Methods of Determining Aquifer Storage Capacity and Fresh-Saline Water Interfaces by Geoelectrical Investigations

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ABSTRACT

Geoelectrical investigations in Grundy County of northwestern Missouri where the groundwater resources of the glacial deposits have already been examined through an extensive drilling program by the Missouri Geological Survey and Water Resources, indicate that water-bearing gravel deposits can be distinguished from glacial deposits containing appreciable amounts of clay and limited amounts of water. The Schlumberger method used for the geoelectric depth soundings in the vicinity of the Survey's drillholes demonstrates the exploratory usefulness of the method in that it can partly replace the more expensive procedure of drilling. The method also provides improved interpretation between drillholes.

Results of the investigation show that, in the area, clay has a resistivity below 20Ωm, that the fresh water-bearing gravel at the bottom of the buried glacial stream channels has a resisitivity of 40 to 50Ωm, and that the near surface glacial gravel deposits have a resistivity above 100Ωm. Interpretation of the depth soundings and the conductivity of water obtained from a local well implies that its water is drawn from the saline water of the bedrock. A recommendation is made for the quality improvement of this particular well.

Keywords: Geoelectrics, Resistivity Method, Groundwater Aquifer, Glacial Stream Channels, Fresh-Salt Water Interface.

Table of Contents

1. Introduction

Industrialization and improvement of the standard of living requires an ever increasing supply of fresh water. Competition for the water between communities and industries together with impoundment for the purpose of conservation and pollution control has an important impact on community planning as well as on agricultural and industrial expansion. For such planning and expansion important questions have to be answered in regard to the amount of water available, its cost of production, and quality. Questions of aquifer recharge and possible pollution from salt water encroachment and infiltration of pollutants from the surface also have to be answered with a certain degree of accuracy and reliability.

Reliable geological, physical, and chemical data are necessary for the proper evaluation of aquifers for water use. These data conventionally are obtained from exploratory drillholes and existing waterwells. Deep aquifers, especially in a multilayered sequence have become important sources of water because of their greater capacity. With the ever increasing depth required for the exploration of these aquifers, the cost per drillhole is likewise increasing. Moreover, if larger amounts of water are needed, the information on the aquifers has to be more accurate because of the necessity for more intensive planning. Consequently, a denser net of drillhole observations is required and this also raises the costs of exploration. It is, therefore, important to look for cheaper methods which can give accurate data at costs lower than is presently required for drilling programs. A combination of geophysical investigations and drillhole observations seems to be the most appropriate approach for lowering present exploration costs because the geophysical methods can substantially reduce the required number of drillholes for determining an aquifer's location and production potential.

2. Measuring and Interpretation Methods of Geoelectrical Depth Sounding:

A geophysical method, which is becoming increasingly popular for the solution of hydrogeological problems is the geoelectric depth sounding method, because its instrumental requirements are comparatively modest and inexpensive. In this method, an electrical current is fed into the soil and the voltage is measured between two electrodes. Successive measurements are made over the same center while the current electrode separations are increased. Thus the electric current is forced to penetrate deeper into the ground. The recorded currents, voltages, and electrode configurations make it possible to calculate apparent resistivities, ρ_a, which are functions of true resistivities and the geometrical boundaries of different materials. The method reveals the change of resistivity with depth.

There are almost as many different opinions concerning successes and/or failures of geoelectrical resistivity methods as there are users. The reason is that in spite of the low cost of the equipment the problems of interpretation are among the most difficult in geophysics. Although the methods have been used for a long time, it has only been in the last 10 to 15 years that some fundamental contributions have been made with respect to interpretations. Geoelectrical resistivity measurements respond to lateral as well as vertical layers of rock resistivity. Theoretical interpretation methods are restricted to either lateral or vertical changes of resistivity. Any subsurface conditions, which favor both types of inhomogeneities at equal intensity, are unsuitable for geoelectrical resistivity surveys. In the case of horizontal bedding, it is possible to calculate theoretically ρ_a as a function of the electrode configuration if a distinct set of n layers with arbitrary thicknesses, m_i, and resistivities, ρ_i, are assumed (Stefanescu, 1930).

The simplest calculation is required for the horizontal, two-layer case. For this problem, two sets of master curves are sufficient to interpret depth sounding results by curve matching. However, for a three-layer problem, the number of master curves that is necessary for accurate curve matching varies between 500 and more than 1000 in accordance to the scope of the field curves. The computation of n-layer depth sounding curves requires sizeable computers and extensive programming facilities which are available only to a few users of the depth sounding method.

Satisfactory master curve catalogs have only lately become available and are mainly restricted to three-layer cases (Orrelana and Mooney, 1966). Hummel's method of approximation allows a multilayer case to be interpreted with two- or three-layer master curves. The first two layers are interpreted and combined to a ficticious layer. The ficticious layer and the true third-layer are treated as a two-layer case and interpreted with respect to depth and resistivity. One can finally combine n-1 layers to a ficticious layer and interpret the nth layer. This approximation method is satisfactory if the deeper layers are successively thicker and if there is only moderate contrast between successive layer resistivities. Even if multilayer curves can be calculated, very different resistivity depth functions can result in the same resistivity curve in certain cases. Also, certain layer combinations cannot be differentiated sufficiently. This is a general feature of the interpretation methods which are based on potential theory.

Concerning the electrode arrangement, it is only lately that one has a better understanding of the application of different arrangements and their implication to the results. (Deppermann, 1954; Kunetz, 1966; Frohlich, 1968).

For hydrogeological exploration, the resistivity of a layer is a complicated function of water saturation, water salinity, clay component, porosity, and of the matrix of pore channels. Even if the composition of the rock material is known, generally it is not possible to predict its resistivity; Therefore, one can decide only on the applicability of the depth sounding method for a specific geological problem if test soundings are made and compared with other evidence such as is provided by drilling.

3. The Hydrogeology of Northwestern Missouri

A rather complicated hydrogeological situation in the northwestern part of Missouri has provided a good test for the application of geoelectrical depth soundings. In this area, there is a comparatively large amount of water in the Paleozoic bedrock (Fuller and Knight, 1967), but it is too saline for drinking purposes. The salinity values range between 1500 to 2000 ppm of dissolved ions, and wherever it is used for drinking, it must be treated. The bedrock in the area is overlain by glacial deposits composed of sand, gravel, and clay. In most of the buried glacial stream channels and also in areas where there are thick glacial deposits, the basal units in immediate contact with the bedrock, contain fresh water. These units are composed of sand and gravel and generally lack clay. Nearer the surface, generally the size of the sand and gravel is smaller and the amount of clay is greater. Some gravel and sand pockets near the surface are limited in extent and are used for local water supplies by farmers. An extensive drilling program has been conducted in this area by the Missouri Geological Survey and Water Resources, and the drillhole records provide excellent comparative material for geoelectrical depth soundings.

The water in the stratigraphic complex of Grundy County is concentrated in aquifers at three different depths. At the bottom, in the upper part of the bedrock is salt water. This is overlain, at the bottom of the glacial drift, by water which has a much lower salinity and is in most cases potable. Near the upper surface of the glacial deposits, scattered pockets of gravel bear limited amounts of fresh water. The purpose of the geoelectrical investigation was to determine how accurately the aquifers at the different depths could be defined, with the middle aquifer at the lower part of the glacial drift being of particular interest because of its potential for yielding large quantities of potable water.

4. Method of Investigation and Instrumentation

The most frequently applied electrode arrangements are those of Wenner and Schlumberger. In both, the outer current electrodes, A and B (Fig. 1) are, after subsequent measurements, expanded symmetrically with respect to the center. In the Wenner method, the potential electrodes are likewise expanded so that their separation is always one third of A B. In the Schlumberger method, the potential electrodes are kept as close together as possible and are only once or twice expanded during a depth sounding to guarantee a sufficiently accurate voltage measurement.

The main advantage of the Wenner method is the greater accuracy of the measured voltage, which is proportional to the potential electrode separation M N. A comparatively large voltage between M and N is especially desirable if the potential electrodes are of low quality, a condition which may cause electrochemical voltages. Normally, nonpolarizable electrodes are used. These consist of copper rods dipped into a saturated copper sulfate solution which penetrates the soil through a porous pot. Often, however, for shallow surveys, the potential electrodes are replaced by metal electrodes, which, predominantly in wet and ion rich soil, give rise to erroneous contact voltages.

Because one has an advantage of choosing a good pair of nonpolarizable electrodes and because high precision and sufficiently sensitive electronic voltmeters are available, the advantages of using the Wenner method are insignificant unless the cost of instruments has to be kept very low. Therefore, the Schlumberger method has gained more interest because it does offer a number of valuable advantages. A very obvious advantage is the reduction of effort expended in field work because the electrodes M and N have to be expanded only once or twice. The greatest advantage, however, is the sensitivity to near surface lateral inhomogeneities of resistivity. Deppermann (1954) proved theoretically that lateral inhomogeneities will affect the Wenner method more than the Schlumberger method. Numerous examples show that measured values of apparent resistivity are widely scattered in the case of the Wenner method, whereas, Schlumberger values lie more nearly on a smooth curve that represents change of resistivity with depth. In the same area of the present study, in northwestern Missouri, earlier geoelectrical investigations had been made with the Wenner method by Meidav (1960). His results showed scattered values and were found to be insufficient for a partial replacement of the drilling program. Figure 2 shows the comparison of a depth sounding after Schlumberger and Wenner (measured by Meidav 1960). In the present study in which Schlumberger depth soundings were used, the influence of lateral changes of resistivity was ineffective.

The instrumentation for Schlumberger depth soundings has to be somewhat more sophisticated than for the Wenner method, especially if large electrode separations are planned for greater depth penetrations. Figure 3 shows the schematic of a unit, which is dimensioned for maximum electrode separations of AB $\simeq$ 800m. The current for this unit is provided by a converter at a maximum of 0.25 amps at 400 V dc. The converter is driven by a 12 V battery. A number of variable and constant resistors can regulate the current, which, while being measured with an ammeter, is fed into the ground. The current can be reversed to avoid corrosion of the electrodes by electrolytic processes. The voltage between M and N is tapped with nonpolarizable copper-copper sulfate electrodes and is measured with a dc Null Voltmeter, Hewlett Packard Model 419A. The instrument indicates zero at the center so that the needle shows in the other direction when the current is reversed. There are a total of 18 sensitivity-ranges, and the most sensitive range at which the voltage is measure of is 100 μV at full scale. At the small electrode separation, MN of 3 m, any noise from variations of the telluric field and of stray industrial currents is comparatively small because their noise level increases with the length of the electrode separation.

5. Presentation and Interpretation of Results

Figure 4 shows the scope of the investigation in Grundy County, where a total of 28 depth soundings were made. Most of the stations for the depth soundings were placed near the drillholes made for the Missouri Geological Survey's earlier investigation. This was done so that the resistivity boundaries could be compared with the changes of rock materials as recorded from the drill cuttings.

Figure 5 shows the results of a depth sounding with ρ_a plotted versus L/2, the half electrode separation of AB. Both values are plotted on a logarithmic scale. The dots are on a smooth curve which represents the resistivity curve. There is no abrupt change of ρ_a versus L/2, and Hummel's method of approximation with two-layer master curves can be demonstrated. Curve matching starts at the left-hand side of the curve for the first two layers. Beginning at A, the thin curve is the two-layer matching curve. The resistivity of the second layer is 0.67 times the resistivity of the first layer. The first and second layers have the same effect as one ficticious layer, whose thickness and resistivity is determined by a point on the auxiliary curve represented by the dotted curve. While the origin of the two-layer master curve is kept on this auxiliary curve, a match is attempted with the ascending part of the field curve, which indicates the third layer. The match defines point B as origin, and this indicates the thickness and resistivity of the ficticious layer. The third layer at a depth of 2.1m and a resistivity of 9 Ωm has the same effect as the two upper layers. This procedure is iteratively repeated, and points C and D give the depth of successive layers with true resistivities being found from curve matching. The resistivity depth function is shown in Figure 5 below the field curve, where the scale of L/2 is used as depth and compared with the drilling. At the top of Figure 5, the thicknesses and ficticious resistivities of the combined effect of the first two, three, and finally four layers are shown.

By not regarding the near surface resistivities of 12 and 7.7 Ωm, one can distinguish a sandy yellow clay of 15.5 Ωm from a sandy gray clay of 12.6 Ωm. At a depth of about 60m, the clay component is no longer evident, and a coarse sand of 47.5 Ωm is encountered. The absolute resistivity values of yellow and gray clays vary in accordance to different proportions of sand. In most cases, however, the gray clay shows a slightly but remarkably lower resistivity than the yellow clay. The yellow clay owes its color to a bleaching process by weathering. The resistivity of the coarse sand is generally four times larger than that of the overburden and can easily be determined with the depth sounding method. Whenever coarse sand and gravel are well developed, their resistivities are between 40 and 50 Ωm. If fine sand is intermixed, lower values between 25 and 35 Ωm are found. Drillholes have confirmed that the coarse sand bodies are excellent aquifers of fresh water.

Surprisingly no resistivity interface was found between the glacial deposits and bedrock where the drillholes had reached bedrock. Whenever the electrode separation was large enough to guarantee sufficient current-penetration into the bedrock, an increase of resistivity was found to occur below the top of the bedrock. This can be seen in Figure 6 where depth sounding No. 3 did not show higher resistivities in the glacial deposits. This agrees with drillhole No. 4, which did not penetrate sand or gravel at the bottom. The increase to higher resistivity from 15 to 55 Ωm occurs between 75 and 85m, which is in the bedrock. Because the upper part of the bedrock frequently contains fissures and cracks which are filled with salt water, one can explain why the upper part of the saline water bearing bedrock has a similar formation resistivity as the fresh water aquifer at the bottom of the glacial drift. The layer resistivity, also known as formation resistivity, decreases both with an increase of porosity as well as with the saline water filling the pore volume. In the fresh water aquifer, the water has low salinity and the deposit has a high porosity, whereas, the upper part of the bedrock has a low porosity but the contained water has a high salinity. This condition results in almost equal formation resistivities for both layers. The increase of resistivity deeper in the bedrock can be explained by a gradual narrowing of the fissures which provides less space for the saltwater at increasing depth.

Two depth soundings are shown in Figure 7 to demonstrate the principle of equivalence. If a low resistant layer is sandwiched between highly resistant beds, the resistivity curve depends on the "horizontal conductivity" of that layer.

This is the quotient m/ρ, m being the layer's thickness, and ρ is arbitrary as long as m/ρ is constant. In the upper part of Figure 7, a minimum of the depth sounding curve No. 23 with $m/\rho = 3.87$ is shown. Underneath the curve are a number of alternate layer cases, all of which can be solutions of the resistivity curve. On the other hand, if a highly resistant layer is sandwiched between less resistant beds (D.S. 28 of Fig. 7) as in the case of a near surface gravel layer, the "transverse resistance" of this layer determines the depth sounding curve. In this case, within limits, the product $m\rho$ must be constant so that again the values of m and ρ need not be specifically defined. In both cases, the ambiguity increases as the resistivity of the intermediate layer approaches the resistivity of the layers above and below. To determine the true resistivity of the intermediate layer, it is necessary to compare depth soundings with the data from drillholes or with the results of other geophysical investigations.

The results of further depth soundings are shown in the Figures 8 to 10. The geological profiles are based on drillhole observations (see Fig. 4). The figures in the columns give the layer resistivity (Formation resistivity), which resulted from an interpretation of the geoelectrical depth soundings. Figure 8 shows at the right hand side a wide bed of gravel deposits filling the base of a stream channel. Left of the center is a comparatively large near surface gravel deposit with a high resistivity between 100 and 135 Ωm. Very characteristic is depth sounding 10 of Figure 8, where no geoelectrical boundary was found at the depth of the bedrock. The indicated change to high resistivities between 270 and 550 Ωm occurs in the bedrock, where the fissures are closed and cannot hold any appreciable amounts of saline water. The same can be seen in Figures 9 and 10. Depth sounding 14, 22 and 23 gave inconclusive data in respect to the intermediate gravel deposit. The reason is that all three aquifers are here represented. The resistivities in the lower bedrock vary considerably at a lateral distance of 1 km (0.7 miles). This means that the depth sounding may be influenced by strong lateral changes of resistivity, in which case this method gives ambiguous results.

Comparing the layer resistivities with the rock material as described in the borehole observations one can assign certain resistivity ranges to different rock materials. Figure 11 shows a classification of the rock material in respect to resistivity. It shows that there is no overlap of resistivity between clay, gravel of the drift and near surface gravel. Thus the depth soundings, consequently, can distinguish layers of this material from each other.

Finally along two profiles a number of depth soundings were placed across the projected stream channel. The location is shown on Figure 4 depth sounding 7, 24, 25, 26 and 27 for the first profile. Figure 12 shows the depth soundings with interpretations of layer thicknesses and resistivities and underneath the resistivity profile. The high resistivity of the near surface gravel distinguishes it from the deeper sand layers of 40 Ωm. The bedrock is found by D. S. 7 (confirmed by drilling), D. S. 25 and D. S. 26. Since D. S. 7 shows a layer of 75 Ωm resistivity at the depth of the bedrock, it is assumed that the bedrock contains here very little salt water. Bedrock resistivities found by D. S. 25 and D. S. 26 are of similar size with 85 and 80 Ωm respectively. The intermediate layer of 40 Ωm under D. S. 7, 24 and 25 indicates a higher gravel-sand component, while resistivities of 24 and 22 Ωm under D. S. 26 and 27 indicate a higher amount of clay or a finer size of the grains. The results indicate a confined stream channel of about 1 km (0.7 miles) width between depth soundings D. S. 7 and D. S. 26. The channel is indicated by the dip of the bedrock as well as by the higher resistivity of 40 Ωm for the basal glacial deposits above the bedrock.

Figure 13 shows another profile across a buried glacial stream channel. The depth soundings are D. S. 13, 13b, and 13c from left to right (compare Figure 4). Depth sounding curves are very smooth and easily to be interpreted. For long electrode separations the resistivity rises due to a layer between 40 and 50 Ωm at a depth of 60-70m below surface. Compared with drilling results this is the sand-gravel fill, which is a potential aquifer for potable water.

6. The Origin of the Well Water

Special consideration has been given to depth sounding No. 1 which was taken next to drillhole No. 1 (Fig. 4) where a well penetrated bedrock. With the aid of a conductivity-bridge, the conductivity of a water sample taken from this well was found to be 2500 μmhos/cm. According to A. Homyk et al. (1967), water with more than 1000 ppm dissolved solids is considered to be saline. Between 2000 and 3000 ppm, the water is "doubtful to unsuitable" for irrigation purposes. Water from this well, which is at the Grundy County Rest Home, was chemically analyzed in 1955 by M. E. Phillips (Fuller and Russell, 1956). The analysis is given in ppm in the second column from the left of Table I. If the concentration of particular ions is known, it is possible to convert them into an equivalent amount of NaCl in solution and one can determine the resistivity of the water. Table I shows the conversion for well No. I.

Water resistivity is a function of the NaCl concentration and of temperature. With the help of standard conversion graphs (Schlumberger Well Logging Manual, 1958), the solution of 1547 ppm NaCl equivalent has a resistivity of 4.2 to 4.3 Ωm at 14^{o}C, the temperature at which conductivity was measured. This calculated resistivity is in good agreement with the measured water resistivity of 4 Ωm. Because the ions of very low concentrations were not considered in Table I, the calculated value of 4.2 to 4.3 Ωm most likely is the result of an accumulative error; therefore, the measured and calculated values are within a five percent margin of error. This means that the salinity of the water has not changed significantly within 16 years, which is the time interval between the chemical water analysis and the conductivity measurement.

It can be shown that the water in this well does not come from a fresh water aquifer in the glacial drift but most likely from bedrock. To do this, it is necessary to relate the layer or formation resistivity to the water resistivity. This relation is known as Archie's Law:

$$Ro = a\,R_w \cdot \Phi^{-m}, \tag{1}$$

where Ro is the formation resistivity, R_w is the water resistivity, Φ is the porosity, and a and m are constants. For consolidated rocks, a is in the vicinity of 1 while m varies between 2.25 and 1.60 having a lower value for less cemented material (Keller and Frischknecht, 1970, p. 21). These values, which represent consolidated rocks, were experimentally obtained to interpret electrical well log measurements for oil and gas explorations. Wyllie and Gregory (1953) investigated the contents of Archie's Law for unconsolidated material, thus their findings would be applicable for the glacial deposits. Their values are: $a = 1$ and $m = 1.3$. The formation factor is defined as $F = Ro/Rw$, where Rw is the resistivity of the pore water and Ro the formation resistivity. Ro is identical with the resistivity found by interpreting the geoelectrical depth sounding.

From depth sounding No. 1, the formation resistivity of the coarse-sand aquifer is Ro = 47.5 Ωm. One can assume that the pore volume in this layer is water saturated. With $F = \Phi^{-1.3}$ for different porosities (Φ in the first line of Table 2), the respective formation factors, F, are shown in the second line of Table 2. With Ro = 47.5 Ωm, the water resistivity can be calculated for each porosity (line 3, Table 2). The result shows that even for a porosity as low as 20 percent, the water resistivity (Rw = 5.6 Ωm) is higher than was actually measured (Rw = 4.0 Ωm). For a good aquifer, the porosity is higher than 20 percent.

This consideration makes it possible to conclude that the water in the well most likely comes from bedrock, while the fresh water in the buried glacial stream channel remains untouched. For improvement of water quality, it is recommended that the bedrock portion of the well be plugged and the water drawn from the glacial drift.

7. Conclusions and Applications

The topic of this investigation was to find geoelectrical prospecting methods, which were capable to distinguish and estimate potential aquifers in a rather complicated structure. Three aquifers were to be distinguished from each other, which appear in a vertical succession. Moreover it is the nature of a glacial deposit near the border of glaciation that the sand-clay relation varies considerably in lateral directions.

With few exceptions geoelectrical depth soundings with the Schlumberger Method indicated resistivities and boundaries, which can be associated with layers observed in drillholes. Unlike the application of the Wenner Method used by Midav (1960), these results can replace a considerable amount of drillholes and thus help to reduce the cost of exploration. A combined interpretation of a limited number of test holes with a detailed geoelectrical survey can give optimum results at minimal expenses. The location of the drill holes should be determined on the basis of the geoelectrical results.

While it is desirable to apply geophysical surface methods instead of many drillholes for cost-saving purposes, this investigation has shown that further important hydrogeological parameters can be obtained from a combined interpretation of geoelectrics and drillhole observations. The comparison of the bedrock-depth obtained from boreholes with the geoelectrically determined boundary at which the resistivity in the bedrock increases, indicates the thickness of a zone in which considerable amounts of saline water are stored. Finally, interesting problems such as salt water contamination of wells by encroachment and possible changes of well water salinity can be investigated.

Preliminary investigations are presently carried out to use gravity measurements for the determination of bedrock depth and of the bulk porosity of the basal gravel deposit in the stream channel.

8. References

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Ions	Concentration (ppm)	Conversion Factor	Equivalent Concentration of NaCl (ppm)
$(HOC_3)^-$	255.0	0.27	68.8
Ca^{++}	249.9	0.95	237.0
Mg^{++}	82.8	2.00	165.0
Na^+, K^+	360.3	1.0	360.3
SO_4^{--}	1,384.4	0.50	692.0
Cl^-	23.5	1.0	23.5
Total	2,355.9		1,546.6

Table 1. Conversion of ion quantities into equivalent amounts of NaCl concentration for well water of test well No. 1 (conversion factors after Dunlap and Hawthorne, 1951).

Φ	0.2	0.3	0.4	0.5	0.6
F	8.10	4.78	3.29	2.70	2.12
Rw(Ωm)	5.6	9.4	13.6	18.3	23. 2

Table 2. Calculation of water resistivity for assumed porosities at a formation resistivity of Ro = 47.5 Ωm.

10. List of Illustrations

11. Publications, Reports, Papers, Talks presented

Papers:

1. Detection of Fresh Water Aquifers in the Glacial Deposits of Northwestern Missouri by Geoelectrical Methods. Submitted for publication.

2. Combined Geoelectrical and Borehole Investigations for the Detection of Totable Water in Northwestern Missouri. Geophysics, in preparation.

3. Geoelectrical Depth Soundings for the Detection of Fresh Water in Glacial Stream Channels.
   Short Paper, Proceedings Series No. 16, Eights American Water Resources Conference, 1972.

Talks presented:

1. Methods of Determining Aquifer Storage Capacity and Fresh-Saline Water Interfaces by Geoelectrical Investigations. 5th Annual Conference, Columbia, Missouri, May 1972, Missouri Water Resources Research Center.

2. Geoelectrical Depth Soundings for the Detection of Fresh Water in Glacial Stream Channels, Eighths American Water Resources Conference, St. Louis, Nov. 1972.

3. Combined geoelectrical and Borehole Investigations for the Detection of Fresh Water Aquifer in Northwestern Missouri. 42nd International Meeting of the Society of Exploration Geophysicists, Anaheim, Calif., Nov. 27-1972.

12. Training Accomplished

One graduate student, Mr. William Head, was supported as graduate research assistant. He received detailed training in all measuring techniques and in geoelectrical interpretation techniques.

The main part ot this work was conducted while the principle investigator offered a course in "Geoelectrical Prospecting Methods". The students were offered application of there methods in conjunction with this grant.

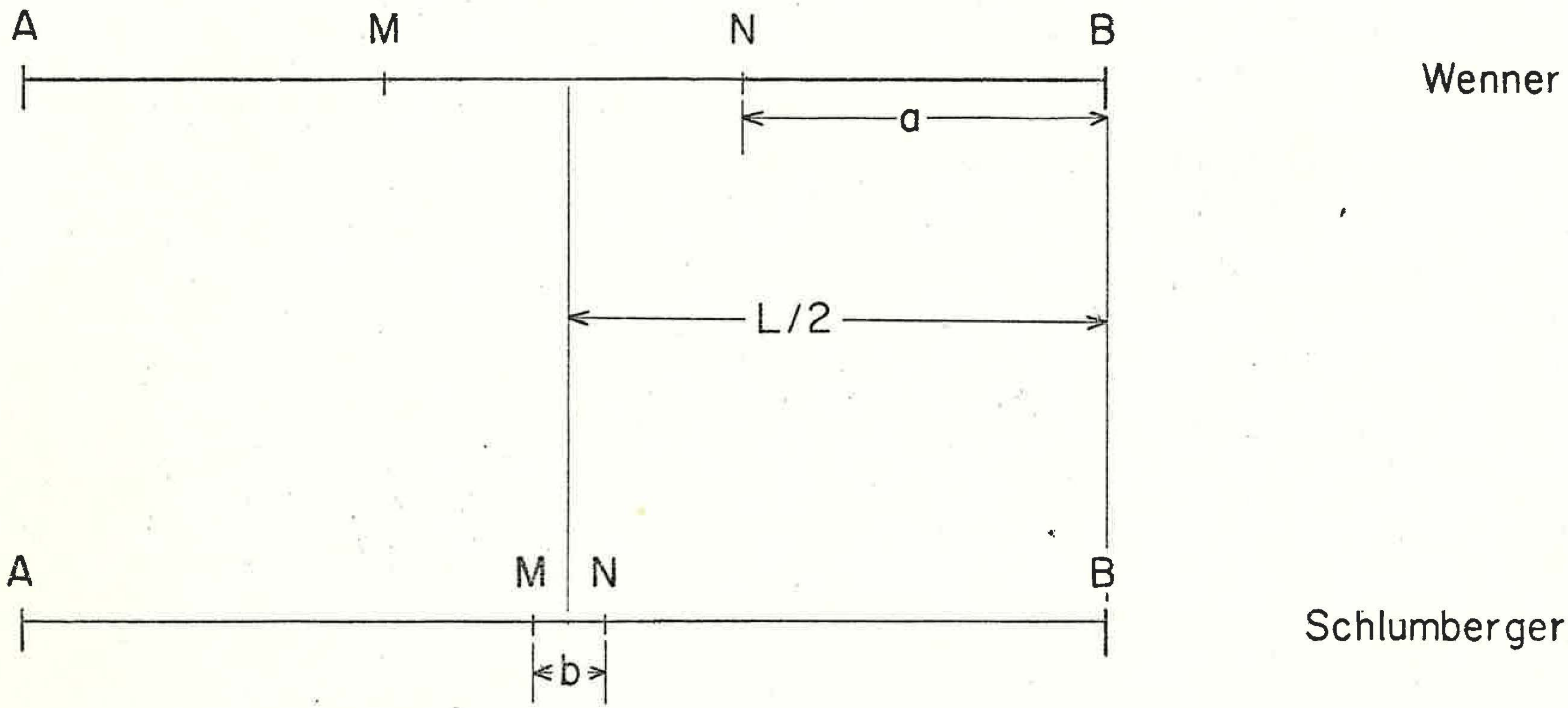
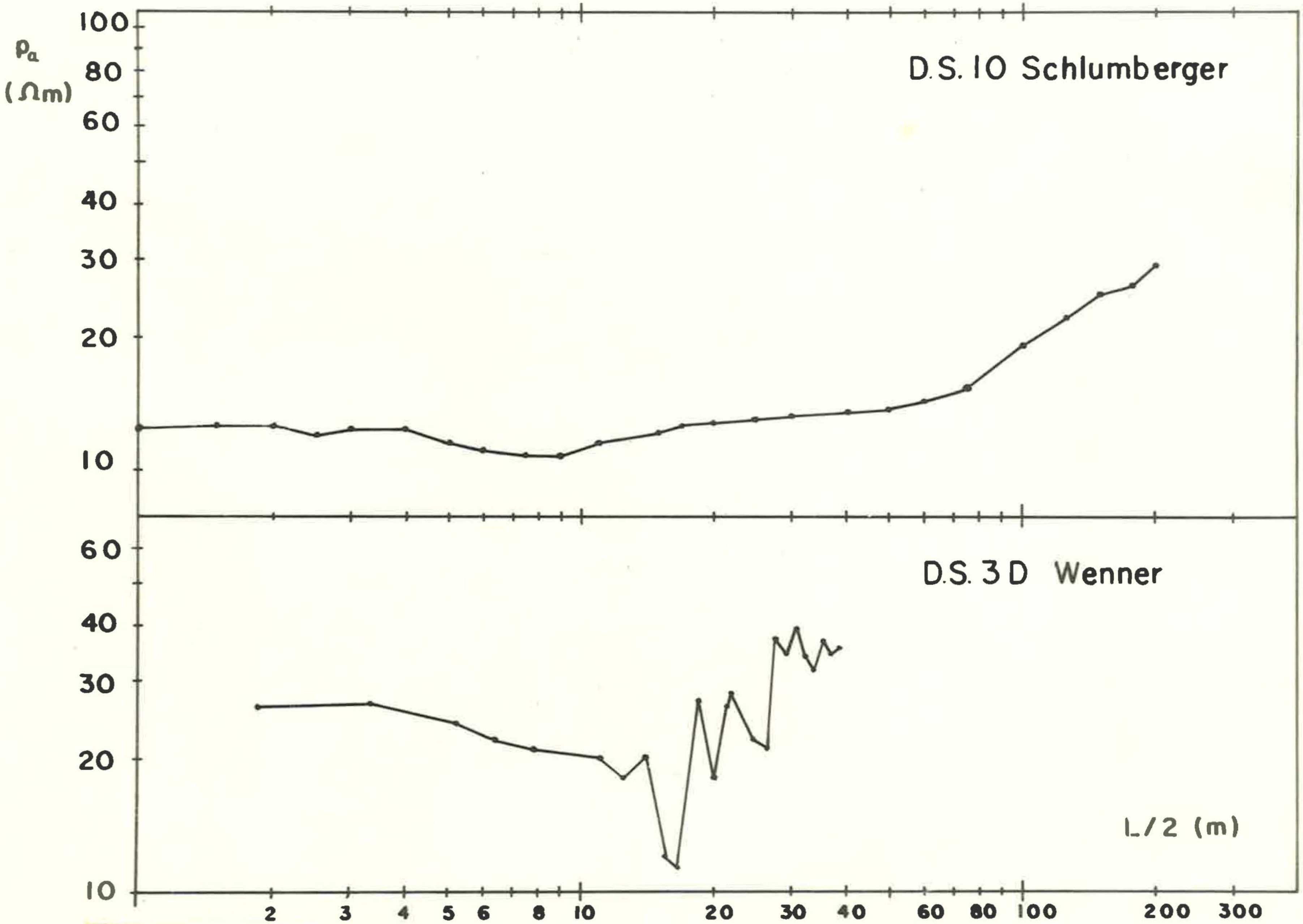


Fig. 1

Fig. 1



D.S. 10 Schlumberger
D.S. 3 D Wenner
ρ_a (Ωm)
100
80
60
40
30
20
10
60
40
30
20
10
L/2 (m)
2
3
4
5
6
8
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80
100
200
300

Fig.2

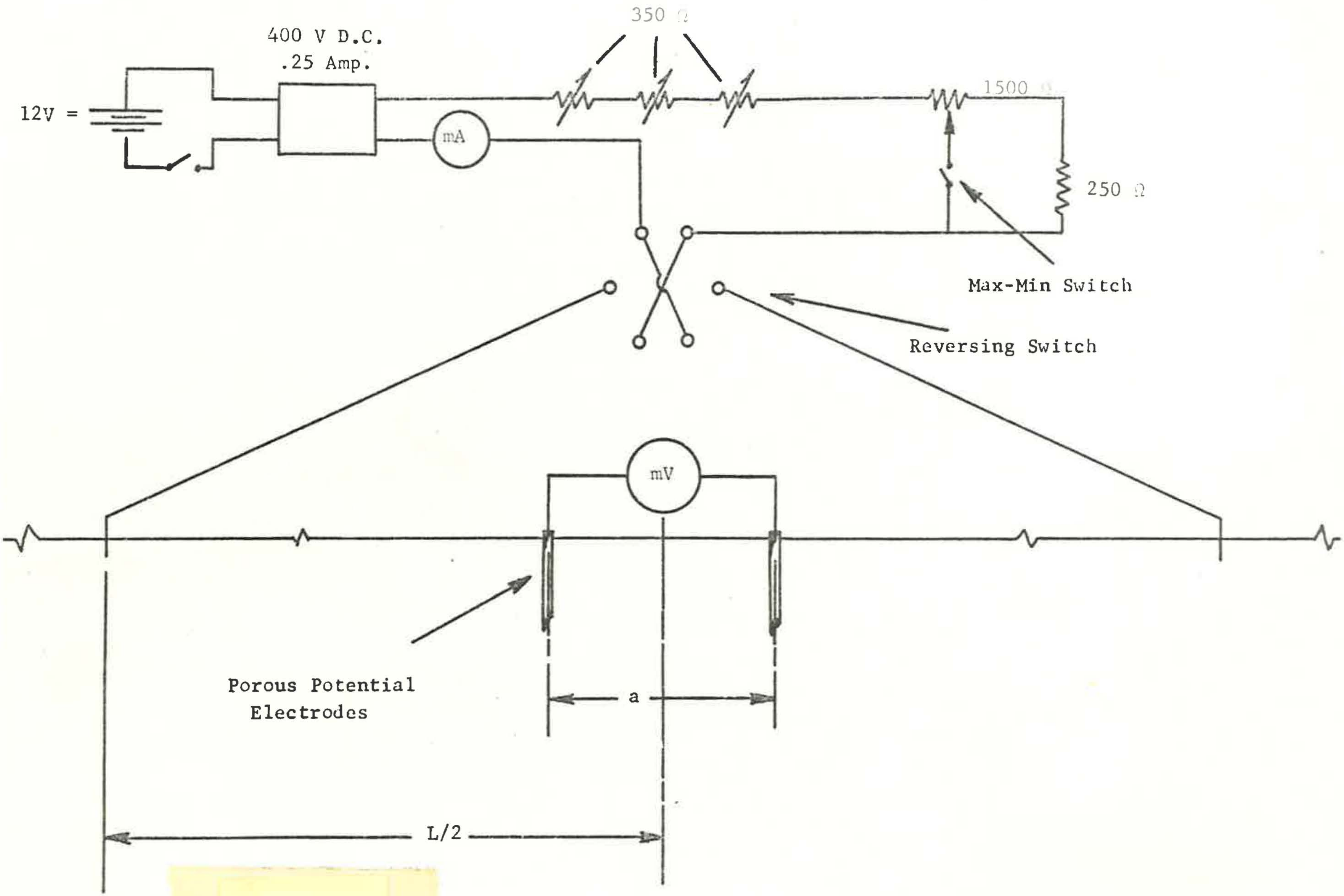

350 Ω
400 V D.C.
.25 Amp.
12V =
mA
1500 Ω
250 Ω
Max-Min Switch
Reversing Switch
mV
Porous Potential
Electrodes
a
L/2

Fig.3

Fig. 3

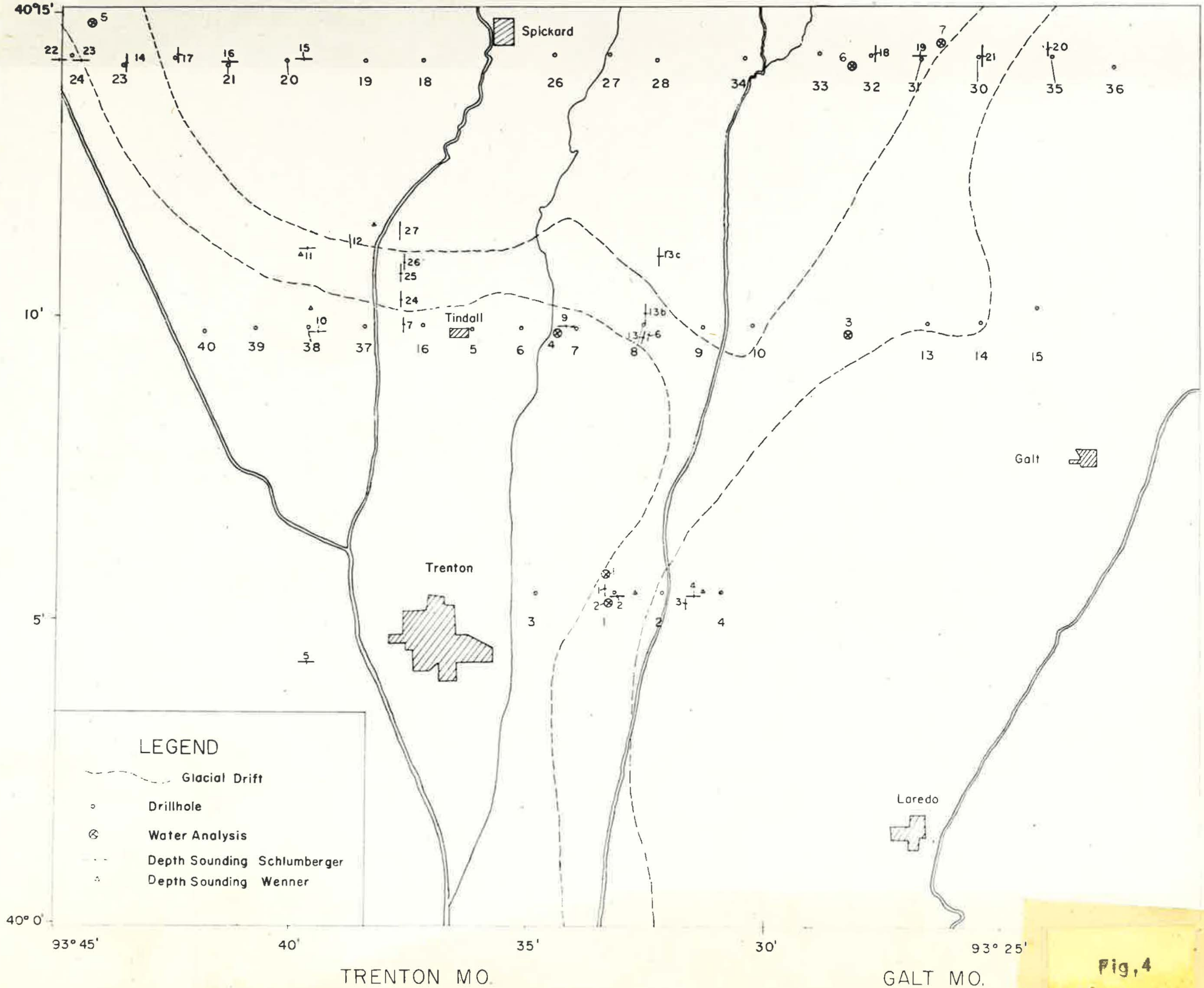
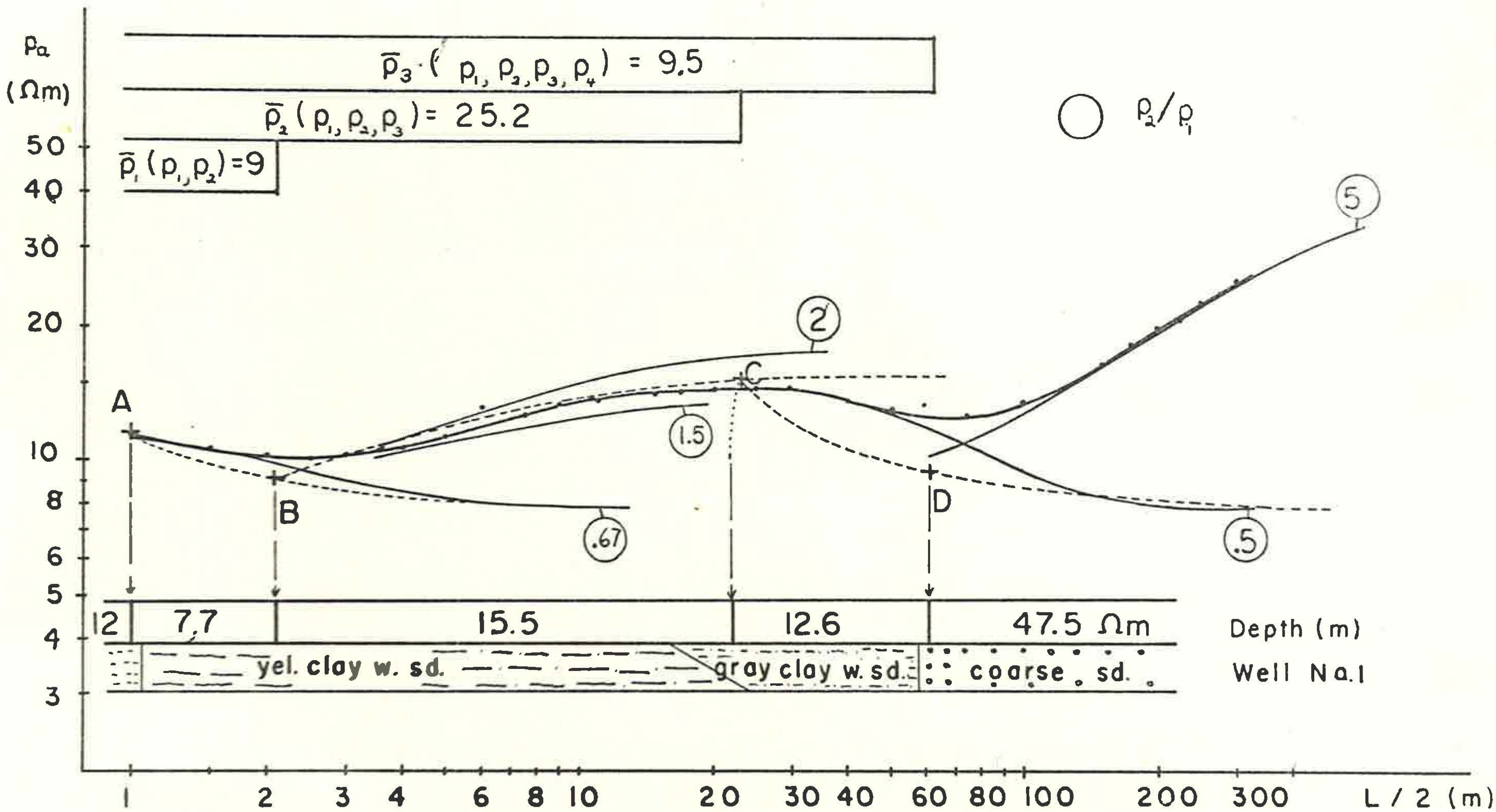

40°15'
10'
5'
40° 0'
93° 45'
40'
35'
30'
93° 25'
Spickard
Tindall
Trenton
Galt
Laredo
LEGEND
Glacial Drift
Drillhole
Water Analysis
Depth Sounding Schlumberger
Depth Sounding Wenner
TRENTON MO.
GALT MO.

Fig. 4



ρa (Ωm)
50
40
30
20
10
8
6
5
4
3
ρ̄3 (ρ1, ρ2, ρ3, ρ4) = 9.5
ρ̄2 (ρ1, ρ2, ρ3) = 25.2
ρ̄1 (ρ1, ρ2) = 9
ρ2/ρ1
5
2
1.5
.67
.5
A
B
C
D
12
7.7
15.5
12.6
47.5 Ωm
Depth (m)
yel. clay w. sd.
gray clay w. sd.
coarse sd.
Well No.1
1
2
3
4
6
8
10
20
30
40
60
80
100
200
300
L / 2 (m)

Fig.5

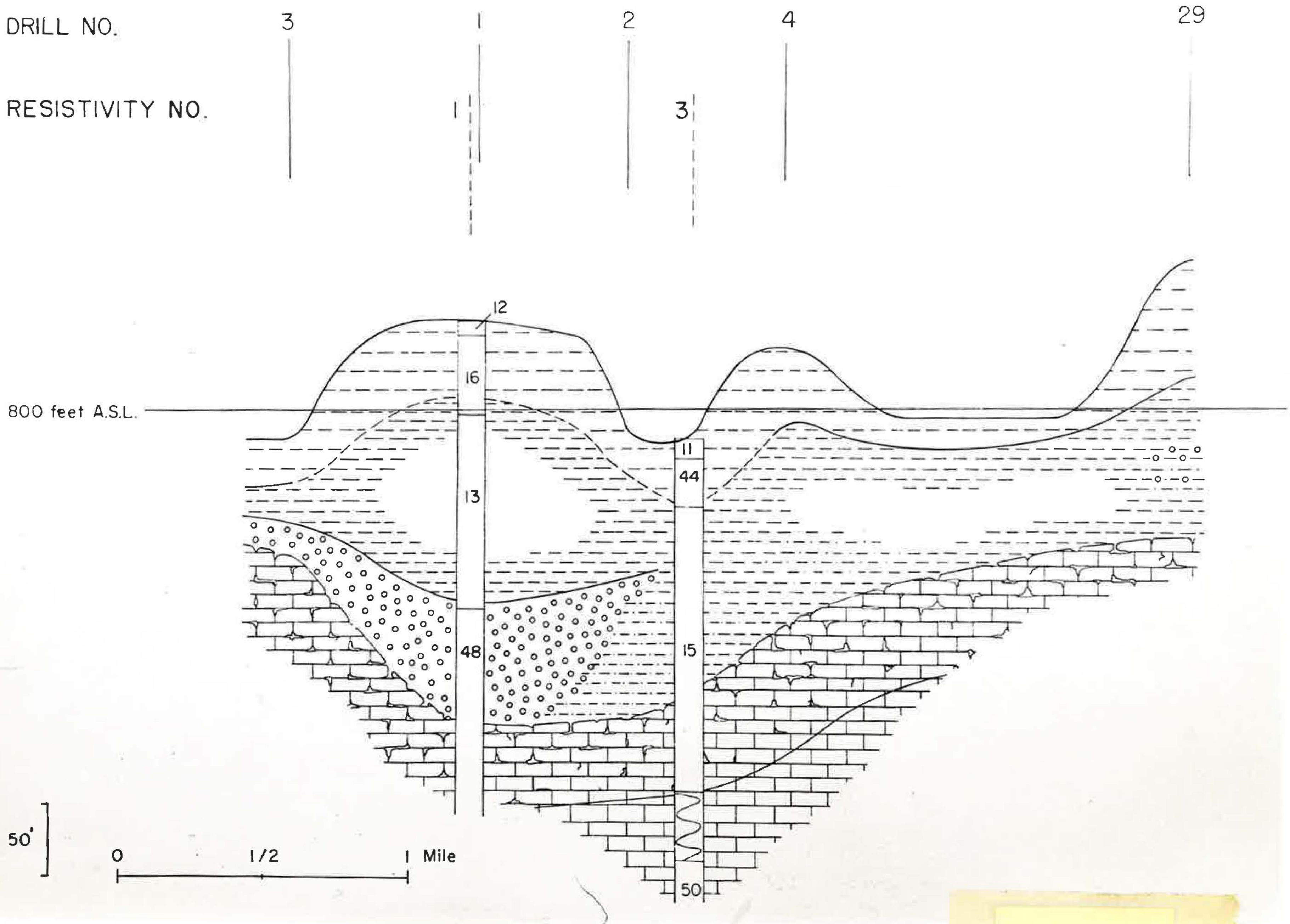
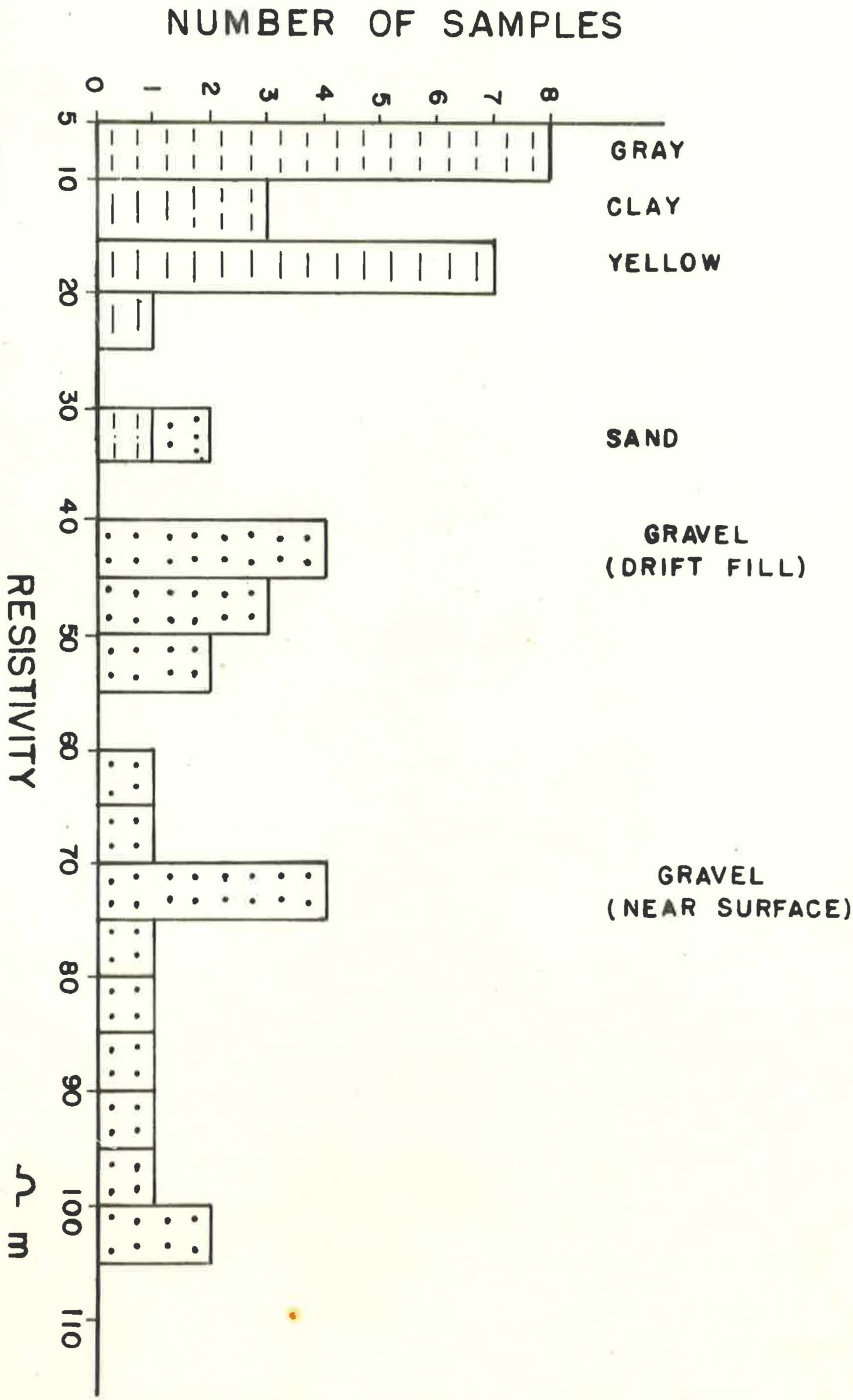
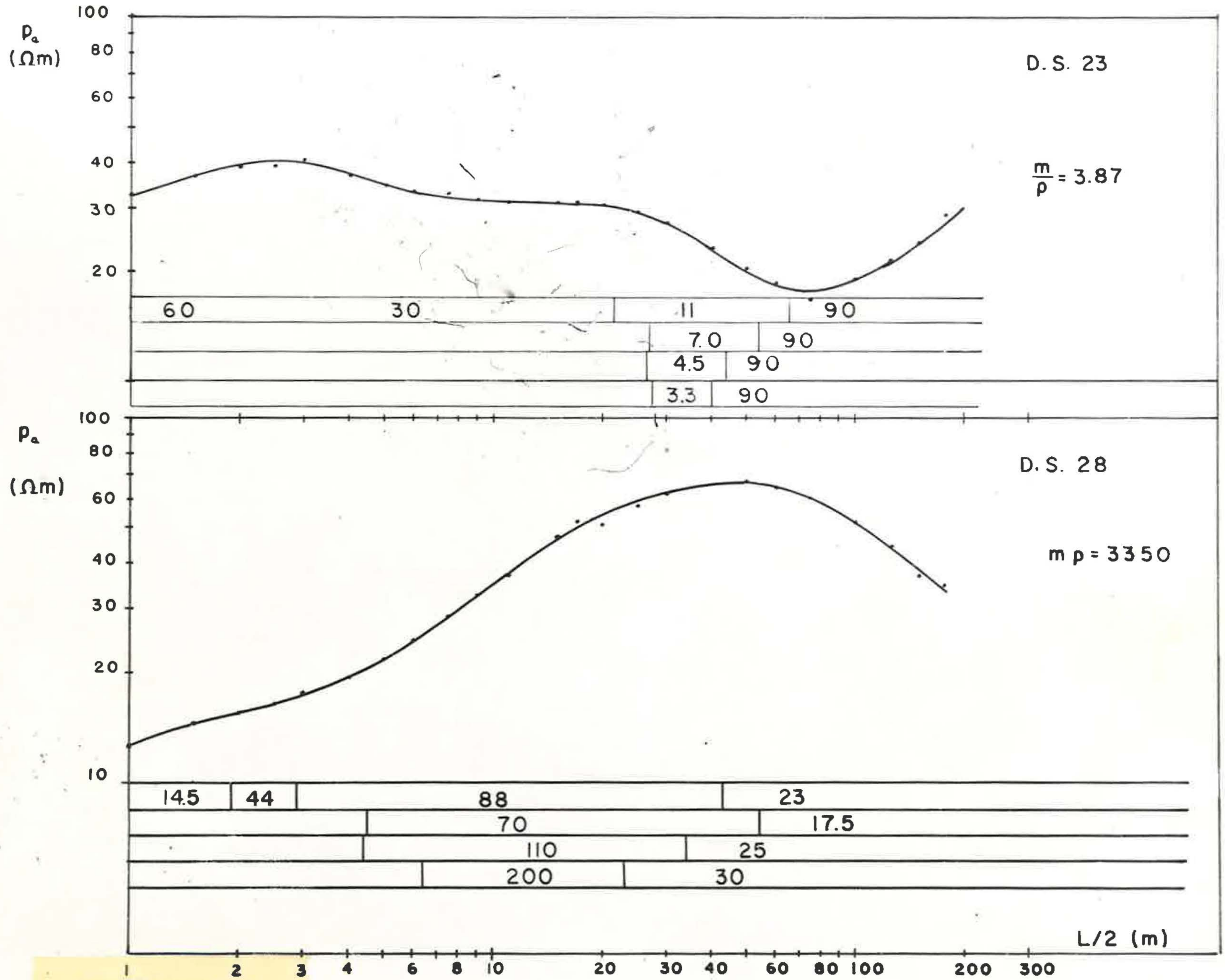

DRILL NO.
3
1
2
4
29
RESISTIVITY NO.
1
3
12
16
13
48
11
44
15
50
800 feet A.S.L.
50'
0
1/2
1 Mile

Fig.6

Fig. 11


NUMBER OF SAMPLES
0
1
2
3
4
5
6
7
8
5
10
20
30
40
50
60
70
80
90
100
110
RESISTIVITY
Ω m
GRAY
CLAY
YELLOW
SAND
GRAVEL
(DRIFT FILL)
GRAVEL
(NEAR SURFACE)



D.S. 23
$\frac{m}{\rho}$ = 3.87
60
30
11
90
7.0
90
4.5
90
3.3
90
D.S. 28
m ρ = 3350
14.5
44
88
23
70
17.5
110
25
200
30
ρ_a (Ωm)
L/2 (m)

Fig, 7

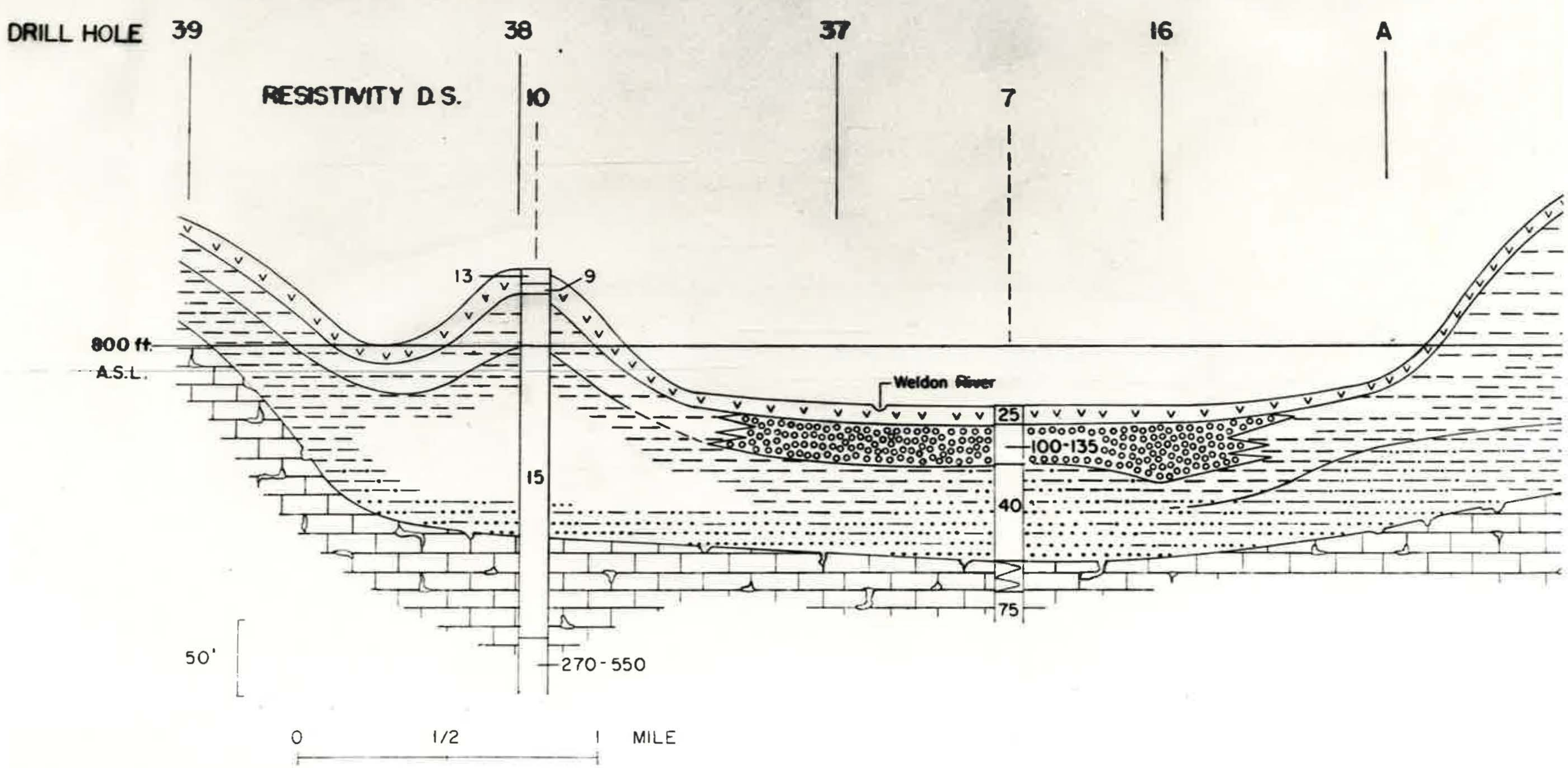

DRILL HOLE
39
38
37
16
A
RESISTIVITY D.S.
10
7
13
9
800 ft.
A.S.L.
Weldon River
25
100-135
40
75
15
270-550
50'
0
1/2
1
MILE

Fig.8a

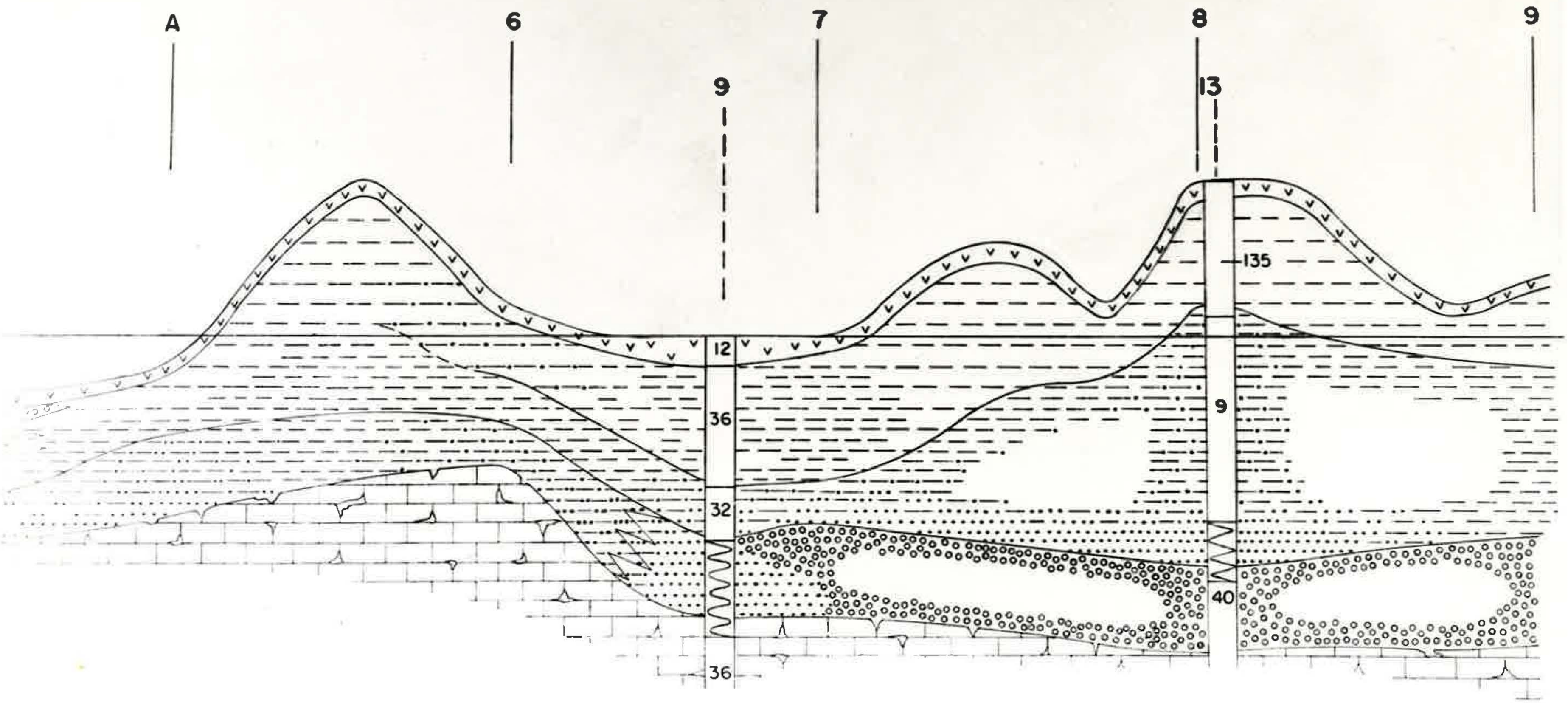


Fig.8b

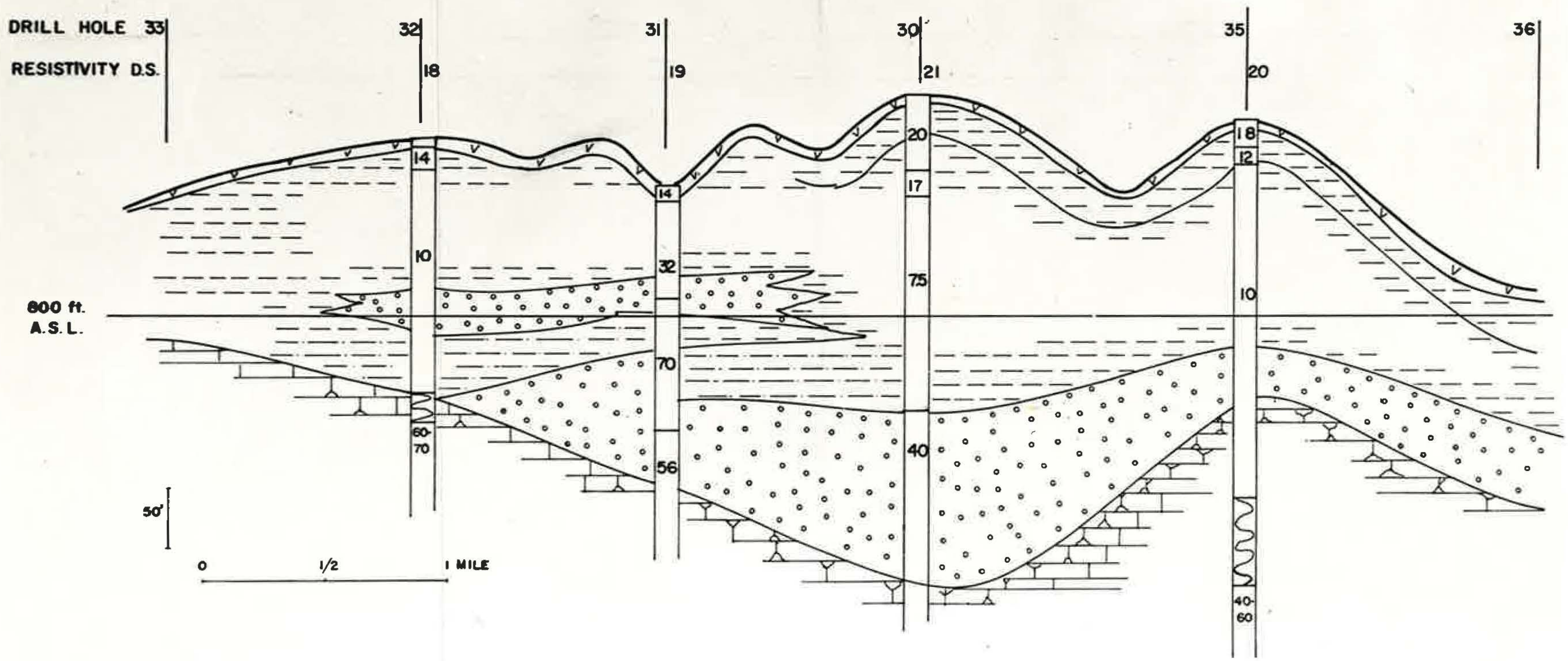

DRILL HOLE 33
32
31
30
35
36
RESISTIVITY D.S.
18
19
21
20
14
14
20
18
12
17
10
32
75
10
70
40
56
60-70
40-60
800 ft.
A.S.L.
50'
0
1/2
1 MILE

Fig.9

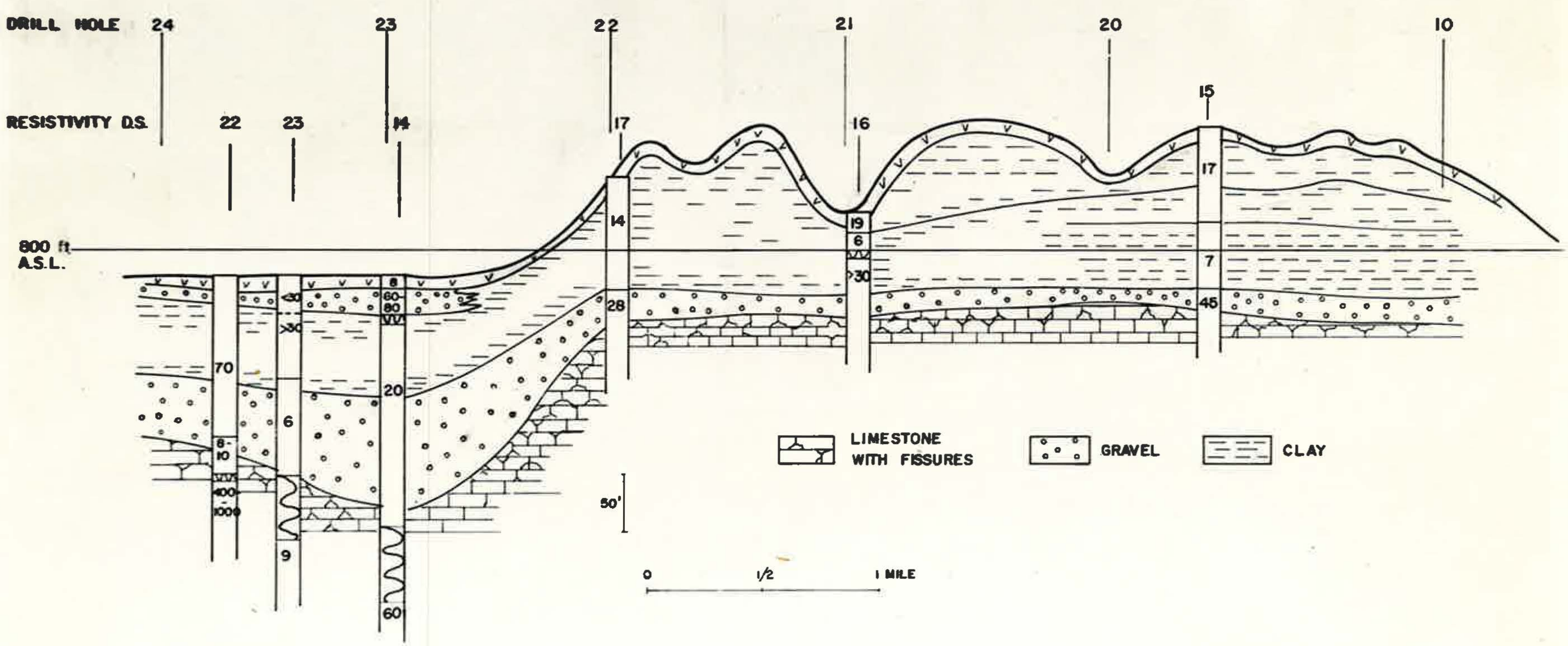

DRILL HOLE
RESISTIVITY D.S.
800 ft
A.S.L.
24
23
22
21
20
10
22
23
14
17
16
15
70
8-10
400-1000
6
9
8
60-80
20
60
14
28
19
6
>30
17
7
45
LIMESTONE WITH FISSURES
GRAVEL
CLAY
50'
0
1/2
1 MILE

Fig. 10

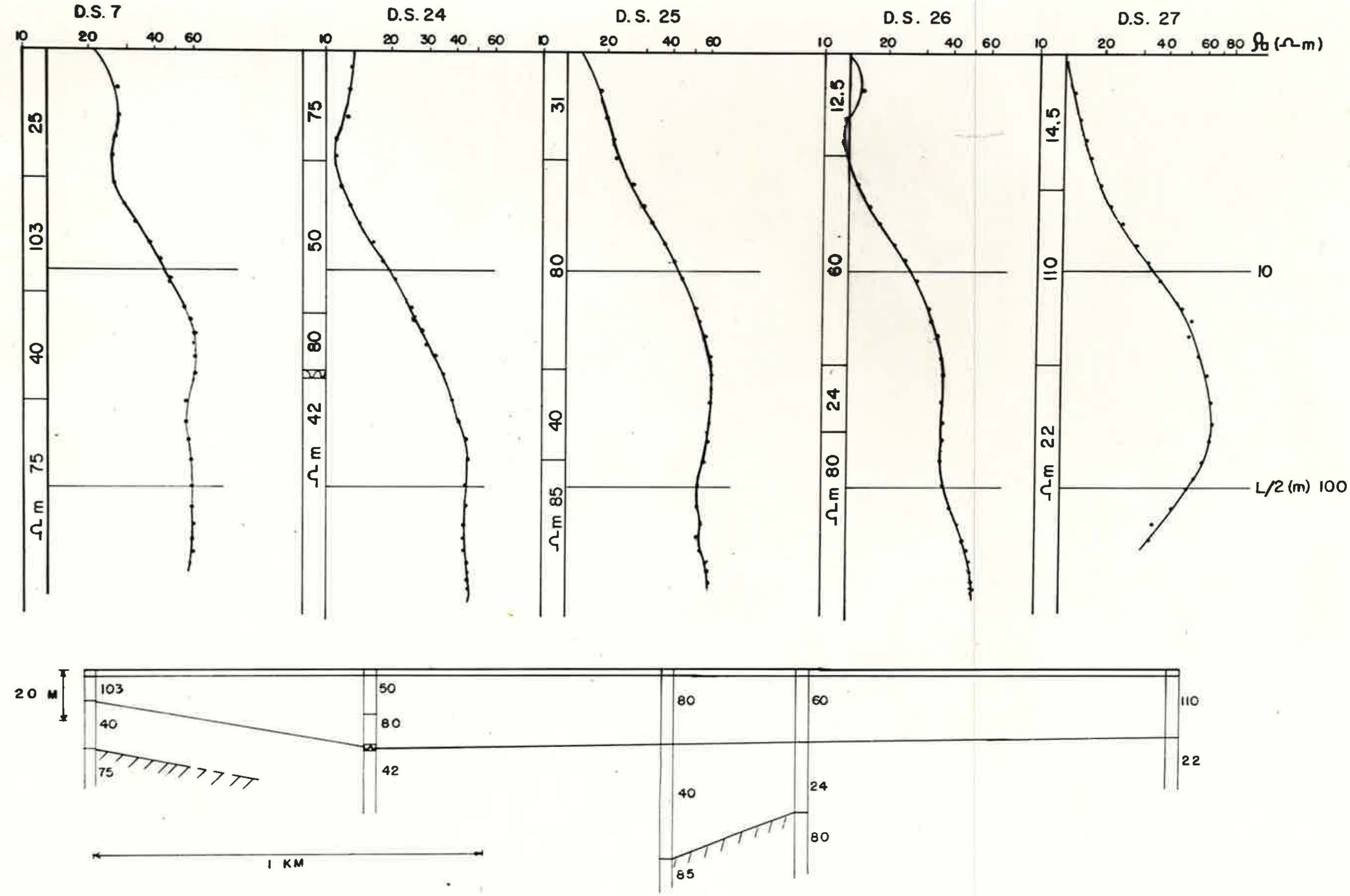
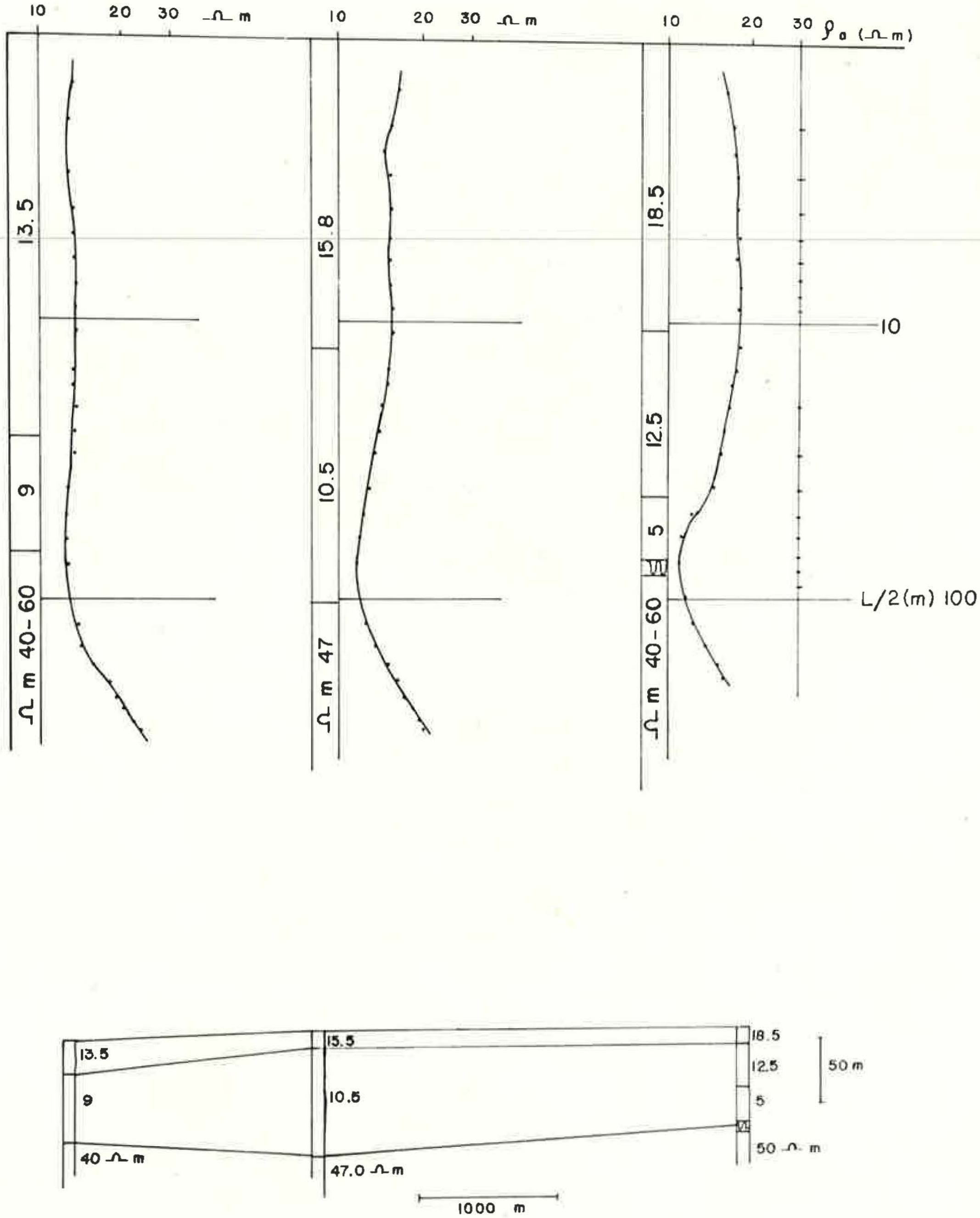


Fig. 12



10
20
30
Ω m
ρa (Ω m)
13.5
9
Ω m 40-60
15.8
10.5
Ω m 47
18.5
12.5
5
10
L/2(m) 100
13.5
9
40 Ω m
15.5
10.5
47.0 Ω m
18.5
12.5
5
50 Ω m
50 m
1000 m

Fig. 13