STATE OF ILLINOIS
OTTO KERNER, *Governor*
DEPARTMENT OF REGISTRATION AND EDUCATION
WILLIAM SYLVESTER WHITE, *Director*

INTEGRATED SEISMIC, RESISTIVITY, and GEOLOGIC STUDIES OF GLACIAL DEPOSITS

Lyle D. McGinnis
John P. Kempton

DIVISION OF THE
ILLINOIS STATE GEOLOGICAL SURVEY
JOHN C. FRYE, *Chief* URBANA

CIRCULAR 323 1961

INTEGRATED SEISMIC, RESISTIVITY, AND GEOLOGIC STUDIES OF GLACIAL DEPOSITS

INTEGRATED SEISMIC, RESISTIVITY, AND GEOLOGIC STUDIES OF GLACIAL DEPOSITS

Lyle D. McGinnis and John P. Kempton

ABSTRACT

A study integrating seismic, resistivity, and geologic methods was made of the glacial deposits in three areas in the northern half of Illinois. Employment of all three methods increased the reliability of the identification of materials and determination of thickness of units by eliminating some of the problems encountered when a single method was used.

Resistivity values are a guide to the identification of, and depth to, low-velocity sand and gravel, and with these layers accounted for, seismic refraction methods give more reliable bedrock depths. Resistivity methods also aid in identifying the sequence and lithologic type of high-velocity drift materials. Depths to the base of high-velocity layers can be determined from resistivity curves, and bedrock depths can then be more accurately calculated from seismic refractions.

A low-velocity surficial layer from 3 to 17 feet thick was indicated in all three areas studied, and some correlation between it and the geologic weathered zone was noted.

Where the drift was 10 to 20 feet thick and the bedrock surface uneven, bedrock depth and velocity could not be determined accurately. Where resistivities in the drift and bedrock were similar, the resistivity method failed to determine depth.

INTRODUCTION

During the summer of 1960, an attempt was made to increase the accuracy of determinations of thickness of glacial deposits and to identify their lithologic variations by the complementary use of seismic and resistivity methods. Nineteen stations in northern Illinois were chosen for the study. This work is part of a broader program to add control to maps of the bedrock topography of Illinois, which yield important information concerning the location and nature of ground-water supplies throughout the state. The results of integrated seismic and resistivity studies at 6 of the 19 stations (fig. 1) are reported here.

Nearly all of Illinois has been covered at some time by one or more Pleistocene glaciers, the only driftless areas occurring in a narrow strip in western Illinois, in the northwestern corner, and in the southern tip of the state. In general, the thickest drift deposits are found in the northeastern third of the state where deposits of the final glaciation (Wisconsinan) cover older drift and bedrock formations.

Pennsylvanian rocks, mainly shale, constitute a large proportion of the bedrock surface in the southern four-fifths of Illinois, whereas Ordovician and Silurian dolomite and shale are the predominant rocks at the surface in the northern fifth. This study deals mainly with the thickness and properties of the drift materials, and is concerned with bedrock only at the drift-bedrock contact.

Geophysical methods have been used extensively in the study of the thickness and character of glacial deposits. Electrical resistivity surveys frequently have been used in drift studies (Hubbert, 1932; Frye, 1938; Bays and Folk, 1944; Bays, 1946; Foster and Buhle, 1951; Buhle, 1953, 1957; Hackett, 1956; and Meidav, 1960) and seismic surveys have been used to determine thickness and variations in the drift material (Johnson, 1954; Warrick and Winslow, 1960). Both refraction and reflection methods were used with success. Woollard and Hanson (1954) applied seismic, electrical, magnetic, and gravity methods to various glacial drift problems in Wisconsin.

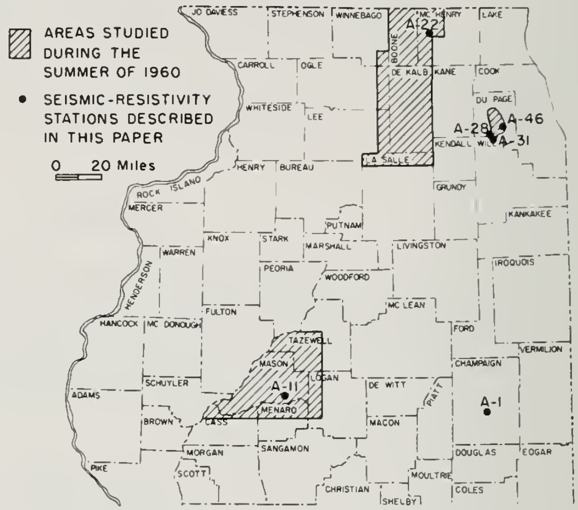

Fig. 1 - Areas studied for this report.

EQUIPMENT

Seismic measurements were made with a Century 12-channel portable refraction seismograph designed primarily for engineering investigations. Geophones used were peaked at a frequency of 12 cycles per second. Paper speed was approximately 27 in/sec. Resistivity studies were made with a modified Gish-Rooney instrument. The potential drop was measured with a specially calibrated potentiometer, and the apparent earth resistivity was read directly in ohm-centimeters.

USES AND LIMITATIONS OF RESISTIVITY AND SEISMIC METHODS IN DRIFT STUDIES

Resistivity Method

Sand and gravel generally are identified quite easily by the resistivity method because their resistivity is higher than that of other glacial drift materials. Identification becomes difficult, if not impossible, where these deposits are thin and deeply buried. Within broad limitations, the water-bearing potential of drift can be determined successfully.

Reliable approximations of drift thickness may be obtained where the resistivities of the drift and bedrock differ substantially (for example, till on dolomite or sand and gravel on shale), but thickness determinations are ambiguous where glacial deposits and the underlying bedrock formations have similar resistivities (sand and gravel on dolomite or till on shale). As perhaps 60 percent of Illinois has this latter type of drift-bedrock contact, resistivity methods must be used with caution for calculations of drift thickness. Resistivity data in this paper are used primarily for the identification of lithologic units of the drift, but where the seismic method is inadequate, shallow boundaries are inferred by using the method described by Moore (1945).

Seismic Method

Seismic velocities also can be used to identify the common lithologic units in glacial deposits. Determinations of thickness generally have proved reliable, particularly where the drift is more or less monolithologic. A considerable contrast in velocity almost always exists between drift and bedrock. Certain lithologic sequences within the drift, however, make depth determinations difficult. For example, the refraction method is seriously limited where high-velocity till covers low-velocity sand and gravel deposits. A refraction spread over such a drift sequence commonly shows the "weathered" layer, the high-velocity till, and the bedrock, but not the low-velocity deposit. An interpretation that does not consider the masked low-velocity layer leads to a thickness determination greater than the real value.

An empirical method used to determine depths to velocity interfaces below the low-velocity layer was developed by Johnson (1954, p. 46), who stated: "Depths to the base of the low velocity zone equal the apparent thickness of the layer overlying the low velocity zone." This empirical method can involve as many inherent errors in the calculation of depths as the theoretical method that gives the total depth as the summation of all the layers.

ANALYSES AT SELECTED SEISMIC-RESISTIVITY STATIONS

Stations A-1, A-11, A-22, A-28, A-31, and A-46 (fig. 1) have been selected for this discussion because they occur in areas of adequate geologic control and because concepts that exert considerable influence on the geophysical study of glacial drift are there well demonstrated. Specific studies were made of

1) The "seismic weathered" layer (Lester, 1932) and its relationship to the "geologic weathered" layer, which is described in detail for station A-1
2) Variations in velocity sequences beneath the weathered layer described in detail at stations A-11 and A-22
3) The integrated use of resistivity and seismic methods at all six stations.

The Seismic Weathered Layer

Seismic refraction or reflection studies may be influenced to a great extent by a surficial, low-velocity layer. This layer is referred to in seismic literature, sometimes erroneously, as the "weathered layer." A surficial, low-velocity layer ranging in thickness from 3 to 17 feet is present in all the areas studied. Where glacial deposits are hundreds of feet thick the layer would have little influence on most drift studies, but where the drift thickness is measured in tens of feet this layer may alter significantly the results of a seismic survey. The

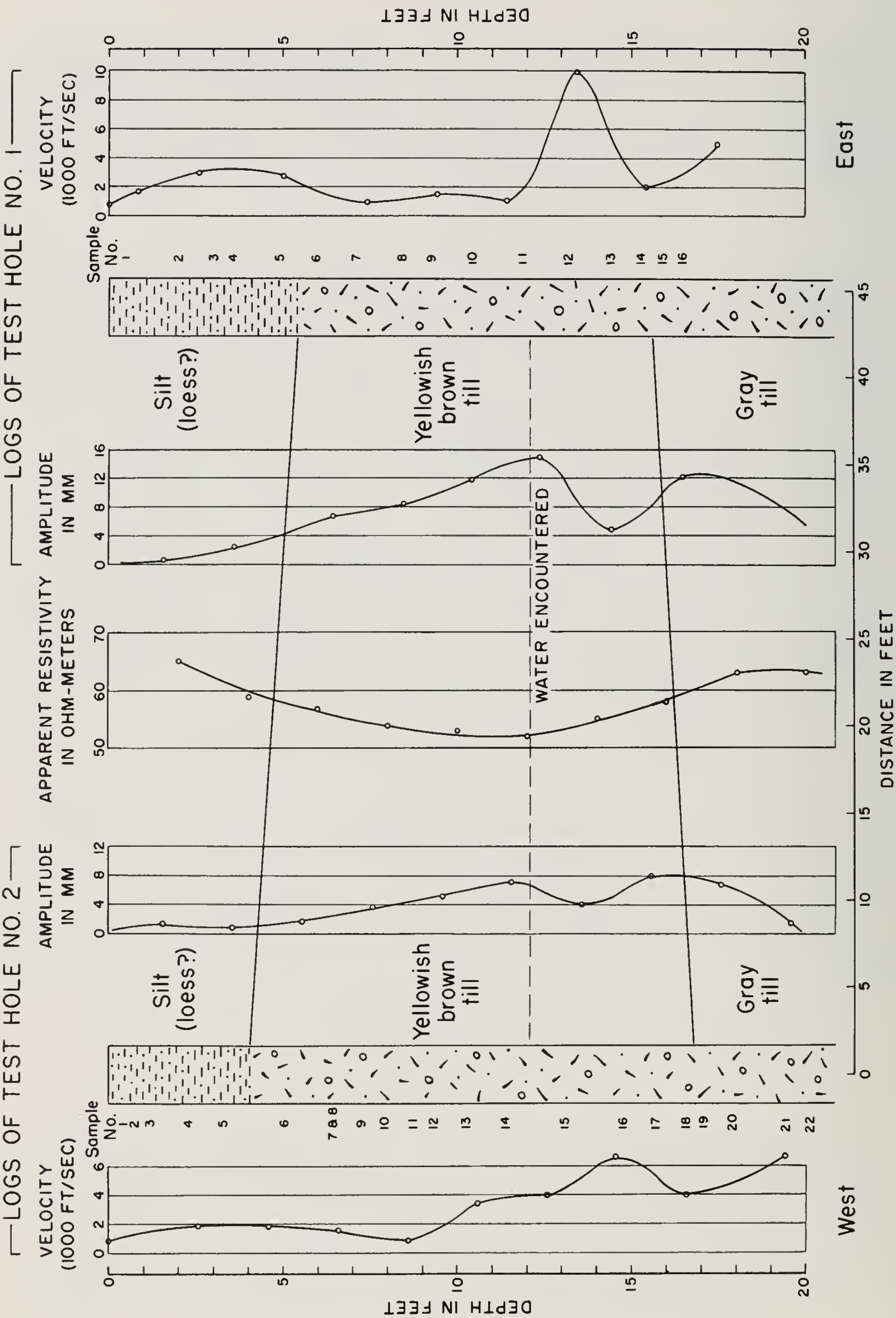

Fig. 2 - Resistivity curve and velocity, amplitude, and geologic logs for two test holes at station A-1.

relationship between the surficial, low-velocity seismic layer and the geologic weathered layer has not been fully explained (Lester, 1932). In order to gain more knowledge of the relationship between the two, a shallow geophysical and geological study was made at station A-1, in the NE corner sec. 30, T. 19 N., R. 9 E., Champaign County, on the front slope of the Champaign moraine.

Two holes were drilled 44 feet apart to depths of 18 feet (hole 1) and 20 feet (hole 2). Samples of the material encountered were taken at short intervals (fig. 2) and studied under a binocular microscope to determine lithologic type and by X-ray diffraction to determine the alteration sequence of the clay minerals. The materials, essentially the same at comparable levels in both holes, included at the surface about 4 to 5 feet of dark yellowish brown, oxidized, noncalcareous silt (possibly loess) (samples 1 to 5) on which a very dark brown to black soil about 3 feet thick is developed. This overlies about 9 to 14 feet of silty, yellowish to olive brown till (samples 6 to 14, hole 1; 6 to 18, hole 2). No carbonates are present in the upper foot of the till (sample 6 of each hole). The remaining samples are dark gray to dark brownish gray calcareous till apparently unaffected by oxidation. No unusual textural variations were noted, although the gray till appears to be more compact and may contain a slightly higher percentage of clay-sized material.

X-ray diffraction analyses were made of the test hole samples by Dr. H. D. Glass of the Illinois State Geological Survey. The diffraction patterns of the clay minerals in the samples clearly substantiate the observed distinction between the yellowish brown and gray tills. Whereas the gray till consists of muscovite-type illite and chlorite in proportions of about two to one, the yellowish brown till samples show a progressive upward degradation of the chlorite. No chlorite can be identified in sample 6 of either hole. No effect was noted on the illite. This alteration sequence is typical of soils developed on tills of Wisconsinan age throughout Illinois. The silt, particularly in sample 5 of each hole, contains about 40 percent montmorillonite plus illite, kaolinite, and chlorite. The weathering profile is developed on the silt and extends into the till.

Because the yellowish brown till is thicker than is considered normal for the weathering profile developed on Wisconsinan tills of the area, alternative interpretations must be considered. The location of station A-1, on the front slope of the Champaign moraine, suggests the possibility of downslope movement of materials. This movement may have increased the thickness of the yellowish brown drift, thus yielding an exaggerated thickness for the oxidized portion of the Champaign drift. The yellowish brown till also may represent a slightly younger drift than the gray till below. The fact that the gray till appears to be slightly more compact may be due to the additional weight of the overriding ice and drift that formed the Champaign moraine. This would not affect the interpretation of the yellowish brown drift as oxidized till but might exaggerate the differences in physical properties of the altered and unaltered till. The thickness of the silt cap also is greater than ordinarily would be expected for loess in this area and may be the result of the addition of some colluvial material of loessal origin from upslope.

Two short refraction spreads were shot at station A-1 to determine the thickness of the low-velocity layer. The travel-time curve of the first spread, shot in May, is shown in figure 3. The velocity in the low-velocity layer was 1130 ft/sec, while the material immediately below had a velocity of 5420 ft/sec. The thickness of the low-velocity layer determined from this shot was approximately 14 feet.

The second spread was shot in October and the thickness again was found to be about 14 feet.

Electric blasting caps, shot at approximately two-foot intervals up the auger holes, were used to plot velocity logs (fig. 2). Geophones were placed at the surface of the two holes and at four-foot intervals between them. The maximum velocity occurred between 13 and 14 feet in hole 1 and between 14 and 15 feet in hole 2. The maximum velocity in hole 1 was much too high for any drift material and may be either somewhat in error or may indicate a real change in the physical properties of the material. Because a high velocity occurred at similar depths in both holes, the latter alternative is believed to be correct. Iida (1940) stated: "The velocity of elastic waves (in clays) tends to decrease with an increase of water until a critical water volume is reached." Samples from this high-velocity zone were wetter than those at any other depth and may represent a water volume greater than the critical volume mentioned by Iida. Velocities below the wetter layer show a marked increase over the velocities in the material above the wet zone, but only one velocity could be obtained at this depth in each hole. The mid-point of the high-velocity wet layer coincides with the depth to the top of the first boundary determined from the short refraction spreads.

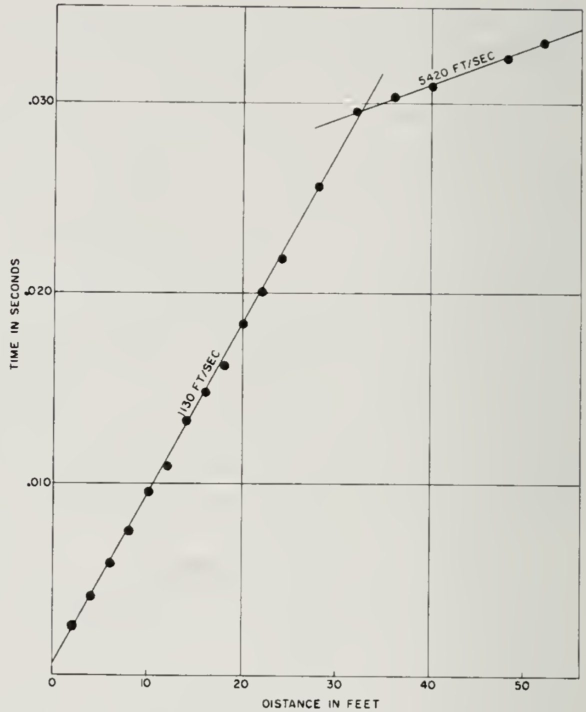

Fig. 3 - Time-distance curve for station A-1.

Another log of the holes with perhaps as much significance as the velocity log is a measure of the amplitude (proportional to the square of the energy) of a refracted first arrival versus the depth of the shot (fig. 3). All shots were made with electric blasting caps to insure a constant source of energy at each shot position (Williams, 1946). Amplifier gains were held constant for all shots. The relative amplitude of the wave refracted off the gray till and recorded on trace 7 of each record was measured. This trace was selected because the geophone was 24 feet from the shot, which allowed part of the ray path for each shot to pass through the gray till and return to the surface at approximately the same angle of incidence for each shot depth.

The amplitude of trace 7 rose with increasing shot depth to a high value until it encountered the wet zone, in which a low amplitude was recorded in each hole. Below this low value the amplitude again increased and reached another high at about 16 feet. The attenuation of a seismic wave generally increases as the shot-detector distance increases. Therefore, the increase in amplitude with increasing shot depth from the surface downward may be attributed to a decrease in the ray path distance from the shot to the receiver. The low-amplitude zone

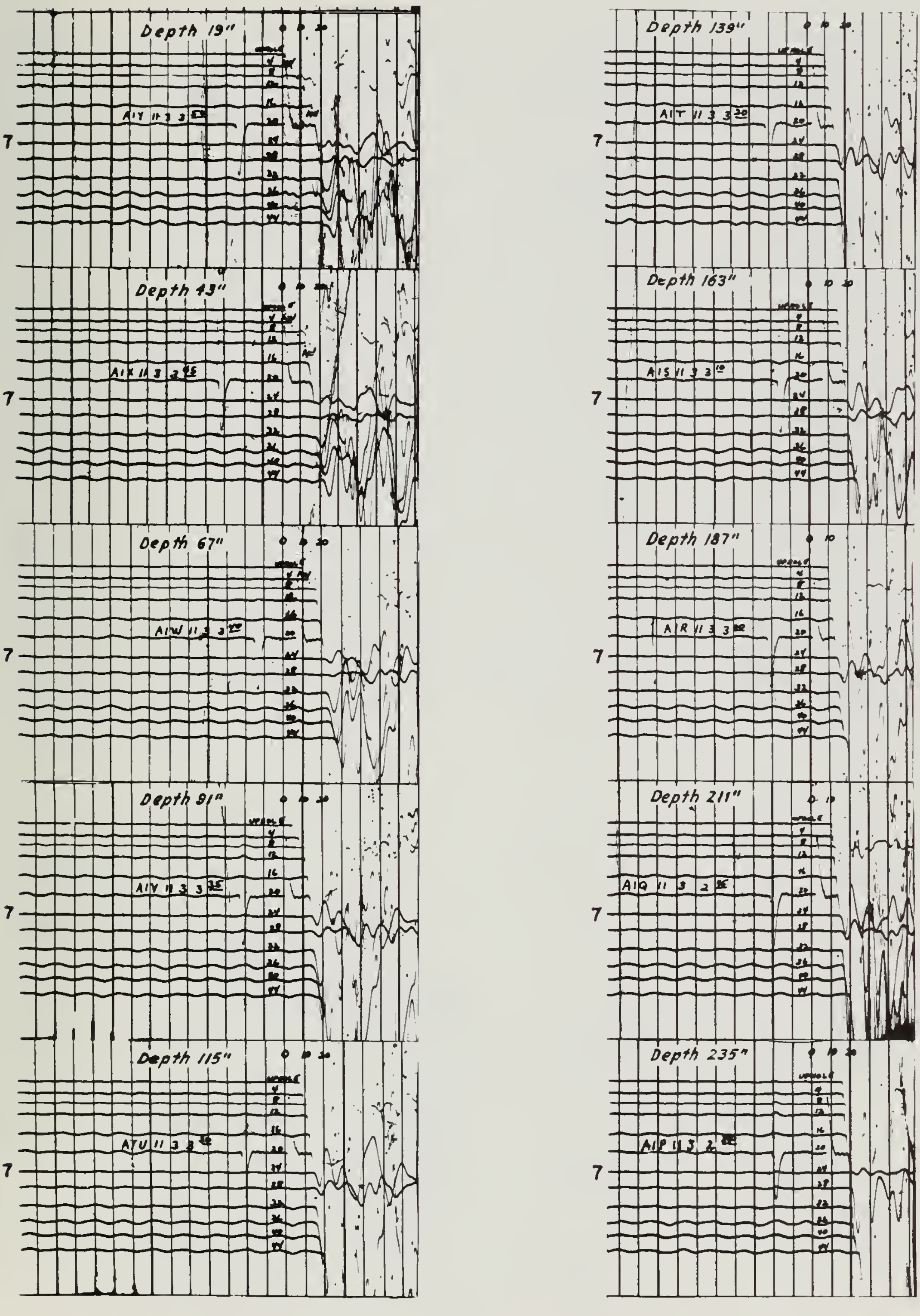

Fig. 4 - Seismic records from test hole 2 for

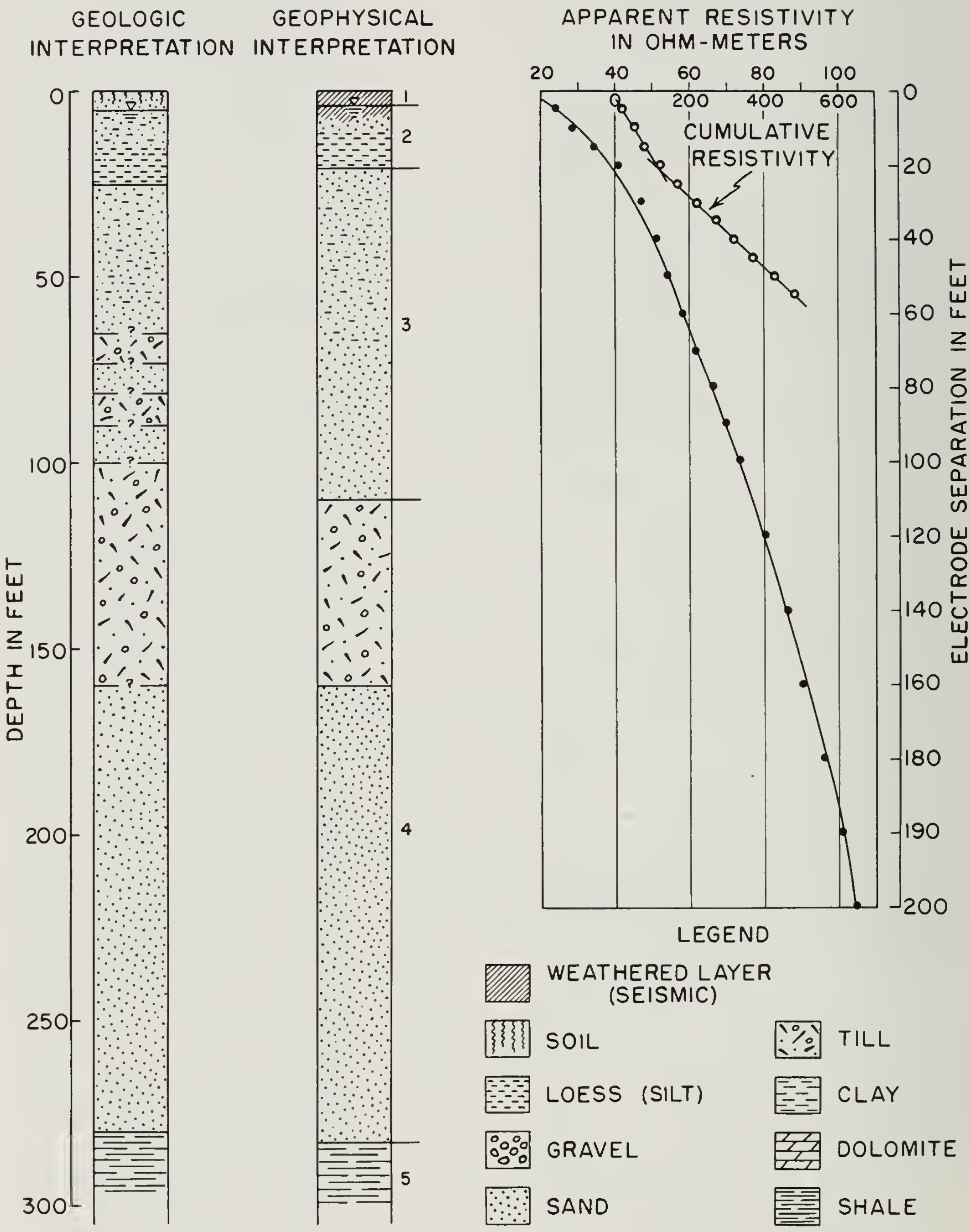

Fig. 5 - Geologic and geophysical interpretations and resistivity curves for station A-11.

may be caused by the wet layer that may decrease the competency of the material and at the same time increase the compressional wave velocity. Increasing the shot depths below 16 feet increased the ray path distance from the shot to the detector, resulting in an increase in the attenuation of the first-arrival compressional waves.

The first arrival of energy at all detectors due to compressional body waves should appear on the traces as downbreaks (upward earth motion), but at small shot-detector distances the first arrival appeared as an upbreak, 180° out of phase (records A1X and A1Y, fig. 4). This upbreak had the velocity of sound in air, indicating a travel path through air and not through soil. The black soil, therefore, had a velocity less than the velocity of sound. Seismic records from hole 2, station A-1, starting with A1Y (19 inches below the uphole geophone) and ending with A1P (235 inches below the geophone) are shown in figure 4.

An expanding electrode resistivity profile between the holes was made with a beginning electrode separation of two feet and two-foot increases thereafter. The resistivity curve (fig. 2) is typical of a three-layer case showing a high resistivity in the surficial silt, low resistivity in the yellowish brown till, and high resistivity in the gray till.

These studies are suggestive of a physical boundary between the yellowish brown weathered till and the gray unweathered till. Geologic studies, shallow refractions, velocity logs, and amplitude logs place a boundary between 14 and 17 feet. From these studies it can be seen that the seismic weathered layer and the geologic weathered layer may coincide, but a more comprehensive study of the relationships between the elastic, chemical, and engineering properties of weathered-unweathered material of a known single till unit must be made before definite conclusions can be drawn. Moisture content may be the cause of the anomalous portion of the geophysical logs, but, again, further study is necessary to be conclusive.

Variations in Velocity Sequences

Station A-11 is in the NE$\frac{1}{4}$ sec. 29, T. 20 N., R. 6 W., Mason County, on the uplands southeast of Havana. The stratigraphy of the drift, as indicated by the scattered well records and outcrops in the region, is shown in figure 5, along with the geophysical interpretation and resistivity curves. About 185 feet of Pennsylvanian rocks, mostly shale, lie below the drift.

Seismic records from refractions shot on drift exhibit many features due to high-speed layers (records C, I, and J, fig. 6), one of which is the long time delay seen on the travel-time curves in figure 7. These time delays are the result of a high-velocity layer (HVL) acting as a high pass filter for waves propagated horizontally and as a low pass filter for waves propagated downward (Press and Dobrin, 1956). Because normal selective attenuation in earth materials occurs more rapidly for high frequencies traveling through the high-speed layer than it does for lower frequencies refracted back up from deeper beds, time delays in a time-distance curve may be produced.

A high-velocity layer caused by a perched zone of saturation is evident at station A-11. The water table is known to be at a considerable depth below the surface, but saturated sand was encountered in the seven-foot seismic shot holes, indicating perched conditions. The perched water probably is caused by a relatively impermeable silt known to exist below the sand.

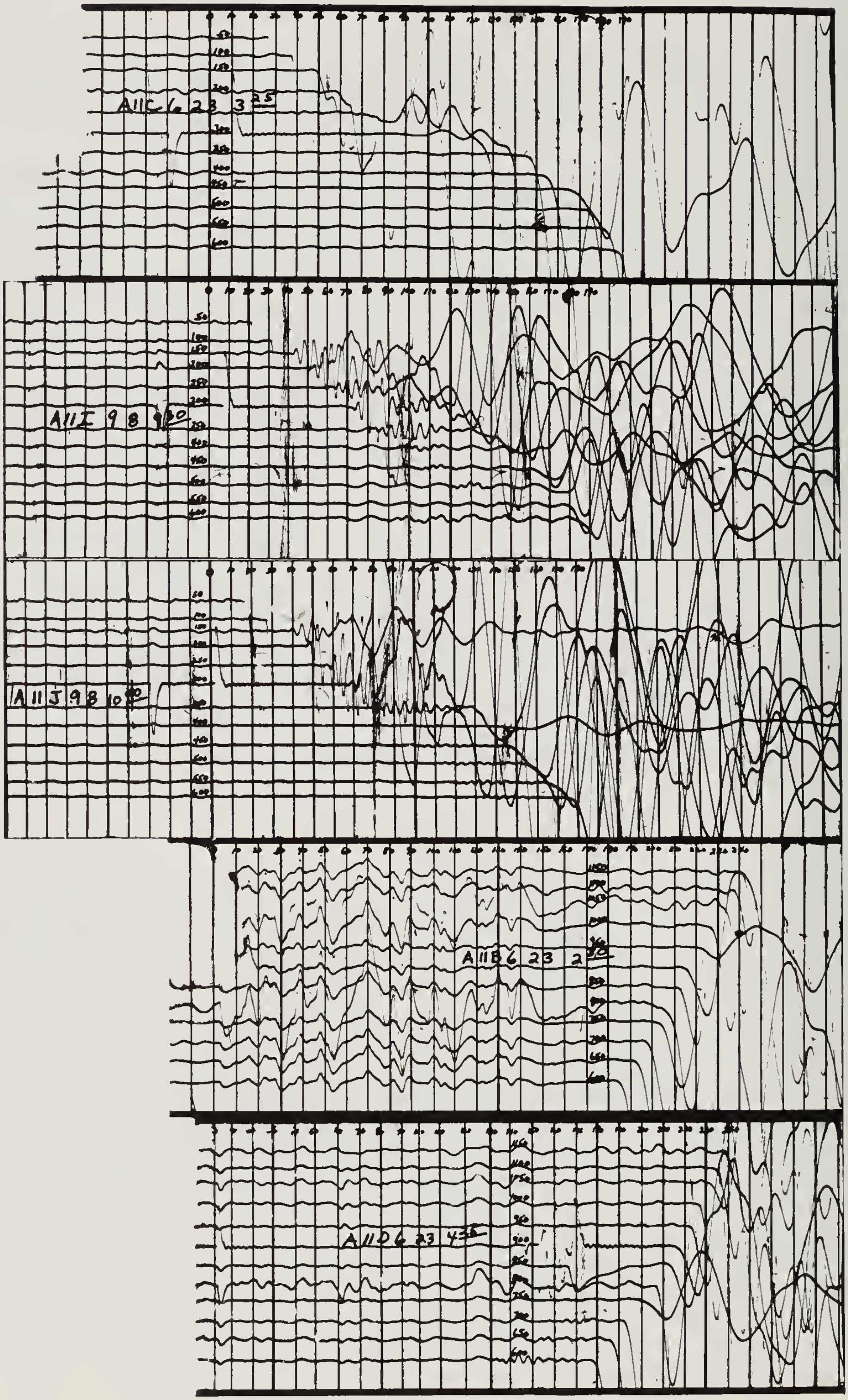

Fig. 6 - Seismic records for station A-11.

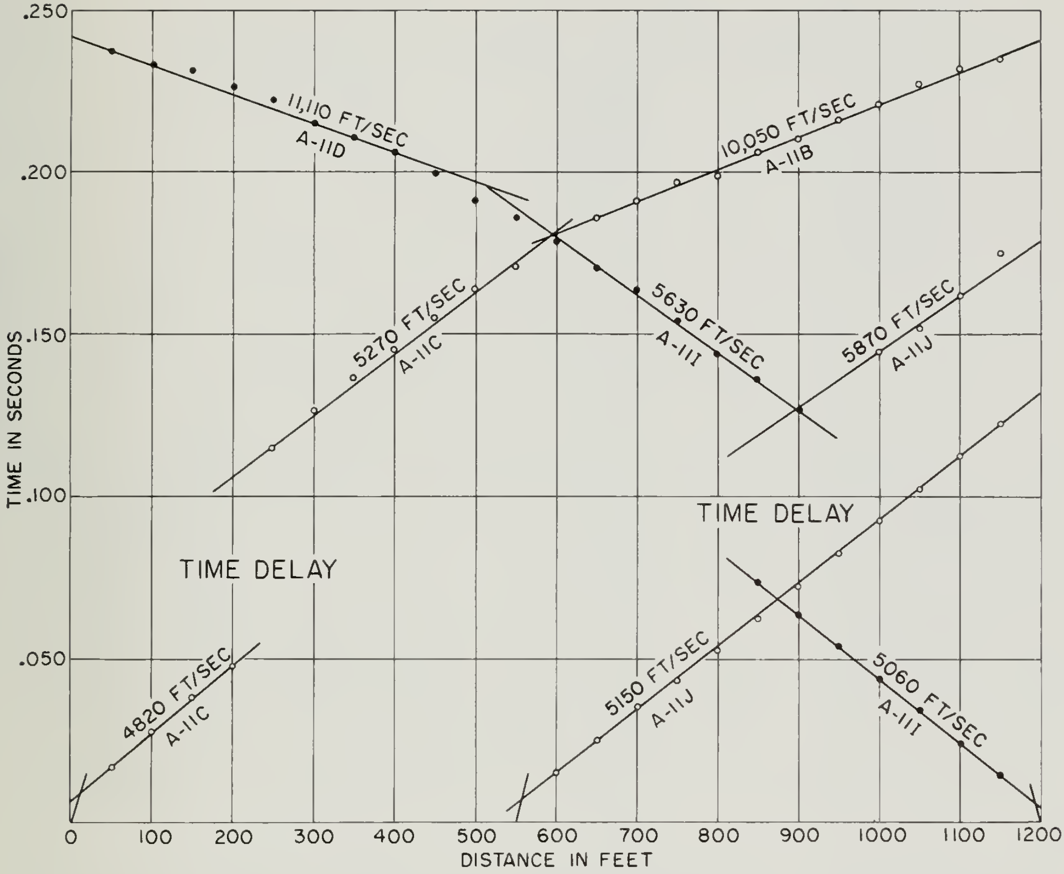

Fig. 7 - Time-distance curves for station A-11.

Record A11C was shot with two pounds of Nitramon at a depth of seven feet. The interval between geophones was 50 feet, with the shot 50 feet from the first detector. Low-frequency energy (60 cps) in the first arrivals appears out to number 6 trace, 300 feet from the shot. Beginning with trace 7, no compressional wave arrivals traveling through the high-velocity layer were recorded.

Compressional body waves traveling through the high-speed layer and continuing out to trace 12 were recorded on record A11J. This record was shot with one pound of Nitramon at a depth of $6\frac{1}{2}$ feet and with the same geophone array as record A11C. Lower predominant frequencies in the first arrivals can be seen on traces nearer the shot than on those farther away, an occurrence common to waves traveling through a high-speed layer. If a velocity reversal did not occur, higher predominant frequencies would be observed on traces near the shot than on those some distance away.

The seismic data indicate five layers are present down to and including bedrock. These layers and their velocities are: 1) a low velocity layer averaging about five feet thick and having a velocity of 1100 ft/sec; 2) the HVL having an average velocity of 5010 ft/sec, probably saturated sand and silt; 3) a low-

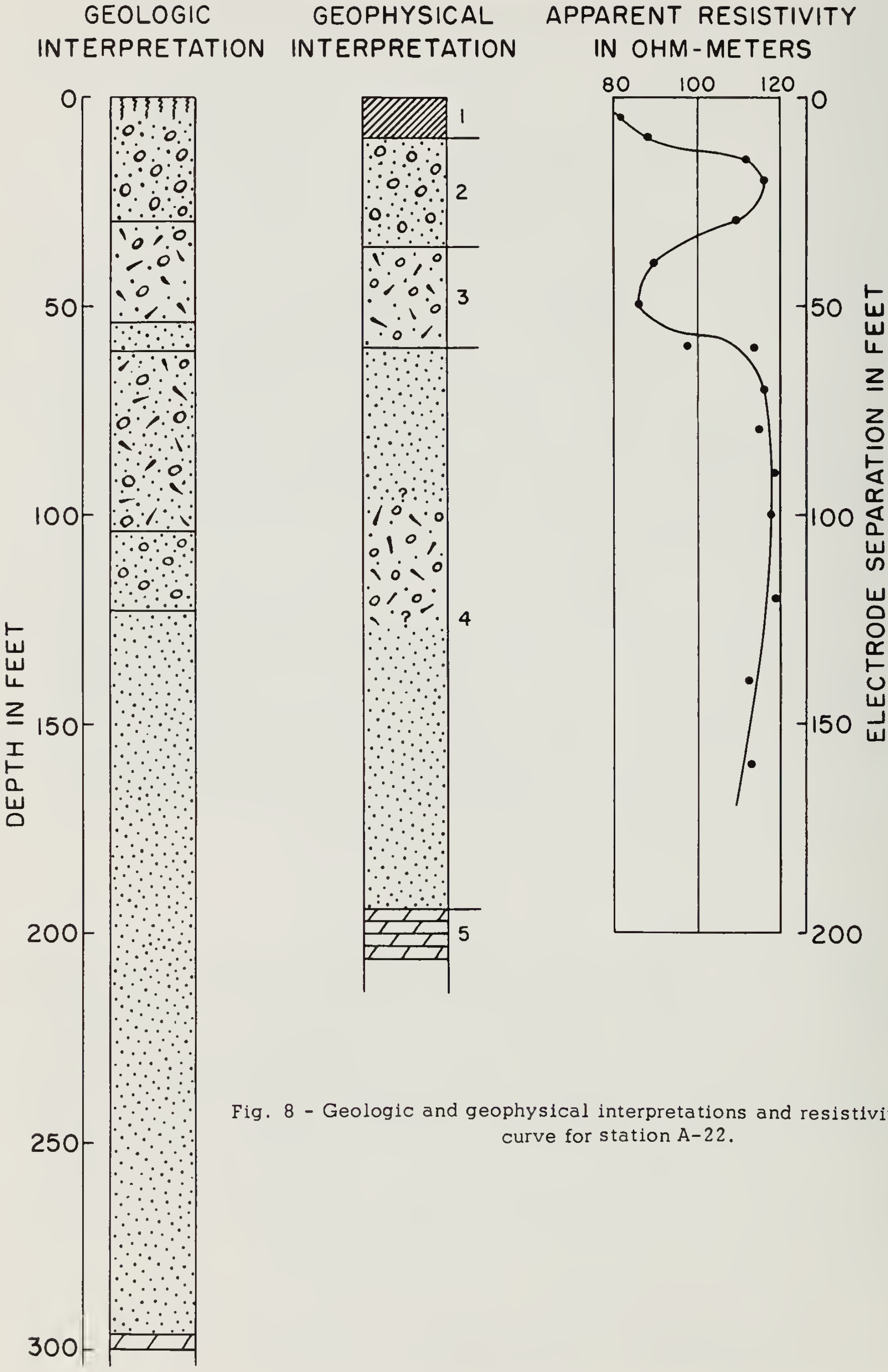

Fig. 8 - Geologic and geophysical interpretations and resistivity curve for station A-22.

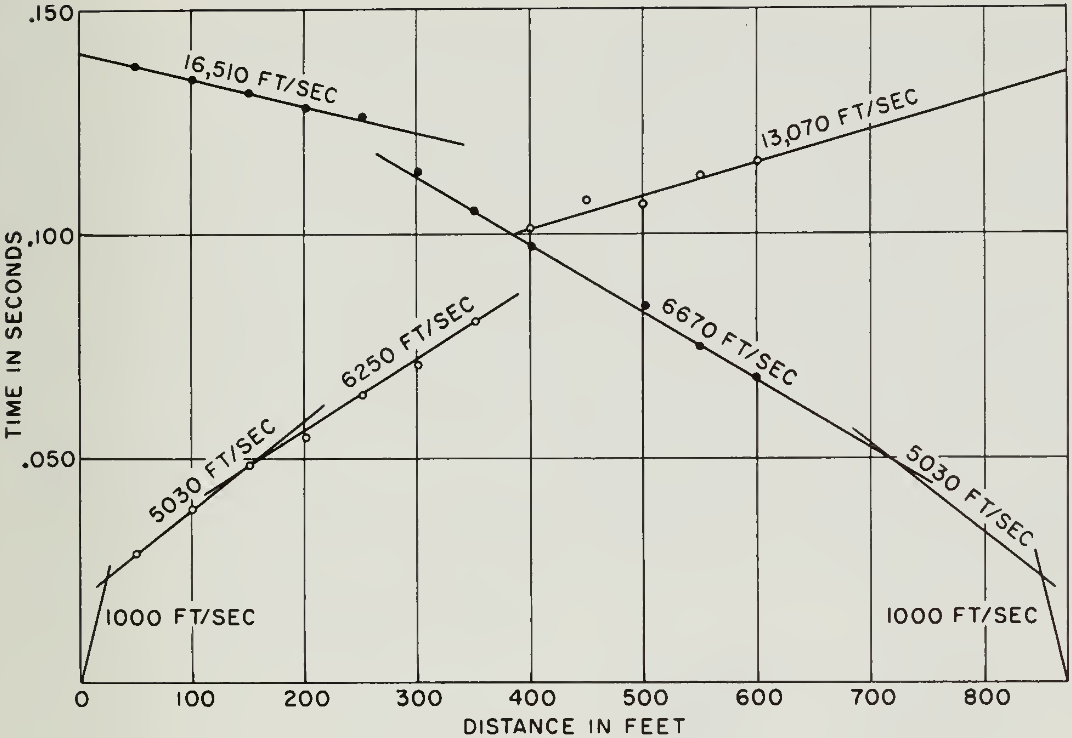

Fig. 9 - Time-distance curves for station A-22.

velocity layer (the LVL) causing a time delay of approximately .055 seconds; 4) a layer having an average velocity of 5590 ft/sec; and 5) the bedrock segment with an average velocity of 10,580 ft/sec. Depths to bedrock and to the HVL-LVL contact cannot be determined from seismic evidence alone because there is no LVL travel-time segment.

Assumptions based on the interpreted geology and on a resistivity curve obtained from this site can be used to find the depth to the base of the saturated sand and silt. Moore's cumulative resistivity method was used to determine a depth of 21 feet to the HVL-LVL boundary. When the thickness of the HVL is known, a travel time through it can be determined. Apparent resistivities and the available geologic information indicate a sand below this boundary. The time delay, therefore, is assumed to be associated with dry sand (LVL) that is overlain by saturated sand and silt (HVL). The time of travel through the LVL can be determined by using an average velocity (3950 ft/sec) of unsaturated sand obtained from other stations throughout the state. A layer with an average velocity of 5590 ft/sec below the LVL probably is a saturated, sandy till. This till was assumed to lie on sand, which in turn lies on bedrock. The combined velocity of the till (5590 ft/sec) and saturated sand (5000 ft/sec) below it is about 5200 ft/sec, because, according to the available geologic information, the basal sand is thicker than the till.

An estimate of the total thickness of the drift can be made if the geologic sequence, the average drift velocity, and the approximate depth to the base of the

HVL as determined by resistivity methods are known. The thickness was found to be approximately 283 feet. A thickness of 280 feet was estimated previously by using well records and other available geologic information.

Station A-22 is in the NW$\frac{1}{2}$ NE$\frac{1}{4}$ sec. 26, T. 45 N., R. 5 E., McHenry County. Stratigraphic control at the station is based on the drillers logs of two water wells that penetrate the bedrock, one 7800 feet north of the station and the other 5280 feet south. The thickness and depth of each drift unit were interpolated from the drillers logs.

The original estimate of drift thickness (fig. 8, geologic interpretation log) was based on an assumption of one of many possible routes of a buried bedrock valley. From a preliminary map of the bedrock topography of the area, the valley was depicted as passing between the two above-mentioned wells. The station was occupied in order to aid in defining the route of the valley more exactly. The resistivity curve and resulting geophysical interpretation also are shown in figure 8.

From seismic, time-distance curves (fig. 9) five layers are evident. These layers can be correlated with part of the geologic section and are, from the surface downward: 1) the low-velocity surface layer, 1000 ft/sec; 2) a gravel deposit

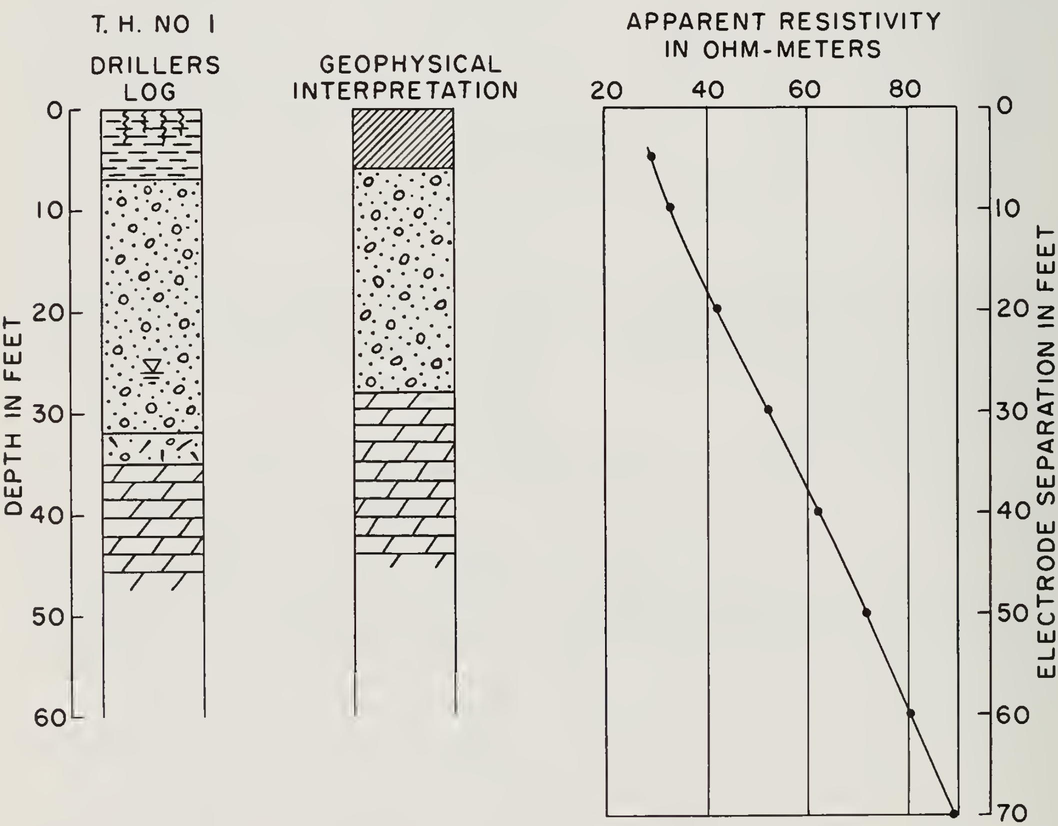

Fig. 10 - Drillers log, geophysical interpretation, and resistivity curve for station A-28.

with a velocity of 5030 ft/sec; 3) a layer of till with an average velocity of 6460 ft/sec; 4) a low-velocity layer indicated by a time delay; and 5) bedrock with an average velocity of 14,790 ft/sec.

The lower till shown on the geologic interpretation (fig. 9) did not transmit a first-arrival refracted wave. This may be because it is thinner and deeper than shown in the original interpretation or because the two tills are not separated by sand at this location. From a geologic point of view, the latter interpretation is preferred. The thick sand below the second till layer is absent from the travel-time curve because of its low velocity.

The thickness of the two units, from the base of the near-surface gravel to bedrock, can be determined by assuming their average velocity is 5500 ft/sec, based on the known geologic sequence of drift materials. Adding the thicknesses of the weathered layer, the near-surface gravel deposit, and the sequence below the gravel deposit, a total depth to bedrock of 194 feet was obtained. This determination, when plotted with other data, indicated that the buried bedrock valley probably lies to the south of station A-22 and that the station is over the northern edge of the buried valley.

Abrupt changes in the apparent resistivity curve (fig. 8) appear to coincide quite well with changes in lithology, but these changes are too great to be caused by lithologic changes. The correlation may be coincidental, with the large changes being caused by some extraneous factor.

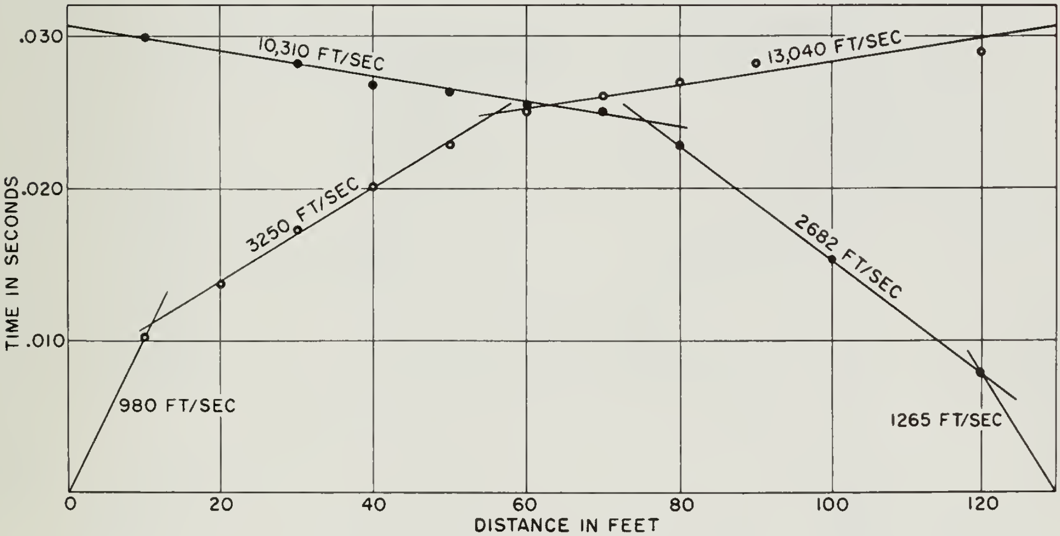

Fig. 11 - Time-distance curves for station A-28.

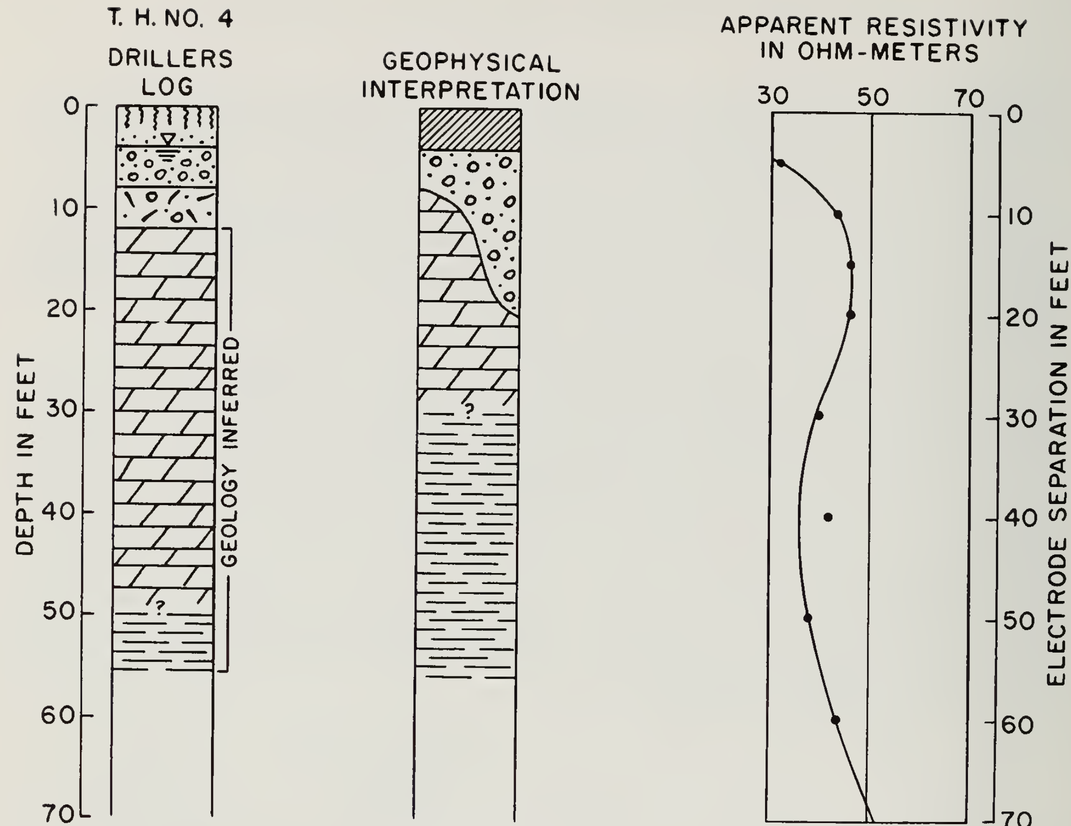

Fig. 12 - Drillers log, geophysical interpretation, and resistivity curve for station A-31.

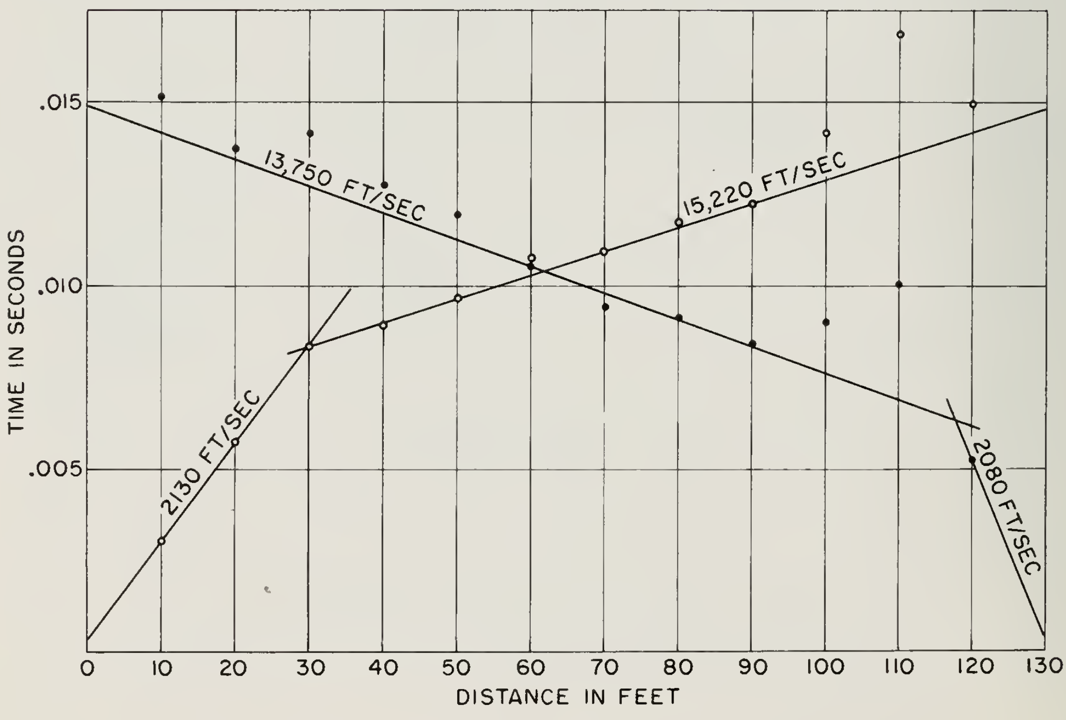

Fig. 13 - Time-distance curves for station A-31.

Other Examples of Integration

Stations A-28, A-31, and A-46 are along the East Branch DuPage River, less than $4\frac{1}{2}$ miles apart and in similar geologic settings. At each station the drift is less than 50 feet thick and is composed chiefly of sand and gravel overlying dolomite of Silurian age. Test wells for hydrologic studies were augered to bedrock near each seismic-resistivity station. The auger was truck-mounted and was incapable of drilling through bedrock. Thin tills found by the drill tests are buried too deeply to be noticeable on any of the time-distance or resistivity curves (Soske, 1959).

Station A-28, near the southeast corner sec. 6, T. 37 N., R. 10 E., Will County, is about 500 feet east of test hole 1. The log of hole 1, the resistivity curve, and the geophysical interpretation are shown in figure 10. The type of drift material and the total drift thickness determined by seismic measurements correlate fairly well with the test hole data. The test hole log shows mainly dry sand and gravel in the drift that cause the low velocities shown on the time-distance curves (fig. 11) and the increasingly high resistivities on the resistivity curve. The depth of the water table cannot be determined by geophysical methods because it lies so near the drift-bedrock boundary. The featureless appearance of the resistivity-depth curve illustrates the hazards in calculating depths in a gravel-dolomite area by resistivity methods.

One of the problems encountered during refraction shooting on thin drift covering an uneven bedrock surface is found at station A-31 (fig. 12). This station

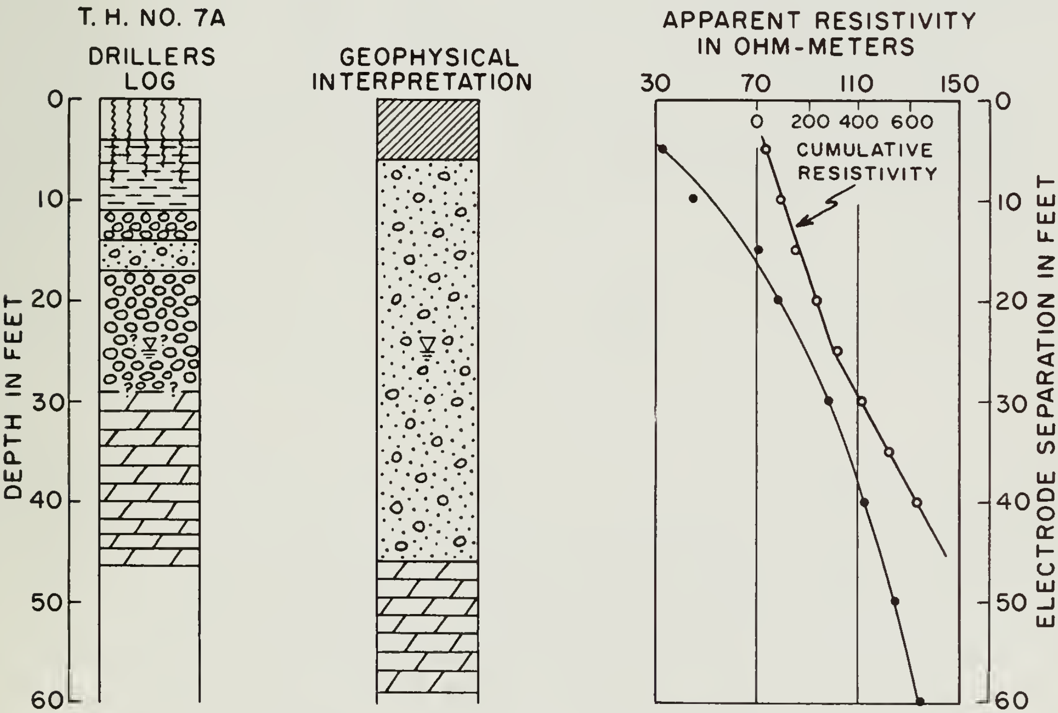

Fig. 14 - Drillers log, geophysical interpretation, and resistivity curves for station A-46.

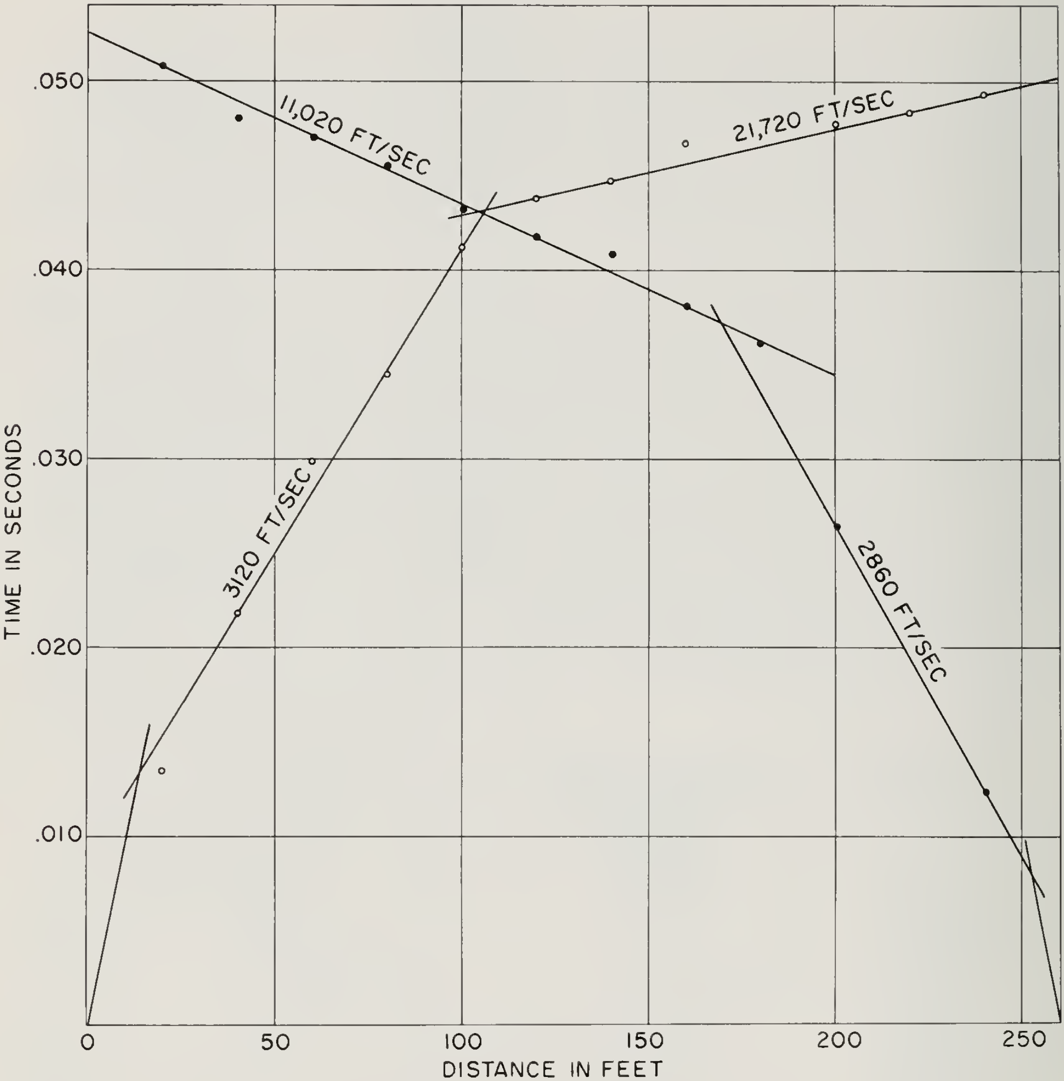

Fig. 15 - Time-distance curves for station A-46.

is located in the SE$\frac{1}{2}$ NE$\frac{1}{2}$ sec. 7, T. 37 N., R.10 E., Will County, with the drill hole located about 800 feet north of A-31 and at about the same land surface elevation. Bedrock was reached at 12 feet in the test hole, below which the geology is inferred from maps and local wells. A static water level was found four feet below the land surface. Very local irregularities in the bedrock surface are indicated by erratic time values on the time-distance curve (fig. 13), causing velocities and depth determinations to be inaccurate (Evison, 1952). High time values on the east end of the reversed spreads could be caused by a small bedrock depression 9 feet deep and 30 feet wide. Drift thicknesses between 8 and 20 feet are indicated by seismic measurements.

The resistivity curve obtained at A-31 (fig. 12) is unusual in that it shows high resistivities in the gravel and upper dolomite and low resistivities deeper in the bedrock. The best explanation for the low bedrock resistivities is the probable presence of the Maquoketa Shale at a relatively shallow depth.

A comparison of the drillers log of test hole 7a and the geophysical log obtained at station A-46 about 200 feet southeast of the test hole shows the lack of agreement sometimes encountered (fig. 14). This station is in the SW$\frac{1}{2}$ SE$\frac{1}{2}$ sec. 22, T. 38 N., R. 10 E., DuPage County, and at approximately the same elevation as the test hole. The test hole log reported impenetrable material at 29 feet that was interpreted as bedrock. Seismic interpretation (fig. 15) shows 46 feet of drift that differs by only four feet from the thickness obtained from a bedrock surface map (Zeizel, unpublished manuscript, 1960) and other wells in the area.

The cumulative resistivity curve (fig. 14) shows a boundary at about 26 feet. This may be caused by the top of the zone of saturation, for the East Branch DuPage River, only 900 feet to the east, is just 30 feet below the level of the station. This also is the approximate depth reached by the auger, which may indicate the presence of a zone of cementation, possibly caused by the water table. The gravel-dolomite contact is masked, as usual, because of similar drift-bedrock resistivities.

CONCLUSION

Glacial drift must be studied in great detail because of its discontinuous, heterogeneous nature. These properties of drift cause difficulties in the application of resistivity, seismic, or geologic methods. Because of their limitations, each method when used alone may yield inaccurate results. If resistivity, seismic, and geologic studies are used jointly, most of the problems associated with each are eliminated because the methods are complementary. The seismic refraction method generally is the most exact of the geophysical methods, but where results are indefinite and geologic control is limited, much can be gained by also using the more rapid resistivity survey.

REFERENCES

Bays, C. A., 1946, Use of electrical geophysical methods in groundwater supply: Jour. Missouri Water and Sewage Conf., v. 16, no. 4, p. 22-35. Reprinted as Illinois Geol. Survey Circ. 122, 14 p.

Bays, C. A., and Folk, S. H., 1944, Developments in the application of geophysics to groundwater problems: Eng. Soc. of W. Pennsylvania, Nov. 1943. Reprinted as Illinois Geol. Survey Circ. 108, 25 p.

Buhle, M. B., 1953, Earth resistivity in groundwater studies in Illinois: AIME Trans., v. 5, no. 4, Mining Engineering, 5 p.

Buhle, M. B., 1957, Uses and limitations of electrical prospecting for water supplies: Illinois Acad. Sci. Trans., v. 50, p. 167-171.

Evison, F. F., 1952, The inadequacy of the standard seismic techniques for shallow surveying: Geophysics, v. 17, no. 4, p. 867-875.

Foster, J. W., and Buhle, M. B., 1951, An integrated geophysical and geological investigation of aquifers in glacial drift near Champaign-Urbana, Illinois: Econ. Geology, v. 46, no. 4, p. 367-397.

Frye, J. C., 1938, Additional studies on the history of Mississippi Valley drainage: unpublished doctoral thesis, Dept. of Geology, State University of Iowa.

Hackett, J. E., 1956, Relation between earth resistivity and glacial deposits near Shelbyville, Illinois: Illinois Geol. Survey Circ. 223, 19 p.

Hubbert, M. K., 1932, Results of earth resistivity survey on various geologic structures in Illinois: AIME Tech. Pub. 463, Class L, Geophysical Prospecting, no. 33, p. 6-7.

Iida, K., 1940, On the elastic properties of soil, particularly in relation to its water content: Tokyo Imp. Univ. Earthquake Res. Inst. Bull., v. 18, no. 4, p. 675-691.

Johnson, R. B., 1954, Use of the refraction seismic method for differentiating pleistocene deposits in the Arcola and Tuscola Quadrangles, Illinois: Illinois Geol. Survey Rept. Inv. 176, 59 p.

Lester, O. C., Jr., 1932, Seismic weathered or aerated surface layer: Am. Assoc. Petroleum Geologists Bull., v. 16, no. 12, p. 1230-1234.

Meidav, T., 1960, An electrical resistivity survey for ground water: Geophysics, v. 25, no. 5, p. 1077-1093.

Moore, R. W., 1945, An empirical method of interpretation of earth-resistivity measurements: AIME Trans., v. 164, Geophysics, p. 197-223.

Press, F., and Dobrin, M. B., 1956, Seismic studies over surface layer: Geophysics, v. 21, no. 2, p. 285-298.

Soske, J. L., 1959, The blind zone problem in engineering geophysics: Geophysics, v. 24, no. 2, p. 359-365.

Warrick, R. E., and Winslow, J. D., 1960, Application of seismic methods to a groundwater problem in northeastern Ohio: Geophysics, v. 25, no. 2, p. 505-519.

Williams, F. J., 1946, Notes on shot point procedure: Geophysics, v. 11, no. 4, p. 443-456.

Woollard, G. P., and Hanson, G. F., 1954, Geophysical methods applied to geologic problems in Wisconsin: Wisconsin Geol. Survey Bull. 78, Scientific Series, no. 15, p. 19-119.

Zeizel, A. J., 1960, Ground-water geology of the shallow aquifers in DuPage County, Illinois: unpublished doctoral thesis, Dept. of Geology, University of Illinois.

Illinois State Geological Survey Circular 323
23 p., 15 figs., 1961

CIRCULAR 323

ILLINOIS STATE GEOLOGICAL SURVEY

URBANA

2M—45752—11-61 572