# Techniques of Water-Resources Investigations of the United States Geological Survey

Chapter D2

# APPLICATION OF SEISMIC-REFRACTION TECHNIQUES TO HYDROLOGIC STUDIES

By F.P. Haeni

Book 2

COLLECTION OF ENVIRONMENTAL DATA

DEPARTMENT OF THE INTERIOR

DONALD PAUL HODEL, *Secretary*

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# PREFACE

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The unit of publication, the Chapter, is limited to a narrow field of subject matter. This format permits flexibility in revision and publication as the need arises. "Application of Seismic-Refraction Techniques to Hydrologic Studies" is Chapter D2 of Book 2.

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This manual is intended to supplement the more general "Application of Surface Geophysics to Ground-Water Investigations," by A.A.R. Zohdy, G.P. Eaton, and D.R. Mabey (U.S. Geological Survey Techniques of Water-Resources Investigations, Book 2, Chapter D1, 1974).

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TWI 3-A10. Discharge ratings at gaging stations, by E.J. Kennedy. 1984. 59 pages.
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[1]Spanish translation also available.

TWI 5-A1. Methods for determination of inorganic substances in water and fluvial sediments, by M.W. Skougstad and others, editors. 1979. 626 pages.

TWI 5-A2. Determination of minor elements in water by emission spectroscopy, by P.R. Barnett and E.C. Mallory, Jr. 1971. 31 pages.

TWI 5-A3. Methods for the determination of organic substances in water and fluvial sediments, edited by R.L. Wershaw, M.J. Fishman, R.R. Grabbe, and L.E. Lowe. 1987. 80 pages. This manual is a revision of "Methods for Analysis of Organic Substances in Water" by Donald F. Goerlitz and Eugene Brown, Book 5, Chapter A3, published in 1972.

TWI 5-A4. Methods for collection and analysis of aquatic biological and microbiological samples, edited by P.E. Greeson, T.A. Ehlke, G.A. Irwin, B.W. Lium, and K.V. Slack. 1977. 332 pages.

TWI 5-A5. Methods for determination of radioactive substances in water and fluvial sediments, by L.L. Thatcher, V.J. Janzer, and K.W. Edwards. 1977. 95 pages.

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TWI 7-C1. Finite difference model for aquifer simulation in two dimensions with results of numerical experiments, by P.C. Trescott, G.F. Pinder, and S.P. Larson. 1976. 116 pages.

TWI 7-C2. Computer model of two-dimensional solute transport and dispersion in ground water, by L.F. Konikow and J.D. Bredehoeft. 1978. 90 pages.

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TWI 8-A1. Methods of measuring water levels in deep wells, by M.S. Garber and F.C. Koopman. 1968. 23 pages.

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TWI 8-B2. Calibration and maintenance of vertical-axis type current meters, by G.F. Smoot and C.E. Novak. 1968. 15 pages.

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# ABBREVIATIONS AND CONVERSION FACTORS

Factors for converting inch-pound units to International System of Units (SI) and abbreviation of units:

| *Multiply inch-pound unit* | *By* | *To obtain SI (metric) unit* |
|---|---|---|
| | *Length* | |
| foot (ft) | 0.3048 | meter (m) |
| mile (mi) | 1.609 | kilometer (km) |
| | *Velocity* | |
| foot per second (ft/s) | 0.3048 | meter per second (m/s) |
| | *Density* | |
| ounce per cubic inch ($oz/in^3$) | 1.7297 | gram per cubic centimeter ($g/cm^3$) |

# GLOSSARY

**Angle of incidence.** The acute angle between a raypath and the normal to an interface.

**Apparent velocity.** The velocity at which a fixed point on a seismic wave, usually its front or beginning, passes an observer.

**Blind zone.** A layer having lower seismic velocity than overlying layers so that it does not carry a head wave.

**Conductivity.** The property of a material that allows the flow of electrical current.

**Critical angle.** The angle of incidence at which a refracted ray just grazes the interface between two media having different seismic velocities; equal to $\sin^{-1} V_1/V_2$.

**Critical distance.** The offset at which reflection occurs at the critical angle.

**Crossover distance.** The source-to-receiver distance at which refracted waves following a deep high-speed marker overtake direct waves, or refracted waves, following shallower markers.

**Geophone spacing.** The distance between adjacent geophones within a spread.

**Geophone spread.** The arrangement of geophones in relation to the position of the energy source.

**Head wave.** A wave characterized by entering and leaving a high-velocity medium at the critical angle.

**Isotropic.** A substance that has the same physical properties regardless of the direction of measurement.

**Reflection.** Energy from a seismic source that has been reflected from an acoustic impedence contrast between layers within the Earth.

**Resistivity.** The property of a material that inhibits the flow of electrical current. Resistivity is the reciprocal of conductivity.

**Stack.** A composite seismic record made by combining traces from different shots.

**Unconsolidated.** Loose material of the Earth's surface; uncemented particles of solid matter.

**Weathered layer.** Zone near the Earth's surface characterized by a low seismic-wave velocity beneath which the velocity abruptly increases, more properly called the low-velocity layer.

# APPLICATION OF SEISMIC-REFRACTION TECHNIQUES TO HYDROLOGIC STUDIES

*By* F.P. Haeni


## Abstract

During the past 30 years, seismic-refraction methods have been used extensively in petroleum, mineral, and engineering investigations and to some extent for hydrologic applications. Recent advances in equipment, sound sources, and computer interpretation techniques make seismic refraction a highly effective and economical means of obtaining subsurface data in hydrologic studies. Aquifers that can be defined by one or more high-seismic-velocity surface, such as (1) alluvial or glacial deposits in consolidated rock valleys, (2) limestone or sandstone underlain by metamorphic or igneous rock, or (3) saturated unconsolidated deposits overlain by unsaturated unconsolidated deposits, are ideally suited for seismic-refraction methods. These methods allow economical collection of subsurface data, provide the basis for more efficient collection of data by test drilling or aquifer tests, and result in improved hydrologic studies.

This manual briefly reviews the basics of seismic-refraction theory and principles. It emphasizes the use of these techniques in hydrologic investigations and describes the planning, equipment, field procedures, and intrepretation techniques needed for this type of study. Furthermore, examples of the use of seismic-refraction techniques in a wide variety of hydrologic studies are presented.


# Introduction

Surface geophysical techniques have been used extensively in the petroleum, mineral, and engineering fields. Hydrologic investigations have used surface geophysical techniques in the past, but to only a limited degree. Recent advances in electronic equipment and computer-interpretation programs and the development of new techniques make surface geophysics a more effective tool for hydrologists. These techniques should be considered in the project planning process and used where appropriate. Treated as a tool, similar to pump tests, simulation modeling, test drilling, geologic maps, borehole geophysical techniques, and so forth, these techniques can be used to help solve hydrologic problems.

Classically, surface geophysical techniques have been used early in the exploration process, prior to use of more expensive data-collection techniques such as drilling (Jakosky, 1950). The use of surface geophysics in this manner minimizes expensive data-collection activities and results in more efficient hydrologic studies.

All surface geophysical methods measure some physical property of subsurface materials or fluids. Selection of the appropriate geophysical method is determined by the specific physical property of a hydrologic unit or by the differences between adjacent hydrologic units. Typical physical properties measured are electrical resistivity, electrical conductivity, velocity of sound, gravity fields, and magnetic fields. Knowledge of the physical properties of a subsurface material is critical for successful application of surface geophysical methods. Aquifers that can be defined by one or more high-seismic-velocity surfaces, such as alluvial or glacial deposits in consolidated rock valleys, limestone or sandstone underlain by metamorphic or igneous rock, or saturated unconsolidated deposits overlain by unsaturated unconsolidated deposits, are ideally suited for seismic-refraction methods. In these hydrogeologic settings, seismic-refraction methods have proved to be the most useful of the surface geophysical techniques (Grant and West, 1965).

Seismic-refraction techniques were among the first geophysical tools used in the exploration for petroleum. In the 1920's, these techniques helped find many structures that were associated with petroleum accumulations. With the introduction and refinement of seismic-reflection techniques during the 1930's, use of refraction methods by the petroleum industry declined, and they are now used primarily in special situations and for weathered-layer velocity determinations.

Use of seismic-refraction techniques in engineering and hydrologic applications, and in coal exploration, has increased over the years, as has the wealth of literature on interpretation procedures. A bibliography by Musgrave (1967, p. 565–594) shows the extent of interest in, and the variety of applications of, seismic-refraction techniques.

Although seismic-reflection techniques have dominated deep-exploration work in recent years, shallow-exploration work has used seismic-refraction techniques extensively. Advances in the miniaturization of electronic equipment and the use of computers for data interpretation have made seismic-refraction techniques a very effective and economical exploration tool for hydrologists.

## Purpose and scope

A brief review of the literature indicates the diversity of seismic-refraction techniques. The purpose of this manual is to help the hydrologist who wishes to apply seismic refraction to a particular project or area of interest. It is intended to help the hydrologist determine if seismic-refraction techniques will work in a particular hydrologic setting. In addition, the manual briefly presents the theory of seismic refraction, identifies advantages and limitations of the techniques, describes the equipment and general field procedures required, and presents several interpretation procedures. Numerous references are cited to provide the reader with additional sources of information which are beyond the scope of this manual.

The techniques presented here are not standardized or rigid, but they have been used effectively in a wide variety of hydrologic studies conducted by the U.S. Geological Survey and others. References are included with each section so that alternative approaches to field procedures and interpretation methods can be investigated.

Ultimately, success in using seismic-refraction methods will depend more on the ability of the hydrologist to apply the principles of the techniques and to extract a hydrologically reasonable answer than on the use of a particular method of interpretation.

## Surface geophysical techniques in hydrologic studies

Surface geophysical techniques are used to obtain information about the subsurface units that control the location and movement of ground water.

A standard approach in exploration investigations is first to assess geologic conditions from available surface and subsurface geological data. From this initial study, the regional or local geologic framework can be hypothesized and the magnitude of the exploration problem defined.

At this point in a study, surface geophysical methods can be used to great advantage. The geologic and hydrologic model developed in this first stage of the study from scattered data points can be verified or, if necessary, modified. The importance of the interdependence of geological data, hydrologic data, and geophysical data cannot be overemphasized. Geophysical data by itself is susceptible to many interpretations. The input of hydrologic or geologic constraints may eliminate unreasonable interpretations and result in the selection of a unique solution.

Commonly, one or more surface geophysical techniques can be used advantageously in a hydrologic investigation. Papers describing the use of individual and combined surface geophysical techniques in hydrologic studies include those of Bonini and Hickok (1958), Eaton and Watkins (1967), Lennox and Carlson (1967), Mabey (1967), Ogiluy (1967), Shiftan (1967), Kent and Sendlein (1972), Zohdy and others (1974), Worthington (1975), and Collett (1978).

The two types of surface geophysical techniques that have been used most widely in hydrologic studies are resistivity methods and seismic-refraction methods. The general use of seismic-refraction methods in hydrologic studies has been discussed in the literature, and in cases in which velocity discontinuities between hydrologic units are present, these methods have proved to be the most useful geophysical technique. The major use of seismic-refraction techniques in hydrologic studies is to assess the hydrogeologic framework and hydrologic boundaries of aquifers. They are generally used early in the investigation, after the preliminary hydrologic assessment and prior to more site-specific data-gathering activities. Another use is for specific data-gathering activities later in the study. Specific information that may be sought during the hydrologic analysis stage of the study, and that can be investigated by seismic-refraction methods, are the depth to water in unconsolidated aquifers at specific locations and the location of aquifer boundaries.

After the geophysical work, the study is ready to enter its final stages when more costly, detailed site-specific data are collected. Generally, these stages of the study involve a drilling program, borehole geophysical studies, detailed hydrologic testing, and data analysis.

## References


Bonini, W.E., and Hickok, E.A., 1958, Seismic refraction method in ground-water exploration: Transactions of the American Institute of Mining, Metallurgical, and Petroleum Engineers, v. 211, p. 485–488.

Collett, L.S., 1978, Introduction to hydrogeophysics: Conference of the International Association of Hydrogeologists, Canadian chapter, Edmonton, September 30-October 4, 1978, Proceedings, p. 16–35.

Eaton, G.P., and Watkins, J.S., 1967, The use of seismic refraction and gravity methods in hydrologic investigations, *in* Morey, L.W., ed., Mining and ground-water geophysics: Geological Survey of Canada Economic Geology Report 26, p. 554–568.

Grant, F.S., and West, G.F., 1965, Interpretation theory in applied geophysics: New York, McGraw Hill, 583 p.

Jakosky, J.J., 1950, Exploration geophysics (2d ed.): Los Angeles, Calif., Times-Mirror Press, 1,195 p.

Kent, D.C., and Sendlein, L.V.A., 1972, A basin study of ground-water discharge from bedrock into glacial drift: Part 1, Definition of ground-water systems: Ground Water, v. 10, no. 4, p. 24–34.

Lennox, D.H., and Carlson, V., 1967, Integration of geophysical methods for ground-water exploration in the prairie provinces, Canada, *in* Morey, L.W., ed., Mining and ground-water geophysics: Geological Survey of Canada Economic Geology Report 26, p. 517–535.

Mabey, D.R., 1967, The role of geophysics in the development of the World's ground-water resources, *in* Morey, L.W., ed., Mining and ground-water geophysics: Geological Survey of Canada Economic Geology Report 26, p. 267–271.

Musgrave, A.W., ed., 1967, Seismic refraction prospecting: Tulsa, Okla., Society of Exploration Geophysicists, 604 p.

Ogiluy, A.A., 1967, Geophysical prospecting for ground water in the Soviet Union, *in* Morey, L.W., ed., Mining and ground-water geophysics: Geological Survey of Canada Economic Geology Report 26, p. 536–543.

Shiftan, Z.L., 1967, Integration of geophysics and hydrogeology in the solution of regional ground-water problems, *in* Morey, L.W., ed., Mining and ground-water geophysics: Geological Survey of Canada Economic Geology Report 26, p. 507–516.

Worthington, P.A., 1975, Procedures for the optimum use of geophysical methods in ground-water development programs: Association of Engineering Geologists Bulletin, v. 12, no. 1, p. 23–38.

Zohdy, A.A.R., Eaton, G.P., and Mabey, D.R., 1974, Application of surface geophysics to ground-water investigations: U.S. Geological Survey Techniques of Water-Resources Investigations, Book 2, Chapter D1, 116 p.

# Seismic-Refraction Theory and Limitations

## Theory

Numerous textbooks and journal articles present the details of seismic-refraction theory (Slotnick, 1959; Grant and West, 1965; Griffiths and King, 1965; Musgrave, 1967; Dobrin, 1976; Telford and others, 1976; Parasnis, 1979; Mooney, 1981). The following discussion reviews only the basic principles and limitations of seismic-refraction methods. The annotated bibliography at the end of this section should be used by hydrologists not familiar with seismic theory to select one or more publications that clearly present a rigorous theoretical development. An encyclopedic dictionary of terms used in exploration geophysics is published by the Society of Exploration Geophysicists (Sheriff, 1973).

It must be emphasized that the absence of an extensive section on the theory of seismic refraction does not minimize the importance of the topic. Hydrologists unfamiliar with geophysics must have a solid understanding of the physics underlying the technique prior to using it.

Seismic-refraction methods measure the time it takes for a compressional sound wave generated by a sound source to travel down through the layers of the Earth and back up to detectors placed on the land surface (fig. 1). By measuring the traveltime of the sound wave and applying the laws of physics that govern the propagation of sound, the subsurface geology can be inferred. The field data, therefore, will consist of measured distances and seismic traveltimes. From this time-distance information, velocity variations and depths to individual layers can be calculated and modeled.

The foundation of seismic-refraction theory is Snell's Law, which governs the refraction of sound or light waves across the boundary between layers having different velocities. As sound propagates through one layer and encounters another layer having faster seismic velocities, part of the energy is refracted, or bent, and part is reflected back into the first layer (see raypath 1 in fig. 1). When the angle of incidence equals the critical angle, the compressional energy is transmitted along the upper surface of the second layer at the velocity of sound in the second layer (see raypath 2 in fig. 1). As this energy propagates along the surface of layer 2, it generates new sound waves in the upper medium according to Huygens' principle, which states that every point on an advancing wave front can be regarded as the source of a sound wave; these new sound waves propagate back to the surface through layer 1 at an angle equal to the critical angle and at the velocity of sound in layer 1. When this refracted wave arrives at the land surface, it activates a geophone and arrival energy is recorded on a seismograph.

If a series of geophones is spread out on the ground in a geometric array, arrival times can be plotted against source-to-geophone distances (fig. 2), which results in a time-distance plot, or time-distance curve. It can be seen from figure 2 that at any distance less than the crossover distance ($x_c$) (sometimes incorrectly called the critical distance), the sound travels directly from the source to the detectors. This compressional wave travels a known distance in a known time, and the velocity of layer 1 can be directly calculated by $V_1 = x/t$, where $V_1$ is the velocity of sound in layer 1 and x is the distance a wave travels in layer 1 in time t. Figure 2 is a plot of time as a function of distance; consequently, $V_1$ is also equal to the inverse slope of the first line segment.

Beyond the crossover distance, the compressional wave that has traveled through layer 1, along the interface with the high-velocity layer, and then back up to the surface through layer 1 arrives before the compressional wave that has been in layer 1 (the low-velocity layer). All first compressional waves arriving at geophones more distant than the crossover distance will be refracted waves, or head waves, from layer 2 (the high-velocity layer). When these points are plotted on the time-distance plot, the inverse slope of this segment will be equal to the apparent velocity of layer 2. The slope of this line does not intersect the time axis at zero, but at some time called the intercept time ($t_i$). The intercept time and the crossover distance are directly dependent on the velocity of sound in the two materials and the thickness of the first layer, and therefore can be used to determine the thickness of the first layer (z).

## Interpretation formulas

Intercept times and crossover distance-depth formulas have been derived in the literature (Grant and West, 1965; Zohdy and others, 1974; Dobrin, 1976; Telford and

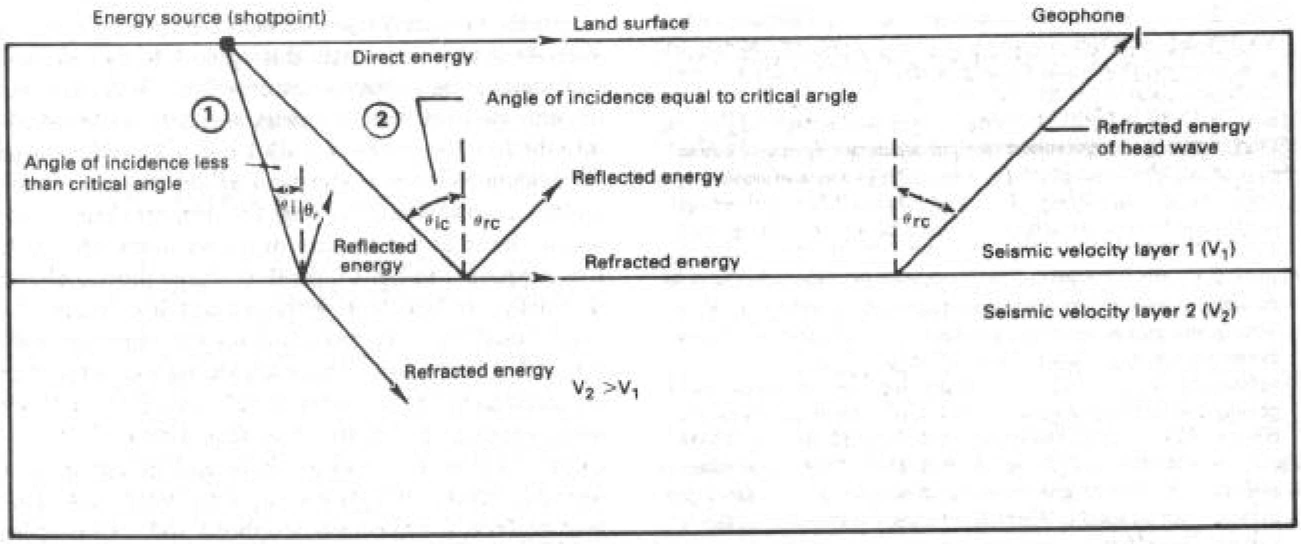



Figure 1.—Raypaths of refracted (1) and reflected (2) sound energy in a two-layer Earth.

others, 1976; Parasnis, 1979; Mooney, 1981), and only the results are given here. These derivations are straightforward inasmuch as the total traveltime of the sound wave is measured, the velocity in each layer is calculated from the time-distance plot, and the raypath geometry is known. The only unknown is the depth to the high-velocity refractor. These interpretation formulas are based on the following assumptions: (1) the boundaries between layers are planes that are either horizontal or dipping at a constant angle, (2) there is no land-surface relief, (3) each layer is homogeneous and isotropic, and (4) the seismic velocity of the layers increases with depth.

### Two-layer parallel-boundary formulas (See figure 3)

1. Intercept-time formula (Dobrin, 1976, p. 297):

$$z=\frac{t_i}{2}\frac{V_2\,V_1}{\sqrt{(V_2)^2-(V_1)^2}}, \tag{1}$$

where

$z$ = depth to layer 2 at point,
$t_i$ = intercept time,
$V_2$ = velocity of sound in layer 2, and
$V_1$ = velocity of sound in layer 1.

2. Crossover-distance formula (Dobrin, 1976, p. 298):

$$z=\frac{x_c}{2}\sqrt{\frac{V_2-V_1}{V_2+V_1}}, \tag{2}$$

where

$z$, $V_2$, and $V_1$ are as defined earlier and
$x_c$ = crossover distance.

### Three-layer parallel-boundary formulas (See figure 4)

1. Intercept-time formulas (Dobrin, 1976, p. 299):

$$z_1=\frac{t_2}{2}\frac{V_2V_1}{\sqrt{(V_2)^2-(V_1)^2}} \text{ (from two-layer formula 1),} \tag{3}$$

$$z_2=\frac{1}{2}\left(t_3-\frac{2z_1\sqrt{(V_3)^2-(V_1)^2}}{V_3V_1}\right)\frac{V_3V_2}{\sqrt{(V_3)^2-(V_2)^2}}, \tag{4}$$

and

$$z_3=z_1+z_2, \tag{5}$$

where

$z_1$ = depth to layer 2, or thickness of layer 1,
$z_2$ = depth from bottom of layer 1 to top of layer 3, or thickness of layer 2,
$z_3$ = depth from surface to top of layer 3,
$t_2$ = intercept time for layer 2,
$t_3$ = intercept time for layer 3,
$V_1$ = velocity of sound in layer 1,
$V_2$ = velocity of sound in layer 2, and
$V_3$ = velocity of sound in layer 3.

2. Crossover-distance formulas (Parasnis, 1979, p. 197–198):

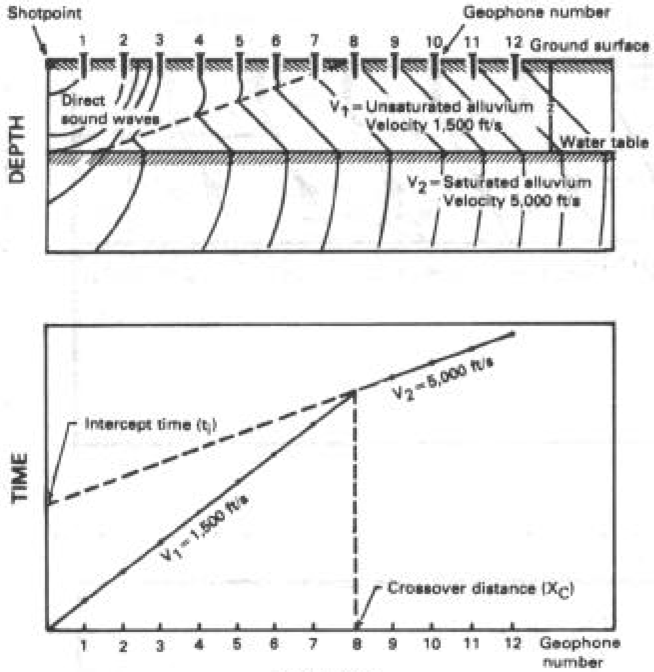



Figure 2.—Seismic wave fronts and raypaths and corresponding time-distance plot.

$$z_1=\frac{x_{c1}}{2}\sqrt{\frac{V_2-V_1}{V_2+V_1}} \text{ (from two-layer formula 2),} \qquad (6)$$

$$z_2=\frac{x_{c2}}{2}\left(\frac{V_3-V_2}{\sqrt{(V_3)^2-(V_2)^2}}\right)$$
$$-z_1\left(\frac{V_2\sqrt{(V_3)^2-(V_1)^2}-V_3\sqrt{(V_2)^2-(V_1)^2}}{V_1\sqrt{(V_3)^2-(V_2)^2}}\right), \qquad (7)$$

and

$$z_3=z_1+z_2\ , \qquad (8)$$

where

$z_1, z_2, z_3, V_1, V_2$, and $V_3$ are as defined earlier,
$x_{c1}$ = crossover distance between layers 1 and 2, and
$x_{c2}$ = crossover distance between layers 2 and 3.

Other forms of this equation are presented by Mooney (1981) and Alsop (1982).

## Two-layer dipping-boundary formulas (See figure 5)

The problem presented by a dipping boundary between layers adds some geometric complexity to the derivation of these formulas. Several important concepts of seismic-refraction theory must be introduced at this point.

To learn about the geometry of a dipping boundary, the refraction profile must be reversed. For a single array, a minimum of two shots must be fired, one from each end of the array. This concept is termed "reversed-profile shooting," and the practice should be followed routinely in all seismic-refraction studies. Failure to reverse seismic profiles leads to invalid results in almost all situations. Figure 5 shows a two-layer dipping-boundary model and the resultant time-distance plot. A fundamental rule of seismic-refraction theory is illustrated in figure 5. The total traveltime of compressional sound waves from shotpoint D to shotpoint U, and in the opposite direction, from shotpoint U to shotpoint D, must be equal; that is, $T_u$ must equal $T_d$ because the same wave path is followed in each case. Comparison of the crossover distances or the intercept times on this plot ($x_{cu} > x_{cd}$ and $t_{2u} > t_{2d}$) shows that layer 2 is deeper at shotpoint 2 than at shotpoint 1, and a dipping-layer analysis must be used. If these values were equal and the segments of the time-distance plots were straight lines, then simple two-layer parallel-boundary formulas could be used.

In the parallel-boundary problems discussed previously, the seismic velocity measured on time-distance plots was in fact the true velocity of the horizontal refracting layer. When the interface is dipping, however, seismic-refraction methods measure the apparent seismic velocity and not the true seismic velocity. The true seismic velocity is the harmonic mean of the measured apparent updip and downdip velocities multiplied by the cosine of the dip angle. It can be determined by the following formula:

$$V_2=\frac{2V_{2u}V_{2d}}{V_{2u}+V_{2d}}\cos\xi \text{ (Redpath, 1973; Mooney, 1981, p. 10–4),} \qquad (9)$$

where

$V_2$ = true velocity of sound in layer 2,
$V_{2u}$ = apparent updip velocity of sound (from time-distance plot),
$V_{2d}$ = apparent downdip velocity of sound (from time-distance plot), and
$\xi$ = dip angle of layer 2.

A good approximation of the velocity of sound in layer 2 is the harmonic mean, since the cosine of small angles is very close to 1.0. Equation 9 reduces to

$$V_2=\frac{2V_{2u}V_{2d}}{V_{2u}+V_{2d}} \text{ (Redpath, 1973, p. 9).} \qquad (10)$$

The depth to the dipping interface can be calculated by using the following formulas:

1. Intercept-time formulas (Dobrin, 1976, p. 304):

$$\text{(a)} \quad \Theta_c=\tfrac{1}{2}\left(\sin^{-1} V_1 m_d+\sin^{-1} V_1 m_u\right), \qquad (11)$$

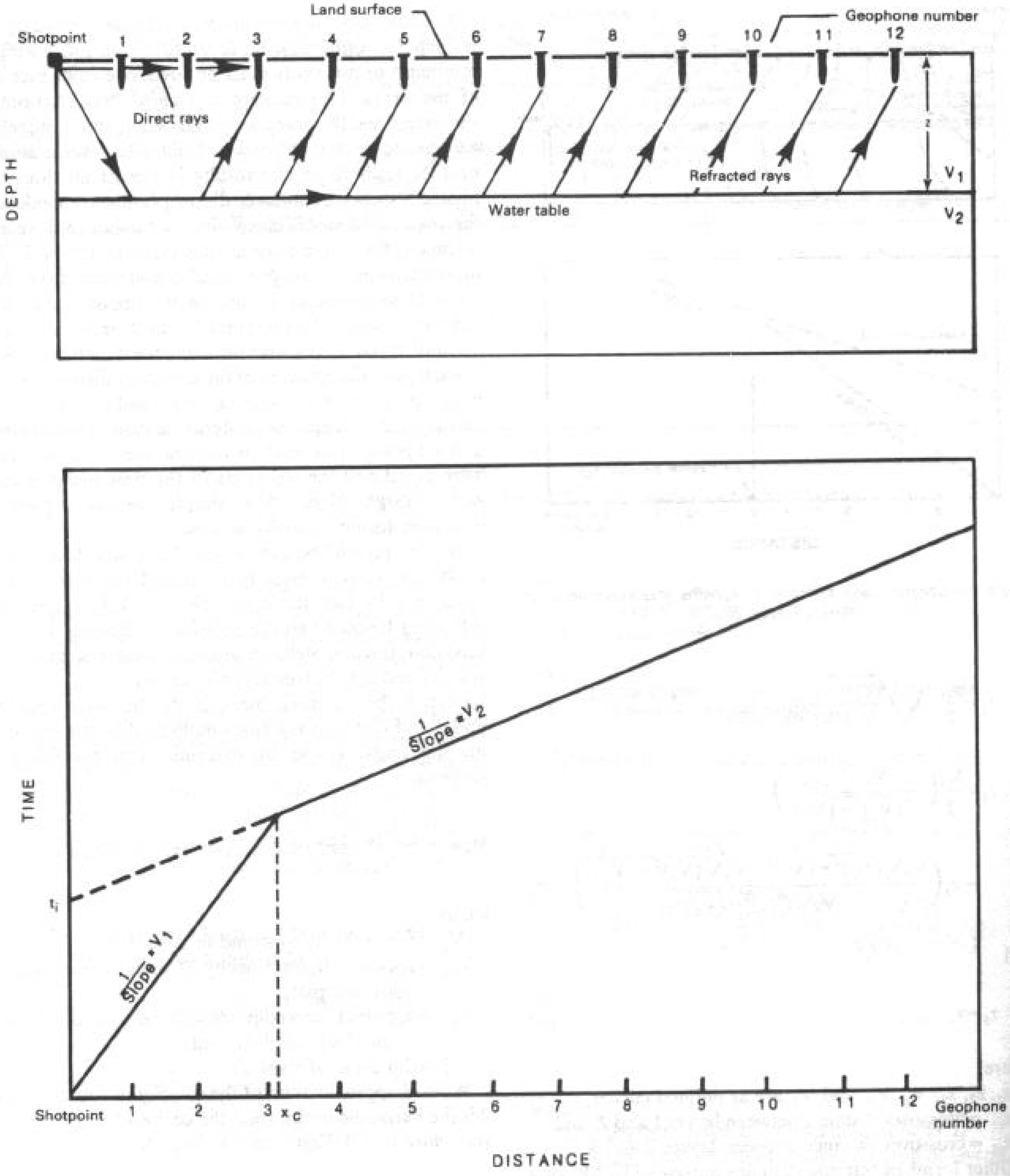



Figure 3.—Seismic raypaths and time-distance plot for a two-layer model with parallel boundaries.

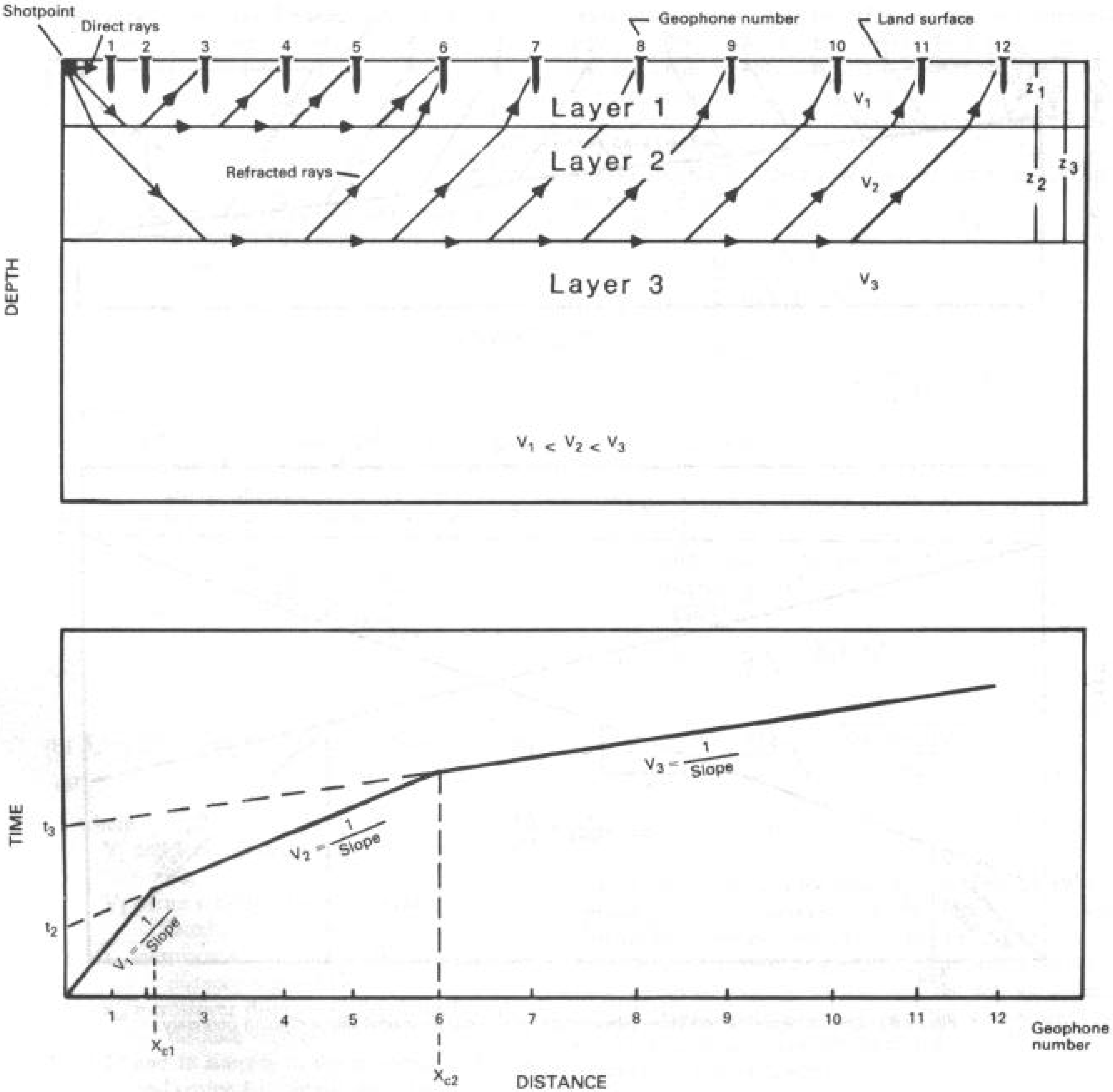



Figure 4.—Seismic raypaths and time-distance plot for a three-layer model with parallel boundaries.

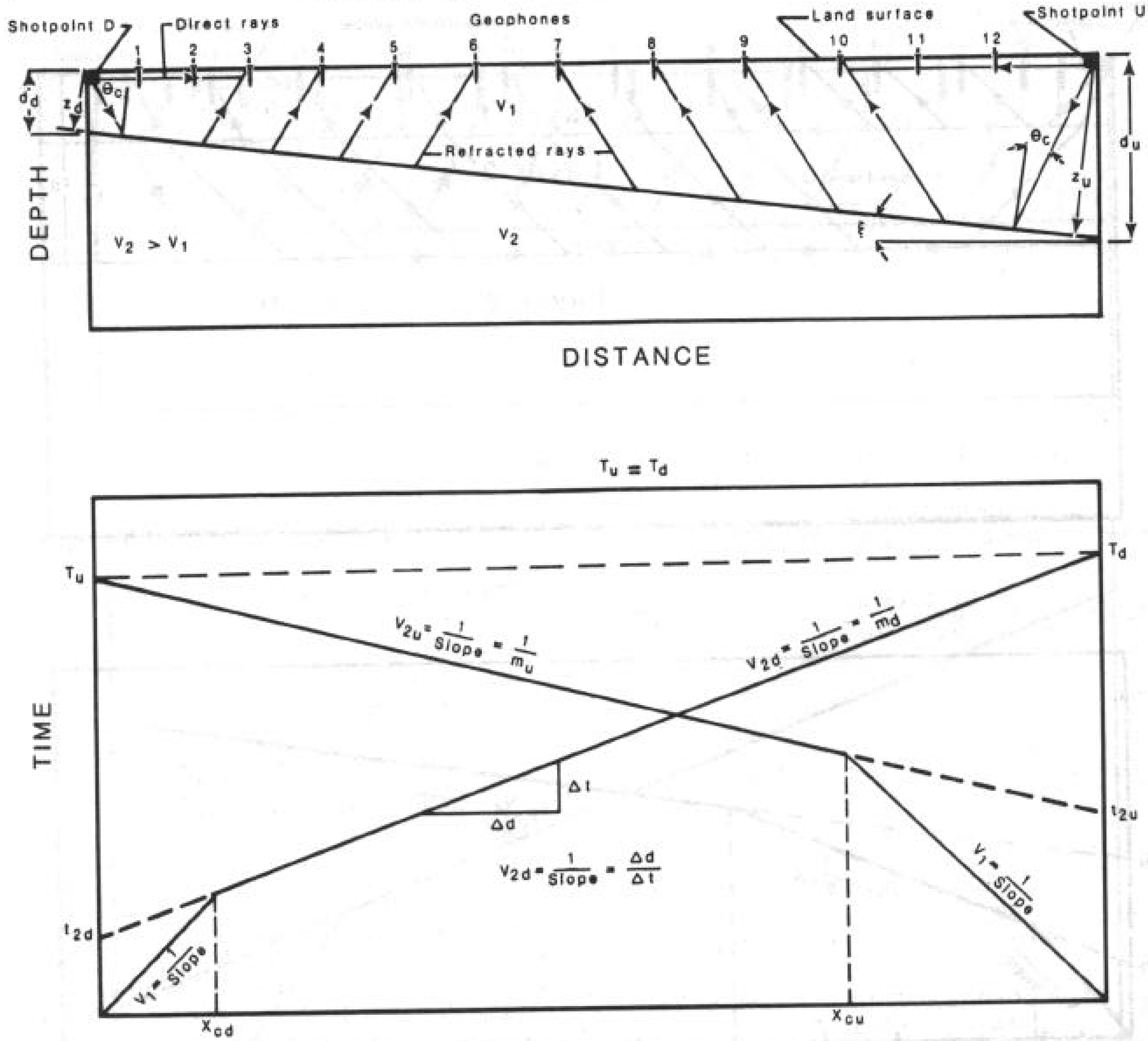



Figure 5.—Seismic raypaths and time-distance plot for a two-layer model with a dipping boundary.

where

$\Theta_c$ = critical angle,

$V_1$ = true velocity of sound in layer 1 (from time-distance plot),

$m_d$ = slope of downdip $V_2$ segment on time-distance plot, and

$m_u$ = slope of updip $V_2$ segment on time-distance plot.

(b) $$\xi = \tfrac{1}{2}\left(\sin^{-1}V_1m_d - \sin^{-1}V_1m_u\right), \tag{12}$$

where

$\xi$ = dip angle of the refractor.

(c) $$z_u = \frac{V_1t_{2u}}{2\cos\Theta_c}, \tag{13}$$

where

$z_u$ = perpendicular distance to refractor at the updip shotpoint (shotpoint 2) and

$t_{2u}$ = intercept time of updip $v_2$ segment of time-distance plot.

(d) $$z_d = \frac{V_1t_{2d}}{2\cos\Theta_c}, \tag{14}$$

where

$z_d$ = perpendicular distance to refractor at downdip shotpoint (shotpoint 1) and

$t_{2d}$ = intercept time of downdip $V_2$ segment of time-distance plot.

(e) $$d_u = \frac{z_u}{\cos \xi}, \tag{15}$$

where
$d_u$ = extrapolated vertical depth to the refractor beneath shotpoint on updip side (shotpoint 2).

(f) $$d_d = \frac{z_d}{\cos \xi}, \tag{16}$$

where
$d_d$ = extrapolated vertical depth to the refractor beneath shotpoint on downdip side (shotpoint 1).

2. Crossover-distance formulas (Mooney, 1981, p. 10–8):

(a) $$d_u = \frac{x_{cu}}{2 \cos \xi} \frac{V_2-(V_1 \cos \xi)}{\sqrt{(V_2)^2-(V_1)^2}} + \frac{x_{cu}}{2} \tan \xi \tag{17}$$

and

(b) $$d_d = \frac{x_{cd}}{2 \cos \xi} \frac{V_2-(V_1 \cos \xi)}{\sqrt{(V_2)^2-(V_1)^2}} - \frac{x_{cd}}{2} \tan \xi, \tag{18}$$

where
$V_1$ and $\xi$ are as defined for equations 11 and 12,
$V_2$ = true velocity of sound in layer 2 (calculated),
$x_{cu}$ = crossover distance of the updip time-distance segment, and
$x_{cd}$ = crossover distance of the downdip time-distance segment.

Equations 17 and 18 simplify to the following if the dip angle is small and cosine $\xi$ is almost equal to 1.0:

(c) $$d_u = \frac{x_{cu}}{2}\sqrt{\frac{V_2-V_1}{V_2+V_1}} + \frac{x_{cu}}{2} \sin \xi \tag{19}$$

and

(d) $$d_d = \frac{x_{cd}}{2}\sqrt{\frac{V_2-V_1}{V_2+V_1}} - \frac{x_{cd}}{2} \sin \xi. \tag{20}$$

**Example problem**

The following example illustrates the use of these formulas and demonstrates the need for choosing the formula most applicable to the field situation.

A. The time-distance plot in figure 6 is obtained in the field by firing only one shot at one end of a seismic-refraction line. If only one shot in one direction is fired, the interpreter would have to use a two-layer horizontal interpretation formula to determine the depth to the refracting layer.

(1) Using the intercept-time formula (eq. 3) to find the depth to the refractor,

$$z = \frac{t_1}{2} \frac{V_2 V_1}{\sqrt{(V_2)^2-(V_1)^2}}$$

$$= \frac{0.0075}{2} \frac{10{,}600(5{,}000)}{\sqrt{(10{,}600)^2-(5{,}000)^2}}$$

$$= 21 \text{ ft.}$$

The depth to rock is determined to be 21 ft along the entire profile.

(2) Similar results are obtained using the crossover-distance formula (eq. 6):

$$z = \frac{x_c}{2}\sqrt{\frac{V_2-V_1}{V_2+V_1}}$$

$$= \frac{70.4}{2}\sqrt{\frac{10{,}600-5{,}000}{10{,}600+5{,}000}}$$

$$= 21 \text{ ft.}$$

B. A shot fired from the opposite end of the geophone spread produces a reversed profile. The time-distance plot shown in figure 7 was plotted from the field data.

(1) Using the two-layer, dipping-interface, intercept-time formulas (eqs. 9, 11–16) and the following data obtained from the time-distance plot, the correct depth to the dipping refractor can be calculated.

From the time-distance plot,

$t_{2u}$ = 0.0448 s
$t_{2d}$ = 0.0075 s
$V_1$ = 5,000 ft/s
$m_u$ = 0.0000375
$m_d$ = 0.0000945
$V_{2u} = \frac{1}{m_u}$ = 26,700 ft/s
$V_{2d} = \frac{1}{m_d}$ = 10,600 ft/s

(a) $$\xi = ½[\sin^{-1}(V_1 m_d) - \sin^{-1}(V_1 m_u)]$$

$$= ½[\sin^{-1} 5{,}000(0.0000945) - \sin^{-1} 5{,}000(0.0000375)]$$

$$= 8.75°$$

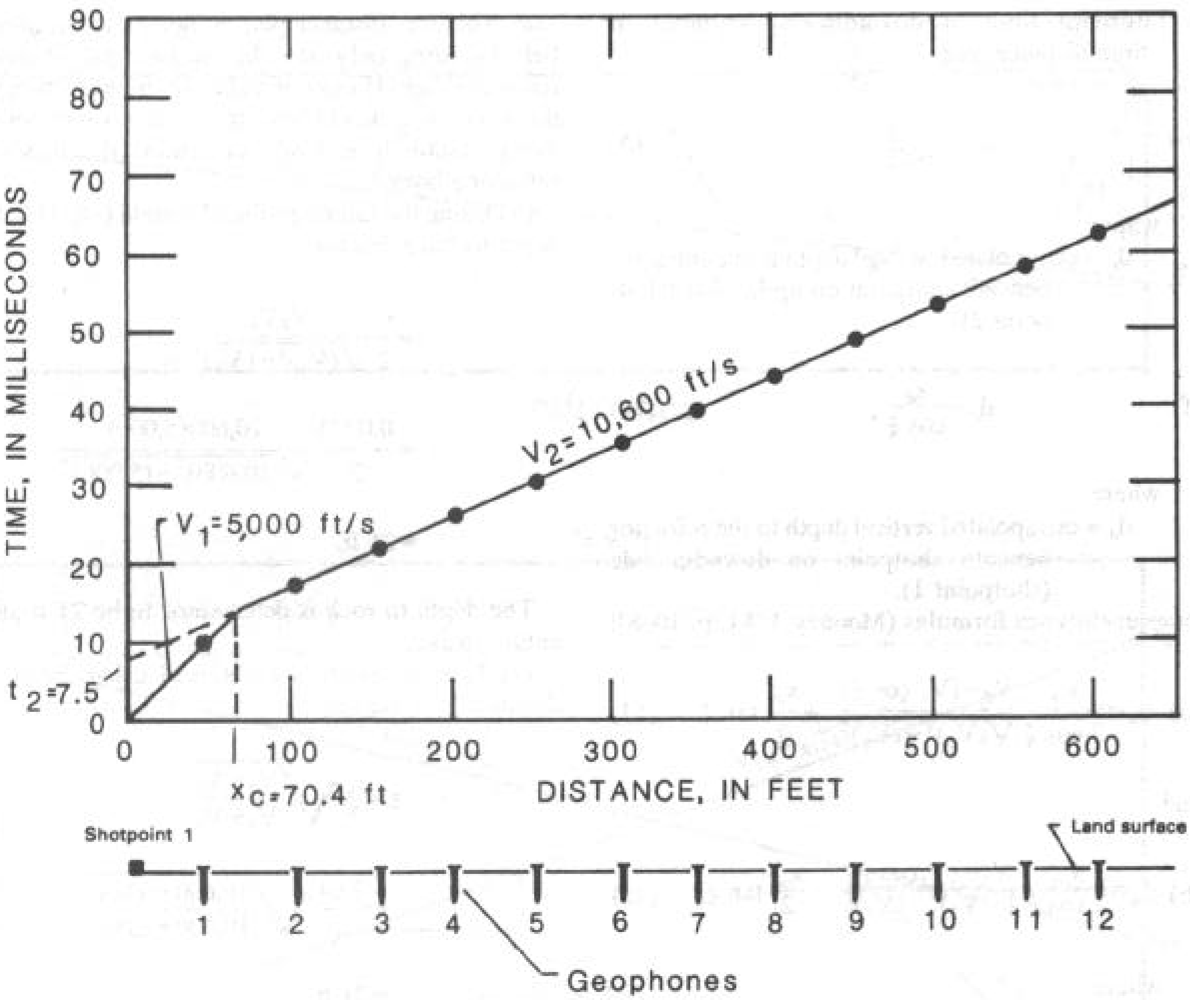



Figure 6.—Time-distance plot resulting from one shotpoint over a two-layer model with a dipping boundary.

(b) $$V_2=\frac{2V_{2u}V_{2d}}{V_{2u}+V_{2d}}\cos\xi$$

$$=\frac{2(26{,}700)(10{,}600)}{26{,}700+10{,}600}\cos 8.75$$

$$=15{,}000 \text{ ft/s}$$

(c) $$\Theta=½[\sin^{-1}(V_1m_d)+\sin^{-1}(V_1m_u)]$$

$$=½[\sin^{-1}5{,}000(0.0000945)+\sin^{-1}5{,}000(0.0000375)]$$

$$=19.5°$$

(d) $$z_u=\frac{V_1t_{2u}}{2\cos\Theta}=\frac{5{,}000(0.0448)}{2\cos 19.5}=118.8 \text{ ft}$$

(e) $$z_d=\frac{V_1\,t_{2d}}{2\cos\Theta}=\frac{5{,}000(0.0075)}{2\cos 19.5}=19.9 \text{ ft}$$

(f) $$d_u=\frac{z_u}{\cos\xi}=\frac{118.8}{\cos 8.7}=120 \text{ ft}$$

(g) $$d_d=\frac{z_d}{\cos\xi}=\frac{19.9}{\cos 8.7}=20 \text{ ft}$$

(2) Using the crossover-distance formulas (eqs. 17, 18) with the same field data, $d_u$ and $d_d$ can again be calculated.

From the time-distance plot,

$x_{cd}$ = 70.4 ft
$x_{cu}$ = 273.8 ft
$V_1$ = 5,000 ft/s

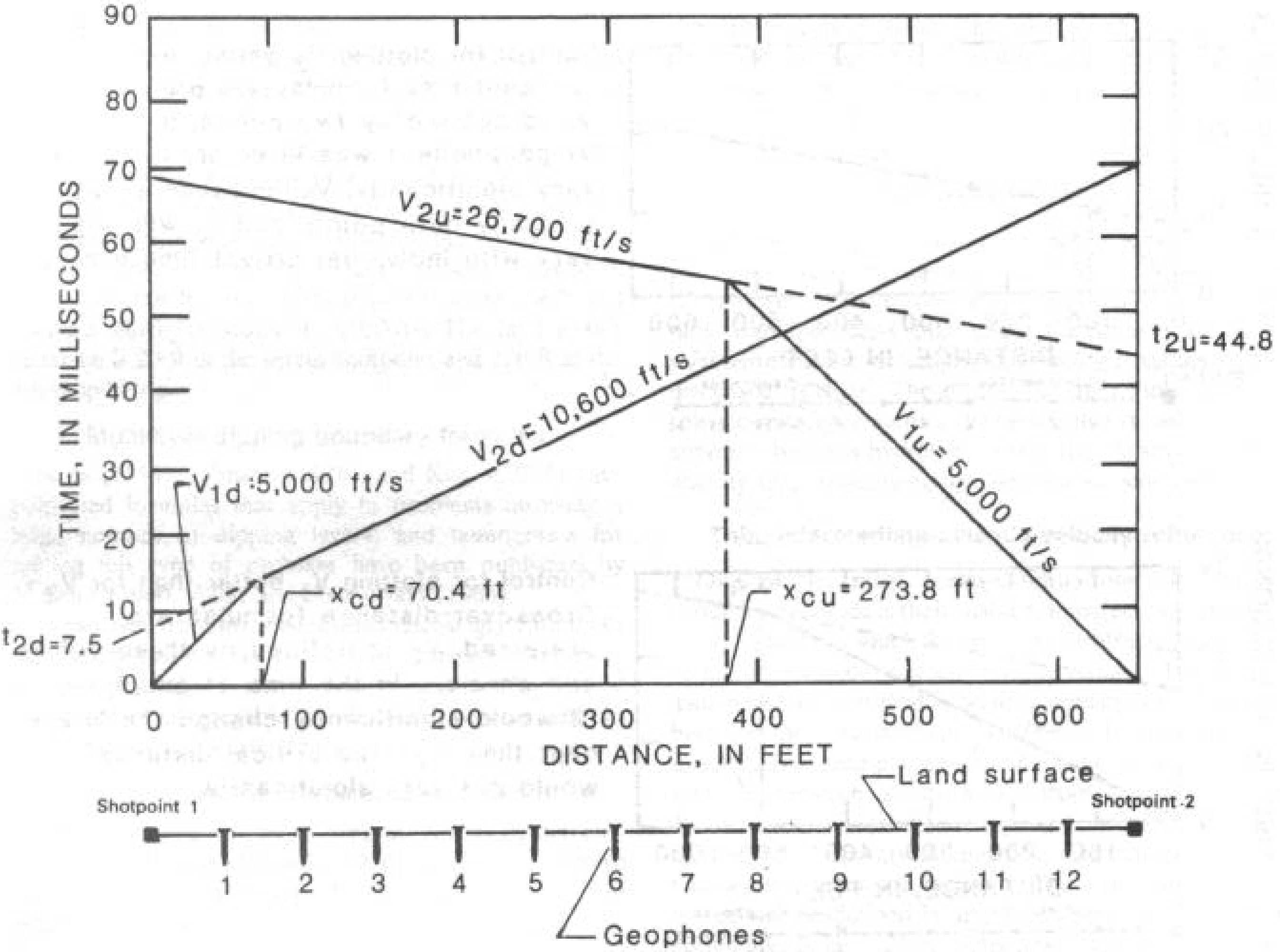



Figure 7.—Time-distance plot resulting from two reversed shots over the two-layer model with a dipping boundary illustrated in figure 5.

(a) $\xi = \frac{1}{2}[\sin^{-1}(V_1 m_d) - \sin^{-1}(V_1 m_u)]$

$= \frac{1}{2}[\sin^{-1} 5{,}000(0.0000945) - \sin^{-1} 5{,}000(0.0000375)]$

$= 8.75°$

(b) $V_2 = \frac{2 V_{2u} V_{2d}}{V_{2u} + V_{2d}} \cos \xi$

$= \frac{2(26{,}700)(10{,}600)}{26{,}700 + 10{,}600} \cos 8.75$

$= 15{,}000$ ft/s

(c) $d_u = \frac{x_{cu}}{2 \cos \xi} \cdot \frac{V_2 - (V_1 \cos \xi)}{\sqrt{(V_2)^2 - (V_1)^2}} + \frac{x_{cu} \tan \xi}{2}$

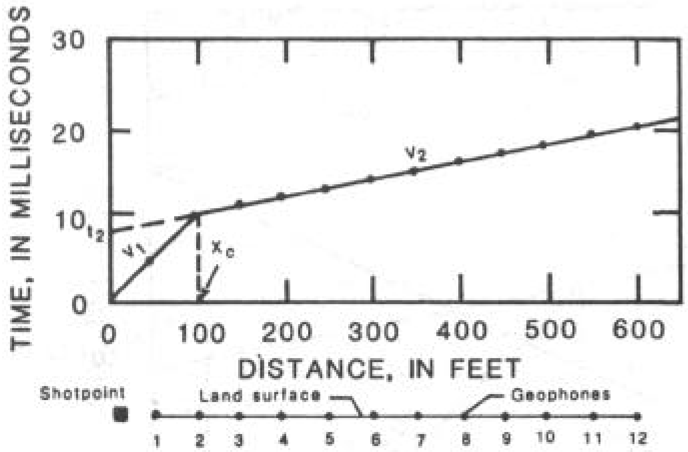
$= \frac{273.8}{2 \cos 8.75} \cdot \frac{15{,}000 - (5{,}000 \cos 8.75)}{\sqrt{(15{,}000)^2 - (5{,}000)^2}} + \frac{273.8 \tan 8.75}{2}$

$= 120$ ft

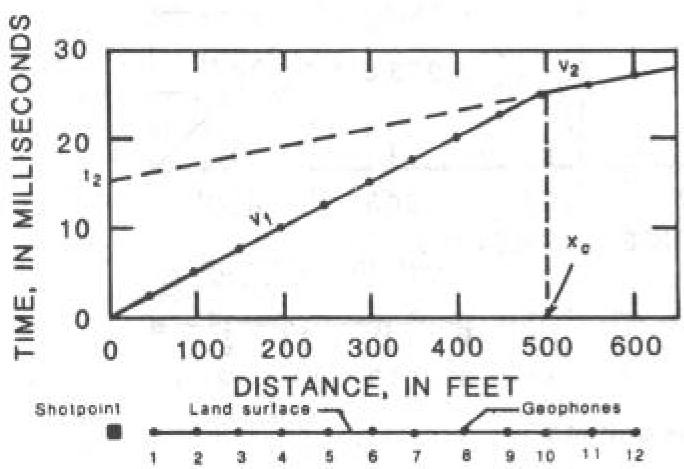
(d) $d_d = \frac{x_{cd}}{2 \cos \xi} \cdot \frac{V_2 - (V_1 \cos \xi)}{(V_2)^2 - (V_1)^2} - \frac{x_{cd} \tan \xi}{2}$

$= \frac{70.4}{2 \cos 8.75} \cdot \frac{15{,}000 - (5{,}000 \cos 8.75)}{(15{,}000)^2 - (5{,}000)^2} - \frac{70.4 \tan 8.75}{2}$

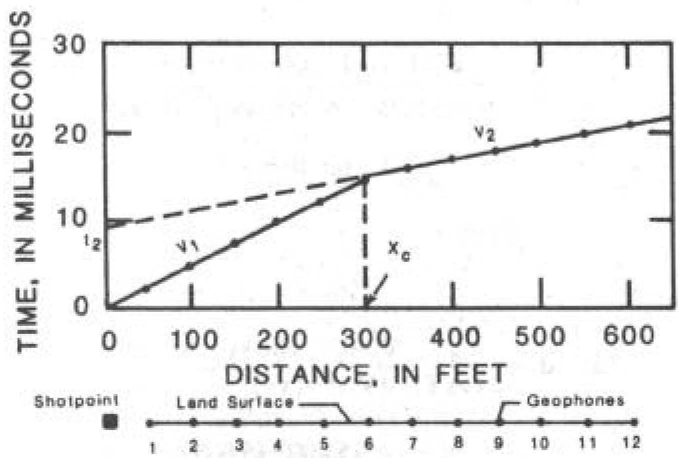
$= 20$ ft




Control for plotting $V_2$ better than for $V_1$-- Intercept-time formulas are preferred. $V_1$ is defined by two points. If the time at geophone 1 was in error, $x_c$ would vary significantly. $V_2$, however, is defined by many data points and $t_2$ will not vary with individual arrival time errors.




Control for plotting $V_1$ better than for $V_2$-- Crossover-distance formulas are preferred. $V_2$ is defined by three points and an error in the time of geophone 12 would significantly change the intercept time ($t_2$). The critical distance would not vary significantly.




Control for plotting $V_1$ and $V_2$ about the same-- Intercept-time and crossover-distance formulas are equal. All line segments are defined by about the same amount of data.

Figure 8.—Advantages and disadvantages of intercept-time versus crossover-distance formulas in determining depth to a refractor under different field conditions.

Summary of example problem:

1. Using a single-shot, nonreversed seismic-refraction profile and the two-layer parallel-boundary formulas, the interpretation gives a subsurface having a velocity of sound in layer 1 of 5,000 ft/s and a second horizontal layer 21 ft deep having a velocity of sound of 10,600 ft/s.

2. Using a reversed seismic-refraction profile and the two-layer dipping-boundary formulas, the correct interpretation gives a subsurface having a velocity of sound in layer 1 of 5,000 ft/s and a second layer dipping at 8.7° and having a velocity of sound of 15,000 ft/s. The depth to this interface is 20 ft at the updip shotpoint and 120 ft at the downdip shotpoint.

### Multilayer dipping-boundary formulas

Mota (1954), Johnson (1976), and Knox (1976) have published formulas that apply to problems involving a large number of dipping layers, and nomograms for solving this type of problem have been published by Meridav (1960, 1968) and Habberjam (1966).

In practice, however, it becomes increasingly difficult to distinguish between small, discrete changes in the time-distance plots that actually indicate different layers and small errors attributable to the field process and to nonhomogeneous Earth layers.

### Formulas for more complex cases

Other solutions for more complex situations are covered in the literature (Dobrin, 1976), but in general these do not apply to hydrologic problems and consequently are not covered here.

### Field corrections

In addition to the theoretical solutions to seismic-refraction problems, corrections for field-related problems have also been developed. The two main types of corrections are elevation corrections and weathering corrections. Both are used to adjust field-derived traveltimes to some selected datum planes, so that straight-line segments on the time-distance plot can be associated with subsurface refractors. These corrections can be applied manually (Dobrin, 1976, p. 335) or by computer (Scott and others, 1972).

### Summary

In this section, formulas for both intercept time and crossover distance were presented for determining the depth to a refractor. Several investigators have shown that, in general, the crossover-distance formulas are less prone to error than the intercept-time formulas (Zirbel, 1954; Meridav, 1960) because of the greater difficulty in determining the correct slope of the segments of the time-distance plot compared with determining the crossover distances. Telford and others (1976, p. 279), however, take the opposite view. The final choice of methods, therefore, depends on the quality and quantity of the data on the time-distance plot (Grant and West, 1965, p. 149–150). The time-distance plots shown in figure 8 illustrate the advantages and disadvantages of each method under several different field conditions.

## Limitations

Prior to using seismic-refraction techniques, certain problems and limitations need to be considered (Domzalski, 1956; Burke, 1967; Wallace, 1970). Three blind-zone problems that affect the success of using seismic-refraction techniques in hydrologic studies will be discussed further. These are (1) thin, intermediate seismic-velocity refractors, (2) insufficient seismic-velocity contrasts between hydrologic units, and (3) slow-seismic-velocity units underlying high-seismic-velocity units.

### Thin, intermediate-seismic-velocity refractor

One of the most serious limitations of seismic-refraction methods is their inability to detect intermediate layers in cases in which the layer has insufficient thickness or seismic-velocity contrast to return first-arrival energy. This problem is critical in water-resources investigations because the intermediate layer may be the zone of interest. For example, saturated unconsolidated aquifer material between unsaturated unconsolidated material and bedrock, or a sandstone aquifer between unconsolidated material and crystalline rock, may not be detected with seismic-refraction methods. These intermediate layers cannot be defined by any alternative location of the geophones or by shallow shotpoints. Deep shotholes may overcome this problem (Soske, 1959), but they are usually impractical under normal field conditions. If the presence of such a layer is suspected, however, calculations can be made to determine its minimum and maximum thickness. Figure 9 shows the wave-front and raypath diagram illustrating a situation in which a 70-ft-thick intermediate-seismic-velocity layer is not detected by first arrivals on the time-distance plot. If the intermediate layer is a thin, intermediate-seismic-velocity layer of till underlying a glacial aquifer, the thickness of the aquifer calculated from the refraction data will be in error (Sander, 1978). Successful interpretation of field data acquired in areas exhibiting this problem is dependent on the correlation of geophysical data with drill holes or knowledge of the local geology.

In the absence of drill-hole data, an unexpected velocity change in the time-distance plot should warn the hydrologist that a thin, intermediate-seismic-velocity layer may be present and that a qualified interpretation is in order. An example of this is shown in figure 10, in which the time-distance plot indicates that a thin, intermediate-seismic-velocity layer may exist, provided the interpreter knows something about the local geology and the speed of sound in the various earth materials near the study area.

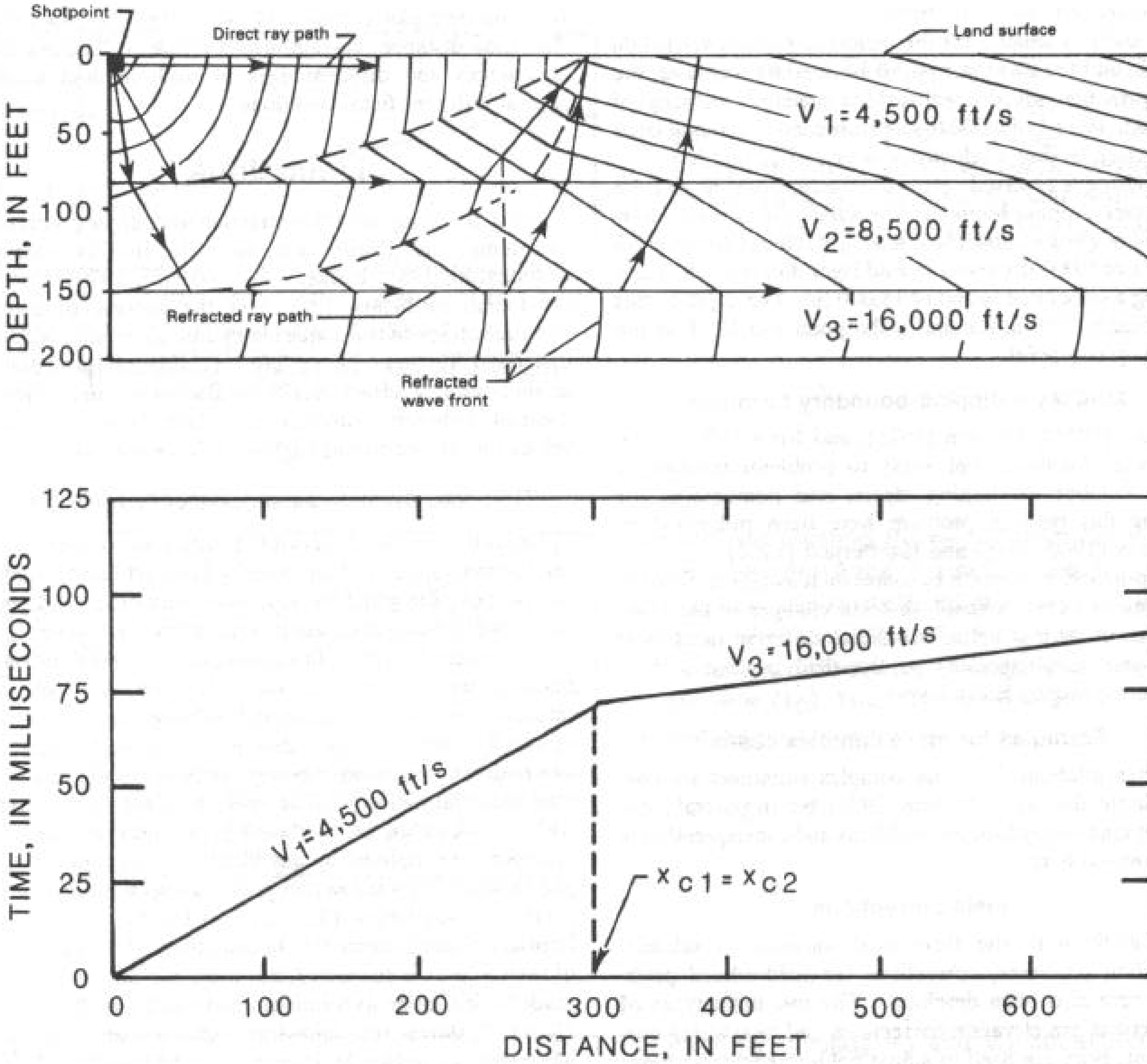



Figure 9.—Seismic wave fronts with selected raypaths and the corresponding time-distance plot for the case of an undetectable intermediate-seismic-velocity layer (modified from Soske, 1959, fig. 4, p. 362).

The case illustrated in figure 10 is very common in hydrologic studies. The unsaturated unconsolidated material has a velocity of 1,000 ft/s, the thin, saturated unconsolidated material has a velocity of about 5,000 to 6,000 ft/s (this layer is not detected by refraction techniques and is not shown in fig. 10), and the crystalline bedrock has a velocity of 15,000 ft/s.

If a thin, intermediate-seismic-velocity layer is suspected, methods are available for determining the maximum thickness of the undetected layer (Soske, 1959; Hawkins and Maggs, 1961; Green, 1962; Redpath, 1973; Mooney, 1981). The following example demonstrates the significance of this problem in water-resources investigations. The calculations in this example and in table 1 are based on a technique described by Mooney (1981, p. 94).

#### Example problem

The time-distance plot shown in figure 11 is plotted from field data, and the following values are obtained:

$x_c$ = 111 ft (from time-distance plot),
$V_1$ = 1,500 ft/s (from time-distance plot),
$V_3$ or $V_2$ = 15,000 ft/s (from time-distance plot), and
$V_2$ = 5,000 ft/s (from previous investigations).

A. Assuming that layer 2 does not exist, we would interpret the time-distance plot as a two-layer subsurface (eq. 2):

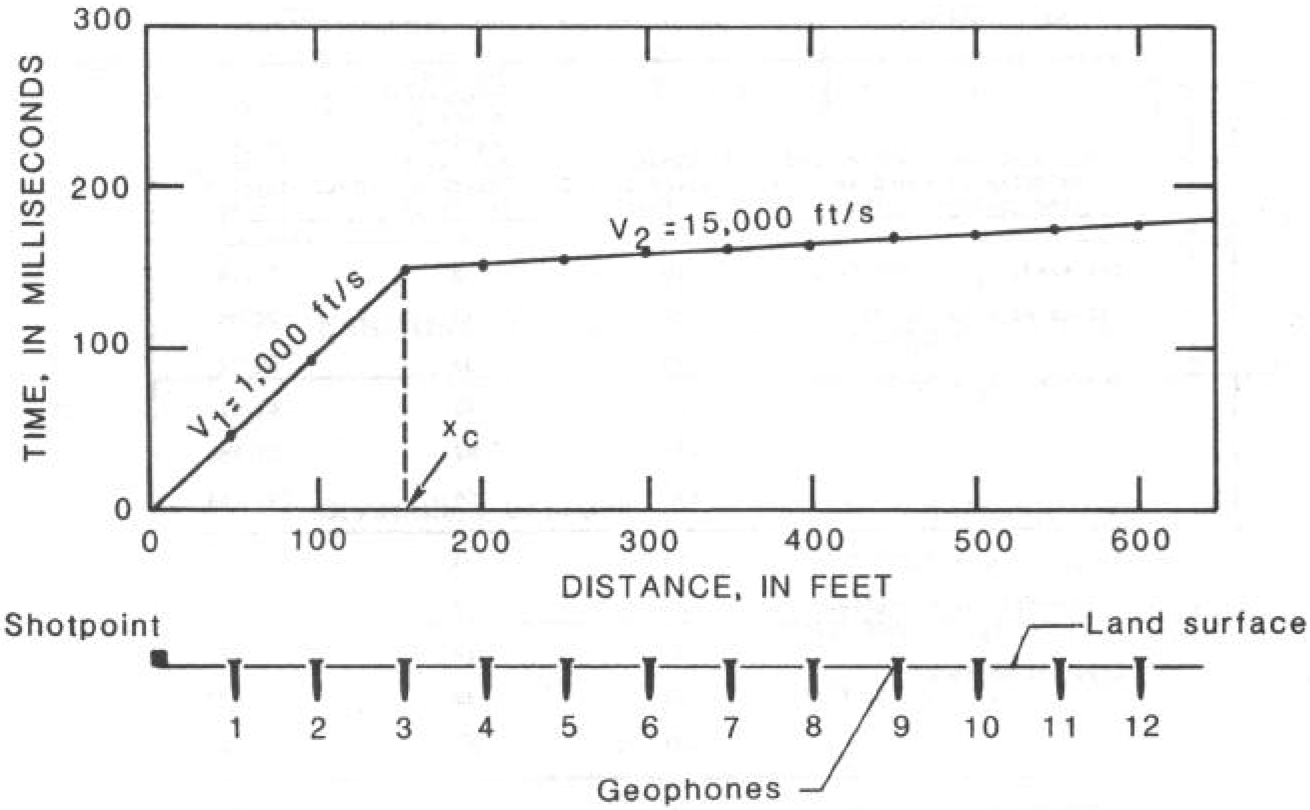



Figure 10.—Time-distance plot showing two layers in an area known to have three layers.

$$\max z_1=\frac{x_c}{2}\sqrt{\frac{V_2-V_1}{V_2+V_1}}=\frac{111}{2}\sqrt{\frac{15{,}000-1{,}500}{15{,}000+1{,}500}}=50 \text{ ft.}$$

The depth to rock using the two-layer interpretation (that is, assuming that there is no saturated material in the geologic section) is, therefore, 50 ft.

B. If the presence of a hidden layer of saturated material is suspected from wells or test holes in the area, the following calculations can be carried out. The minimum depth to layer 2 (the water table) and the maximum possible thickness of undetectable saturated material can be calculated when $x_{c1}=x_{c2}$. (See figs. 9, 11.) In order to calculate these values we assume that a three-layer subsurface exists and proceed with a normal three-layer interpretation using either the time-intercept formulas (eqs. 3–5) or the crossover-distance formulas (eqs. 6–8). A method described by Mooney (1981) using crossover-distance formulas is used in the following calculations.

1. For the depth to layer 2 (the water table),

$$\min z_1=\frac{x_{c1}}{2}\sqrt{\frac{V_2-V_1}{V_2+V_1}}=\frac{111}{2}\sqrt{\frac{5{,}000-1{,}500}{5{,}000+1{,}500}}=41 \text{ ft.}$$

That is, the minimum depth to the water table in the three-layer subsurface is 41 ft.

2. For the depth to layer 3 (the bedrock surface),

$$\max z_3=P(z_1)+\frac{x_{c2}}{2}\sqrt{\frac{V_3-V_2}{V_3+V_2}},$$

where P is defined as

$$P=1-\left(\frac{\frac{V_2}{V_1}\sqrt{\left(\frac{V_3}{V_1}\right)^2-1}-\frac{V_3}{V_1}\sqrt{\frac{V_2^2}{V_1}-1}}{\sqrt{\left(\frac{V_3}{V_1}\right)^2-\left(\frac{V_2}{V_1}\right)^2}}\right)$$

$$P=.86$$

$$\max z_3=.86(40.7)+\frac{111}{2}\sqrt{\frac{15{,}000-5{,}000}{15{,}000+5{,}000}}=74 \text{ ft.}$$

The maximum depth to the bedrock surface is 74 ft.

3. For the maximum undetected thickness of layer 2 (that is, the saturated thickness of the unconsolidated material),

$$\max z_2=z_3-z_1=74-41=33 \text{ ft.}$$

The maximum thickness of an undetected layer 2 in a three-layer subsurface is 33 ft.

In summary, a maximum of 33 ft of saturated sand and gravel under a minimum of 41 ft of unsaturated sand and

Table 1.—*Maximum thickness of an undetectable layer in various hydrogeologic settings*

| Hydrogeologic setting and velocity of sound in the geologic units | Thickness of layer 1 (in feet) | Maximum thickness of undetected aquifer material in layer 2 (in feet) | Range in depth to layer 3 (in feet) |
|---|---|---|---|
| Dry sand, $V_1$ = 1,500 ft/s | 10 | 8 | 12-18 |
| Saturated sand aquifer, $V_2$ = 5,000 ft/s | 20 | 16 | 24-36 |
| | 40 | 33 | 50-74 |
| Bedrock, $V_3$ = 15,000 ft/s | 50 | 41 | 61-91 |
| | 100 | 82 | 123-182 |
| | 200 | 164 | 243-364 |
| Till, $V_1$ = 7,000 ft/s | 10 | 3 | 11-13 |
| Sedimentary rock aquifer, $V_2$ = 13,000 ft/s | 20 | 7 | 22-26 |
| | 50 | 17 | 55-67 |
| Crystalline rock, $V_3$ = 15,000 ft/s | 100 | 33 | 110-133 |
| | 200 | 67 | 219-267 |
| Saturated sand and gravel, $V_1$ = 5,000 ft/s | 10 | 6 | 12-16 |
| | 20 | 12 | 24-32 |
| Limestone aquifer, $V_2$ = 10,000 ft/s | 50 | 29 | 61-79 |
| Crystalline rock, $V_3$ = 15,000 ft/s | 100 | 58 | 122-158 |
| | 200 | 115 | 245-315 |

gravel could not be detected with the seismic-refraction method in the above example. The depth to rock is between 50 and 74 ft depending on the thickness of the saturated zone. The saturated thickness of undetected sand and gravel is between 0 and 33 ft. The minimum depth to the water table is 41 ft.

### Insufficient seismic-velocity contrasts between hydrogeologic units

In many studies, significant hydrogeologic materials may not have detectable seismic-velocity contrasts. Many rock surfaces are not fresh and exhibit different degrees of weathering. As the rock surface weathers, the seismic velocity decreases and is no longer indicative of the unweathered bedrock. In these cases, seismic-refraction techniques may not differentiate the weathered surface from the overlying low-velocity material.

Some significant hydrologic boundaries may have no field-measurable velocity contrast across them and, consequently, cannot be differentiated with these techniques. For example, saturated unconsolidated gravel deposits may have approximately the same seismic velocity as saturated unconsolidated silt and clay deposits (Burwell, 1940).

### Low-seismic-velocity units underlying high-seismic-velocity units

In some hydrogeologic settings, the velocity of sound in each of the Earth's layers does not increase with depth, and low-seismic-velocity units underlie high-seismic-velocity units. Examples of this are (1) an unconsolidated sand and gravel aquifer underlying compact glacial tills, (2) semiconsolidated rubble zones beneath dense basalt flows, and (3) dense limestone overlying a poorly cemented sandstone.

In all of these cases, the low-velocity unit will not be detected by seismic-refraction techniques and the calculated depth to the deep refractor will be in error. The reason for this problem is found in Snell's Law, which says that a sound wave will be refracted toward the low-velocity medium. When a low-velocity layer underlies a high-velocity layer, the seismic raypaths are refracted downward or away from the land surface. The sound wave, therefore, would not be detected at the surface until it

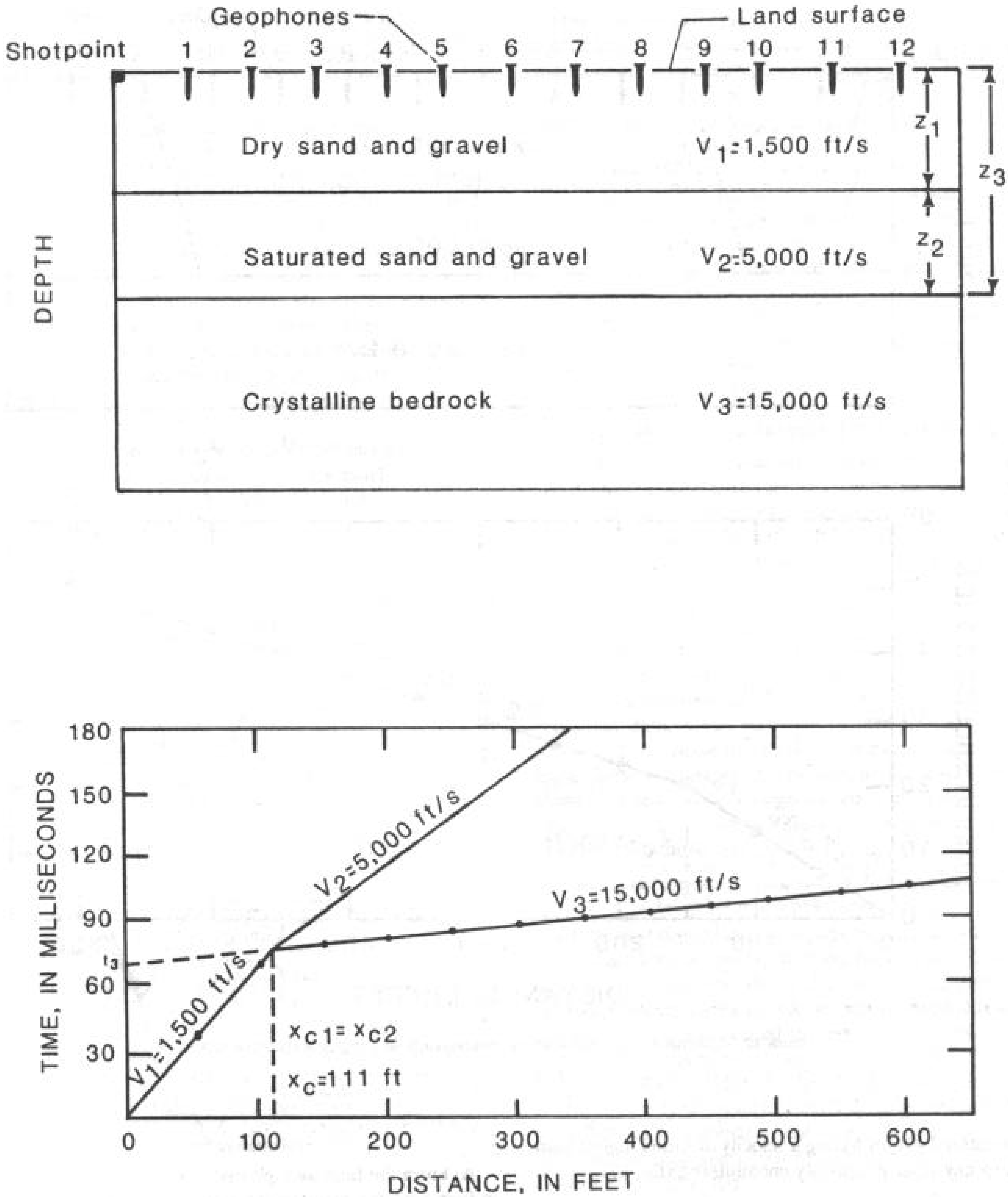



Figure 11.—Seismic section with hidden layer (layer 2) and resulting time-distance plot.

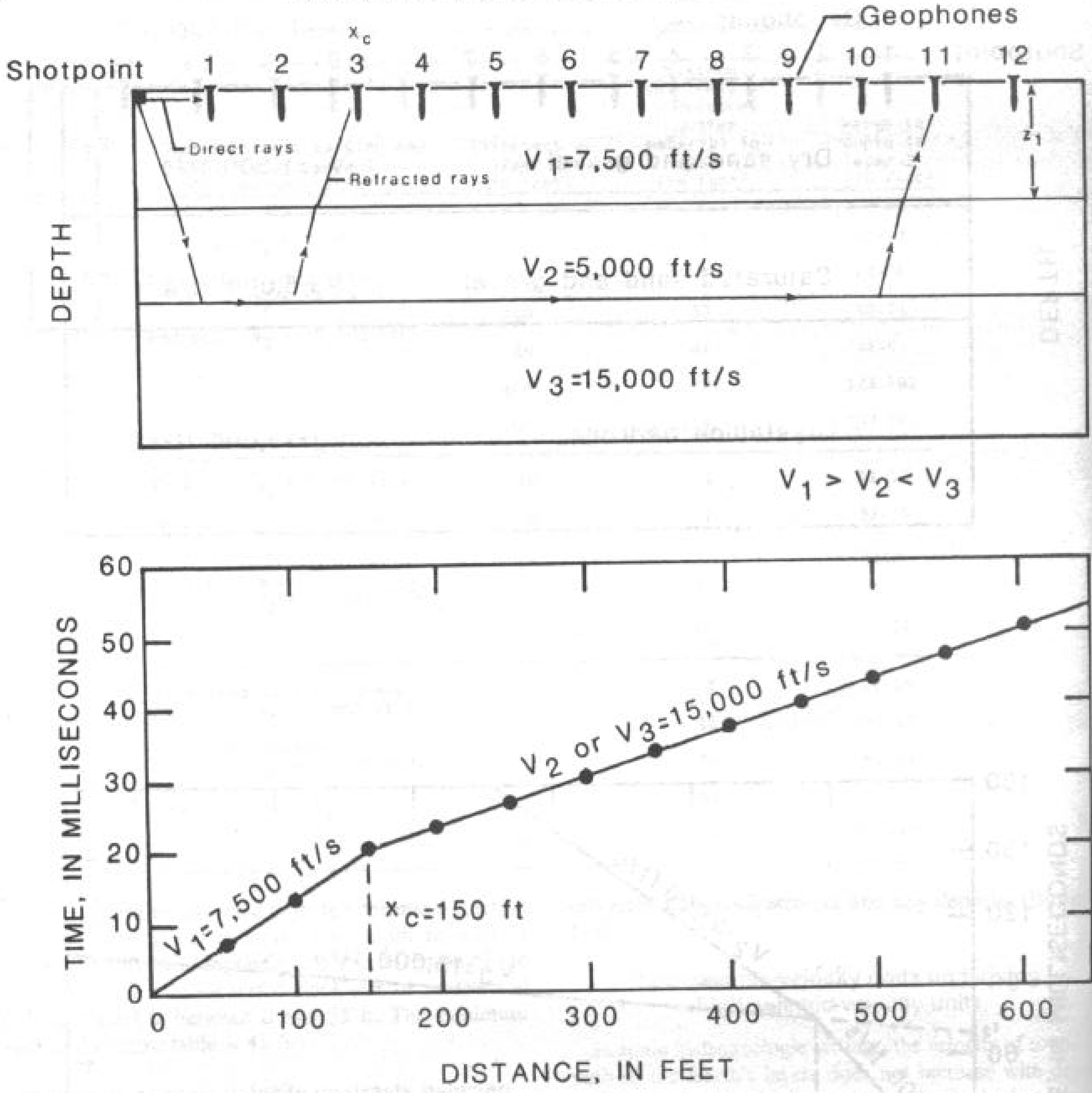



Figure 12.—Seismic section with velocity reversal and resulting time-distance plot.

encountered a layer having a velocity of sound higher than that of any layer previously encountered (fig. 12).

If a low-seismic-velocity unit is known to exist beneath a high-seismic-velocity unit from drill-hole or geologic data, and if its depth and seismic velocity are approximately known, the depth to a deeper refractor can be estimated (Mooney, 1981; Morgan, 1967). Without this information, the depth calculated from the seismic-refraction data will be greater than the actual depth.

**Example problem**

A. From the field data plotted in the time-distance plot in figure 12, the existence of layer 2 would not be known and an erroneous depth to layer 3 would be calculated if one used the two-layer parallel-boundary formulas (eqs. 3–5):

$V_1$ = 7,500 ft/s (from time-distance plot),
$V_2$ = 15,000 ft/s (from time-distance plot),
$z_2'$ = erroneous depth to layer 3, and
$x_c$ = 150 ft (from time-distance plot).

$$z_2'=\frac{X_c}{2}\sqrt{\frac{V_2-V_1}{V_2+V_1}}=\frac{150}{2}\sqrt{\frac{15{,}000-7{,}500}{15{,}000+7{,}500}}=43 \text{ ft.}$$

The depth to rock using the two-layer interpretation is, therefore, 43 ft. If the thickness and the velocity of sound in layer 2 are known or can be estimated from drill-hole or other data, a more accurate depth can be calculated.

B. From a nearby drill hole and a previous seismic-refraction investigation in a nearby area, it is determined that layer 1 is glacial till approximately 20 ft thick and having a seismic velocity of approximately 7,500 ft/s. It is underlain by saturated sand and gravel having a velocity of about 5,000 ft/s. Now, a more realistic value for the depth to layer 3 ($z_2$) can be calculated using the following method described by Mooney (1981, p. 9–17):

$V_1$ =7,500 ft/s,
$V_2$ =5,000 ft/s (from previous investigation),
$V_3$ =15,000 ft/s (from time-distance plot),
$z_1$ =20 ft (from nearby drill hole), and
$z_2$ =true depth to layer 3.

$$z_2=(Q+1)\frac{X_c}{2}\sqrt{\frac{V_3-V_1}{V_3+V_1}}-z_1Q, \tag{21}$$

where Q is defined as

$$Q=\sqrt{\frac{\left(\frac{V_3}{V_1}\right)^2-1}{\left(\frac{V_3}{V_2}\right)^2-1}}-1. \tag{22}$$

Now substituting,

$$Q=\sqrt{\frac{\left(\frac{15{,}000}{7{,}500}\right)^2-1}{\left(\frac{15{,}000}{5{,}000}\right)^2-1}}-1=-0.39$$

and

$$z_2=(-0.39+1)\frac{150}{2}\sqrt{\frac{15{,}000-7{,}500}{15{,}000+7{,}500}}-20(-0.39)$$

$$=34 \text{ ft.}$$

In summary, without any external data, a two-layer subsurface with rock at 43 ft was interpreted from the seismic data. Using data from a nearby test hole and the results from a previous seismic-refraction study, a three-layer subsurface with rock at 34 ft was interpreted from the same field data.

One special example of a hidden-layer problem is encountered when seismic-refraction surveys are conducted in areas where the surface of the ground is frozen. The velocity of sound in frozen ground is about 12,000 ft/s (Bush and Schwarz, 1965), and the frozen zone can act as a high-velocity surficial layer. Any layers under the frozen ground cannot be detected unless the velocity of sound in them is greater than 12,000 ft/s. The hydrologist must be careful in interpreting data gathered under these field conditions. Figure 13 shows the time-distance plot that would be obtained in a stratified-drift valley with frozen ground at the surface.

One way to eliminate this problem is to bury both the sound source and the geophones beneath the frozen layer. This usually involves considerable effort and is not economical in most hydrologic programs.

## Other limitations of seismic-refraction techniques

The following limitations are mentioned not to discourage the use of seismic-refraction techniques, but rather to make hydrologists aware of potential pitfalls. These situations, recognized early in the study, can be accounted for in the planning, data-acquisition, and interpretation phases of the study.

### Ambient noise

Ambient noise, that is, the noise produced by vehicular traffic, construction equipment, railroads, wind, and so forth, has a detrimental effect on the quality of seismic-refraction data. Some solutions to this problem are as follows: (1) decrease the amplifier gains and increase the input signal by using more explosives or repeated hammer blows, (2) reschedule operations for a quiet part of the day, and (3) use selective filters on the seismograph to eliminate unwanted frequencies.

### Horizontal variations in the velocity of sound and the thickness of the weathered zone

Horizontal discontinuities in the low-velocity zone near the surface have a significant effect on seismic-refraction studies. This zone usually is the unsaturated zone and typically has velocities of 400 to 1,600 ft/s. Short geographic spreads are needed to determine the velocity of sound and the thickness of this layer. A variation of 1 ft in the thickness of a weathered layer consisting of material having a velocity of sound of 1,000 ft/s causes the refracted sound ray to be delayed or sped up by 1 ms. This same time interval represents 10 ft of material having a velocity of sound of 10,000 ft/s.

### Accuracy of seismic-refraction measurements

The accuracy with which the depth to a refractor can be determined by seismic-refraction methods depends on many factors. Some of these factors are

- Type and accuracy of seismic equipment,
- Number and type of corrections made to field data,
- Quality of field procedures,

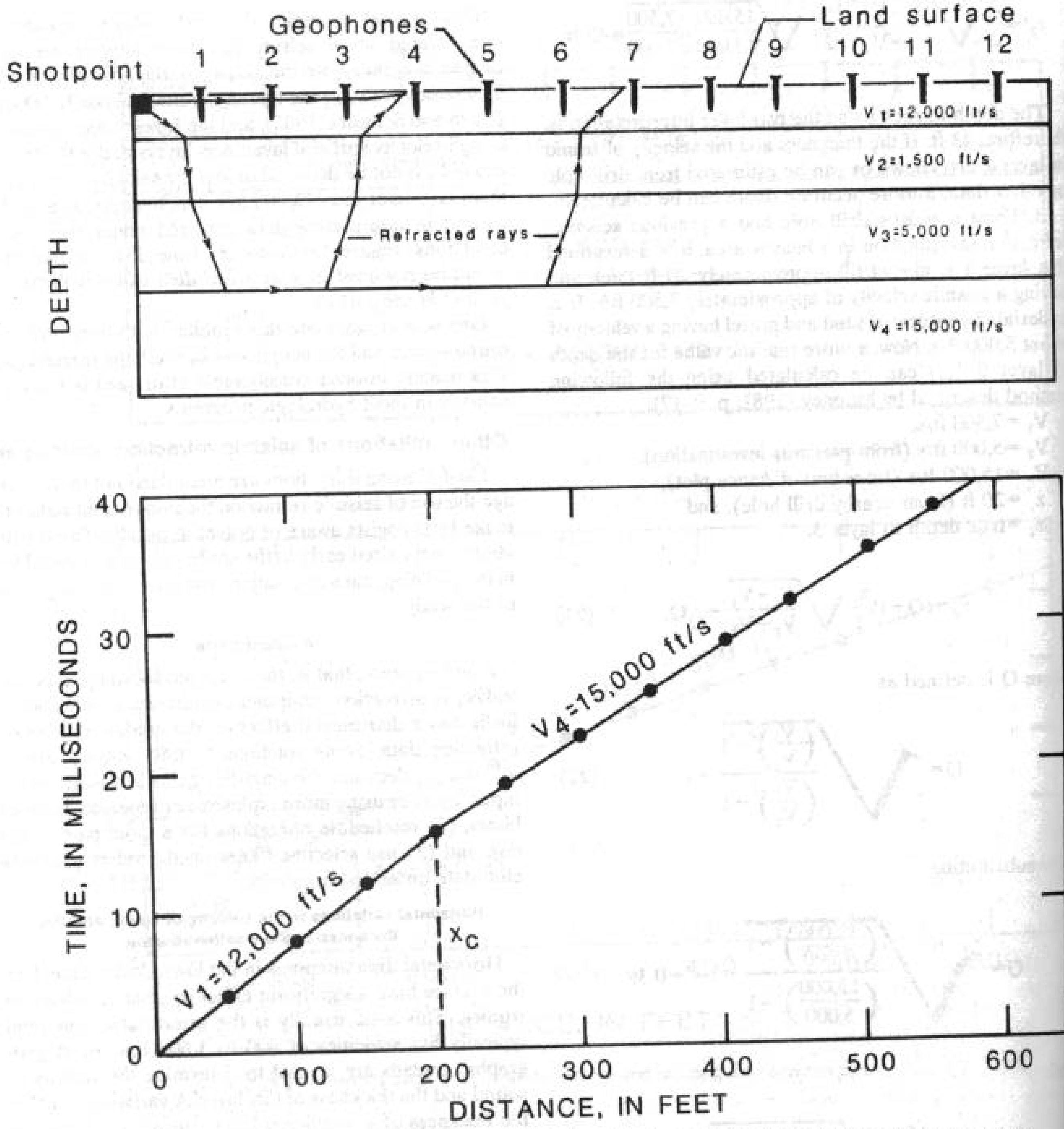



Figure 13.—Interpreted seismic section and time-distance plot for a four-layer model having frozen ground at the surface.

- Type of interpretation method used,
- Variation of the Earth from simplifying assumptions used in the interpretation procedure, and
- Ability and experience of the interpreter.

Published references (Griffiths and King, 1965; Eaton and Watkins, 1967; Wallace, 1970; Zohdy and others, 1974) and the author's unpublished data indicate that the depth to a refractor can reasonably be determined to within 10 percent of the true depth. Larger errors usually are due to improper interpretation of difficult field situations.

## Annotated references

Alsop, S.A., 1982, Engineering geophysics: Association of Engineering Geologists Bulletin, v. 19, no. 2, p. 181–186.
[Brief review of the engineering application of seismic-refraction techniques.]

Burke, K.B.S., 1967, A review of some problems of seismic prospecting for ground water in surficial deposits, *in* Morey, L.W., ed., Mining and ground-water geophysics: Canada Geological Survey Economic Geology Report 26, p. 569–579.

[Review and evaluation of three problems encountered while using seismic prospecting for ground water: (1) the difficulty in generating simple seismic waves in unconsolidated materials, (2) the lack of a definitive relationship between seismic velocity and a particular deposit, and (3) the much greater significance of error in small-scale seismic investigations.]

Burwell, E.B., 1940, Determination of ground-water levels by the seismic method: Transactions of the American Geophysical Union, v. 21, p. 439–440.

[Changes in velocity of sound in saturated material shown to be independent of alluvial material present.]

Bush, B.O., and Schwarz, S.D., 1965, Seismic refraction and electrical resistivity measurements over frozen ground, *in* Brown, R.J.E., ed., Proceedings of the Canadian Regional Permafrost Conference, 1 and 2 December 1964: Ottawa, National Research Council of Canada, Associate Committee on Soil and Snow Mechanics, Technical Memorandum 86, p. 32–40.

[Seismic-refraction techniques were evaluated for predicting the depth to rock with frozen ground at the surface.]

Dobrin, M.B., 1976, Introduction to geophysical prospecting (3d ed.): New York, McGraw-Hill, 630 p.

[A basic reference text that covers theoretical and practical aspects of the major surface geophysical methods, with emphasis on deep exploration.]

Domzalski, W., 1956, Some problems of shallow refraction investigations: Geophysical Prospecting, v. 4, no. 2, p. 140–166.

[Detailed discussion and examples of problems encountered in shallow seismic-refraction studies.]

Eaton, G.P., and Watkins, J.S., 1967, The use of seismic-refraction and gravity methods in hydrologic investigations, *in* Morey, L.W., ed., Mining and ground-water geophysics: Geological Survey of Canada Economic Geology Report 26, p. 554–568.

[A review of seismic-refraction methods with case histories pertaining to ground-water studies. Includes a comprehensive bibliography of work prior to 1967.]

Grant, F.S., and West, G.F., 1965, Interpretation theory in applied geophysics: New York, McGraw-Hill, 583 p.

[Detailed discussions of the basic theory behind surface geophysical techniques.]

Green, R., 1962, The hidden layer problem: Geophysical Prospecting, v. 10, no. 2, p. 166–170.

[Discusses how to determine the range in thickness of a hidden layer given a refraction time-distance curve.]

Griffiths, D.H., and King, R.F., 1965, Applied geophsyics for engineers and geologists (2d ed.): Oxford, England, Pergamon Press, 223 p.

[Clear, short text explaining various geophysical techniques.]

Habberjam, G.M., 1966, A nomogram for investigation of the three-layer refraction problem: Geoexploration, v. 4, no. 4, p. 219–225.

[Describes construction of a single-sheet nomogram for investigating the three-horizontal-layer refraction problem when the deepest layer has the highest seismic velocity.]

Hawkins, L.V., and Maggs, D., 1961, Nomographs for determining maximum error problem: Geophysical Prospecting, v. 9, no. 4, p. 526–532.

[Has nomograms for solution of the thin, intermediate-seismic-velocity-layer problem for critical-distance and time-intercept formulas.]

Johnson, S.H., 1976, Interpretation of split-spread refraction data in terms of plane dipping layers: Geophysics, v. 41, no. 3, p. 418–424.

[Develops a computing procedure to interpret split-spread refraction data assuming that the geologic formation may be approximated by planar, dipping velocity layers.]

Knox, W.A., 1976, Multi-layer near surface refraction computations, *in* Musgrave, A.W., ed., Seismic-refraction prospecting: Tulsa, Okla., Society of Exploration Geophysicists, p. 197–216.

[Discussion of near-surface geologic conditions and the effects of varying shot depths on the interpretation of shallow refraction formulas.]

Meridav, Tsvi, 1960, Nomograms to speed up seismic refraction computations: Geophysics, v. 25, no. 5, p. 1035–1053.

[Includes nomograms for the solution of two-layer crossover-distance formulas, critical-angle and offset-distance formulas, and two-layer intercept-time formulas, and calculation of true velocity from apparent updip and downdip velocity.]

———1968, A multi-layer seismic refraction nomogram: Geophysics, v. 33, no. 3, p. 524–526.

[A nomogram for solving multilayer refraction problems using crossover distances and layer velocities.]

Mooney, H.M., 1981, Handbook of engineering geophysics, v. 1: Seismic: Minneapolis, Bison Instruments, Inc., 220 p.

[A thorough handbook covering the theory and practical aspects of shallow refraction and reflection techniques.]

Morgan, N.A., 1967, The use of continuous seismic profiles to solve hidden-layer problems: Geophysical Prospecting, v. 15, no. 1, p. 35–43.

[Shows that use of a continuous seismic-reflection profiler may overcome the hidden-layer problem in refraction surveys. This technique allows calculation of the velocity and thickness of the hidden layer.]

Mota, L., 1954, Determination of dips and depths of geological layers by the seismic refraction method: Geophysics, v. 19, p. 242–254.

[Equations developed for computing depths and dips of inclined interfaces from seismic-refraction data.]

Musgrave, A.W., ed., 1967, Seismic-refraction prospecting: Tulsa, Okla., Society of Exploration Geophysicists, 604 p.

[A volume of papers on the refraction method with an extensive bibliography of seismic-refraction techniques.]

Parasnis, D.S., 1979, Principles of applied geophysics (3d ed.): London, Chapman and Hall (distributed in the United States by John Wiley and Sons, New York), 275 p.

[A very brief description of the major surface geophysical methods.]

Redpath, B.B., 1973, Seismic refraction exploration for engineering site investigations: National Technical Information Service AD-768710, 51 p.

[A summary of the theory and practice of using the refraction seismograph for shallow, subsurface investigations.]

Sander, J.E., 1978, The blind zone in seismic ground-water exploration: Ground Water, v. 16, no. 6, p. 394–397.

[Refraction techniques were used to map areas of thick, compacted till beneath an unconfined glacial aquifer in northern Minnesota. Where this unit is thin, a blind-zone layer exists. Treatment of this case is discussed.]

Scott, J.H., Tibbets, B.L., and Burdick, R.G., 1972, Computer analysis of seismic-refraction data: U.S. Bureau of Mines Report of Investigation 7595, 95 p.

[Presents a computer program that uses seismic-refraction data to generate a two-dimensional model representing a layered geologic section.]

Sheriff, R.E., 1973, Encyclopedic dictionary of exploration geophysics: Tulsa, Okla., Society of Exploration Geophysicists, 266 p.

[Exhaustive glossary of exploration geophysics and its related disciplines.]

Slotnick M.M., 1959, Lessons in seismic computing: Tulsa, Okla., Society of Exploration Geophysicists, 268 p.

[An in-depth theoretical discussion of the geometrical aspects of refraction prospecting.]

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

[Wave-front diagrams illustrate the cause and results of, and solutions for, the blind-zone problem in shallow refraction studies.]

Telford, W.M., Geldart, L.P., Sherriff, R.E., and Keys, D.A., 1976, Applied geophysics: New York, Cambridge University Press, 806 p.
[A basic reference text that covers the theoretical and practical aspects of the major geophysical methods, with emphasis on deep exploration.]

Wallace, D.E., 1970, Some limitations of seismic refraction methods in geohydrological surveys of deep alluvial basins: Ground Water, v. 8, no. 6, p. 8–13.
[A case history illustrating some limitations of seismic-refraction techniques in deep alluvial basins.]

Zirbel, N.N., 1954, Comparison of break-point and time-intercept methods in refraction calculation: Geophysics, v. 19, no. 4, p. 716–721.
[Shows that break-point (crossover-distance) formulas in some instances are more accurate than intercept-time formulas for determining the depth to a refractor.]

Zohdy, A.A.R., Eaton, G.P., and Mabey, D.R., 1974, Application of surface geophysics to ground-water investigations: U.S. Geological Survey Techniques of Water-Resources Investigations, Book 2, Chapter D1, 116 p.
[An overview of surface geophysical techniques and their applications to hydrologic studies.]

# Applications of Seismic-Refraction Techniques to Hydrology

Seismic-refraction techniques have been used for a variety of studies conducted in many different hydrogeologic settings. This section describes the results of some recent studies involving typical hydrogeologic problems that demonstrate where the techniques (1) can be used successfully, (2) may work but with some difficulty either in the field procedures or in the interpretation process, and (3) cannot be used. In addition to the discussion of individual case histories, references to other studies that have applied seismic-refraction techniques to similar hydrogeologic problems are provided. This section is intended as an initial guide for the hydrologist considering the use of geophysical techniques. Specific applications of the techniques should be tested in the field, in areas where adequate geologic and hydrologic controls are available.

## Hydrogeologic settings in which seismic-refraction techniques can be used successfully

Hydrogeologic settings in which each successively deeper layer has a higher seismic velocity, no thin layers are present, and a significant seismic-velocity change occurs at each hydrogeologic interface are ideally suited for the application of seismic-refraction techniques. The five case histories presented below illustrate successful application of seismic-refraction techniques in hydrogeologic settings that satisfy these conditions.

### Unconsolidated unsaturated glacial or alluvial material overlying glacial or alluvial aquifers

Determining the depth to a shallow water table within this type of setting is a common hydrologic goal. Because the velocity of sound in unconsolidated, unsaturated sands and gravels ranges from 400 to 1,600 ft/s, and because the velocity of sound in unconsolidated, saturated sands and gravels ranges from 4,000 to 6,000 ft/s, seismic-refraction methods will generally be successful in determining the depth to water. The seismic-velocity contrast between the unsaturated and saturated material, however, will decrease as the grain size of the aquifer decreases and the depth to water increases (White and Sengbush, 1953).

To determine the depth to a shallow water table, short geophone spreads must be used so that the velocity of sound in the unsaturated zone is accurately determined. Lateral changes in the seismic velocity of this layer are common and must be measured in the field and accounted for in the interpretation process. However, because the seismic velocity of the unsaturated zone exhibits a gradual increase with depth (Emerson, 1968), it can only be approximated as a constant velocity layer.

Galfi and Palos (1970) demonstrated that in sandy areas, seismic-refraction techniques can accurately determine the depth to water. Their study used a single-channel seismograph, a sledge hammer for the sound source, and a 3.3-ft geophone spacing. The results of one seismic profile and the well control data are shown in figure 14. The seismically determined depth to the water table of 13.3 ft agreed with the well data, 13.1 ft. The use of the sledge hammer as a sound source provided sufficient first-arrival energy to a distance of only 75 ft from the source and, consequently, limited the penetration depth to about 25 ft. To determine greater depths to water, other, more powerful sound sources would be needed. In this study, the unsaturated zone was interpreted using a continuous-velocity-distribution formula (Dobrin, 1976).

Many seismic-refraction studies have been conducted in Connecticut as part of water-resources investigations. A comparison of the seismically determined depths to water and the subsequent drill-hole data for four studies is presented in table 2. In these studies, the velocity of the unsaturated zone was considered constant and the depth to water was calculated by a delay-time and ray-tracing modeling process described by Scott and others (1972).

Other studies that have used seismic-refraction techniques for determining the depth to water in unconsolidated aquifers include those of Burwell (1940), Emerson (1968), Sjogren and Wagner (1969), and Followill (1971).

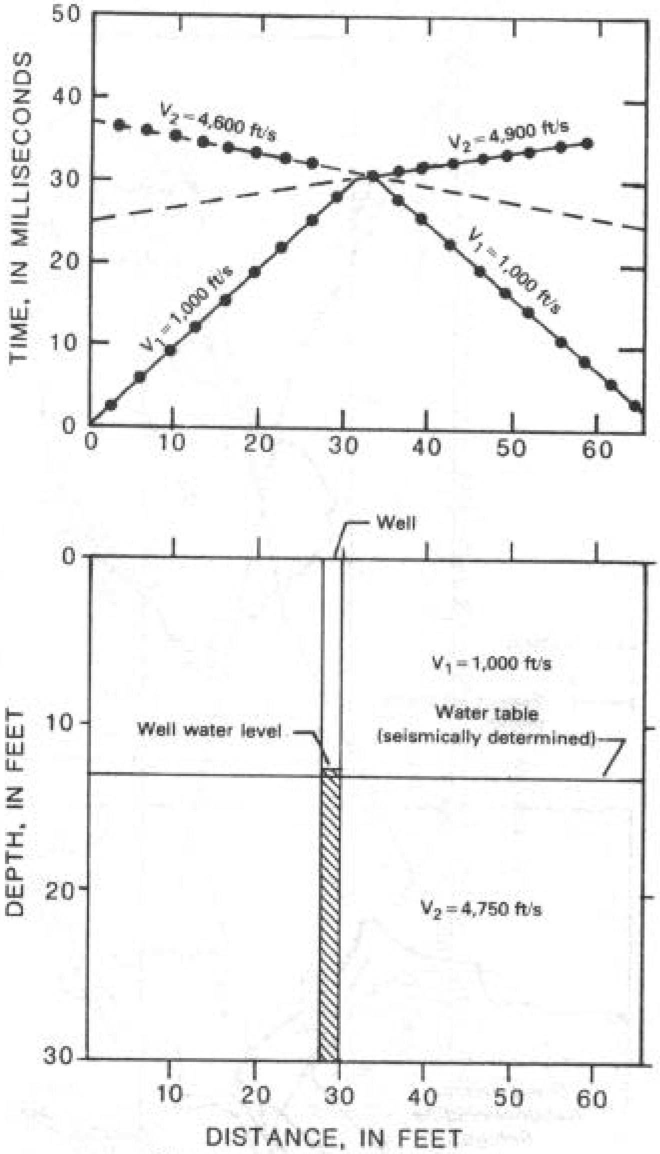



Figure 14.—Time-distance plot and interpreted seismic section from a ground-water study in Vertessomto, Hungary (modified from Galfi and Palos, 1970, p. 45).

### Unconsolidated glacial or alluvial material overlying consolidated bedrock

Determination of the saturated thickness of the aquifer material and (or) the shape of the bedrock surface in this setting is a common hydrologic problem. The velocity of sound in both the unsaturated and saturated material is the same as in the previous problem (400–1,600 ft/s and 4,000–6,000 ft/s, respectively). The velocity of sound in the consolidated bedrock should be between 10,000 and 20,000 ft/s. The velocity constraints of the refraction technique are met, as the velocity of sound in each layer increases with depth. Seismic-refraction techniques can define the top of the water table and the top of the bedrock, provided the saturated zone does not get too thin (see section on thin, intermediate-seismic-velocity layer problems).

To map both a shallow refractor, such as the water table, and a deep refractor, such as the bedrock surface, careful consideration must be given to the choice of shotpoints, geophone spacing, and interpretation method used. Multiple shots, variable geophone spacings, and (or) test-hole data will be needed, depending on the geometry of the problem.

A reconnaissance seismic-refraction survey was conducted by the U.S. Geological Survey near the Great Swamp National Wildlife Refuge, Morristown, N.J. (fig. 15). To determine the depth to bedrock, several profiles with two or three geophone spreads were run along roads and paths in the area. A typical time-distance plot and the interpreted seismic section are shown in figure 16.

Because the primary purpose of this study was of a reconnaissance nature, and because the water table was known to be close to the surface, only one shotpoint on each end of each geophone spread was used. The shots were placed in the saturated layer so that small explosive charges could be used and the depth to water measured directly. The measured depths to water were used in the interpretation procedure to estimate, or "back out," the velocity of the thin unsaturated zone. The geophone spreads were overlapped in order to obtain a continuous bedrock profile. The depth to water in the study area averaged about 5 ft, and the depth to rock ranged from 75 to 200 ft.

Other studies in similar hydrogeologic settings that have successfully used this technique include those of Gill

Table 2.—*Comparison of the depth to water determined by seismic-refraction methods and by drilling*

| Location in Connecticut | Depth to water determined by seismic-refraction methods (feet) | Depth to water determined by drilling (feet) |
|---|---|---|
| Plainville | 25 | 26 |
| Newtown | 12 | 9 |
| | 5 | 3 |
| | 10 | 12 |
| | 12 | 7 |
| | 25 | 27 |
| | 35 | 45 |
| | 10 | 5 |
| | 9 | 6 |
| Farmington | 10 | 11 |
| | 55 | 56 |
| | 5 | 3 |
| Stonington | 16 | 12 |
| | 6 | 5 |
| | 8 | 7 |

and others (1965), Lennox and Carlson (1967), Duguid (1968), Joiner and others (1968), Peterson and others (1968), Mercer and Lappala (1970), and Wachs and others (1979).

### Thick, unconsolidated alluvial or sedimentary materials overlying consolidated sediments and (or) basement rock in large structural basins

This problem is similar to the preceding one, except that the geologic section can be more complex and the unsaturated and saturated layers are much thicker. As long as the successively deeper layers have a higher seismic velocity and are not thin, seismic-refraction techniques will work. As the depth to the water-table increases, however, the seismic velocity of the unsaturated layer increases, and this may prevent identification of the saturated zone as a separate refracting layer.

The U.S. Geological Survey conducted a seismic-refraction study near Tucson, Ariz. (H.D. Ackermann, U.S. Geological Survey, written commun., 1980), to determine the saturated thickness of the aquifer near the outlet of ground-water flow from the Aura-Altar basin (fig. 17). Figure 18 shows the results of the interpreted seismic data. The small seismic-velocity contrast between the unsaturated and saturated alluvium made detection of the water table very difficult. It was finally delineated with the use of available well data in conjunction with a comprehensive seismic-refraction modeling program (Ackermann and others, 1983). The 4-mi profile shown in figure 18 was obtained using two spreads of 24 geophones with the geophones spaced 400 ft apart and one spread of 24 geophones with the geophones spaced 200 ft apart. Five to seven shots, each consisting of 15 to 80 lb of explosives buried 30 ft below the surface, were used as a sound source.

Other hydrogeologic studies of deep alluvial basins that have used seismic-refraction techniques are described by Dudley and McGinnis (1962), Arnow and Mattick (1968), Mower (1968), Libby and others (1970), Wallace (1970), Marshall (1971), Robinson and Costain (1971), Mattick and others (1973), Crosby (1976), and Pankratz and others (1978).

### Unconsolidated alluvial material overlying sedimentary rock, which in turn overlies volcanic or crystalline bedrock

In this type of setting, mapping the saturated thickness of the unconsolidated sand aquifer and the thickness of the sedimentary rock aquifer is a common exploration goal. Such goals can be achieved using seismic-refraction techniques when the velocity of sound in the sedimentary rock aquifer is greater than that in the saturated alluvium and less than that in the underlying volcanic or crystalline rock. Again, the intermediate layer (in this case the sedimentary rock) must not be too thin (see section on limitations of seismic-refraction techniques). Figure 19 shows the location of a study conducted in the Guanajibo area, Puerto Rico (Colon-Dieppa and Quinones-Marquez, 1985). Figure 20 shows a typical time-distance plot and the interpreted seismic section from one seismic

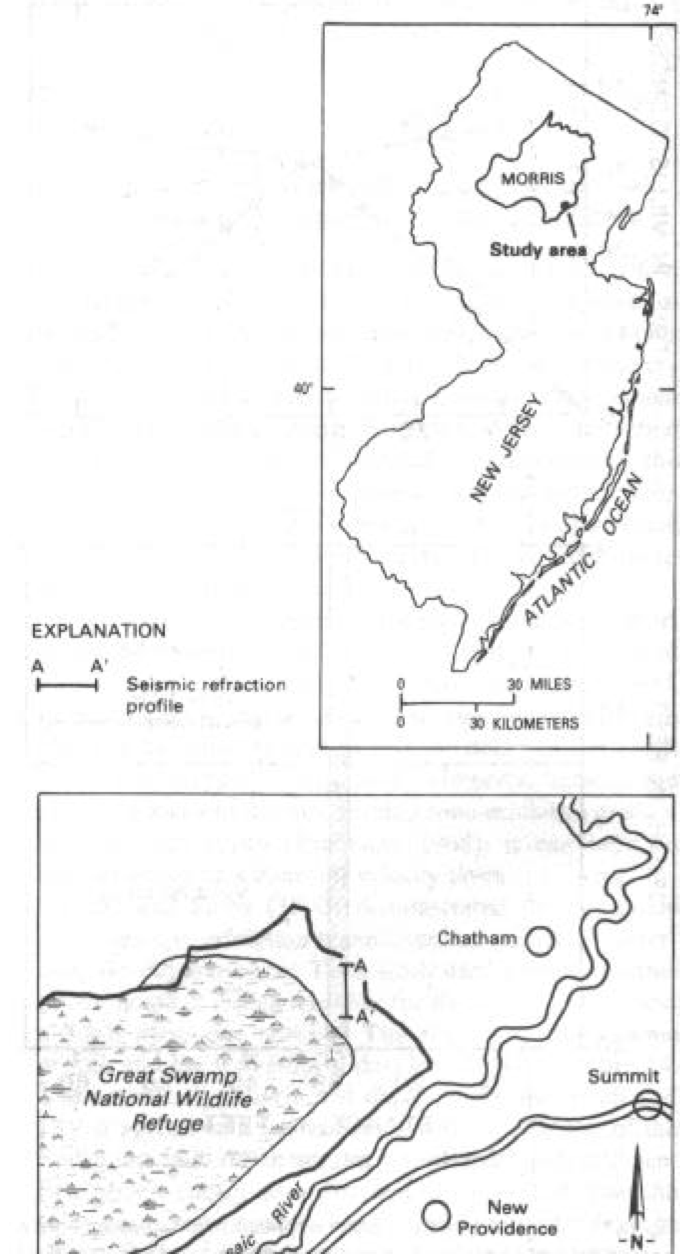



Figure 15.—Generalized location map of Great Swamp National Wildlife Refuge, N.J., and location of seismic-refraction profile A-A'.

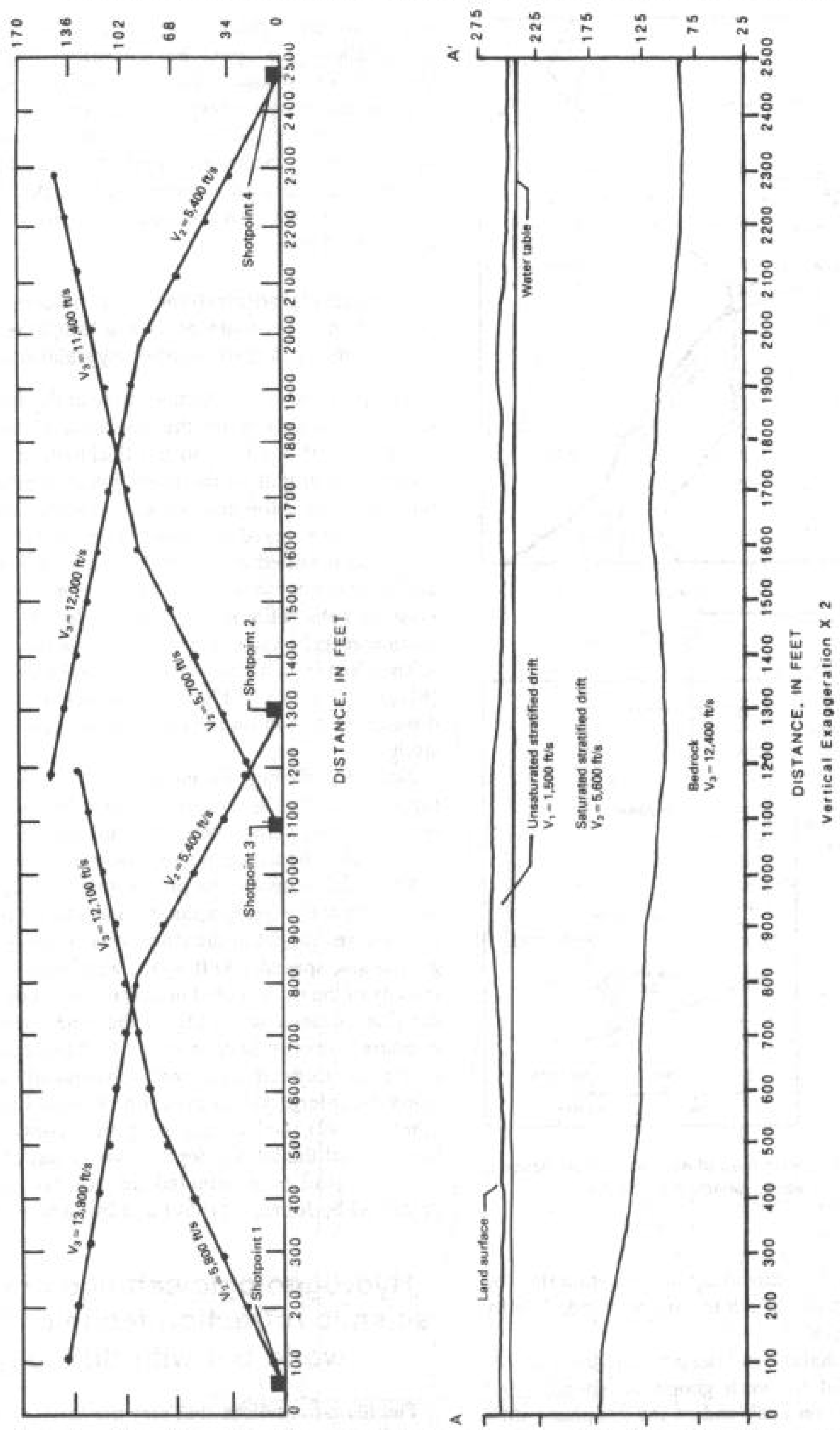



Figure 16.—Time-distance plot and interpreted seismic section near Great Swamp National Wildlife Refuge, Morristown, N.J.

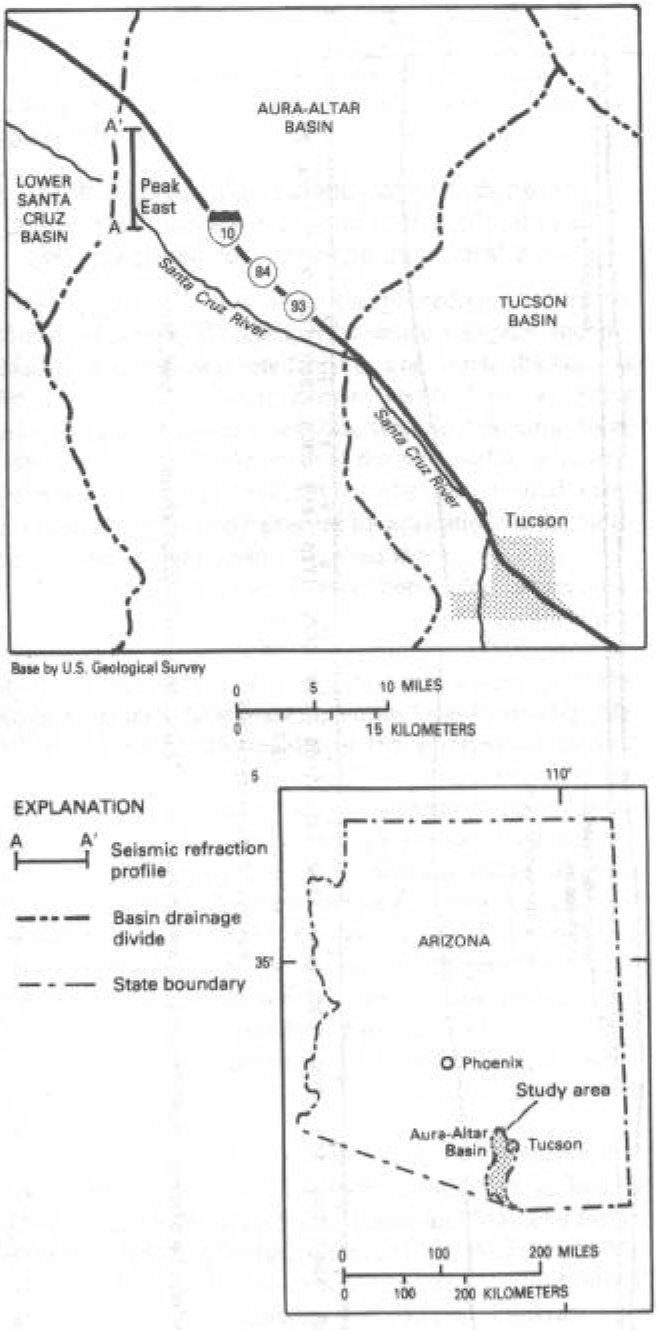



Figure 17.—Generalized location map of Aura-Altar basin, Arizona, and location of seismic-refraction profile A-A'.

profile. In this study, the alluvial aquifer was underlain by a thick limestone aquifer which in turn was underlain by volcanic basement rock.

To map both the shallow and deep refractors, multiple shotpoints were used for each geophone spread. One shotpoint was placed on each end of the geophone line, while others were offset 1,000 ft from each end. Each geophone spread consisted of 12 geophones spaced 100 ft apart. The seismic velocity of the unsaturated layer was not measured in the field because the water-table depth was shallow and could be measured directly in each shothole. The seismic velocity of this layer was eventually determined in the interpretation program described by Scott and others (1972) by adjusting the seismic velocity of layer 1 until the known depth to water was matched.

Other studies in similar hydrologic settings are described by Visarion and others (1976) and by Torres-Gonzalez, 1984.

### Unconsolidated stratified-drift material overlying significant deposits of dense lodgement glacial till, which in turn overlie crystalline bedrock

The purpose of a refraction study in this hydrogeologic setting is to determine the thickness of the saturated stratified-drift aquifer and the thickness of the till. The velocity constraints of the refraction technique are again satisfied. The estimated seismic velocities are 1,000 ft/s for the unsaturated stratified drift, 5,000 ft/s for the saturated stratified drift, 7,500 ft/s for the lodgement till, and 15,000 ft/s for the bedrock. The thickness of the till must be substantial in order to be detected by seismic-refraction techniques. Figure 21 shows the location of a seismic line from a study conducted in Farmington, Conn. (Mazzaferro, 1980). Figure 22 shows one of the time-distance plots and interpreted seismic sections from this study.

Note that the significant thickness of till at this site (approximately 250 ft) is represented by a short segment on the time-distance plot. The till layer is an almost undetectable intermediate-seismic-velocity layer.

The field setup for the profile shown in figure 22 was limited by the physiographic setting and by proximity to urban development of the study area. Three shots and 12 geophones, spaced 100 ft apart, were used. The seismic velocity of the unsaturated material was not determined in the field because the depth to the water table could be measured directly in each shothole. The seismic velocity of the unsaturated layer was subsequently determined using the interpretation program described by Scott and others (1972), and by adjusting the seismic velocity of layer 1 until the known depth to water was obtained.

Other studies conducted in similar settings are described by Johnson (1954) and by Sander (1978).

## Hydrogeologic settings in which seismic-refraction techniques may work, but with difficulty

The main limitations that may prevent successful completion of a seismic-refraction survey are (1) the lack of seismic-velocity contrasts between geologic units or hydrologic boundaries, (2) the presence of a thin, intermediate-

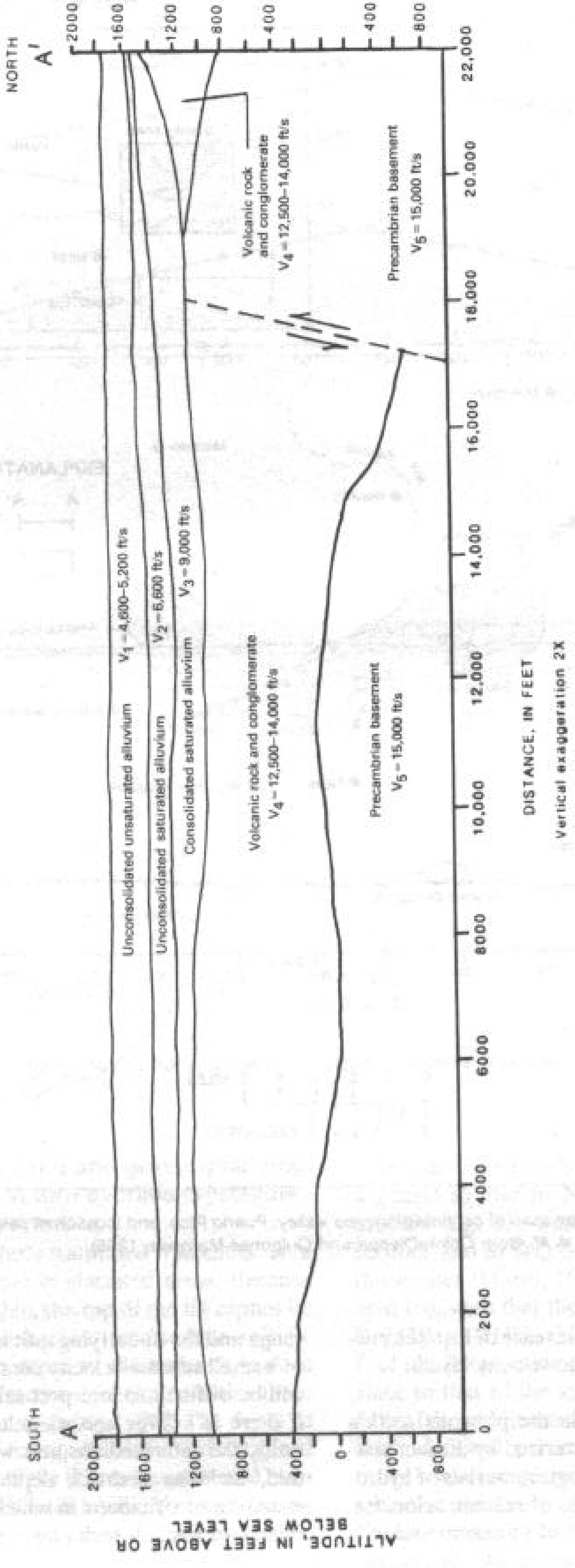



Figure 18.—Interpreted seismic section A-A' in Aura-Altar basin, near Tucson, Ariz. (Patrick Tucci, written commun., 1981).

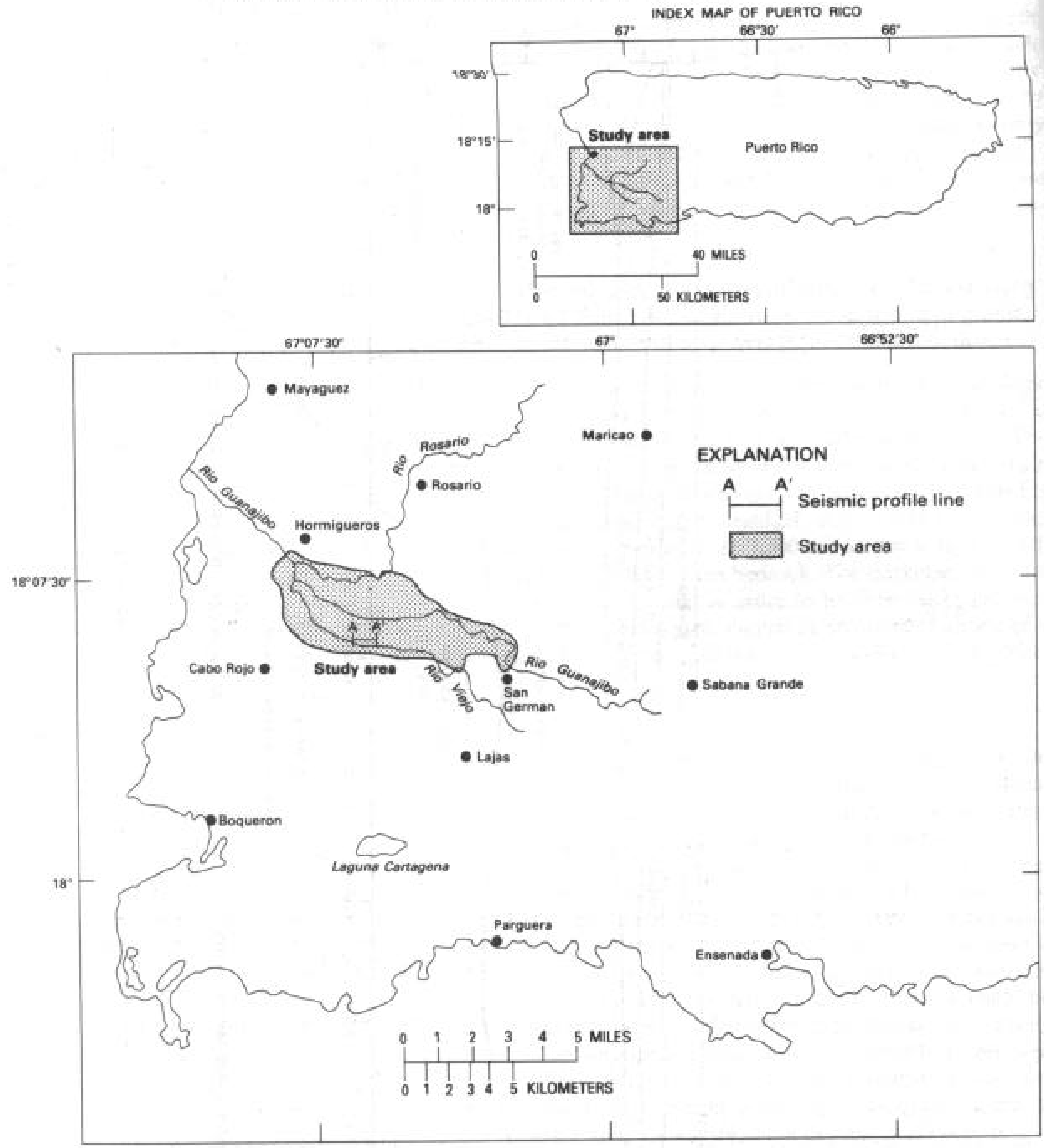



Figure 19.—Generalized location map of central Guanajibo Valley, Puerto Rico, and location of seismic-refraction profile A-A' (from Colon-Dieppa and Quinones-Marquez, 1985).

seismic-velocity layer, and (3) the presence of low- seismic-velocity layers beneath high-seismic-velocity layers.

All of the examples discussed in the previous section describe geologic materials characterized by distinct seismic velocities. However, some geologic materials or hydrogeologic units display a wide range of seismic velocities. When one unit is at the upper end of its seismic-velocity range and the underlying unit is at the lower end, resulting in a small seismic-velocity contrast across the boundary, it will be difficult to interpret seismic-refraction data. Even if there is a large seismic-velocity contrast between two units, the intermediate unit will not be detected if it is thin, and the bedrock depth will be in error. Seven examples of situations in which it may be difficult to use seismic-refraction techniques are presented below.

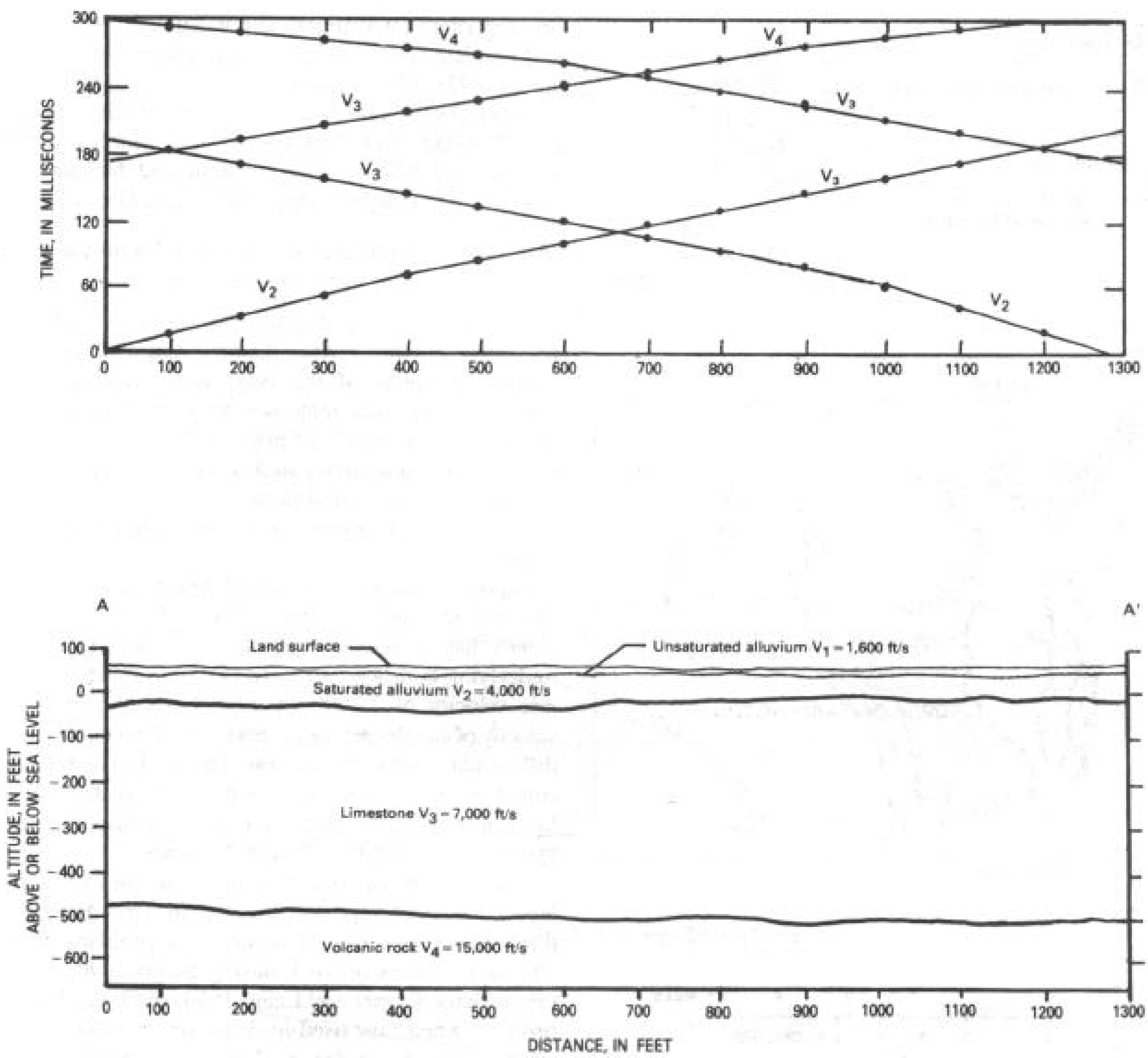



Figure 20.—Time-distance plot and interpreted seismic section at Guanajibo Valley, Puerto Rico.

### Unconsolidated glacial sand and gravel overlying a thin till layer, which in turn overlies crystalline bedrock

Determining the aquifer's saturated thickness is a common hydrogeologic goal in glaciated areas. Because many basal till layers are thin, the top of the till cannot be determined even though it has an intermediate seismic velocity of 7,000 ft/s. The depth to the bedrock surface determined by seismic-refraction techniques under these conditions will be incorrect (Sander, 1978). The depth to bedrock, and thickness of the aquifer, can be determined accurately if the thickness of the till can be estimated from drill-hole or other data. Thin till layers, however, can be considered negligible for the purpose of many hydrologic studies.

In a modeling study of the ground-water availability of a glacial aquifer in Newtown, Conn., seismic-refraction profiles (fig. 23) were used to determine the depth to bedrock and to help determine the saturated thickness of the aquifer (Haeni, 1978). Existing drill-hole data in this area indicated that the saturated aquifer material ranged from 10 to 100 ft in thickness and was underlain by 5 to 10 ft of till. Because the till was thin, its seismic velocity was close to that of the saturated material, 7,500 ft/s versus 5,000 ft/s, and because the accuracy of seismic-refraction methods is ±10 percent, the seismically determined depth to rock was considered to be the true depth to rock. The saturated thickness of the aquifer, determined from the refraction results, was arbitrarily decreased by 5 ft to account for the presence of the till.

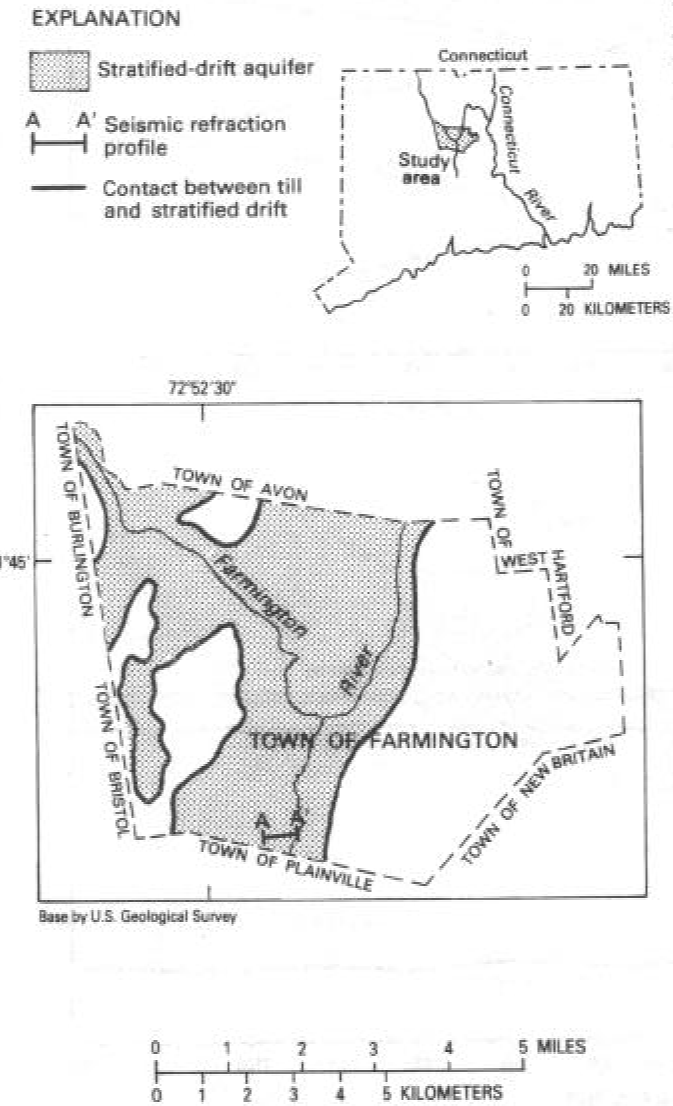



Figure 21.—Generalized location map of Farmington, Conn., and location of seismic-refraction profile A–A'.

Figure 24 shows a time-distance plot and the interpreted seismic section of one of the seismic-refraction profiles conducted for this study. In this profile, three overlapping geophone spreads with a geophone spacing of 50 ft and a total of seven shotpoints were used. Small explosive charges, weighing from 1/3 to 2 lb and placed at the water table, were used as energy sources. The depth to water was recorded in each shothole and the seismic velocity of the unsaturated zone was determined by the interpretation process described by Scott and others (1972), by adjusting the seismic velocity of layer 1 until the known depth to water was matched. Figure 23 shows a map of the saturated thickness of the aquifer as determined by the refraction survey and drill-hole control.

Other hydrologic studies using seismic-refraction techniques, and conducted in similar hydrogeologic settings, are described by Warrick and Winslow (1960), Watkins and Spieker (1971), Birch (1976), Dickerman and Johnston (1977), Sharp and others (1977), Sander (1978), Frohlick (1979), Haeni and Anderson (1980), Mazzaferro (1980), Grady and Handman (1983), Morrissey (1983), Tolman and others (1983), Haeni and Melvin (1984), Mazzaferro (1984), Winter (1984), and Haeni (1986).

### An aquifer underlain by bedrock having a similar seismic velocity

The exploration goal in this hydrogeologic setting is to determine the thickness of the upper aquifer. Because the seismic velocities of the two layers overlap, seismic-refraction methods may not yield useful information about the thickness of the upper aquifer. The success of a seismic-refraction survey in this setting will depend on the actual velocity of sound in the subsurface materials and the accuracy of seismograph and field data-collection activities.

Figure 25 shows hypothetical time-distance plots for a situation in which the upper aquifer (for example, sandstone) has a seismic velocity of 10,000 ft/s and the underlying bedrock (for example, limestone) has a seismic velocity of 10,000 to 20,000 ft/s. As the seismic velocity of the deeper layer increases, it becomes easier to differentiate between the two layers. If the velocity of sound in the second layer approaches that of the first layer, it may not be possible to differentiate between the two using seismic-refraction techniques.

The problem of similar seismic velocities for adjacent layers has been reported for several hydrogeologic settings. Broadbent (1978) describes a problem in which alluvium overlies bedrock having an unusually low seismic velocity. Topper and Legg (1974) discovered a similar problem when they tried to determine the thickness of a weathered rock aquifer overlying unweathered rock.

### A study area having a surface layer that varies significantly in thickness or material composition

The exploration goal is to map the depth to the undulating surface of a high-velocity layer in an area that has discontinuous, shallow, low-seismic-velocity materials. Seismic-refraction techniques may work here, but with some difficulty. It will be difficult to differentiate between the effects of the discontinuous surficial material and the effects of the undulating refractor. Pakiser and Black (1957) describe how to differentiate between these effects in a simple geologic setting.

Figure 26 shows a seismic section and the resulting time-distance plot in an area that has relief on a refracting surface and seismic-velocity discontinuities in the upper unit. The delay time in first arrival energy at a particular geophone, caused by a surficial low-velocity unit, will be equal for shots from both ends of the spread. The delay time at any geophone caused by relief on the refracting

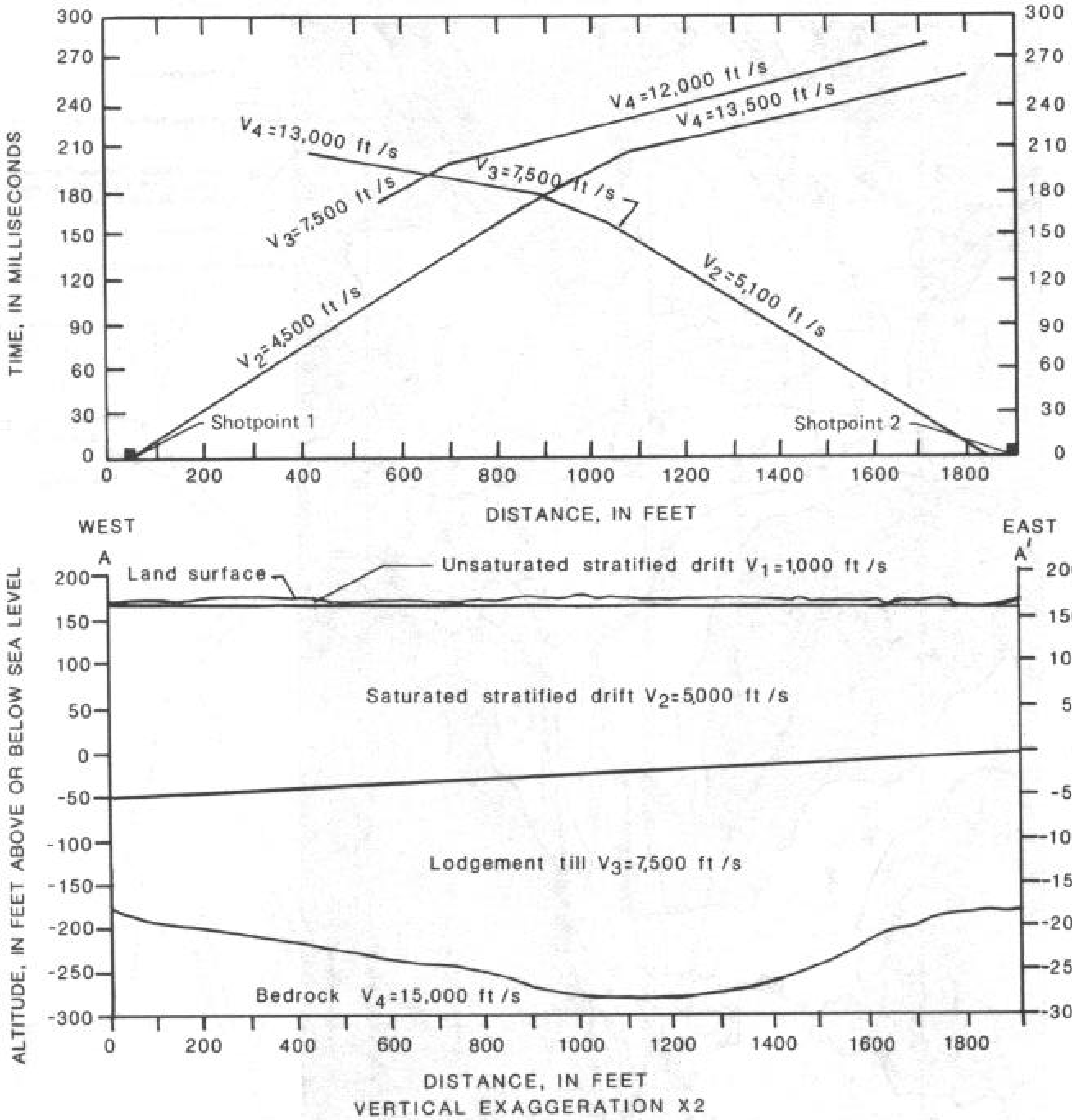



Figure 22.—Time-distance plot and interpreted seismic section near Farmington, Conn.

surface, on the other hand, will be different for shots from opposite ends of the spread. Shown is a very simple example; as the relief on the refracting surface and the number of shallow discontinuities increases, the problem becomes more difficult to solve.

### Quantitative estimation of aquifer hydraulic properties

The purpose of some seismic-refraction studies is to obtain estimates of aquifer hydraulic properties. Seismic-refraction methods do not provide a direct measurement of such aquifer properties as permeability or porosity. However, an empirical relationship may be developed and used in areas where the hydrologic setting is known. Although this use of seismic-refraction methods has been demonstrated in some studies (Eaton and Watkins, 1967; Wallace and Spangler, 1970; Watkins and Spieker, 1971; van Zijl and Huyssen, 1971; Barker and Worthington, 1973; Worthington, 1975; Worthington and Griffiths, 1975;

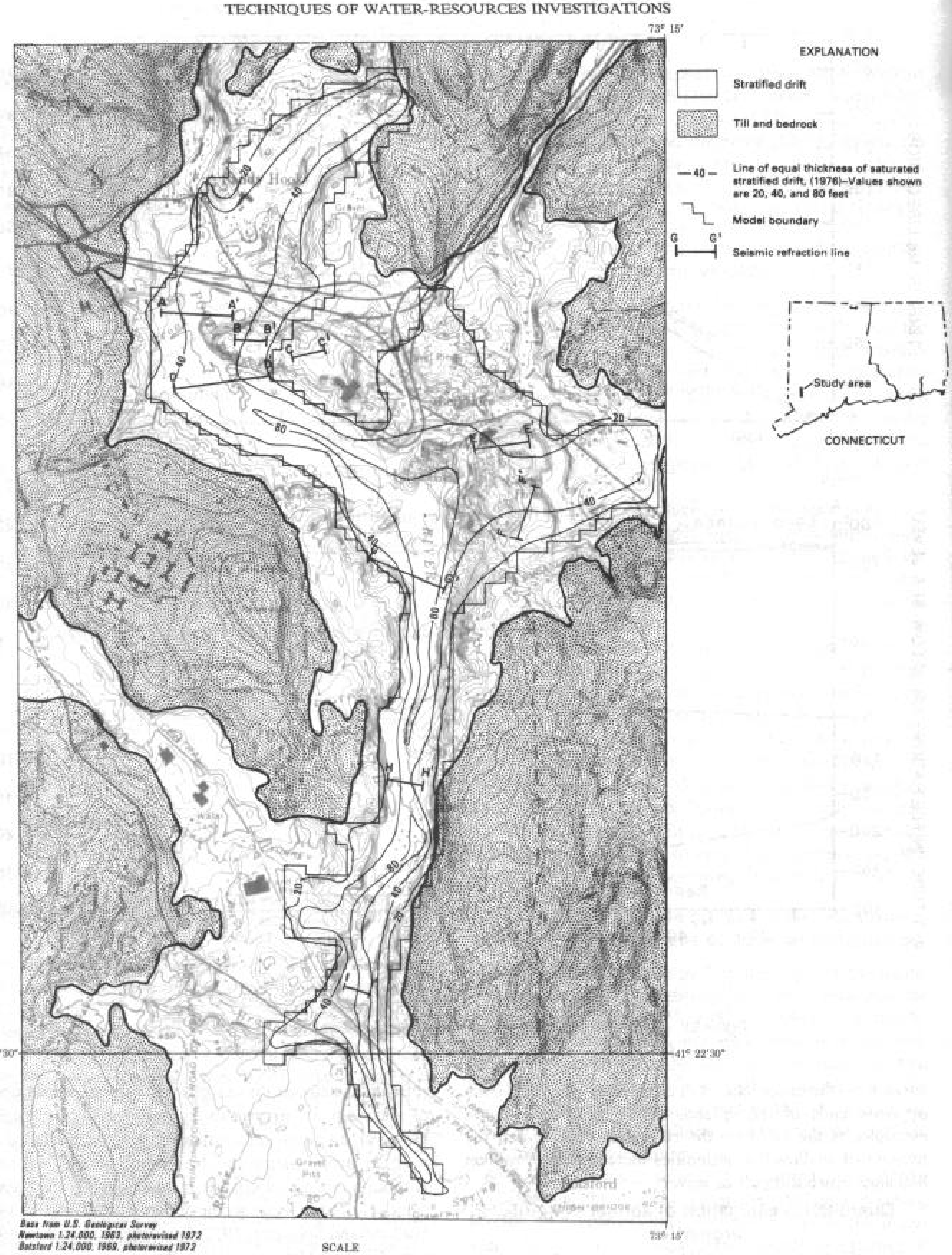



Figure 23.—Saturated thickness of stratified drift and location of seismic-refraction lines in the Pootatuck River valley, Newtown, Conn. (from Haeni, 1978).

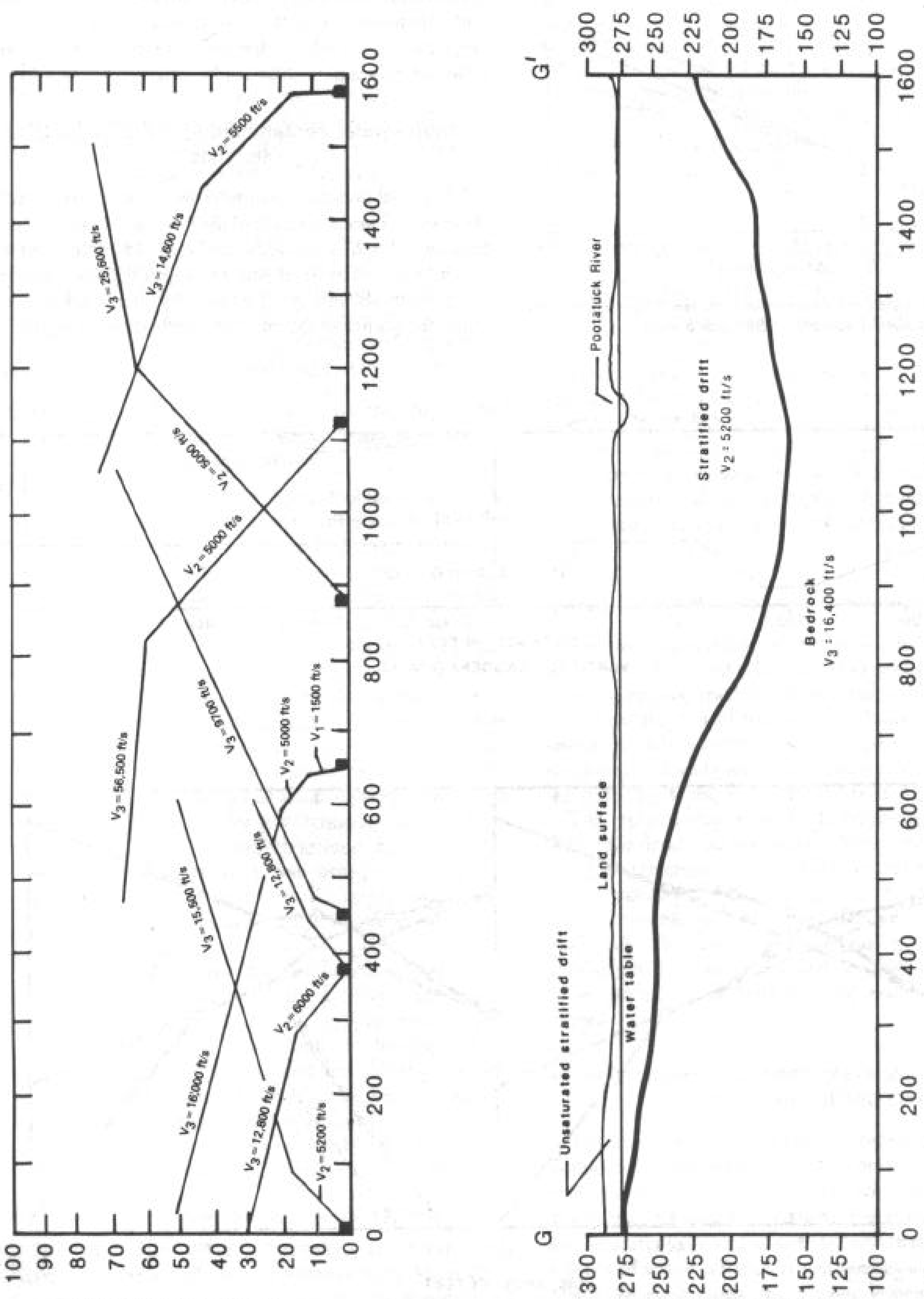



Figure 24.—Time-distance plot and interpreted seismic section of Pootatuck River valley, Newtown, Conn. (from Haeni, 1978).

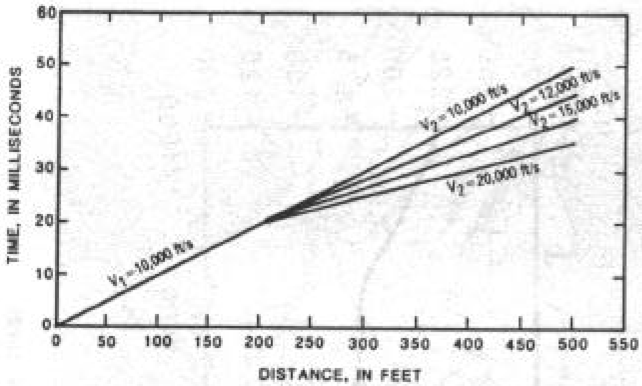



Figure 25.—Hypothetical time-distance plots resulting from different seismic velocities in the second layer.

Duffin and Elder, 1979), much remains to be investigated and documented. It must be emphasized that most of the empirical relationships developed in these studies are valid for only a particular study area.

## Ground-water contamination in unconsolidated materials

The initial phases of ground-water-contamination studies involve characterization of the hydrogeology at the site. Seismic-refraction methods can be used to determine the depth to the water table and the depth to rock, although these methods will not provide any direct information about the nature or extent of contamination of the ground

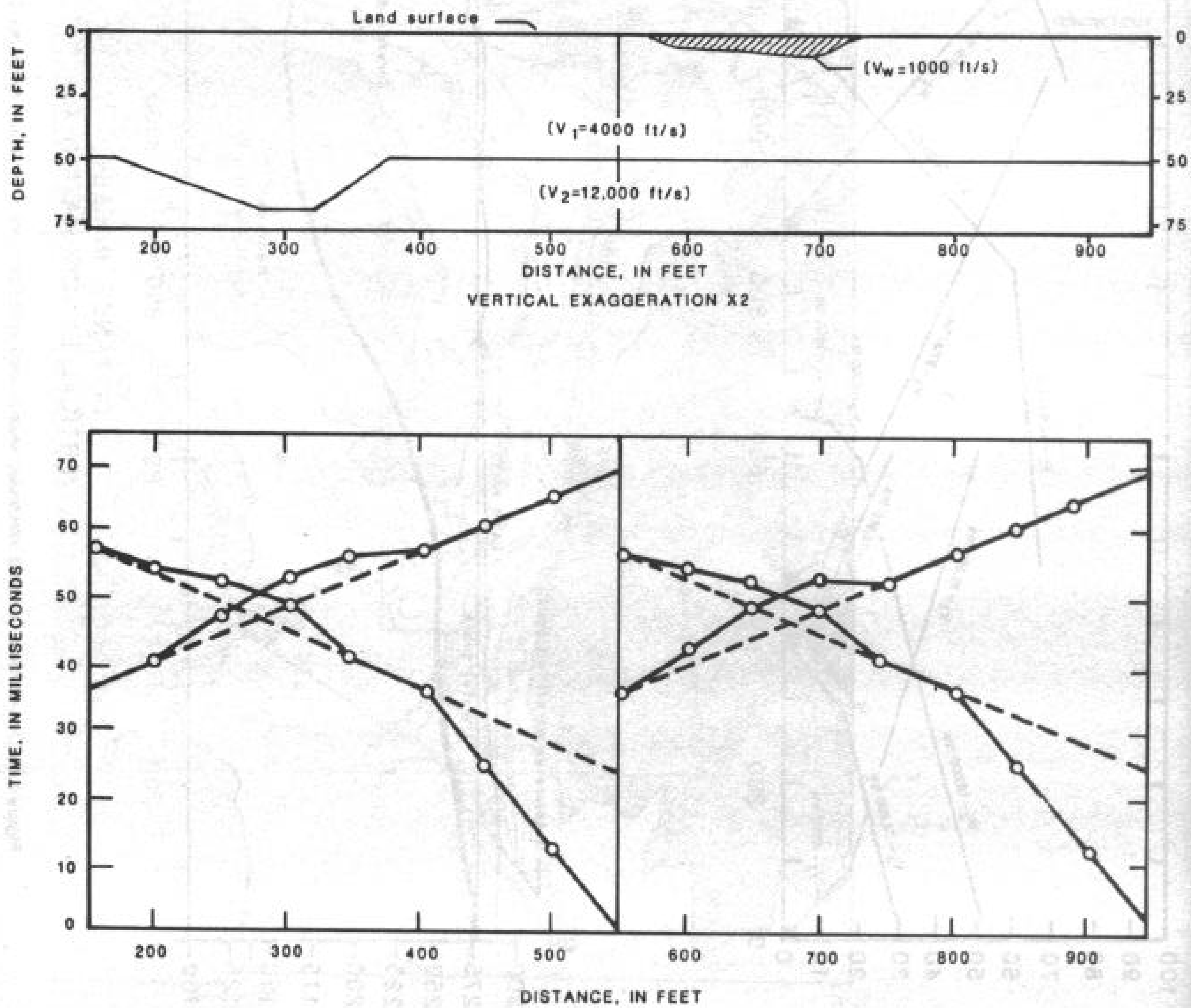



Figure 26.—Seismic section with shallow seismic-velocity discontinuities and relief on a refracting surface, and the resulting time-distance plot, Monument Valley area of Arizona and Utah (modified from Pakiser and Black, 1957).

water. This information must be obtained from other surface geophysical methods such as electrical-resistivity or electromagnetic methods.

In a ground-water-contamination study of a municipal landfill site in Farmington, Conn., Grady and Haeni (1984) used three seismic-refraction profiles to define the water table and the bedrock surface at the site. Figure 27 shows the landfill, the location of the seismic-refraction lines, and one interpreted seismic section. Multiple overlapping geophone spreads and multiple shotpoints were used to provide tight control on the depth of the water table and to provide a continuous bedrock profile.

Other ground-water-contamination studies that used seismic-refraction methods to characterize the hydrogeology of the site include studies by Bianchi and Nightingale (1975), Leisch (1976), and Yaffe and others (1981).

### A multilayered Earth with a shallow, thin layer that has a seismic velocity greater than the layers below it

The exploration goal in this hydrogeologic setting is to determine the depth to a particular refractor through the high-seismic-velocity layer. In most cases, the presence of a shallow high-seismic-velocity layer prevents accurate determination of the depth of a deep refractor underlain by a low-seismic-velocity refractor (see section on "Limitations"). If the high-seismic-velocity layer is very thin, however, seismic- refraction techniques may work.

Bush and Schwarz (1965) found that a thin layer of frozen unconsolidated material did not prevent accurate determination of the depth of the underlying rock surface. The velocity of the frozen material was 14,000 ft/s, and the seismograph records contained some high-frequency early energy arrivals followed by low-frequency arrivals from bedrock. In areas of thick frozen ground, however, calculation of the depth to rock was usually not possible. Ackermann (1976) also used seismic-refraction methods to locate unfrozen materials for water supplies in permafrost areas in Alaska.

Morony (1977) found that a shallow high-seismic-velocity (9,500 ft/s) limestone 33 ft thick underlain by lower seismic-velocity (6,600 ft/s) aquifer material prevented determination of the depth to basement rock (seismic velocity 16,000 ft/s) and the thickness of the limestone unit. Using drill-hole data for the thickness of the limestone, and assuming a velocity of the underlying saturated aquifer material, a reasonable depth to basement rock of 450 ft was calculated from the seismic data.

### Miscellaneous hydrogeologic settings

There are several other hydrogeologic settings in which seismic-refraction techniques have been used. Shields and Sopper (1969) used these techniques in a watershed hydrology study. Depth to rock and depth to water, determined from seismic-refraction profiles, were used to help characterize the hydrologic properties of the watershed.

Winter (1984) used seismic-refraction methods in a lake hydrology study of Mirror Lake, N.H. In this study, the interaction of the ground-water system and the water in the lake was studied, and seismic-refraction methods were used to map the saturated thickness of unconsolidated materials around the lake and in the surrounding watershed.

## Hydrogeologic settings in which seismic-refraction techniques cannot be used

Seismic-refraction methods cannot be used successfully to detect (1) low-seismic-velocity layers overlain by high-seismic-velocity layers, (2) two hydrologically different units having the same seismic velocity, or (3) thin beds of intermediate seismic velocity in a sequence of beds whose seismic velocities increase with depth. Three examples of situations in which these limitations apply are cited below.

### Basalt flows with interflow zones that are aquifers

The most important aquifers in layered basalt formations or other layered volcanic rocks generally occur in the zones of rubbly, vesicular, brecciated, or weathered rock that form the top of many of the lava flows, or in the sediments that accumulate on the surface of a flow prior to successive lava flows. These interflow zones are usually separated by dense, unfractured basalt.

The exploration goal in this hydrogeologic setting is to define the depth and thickness of these interflow aquifers. Seismic-refraction techniques will not work, because the seismic velocity of the dense basalt is 15,000 to 20,000 ft/s and the seismic velocity of the interflow zone is 5,000 to 7,000 ft/s. The condition of increasing seismic velocity with depth does not hold, and the low-seismic-velocity layer cannot be defined with seismic-refraction techniques.

### Unconsolidated sand and gravel aquifer material underlain by silt and clay

The exploration goal in this hydrogeologic setting is to define the areal extent and thickness of the sand and gravel aquifer. Seismic-refraction techniques usually cannot be used to solve this problem. The velocity of sound in the saturated clay and silt will be almost the same as the velocity of sound in the saturated sand and gravel (Burwell, 1940). In most cases, the seismic velocities of the two hydrogeologic units cannot be differentiated on the time-distance plot. Resisitivity techniques may work in this setting.