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Shallow Seismic Investigation of Hydrogeologic Problems in the Brazos River Alluvium, Texas A&M Plantation, Burleson County, Texas

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SHALLOW SEISMIC INVESTIGATION OF HYDROGEOLOGIC PROBLEMS IN THE BRAZOS RIVER ALLUVIUM, TEXAS A&M PLANTATION, BURLESON COUNTY, TEXAS

By

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ABSTRACT

Twenty-two shallow, reversed, seismic refraction profiles were conducted in the Brazos River floodplain to test the feasibility of using seismic methods to provide hydrogeologic information in this province. The specific objectives were to map the total and saturated thickness of the alluvial deposits and to outline gravel lenses within the alluvium.

It was found that the water table was the only interface at which the acoustical properties of the deposits above and below changed sufficiently to be mapped by seismic methods. The alluvial deposits above the water transmitted compressional waves at an average velocity just greater than the velocity of sound in air. The saturated alluvial deposits transmitted seismic waves at an average velocity of slightly greater than the velocity of sound in water.

The saturated alluvial deposits and the bedrock appear to have a continuous increase in velocity with depth rather than a significant change in acoustical characteristics.

The seismic measurements in themselves could not delineate gravel lenses within the alluvium. However, the zones of greater permeability were indicated on the resulting contour map of the water table by the areas of gentle

gradients. Zones of greater permeability in the alluvial deposits are probably gravel lenses.

INTRODUCTION

The object of this study was to investigate the hydrogeology of the alluvial deposits of a portion of the Texas A&M Plantation using the techniques of shallow refraction seismology. An attempt was made to use these methods to map the total thickness and the saturated thickness of the alluvial deposits and to outline zones of gravel lenses within the alluvium. Domestic water wells can be made in the alluvium almost anywhere in the floodplain, but the high discharge irrigation wells are limited in location by the erratic distribution of gravel lenses. Although geophysical techniques do not result in as much detailed or as accurate information as does test drilling, they may outline the more favorable areas for well locations without the necessity of the present procedure of extensive test drilling.

The area chosen for this investigation was located on the Texas A&M Plantation, Burleson County, Texas. The Plantation is that portion of the University farm in the Brazos River floodplain, locally known as the "Brazos Bottoms." The test site, an area approximately 3,000 feet by 1,200 feet, was bordered by the Brazos River on the northeast side (figure 1). The topography is fairly flat, the maximum relief being about 6 feet.

The standard seismic refraction technique was followed

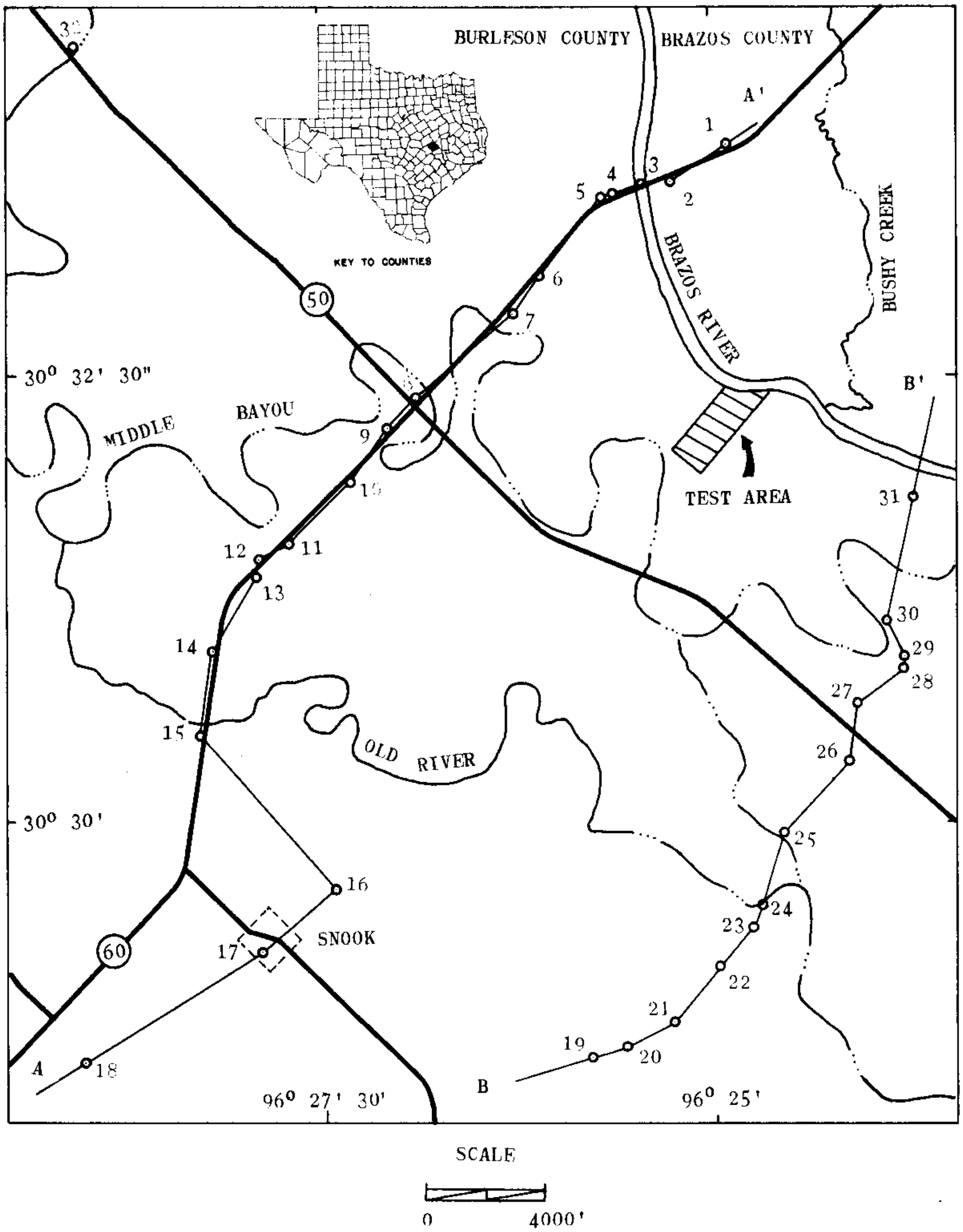


FIG. 1.—INDEX MAP SHOWING LOCATION OF TEST AREA AND POSITIONS OF SECTIONS A-A' AND B-B' (FIG. 3).

and the velocities and depths were calculated by standard mathematical solutions of the travel-time plots. A discussion of the general theory and methods are given in Jakosky (1950), Dobrin (1960), and Griffiths and King (1965). Briefly, the seismic method consists of the generation of an elastic pulse near the surface of the ground and recording the resulting motion of the ground at nearby points on the surface. Measurements of the time intervals between the generation of the pulse and its detection at the geophones at various distances give the velocity of propagation of the energy in the ground. Where the subsurface structure is simple, the values of elastic wave velocities and the positions of boundaries between regions of differing velocity can be calculated from the time-distance data. These calculations require a simple model of the structure; that is, the velocity zones should be homogeneous and isotropic with planar interfaces between each zone. It should be remembered that a seismic profile is an average profile of the refracting horizon, with the highs and lows being smoothed out. Hence, the difference in depth between the true surface and the seismic interfaces may become a relatively large percentage of the total depth in shallow studies.

PREVIOUS INVESTIGATIONS

Brazos River Alluvium

Geological investigations of the alluvial deposits of the Brazos River valley are limited. The valley in the area of investigation is a typical alluvial valley developed in the West Gulf Coastal Plain. This section of the Coastal Plains Province is bounded upstream by the Balcones fault zone, approximately located along the Bosque-McLennan county line. Northwest of this line, the Brazos valley has been characterized by Stricklin (1961) as a bedrock valley, which has been developed across an upland of Paleozoic and Cretaceous sedimentary rocks.

The alluvial valley portion has been developed in strata ranging in age from the Late Cretaceous to Recent. The Cretaceous rocks are composed predominately of marl, shale, and sandstone; the Tertiary rocks are chiefly shale, clay, and sand. The portion of the valley in the Coastal Plain is characterized by its width and gentle topography as compared to the narrow, hilly bedrock valley portion which lies upstream from the Balcones fault zone. The meandering stream pattern and the alluvial walls and floor of the river are also distinctive features of the alluvial valley portion. This type of valley has been geologically investigated along other streams, such as the Red River in Louisiana (Newcome, 1960) and the Mississippi River

(Fisk, 1944, 1952).

Deussen (1924) described six terraces on the Brazos River between the inner and outer margins of the Coastal Plain. Terrace No. 1 is present along most of the length of the valley south of Waco, ranging from 30 to 35 feet in height above the river bed. This terrace, composed of red, sandy clay, is subject to overflow in times of flood. Terrace No. 2 ranges from 40 to 55 feet above the river bed and is composed of red, sandy clay with more or less gravel in its basal part. Terrace No. 2 has been dated as Early Aftonian Age of the Pleistocene, based on a vertebrate fauna found at Munson Shoals in Brazos County. Terraces numbers 3 through 6 range in height from 70 to 220 feet above the bed of the river. These older terraces are generally composed of sand or gravel and are areally limited in their present distribution.

A generalized study of the ground-water geology of the Brazos alluvium in the Coastal Plain was made by Cronin and others (1963). Recharge to the alluvium principally is from irrigation, and by underflow from adjacent and underlying formations. The annual average rainfall in the floodplain and lower terraces varies from about 34 inches in the northern portion to about 48 inches in the coastal area. Part of this precipitation is added to the water in storage by infiltration. Recharge also occurs in places during high water or floods when surface water moves into the alluvium as bank storage. The alluvial deposits are also recharged by water moving from the underlying formations to the river. At present the quantity of underflow moving from the formations is unknown. Based on scattered and rather meager data, the recharge to the alluvium in excess of the pumpage between 1957 and 1960 was in the order of 250,000 acre-feet.

The movement of ground water in the alluvium was found to be toward the Brazos River and slightly downstream. As best as could be determined, at no area along the river could flow in the opposite direction be determined during normal water stages. The rate of water movement is unknown according to Cronin and others (1963).

Discharge from the alluvium is by springs and seeps along the Brazos River, by transpiration and evaporation in areas where the water table is at or near the surface, and by pumpage from wells. The major portion of the water produced from wells is used for irrigation, as no public water supplies and few domestic wells obtain water from the alluvium. Irrigation pumpage varied from about 200,000 acre-feet a year during the 1953 to 1956 drought to about 20,000 acre-feet a year in 1959 and 1960 (Cronin and others, 1963, p. 116).

Brazos River Discharge

The flow of the Brazos River in the Burleson-Brazos County Area is regulated by upstream reservoirs on the Brazos and by the Belton Reservoir on the Leon River. A hydrograph of the maximum, mean, and minimum discharge by months for the calendar year 1965 is shown in figure 2. During this period, the peak flow was measured on May 19 at 134,000 cubic feet per second and a river elevation of 234.3 feet. The low flow for this same period was measured on September 10 at 680 cubic feet per second and a river elevation of 195.6 feet (Twichell, 1966). These observations were made approximately 10.4 miles upstream from the test area at the U. S. Geological Survey streamflow station 8-1090. The drainage area of the river at this location is 38,400 square miles, of which approximately 9,240 square miles are probably non-contributing (Eisenhuth, 1965).

Shallow Refraction Seismology

The first refraction seismograph designed for shallow exploration was built by the Bureau of Public Roads in 1933. Shepard (1935) reported on the experimental use of this instrument in determining the presence and location of consolidated rock at two bridge sites in Washington, D. C. and the depth of overburden at a quarry.

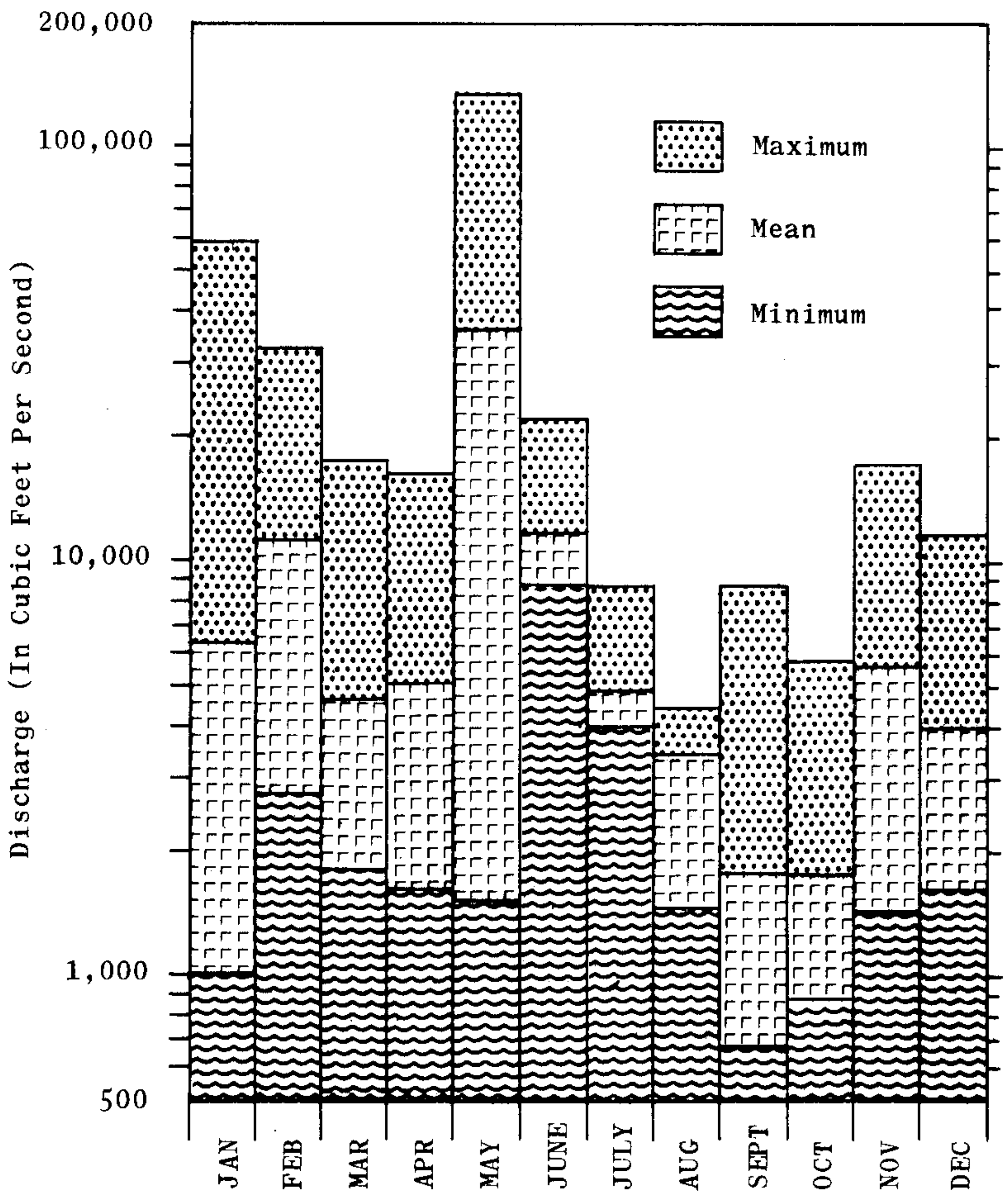


Fig. 2.—Hydrograph of the maximum, mean, and minimum discharge of the Brazos River near Bryan for the calendar year 1965. From Twichell (1966).

This instrument was a four-channel seismograph; dynamite was used as the energy source.

In 1934, Partlo and Service built a single-channel seismograph to determine the depth of overburdens of 100 feet or less. They used both a sledgehammer and an explosive energy source to determine the overburden thickness at three sites and the location of a concealed fault.

Shepard (1939) reported on the results of seismic methods as applied to construction projects. Using the Bureau of Public Roads seismograph, profiles were made at several locations in the United States to determine overburden thickness, depth to the water table in sandy soil, depth to fresh igneous rock, and to explore for damsite foundations.

In 1940, the Waterways Experiment Station explored eight damsites in the Ozark highlands of Arkansas and Missouri. These studies were made to determine the thickness of the alluvial deposits and of the weathered portion of the bedrock. On the uplands, a velocity increase of approximately 2,000 feet per second within the unconsolidated deposits was attributed to a transition from loose, aerated, surface material to the moist, more compacted, residual overburden. In the floodplain, a velocity increase on the order of 3,500 feet per second was found at a depth which corresponded closely to the water table. It was concluded that the refraction method could, in general, outline the rock surface, but not define the "firm rock line" or channels within the unweathered rock.

In connection with the Corps of Engineers exploration program for flood control and navigation improvements during the period 1936 to 1939 (Shepard and Wood, 1940), Burwell (1940) was apparently the first to report on the determination of ground-water levels in alluvial deposits. During preliminary foundation investigations for damsites in the Ohio River Basin, it was found that the elevation of the water table could be determined within small limits of error. The velocity varied from 1,100 to 3,170 feet per second in the dry alluvium and from 4,000 to 6,000 feet per second in the saturated alluvium. The bedrock velocity ranged from 8,210 to 15,000 feet per second. Burwell noted that the velocity in the saturated alluvium averaged slightly greater than the average seismic velocity in water. Because the velocity obtained in the saturated material appeared to be influenced primarily by the properties of the water rather than by the properties of the material, it was concluded that, generally, seismic methods could not differentiate satisfactorily between good and poor aquifers.

The Waterways Experiment Station (1943) investigated the use of shallow refraction seismology for determining the character and depth of the alluvium in the lower Mississippi valley. A total of twelve seismic profiles were made at three sites in the floodplain at which borings had been completed into the underlying Tertiary formation. At all three sites, the Tertiary deposits were marine silts, clays, or silty clays. On all of the profiles, the depth to the top of sand or water table could be determined; however, only two profiles at one site yielded data from which calculations of the alluvial thickness could be correlated with the actual depth to the Tertiary formations. The seismic velocities ranged from 600 to 1,000 feet per second in the dry alluvium and from 5,000 to 6,000 feet per second in the saturated alluvium. The upper Tertiary materials were found to have seismic velocities of about 4,000 feet per second, a velocity comparable with that determined for other relatively stiff marine clays, except for the localized 7,000 and 8,000 feet per second velocities found on two profiles. It was concluded that this method was not suitable as a supplement to boring exploration in the alluvial valley.

Linehan and Keith (1949) used seismic refraction in exploration for ground water at five sites in Massachusetts and Connecticut. Their work consisted mainly of determining the thickness of glacial drift overlying a crystalline bedrock, with velocity measurements made in the unconsolidated deposits used to determine the location of potential aquifers.

In 1952, Stickle and others discussed the general use of refraction seismology in water-table location and bedrock-depth determination.

In a series of papers, Gough (1952; 1953; Gough and van Niekerk, 1957) reported on a shallow seismic refraction instrument and the results of several investigations made in South West Africa. These investigations included five bedrock-depth determinations in river valleys and a study of one building foundation site. The velocity increase between the alluvium and the bedrock was of the order of 10,000 to 15,000 feet per second. In all but one of these surveys, the initial straight line on the travel-time graph does not pass through the origin. This positive time intercept was attributed to a thin, very low velocity, aerated, soil zone (Gough, 1953; Gough and van Nieker, 1957). At one site, the first arrivals corresponded to the velocity of sound in air, hence only a maximum depth could be computed (Gough and van Nieker, 1957). At all of the sites, the depths known from borings agreed closely with the seismic depths.

In 1954, Woollard and Hanson published the results of a series of geophysical surveys made in Wisconsin between 1948 and 1954. These surveys involved the use of several geophysical methods to study mineral exploration problems, engineering problems, and ground-water supply investigations. Shallow refraction seismology was used in eleven ground-water supply studies which were mainly concerned with the location of buried valleys in crystalline rocks or in Paleozoic limestones or sandstones. In seven of the sites studied, a velocity increase of approximately 1,000 to 3,000 feet per second was found between the dry and saturated glacial deposits.

Shallow refraction methods were used by Pakiser and Black (1957) to explore for channels in the Monument Valley of Arizona and Utah. The velocity increase between the Moenkopi Formation bedrock and the Shinarump fill was of the order of 8,000 feet per second. A delay time analysis was used to outline small channels that might have localized accumulations of uranium ore. Refractions from saturated rocks were not observed.

Both reflection and refraction techniques were used by Pakiser and Warrick (1956) and Warrick and Winslow (1960) to determine the location, depth, and cross-sections of buried valleys in northeastern Ohio. These valleys are cut in Pennsylvanian, Mississippian, and Devonian sedimentary rocks and are buried by glacial drift. The increase in velocity from drift to bedrock was about 5,000 feet per second. The presence of a water table was not indicated, but a low velocity surface zone (about 2,000 feet per second) was noted.

Bonini and Hickok (1958) used refraction techniques in New Jersey to outline a Pleistocene sand and gravel aquifer which occurs as a channel in the Triassic Brunswick Formation. A velocity increase of approximately 6,000 feet per second marks the interface between the unconsolidated material and the bedrock. A velocity increase at the water table was not noted. Gill and others (1965) extended this survey westward. Their experiments resulted in the outlining of two separate channels. The velocity in the dry, unconsolidated deposits varied from 1,100 to 3,000 feet per second; in the saturated, unconsolidated deposits it ranged from 4,500 to 8,300; and in the bedrock it ranged from 10,700 to 17,900.

Moore (1961) briefly described the refraction seismic and electrical resistivity methods as applied to highway design and maintenance. Examples of the use of the seismic method in investigations of slope design, tunnel sites, bridge foundations, damsites, and potential quarry sites were reported.

The Desert Research Institute of the University of Nevada has investigated the use of geoelectrical and seismic methods as applied to hydrogeologic problems. Resistivity measurements were used to determine the lithology of valley-fill sediments, and seismic methods were used to determine the shape of the bedrock valley and the depth to the water table. Refraction profiles were made at eight localities in the Humboldt River basin (Dudley and McGinnis, 1962) and at three other sites in northern Nevada (McGinnis and Dudley, 1964). In general, the refraction method was very successful in determining the depth to the water table, where the water table was present, and to the bedrock. Failure to obtain primary arrivals from the saturated alluvium was attributed to the relative thickness of the layers and the effect of the extent to which the velocity of the third layer exceeded that of the second. This problem of intermediate "blind zones" in three-layer models has been discussed by Soske (1959) and Green (1962). Where the water table was detected, the seismic velocity in the dry alluvium was approximately 2,000 feet per second, and in the saturated alluvium it was approximately 5,000 to 6,000 feet per second. The bedrock velocities varied from 10,000 to 16,500 feet per second.

Shaw (1963) outlined the use of geophysical methods in ground-water problems and the extent to which the various methods were being used in several countries outside of Great Britain, mainly in the Tropics. He concluded that the depth to the water table may be obtained in areas limited to relatively small thicknesses of unconsolidated material, with the most important use of shallow refraction seismology being in its ability to delineate hydrologic features with a greater degree of accuracy than other methods.

In conjunction with geoelectric investigations in Santa Clara County, California, Zohdy (1965) investigated the use of refraction seismology to locate permeable strata for the artificial recharge of ground water. Within the depths of interest however, the alluvial strata lacked a sufficient velocity contrast.

Burroughs and others (1965) used shallow refraction seismology to explore for stream gaging sites and to predict the type of deposits to be encountered during construction. They also used refraction methods to test for uniform subsurface conditions in the selection of large plots of forest lands for experimental treatment. In a ground-water study in the Bitterroot Mountains of Idaho, refraction methods were used to profile the basement surface. Results of this work indicated average velocities of 1,612 feet per second in dry soil, 4,884 in saturated deposits, and 9,629 in the crystalline basement.

GEOLOGY

The Brazos River floodplain is approximately nine miles wide to the north of the area of investigation, narrowing to about 5.5 miles just to the south of the test area. The Brazos River borders the east side of the floodplain. The alluvium in this area varies in thickness across the valley from a featheredge to 72 feet, with the thinner section generally on the west side. In the test area the alluvium varies from 50 to 75 feet in thickness. The test area was probably on Deussen's (1924) Terrace No. 2.

East-west trending cross sections (Cronin and Wilson, in press) based on well logs and supplementary drillings are shown in figure 3. Section A-A' is along Farm Road 60, approximately 8,000 feet upstream from the test area; section B-B' is across the Chance Farm, about 6,500 feet downstream. The locations of these sections are shown on figure 1.

The alluvial deposits consist of clay, silt, sand, and gravel in proportions that vary locally but remain reasonably constant throughout the flood plain. The alluvium is generally gradational from clay or silt at the top to sand and gravel at the base. In general, two to three feet of red to brown clay soil is developed over most of the river valley. Below this soil zone, there is

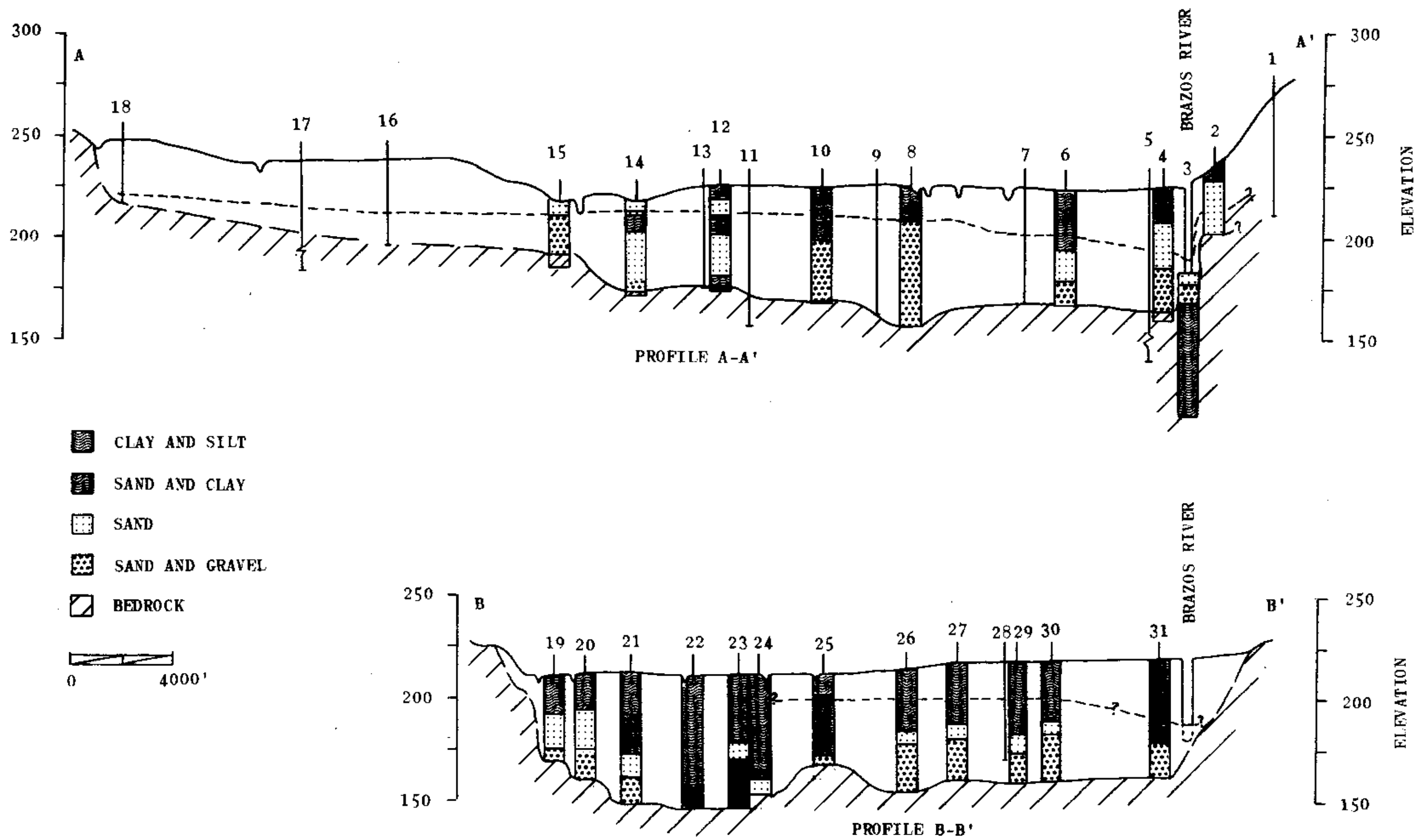


FIG. 3.—GEOLOGIC SECTIONS OF BRAZOS RIVER ALLUVIUM BASED ON WELLS AND TEST BORINGS. TRACES ARE SHOWN ON FIG. 1. FROM CRONIN AND WILSON (IN PRESS).

a layer of red to reddish-brown clay which is usually slightly sandy or silty. This zone varies in thickness from approximately 5 to 55 feet, where it is encountered. Below the clay, a 5- to 25-foot layer of friable, red brown, locally argillaceous sand is usually present. This sand is generally very fine grained at the top and grades to medium grained at the base. The lowest alluvial unit is a zone of sand and gravel, which varies in thickness from 5 to 50 feet, where it is present. The gravel fraction is generally 50 percent or less, and most of the particles are one-half inch or less in diameter. The sand in this unit varies from fine to coarse grained.

The alluvial deposits in the Plantation area are underlain by the Eocene Yegua Formation of the Claiborne Group. The Yegua Formation (Smith, 1958) is nonmarine and consists of about 50 percent sand which is fine to medium grained and laminated to massive; 48 percent sandy clay and clay which is dark chocolate-brown, to gray, to greenish-gray; 1 percent lignite; and 1 percent bentonite. The sands were deposited as alluvial fans built up by the coalescing of stream levees and deltas; the clays were deposited in fresh-water coastal lakes and back swamps; and the lignites were deposited in fresh-water swamps.

Auger samples of the Yegua Formation underlying the alluvium near the test area were taken by Cronin and Wilson (in press). The samples were taken in conjunction with the test borings at the locations shown on figure 1. All of the six samples available were laminated, light and dark gray, clayey, silty, very-fine-grained sandstones, with more or less carbonaceous material. Most of the samples showed considerable cohesion when dry, were poorly sorted, and swelled in water, indicating the presence of montmorillonite. Samples 11 and 12 were composed of about 15 percent clay, 60 percent silt, and 25 percent sand, with sample 11 containing a larger amount of carbonaceous material. Sample 14 was a well sorted, fine sand, composed of approximately 15 percent clay and silt and 85 percent sand. At location 15, the sample was highly porous and was composed of approximately 10 percent clay, 15 percent silt, and 75 percent sand, with very minor amounts of carbonaceous material. Sample 25 was composed of approximately 5 percent carbonaceous material, 5 percent clay, 10 percent silt, and 80 percent sand, with a few well rounded and highly angular quartz grains that attained a maximum size of 0.5 mm in diameter. Sample 32 was slightly calcareous and was composed of approximately 15 percent clay, 25 percent silt, and 60 percent sand, with abundant carbonaceous material.

EQUIPMENT

The seismic profiles were made with a Geo Space Corporation GT-2 portable refraction system. The seismograph was equipped with six information recording channels and a time-break trace. The signals were recorded on Polaroid type 47 photographic film (3,000 ASA speed) with a standard polaroid #95 camera back. The seismic record was projected on the film by a moving optics system. Each data trace had a separate high-gain, printed circuit amplifier and gain control. The frequency response curves for the amplifiers are shown in figure 4. Time lines were recorded at 10 millisecond intervals with an accuracy of better than 0.5 millisecond. The standard recording times of the instrument were 2, 3, or 4 tenths of a second. These times are approximately equivalent to 17, $13\frac{1}{2}$, or 10 inches per second paper speed. The power supply for the amplifiers and the optic system consisted of eight $1\frac{1}{2}$-volt dry cell batteries for each. A 90-volt battery charged a blaster condenser during the operating cycle of the optics system. The total weight of the seismograph is 39 pounds (Geo Space Corporation Catalog).

The geophones used were HS-1 miniature transducers manufactured by the Geo Space Corporation. These vertical detectors were of a floating-coil type, which measure the velocity of ground motion and were virtually insensitive

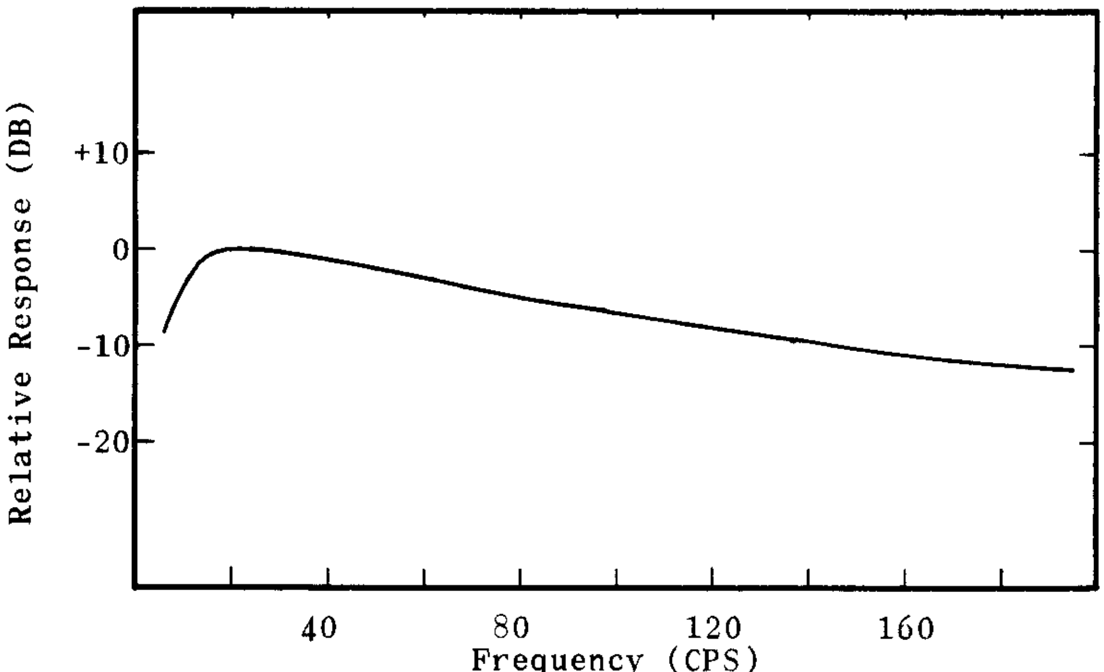


Fig. 3.—Frequency response curve for the GT-2 Refraction System amplifiers. From information furnished by the Geo Space Corporation.

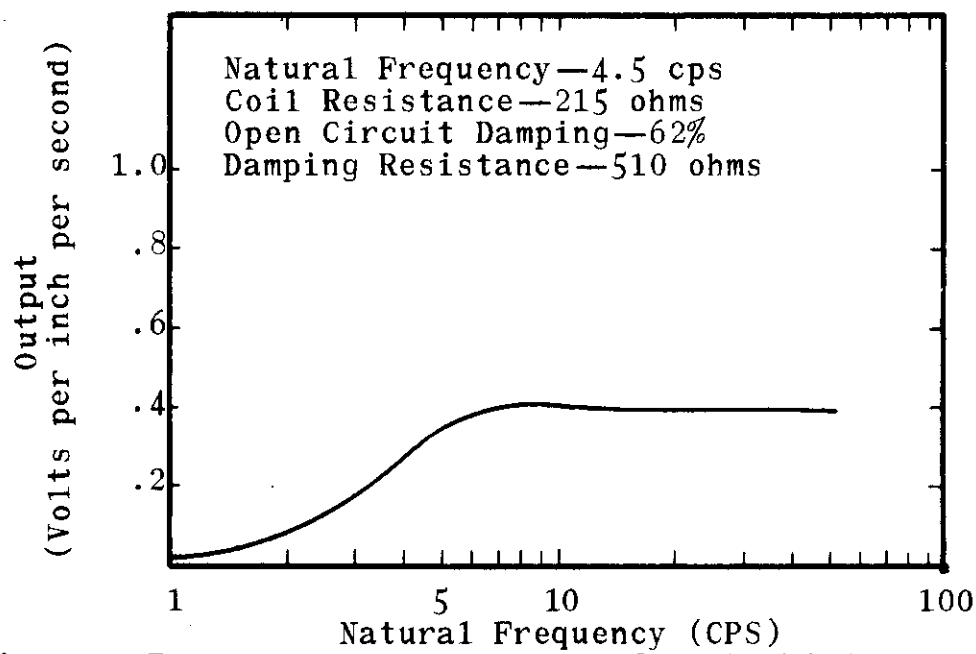


Fig. 4.—Frequency response curve for the HS-1 geophone. From the catalog of the Geo Space Corporation.

to unbalance in the seismometer-cable system and external field pickup. The geophones had a natural frequency of 4.5 cps and an impedance of 500 ohms, which resulted in a nearly linear frequency response curve over a frequency range of 7 to 50 cps. The frequency response curve for this detector is shown in figure 5.

A specially designed 12-conductor geophone cable was supplied with the refraction system. The takeouts were of a fixed, plastic type spaced at 30-foot intervals.

DuPont SSS seismograph electric blasting caps and boosters were used for the energy source. Seismograph caps were used because they have no time lag between open circuit and detonation, insuring an accurate shot instant on the seismic records.

PROCEDURE

Preliminary Considerations

Test borings in the vicinity of the Texas A&M Plantation indicate the water table is approximately 15 to 20 feet below the surface over most of the floodplain, increasing to 35 to 45 feet adjacent to the Brazos River (Cronin and Wilson, in press).

An un-reversed, 225-foot profile was conducted to determine the feasibility of using the seismic refraction method to determine water table and bedrock depths. Records were made with both a sledgehammer and an explosive energy source. The hammer source failed to produce an initial disturbance on the records as sharp as those produced with an explosive source at comparable distances. The travel-time curve for this profile indicated a velocity of about 1,300 feet per second for the dry alluvium and 4,100 feet per second for the saturated alluvium. Using a depth of 20 feet as an average from the test borings and the velocities obtained from this initial study, a critical distance of 55 feet was determined for the expected first refracted arrival from the water table. With the assumption that the saturated alluvium-bedrock velocity contrast was approximately 1,000 feet per second, the critical distance for refraction from this interface would be about 225 feet. The failure to detect a third velocity layer was therefore thought to be due to the short spread length.

Field Procedure

Using the information obtained from the un-reversed profile, a series of reversed profiles were made. A reversed profile consists of two profiles shot in opposite directions along the line joining the shot points. These profiles were made with a symmetric geophone spacing so that at least two geophones were within 40 feet of both shot points. On windy days, the geophones were buried to reduce the background noise level.

Each profile was centered on approximately a 500-foot grid, with four supplementary profiles near the river. The grid was measured by pacing in the field to reduce the amount of time between stations. The geophone spacings were measured with a 50- or 100-foot tape. From the test boring data, the water table appeared to slope toward the river, therefore the profiles were arranged nearly at right angles to the river. Figure 6 is a map of the locations of the profiles.

The shot holes were made by driving a one-inch steel rod to a depth of 1.5 to 2 feet with a sledgehammer. After the charge was emplaced, the hole was filled with water, or backfilled with soil if it would not hold the water. The shot points were marked and later mapped with a plane table to determine their locations and elevations.

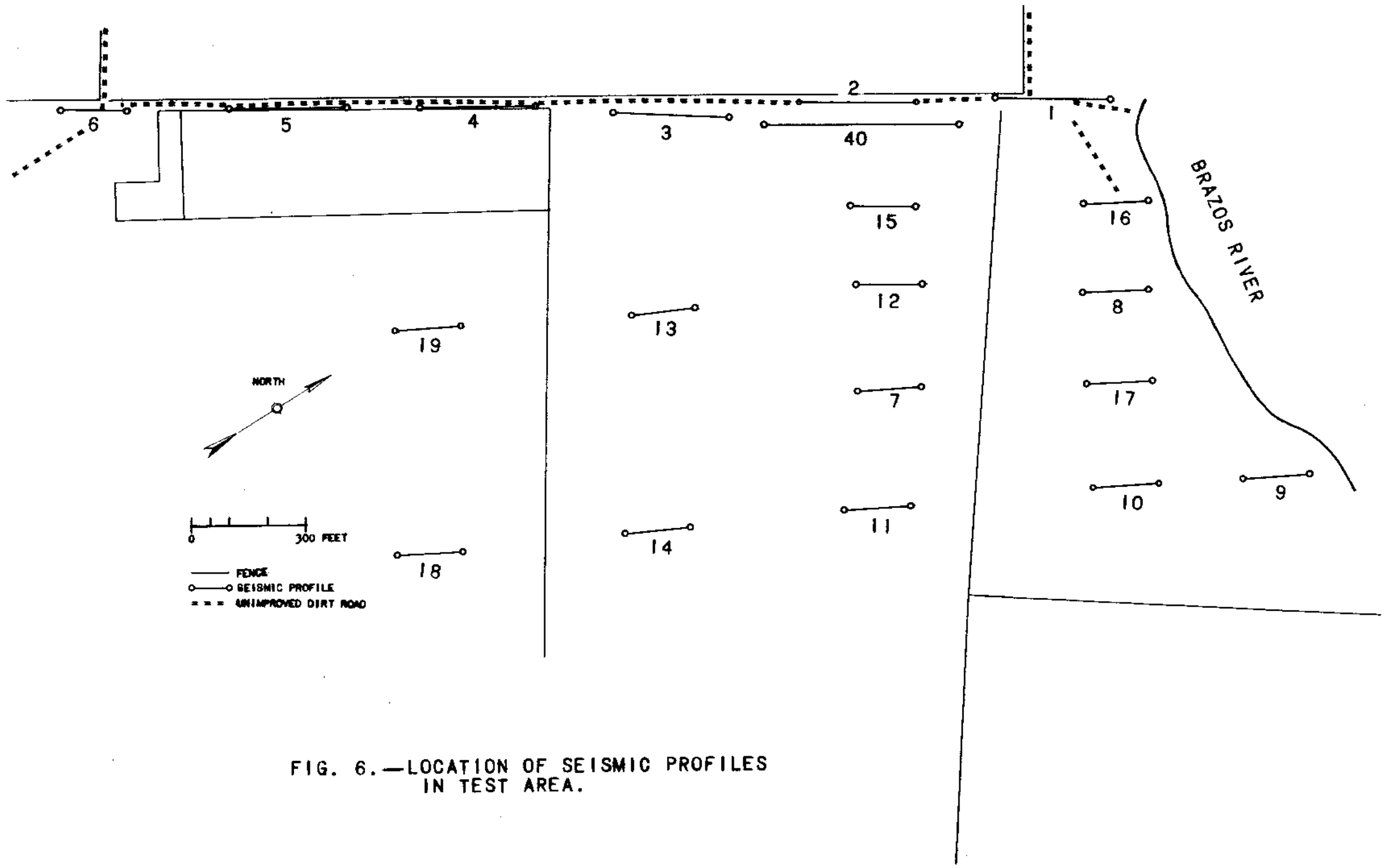


FIG. 6.—LOCATION OF SEISMIC PROFILES IN TEST AREA.

The first five profiles were made with a total spread length of 300 feet between the shot points and 11 geophone positions, which necessitated moving the geophones once and shooting at each shot point twice (see figure 7). These records did not indicate the saturated alluvium-bedrock velocity contrast expected. To investigate the possibility of a smaller velocity difference (hence, a larger critical distance), profile number 40 was made with a total spread length of 530 feet (see figure 7). This profile required moving the geophone spread twice and shooting three times at each shot point. For the spreads in which the shot point was 370 feet from the first geophone, two boosters were taped together to provide sufficient energy. This profile indicated that, if the three-layer case was applicable, the critical distance for refraction from bedrock was on the order of 400 feet.

Due to the limited amount of time the seismograph was available and the threat of unfavorable weather, it was decided to reduce the spread length to 170 feet (see figure 7). This procedure consisted of placing the geophones once and shooting 20 feet from each end. To check on the presence of a third layer, one shot point was placed 395 feet west of the geophones on profiles 6 through 19. This distant shot was made with two boosters taped together.

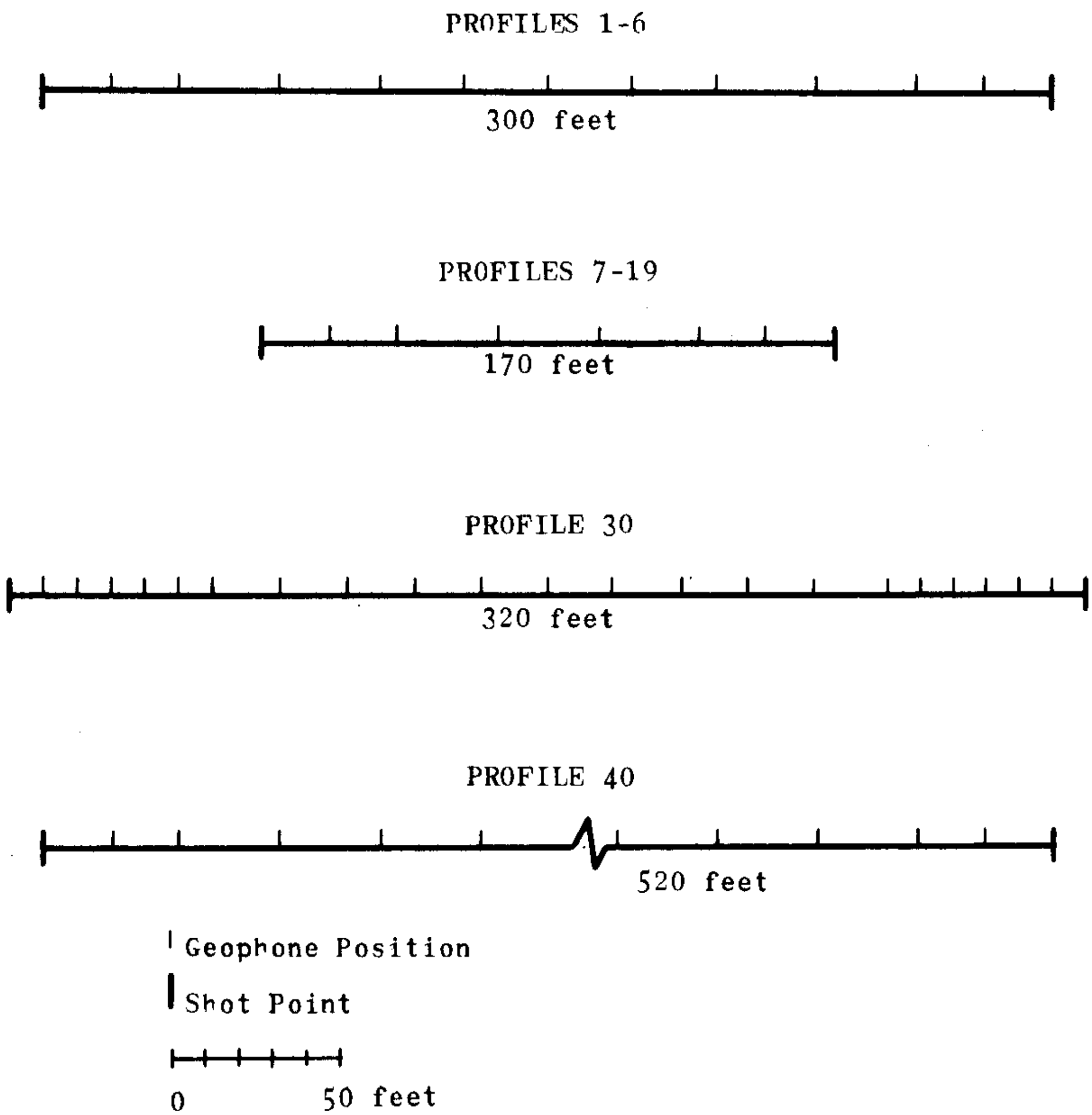


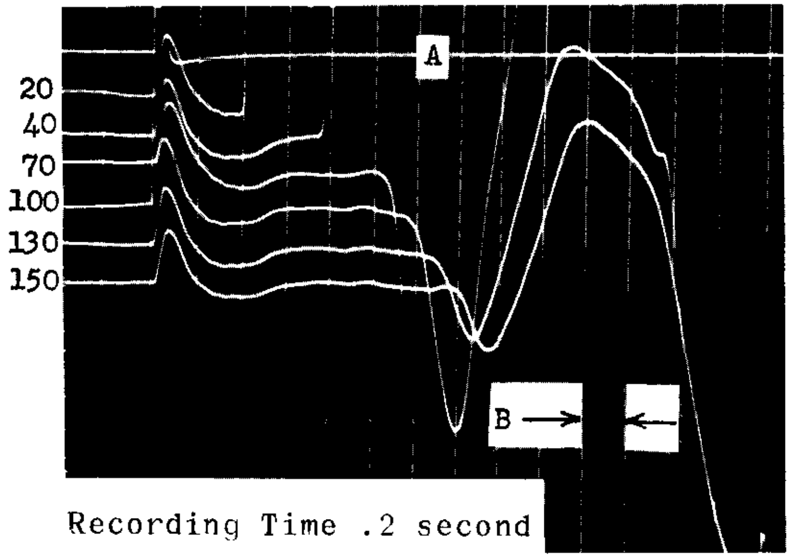
Fig. 7.—Shot and Detector Arrangements Used.

Profile number 30 was centered adjacent to an irrigation well to investigate the effects of a gravel lense on the seismic records. This profile was composed of four separate geophone spreads totaling 320 feet. In order to establish several data points on the regression line for the first layer, the geophone positions were symmetrically arranged so that six geophones would be within the first critical distance (see figure 7).

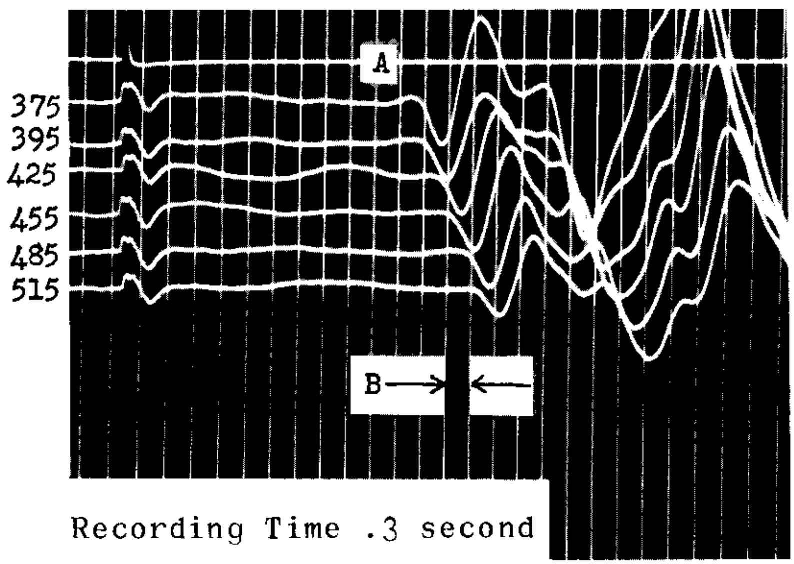
Analysis of Seismic Records

The shot instant and first breaks for each geophone trace were measured with a variable scale rule to an accuracy of about one-half millisecond. Figure 8 shows a typical seismogram of one short and one long profile. The data points for each profile were plotted on a time-distance graph and the number of points to be included in each velocity layer was chosen.

A computer program was written to determine a "best fitting" velocity line for each segment of the travel-time curve by the least-squares method. The program then computed the apparent velocities of the layers and the apparent depth to the first and second interfaces using the time-intercept method. This program was written in Fortran IV language for an IBM 7090/7094 computer and was designed for reversed seismic profiles. The program was written to compute the apparent velocity lines for each



Recording Time .2 second



Recording Time .3 second

A-Time Break Trace B-10 milliseconds

Fig. 8.—Typical Seismograms.

layer and the apparent depth to the first interface for each profile independently. The arithmetic mean of the apparent velocities of the first layer was used as the true velocity for that layer. With a velocity established for the first layer and the apparent measured velocities from the travel-time graph for the second layer, the velocity of the second layer, the dip angle, and the critical angle of refraction for the interface were computed. The true velocity of the second layer was then computed. The equations used are given in Appendix B. That portion of the travel-time curve which resulted from the distant shot and was un-reversed was treated as having the same shot point as the shorter, reversed profile for each station. That is, the shallow velocity measurements determined by the short, reversed profiles were assumed to hold in the vicinity of the distant shot. A print-out of the computer program is given in Appendix A.

The program was later expanded to include the calculation of the critical distance for each interface, 95 percent confidence limits for both the slope and time intercept of each velocity regression equation, true depths to the interfaces, and the travel times at the reversed points for the second layer for each profile. Both the initial time-distance data and the information listed above are output. A simplified flow chart of the major steps of the program is shown in figure 9.

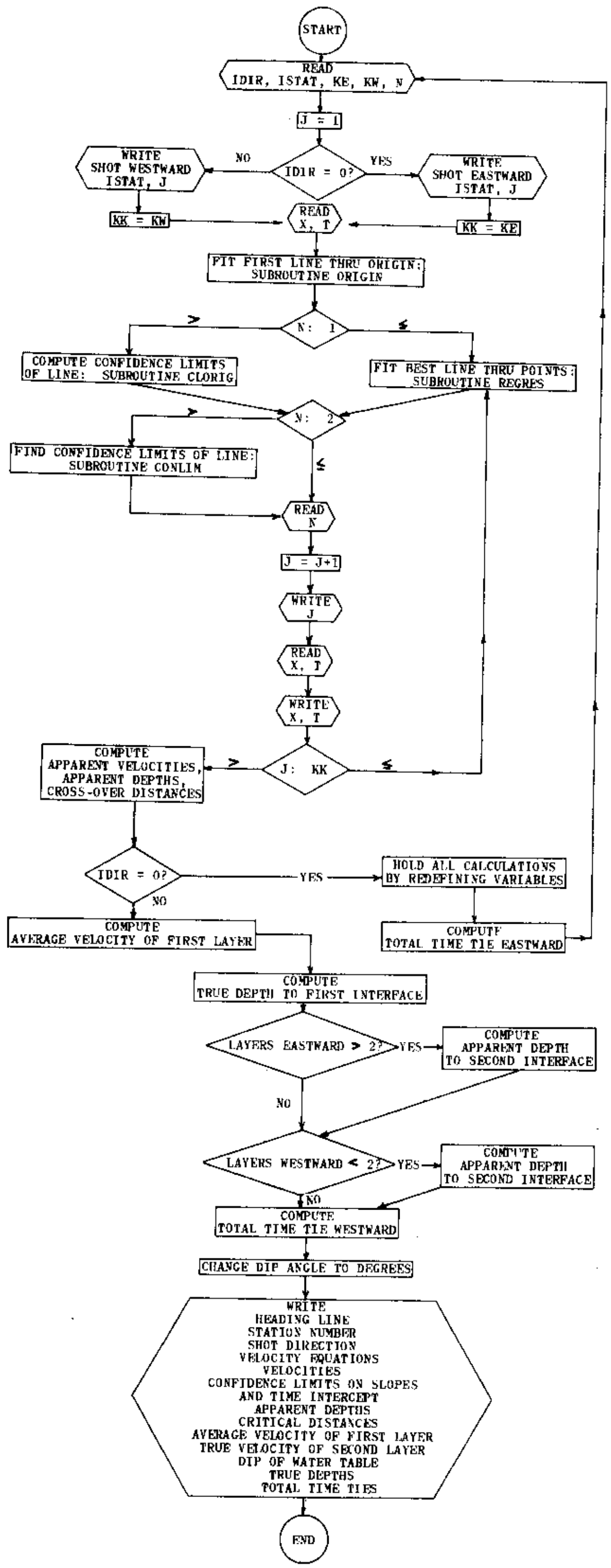


FIG. 9.—SIMPLIFIED FLOW DIAGRAM OF COMPUTER PROGRAM.

The calculation of the depth was made by a non-linear function involving the time intercept and the first and second layer velocities (equation 4, Appendix B). An estimate of the variance of the depth calculation could be made if the non-linear function can be made linear, using the additive property of the variance. Estimates of the variance for the time intercept and the velocities were already calculated. The function cannot be linearized by any simple transformation. However, an effective linearization can be accomplished by expanding the function into a Taylor's series in the region of interest. The three variables in this equation were considered as random variables, and the best estimate of depth was obtained by using their mean values. The estimate of the variance of the depth to the water table was approximated using a Taylor's series with only first-order terms. The equations used to compute the estimated variance of the depth are also found in Appendix B. The actual calculations were performed on the computer. The depths and estimated variance to the water table, and the dry and saturated alluvial velocities are tabulated in tables I and II in Appendix C.

A separate method of analysis of the travel-time data using the equations given by Steinhart and Meyer (1961) was also investigated. This method requires that the two line segments representing the velocity of each layer must have an equal time intercept over the distance between the two shot points. The equations involved are again least-squares estimates of the time tie and slopes for each layer. These equations were also programed for the computer. It was found, however, that for fewer than five time-distance points in each layer, the regression equations could not be accurately determined. This method involved the solution of three simultaneous equations with three unknowns. With only a few data points, the equations were ill-conditioned. An ill conditioned system is very sensitive to small variations in the values of the data points. Using this method, the slopes of the velocity lines were too large, and hence, the time intercepts for the first layer were calculated as negative. At station number 2 for example, the eastward slope for the first layer was computed at 2.81 and the time tie at 808.91 milliseconds, which results in a time intercept of -32.76 milliseconds.

RESULTS

The increase in seismic velocity between the first and second layers was generally on the order of 3,500 to 4,000 feet per second. The depth to this first interface in the preliminary tests was calculated at 19.5 to 21 feet, which agreed with the depth to the water table of 20 feet that was found in a test boring made at that location. The first seismic interface was therefore interpreted as the water table. Since the elevations of the shot points were known from the plane table survey, a contour map of the water table (figure 10) was drawn rather than a map depicting the depth to the water table. The river elevation during the test period was approximately 193 feet (Twichell, 1966).

Travel-time graphs were prepared for each profile and are shown in figures 11, 12, 13, and 14. The ordinate of these curves is the travel time in milliseconds and the abscissa is the distance in feet.

The velocity lines through the points representing direct arrivals through the first layer had a positive time intercept which ranged from 0.5 to 7 milliseconds, with an average of 2 to 3 milliseconds. A time delay such as this could possibly be introduced in the recording instrument or introduced between the moment of explosion and the moment when the blasting circuit current is

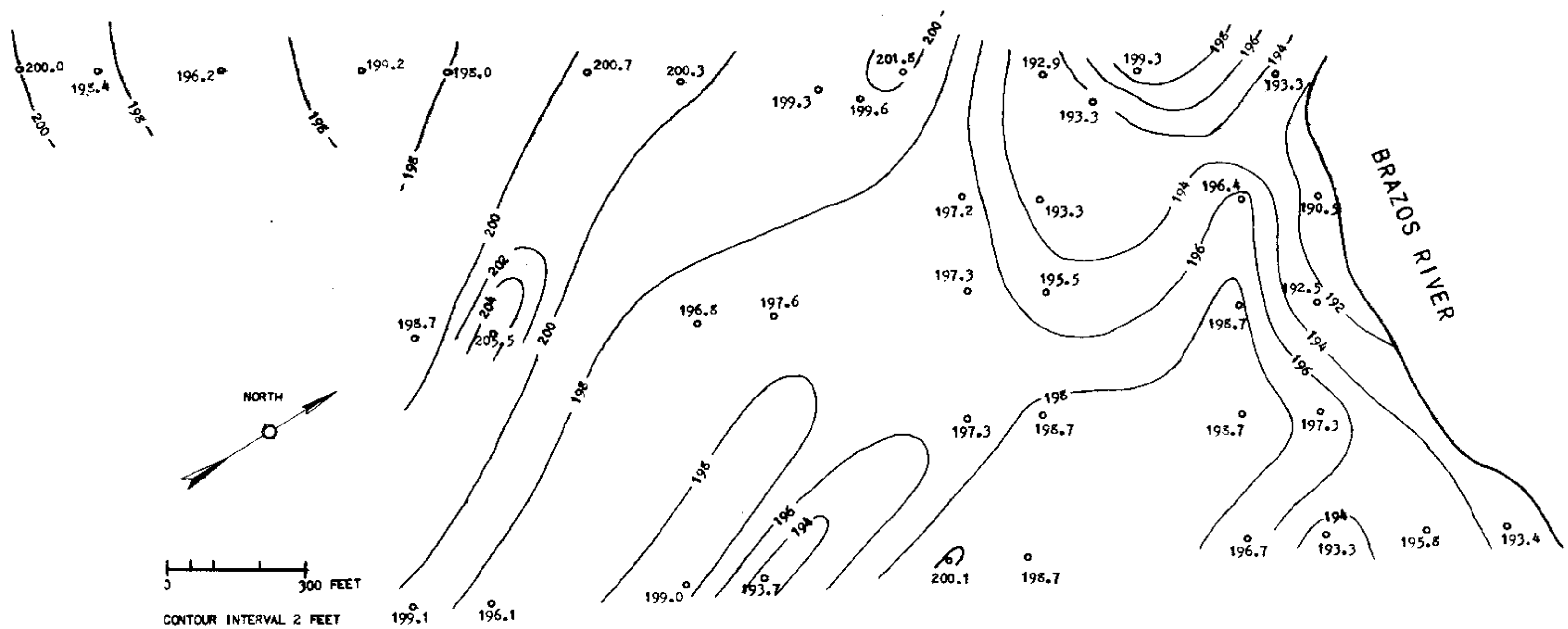


FIG. 10.—CONTOUR MAP OF WATER TABLE BASED ON SEISMIC DATA.

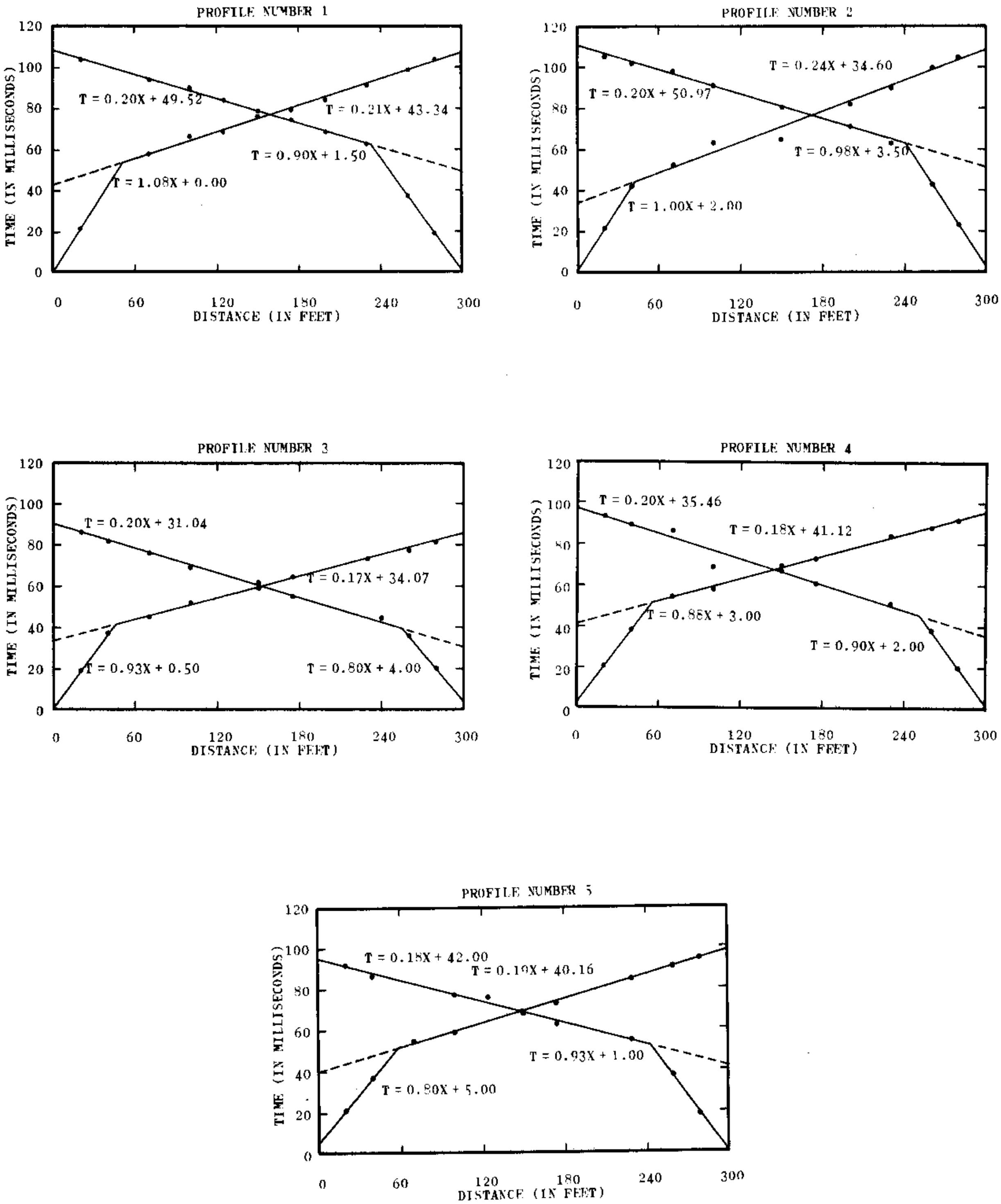


FIG. 11.—TRAVEL-TIME GRAPHS (PROFILES 1 THROUGH 5).

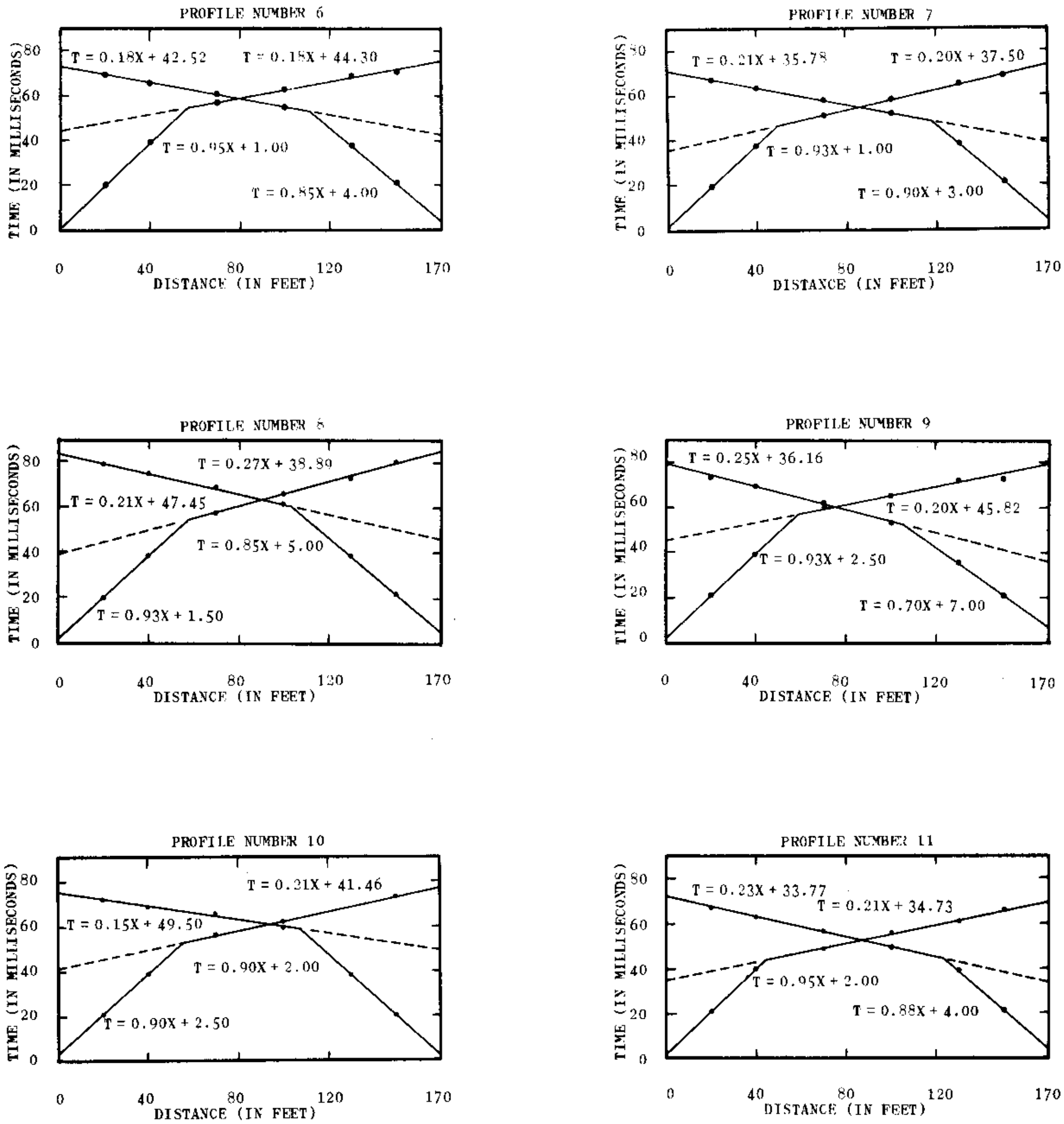


FIG. 12.—TRAVEL-TIME GRAPHS (PROFILES 6 THROUGH 11).

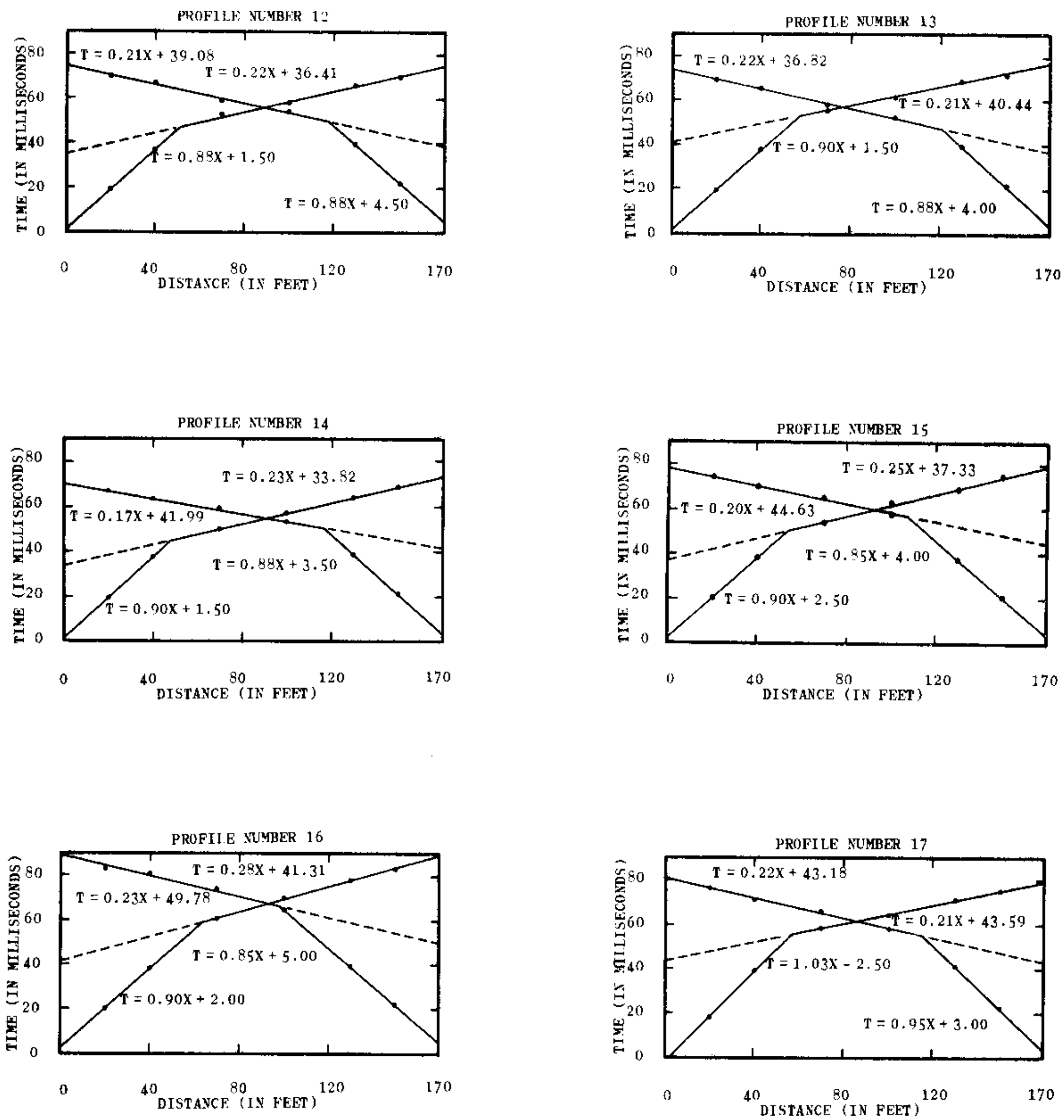


FIG. 13.—TRAVEL-TIME GRAPHS (PROFILES 12 THROUGH 17).

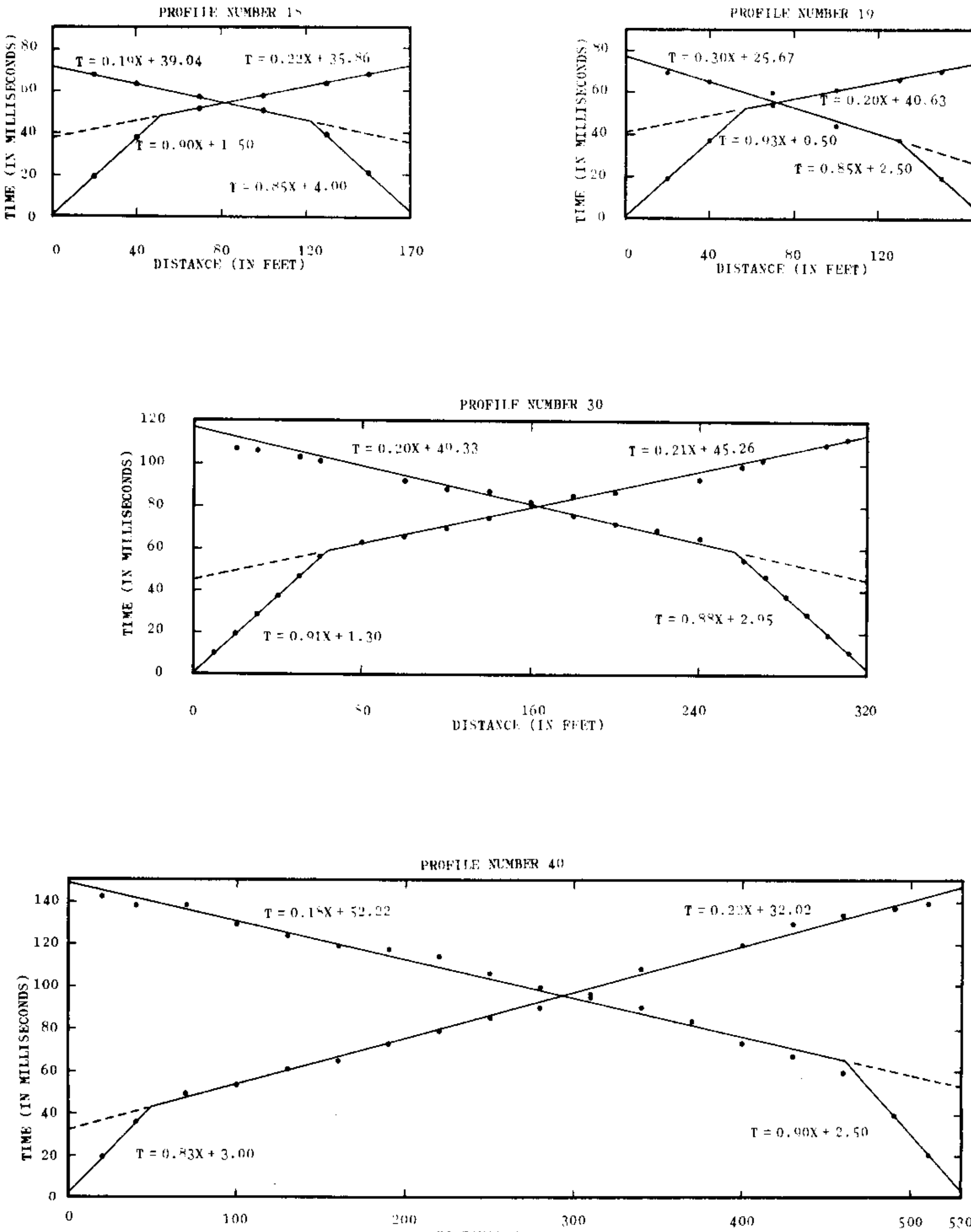


FIG. 14.—TRAVEL-TIME GRAPHS (PROFILES 18, 19, 30, 40).

interrupted. The accuracy of the equipment used, however, indicates that this observed time delay is real and is not an instrumental error. This initial positive time intercept has been reported by several other authors (Dobrin, 1942; Sharpe, 1942; Gough, 1952; Gough and van Niekerk, 1957; and Koefoed, 1954); none of them regarded it as an error in measurement. Koefoed and Gough show profiles in which the first arrivals were the sound wave through air, and the time intercept was very small or zero. Gough reported the positive time intercept to be due to a thin layer of aerated soil in which the seismic velocity was very low. This theory was based on Lester's (1932) investigations of the effect of various proportions of air to earth on the seismic velocity. Koefoed proposed that the probable explanation of this phenomenon was found in a theory developed by Gassmann (1951), who derived the velocity distribution for a hexagonal packing of solid, spherical bodies. According to this theory, velocity is proportional to the sixth root of the depth, hence extremely low velocities would occur in the first few inches of the sedimentary deposit.

The compressional wave velocities in the dry alluvium showed some fluctuation in the horizontal direction, the extreme values being of the order of 1,012 to 1,254 feet per second, with an average velocity of 1,131 feet per second. This variation in velocity was probably due to anisotropy in the dry alluvium caused by a variation in the grain size. Over most of the test area a clay soil is present at the surface, but it grades horizontally to a sandy soil at some of the stations. The shot holes made in the clay would hold water for up to 30 minutes, while some of the shot holes in sand would not hold water. The elastic and transmitting characteristics of the material immediately below the geophone have a substantial influence on the arrival times. Domzalski (1956) and Koefoed (1954) demonstrated that the arrival times are influenced by even small thicknesses of low velocity material. Koefoed (1954) reported on a profile made with three geophones. After an initial shot, the middle geophone was removed, the hole was filled with 5 to 10 centimeters of loose sand, and the geophone was replaced. A second shot at the same shot point as in the first profile resulted in a two- to three-millisecond increase in the arrival time at the middle geophone. Domzalski (1956) reported on a seismic profile made in a fan spread in which all of the geophones were 200 feet from the shot point. One geophone was on undisturbed ground, one on seven inches of disturbed ground, one on 3.5 inches of fine gravel, and one on six inches of sand. The travel times varied from three to four milliseconds. The grain size variation at the surface in the test area is, therefore, thought to be the probable cause for the variation in the seismic velocity within the dry alluvium.

If the assumption is made that the data fits a three-layer model, then the velocity increase between the saturated alluvium and the saturated bedrock was generally less than 500 feet per second. On a few of the profiles, the velocities determined from the distant shot were less than those computed for the saturated alluvium. The confidence limits of the velocity regression equations for the second and third layers overlapped at all stations, indicating that on a statistical basis the two lines should be considered as only one line. The slight increase in velocity observed on most of the distant profiles could be due to a slight velocity contrast at the alluvium-bedrock interface or the result of a continuous increase of velocity with depth below the water table, possibly due to increased compaction. White and Sengbush (1953) investigated seismic velocities in shallow deposits at several locations where the near-surface material was loose sand. They found that all velocities were low at the surface and increased smoothly with depth, except for an abrupt increase in compressional velocity, without a perceptible change in shear properties, at the water table. Their results for profiles shot in loose sand confirmed the general shape of Gassmann's curve; that is, the velocity is proportional to the sixth root of the depth.

The seismic velocities in the saturated alluvium, or at least the upper portion of the saturated deposits, showed considerable fluctuation, the extreme values being of the order of 3,934 to 5,608 feet per second, with an average velocity of 4,805 feet per second. This average velocity is slightly greater than the average velocity in water, as Burwell noted. The most probable cause for the variation in the velocity is due to the amount of gravel mixed with the sand. Lenses of clay or silt could also cause these velocity variations. In the absence of an independent check, these interpretations are only speculations based on published velocities. White and Sengbush (1953) report a velocity of 5,500 feet per second for a loose sand below the water table; and Woollard and Hanson (1954) reported a velocity of 5,090 feet per second for a saturated gravelly sand.

The variations in the elevation of the water table are probably partially due to the capillary fringe, which extends from the water table to the limit of the capillary rise of water. According to Tolman (1937, p. 155), most authorities give the maximum capillary lift as approximately 10 feet. The maximum relief of the water table in the test area is 15 feet. Tolman also listed the capillary rise in sediments of different grain size. His results showed the fine-granular material (silts) to have a greater extent of lift than the sands and clays. Using a fine to very fine sand which had a 60 percent size of 0.19 mm and a 10 percent size of 0.08 mm, Lambe (1951, p. 411) reported a capillary rise of 35 inches over an elapsed time of 22 days, and concluded that the total rise should be greater than 39 inches over a longer period of time. Smith (1933, p. 438) summarizes the results of King's (1897-1898) experiments in measuring the extent of capillary rise over a period of 2.5 years. The maximum rise recorded was 102.9 inches for a very fine-grained sand (median diameter = 0.082 mm). The rise in coarser sands was less, being 18.7 inches for a medium-grained sand (median diameter = 0.474 mm).

The seismic velocities, confidence limits, and variance of depth to the water table of the profile shot adjacent to an irrigation well located 1,000 feet north of the test area did not indicate any characteristic values that would delineate similar gravel lenses in the test area. Thus, the actual refraction profile by itself cannot outline the gravel zones within the alluvium.

Todd (1960, p. 67-68) gave a procedure which utilized a contour map of the water table together with the flow lines to outline zones of greater permeability. The resulting equations could be interpreted as indicating, for an area of uniform ground-water flow, that portions having flat gradients (wide contour spacing) would have greater permeabilities than those with steep gradients (narrow spacing). Zones of greater permeability in the alluvial deposits of the test area are probably zones of thicker gravel deposits.

No clear correlation could be made between the seismic velocities in the saturated alluvium and the areas of flat gradients. There was a similar correlation, however, between the areas of gentle gradients on the contour map and the areas in which the profiles showed a lesser amount of "scatter" on the travel-time curves. The "scatter" was measured by the mean square error of the time-distance points about the velocity regression line for the saturated layer. This "scatter" results from the failure of the seismic model to fit the geologic conditions, principally due to inhomogeneity of the transmitting medium and to the deviations of the interfaces from planes. A speculative reason for the similar trend of the proposed zone of higher permeability and the area of least "scatter" might be the approximately planar seismic interface.

CONCLUSIONS

As the depth of seismic refraction investigations becomes shallower, the differences between the theoretical model and the geologic conditions become pronounced. Observational errors, errors due to assumptions, and errors due to rapid changes in conditions within short vertical and horizontal distances may result in a relatively large error in the depth estimate when making short refraction profiles. The disturbing factors are due to the method of generating the seismic wave, the shot hole conditions, the topography of the ground surface, inhomogeneity and horizontal and vertical changes in the stratification, and irregularities in the refracting surfaces.

Shallow seismic studies are useful because they rapidly provide a picture of the refracting interfaces to serve as a guide for any subsequent drilling. The accuracy of the depths to and nature of the refracting layers would be greatly improved if some preliminary test drilling information were available.

Shallow seismic refraction methods can be used to map the top of the saturated zone in the unconsolidated or poorly indurated sediments of the Gulf Coastal Plain with some accuracy. Although the seismic method cannot differentiate sediment types below the water table, the contour spacing on the resulting map of the water table should outline zones of higher permeability. To test this assumption, a profile could be made across this zone before and during a period of heavy irrigation withdrawal. If these are zones of greater permeability, the water table should decline during the pumping period.

The elastic properties of the saturated alluvium and the bedrock did not contrast sufficiently for this interface to be mapped by seismic methods. The poor induration of most of the Tertiary sediments in the Coastal Plain would hinder mapping of the alluvium-bedrock interface by seismic methods. Upstream, however, where the alluvium overlies more indurated Cretaceous deposits, shallow refraction methods could probably map this interface.

The range of uncertainty in the depth determinations is probably an indication of the combined effects of inhomogeneity of the deposits and the departure from flatness of the seismic interfaces. This range is not a criterion for judging the correctness of the interpretation as the interpretation was made in the original grouping of the data before making the least-squares solutions.

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APPENDIX A

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$EXECUTE       IBJOB
$IBJOB THESIS
$IBFTC MAIN
C   ---------------
C   ---------------
C    IBM 7090/7094 COMPUTER PROGRAM IN FORTRAN 4, IBJOB VERSION 5 (VERSION 13)
C    FOR SHALLOW, REVERSED, DIPPING, SEISMIC REFRACTION PROFILES
C   ---------------
C   ---------------
C    MAIN PROGRAM
C       T VAL(X)=STUDENTS T DISTRIBUTION AT THE DESIRED LEVEL
C       IDIR=DIRECTION OF GEOPHONES FROM SHOT POINT, O=WESTWARD,1=EASTWARD
C       ISTAT=STATION NUMBER    KE=NUMBER OF LAYERS IN EASTWARD DIRECTION
C       KW=NUMBER OF LAYERS IN WESTWARD DIRECTION    N=NUMBER OF POINTS IN EACH
C       LAYER    J=LAYER NUMBER    X(I)=DISTANCE FROM SHOT POINT
C       T(I)=TRAVEL TIME AT DISTANCE X(I)
C   ---------------
      DIMENSION T(40),X(40),B(8),A(8),Z(8),V(8),V HOLD(8),B HOLD(8),
     1Z HOLD(8),A HOLD(8),VAVE(8),D UP(2),D DOWN(2),
     2T VAL(13), BUL(8), BLL(8), AUL(8), ALL(8),
     3BUL HO(8), BLL HO(8), AUL HO(8), ALL HO(8),
     4WINT(8), WINTHL(8), VOH(4), AOH(4), BOH(4),
     5AO(4), BO(4), VO(4), BULO(4), BLLO(4), BULOH(4), BLLOH(4),VT(4)
      COMMON X, T, A, B, Z, V, V HOLD, B HOLD, Z HOLD, A HOLD, VAVE,
     1D UP, D DOWN, BUL, BLL, AUL, ALL, BUL HO, BLL HO, AUL HO,
     2ALL HO, T VAL, WINT , WINTHL, N, XX, J, KK, KE, KW,
     3BO, AO, VO, AOH, BOH, VOH, BULO, BLLO, BULOH, BLLOH, VT
      T VAL(1)=12.706
      T VAL(2)=4.303
      T VAL(3)=3.182
      T VAL(4)=2.776
      T VAL(5)=2.571
      T VAL(6)=2.447
      T VAL(7)=2.365
      T VAL(8)=2.306
```

```
      T VAL(9)=2.262
      T VAL(10)=2.228
      T VAL(11)=2.201
      T VAL(12)=2.179
      T VAL(13)=2.160
   20 READ (5,111) IDIR, ISTAT, KE, KW
  111 FORMAT (4I2)
      READ (5,100) N
      BUL(I)=0.0
      BLL(I)=0.0
      AUL(I)=0.0
      ALL(I)=0.0
      BULO(I)=0.0
      BLLO(I)=0.0
      J=1
      IF (IDIR) 378,379,378
  379 WRITE (6,200) ISTAT, J
  200 FORMAT (1H1/20X49HM. A. MCBRAYER THESIS, SEISMIC REFRACTION PORTIO
     1N/7X15HSTATION NUMBER I2//10X13HSHOT EASTWARD/20X6HLAYER I2/)
      KK = KE
      GO TO 167
  378 WRITE (6,376) J
  376 FORMAT (///10X13HSHOT WESTWARD/20X6HLAYER I2/)
      KK = KW
  167 READ (5,101)(X(I), T(I), I=1,N)
      WRITE (6,201)(I,X(I),I,T(I),I=1,N)
  201 FORMAT(10X2HX(I2,2H)=F8.2,5X2HT(I2,2H)=F8.2)
  100 FORMAT (I2)
  101 FORMAT (2F15.2)
      XX=N
C     ---------------
C      SUBROUTINE ORIGIN IS A LEAST SQUARES FIT FOR THE 1ST LAYER WHICH
C      FORCES THE LINE THRU THE ORIGIN      BO=SLOPE OF LINE
C     ---------------
      CALL ORIGIN (X, T, BO, AO, N)
```

```
      Q = N-2
      IF (N-1) 771,771,772
C   ---------------
C    SUBROUTINE CLORIG COMPUTES THE CONFIDENCE LIMITS AT THE DESIRED LEVEL
C    FOR THE LINE THRU THE ORIGIN   BULO=UPPER LIMITS ON SLOPE, BLLO=LOWER
C   ---------------
  772 CALL CLORIG (X, T, T VAL, N, J, BO, AO, XX, BULO, BLLO)
C   ---------------
C    SUBROUTINE REGRES IS A LEAST SQUARES FIT OF THE REMAINING VELOCITY LINES
C       A=TIME INTERECPT OF LINE       B=SLOPE
C   ---------------
  771 CALL REGRES (X, T, A, B, N, XX, J)
      IF (N-2) 871,871,770
C   ---------------
C    SUBROUTINE CONLIM COMPUTES THE CONFIDENCE LIMITS AT THE DESIRED LEVEL
C    FOR VELOCITY LINES FITTING 3 OR MORE DATA POINTS
C       BUL=UPPER CONFIDENCE LIMITS ON SLOPE, BLL=LOWER
C       AUL=UPPER CONFIDENCE LIMITS ON TIME INTERECPT, ALL=LOWER
C   ---------------
  770 CALL CONLIM (X, T, A, B, BUL, BLL, AUL, ALL, T VAL, N, XX, KK, J)
  871 DO 3 J=2,KK
      READ (5,100) N
      WRITE (6,709) J
  709 FORMAT (//20X6HLAYER I2/)
      READ (5,101)(X(I), T(I), I=1,N)
      WRITE (6,201)(I,X(I),I,T(I),I=1,N)
      XX = N
      CALL REGRES (X, T, A, B, N, XX, J)
      CALL CONLIM (X, T, A, B, BUL, BLL, AUL, ALL, T VAL, N, XX, KK, J)
    3 CONTINUE
      DO 4 J=1,KK
C   ---------------
C       V(J)=VELOCITY OF LAYER J,VO(1)=VELOCITY 1ST LAYER WHEN FORCED THRU ORIGIN
C       Z(1)=APPARENT DEPTH TO 1ST INTERFACE, WINT(1)=CROSS OVER DISTANCE FOR
C       1ST INTERFACE REFRACTIONS
```

```
C  ---------------
    4 V(J) = 1.0/B(J)
      VO(1) = 1./BO(1)
      Z(1)=0.5*A(2)*V(2)*V(1)/SQRT(V(2)*V(2)-V(1)*V(1))
      WINT(1) = A(2)/(B(1)-B(2))
      IF (IDIR) 13,11,13
C  ---------------
C   ALL EASTWARD COMPUTATIONS ARE HELD (STORED) AND WESTWARD PROFILE IS COMPUTED
C  ---------------
   11 DO 10 I=1,KK
      AOH(1) = AO(1)
      BOH(1) = BO(1)
      VOH(1) = VO(1)
      QH = Q
      KKH=KK
      BULOH(1) = BULO(1)
      BLLOH(1) = BLLO(1)
      WINTHL(1) = WINT(1)
      V HOLD(I) = V(I)
      Z HOLD(I) = Z(I)
      B HOLD(I) = B(I)
      BUL HO(I) = BUL(I)
      BLL HO(I) = BLL(I)
      AUL HO(I) = AUL(I)
      ALL HO(I) = ALL(I)
   10 A HOLD(I) = A(I)
      IF (ISTAT-5) 124,124,125
C  ---------------
C       TTE=TOTAL TIME TIE SHOOTING EASTWARD, TTW=TIME TIE WESTWARD
C       VAVE(1)=AVERAGE VELOCITY OF 1ST LAYER, VT(2)=TRUE VELOCITY OF 2ND
C       CI=CRITICAL ANGLE OF REFRACTION, ALPHA=DIP ANGLE OF INTERFACE
C       CHK DETERMINES UP DIP DIRECTION
C       D UP(1)=TRUE DEPTH UP DIP, D DOWN(1)=TRUE DEPTH DOWN DIP
C  ---------------
  124 TTE = A HOLD(2) + B HOLD(2)*300.
```

```
  125 TTE = A HOLD(2) + B HOLD(2)*170.
      GO TO 20
   13 VAVE(1) = 0.5*(V HOLD(1)+V(1))
      CI=0.5*(ASIN(VAVE(1)/V HOLD(2))+ASIN(VAVE(1)/V(2)))
      CHK = V HOLD(2)-V(2)
      IF (CHK) 1000,1000,1001
 1000 ALPHA=0.5*(ASIN(VAVE(1)/V HOLD(2))-ASIN(VAVE(1)/V(2)))
      GO TO 1002
 1001 ALPHA=0.5*(ASIN(VAVE(1)/V(2))-ASIN(VAVE(1)/V HOLD(2)))
 1002 VT(2) = VAVE(1)/SIN(CI)
      D UP(1) = Z(1)/COS (ALPHA)
      D DOWN(1) = Z HOLD(1)/COS (ALPHA)
      IF (KKH-2) 442,443,442
  442 DC=D DOWN(1)*B HOLD(1)*B HOLD(3)
      DD=SQRT(V HOLD(3)*V HOLD(3)-V HOLD(1)*V HOLD(1))
      DE=V HOLD(3)*V HOLD(2)
      DF=SQRT(V HOLD(3)*V HOLD(3)-V HOLD(2)*V HOLD(2))
C   ---------------
C        Z HOLD(2)=APPARENT DEPTH TO 2ND LAYER WHEN SHOOTING EASTWARD
C        Z(2)=APPARENT DEPTH WHEN SHOOTING WESTWARD
C        WINTHL(2)=CROSS OVER DISTANCE SHOOTING EASTWARD, WINT(2)=WESTWARD
C   ---------------
      Z HOLD(2)=0.5*(A HOLD(3)-2.*(DC*DD*DE/DF))
      WINTHL(2)=(A HOLD(3)-A HOLD(2))/(B HOLD(2)-B HOLD(3))
  443 IF (KK-2) 444,555,444
  444 DA=D DOWN(1)*B(1)*B(3)*SQRT(V(3)*V(3)-V(1)*V(1))
      DB=V(3)*V(2)/SQRT(V(3)*V(3)-V(2)*V(2))
      Z(2)=0.5*(A(3)-2.*(DA*DB))
      WINT(2) = (A(3)-A(2))/(B(2)-B(3))
  555 CONTINUE
      IF (ISTAT-5) 126,126,127
  126 TTW = A(2) + B(2)*300.
      GO TO 1010
  127 TTW = A(2) + B(2)*170.
 1010 ALPHA = (ALPHA)*(180./3.1416)
```

```
      DO 37 J=1,KE
   37 V HOLD(J)=V HOLD(J)*1000.
      DO 73 J=1,KW
   73 V(J) = V(J)*1000.
      VO(1) = VO(1)*1000.
      VOH(1)=VOH(1)*1000.
      VAVE(1) = VAVE(1)*1000.
      VT(2) = VT(2)*1000.
      WRITE (6,207) ISTAT
  207 FORMAT (1H1/20X48HREGRESSION EQUATIONS AND DEPTHS, STATION NUMBER
     1 I2/)
      WRITE (6,203)
  203 FORMAT (/// 5X13HSHOT EASTWARD//)
      WRITE (6,832)  AOH(1), BOH(1), VOH(1)
  832 FORMAT (10X14HTHRU ORIGIN    ,3HT =F8.4,1H+F8.4,1HX,15X6HV( 1)=F9.
     13,10HFT PER SEC)
      WRITE (6,812)  A HOLD(1), B HOLD(1), V HOLD(1)
  812 FORMAT ( 10X14HBEST FIT      ,3HT =F8.4,1H+F8.4,1HX,15X6HV( 1)=F9.
     13,10HFT PER SEC//)
      WRITE (6,202)(A HOLD(I), B HOLD(I), I, V HOLD(I), I=2,KE)
  202 FORMAT ( 24X3HT =F8.4,1H+F8.4,1HX,15X2HV(I2,2H)=F9.3,10HFT PER SEC
     1/)
      IY = KE-1
      WRITE (6,204)(I,Z HOLD(I),I,WINTHL(I), I=1,IY)
  204 FORMAT (/10X2HZ(I2,2H)=F8.3,2HFT,9X19HCRITICAL DISTANCE (I2,2H)=
     1F9.3,2HFT)
   71 WRITE (6,95)
   95 FORMAT (// 7X58H95 PERCENT CONFIDENCE LIMITS ON SLOPES AND TIME IN
     1TERCEPTS/)
      IF (QH) 89, 89, 90
   90 WRITE (6,93)  BULOH(1), BLLOH(1)
   93 FORMAT (10X9HBUL ( 1)=F6.3,5X9HBLL ( 1)=F6.3/)
   89 WRITE (6,92)(I, BUL HO(I), I, BLL HO(I), I, AUL HO(I), I, ALL HO
     1(I), I=1,KE)
   69 WRITE (6,205)
```

```
  205 FORMAT (// 5X13HSHOT WESTWARD//)
      WRITE (6,832)  AO(1), BO(1), VO(1)
      WRITE (6,812)  A(1), B(1), V(1)
      WRITE (6,202)(A(I), B(I), I, V(I), I=2,KW)
      IZ = KW-1
      WRITE (6,204)(I,Z(I),I,WINT(I), I=1,IZ)
      WRITE (6,95)
      IF (Q) 99,99,113
  113 WRITE (6,93)  BULO(1), BLLO(1)
   99 WRITE (6,92) (I, BUL(I), I, BLL(I), I, AUL(I), I, ALL(I), I=1,KW)
   92 FORMAT ( 10X5HBUL (I2,2H)=F6.3,5X5HBLL (I2,2H)=F6.3,10X5HAUL (I2,2
     1H)=F7.3,5X5HALL (I2,2H)=F7.3)
   97 WRITE (6,206) VAVE(1), VT(2)
  206 FORMAT (//10X22HAVERAGE VELOCITY (1) =F9.3,10HFT PER SEC,10X19HTRU
     1E VELOCITY (2) =F9.3,10HFT PER SEC)
      WRITE (6,208) ALPHA, D UP(1), D DOWN(1)
  208 FORMAT (//10X7HALPHA =F8.5,10X9HD UP(1) = F8.3,2HFT,10X11HD DOWN(1
     1) =F8.3,2HFT)
      WRITE (6,128) TTE, TTW
  128 FORMAT (//10X15HTIME TIE EAST =F8.3,8HMILLISEC,10X15HTIME TIE WEST
     1 =F8.3,8HMILLISEC)
      GO TO 20
      END
$IBFTC ONE
      SUBROUTINE ORIGIN (X, T, BO, AO, N)
      DIMENSION X(10), T(10), BO(3), AO(3)
      COMMON X, T, N, AO, BO
      SUM 1=0.0
      SUM 2*0.0
      DO 1 I=1,N
      SUM 1= SUM 1+X(I)*T(I)
    1 SUM 2= SUM 2+X(I)*X(I)
      BO(1) = SUM 1/SUM 2
      AO(1) = 0.0
      RETURN
```

```
      END
$IBFTC TWO
      SUBROUTINE CLORIG (X, T, T VAL, N, J, BO, AO, XX, BULO, BLLO)
      DIMENSION X(40), T(40), BO(4), BULO(4), BLLO(4), T VAL(13)
      COMMON X, T, BO, T VAL, N, J, BULO, BLLO, XX
      D = 1.0/(XX-1.0)
      P = 0.0
      Q = 0.0
      R = 0.0
      DO 42 LJ = 1,N
      P = P + T(LJ)*T(LJ)
      Q = Q + T(LJ)*X(LJ)
   42 R = R + X(LJ)*X(LJ)
      S2YX = (P-(Q*Q/R))*D
      SB = SQRT(S2YX/R)
      BULO(1) = BO(1) + T VAL(N-1)*SB
      BLLO(1) = BO(1) - T VAL(N-1)*SB
      RETURN
      END
$IBFTC THREE
      SUBROUTINE REGRES (X, T, A, B, N, XX, J)
      DIMENSION X(40), T(40), A(4), B(4), T VAL(13)
      COMMON X, T, A, B, T VAL, N, XX, J
      SUM 1=0.0
      SUM 2=0.0
      SUM 3=0.0
      SUM 4=0.0
      DO 2 I=1,N
      SUM 1=SUM 1+X(I)
      SUM 2=SUM 2+T(I)
      SUM 3= SUM 3+X(I)*T(I)
    2 SUM 4 = SUM 4+X(I)*X(I)
      A(J)=(SUM 1*SUM 3-SUM 4*SUM 2)/(SUM 1*SUM1-XX*SUM 4)
      B(J)=(SUM 2-A(J)*XX)/SUM 1
      RETURN
```

```
      END
$IBFTC FOUR
      SUBROUTINE CONLIM (X, T, A, B, BUL, BLL, AUL, ALL, T VAL, N, XX,
     1KK, J)
      DIMENSION X(40), T(40), BUL(5), BLL(5), AUL(5), ALL(5), A(8),
     1B(8), T VAL(13)
      COMMON X, T, A, B, XX, T VAL, N, KK, BUL, BLL, AUL, ALL, J
      D=1.0/(XX-2.0)
      T BAR =0.0
      X BAR =0.0
      E =0.0
      F =0.0
      G =0.0
      DO 222 LL=1,N
      X BAR=X BAR+X(LL)
  222 T BAR=T BAR+T(LL)
      X BAR=X BAR/XX
      T BAR=T BAR/XX
      DO 541 LL=1,N
      E=E+(T(LL)-T BAR)*(T(LL)-T BAR)
      F=F+(X(LL)-X BAR)*(T(LL)-T BAR)
  541 G=G+(X(LL)-X BAR)*(X(LL)-X BAR)
      S2YX=(E-(F*F/G))*D
      SB=SQRT(S2YX/G)
      BUL(J)=B(J)+T VAL(N-2)*SB
      BLL(J)=B(J)-T VAL(N-2)*SB
      SA=SQRT(S2YX*((1.0/XX)+(X BAR*X BAR)/G))
      AUL(J)=A(J)+T VAL(N-2)*SA
      ALL(J)=A(J)-T VAL(N-2)*SA
      RETURN
      END
$DATA
```

Least Square Fitting the Line Segments and Depth Equations

The geophysical model used in this analysis is based on the assumption that the earth is composed of isotropic, homogeneous layers bounded by planar interfaces. It is also assumed that the seismic velocity is constant in each layer and that the velocity for each successive layer increases. For such a model, a plot of the time of arrival of the seismic energy at each geophone versus the distance to that geophone will be a series of straight line segments. Therefore, to compute the velocities and depths of each layer, it is necessary to fit the field data to a series of straight lines.

For a line $T = A + BX$ from N points (x_i, t_i), the least squares estimates of the time intercept (A) and the slope (B) parameters are given by

$$A = \frac{\Sigma x_i \Sigma x_i t_i - \Sigma x_i^2 \Sigma t_i}{[\Sigma x_i]^2 - N \Sigma x_i^2}, \text{ and} \qquad (1)$$

$$B = \frac{\Sigma t_i - A \cdot N}{\Sigma x_i}. \qquad (2)$$

For the special case of a line passing through the origin, these equations reduce to

$$A = 0.0 \text{ , and}$$
$$B = \frac{\sum x_i \sum t_i}{\left[\sum x_i\right]^2} \quad . \qquad (3)$$

The velocities for each layer are found by taking the reciprocal of the slope of the regression line. The depth to each refracting layer can then be found by two computational methods: the critical distance method or the time intercept method. Since the regression equations parameters were already stored in the computer, the time intercept method was used. As only two interfaces were of interest, the equations to compute these two depths were programed rather than a general, series equation used for a number of layers. The equations used (Dobrin, 1960, p. 73, 75) were

$$Z(1) = \frac{T_{i2}}{2} \frac{V(2) \cdot V(1)}{\sqrt{V(2)^2 - V(1)^2}} \text{ , and}$$
$$Z(2) = \frac{1}{2}\left[T_{i3} - 2\,Z(1)\frac{\sqrt{V(3)^2 - V(1)^2}}{V(3) \cdot V(1)}\right]\frac{V(3) \cdot V(2)}{\sqrt{V(3)^2 - V(2)^2}} \quad . \qquad (4)$$

where Z (1) = depth to first interface,

Z (2) = depth to second interface below Z (1),

T_{i2} = time intercept for the second layer,

T_{i3} = time intercept for the third layer,

V (1) = seismic velocity in the first layer, V (2) = seismic velocity in the second layer,

V (3) = seismic velocity in the third layer.

Estimate of the Uncertainty in the Slopes and Time Intercepts

Ninety-five percent confidence limits were used to estimate the error in the slope and time intercept determination. To compute the confidence limits, the mean square of the deviations about the regression line was used to estimate the variance. The equation used to compute this residual mean square was

$$S^2_{t|x} = \frac{1}{(N-2)}\left[\frac{\sum(t_i-\bar{t})^2 - \sum\left[(x_i-\bar{x})(t_i-\bar{t})\right]^2}{\sum(x_i-\bar{x})^2}\right], \text{ where} \qquad (5)$$

N = number of points,

t_i = individual times where i = 1, 2, 3, ..., N,

X_i = individual distances where i = 1, 2, 3, ..., N,

$\bar{t}$ = average of the individual times,

$\bar{x}$ = average of the individual distances.

The standard deviation of the slope (S_B) was computed as

$$S_B = \sqrt{\frac{S^2_{t|x}}{\sum(x_i-\bar{x})^2}}, \qquad (6)$$

and the standard deviation of the time intercept (S_A) as

$$S_A = \sqrt{S^2_{t|x}\left[\frac{1}{N} + \frac{\bar{x}^2}{\sum(x_i - \bar{x})^2}\right]}\,. \qquad (7)$$

The confidence limits (CL) on the slope and time intercept parameters of the velocity lines were then computed as

$$\begin{aligned} CL &= A \pm t_{(N-2),0.05}\, S_A \text{ , and} \\ CL &= B \pm t_{(N-2),0.05}\, S_B \text{ , where} \end{aligned} \qquad (8)$$

$t_{(N-2),\ 0.05}$ is the Students's t-distribution with (N-2) degrees of freedom.

Estimate of the Variance of the Depth to the First Interface

Because the equations used to compute the depth are non-linear, the variance in the depth can only be estimated. The estimate of the variance of the depth to the water table was approximated with a Taylor's series of only first order terms. As the variance of the time intercept and slopes were already calculated, the depth equation for the first interface was written in terms of these paramaters. A discussion of the general method for estimating this type of variance can be found in Bowker and Lieberman (1959). The equation used to estimate the variance in the depth was

$$V(Z) = \frac{\partial Z}{\partial t} V(t) + \frac{\partial Z}{\partial B(1)} V\left[B(1)\right] + \frac{\partial Z}{\partial B(2)} V\left[B(2)\right], \qquad (9)$$

where V (Z) = estimated variance of depth,

V (t) = variance in time intercept,

V [B (1)] = variance in slope of velocity line in first layer,

V [B (2)] = variance in slope of velocity line in second layer.

The equations used to compute the partial derivatives were

$$\frac{\partial Z}{\partial t} = \left[2 \cdot B(1) \cdot B(2) \sqrt{\frac{1}{B(2)^2} - \frac{1}{B(1)^2}}\right]^{-1},$$

$$\frac{\partial Z}{\partial B(1)} = \frac{t}{2} \; \frac{B(2)\sqrt{\frac{1}{B(2)^2} - \frac{1}{B(1)^2}} + \frac{B(2)}{B(1)^2\sqrt{\frac{1}{B(2)^2} - \frac{1}{B(1)^2}}}}{\frac{1}{B(2)^2} - \frac{1}{B(1)^2}}, \qquad (10)$$

$$\frac{\partial Z}{\partial B(2)} = \frac{t}{2} \; \frac{B(1)\sqrt{\frac{1}{B(2)^2} - \frac{1}{B(1)^2}} - \frac{B(1)}{B(2)^2\sqrt{\frac{1}{B(2)^2} - \frac{1}{B(1)^2}}}}{\frac{1}{B(2)^2} - \frac{1}{B(1)^2}},$$

where t = time intercept,

B (1) = slope of velocity line in first layer,

B (2) = slope of velocity line in second layer.

APPENDIX C

TABLE I

DEPTHS AND ESTIMATED VARIANCE OF THE
DEPTHS TO THE WATER TABLE

Station	Depth (feet)	Variance (feet)*
1E	28.19	0.38
1W	20.57	0.58
2E	26.71	0.71
2W	17.85	1.95
3E	20.01	1.26
3W	18.73	0.44
4E	20.23	2.45
4W	24.03	0.57
5E	23.13	0.46
5W	25.88	1.34
6E	25.57	0.27
6W	23.74	0.89
7E	20.45	0.20
7W	20.76	0.00
8E	28.84	0.42
8W	22.01	1.20
9E	27.67	1.36
9W	25.37	0.62
10E	27.90	0.94
10W	23.68	0.50
11E	19.98	0.33
11W	18.73	0.67
12E	22.99	1.06
12W	21.52	0.82
13E	21.70	0.34
13W	23.11	0.69
14E	24.46	0.92
14W	19.46	0.08
15E	27.00	0.69
15W	21.59	0.88
16E	30.40	2.19
16W	24.16	0.72
17E	23.35	0.91
17W	21.72	0.42
18E	23.55	0.27
18W	20.54	0.32
19E	16.18	4.61
19W	22.52	0.63

TABLE I (Continued)

Station	Depth (feet)	Variance (feet)*
30S	25.69	0.73
30N	28.70	0.47
40E	27.13	0.50
40W	20.65	1.26

*-The variance is plus or minus.
E-Eastern shot point of profile
W-Western shot point of profile
N-Northern shot point of profile
S-Southern shot point of profile

TABLE II

SEISMIC VELOCITIES-COMPRESSIONAL
(Feet per second)

Profile	Dry Alluvium	Saturated Alluvium
1	1020	4894
2	1012	4504
3	1165	5470
4	1126	5176
5	1165	5412
6	1114	5608
7	1096	4867
8	1128	4132
9	1254	4462
10	1111	5605
11	1097	4615
12	1142	4626
13	1126	4700
14	1126	5030
15	1143	4482
16	1143	3933
17	1041	4677
18	1143	4919
19	1128	4014
30	1119	4915
40	1161	5023