flovenz et al, 1984).

### 5.1.4 Discussion

The aim of the resistivity measurements is to delineate low resistivity barriers as well as to determine resistivity changes which might be related to the country rock and possible aquifers. In Gleradalur, the geological picture is complicated by the many faults and dykes cutting one another. The Schlumberger sounding alone cannot resolve clearly such structures, especially the shallower ones. A combination with Headon will aid the interpretation greatly; this type of survey seems to be working very well in Iceland especially in detecting concealed faults and dykes. However its success will depend on several factors notably the ability to minimise equivalence.

## 5.2 Olkaria, Kenya

### 5.2.1 Introduction

Interest in the geothermal potential at Olkaria region has been indicated since the early 1950's. However a more ambitious programme of exploration and eventual exploitation started in the mid-seventies. The siting of both exploration and production wells was based on the interpretation of geophysical data and, in part, on structural and geochemical interpretation of surface manifestations.

Up to date, many geophysical methods have been used in geothermal survey at Olkaria. These include micro-earthquake studies, Roving dipole, M.T, Schlumberger sounding and Headon profiling; of all these the Schlumberger sounding has proved to be the most useful tool in understanding the geothermal system (Genzil, 1980).

### 5.2.2 Geological setting

A detailed geology of the area has been described by Naylor (1972). The Olkaria area is seen as the remnants of a caldera, now only recognised by the southern section of the peripheral pyroclastic cones. In the west, the caldera margin is indicated by a series of low hills. Periods of little or no volcanic activity have allowed the deposition of lake sediments. Two sets of fault lines are dominant in the area: an older set trending N-W related to the rift faulting and the other set is younger and trends NE to SW. Most of the geothermal manifestations are found along these recent volcanic faults.

### 5.2.3 Resistivity survey

Most of the geophysical methods have been tried out at Olkaria. However results from these surveys were often at such a contradiction from each other and the geology of the area that they were uninterpretable and hence abandoned. Bhogal

(1978) carried out a dipole-dipole survey but this time used the available Schlumberger sounding data to control the model. The results were very encouraging.

Many Schlumberger soundings have been carried out since 1972. Most of them provide sufficient penetration to examine the underlying geothermal system. The resistivity structure at Olkaria is dominated by two coherent low resistivity layers that underlie most of the field. The shallower layer (3-20 Ωm) lies within 300 M depth while the deeper one (≤12 Ωm) is at about 1 Km depth (Karta, 1984).

Headon profiling has been carried out in order to investigate the dips of the already known faults. Analysis of the Headon data was complicated by lateral resistivity variations which makes the use of this technique to locate faults difficult and imprecise (Mwangi, 1983). It was however recommended in the 1984 review meeting to continue with the method.

### 5.2.4 Discussion

The application of Schlumberger sounding method at Olkaria has produced tangible results. It should still be undertaken to extend research into the surrounding areas. Nevertheless it becomes apparent that the Schlumberger VES tend to be used as a constraint on most models. The failure of Headon to produce the expected results could be ascribed to a number of causes. First, there is a basic ambiguity in the technique with a number of different models producing similar fits to the measured data. Secondly, the sensitivity of the measured apparent resistivity to lateral near-surface inhomogenity is often of 3-dimensional nature. Further these near-surface noise tend to obscure deeper effects. Perhaps further application of the Headon method should be continued in order to reveal the apparent failure.

# 6 INTERPRETATION OF SCHLUMBERGER SOUNDINGS FROM KRÍSUVÍK

## 6.1 Introduction

Several high temperature areas are found along the Reykjanes Peninsula (Figure 8); the Krísuvík high temperature field is one of these. Up to date, there has been systematic exploration in this field involving geological, geochemical and geophysical studies, including the sinking of several deep wells. Surface manifestations cover an area of about 10-20 $Km^2$. The surface manifestations consists of hot ground where bedrock has been intensely altered by acid surface leaching. Steam vents and mud pools are common. Just like the rest of the Reykjanes Peninsula the ground water-table is shallow. This may be reflecting high permeability of the bedrock but not so much the local topography (Arnorsson et al, 1975). Information from resistivity survey reveal surface layers of high resistivity (over 10,000 $\Omega$m) which drop to less than 10 $\Omega$m in a few hundred metres depth. Boreholes show temperatures with inverse gradient with the highest temperature of 262°C. There is considerable discrepancy between measured temperatures and those from geochemical analysis. Various models have been proposed to try and explain the inverse temperature gradient. None of these, however, seems to be consistent with most of the observed characteristics of the geothermal field. The one currently held depicts a system gradually cooling from above and below of the reservoir by relatively fresh water which is replacing an originally more saline water.

Since the production characteristics of the field are not clearly understood and in order to determine the true model further research is deemed necessary. The author participated in the collection of data, using Schlumberger sounding method, from Krísuvík. Their treatment and possible interpretation is given below in sections 6.4 through 6.6.

## 6.2 Geological setting

The Reykjanes peninsula is the landward extension of the Mid-Atlantic Ridge. It is part of the zone which is shifting eastwards from the plate boundary. Rifting and transform fault characteristics can be seen. There are post-glacial lava fields with steep sided ridges which protrude through the lava fields. Basaltic rocks are predominant. There have been many volcanic eruptions in the Krísuvík area giving rise to volcanic edifices on fissures. The Krísuvík high temperature area is characterised by two major SW-striking hyaloclastite ridges. The valley between the two ridges is covered with lava flows; most of the geothermal activity is found within the hyaloclastite ridges and on their 'outer' sides but none in the lava covered valley between the two ridges. The fissure swarm that cuts the Krísuvík ridges is more intense at the ridges than at the valley. It appears, therefore, that there has been more tectonic activity on zones occupied by the ridges than at the young lava on the valley. It is possible that hot water rises to the surface under the ridges. This could explain the inverse thermal gradient but this has to be verified by drilling at the ridges. The thermal water in the Krísuvík area has high salt content, though not as high as at Reykjanes or Svartsengi. The sea water seems to percolate through the highly permeable bedrock and mixes with the thermal waters of the Krísuvík system.

## 6.3 Resistivity survey

Much of the Reykjanes Peninsula is generally characterised by low relief. In spite of this, surveying is often hampered by rough lava terrain with no soil or vegetation and only accessible on foot. Schlumberger soundings have been carried out to try and estimate the bedrock resistivity at different depths. Dipole-dipole profiling measurements with 1500 M centre spacing along the eastern ridge have been tried. Surface resistivities are quite high, decreasing rapidly with depth. Figure 9 (after Arnorsson, 1975) shows an area of about 70

$Km^2$ which encloses all the surface thermal manifestations where resistivity decrease to between 50-5 Ωm at a few hundred metres depth. Below 3 Km (from MT measurements) resistivity increases rapidly. Despite the results from drilling it is still not certain to what extent the low resistivity is due to temperature, porosity or salinity of the water.

## 6.4 Treatment of data from the field work

Figure 10 shows the location where the soundings were made. The position of the exploration wells is also marked. The line XY is the general direction along which the measurements were taken. The data was first fed into the computer and then modelled one-dimensionally, using the programme 'Ellipse'. The type of curves obtained are shown in Appendix 1. Many of the proposed parameters for the models of the various soundings appear to fit well with the measured data. However there are several exceptions, notably sounding number TD74 and TD75. This necessitated the application of two-dimensional interpretation using the programme 'Twodim' followed by the construction of both the measured and calculated apparent resistivity pseudo-sections (Figure 11). The results were then compared with previously collected data from the region.

## 6.5 Results from treatment of data

The pseudo-section of the computed apparent resistivity seem to match with that from the measured data. Further, comparing the results from 1-D and 2-D curves the agreement is close enough giving allowance, of course, for the limited accuracy imposed by the measurement and the modelling process. Hence, one can assume that the proposed model (Figure 12) reflects the geophysical conditions of this region.

Soundings TD74 and TD75 were the most difficult to fit into the profile. It became desirable to include a thin vertical block of low resistivity between TD74 and TD83. This made some improvement, but not enough. To avoid great contrasts, between the various column blocks intermediate resistivities were introduced between sounding TD84 and TD87, similarly between TD76 and TD81. All the time, it was kept in mind that the model had to be as simple as possible but yet close enough to the expected geophysical picture of the region.

Several prominent features can be seen on the model. Generally, the resistivities near the surface at both ends of the profile tend to be high. This is especially true of the first 100-200 m depth. In the middle of the profile the resistivities are low (≤50 Ωm). In all soundings, except TD81 and TD82, the resistivities decrease rapidly with depth, going as low as 2 Ωm within a couple of hundred metres depth. Under TD82 the resistivity remains fairly constant with depth, around 30 Ωm while the adjacent ones have an average low of 4 Ωm. Though resistivity under TD81 decreases quite rapidly with depth its values reach only 50 Ωm at 1000 M depth while others at the same depth are less than 15 Ωm.

## 6.6 Discussion

As already discussed earlier on, interpretation of resistivity measurements has that inherent characteristic of equivalence. Unless one has other extra information this phenomenon can be of great significance. In this discussion an effort will be made to correlate the model with the already available data from earlier works. The high surface resistivity under TD81 and TD84 are due to post-glacial lava fields. By contrast, the low near surface resistivity values in the middle of the profile could be due to alteration from surface leaching. The water-table at Reykjanes peninsula is generally shallow. This could explain the rapid decrease in resistivity values in just a few hundred metres, especially when the water has high salt content. There is a big fumarole, Austurengjahver, next to TD75. As expected, the sounding point is underlain by a low resistive structure which is 'sandwiched' between relatively high resistive ones under TD82 and TD76, down to about 700 M deep; beyond that, the low resistive structure spreads eastwards.

From the geological map (Appendix 2) the sounding TD75 and the fumarole are situated within a fault system. Hence one could suggest that the conduit for the geothermal fluids here is the fault, which tends to extend eastwards with depth. This could also explain why modelling for TD75 could not fit into the profile properly.

There is another surface manifestation between TD74 and TD83. Here, mud pools and steam vents exist. In order to account for the conduit for the fluids and in order to try and match the 1-D and 2-D models a dyke of low resistivity, 2 Ωm, was inserted. The improvement was only slight, hence it is probable that the high resistivity barrier between 50 and 100 M deep below TD74 is an impermeable structure through which the water can not percolate.
The general trend of the resistivity anomaly is a high surface resistivity which decreases to less than 10 Ωm at about 100-200 M depth and gradually increases again. This trend may be explained by the inverse temperature gradient. From present drill-hole data it is hard to judge if the low resistivity anomalies are due to temperature of the geothermal fluid or porosity. Nevertheless, the model parameters are comparable to those values obtained by Arnorsson et al, 1975.

## 6.7 Conclusions

This work has demonstrated that using both 1-D and 2-D modelling encouraging results can be obtained. It is important, however, to distinguish between low resistivity anomaly due to low surface resistivity layer and low resistivity layer at depth. This holds true especially when one moves from 1-D to 2-D interpretation. The resistivity pattern at depth in the Krísuvík area shows a low resistivity zone, less than 10 Ωm in the middle regions of the area studied. The geothermal manifestations tend to be associated with regions of low resistivity. Though surface resistivities may be high, these values decrease rapidly with depth; the shallow water-table playing a substantial role in this decrease. The conduits for the geothermal fluids appearing on the surface can be both dykes or faults. The inverse temperature gradient can be inferred from the resistivity anomaly pattern. Studies already done don't provide enough information to indicate if the geothermal field at Krísuvík could be a 'dying' source of heat or not. Further investigations are still necessary to fully understand the geothermal potential of the area.

## 6.8 Recommendations

More research should be put on defining the reservoir model which will hopefully explain the inverse temperature gradient which in turn will cast some light on how much the low resistivity is a function of temperature, salinity or porosity.

ACKNOWLEDGEMENTS

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Kærar þakkir.

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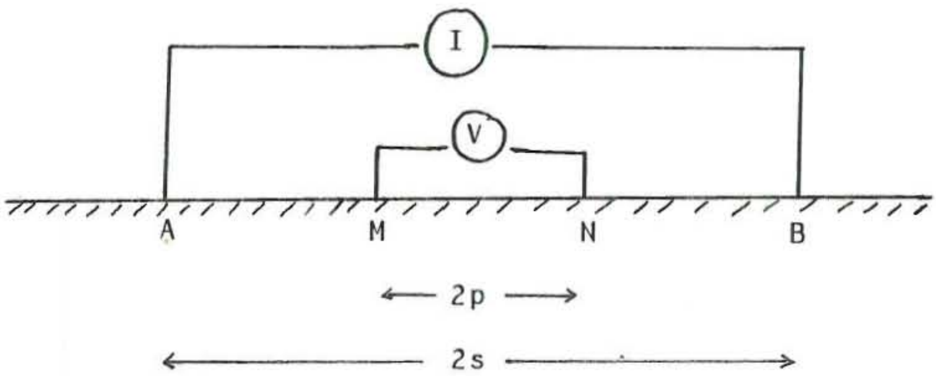



Figure 1: The Schlumberger array configuration

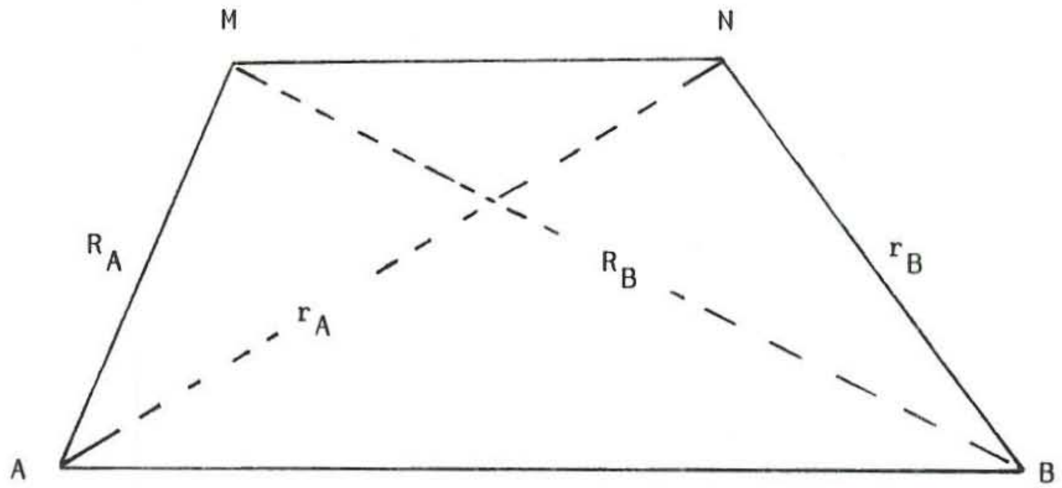



Figure 2: General four electrode array

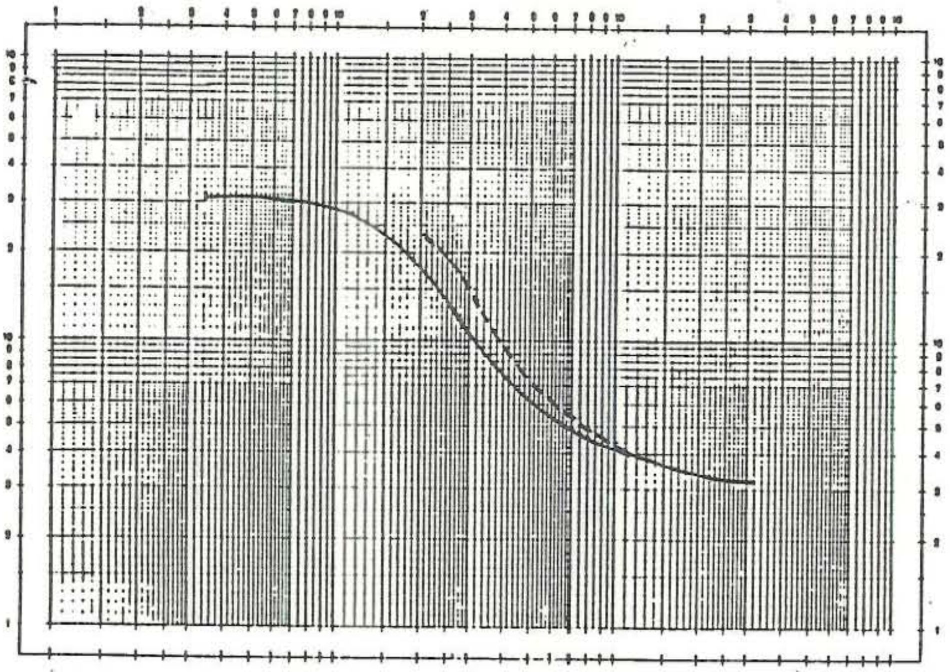


Figure 3: Converging shift in apparent resistivity curve

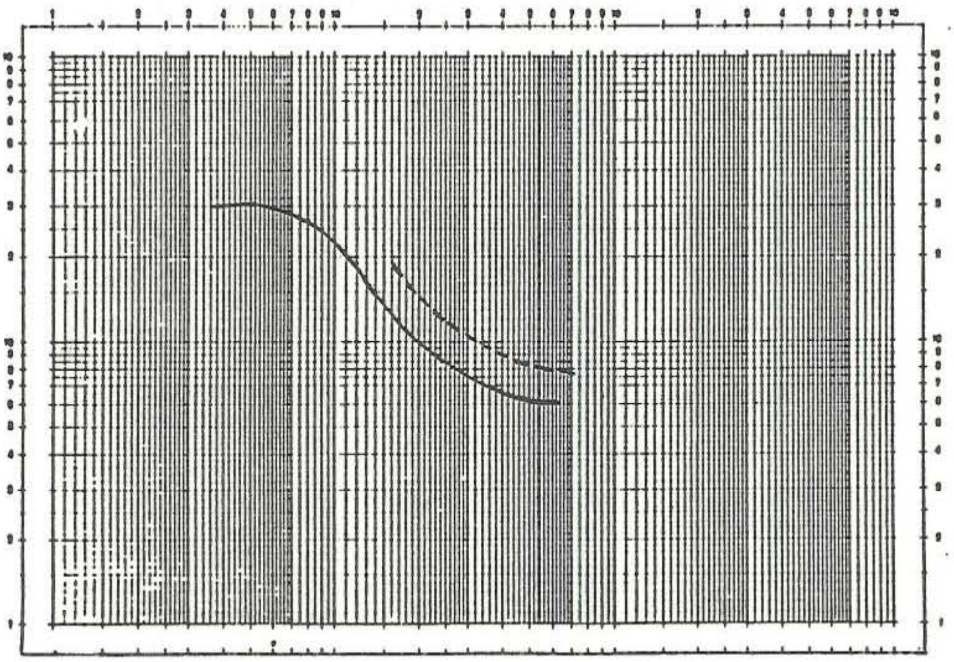


Figure 4: Nonconverging shift in apparent resistivity curve

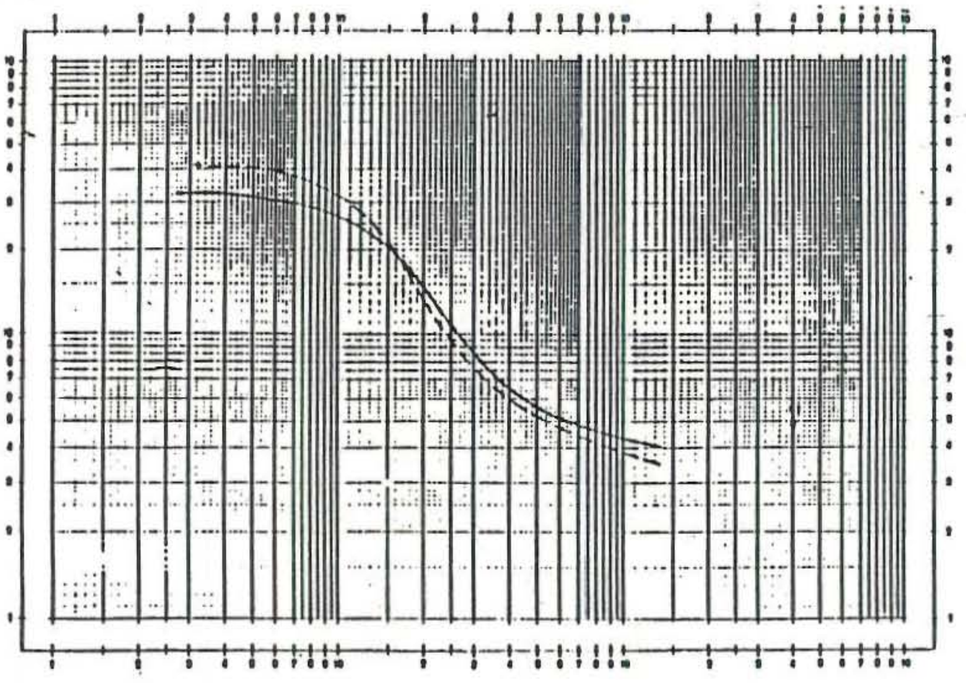


Figure 5: Combination of converging and nonconverging shifts in apparent resistivity curve.

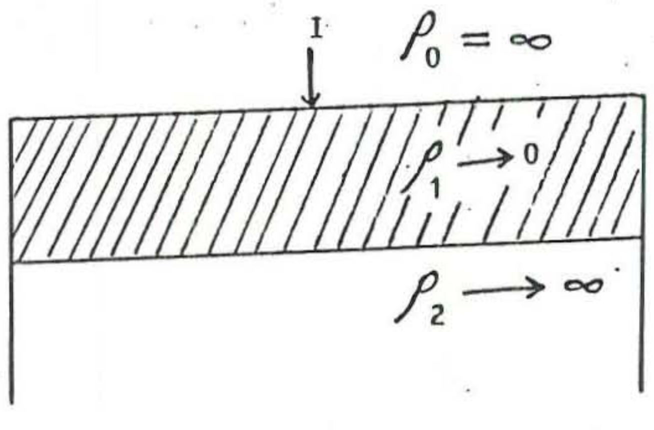



Figure 6: Single overburden problem, upper layer conductive.

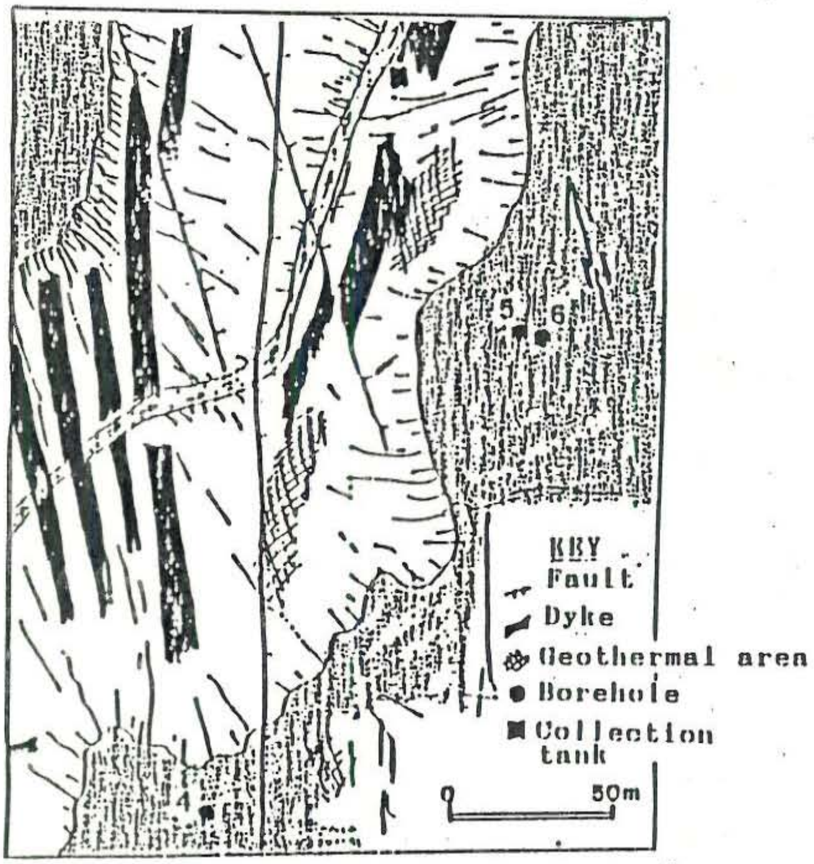



Figure 7: A simplified geological picture of Glerargil (Flovenz, 1984)

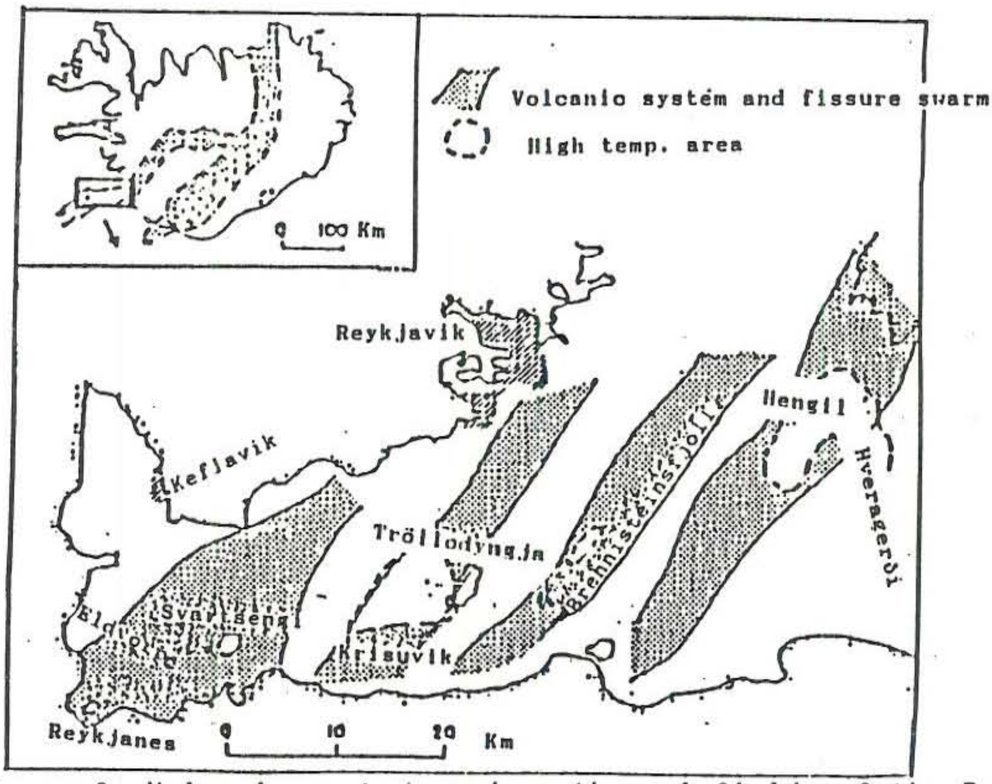



Figure 8: Volcanic systems and geothermal fields of the Reykjanes Peninsular. (After Georgsson, 1984)

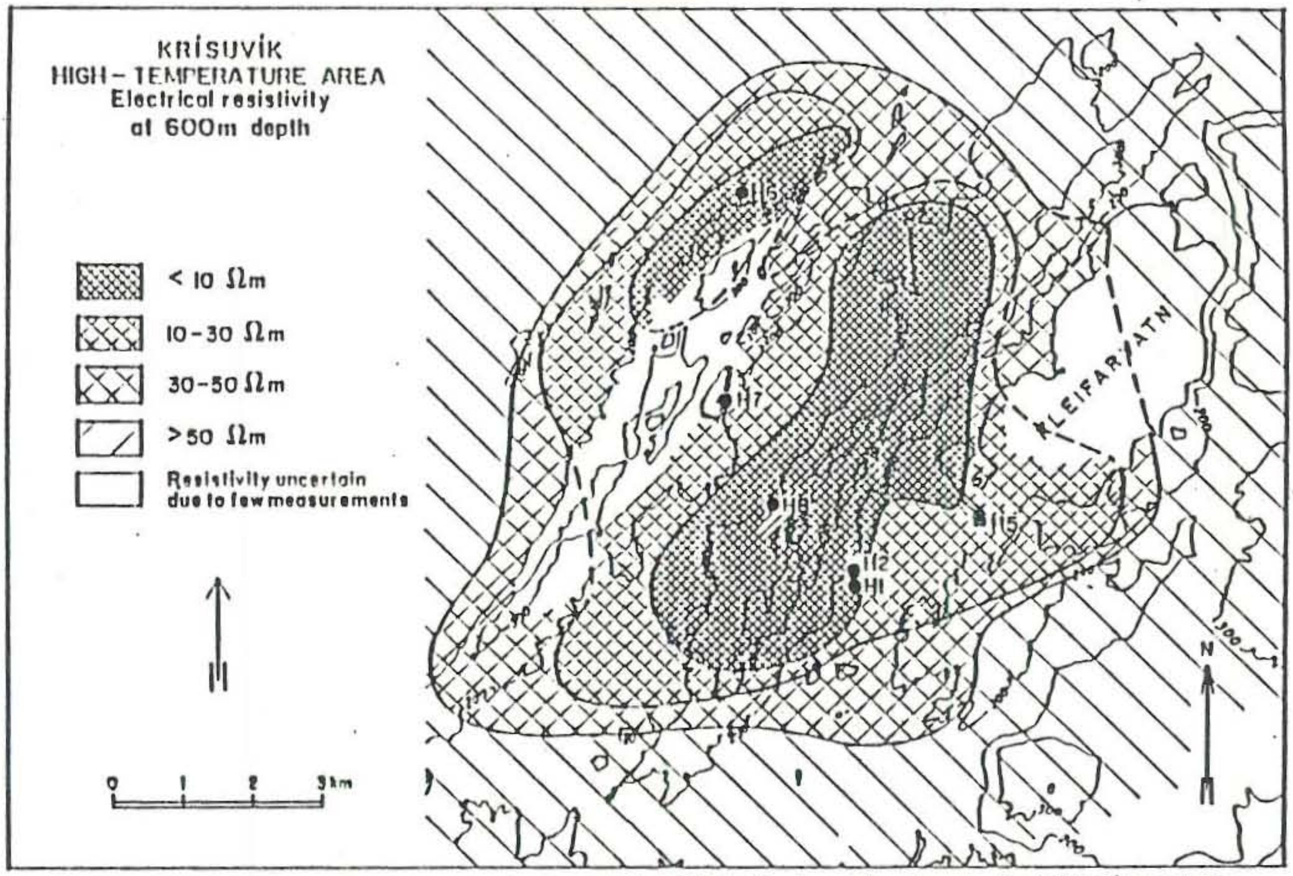



Figure 9: Apparent resistivity at 600 M depth from Schlimberger configuration. Location of wells are also shown. (After Arnórsson, 1975)

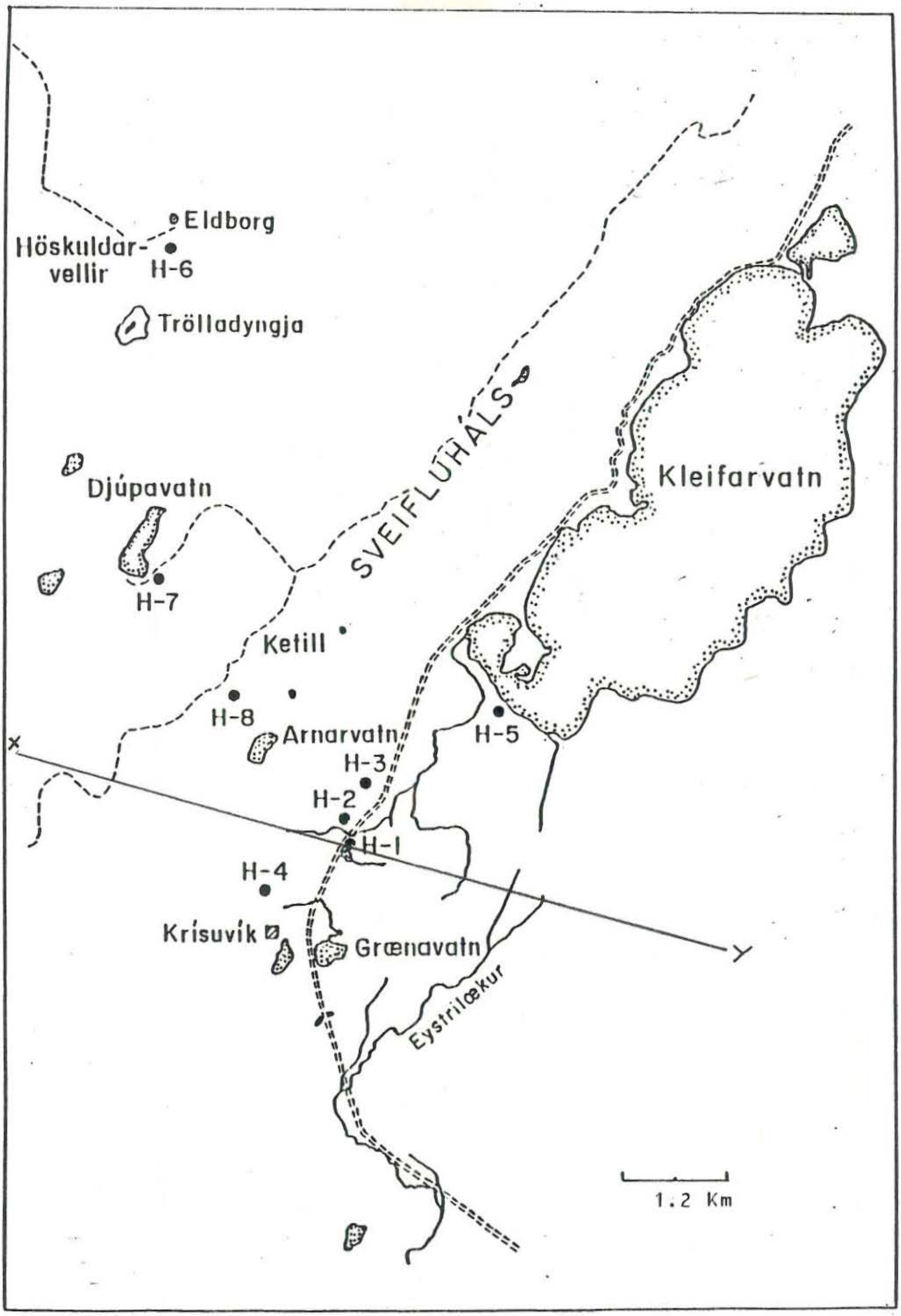
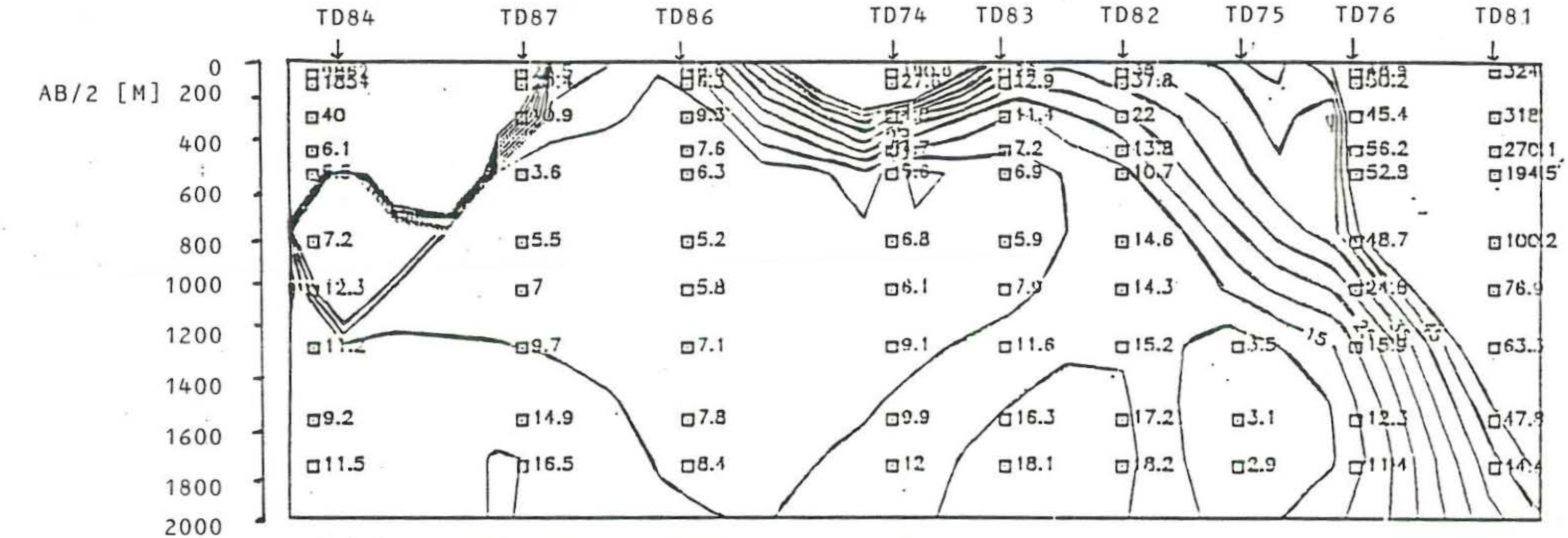
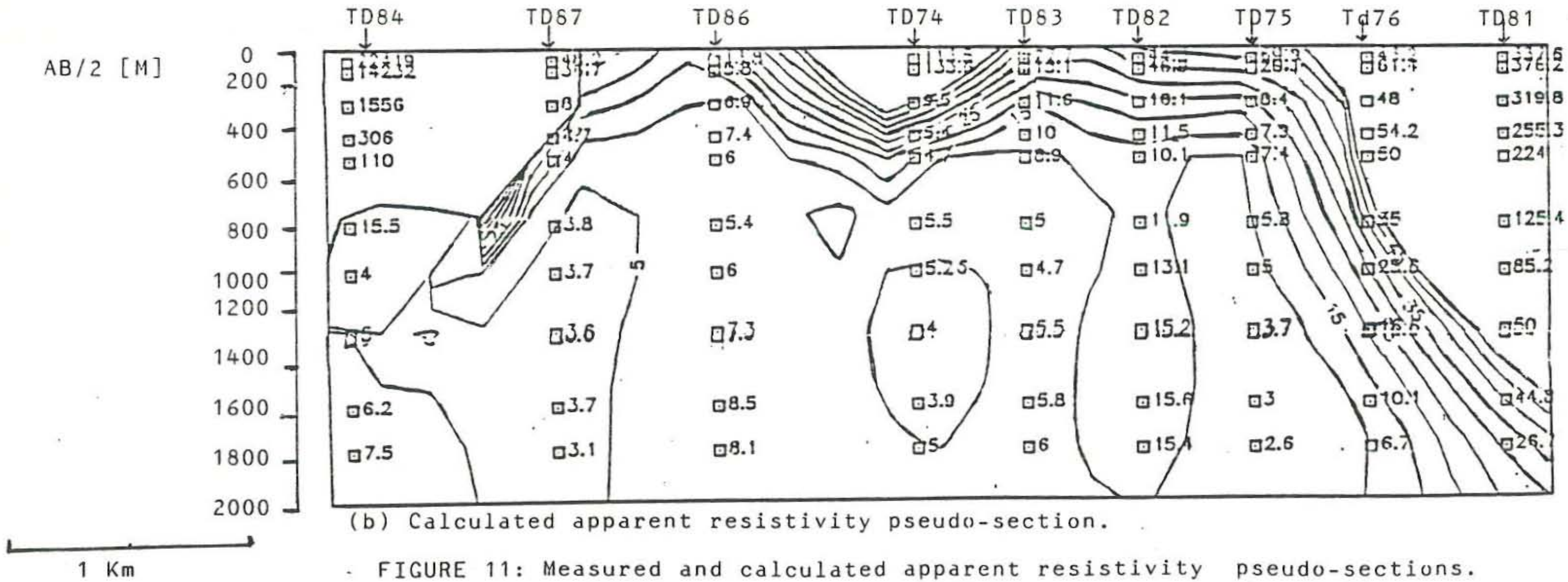



Figure 10: Sounding line




(a) Measured apparent resistivity pseudo-section




(b) Calculated apparent resistivity pseudo-section.

FIGURE 11: Measured and calculated apparent resistivity pseudo-sections.

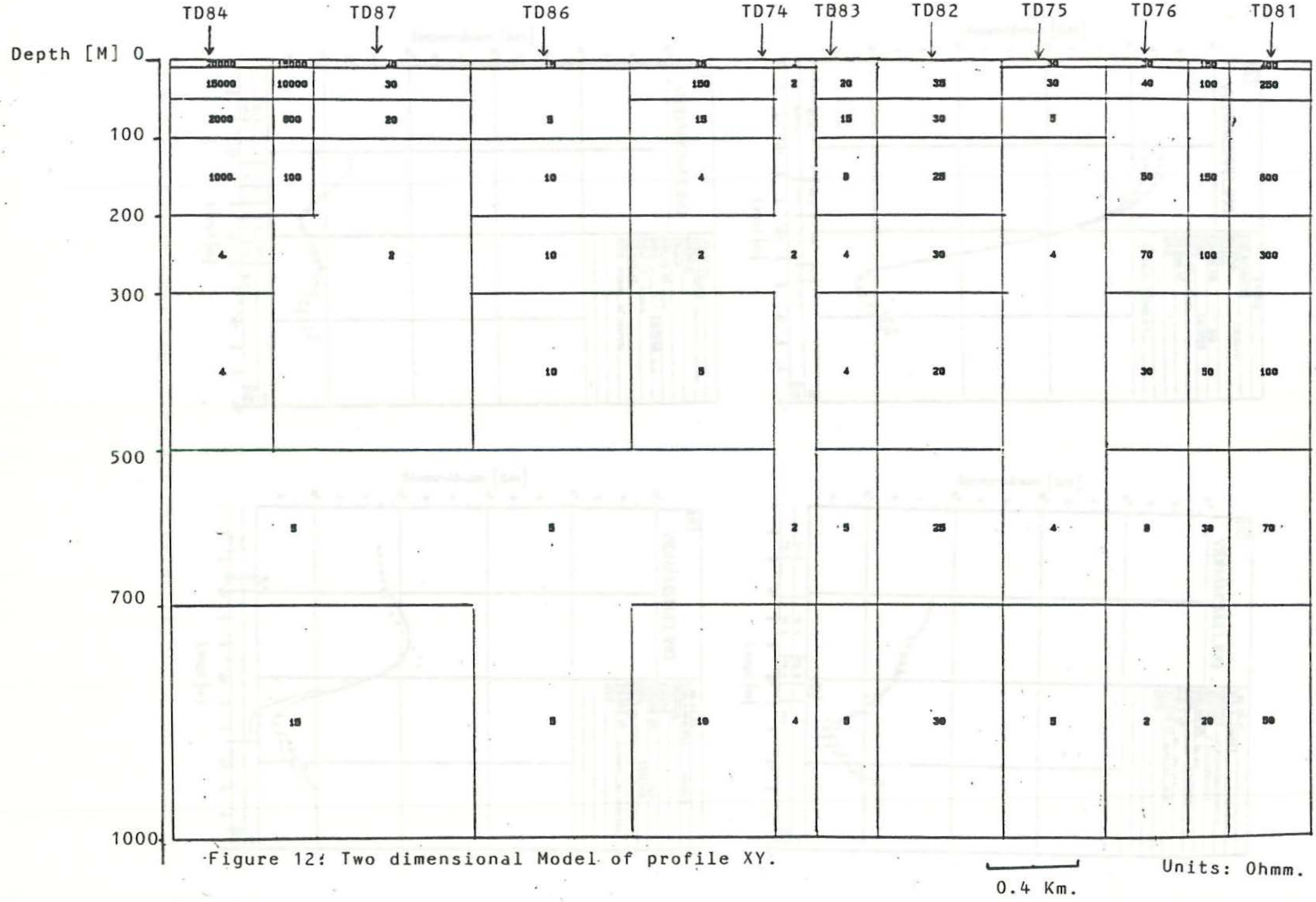
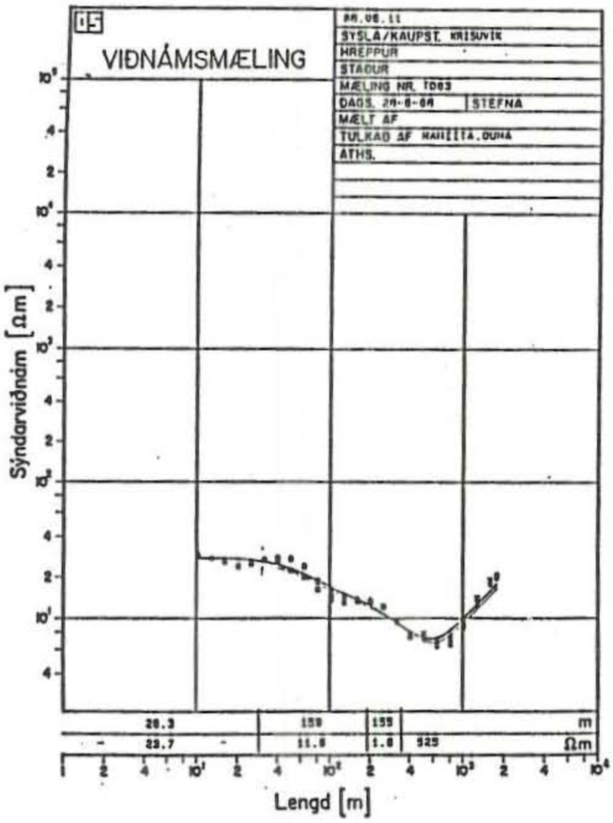



Figure 12: Two dimensional Model of profile XY.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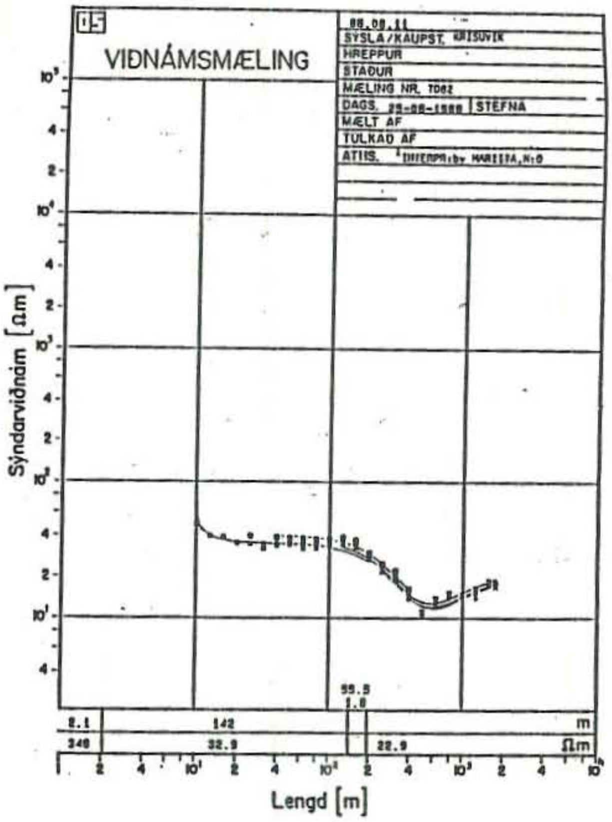
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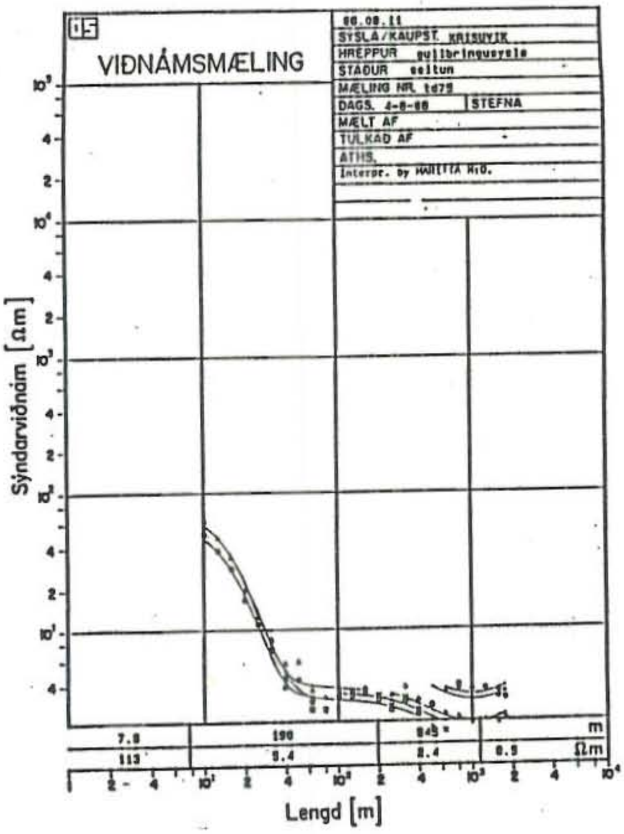
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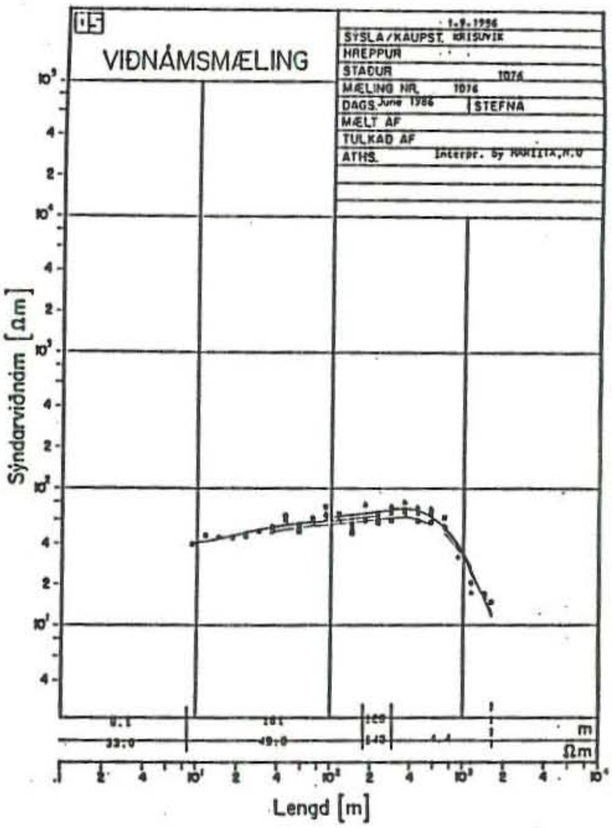
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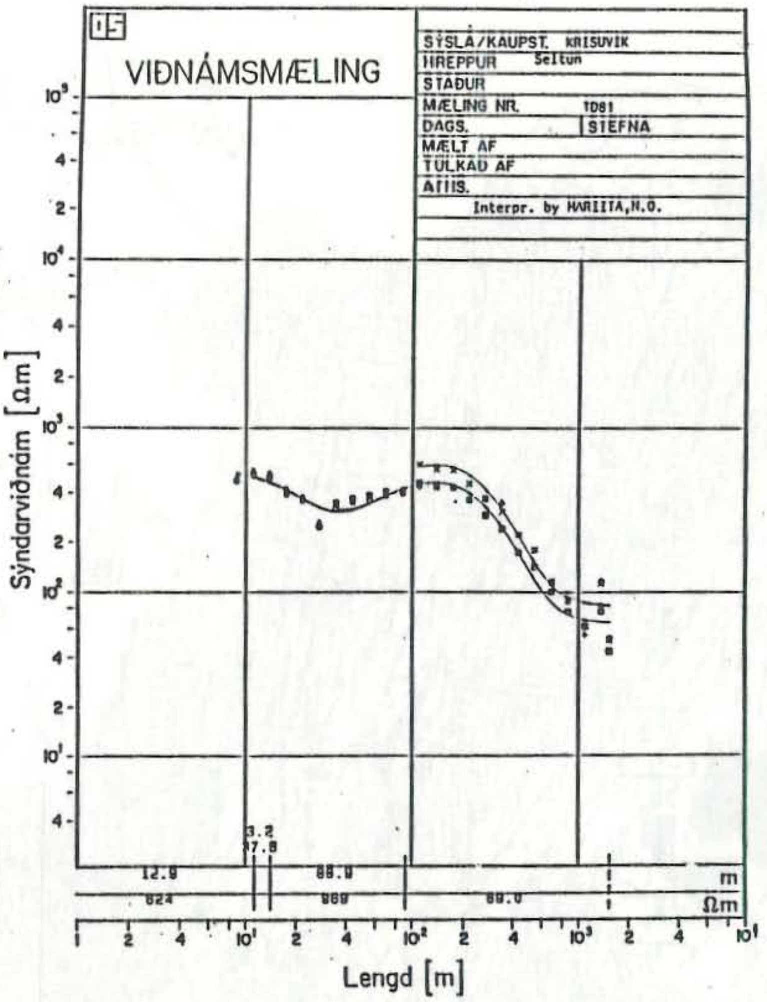
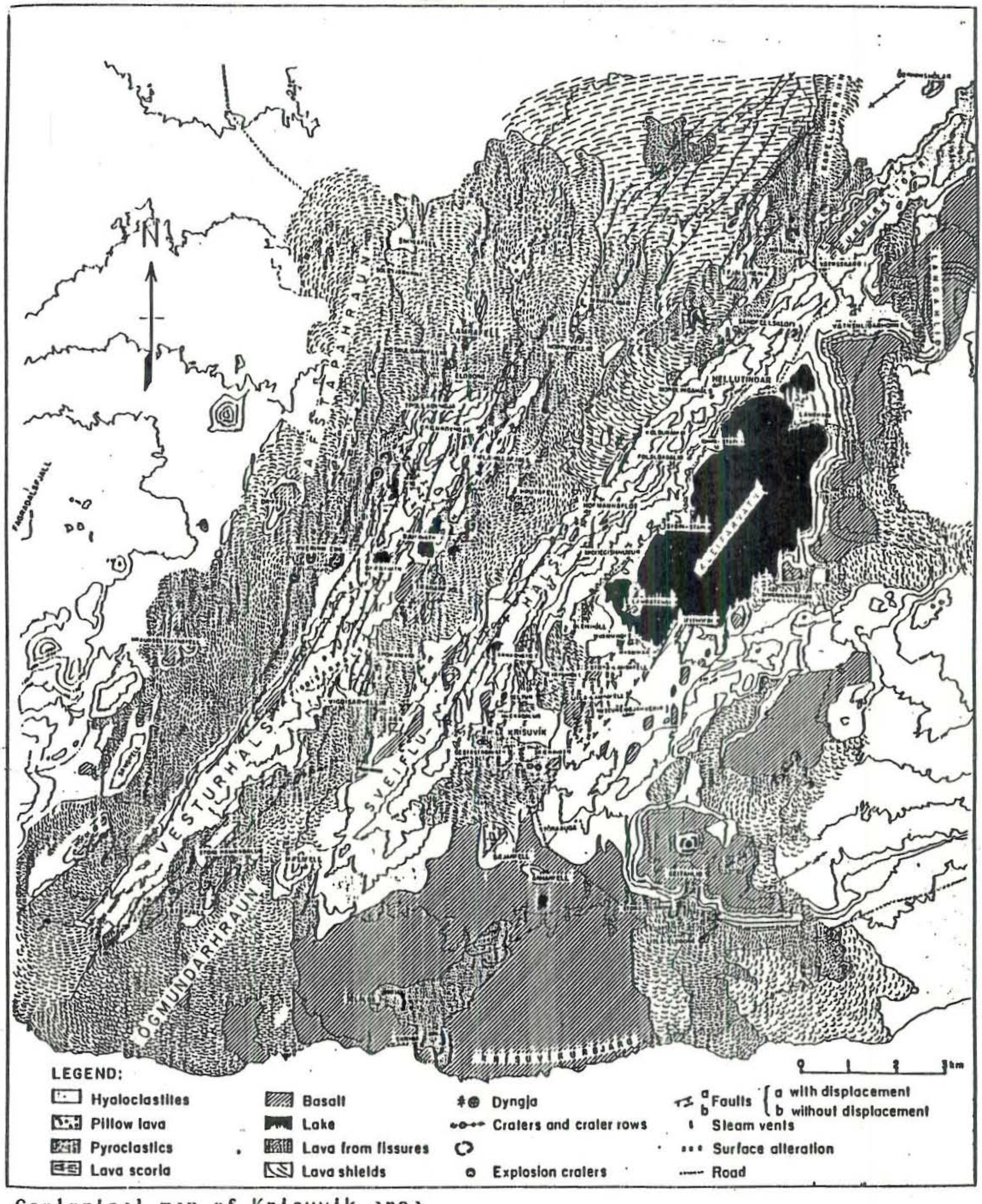
APPENDIX 1 (continued)






Geological map of Krisuvik area