NATIONAL OPEN UNIVERSITY OF NIGERIA

SCHOOL OF SCIENCE AND TECHNOLOGY

COURSE CODE: PHY 361

COURSE TITLE: GEOPHYSICS II

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UNIT 1: Seismic Methods

1.0 Introduction

In this unit, you will be introduced to the basic concept of Seismic Methods. The geophysical techniques mostly widely employed for exploration work are the seismic, gravity, magnetic, electrical, and electromagnetic methods. Less common methods involve the measurement of radioactivity and temperature at or near the earth's surface and the air. Some of these methods are used entirely in the search for oil and gas.
Seismic methods are the most effective, and the most expensive, of all the geophysical techniques used to investigate layered media. Features common to reflection and refraction surveys are discussed in this section.

2.0 Objectives

At the end of this unit, readers should be able to:

(i) Understand that seismic wave is the basic measuring rod used in seismic prospecting.
(ii) Be familiar with the basic physical principles governing its propagation characteristics and seismic velocities.

(iii) Understand that seismic waves are generally referred to as elastic waves.

(iv) Describe the characteristics of seismic waves and formation of other types of waves

3.0 Main Contents

3.1 Seismic Waves

A seismic wave is acoustic energy transmitted by vibration of rock particles. Low-energy waves are approximately elastic; leaving the rock mass unchanged by their passage, but close to a seismic source the rock may be shattered and permanently distorted.

3.1.1 Types of elastic wave

When a sound wave travels in air, the molecules oscillate backwards and forwards in the direction of energy transport. This *pressure* or 'push' wave thus travels as a series of compressions and rarefactions. The pressure wave in a solid medium has the highest velocity of any of the possible wave motions and is therefore also known as the *primary* wave or simply the *P wave*. Particles vibrating at right angles to the direction of energy flow (which can only happen in a solid) create an *S* (*shear*, 'shake' or, because of its relatively slow velocity, *secondary*) wave. The velocity in many consolidated rocks is roughly half the P-wave velocity. It depends slightly on the plane in which the particles vibrate but these differences are not significant in small-scale surveys.

P and S waves are *body waves* and expand within the main rock mass. Other waves, known as *Love waves*, are generated at interfaces, while particles at the Earth's surface can follow elliptical paths to create *Rayleigh waves*. Love and Rayleigh waves may carry a considerable proportion of the source energy but travel very slowly. In many surveys they are simply lumped together as the *ground roll*.

3.1.2 Seismic velocities

The 'seismic velocities' of rocks are the velocities at which wave motions travel through them. They are quite distinct from the continually varying velocities of the individual oscillating rock particles. Any elastic-wave velocity (V) can be expressed as the square root of an elastic modulus divided by the square root of density (ρ). For P waves the elongational elasticity, j is appropriate, for S waves the shear modulus, μ. The equations:

$Vp = (j/\rho)$ $Vs = (\mu/\rho)$

Suggests that high density rocks should have low seismic velocities, but because elastic constants normally increase rapidly with density, the reverse is usually true. Salt is the only common rock having a high velocity but a low density.
If the density and P and S wave velocities of a rock mass are known, all the elastic constants can be calculated, since they are related by the equations:

$(Vp/Vs)2 = 2(1 - \sigma)/(1 - 2\sigma)$ $\sigma = [2 - (Vp/Vs)2]/2[1 - (Vp/Vs)2]$

$j = q(1 - \sigma)/(1 + \sigma)(1 - 2\sigma)$ $\mu = q/2(1 + \sigma)$ $K = q/3(1 - 2\sigma)$

Where σ is the Poisson ratio, q is the Young's modulus and K is the bulk modulus. It follows that $j = K + 4\mu/3$ and that a P wave always travels faster than an S wave in the same medium. The Poisson ratio is always less than 0.5. At this limit, Vp/Vs is infinite.

Most seismic surveys provide estimates only of P-wave velocities, which are rather rough guides to rock quality. Figure 1.1 shows ranges of velocity for common rocks and also their *rippabilities*, defined by whether they can be ripped apart by a spike mounted on the back of a bulldozer.

3.1.3 Velocities and the time-average equation
Within quite broad limits, the velocity of a mixture of different materials can be obtained by averaging the transit times (the reciprocals of velocities) through the pure constituents, weighted according to the relative amounts present. The principle can be used even when, as in Example 1.1, one of the constituents is a liquid.

Example 1.1
Vp (quartz) = 5200 m s−1

Vp (water) = 1500 m s−1

P-wave velocity in a sandstone, 80% quartz, 20% water-filled porosity, is given by:

$1/Vp = 0.8/5200 + 0.2/1500$

$= 0.000287$

i.e. $Vp = 3480$ m s−1

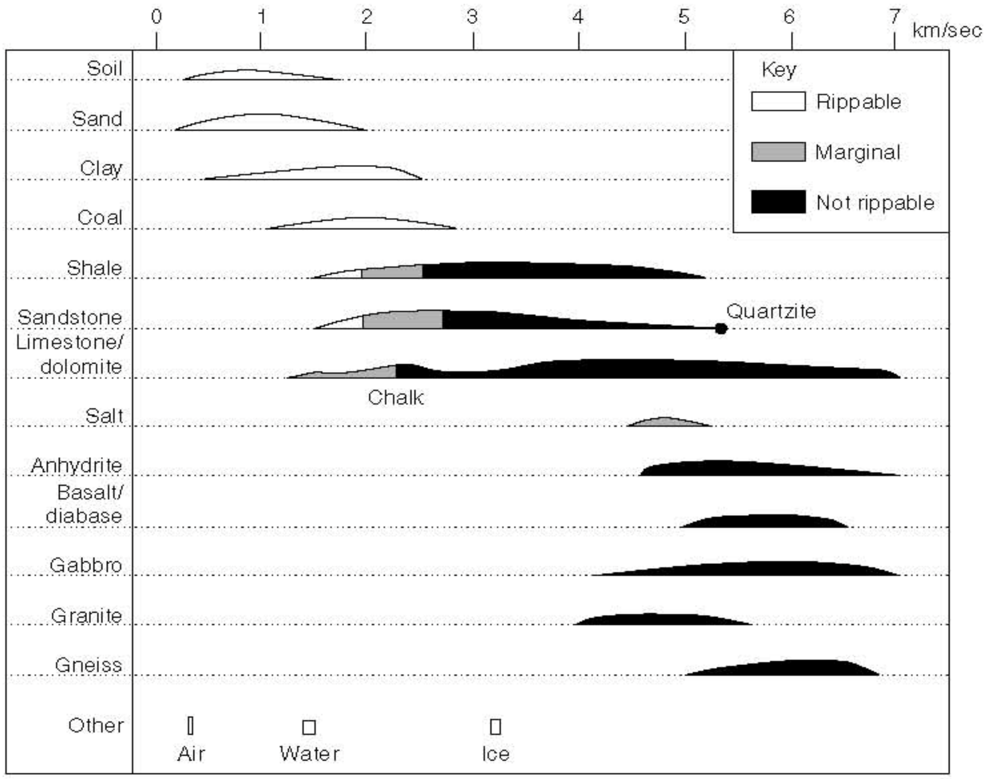

Figure 1.1: *Ranges of P-wave velocities and rippabilities in common rocks. The vertical axis, for each rock type, is intended to show approximately the relative numbers of samples that would show a given velocity.*

In dry rocks, the pore spaces are filled with air (V = 330 ms−1) rather than water. Time averaging cannot be applied quantitatively to gas-filled pores, but dry materials generally have very low P-wave velocities. If they are poorly consolidated and do not respond elastically, they may also strongly absorb S waves. Poorly consolidated water-saturated materials generally have velocities slightly greater than that of water, and the water table is often a prominent seismic interface.

Weathering normally increases porosity, and therefore reduces rock velocities. This fact underlies the rippability ranges shown in Figure 1.1. Few fresh, consolidated rocks have velocities of less than about 2200 ms−1, and rocks that are rippable are generally also at least partly weathered.

3.1.4 Ray-path diagrams

It is convenient to identify the important travel paths by drawing seismic *rays*, to which the laws of geometrical optics can be applied, at right angles to the corresponding wavefronts. Ray-path theory works less well in seismology than in optics because the most useful seismic wavelengths are between 25 and 200 m, and thus comparable with survey dimensions and interface depths. Wave effects can be significant under these circumstances but field interpretation can nonetheless be based on ray-path approximations.

3.1.5 Reflection and refraction

When a seismic wave encounters an interface between two different rock types, some of the energy is reflected and the remainder continues on its way at a different angle, i.e. is *refracted*. The law of reflection is very simple; the angle of reflection is equal to the angle of incidence (Figure 1.2a). Refraction is governed by *Snell's law,* which relates the angles of incidence and refraction to the seismic velocities in the two media:

$$\sin i / \sin r = V1/V2$$

If $V2$ is greater than $V1$, refraction will be towards the interface. If sin i equals $V1/V2$, the refracted ray will be parallel to the interface and some of its energy will return to the surface as a *head wave* that leaves the interface at the original angle of incidence (Figure 1.2b). At greater angles of incidence there can be no refracted ray and all the energy is reflected. When drawing ray paths for either reflected or critically refracted waves, allowance must be made for refraction at all shallower interfaces. Only the *normal-incidence* ray, which meets all interfaces at right angles, is not refracted.

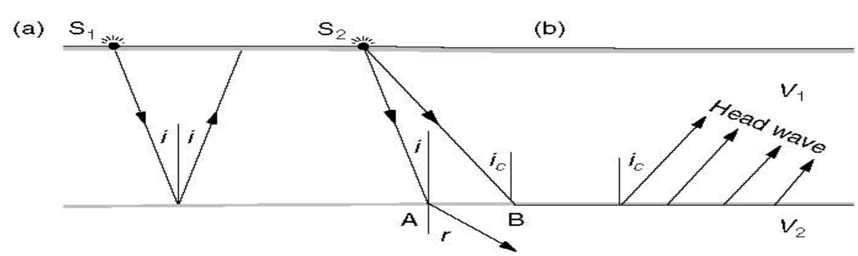

Figure 1.2 *(a) Reflection and (b) refraction. Simple refraction occurs at A, critical refraction at B.*

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4.0 Conclusion

A seismic wave is properly described in terms of *wavefronts*, which define the points that the wave has reached at a given instant. However, only a small part of a wavefront is of interest in any geophysical survey, since only a small part of the energy returns to the surface at points where detectors have been placed.

5.0 Summary

When seismic energy is released suddenly at a point P near the surface of homogeneous medium, part of the energy propagates through the body of the medium as seismic body waves. The remaining part of the seismic energy spreads out over the surface as a seismic surface wave, analogous to the ripples on the surface of a pool of water into which the stone has been thrown.

When a body waves reaches a distance 'r' from it's source in a homogeneous medium, the wavefronts (defined as the surface in which all particles vibrate with the same phase) has a spherical shape and the wave is called aspherical wave. As the distance from the source increases, the curvature of the spherical wavefront decreases. At great distances from the source the wavefront is so flat that it can be considered to be a plane and the seismic wave is called Plane wave. The direction perpendicular to the wavefront is called the seismic ray path, The description of harmonic waves in plane waves is simpler than for spherical waves, because for plane waves we can use arthogonal Cartesian coordinates. Even for plane waves the mathematical description of the three dimensional displacements of the medium is fairly complex. However, we can learn quite a lot about body wave propagation from a simpler, less rigorous description.

6.0 Tutor Marked Assignment:

Q 1.

a) What is Seismic Waves?
b) State the law of reflection
c) State the Snell's law

Q 2. With the aid of a simple diagram explain and show the following concept in seismic theory.

a) Reflection
b) Refraction
c) Head waves

Q3. Explain the following types of waves

a) Primary wave
b) Shear wave
c) Love wave
d) Body wave
e) Rayleigh wave

7.0 References/ Further Readings:

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Unit 2: Seismic Sources

1.0 Introduction

The traditional seismic source is a small charge of dynamite. Impact and vibratory sources are now more popular but explosives are still quite commonly used.

2.0 Objectives

At the end of this unit, readers should be able to:

(i) Understand clearly seismic sources
(ii) Understand the usefulness of hammers in application of seismic waves
(iii) Know other impact sources.
(iv) Understand the uses of explosives in seismic works

3.0 Main Content:

3.1 Hammers

A 4- or 6-pound sledgehammer provides a versatile source for small-scale surveys. The useful energy produced depends on ground conditions as well as on strength and skill. Hammers can nearly always be used in refraction work on spreads 10 to 20 m long but very seldom where energy has to travel more than 50 m.

The hammer is aimed at a flat plate, the purpose of which is not so much to improve the pulse (hitting the ground directly can sometimes provide more seismic energy) but to stop the hammer abruptly and so provide a definite and repeatable shot instant. Inch-thick aluminium or steel plates used to be favoured, but are now being replaced by thick rubber discs that last longer and are less painfully noisy. The first few hammer blows are often rather ineffective, as the plate needs to 'bed down' in the soil. Too much enthusiasm may later embed it so deeply that it has to be dug out.

3.2 Other impact sources

More powerful impact sources must be used in larger surveys. Weights of hundreds of kilograms can be raised by portable hoists or cranes and then dropped (Figure 1.3). The minimum release height is about 4 m, even if a shorter drop would provide ample energy, since rebound of the support when the weight is released creates its own seismic wavestrain. A long drop allows these vibrations to die away before the impact occurs. Tractor-mounted posthole drivers, common in farming areas, are also convenient sources. The weight drops down a guide and is raised by a pulley system connected to the tractor power take-off.

Relatively small (70 kg) weights falling in evacuated tubes have sometimes been used. The upper surface of the weight is exposed to the air, and effectively several hundred extra kilograms of atmosphere are also dropped. The idea is elegant but the source is difficult to transport because the tube must be strong and therefore heavy and must be mounted on a trailer, together with a motor-driven compressor to pump out the air. Vibration sources are widely used in large-scale reflection surveys but produce data that need extensive and complex processing.

3.3 Explosives

Almost any type of (safe) explosive can be used for seismic work, particularly if the shot holes are shallow and the charges will not be subject to usual

Figure1.3 *Impact source. A half-ton weight being dropped from a portable crane during a survey of the low-velocity layer.*

temperatures or pressures. Cord explosives, used in quarry blasting to introduce delays into firing sequences, are rather safer to handle than normal gelignite and can be fed into shot holes prepared by driving metal rods or crowbars into the ground. Detonators used on their own are excellent sources for shallow reflection surveys where high resolution is needed. Often, much of the energy delivered by an explosion is wasted in shattering rock near the shot point, and seismic waves are produced much more efficiently by shots fired in a metre or so of water. This effect is so marked that, if shot position is not critical, it can be worth going tens or even hundreds of metres away from the recording spread in order to put the charge in a river. In dry areas, significant improvements can be obtained by pouring water down shot holes.

Electrical firing is normal when using explosives but with ordinary detonators there is a short delay between the instant at which the filament burns through, which provides a time reference, and the time at which the main charge explodes. *Zero-delay* detonators should be used for seismic work and total delays through the entire system, including the recorders, should be routinely checked using a single detonator buried a few inches away from a geophone.

Explosives involve problems with safety, security and bureaucracy. They must be used in conformity with local regulations, which usually require separate secure and licensed stores for detonators and gelignite. In many countries the work must be supervised by a licensed shot-firer, and police permission is required almost everywhere. Despite these disadvantages, and despite the headaches that are instantly produced if gelignite comes into contact with bare skin, explosives are still used. They represent potential seismic energy in its most portable form and are virtually essential if signals are to be detected at distances of more than 50 m.

A variety of explosive-based methods are available which reduce the risks. Seismic waves can be generated by devices which fire lead slugs into the ground from shotgun-sized cartridges, but the energy supplied is relatively small, and a firearms certificate may be needed, at least in the UK. Another approach is to use blank shotgun cartridges in a small auger which incorporates a firing chamber, combining the shot hole and the shot. Even this seldom provides more energy than a blow from a well-swung hammer, and is less easily repeated.

3.4 Safety

Large amounts of energy must be supplied to the ground if refractions are to be observed from depths of more than a few metres or reflections from depths of more than a few tens of metres, and such operations are inherently risky. The dangers are greatest with explosives but nor is it safe to stand beneath a half-ton weight dropping from a height of 4 m.

Explosives should only be used by experienced (and properly licensed) personnel. Even this does necessarily eliminate danger, since experts in quarry blasting often lack experience in the special conditions of seismic surveys. If there is an accident, much of the blame will inevitably fall on the party chief who will, if he is wise, keep his own eye on safety.

The basic security principle is that the shot-firer must be able to see the shot point. Unfortunately, some seismographs have been designed so that the shot is triggered by the instrument operator, who can seldom see anything and who is in any case preoccupied with checking noise levels. If such an instrument is being used, it must at least be possible for firing to be prevented by someone who is far enough from the shotpoint to be safe but close enough to see what is happening. This can be achieved if, after the shot hole has been charged, the detonator is first connected to one end of an expendable cable 20 or 30 m long. Only when the shot point is clear should the other end of this cable be connected to the cable from the firing unit. Firing can then be prevented at any time by pulling the two cables apart.

Unless ‘sweaty’ gelignite is being used (and the sight of oily nitro-glycerine oozing out of the packets should be sufficient warning to even the least experienced), modern explosives are reasonably insensitive to both heat and shock. Detonators are the commonest causes of accidents. Although their explosive power is small, they have caused loss of fingers and even hands. If fired on their own as low energy sources, they should always be placed in well-tamped holes, since damage or serious injury can be caused by fragments of the metal casing.

3.5 Time breaks

In any seismic survey, the time at which the seismic wave is initiated must be known. In some instruments this appears on the record as a break in one of the traces (the *shot break* or *time break*). On most modern instruments it actually defines the start of the record. Time-break pulses may be produced in many different ways. A geophone may be placed close to the source, although this is very hard on the geophone.

Explosive sources are usually fired electrically, and the cessation of current flow in the detonator circuit can provide the required signal. Alternatively, a wire can be looped around the main explosive charge, to be broken at the shot instant. This technique can be used on the rare occasions when charges are fired using lit fuses.

Hammer surveys usually rely on making rather than breaking circuits. One method is to connect the hammer head to one side of the trigger circuit and the plate (assuming it is metal, not rubber) to the other. Although this sounds simple and foolproof, in practice the repeated shocks suffered by the various connections are too severe for long-term reliability. In any case, the plates themselves have rather short lives, after which new connections have to be made. It is more practical to mount a relay on the back of the hammer handle, just behind the head, that closes momentarily when the hammer hits the plate (Figure 1.4).

It will close late, or not at all, if the hammer is used the wrong way round. Solid-state switches sold by some seismograph manufacturers give more repeatable results but are expensive and rather easily damaged. The cable linking the trigger switch on a hammer to the recorder is always vulnerable, tending to snake across the plate just before impact. If it is cut, the culprit is traditionally required both to repair the damage and ease the thirst of all the witnesses!

Where the source is a heavy weight dropped from a considerable height, a relay switch can be attached to its top surface but may not trigger if the drop is not absolutely straight. A crude but more reliable home-made device which can be attached to any dropping weight is shown in Figure 1.5.

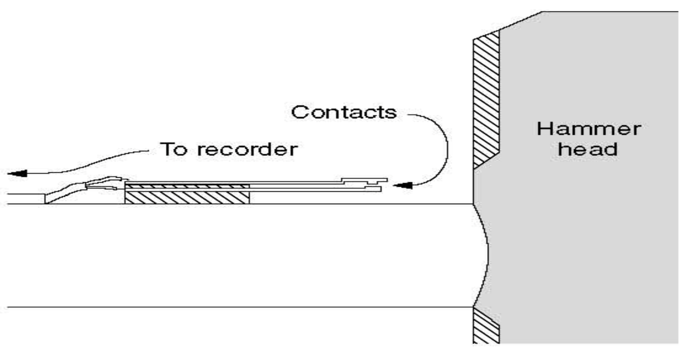

Figure 1.4 *'Post-office relay' impact switch on the back of a sledgehammer handle.*

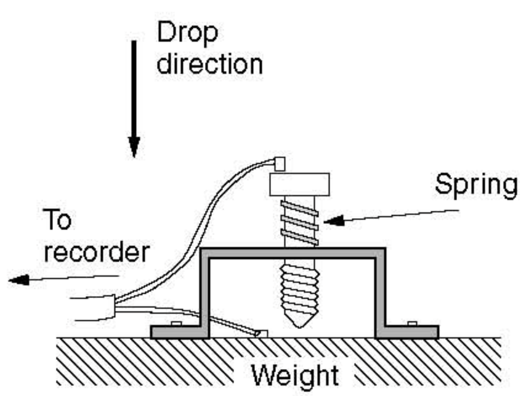

Figure 1.5 *Weight-drop contact switch. On impact the inertia of the bolt compresses the spring and contact is made with the upper surface of the weight.*

Time-break pulses may be strong enough to produce interference on other channels (*cross-talk* ; Section 1.3.5). Trigger cables and circuits should therefore be kept well away from data lines.

4.0 Conclusion

It is possible (although not common) for a detonator to be triggered by currents induced by power lines or radio transmissions but this is less likely if the leads are twisted together. Triggering by static electricity is prevented if the circuit is closed. The shorted, twisted, ends of detonator leads should be parted only when the time comes to make the connection to the firing cable, which should itself be shorted at the far end. Explosives should not be handled at all when thunderstorms are about.

5.0 Summary

Explosive charges need to be matched to the holes available. Large charges may be used in deep holes with little obvious effect at the surface, but a hole less than 2 m deep will often blow out, scattering debris over a wide area. Only experience will allow safe distances to be estimated and even experienced users can make mistakes; safety helmets should be worn and physical shelter such as a wall, a truck or a large tree should be available. Heavy blasting mats can reduce blow-outs, but their useful lives tend to be short and it is unwise to rely on them alone.

A point where a shot has been fired but no crater has formed should be regarded with suspicion. The concealed cavity may later collapse under the weight of a person, animal or vehicle, leading to interesting litigation.

6.0 Tutor Marked Assignment

Q1. Write short notes on the following in relation to seismic work.

a) Hammer
b) Explosives
c) Safety
d) Time Breaks.
e) crater

7.0 References/ Further Readings:

Kearey, P., Brooks, M. and Hill, I. (2002) An Introduction to Geophysical Exploration (Third Edition), Blackwell Science, Oxford, 262 pp.
John, M. (2003) Field Geophysics (Third Edition). John Wiley and Sons Ltd. England, 249pp
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Unit 3: Detection of Seismic Waves

1.0 Introduction

In this unit, detection of seismic waves are discussed. Land seismic detectors are known as *geophones*, marine detectors as *hydrophones*. Both convert mechanical energy into electrical signals. Geophones are usually positioned by pushing a spike screwed to the casing firmly into the ground but it may be necessary to unscrew the spike and use some form of adhesive pad or putty when working on bare rock.

2.0 Objectives

At the end of this unit, readers should be able to:

(i) Differentiate between geophones and hydrophones
(ii) Apply geophones on land and hydrophones on water
(iii) Know that geophones are referred to as seismometer or detectors.

3.0 Main Contents

3.1 Geophones

A geophone consists of a coil wound on a high-permeability magnetic core and suspended by leaf springs in the field of a permanent magnet (Figure 1.6). If the coil moves relative to the magnet, voltages are induced and current will flow in any external circuit. The current is proportional to the velocity of the coil through the magnetic field, so that ground movements are recorded, not ground displacements. In most cases the coil is mounted so that it is free to vibrate vertically, since this gives the maximum sensitivity to P waves rising steeply from subsurface interfaces, i.e. to reflected and refracted (but not direct) P waves. P-wave geophones that have been normally connected give negative first-arrival pulses (*breaks*) for refractions and reflections, but may break either way for direct waves.

In reflection work using large offsets, or in refraction work where the velocity contrasts between overburden and deeper refractors are small, the rising wavefronts make relatively large angles with the ground surface and the discrimination by the geophones between S waves and P waves will be less good.

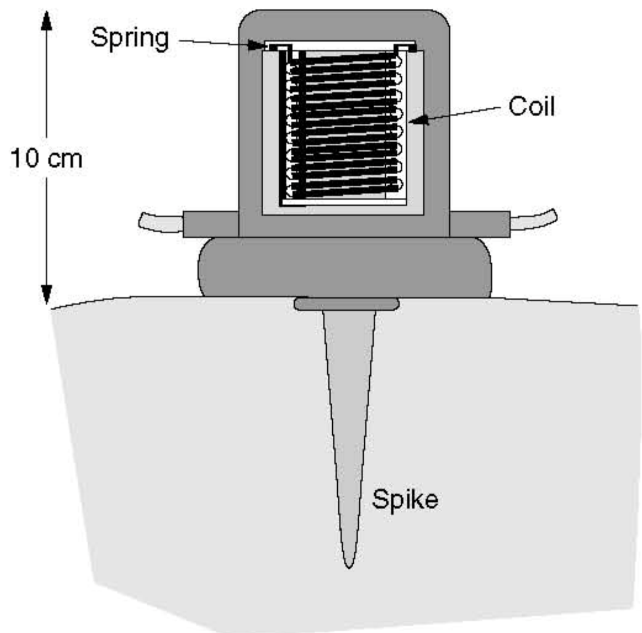

Figure 1.6 *Moving coil geophone.*

Geophone coils have resistances of the order of 400 ohms and damping is largely determined by the impedance of the circuits to which they are linked. The relative motion between coil and casing is also influenced by the natural vibration frequency of the suspended system. At frequencies above resonance the response approximately replicates the ground motion, but signals below the resonant frequency are heavily attenuated. Standard geophones usually resonate at or below 10 Hz, i.e. well below the frequencies useful in small-scale surveys. Response curves for a typical 10 Hz phone are shown in Figure 1.7.

Geophones are remarkably rugged, which is just as well considering the ways in which they are often treated. Even so, their useful lives will be reduced if they are dumped unceremoniously from trucks into tangled heaps on the ground. Frames can be bought or made to which they can be clipped for carrying (Figure 1.8) and these can be good investments, but only if actually used.

3.2 Detection of S waves

Although S waves are regarded as noise in most seismic work, there areoccasions when S-wave information is specifically sought. For example, both S- and P-wave velocities are required to determine elastic properties.

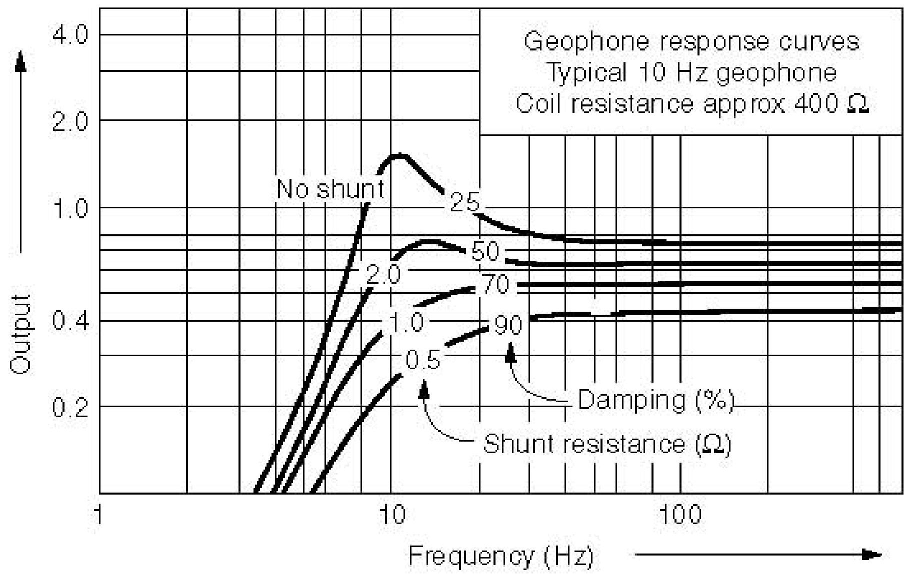

Figure 1.7 *Frequency response of a typical moving-coil geophone. The degree of damping depends on the value of the shunt resistance connected in parallel with the geophone, and also on the input resistance of the recorder's No shunt' corresponds to infinite shunt resistance.*

Figure 1.8 *Geophone carrying frame in use, Papua New Guinea.*

‘S-wave’ geophones have coils that move horizontally rather than vertically, the assumption being that wavefronts of interest will be rising more or less vertically and the S-wave vibrations will therefore be in the plane of the ground surface. Because direct waves travel parallel to the ground surface, S-wave geophones are more sensitive to direct P waves than direct S waves, just as P-wave geophones are sensitive to vertically polarized direct S waves.

3.3 Detection in swamps and water

Normal geophones are rainproof rather than waterproof, and are connected to cables by open crocodile clips. Geophones are also available that are completely enclosed and sealed into waterproof cases, for use in swamps. These do not have external spikes but are shaped so that they can be easily pushed into mud.
Motion-sensitive instruments cannot be used in water. Piezo-electric hydrophones respond to variations in pressure rather than motion and are equally sensitive in all directions. Discrimination between P and S waves is not required since S waves cannot travel through fluids.

3.4 Noise

Any vibration that is not part of the signal is *noise*. Noise is inevitable and *coherent* noise is generated by the shot itself. S waves, Love and Rayleigh waves and reflections from surface irregularities are all forms of coherent noise. In shallow refraction work these slow, and therefore late-arriving, waves usually prevent the use of any event other than the first arrival of energy.

Noise which is not generated by the shot is termed *random*. Movements of traffic, animals and people all generate random noise and can, to varying extents, be controlled. It should at least be possible to prevent the survey team contributing, by giving warning using a whistle or hooter. Random noise is also produced by vegetation moving in the wind and disturbing the ground. The effects can be reduced by siting geophones away from trees and bushes, and sometimes by clearing away smaller plants. Significant improvements can often be achieved by moving particularly noisy geophones a few inches. Placement is also important. It may not be easy to push a spike fully home in hard ground but a geophone an inch above the ground vibrates in the wind.

3.5 Seismic cables

Seismic signals are carried from geophones to recorders as varying electric currents, in cables which must contain twice as many individual wires as there are geophones. Wires are necessarily packed very closely and not only can external current carriers such as power and telephone cables induce currents, but a very strong signal in one wire can be passed inductively to all the others.

Cross-talk can be particularly severe from the strong signals produced by geophones close to the shot point, and it may even be necessary to disconnect these to obtain good records on other channels.

The amount of cross-talk generally increases with the age of the cable, probably because of a gradual build-up of moisture inside the outer insulating cover. Eventually the cable has to be discarded. Cables and plugs are the most vulnerable parts of a seismic system and are most at risk where they join. It is worthwhile being very careful. Resoldering wires to a plug with 24 or more connections is neither easy nor interesting. Most cables are double-ended, allowing either end to be connected to the receiver. If a wire is broken, only the connection to one end will be affected and the 'dead' channel may revive if the cable is reversed. All too often, however, other dead channels are discovered when this is done.

4.0 Conclusion

The Geophones, sometimes referred to as seismometers or detectors is the unit in direct contact with the earth that converts the motion of the earth resulting from the shot into electric signals. These signals constitutes the inputs into an instrumentals system, the end product of which is the presentation of subsurface geological information in some viable form, usually as a record section, which, except for distortion of scale, is comparable to geologic cross section.

5.0 Summary

Geophones are designed to react to motion of the earth in a given direction. Mechanical instruments record the amplified displacement of the ground; electromagnetic instruments respond to the velocity of ground motion. Depending on the designed, either type may respond to vertical or horizontal motion. Some modern electromagnetic instruments are constructed so as to record simultaneously three orthogonal components of motion. Most designs employ variations on the pendulum principle.

6.0 Tutor Marked Assignment:

Q1. Define the following

a) Seismometers
b) Hydrophones
c) Noise

Q2. Describe with the aid of a simple diagram Geophones.

Q3 Differentiate between a geophones and hydrophones.
Q4 Differentiate with examples between coherent noise and random noise

7.0 References/ Further Readings:

Kearey, P., Brooks, M. and Hill, I. (2002) An Introduction to GeophysicalExploration (Third Edition), Blackwell Science, Oxford, 262 pp.**John, M.** (2003) Field Geophysics (Third Edition). John Wiley and Sons Ltd. England, 249pp

McCann, D.M., Fenning, P. and Cripps, J. (Eds) (1995) Modern Geophysicsin Engineering Geology, Engineering Group of the Geological Society, London, 519 pp.

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Whitely, R.J. (Ed.) (1981) Geophysical Case Study of the Woodlawn Orebody, New South Wales, Australia, Pergamon Press, Oxford, 588 pp.

Hawkins, L.V. (1961) The reciprocal method of routine shallow seismic refraction investigations. Geophysics, **26**, 806–19.

Unit 4: Recording Seismic Signals

1.0 Introduction

Instruments that record seismic signals are known as *seismographs*. They range from timers which record only single events to complex units which digitize, filter and store signals from a number of detectors simultaneously.

2.0 Objectives

At the end of this unit, readers should be able to:

(i) Understand simple seismic recorders
(ii) Familiar with seismographs.
(iii) Understand common instruments used in seismic signals recording

3.0 Main Content:

1.4.1 Single-channel seismographs
Most single-channel seismographs have graphic displays, although rudimentary seismic 'timers' which simply displayed the arrival time of the first significant energy pulse numerically were once popular. On a visual display, the time range is switch or key-pad selected and the left-hand edge of the screen defines the shot or impact instant. Hard copy is not usually obtainable and times are measured directly. In some models a cursor can be moved across the screen while the time corresponding to its position is displayed. Noise levels can be monitored by observing the trace in the absence of a source pulse.

Modern single-channel instruments use enhancement principles. A digital version of the signal is stored in solid-state memory, as well as being displayed on the screen. A second signal can either replace this or be added to it. Any number n of signals can be summed (*stacked*) in this way for a theoretical $\sqrt{n}$ improvement in signal/noise ratio.

1.4.2 Multi-channel seismographs
Seismographs with 12 or 24 channels are generally used in shallow surveys, whereas a minimum of 48 channels is now the norm in deep reflection work. With multiple channels, both refraction and reflection work can be done and

explosives can reasonably be used since the cost per shot is less important when each shot produces many traces. Enhancement is used very widely and most instruments now provide graphic displays, optional hard copy and digital recording.

The enhancement seismographs now in use (Figure 1.9) are very sophisticated and versatile instruments. Display formats can be varied and individual traces can be selected for enhancement, replacement or preservation. Traces can be amplified after as well as before storage in memory, and time offsets can be used to display events that occur after long delay times. Digital recording has virtually eliminated the need for amplification before recording, because of the inherently very large *dynamic range* associated with storage of data as fixed precision numbers plus exponents (Example 1.2). Filters can also be applied, to reduce both high frequency random noise and also the long-period noise, of uncertain origin, that sometimes drives the traces from one or two geophones across the display, obscuring other traces.

Figure 1.9 *Enhancement seismographs. The instrument on the right is the now obsolete knob and switch controlled Geometrics 1210F. The instrument on the left is one of its successors, the Smartseis, which is entirely menu-driven Note the hard-copy record just emerging from the Smartseis, and the much greater size of the display 'window'.*

Example 1.2

Dynamic range is concerned with the range over which data can be recorded with roughly uniform *percentage* accuracies. When seismic amplitudes were recorded in ‘analogue’ form on magnetic tape, in which magnetization was proportional to signal strength, the dynamic range was limited at low amplitudes by tape noise and at high amplitudes by magnetic saturation. Automatic gain control (AGC) was therefore applied before recording, and inevitably distorted the signals. In digital systems, data are recorded as numerical values plus *exponents*, which are the powers of some other number by which the numerical value must be multiplied.

Thus, the values

46 789 and 0.0000046789

can be written in the familiar engineering notation, which uses powers of 10, as:

4.6789E +4 and 4.6789E − 6

The two quantities are thus recorded to the same percentage accuracy. In digital systems, data are usually recorded in binary formats and the exponent uses powers of 2. It is commonly allowed to range between −128 and +127, which is roughly equivalent to a range from 10^{-38} to 10^{+38}

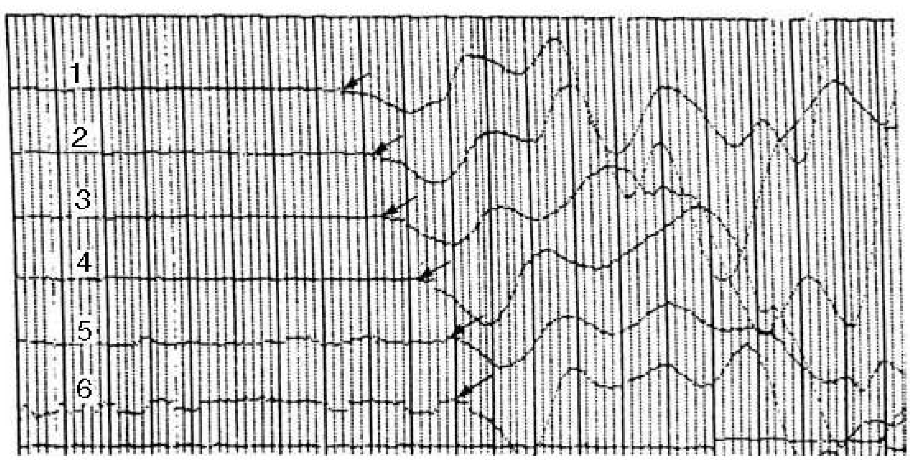

Figure 1.10 *Six-channel refraction record showing refraction ‘picks’. Noise prior to these picks is increasingly obvious as amplification is increased to record signals from the further geophones.*

The refraction survey example in Figure 1.10 shows the signals recorded by six geophones at points successively further from the source, with the traces from distant geophones amplified more to compensate for attenuation. Inevitably, amplifying the signal also amplifies the noise. In the field, arrival times can be estimated from the screen but this is never easy and seldom convenient. On the other hand, hard copies produced directly from the instrument are often of rather poor quality. This is especially true of dot-matrix outputs, because the matrix size causes irregularities in what should be smooth curves (Figure 1.10). Where these are the built-in printers, and assuming the instrument also has the capacity to store data digitally, it is worthwhile having a separate lap-top computer coupled to a reasonable printer at the field base. It would be foolhardy, however, not to produce, and preserve, the field hard-copy.
Powerful microcomputers are incorporated into most modern instruments, with high-capacity hard drives for data storage. Bewildering numbers of acquisition and processing options are available via menu-driven software. So versatile are these instruments that it is sometimes difficult, or at least time consuming, to persuade them to carry out routine, straightforward survey work.

4.0 Conclusion

Seismographs that allow signals to be displayed and summed are obviously superior to mere timers, and can be used to study events other than first arrivals. However, they are generally only useful in shallow refraction work since it is difficult to distinguish between direct waves, refractions and reflections on a single trace. Hammer sources are universal, since it would be expensive and inefficient to use an explosive charge to obtain such a small amount of data.

5.0 Summary

The seismograph

The earliest known instrument for indicating the arrival of seismic tremor from a distant source is reputed to have been invented by a Chinese astronomer called Chang Heng in 13AD. The device consisted of eight invented dragons placed at equal intervals around the rim of vase. Under each dragon sat an open-mouthed metal toad. Each dragon held a bronze ball in its mouth. When a slight tremor shook the device, an internal mechanism opened the mouth of one dragon, releasing its bronze balls, which fell into the open mouth of the metal toad beneath, thereby marking the direction of arrival of the tremor. The principle of this instrument was used in the 18th century European devices that consist of brimful bowl or mercury with grooved rims under which tiny collector bowl were placed to collect the overflow occasioned by a seismic tremor. These instruments gave viable evidence of a seismic event but were unable to trace a permanent record of the seismic wave itself. They are classified as seismoscopes.

6.0 Tutor Marked Assignment

Q1. What are seismographs?

Q2. List and Explain different types of seismograph you have learnt in this unit**.**

7.0 References/ Further Readings

Kearey, P., Brooks, M. and Hill, I. (2002) An Introduction toGeophysicalExploration (Third Edition), Blackwell Science, Oxford, 262 pp.
John, M. (2003) Field Geophysics (Third Edition). John Wiley and Sons Ltd. England, 249pp
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Hawkins, L.V. (1961) The reciprocal method of routine shallow seismic refraction investigations. Geophysics, **26**, 806–19.

MODULE 2

Unit 1 Seismic reflection

Unit 2 Reflection Survey

UNIT 1: SEISMIC REFLECTION

1.0 Introduction

The seismic reflection method absorbs more than 90% of the money spent world-wide on applied geophysics. Most surveys are aimed at defining oil bearing structures at depths of thousands of metres using hundreds or even thousands of detectors. However, some reflection work is done by small field crews probing to depths of, at most, a few hundred metres. The instruments used in these surveys were originally very simple but may now have as much in-built processing power as the massive processing laboratories of 20 years ago. Field operators need to have some understanding of the theory behind the options available.

2.0 Objectives

At the end of the unit, readers should be able to

(i) Give a descriptive treatment of the more important aspects, concentrating on the developing fundamental understanding of these methods and the physical principles on which they are based.
(ii) Know the instruments used in this seismic survey .
(iii) Have understanding of the theory behind seismic reflection.

3.0 Main Contents

3.1 Reflection Theory

Ray-path diagrams, as used previously, provide useful insights into the timing of reflection events but give no indication of amplitudes. 2.1.1 Reflection coefficients and acoustic impedances The *acoustic impedance* of a rock, usually denoted by **I**, is equal to its density multiplied by the seismic P-wave velocity. If a seismic wavefront strikes a planar interface between two rock layers with impedances $\mathbf{I}_1$ and $\mathbf{I}_2$ at right angles (*normal incidence*), the amplitude of the reflected wave, as a percentage of the amplitude of the incident wave (the *reflection coefficient*, *RC*) is given by:

$$RC = (\mathbf{I}_2 - \mathbf{I}_1) / (\mathbf{I}_2 + \mathbf{I}_1)$$

If $\mathbf{I}_1$ is greater than $\mathbf{I}_2$, the coefficient is negative and the wave is reflected with phase reversed, i.e. a negative pulse will be returned where a positive pulse was transmitted and vice versa. The amount of energy reflected first decreases and then increases as the angle of incidence increases. If the velocity is greater in the second medium than in the first, there is ultimately total reflection and no transmitted wave (Section 1.1.5). However, most small-scale surveys use waves reflected at nearly normal incidence.

3.2 Normal moveout

The true normal-incidence ray cannot be used in survey work, since a detector at the shot point would probably be damaged and would certainly be set into such violent oscillation that the whole record would be unusable. Geophones

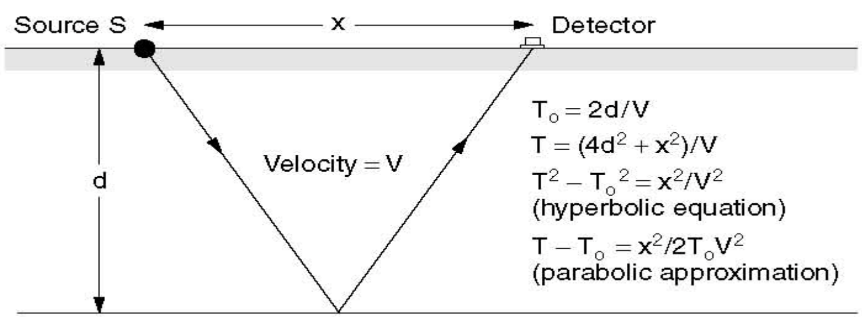

Figure 2.1 *Derivation of the normal moveout equation for a horizontal reflector.* T_o *is the normal incidence time*.

are therefore offset from sources and geometric corrections must be made to travel times.

Figure 2.1 shows reflection from a horizontal interface, depth *d*, to a geophone at a distance *x* from the source. The exact *hyperbolic* equation linking the travel time *T* and the normal incidence time T_o is established by application of the Pythagoras theorem. For small offsets, the exact equation can be replaced by the *parabolic* approximation, which gives the *normal moveout* (NMO), $T - T_o$, directly as a function of velocity, reflection time and offset.

$$T - T_o = x^2/2V^2T_o$$

Since V usually increases with depth and To always does, NMO decreases (i.e. NMO curves flatten) with depth. Curved alignments of reflection events can be seen on many multi-channel records (Figure 2.2). Curvature is the most reliable way of distinguishing shallow reflections from refractions.

3.3 Dix velocity

If there are several different layers above a reflector, the NMO equation will give the 'root-mean-square' (*RMS*) velocity defined as:

$$V^2_{RMS} = (V^2_1 t_1 + V^2_2 t_2 \cdots + \cdots V^2_n t_n)/\mathbf{T_n}$$

where t_n is the transit time through the nth layer, velocity V_n, and $\mathbf{T_n}$ is the total transit time to the base of the nth layer. Interval velocities can be calculated from RMS velocities using the *Dix formula:*

$$V^2_{DIX} = (V^2_{n-1}\mathbf{T}_{n-1} - V^2_n \mathbf{T}_n)/(\mathbf{T}_{n-1} - \mathbf{T}_n)$$

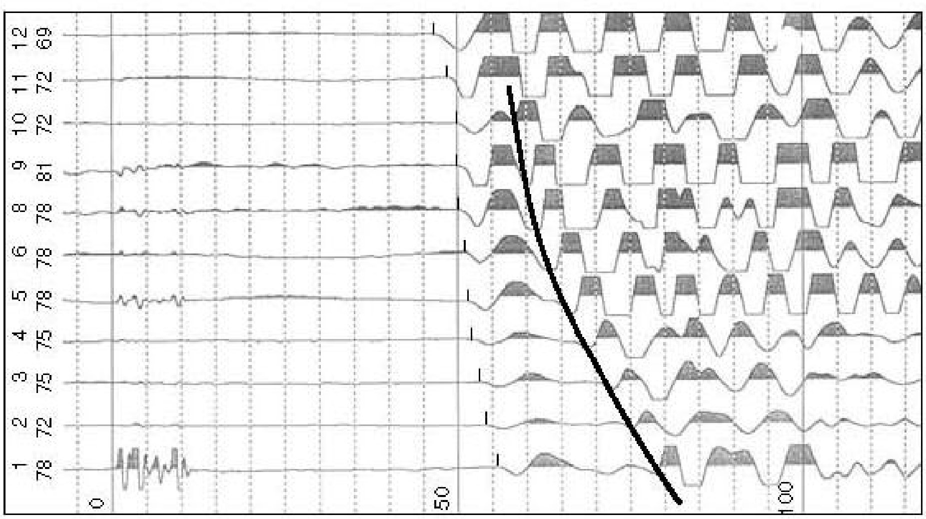

***Figure 2.2** Enhancement seismograph record showing curved alignment of reflections (thick line). The earlier events were produced by refractions. Note that on Channels 11 and 12 the strong refracted wave completely overwrites the reflection. The variable area presentation used is popular for reflection work since it emphasizes trace-to-trace correlations, although some information is lost where traces overlap.*

The subscripts $n-1$ and n denote, respectively, the top and bottom of the nth layer. *RMS* velocities are normally slightly higher than true average velocities, since squaring the high velocities increases their influence on the average. Significant errors can arise if RMS velocities are used directly to make depth estimates but these are generally less than the errors introduced by the use of the NMO equation to estimate velocity using reflections from interfaces that may well not be horizontal. Dix conversion may not help very much in these cases.

3.4 Effect of dip

If the source is placed at the centre of the geophone spread, the curves obtained over horizontal interfaces will be symmetrical about the source point. If, however, the reflector has a uniform dip α, the reduction in travel path on the up-dip side of the shot compensates to some extent for the offset, and some travel times will be less than the normal-incidence time (Figure 2.3). The minimum time $2d\cdot \cos(\alpha)/V$ is recorded at a distance $2d\cdot \sin(\alpha)$ from the shot, on the up-dip side. The reflected ray rises vertically to this point, about which the move-out curve is symmetrical. Dip effects in shallow reflection surveys are detectable only for very large dips or very long spreads.

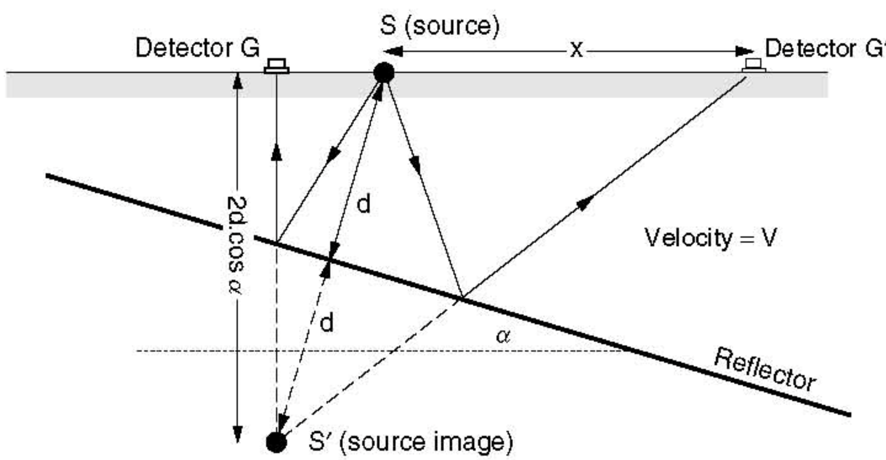

Figure 2.3 *Effect of dip on a single-fold record. Rays are reflected from the dipping interface as if derived from the image point S_ at depth* 2*d*· cos *α below the surface, where d is the perpendicular distance from the shot-point to the interface. The normal incidence travel time is* 2*d/V but the shortest travel time is for the ray which is vertical after reflection. An identical move-out hyperbola would be produced by a shot at point G and a horizontal interface at depth d·* cos *α*.

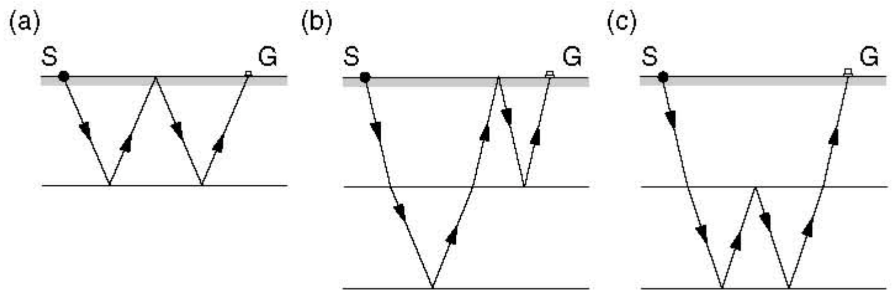

Figure 2.4 *Paths for multiple reflections. (a) Simple multiple. (b) Peg-leg. (c) Intra-formational multiple*.

3.5 Multiple reflections

A wave reflected upwards with high amplitude from a subsurface interface can be reflected down again from the ground surface and then back from the same interface. This is a simple *multiple*. Two strong reflectors can generate *peg-leg* and *intraformational* multiples (Figure 2.4). Multiples are difficult to identify with certainty on single traces. They can sometimes be recognized on multi-channel records because they have move-outs appropriate to shallow reflectors and simple time relationships with their primaries.

4.0 Conclusion

This method is by far the most widely used geophysical technique - the struc ture of the subsurface formations is mapped by measuring the times required for a seismic wave (or pulse), generated in the earth by a near-surface explosion, mechanical impact, or vibration, to return to the surface after reflection from interfaces between formations having different physical properties. The reflections are recorded by detecting instruments responsive to ground motion. They are laid along the ground at distances from the point of generation, which are generally small compared with the depth of the reflector. Variations in the reflection times from place to place on the surface usually indicate structural features in the strata below. Depth to reflecting interfaces can be estimated from the recorded times and velocity information that can be obtained either from the reflected signals themselves or from surveys in wells. Reflections from depths of 30, 000 ft or more can normally be observed by combining the reflections from the repeated source applications, so in most areas geologic structure can be determined throughout the sedimentary section.

5.0 Summary

Reflection data can be used to determine the average velocities of seismic waves between the surface and the reflector. More important from a geological viewpoint, the velocities of seismic waves through depth intervals of a few percent of depth from the surface can now be obtained and often provide a good indication of lithology. The usefulness of such information depends on the layering as well as on the problem at hand.

With reflection methods, one can locate and map such features as anticlines, faults, salt domes, and reef. Many of these are associated with the accumulation of oil and gas. Major convergences caused by depositional thinning can be detected from reflection sections. The resolution of the method is now approaching a fineness adequate for finding stratigraphic traps such as pinch outs or faces changes. However, successful exploration for stratigraphic oil accumulations by reflection techniques requires skilful coordination of geological and seismic information.

6.0Tutor Marked Assignment

Q1.

a) What is seismic reflection?
b) What do you understand by the term *acoustic impedance of a rock?*

Q2. Explain with the aid of diagram the following

a) Multiple reflection
b) Simple multiple
c) Peg leg
d) Intraformational multiples

Q3 State the dix formular and define all the parameters and variable.

Q4. What is normal moveout? Derive a moveout equation for a horizontal reflector.

7.0References/ Further Readings

Kearey, P., Brooks, M. and Hill, I. (2002) An Introduction to Geophysical Exploration (Third Edition), Blackwell Science, Oxford, 262 pp.
John, M. (2003) Field Geophysics (Third Edition). John Wiley and Sons Ltd. England, 249pp

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Telford, W.M., Geldart, L.P., Sheriff, R.E. and Keys, D.A. (1990) Applied Geophysics (Second Edition), Cambridge University Press, Cambridge, 770 pp.
Whitely, R.J. (Ed.) (1981) Geophysical Case Study of the Woodlawn Orebody, New South Wales, Australia, Pergamon Press, Oxford, 588 pp.
Hawkins, L.V. (1961) The reciprocal method of routine shallow seismic refraction investigations. Geophysics, **26**, 806–19.

Unit 2 Reflection Surveys

1.0 Introduction

Reflected waves are never first arrivals, so clear-cut reflection events are seldom seen. Oil-industry techniques for improving signal-to-noise ratios can be used for shallow work and simple versions of the programs used are incorporated in the software supplied with the latest generation of 12- and 24-channel seismographs.

2.0 Objectives

At the end of the unit, readers should be able to

(i) Give a descriptive treatment of the more important aspects, concentrating on the developing fundamental of reflection surveys.

(ii) Know that reflection survey techniques are therefore widely used in oil industry for exploring oil.

3.0 Main Contents

3.1 Spread lengths

The distance from the source to the nearest geophone in a shallow reflection survey is usually dictated by the strength of the source (and the need to protect the geophone) and may be as little as 2 m when a hammer is being used. Even with explosives or heavy weight drops, minimum offsets of more than about 10 m are unusual when observing shallow reflections. A reflection spread can be much shorter than a refraction spread used to probe to similar depths, but with powerful sources and multi-channel recording, the furthest geophone may be more than 100 m from the source. The optimum spread length can be determined only by experiment, since the most important factors are the arrival times of the noise trains associated with the direct wave and any strong refracted waves. Field work should begin with tests specifically designed to examine these arrivals, generally by using elongated spreads.

3.2 Arrays

Ideally, reflected energy should arrive after the near-surface waves (ground-roll and refractions) have passed but this may not be possible if the depth of investigation is very small. In such cases, geophones may be connected in arrays to each recording channel. Reflected waves, which travel almost vertically, will reach all the geophones in an array almost simultaneously but the direct waves will arrive at different times and produce signals that can interfere destructively.

The efficiency with which a wave is attenuated by an array is defined by its *relative effect* (RE) compared to the effect of the same number of geophones placed together at the array centre. The variation of the RE with *apparent wavelength* (which for the direct wave is equal to the true wavelength), for a linear array of five geophones equally spaced on a line directed towards the shot point, is shown in Figure 2.5. Non-linear arrays produce more complex curves.

Simple arrays are preferred in the field, since mistakes are easily made in setting out complicated ones. The range of frequencies over which attenuation of the direct wave occurs is proportional to array length and it may be necessary to overlap the geophones in adjacent arrays. It would be unusual in a shallow survey to use more than five geophones per array.

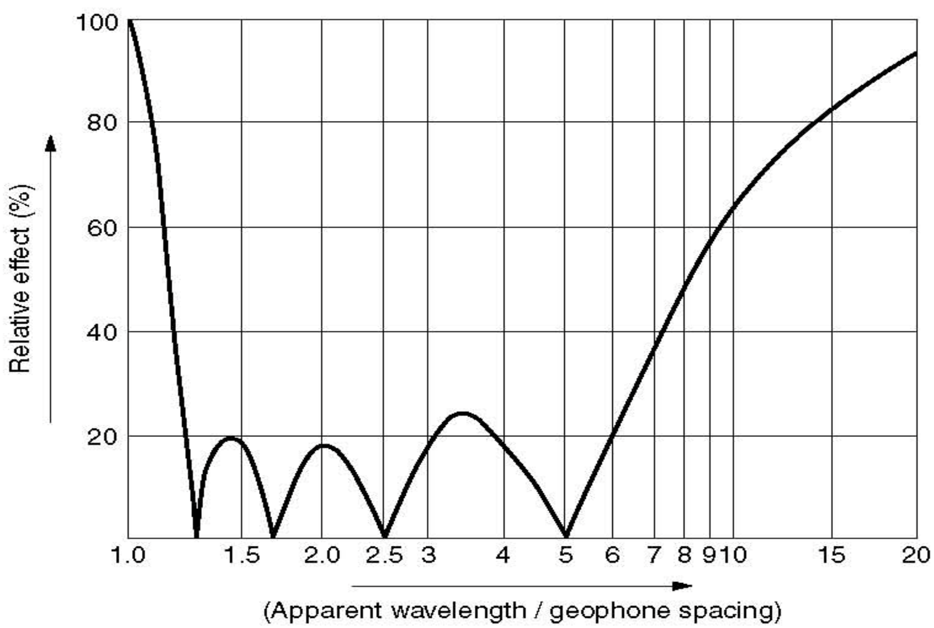

Figure 2.5 *Relative effect (RE) of an array of five equi-spaced in-line geophones. The 100% level would be attained with zero spacing between the geophones. The apparent wavelength is equal to the actual wavelength divided by the sine of the angle between the wavefront and the ground surface, and is infinite for a wave rising vertically and equal to the true wavelength for the direct wave. Attenuation is concentrated between values of apparent wavelength*

divided by geophone spacing of about 1.2 and 7. With 2 m spacing, a 500 m·s−1 *wave would be attenuated at frequencies of between about 35 and 200 Hz.*

3.3 Shot arrays

Seismic cables for use with only 12 or 24 channels are not designed with arrays in mind, and non-standard connectors may have to be fabricated to link the geophones to each other and to the cable. It may be easier to use arrays of shots instead.

A shot array using explosives usually involves simultaneous detonation of charges laid out in a pattern resembling that of a conventional geophone array. If an impact source is used with an enhancement instrument, the same effect can be obtained by adding together results obtained with the impact at different points. This is the simplest way of reducing the effects of surface waves when using a hammer.

3.4 Common mid-point shooting

Improving signal-to-noise ratios by adding together several traces *(stacking)* is fundamental to deep reflection surveys. In shallow surveys this technique is normally used only to stack (enhance) results obtained with identical source and detector positions. If, however, the data are recorded digitally, NMO corrections can be made (although not in the field) to traces produced with different source–receiver combinations. The technique normally used is to collect together a number of traces that have the same mid-point between source and receiver (*common midpoint* or CMP traces), apply the corrections and then stack.

The number of traces gathered together in a CMP stack defines the *fold* of coverage. Three traces forming a single synthetic zero-offset trace constitute a 3-fold stack and are said to provide *300% cover*. The maximum fold obtainable, unless the shot point and geophone line are moved together by fractions of a geophone interval, is equal to half the number of data channels.

Figure 2.6 shows the successive geophone and source positions when a six-channel instrument is used to obtain 300% cover. Special cables and switching circuits are available for use in deep reflection surveys, but CMP fieldwork with the instruments used for shallow surveys is very slow and laborious. The need to combine traces from several different shots makes it difficult to do CMP processing in the field.

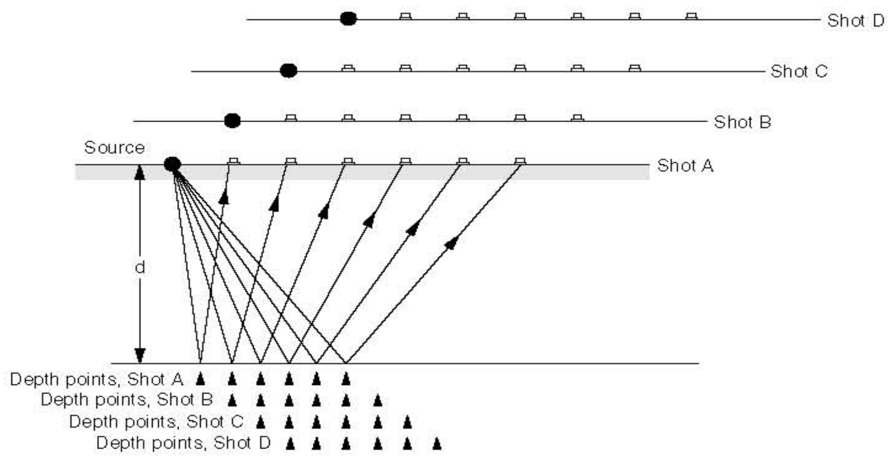

Figure 2.6 *CMP schematic, for 3-fold cover with a 6-channel system. Shot points A, B, C and D are progressively one geophone group interval further to the right. Note that the distance between reflection points (depth points) on the interface is only half that between geophone groups on the surface. Shots A and D have no depth points in common.*

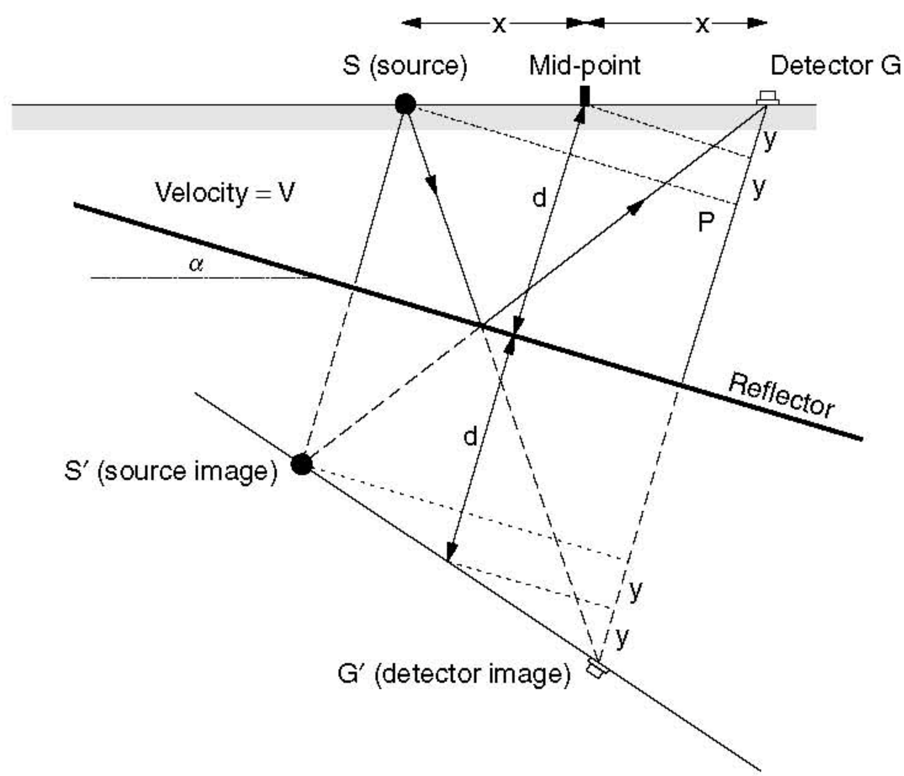

Figure 2.7 *Effect of dip in CMP shooting. In contrast to single-fold shooting (Figure 2.3), the shot points as well as the geophone locations are different for the different traces. Shot points and detector locations are equivalent and the 'depth point' on the reflector moves up dip as the offset increases. The move-out equation is most easily derived by noting that the path from source to detector is equal in length to the path* SG^{I} *from the source to the detector 'image point', and that the geometric relationships between similar triangles imply the equality of all the lengths marked 'y'. The Pythagoras relationship can be applied to the triangle* $SG^{I}P$*, and the times can be obtained by dividing the distances by V . Thus,* $T_o = 2d/V$ *and* $T = SG^{I}/V$.

The geometry of a CMP shoot differs from that for single-fold coverage, and the effect of dip is therefore different (Figure 2.7). If the interface dips at an angle α, the velocity deduced from a CMP stack is equal to $V/\cos\alpha$ and the 'depth' is equal to the length of the normal incidence ray from the common mid-point to the interface. In contrast to the single-fold gather of Section 2.1.4, the minimum time is associated with the normal incidence ray. The aim of stacking is to produce a noise-reduced seismic trace that approximates to the normal incidence trace, i.e. to the trace that would have been produced had the source and detector been coincident. The initials CMP replaced an earlier acronym, CDP (*common depth point*) used for the same method. Referring to depth points (reflection points) as 'common' implies that all the reflections in a gather have come from the same point on the subsurface interface, which is true only for horizontal interfaces.

3.5 Depth conversion

Reflection events are recorded not in depth but in *two-way time* (TWT). Velocities are needed to convert times into depths, but the Dix velocities (Section 2.1.3) obtained from NMO curves may be 10–20% in error, even

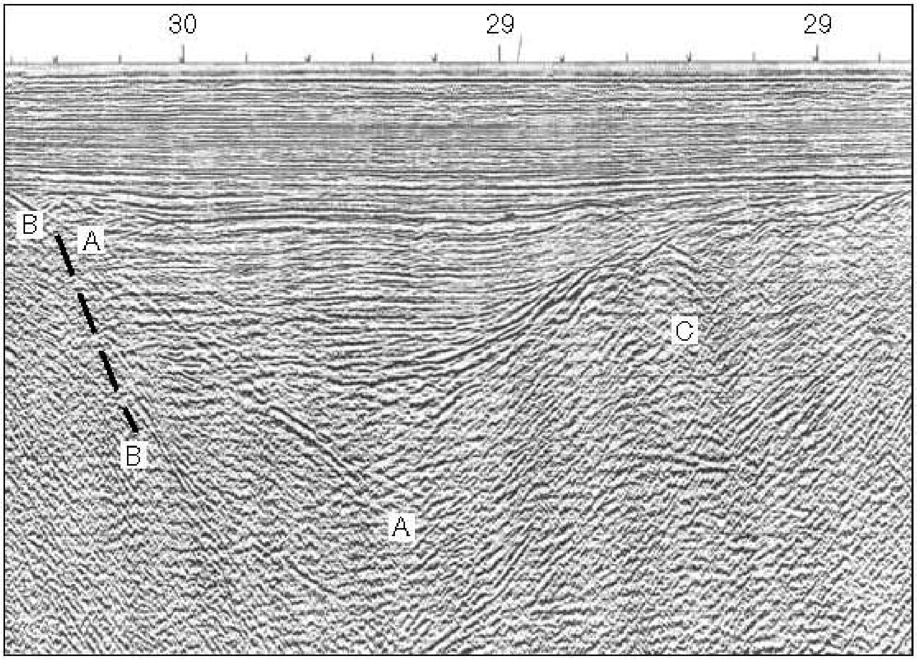

Figure 2.8 *Geometric distortion on seismic sections. The image is of a small graben structure beneath an unconformity. The position of the true fault plane BB (indicated by the dashed line) can be estimated from the positions of the terminations of the sub-horizontal reflectors representing the sediment fill within the graben (although care must be exercised because many of the deeper sub-horizontal events are multiples). The event AA is the seismic image of BB. It is displaced because the techniques used to display the data assume that reflections are generated from points vertically beneath the surface points, whereas they are actually generated by normal-incidence rays that are inclined to the vertical if reflected from dipping interfaces. The reflections from the fault and the opposite side of the graben cross over near the lower symbol 'A', forming a 'bow-tie'. Convex-upward reflections near point C are diffraction patterns generated by faulting.*

for horizontal reflectors. Interpretations should be calibrated against borehole data wherever possible, and field crews should always be on the lookout for opportunities to measure vertical velocities directly.

3.6 Geometric distortion

Seismic reflection data are normally presented as sections prepared by playing out, next to each other and vertically down the sheet of paper, the traces from adjacent CMP gathers. Such sections are subject to geometric distortion. Artefacts such as displaced reflectors, diffraction patterns and 'bow-ties', described in Section 3.2 as affecting radar sections, also appear on seismic imagery, as is shown in Figure 2.8. Refraction surveys are widely used to study the water table and, for engineering purposes, the poorly consolidated layers near the ground surface, and also in determining near-surface corrections for deep reflection traces. Travel times are usually only a few tens of milliseconds and there is little separation between arrivals of different types of wave or of waves that have travelled by different paths. Only the first arrivals, which are always of a P wave, can be ‘picked’ with any confidence.

4.0 Conclusion

In recent years, reflection data have also been for identifying lithology, generally from velocity and attenuation characteristics of the transmitted and reflected seismic waves, and for detecting hydrocarbons, primarily gas, directly on the basis of reflection amplitudes and other seismic indicators.

The reflection method comes closer than any other prospecting technique to providing a structural picture of the subsurface comparable to what could be obtained from a great number of boreholes in close proximity. Modern reflection record sections are similar in appearance to geologic cross sections, and geologists must sometimes be cautioned not to use them as such without taking into consideration some potential hazards that might lead to erroneous interpretation, even with good-quality reflection data. Under ideal conditions, structural relief can be determined with a precision of about ½ percent of depth below the surface. This method makes it possible to produce structural maps of any geologic horizons that yield reflections, but the horizons themselves usually cannot be identified without independent geological information such as might be obtained from wells.

This method makes it possible to produce structural maps of any geologic horizons that yield reflections, but the horizons themselves usually cannot be identified without independent geological information such as might be obtained from wells.

5.0 Summary

While current technological improvements have made it possible to obtain usable reflection data in many areas where reflections were formerly too poor to map, there are still places where reflection does not yield reliable information even though highly sophisticated data acquisition and processing techniques are used. In such intractable areas, other geophysical and geological methods must be employed

6.0Tutor Marked Assignment

Q1.Explain the following terminologies

a) Reflection Survey
b) Geometric distortion in relation to reflection survey.

Q2 Write short notes on the following in relation to reflection

a) Spread length
b) CMS - Common mid-point Shooting
c) Array
d) Depth Conversion

7.0References/ Further Readings

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MODULE 3

Unit 1 Seismic Refraction

Unit 2 Field Interpretation

Unit 3 Limitation of Refraction Methods

UNIT 1: SEISMIC REFRACTION

1.0 Introduction

Large primary voltages are needed to produce measurable IP effects. Current electrodes can be plan metal stakes but non- polarising electrodes must be used to detect the few millivolts of transient signal.

2.0 Objectives

At the end of the unit, readers should be able to;

(i) Understand the importance of power source in seismic refraction.
(ii) Know that important fundamentals in refraction seismic refraction.

3.0 Main Contents

3.1 Refraction Surveys

Ideally the interfaces studied in a small refraction survey should be shallow, roughly planar and dip at less than 15◦. Velocity must increase with depth at each interface. The first arrivals at the surface will then come from successively deeper interfaces as distance from the shot point increases.

3.1.1 The principal refractors

P-wave velocities for common rocks were shown in Figure 1.1. In shallow refraction work it is often sufficient to consider the ground in terms of dry overburden, wet overburden and weathered and fresh bedrock. It is very difficult to deal with more than three interfaces. The P-wave velocity of dry overburden is sometimes as low as 350 ms−1, the velocity of sound in air, and is seldom more than 800 ms−1. There is usually a slow increase with depth, which is almost impossible to measure, followed by an abrupt increase to 1500–1800 ms−1 at the water table. Fresh bedrock generally has a P-wave velocity of more than 2500 ms^{-1} but is likely to be overlain by a transitional weathered layer where the velocity, which may initially be less than 2000 ms^{-1}, usually increases steadily with depth and the accompanying reduction in weathering.

3.1.2 Critical refraction and the head wave Snell's law

This implies that if, in Figure 1.2, $V2$ is greater than $V1$ and if sin $i = V1/V2$, the refracted ray will travel parallel to the interface at velocity $V2$. This is *critical refraction*. After critical refraction, some energy will return to the ground surface as a *head wave* represented by rays which leave the interface at the critical angle. The head wave travels through the upper layer at velocity $V1$ but, because of its inclination, appears to move across the ground at the $V2$ velocity with which the wave-front expands below the interface. It will therefore eventually overtake the direct wave, despite the longer travel path. The *cross-over* or *critical* distance for which the travel times of the direct and refracted waves are equal is:

$$xc = 2d\,[(V2 + V1)/(V2 - V1)]$$

This equation forms the basis of a simple method of refraction interpretation.
xc is always more than double the interface depth and is large if the depth is large or the difference in velocities is small. The *critical time*, obtained by dividing the critical distance by the direct-wave velocity, is also sometimes used.

The term 'critical distance' is also sometimes used for the minimum distance at which refractions return to the surface, i.e. the distance from the shot point at which energy arrives after reflection at the critical angle. This usage is not common amongst field crews since the refractions arrive after the direct wave at this point, and for some distance beyond, and are difficult to observe. If more than one interface is involved (as in Figure 3.1), the ray that is critically refracted at the lowermost interface leaves the ground surface at an angle in given by:

$$\sin in = V1/Vn$$

Thus, the angle at which energy leaves the ground surface for ultimate critical refraction at a deep interface depends only on the velocities in the uppermost

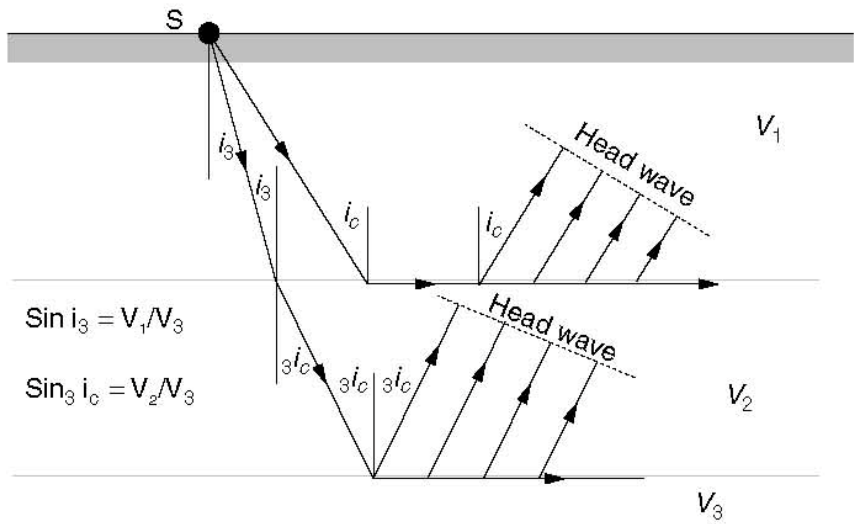

Figure 3.1 *Critical refraction at two interfaces: sin ic = V*1/*V*2.

and lowermost layers involved, and not on the velocities in between. Eventhough this is a surprisingly simple result, cross-over interpretation becomes rather complicated for multiple layers and the intercept-time method discussed below (Section 3.2) is generally preferred.

3.1.3 Lengths of refraction spreads

A line of geophones laid out for a refraction survey is known as a *spread*, the term *array* being reserved for geophones feeding a single recording channel. Arrays are common in reflection work but are almost unknown in refraction surveys where the sharpest possible arrivals are needed. Sufficient information on the direct wave and reasonable coverage of the refractor is obtained if the length of the spread is about three times the crossover distance. A simple but often inaccurate rule of thumb states that the spread length should be eight times the expected refractor depth.

3.1.4 Positioning shots

In most refraction surveys, *short shots* are fired very close to the ends of the spread. Interpretation is simplified if these shots are actually at the end geophone positions so that travel times between shot points are recorded directly. If this system is used, the geophone normally at the short shot location should be moved half-way towards the next in line before the shot is actually fired (and replaced afterwards). Damage to the geophone is avoided and some extra information is obtained on the direct wave. *Long shots* are placed sufficiently far from the spread for all first arrivals to have come via the refractor, and short-shot data may therefore be needed before long-shot offsets can be decided. Distances to long shots need be measured accurately only if continuous coverage is being obtained and the long-shot to one spread is to be in the same place as a short or centre shot to another. If explosives are being used, it may be worthwhile using a very long offset if this will allow firing in water (see Section 1.2.1).

3.1.5 Centre shots

The information provided by a conventional four-shot pattern may be supplemented by a centre shot. Centre shots are especially useful if there are considerable differences in interpretation at opposite ends of the spread, and especially if these seem to imply different numbers of refractors. They may make it possible to obtain a more reliable estimate of the velocity along an intermediate refractor or to monitor the thinning of an intermediate layer that is hidden, at one end of the spread, by refractions from greater depths. An additional reliable depth estimate is obtained that does not depend on assumptions about the ways in which the thicknesses of the various layers vary along the spread, and there will be extra data on the direct wave velocity. Centre shots are used less than they deserve. The extra effort is generally trivial compared to the work done in laying out the spread, and the additional and possibly vital information is cheaply obtained.

3.1.6 Annotation of field records

Hard-copy records can be (and should be) produced in the field from most of the seismographs now used for shallow refraction surveys. The several dozen records produced in a day's work that includes repeats, checks and tests as well as the completion of a number of different spreads must be carefully annotated if confusion is to be avoided. Annotations should obviously include the date and the name of the observer-in-charge, along with the survey location and spread number. Orientation should be noted, and the position of Geophone 1 should be defined. Unless the geophone spacing is absolutely uniform, a sketch showing shot and geophone locations should be added. If the interval between timing lines on the records can be varied and/or variable time offsets can be applied, these settings must also be noted. In many cases this is now done automatically.
Other items are optional. Amplifier gains and filter settings are not often recorded but such information may be useful. The number of shots or impacts combined in a single record can also be important with enhancement instruments.

And, of course, features such as the use of S-wave geophones at some points or peculiarities in the locations of some of the geophones should always be noted. Many of the items listed above can be printed directly on to the hard-copy record, provided they are first entered into the machine. This is often a more tedious, and more error-prone, process than simply writing the information on each record by hand.

3.1.7 Picking refraction arrivals

Picking first arrivals on refraction records relies on subjective estimates of first break positions (Figure 3.2) and may be difficult at remote geophones where the signal-to-noise ratio is poor. Some of the later peaks and troughs in the same wave train are likely to be stronger, and it is sometimes possible to work back from these to estimate the position of the first break. However, because high frequencies are selectively absorbed in the ground, the distance between the first break and any later peak gradually increases with increasing distance from the source. Furthermore, the trace beyond the first break is affected by many other arrivals as well as by later parts of the primary wave train, and these will modify peak and trough locations. Using later features to estimate first-arrival times should always be regarded as a poor substitute for direct picking.

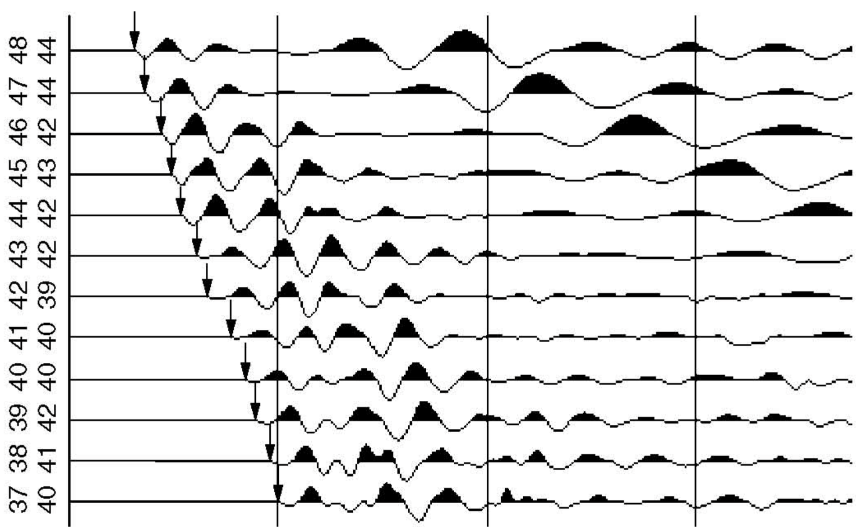

Figure 3.2 *Portion of a multi-channel refraction record, with first-break 'picks' identified by arrows. These become more difficult to make as signal to-noise ratio worsens. This record would be considered good, and much more difficult decisions usually have to be made*.

3.1.8 Time–distance plots

The data extracted from a refraction survey consist of sets of times (usually first-arrival times) measured at geophones at various distances from the source positions. Since these are plotted against vertical time axes and horizontal distance axes, the gradient of any line is equal to the reciprocal of a velocity, i.e.

steep slopes correspond to slow velocities. All the data for a spread are plotted on a single sheet that has a working area covering only the ground where there are actually geophones (see Figure 3.8 accompanying Example 3.1). It is not necessary to show the long-shot positions. Since as many as five sets of arrivals may have to be plotted, as well as a set of time differences, different colours or symbols are needed to distinguish between data sets. If the arrival times lie on a number of clearly defined straight-line segments, best-fit lines may be drawn. These are not actually necessary if the intercept-time interpretation method described below is used, and will be difficult to define if the arrival times are irregular because of variations in refractor depth. It is often best to draw lines through only the direct-wave arrivals (which should plot on straight lines), leaving refracted arrivals either un-joined or linked only by faint lines between adjacent points.

4.0 Conclusion

In reflection surveying, the detecting instruments record seismic signals at a distance from the shot point that is large compared with the depth of the horizon to be mapped. The seismic waves must thus travel large horizontal distances through the earth, and the times required for the travel at various source-receiver distances give information on the velocities and depths of the subsurface formations along which they propagate. Although the refraction method does not give as much information or as precise and unambiguous a structural picture as reflection, it provides data on the velocity of the refracting beds. The method made it possible to cover a given area more quickly and economically than with the reflection method, though with a significant loss of detail and accuracy.

Despite its disadvantages, refraction is now rarely employed in oil exploration because of the larger-scale field operations required. Also, the reflection method has developed to the point that it can now yield nearly all of the information that refraction shooting could produce as well as relatively unambiguous and precise structural information unavailable from refracted waves.

5.0 Summary

Refraction is particularly suitable where the structure of a high-speed surface, such as the basement or top of a limestone layer, is the target of geological interest. If the problem is to determine the depth and shape of a sedimentary basin by mapping the basement surface, and if the sedimentary rocks have a consistently lower seismic velocity than do the basement formations, refraction was in the past an effective and economical approach for achieving this objective. Airborne magnetics and, to some extent, gravity have replaced seismic refraction for such purposes. Because velocities in salt and evaporites are often greater than in surrounding formations, refraction has been useful in mapping diapiric features such as salt domes. Under favourable circumstances, this technique has been used to detect and determine the throw of faults in high-speed formations, such as dense limestone and basement materials.

6.0 Tutor Marked Assignment

Q1. What do you understand by the Principal Refractors

Q2. Explain with the aid of a diagram critical refraction at two interfaces.

Q3. Write short notes on the following

a) Critical distance
b) Annotation of field records
c) Length of refraction spread
d) A spread

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UNIT 2 FIELD INTERPRETATION

1.0 Introduction

Interpretation is an essential part of refraction fieldwork because the success of a survey depends on parameters such as line orientation, geophone spacing, shot positions and spread lengths, that can be varied almost at will. Only if analysis keeps pace with data collection will the right choices be made. Field interpretation has been made easier by computer programs that can be implemented on portable PCs or on the seismographs themselves but such programs are based on very simple models and are no substitute for actually thinking about the data.

2.0 Objectives

At the end of the unit, readers should be able to

(i) Give a descriptive interpretations of the more important field aspects
(ii) show prominent practical interpretations of geophysical methods .
(iii) Practicalised seismic refraction techniques.

3.0 Main Contents

3.1.1 Intercept times

The intercept time *t*i is defined as the time at which the back-extrapolated refracted arrival line cuts the time axis. For a single refractor, it is related to the velocities and the refractor depth by the equation:

$$t_i = 2d\sqrt{(V_2^2 - V_1^2)}/V_1V_2 = 2d/V_{1,2}$$

The quantity $V_{1,2}$ is defined by this equation. It has the units of a velocity and is approximately equal to V_1 if V_2 is very much larger than V_1. The critical angle is then almost 90 ◦ and the delay suffered by the refracted ray in travelling between the surface and the refractor is close to double the vertical travel time. If the difference between V_1 and V_2 is small, $V_{1,2}$ can be very large. Intercept times are conventionally obtained by drawing best-fit lines through the refracted arrival times but even a very good fit is no guarantee that the depth of the refractor does not change in the region near the shot point, from which no refractions are observed. If, however, a long shot is used, there should be a constant difference between long-shot and short-shot arrival times at points towards the far end of the spread (Figure 3.3). An intercept time can then be obtained by subtracting this difference from the long-shot arrival time at the

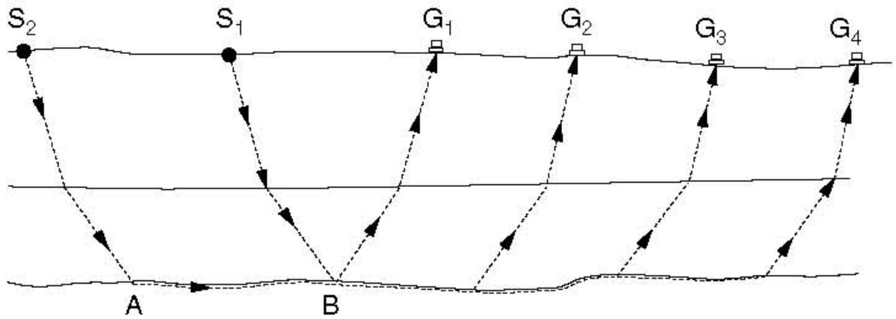

Figure 3.3 *Long-shot and short-shot travel paths for a three-layer case. The paths for energy travelling to the geophones from S*1 *and S*2 *via the lower refractor are identical from point B onwards. Upper refractor paths have been omitted for clarity*.

short-shot location, and this can be done exactly if there is a geophone in this position when the long shot is fired (Example 3.1). Otherwise, use of the nearest long-shot arrival at least reduces the distance over which extrapolation must be made.

3.2 Multiple layers

The intercept-time equation can be extended to cases involving a number of critically refracting layers. If the intercept time associated with the nth refractor is t_n, then:

$$t_n = 2d_1/V_{1,n+1} + 2d_2/V_{2,n+1} \cdot \cdot \cdot + \cdot \cdot \cdot 2d_n/V_{n,n+1}$$

where d_n is the thickness of the nth layer, which overlies the nth refracting interface, at which velocity increases from V_n to V_{n+1}. The definition of the various quantities $V_{m,n}$ is exactly analogous to the definition of $V_{1,2}$ citedabove. The presence of intermediate layers may be recognized by comparinglong- and short-shot data, but at least two points are needed to define a velocityand three for any confidence to be placed in the estimate. At best, therefore,only four layers can be easily investigated with a 12-channel system. Complicated field procedures can be devised to overcome this limitation; geophones may, for example, be moved one half-interval after a shot has been fired and the same shot-point can then be reused. Progress is extremely slow and the problems presented by refractor topography, hidden layers and blind zones (Section 3.3) still exist. In most circumstances, firing multiple shots into modified spreads represents an attempt to extract more from the method than is really obtainable.

3.3 Effect of dip

Depths estimated from refraction surveys are related to geophone and shot point elevations, which must therefore be measured to obtain a true picture of the subsurface refractor. Furthermore, the 'depths' determined are the perpendicular, not the vertical, distances to interfaces from shot points or geophones. With this proviso, 'horizontal' formulae can be applied without modification wherever the ground surface and the refractor are parallel. More usually their slopes will be different. Formulae are then most commonly quoted in terms of a horizontal ground and dipping refractors, but can equally well be applied if the ground slopes above, for example, a horizontal water table.
The intercept-time equations require the true value of $V2$ to be used. However, a wave that travels down-dip not only has to travel further at velocity $V2$ to reach more distant geophones, but also further at the slow velocity $V1$ in the upper layer (Figure 3.4). It therefore arrives late, with a low *apparent velocity*. The reverse is true shooting up-dip, when arrivals at further geophones may actually precede those at nearer ones. The slope of the line through the refracted arrivals on a time–distance plot depends on the dip angle, α, according to:

$$V\text{app} = V2/(1 + \sin \alpha)$$

If shots are fired from both ends of the spread, different apparent velocities will be measured because the sign of the dip angle will differ. For dips of less than about 10°, the true velocity is given by the dip-velocity equation:

$$2/V2 = 1/V\text{up} + 1/V\text{down}$$

3.4 Refractor relief and true velocities

Most refractors, except the water table, are irregular. If there were only a single local depression in an otherwise flat refractor, refracted arrivals from shots in opposite directions would plot on straight lines of equal slope, and the differences between the two arrival times at each geophone would plot on a line with double this slope. The exception to this rule would seem to be the geophone immediately above the depression.

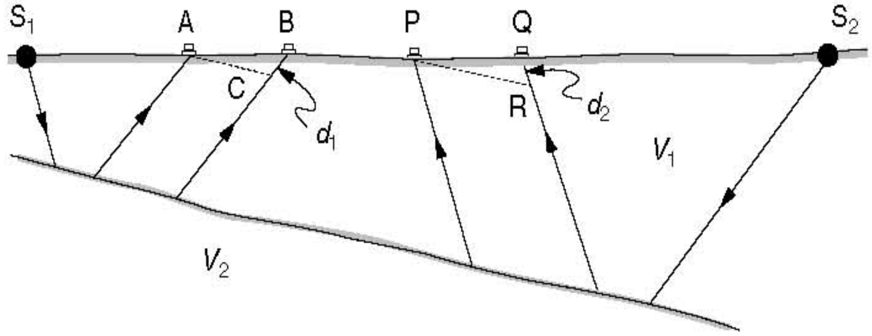

Figure 3.4 *Refraction at a dipping interface. The refracted energy from S*1 *arrives later at B than at A not only because of the greater distance travelled along the refractor but also because of the extra distance* d1 *travelled in the low velocity layer. Energy from S*2 *arrives earlier at P than would be predicted from the time of arrival at Q, by the time taken to travel* d2 *at velocity* V1. *The lines AC and PR are parallel to the refractor*.

However, both waves would arrive late at this point (Figure 3.5) and, for small dips, the delays would be very similar. The difference between the arrival times would thus be almost the same as if no depression existed, plotting on the straight line generated by the horizontal parts of the interface. The argument can be extended to a refractor with a series of depressions and highs. Provided that the dip angles are low, the difference points will plot along a straight line with slope corresponding to half the refractor velocity. Where a refractor has a constant dip, the slope of the difference line will equal to the dip velocity equation.

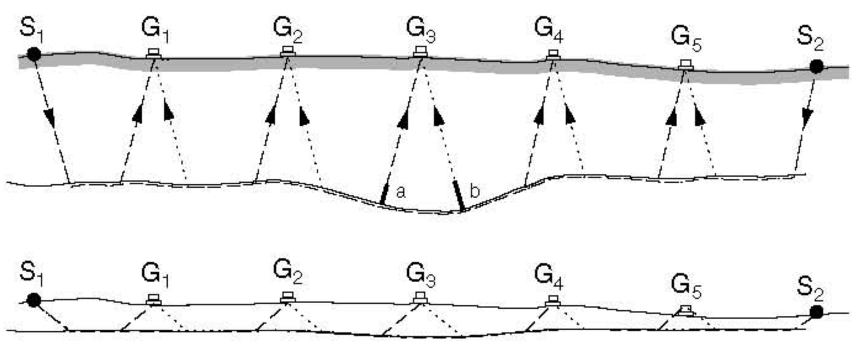

Figure 3.5 *Effect on travel times of a bedrock depression. The arrivals at G*3 *of energy from S*1 *and S*2 *are delayed by approximately the same amounts 'a'*

and 'b'). Note that, as in all the diagrams in this chapter, vertical exaggeration has been used to clarify travel paths. The lower version gives a more realistic picture of the likely relationships between geophone spacing and refractor depths and gradients.

the sum of the slopes of the individual lines, giving a graphical expression

The approach described above generally works far better than the very qualitative 'proof' (and the rather contrived positioning of the geophones in Figure 3.5) might suggest. Changes in slopes of difference lines correspond to real changes in refractor velocity, so that zones of weak bedrock can be identified. The importance of long shots is obvious, since the part of the spread over which the first arrivals from the short shots at both ends have come via the refractor is likely to be rather short and may not even exist. It is even sometimes possible, especially when centre shots have been used, for the differencing technique to be applied to an intermediate refractor. Differences are easily obtained directly from the plot using dividers, or a pencil and a straight-edged piece of paper. They are plotted using an arbitrary time zero line placed where it will cause the least confusion with other data (see Figure 3.8).

3.5 Reciprocal time interpretation

The *reciprocal time, t*R, is defined as the time taken for seismic energy to travel between the two long-shot positions. The difference between *t*R and the sum of the travel times *t*A and *t*B from the two long-shots to any given geophone, G, is:

$$tA + tB - tR = 2D/\mathbf{F}$$

where D is the depth of the refractor beneath G and **F** is a *depth conversion factor* (Figure 3.6). If there is only a single interface, D is equal to the

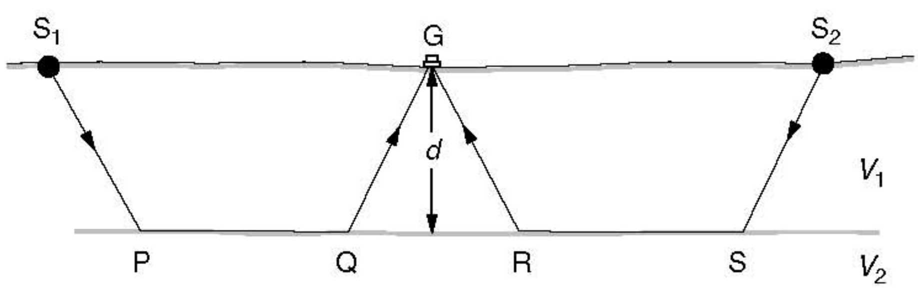

Figure 3.6 *Reciprocal time interpretation. The sum of the travel times from S*1 *and S*2 *to G differs from the reciprocal time,* tR, *taken to travel from S*1 *to S*2 *by*

the difference between the times taken to travel QR at velocity V2 *and QGR at velocity* V1.

thickness, *d*, of the upper layer and **F** is equal to *V*1,2. If there are more interfaces, **F** is a composite of all the velocities involved, weighted according to the layer thicknesses. At the short shots 2*D/***F** = ti (the intercept time) and the **F** values can be calculated. The ways in which **F** varies between these points may be very complex, but linear interpolation is usually adequate in the field (Example 3.1).

Although *t*R can be measured directly, it is more convenient to calculate it from the equation above using the intercept times. This can be done provided that geophones are located at the short-shot points when the long shots are fired (so that *t*A + *t*B at those points can be measured). The estimates of *t*R made using the data from the two ends should agree within the limits of error of the method (i.e. within 1–2 msec). If they are not, the raw data and the calculations should be thoroughly checked to find the reason for the discrepancy. Short-shot reciprocal times are measured directly if short-shots are fired at end-geophone positions, and the fact that they should be equal may help in picking arrivals. However, they have little interpretational significance.

4.0**Conclusion**

The science of seismology dates from the invention of the seismograph by the English scientist John Milne in 1892. Its name derives from its abilityto convert an unfelt ground vibration into a visible record. The seismograph consist of a receiver and a recorder. The ground vibration is detected and amplified by a sensor, called the seismometer or, in exploration seismology, the geophone. In modern instruments the vibration is amplified and filtered electronically. The amplified ground motion is converted to a visible record, called the seismogram.

5.0 Summary

Early seismographs were undamped and reacted only to limited band of seismic frequencies. Seismic waves with inappropriate frequencies were barely recorded at all, but strong waves could set the instrument into resonant vibration. In 1903, the Germen seismologist Emil Wiechart substantially increased the accuracy of the seismograph by improving the amplification method and by damping the instrument. These early instruments relied on mechanical levers for amplification and recording signals on smoked paper. This made them both bulky and heavy, which severely restricted their application.

6.0Tutor Marked Assignment

Q1. What are intercept times?
Q2. Explain by the use of diagram the reciprocal time interpretations
Q3 Differentiates between refractor relief and true velocity.

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Unit 3 LIMITATIONS OF THE REFRACTION METHOD

1.0 Introduction

First-arrival refraction work uses only a small proportion of the information contained in the seismic traces, and it is not surprising that interpretation is subject to severe limitations. These are especially important in engineering work; in low-velocity-layer studies only a time delay estimate is sought and short shots alone are often sufficient.

2.0 Objectives

At the end of the unit, readers should be able to

(i) Understand prominent practical limitations of the geophysical methods.
(ii) Be exposed to different techniques applicable in low velocity layer environment.

3.0 Main Contents

3.1 Direct waves
The *ground roll* consists of a complex of P and S body waves and Love and Rayleigh surface waves travelling with different but generally slow velocities. There is often some doubt as to which component actually produces the first break, since conventional geophones respond only poorly to the horizontal ground motions of direct P-waves. Close to the source, enough energy is associated with the P-waves for the response to be measurable, but at greater distances the first breaks may record the arrival of S-waves, surface waves or even the air wave.

The complex character of the direct wave may be among the reasons for the commonly observed failure of the best-fit arrival line to pass through the origin. Delays in the timing circuits may also play a part but can be determined by direct experiment, with a detonator or a light hammer blow close to a geophone. A more important reason may be that the amplifier gains at geophones close to the shot point have been set so low that the true first arrivals have been overlooked (Figure 3.7). Full digital storage of the incoming signals should allow the traces to be examined individually over a range of amplifications, but if this is not possible, then the most reliable velocity estimates will be those that do not treat the origin as a point on the line.

3.2 Vertical velocities

However much care is taken to obtain valid direct-wave or refracted-wave velocities, the refraction method is fundamentally flawed in that the depth equations require vertical velocities but what are actually measured are horizontal velocities. If there is significant anisotropy, errors will be introduced.

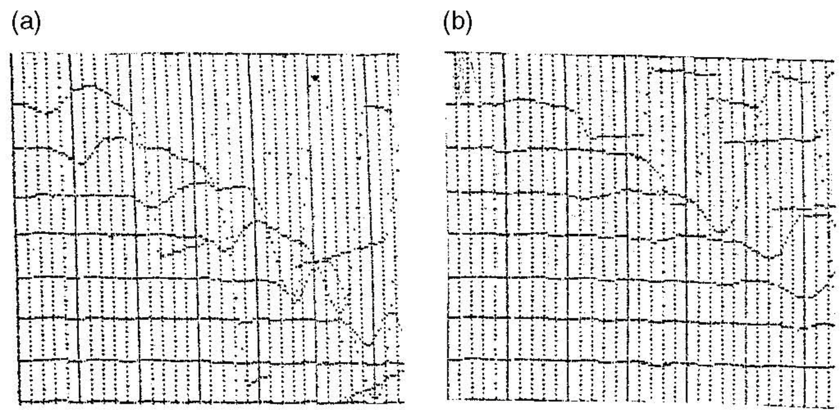

Figure 3.7 *Hard copy of a single stored data set played back at two different amplifications. The first arrivals clearly visible on (a) would probably be overlooked or dismissed as noise on (b). A direct wave velocity based on (b) would be roughly correct provided that the best-fit line was not forced through the origin. The cross-over distance would also be wrong but the intercept time would not be affected, provided that the refracted arrivals were amplified sufficiently.*

This is a problem for interpreters rather than field observers but the latter should at least be aware of the importance of using any boreholes or recent excavations for calibration or to measure vertical velocities directly.

3.3 Hidden layers

A refractor that does not give rise to any first arrivals is said to be hidden. A layer is likely to be hidden if it is much thinner than the layer above and has a much lower seismic velocity than the layer below. Weathered layers immediately above basement are often hidden. The presence of a hidden layer can sometimes be recognized from second arrivals but this is only occasionally possible, in part because refracted waves are strongly attenuated in thin layers.

A layer may also be hidden even if the head wave that it produces does arrive first over some part of the ground surface, if there are no appropriately located geophones. Concentrating geophones in the critical region can sometimes be useful (although never convenient) but the need to do so will only be recognized if preliminary interpretations are being made on a daily basis.

3.4 Blind zones

If velocity decreases at an interface, critical refraction cannot occur and no refracted energy returns to the surface. Little can be done about these *blind* interfaces unless vertical velocities can be measured directly. Thin high-velocity layers such as perched water tables and buried terraces often create blind zones. The refracted waves within them lose energy rapidly with increasing distance from the source and ultimately become undetectable. Much later events may then be picked as first arrivals, producing discontinuities in the time–distance plot. A similar effect is seen if the layer itself ends abruptly.

Example 3.1

Field interpretation of a four-shot refraction spread with long shot (LS) and short shot (SS) arrivals from west (W) and east (E) ends plotted on same set of axes (Figure 3.8). After plotting the data, interpretation proceeds in the following stages.

Stage 1 Base refractor intercept times

Measure LS(W)–SS(W) time differences. These are roughly constant and close to 41 ms from G6 to G12, indicating that in this region the SS(W) arrivals have come from base refractor. Similarly, LS(E)–SS(E) time differences are close to 59 ms, from G1 to G4.

Intercept times: LS(W) time at W end = 101 ms.
Intercept time = 101 − 41 = 60 ms.
LS(E) time at E end = 208 ms.
Intercept time = 208 − 59 = 149 ms.

Note the LS(E) difference from the extrapolated intercept time of about 170 ms.

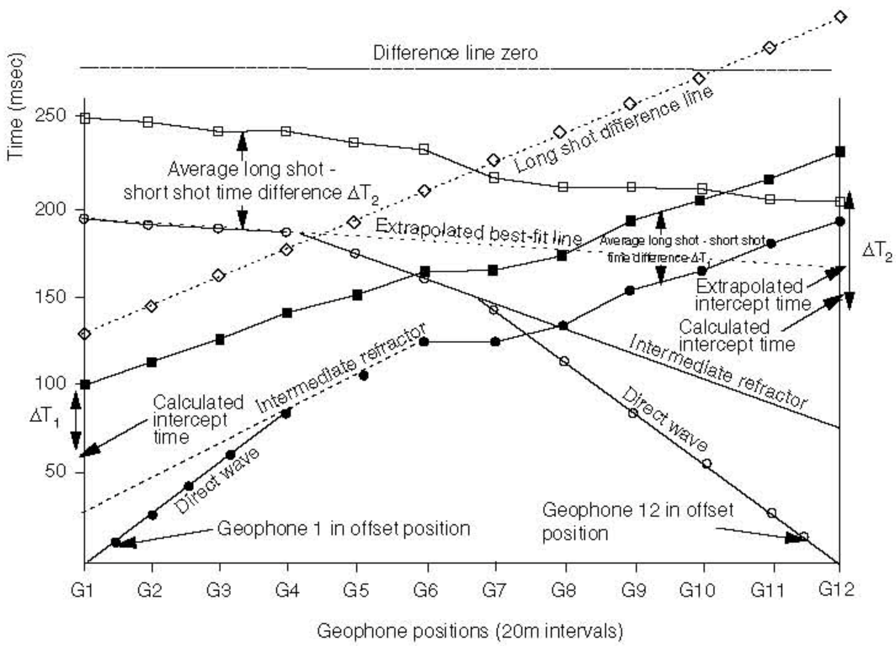

Figure 3.8 *Time–distance plot for a four-shot refraction spread. The long shot difference times, indicated by open circles, are referred to a zero line arbitrarily placed at* t = 280 *ms*−1. *Note the difference between the intercept time obtained by extrapolation of short-shot data and by using long-shot–short-shot difference times, for the G12 position. Extrapolation of the refracted arrival line back to zero time would be even more difficult for the G1 position and could lead to an even more erroneous interpretation. West is to the left of the plot (Example 3.1).*

Stage 2 Velocities

Direct-wave velocity:

Straight line from W origin through nearby SS(W) arrivals extends 60 m to G4.

Velocity *V*1 = 60/0.079 = 759 ms−1

Straight line from E origin through nearby SS(E) arrivals extends 100 m to G7.

Velocity *V*1 = 100/0.134 = 746 ms−1 Average *V*1 value = 750 ms−1

Intermediate refractor:

Arrivals at G5 from SS(W) and at G5 and G6 from SS(E) do not belong to the ‘base refractor’ sets (see Stage 1) nor do they fall on the direct-wave arrival line, suggesting the presence of an intermediate, ‘*V*2’, refractor. The *V*2 velocity is poorly controlled but the arrivals lines should pass above all direct wave (*V*1) and base refractor first arrivals. For the most likely positions, as shown;

SS(W): V2 = 1470 ms−1 Intercept time = 29 ms

SS(E): V2 = 1560 ms−1 Intercept time = 77 ms

These velocities suggest that the interface is probably the water table, with a velocity of about 1500 ms−1.

Base refractor velocity:
Plot LS(W)–LS(E) time differences at each geophone, using a convenient (280 ms) line as time zero.

*V*3 = 2/(slope of difference line) = 2 × 220/00.182 = 2420 ms−1

Velocity functions:

*V*1,2 = *V*1 × *V*2/ (*V* 22− *V* 21) = 750 × 1500/ (15002 − 7502) = 870 ms−1

*V*1,3 = *V*1 × *V*3/ (*V* 23− *V* 21) = 750 × 2420/ (24202 − 7502) = 790 ms−1

*V*2,3 = *V*2 × *V*3/ (*V* 22− *V* 21) = 1500 × 2420/ (24202 − 15002) = 1910 ms−1

Stage 3 Depths at shot points

Depths to intermediate refractor (*d*1= 12*tiV*1,2):**
W end: *d*1 = 12× 0.029 × 870 = 12.6 m
E end: *d*1 = 12× 0.077 × 870 = 33.5 m

**Thickness of intermediate layer (*d*2= 12
[*t*i −2*d*1/*V*1,3]×*V*2,3):**
W end: *d*2 = 12× {0.060 − 25.2/790} × 1910 = 26.8 m
D = 26.8 + 12.6 = 39.4 m
E end: *d*2 = 12× {0.149 − 67.0/790} × 1910 = 61.3 m
D = 33.5 + 61.3 = 94.8 m

Stage (4) Reciprocal time interpretation (example using Geophone 8)

Reciprocal time (*t*A +*t*B −*t*i)
W end: *t*R = 101 + 254 = 295 ms
E end: *t*R = 233 + 208 − 149 = 292 ms
Average = 293 ms

Depth conversion factors at short shots (F = 2 × D/*t* i)
W end: **F** = 2 × 39.4/0.060 = 1310 ms−1
E end: **F** = 2 × 94.8/0.149 = 1270 ms−1
F at G8 (by interpolation) = 1280 ms−1

Depth at G8 (D = tA +*t*B −*t*R)
D = 12× *(*0.174 + 0.213 − 0.293*)* × 1280 = 60.2 m

4.0 Conclusion

Despite the limitations of refraction surveys, interpretations are not always wrong when they disagree with drill hole data. Only a very small subsurface volume is sampled by the drill, and many drill tests of drift thickness have been terminated in isolated boulders some distance above the true top of bedrock. It is always important that explanations are found for any differences between drilling and seismic results.

5.0 Summary

The complex character of the direct wave may be among the reasons for the commonly observed failure of the best-fit arrival line to pass through the origin. Delays in the timing circuits may also play a part but can be determined by direct experiment, with a detonator or a light hammer blow close to a geophone.

6.0 Tutor Marked Assignments

Q1. Write short notes on the following

a) Direct Waves
b) Hidden layers
c) Blind zones
d) Intermediate refractor

7.0 References/ Further Readings

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