

Seismic Reflection and Refraction Methods

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Introduction

Seismic reflection and refraction is the principal seismic method by which the petroleum industry explores hydrocarbon-trapping structures in sedimentary basins. Its extension to deep crustal studies began in the 1960s, and since the late 1970s these methods have become the principal techniques for detailed studies of the deep crust. These methods are by far the most important geophysical methods and the predominance of these methods over other geophysical methods is due to various factors such as the higher accuracy, higher resolution and greater penetration. Further, the importance of the seismic methods lies in the fact that the data, if properly handled, yield an almost unique and unambiguous interpretation. These methods utilize the principle of elastic waves travelling with different velocities in different layer formations of the Earth. An energy source produces seismic waves that are directed into the ground. These waves pass through the earth and are reflected / refracted at every boundary between rocks of different types. The response to this reflection / refraction sequences is received by instruments placed on or near the surface. Onshore seismic data is recorded using a simple (normally electro-magnetic) device known as a geophone, whereas hydrophones are used in marine recording which normally uses a piezo-electric device to record the incoming energy. Thus, the data set derived from seismic methods consists of a series of travel-times versus distances. The knowledge of travel times to the various receivers and the velocity of waves in various media enable us to reconstruct the paths of seismic waves. Structural information is derived principally from paths that fall into two main categories, viz., reflected paths and refracted paths. These paths can be obtained using seismic reflection and refraction methods. In seismic reflection method the waves travel downward initially and are reflected at some point back to the surface, the overall path being essentially vertical. In this sense, reflection method is a very sophisticated version of the echosounding used in submarines, ships,

and radar systems. Whereas, in seismic refraction method, principal portion of the wave-path is along the interface between the two layers and hence approximately horizontal. The travel times of the refracted wave paths are then interpreted in terms of the depths to subsurface interfaces and the speeds at which wave travels through the subsurface within each layer. For both types of paths the travel time depends upon the physical property, called elastic parameters, of the layered Earth and the attitudes of the beds. The objective of the seismic exploration is to deduce information about such beds especially about their attitudes from the observed arrival times and from variations in amplitude, frequency and wave form.

Both the seismic techniques have specific advantages and disadvantages when compared to each other and when compared to other geophysical techniques. For these reasons, different industries apply these techniques to differing degrees. For example, the oil and gas industries use the seismic reflection technique almost to the exclusion of other geophysical techniques. The environmental and engineering communities use seismic techniques less frequently than other geophysical techniques. When seismic methods are used in these communities, they tend to emphasize the refraction methods over the reflection methods.

Advantages and disadvantages of seismic methods

Seismic methods have several distinct advantages as well as disadvantages compared to other geophysical methods (Table 1). The seismic methods are more expensive to undertake than other geophysical methods. It can produce remarkable images of the subsurface, but this comes at a relatively high economic cost. Thus, when selecting the appropriate geophysical survey, one must determine whether increased resolution of the survey is justified in terms of the cost of conducting and interpreting observations from the survey.

Table 1. Advantage and disadvantage of seismic methods.

Seismic Methods	
Advantage	Disadvantage
Can detect both lateral and depth variations in a physically relevant parameter: seismic velocity.	Amount of data collected in a survey can rapidly become overwhelming.
Can produce detailed images of structural features present in the subsurface.	Data is expensive to acquire and the logistics of data acquisition are more intense than other geophysical methods.

Can be used to delineate stratigraphic and, in some instances, depositional features.	Data reduction and processing can be time consuming, require sophisticated computer hardware, and demand considerable expertise.
Response to seismic wave propagation is dependent on rock density and a variety of physical (elastic) constants. Thus, any mechanism for changing these constants (porosity changes, permeability changes, compaction, etc.) can, in principle, be delineated via the seismic methods.	Equipment for the acquisition of seismic observations is, in general, more expensive than equipment required for the other geophysical surveys.
Direct detection of hydrocarbons, in some instances, is possible.	Direct detection of common contaminants present at levels commonly seen in hazardous waste spills is not possible.

Elastic waves

If the stress applied to an elastic medium is released suddenly, the energy imparted into the Earth by the source is transmitted in the form of elastic waves. A wave is defined as a disturbance that propagates through, or on the surface of, a medium. Elastic waves satisfy this condition and also propagate through the medium without causing permanent deformation of any point in the medium. Waves that propagate through the Earth as elastic waves are referred to as seismic waves. There are two broad categories of seismic waves: body waves and surface waves.

Body waves - Body waves are elastic waves that propagate through the Earth's interior. In reflection and refraction prospecting, body waves are the source of information used to image the Earth's interior. Seismic body waves can be further subdivided into two classes of waves: Longitudinal or P-waves and Transverse or S-waves.

Longitudinal or P-waves - P-waves are also called primary waves, because they propagate through the medium faster than the other wave types. Most exploration seismic surveys use P-waves as their primary source of information. In P-wave, particles constituting the medium are displaced in the same direction as the direction of the wave propagation. Thus P-wave pushes the particles of material ahead of it, causing compression and expansion of the material. P waves are analogous to sound waves propagating through the air. The velocity of P-waves (V_P) is given by:

$$V_P = \sqrt{(\lambda + 2\mu)/\rho}$$

where, ρ is the density of the medium, μ is the measure of resistance to shearing strain often referred to as the modulus of rigidity or shear modulus and λ is the Lamé's constant. P-waves are also known as compressional wave, longitudinal wave, push-pull wave, pressure wave, dilatational wave, rarefaction wave and irrotational wave.

Transverse or S-waves - S-waves are sometimes called secondary waves, because they propagate

through the medium slower than P-waves. In S-waves, particles constituting the medium are displaced at right angles to the direction of wave propagation. The velocity of these waves (V_S) is given by:

$$V_S = \sqrt{(\mu/\rho)}$$

S-Waves are used for some forms of seismic exploration but they do not propagate through fluids. Therefore, it is not recorded (directly) in conventional marine acquisition. It is evident that $V_S < V_P$. S-wave is also known as shear wave, transverse wave, rotational wave, distortional wave and tangential wave.

Surface waves

Surface waves are the waves that propagate along the Earth's surface. Their amplitude at the surface of the Earth can be very large and decay with distance more slowly compared to body waves. Surface waves propagate at speeds that are slower than S-waves and are less efficiently generated by buried sources. Like body waves, there are two classes of surface waves, Love-waves and Rayleigh-waves that are distinguished by the type of particle motion they impose on the medium. Surface waves are confined to the vicinity of one of the surfaces, which bound the medium. In Rayleigh waves, the particles describe ellipses in the vertical plane that contain the direction of propagation. At the surface, the motion of the particle is retrograde with respect to that of the waves. The velocity of Rayleigh waves is about $(0.9 \times V_S)$. Love waves are observed when the velocity in the top of the medium is less than that of the sub-stratum. The particles oscillate transversely to the direction of the wave and in a plane parallel to the surface. Love waves have velocities intermediate between the S-wave velocity at the surface and that in the deeper layers, and exhibit dispersion. The spectrum of the body waves in the Earth generally extends from about 15 Hz whereas the surface waves have frequencies lower than 15 Hz. It is worth mentioning here that for virtually all exploration surveys, surface waves are a form of noise that we attempt to suppress. In all of the remaining discussion about seismic waves, we will consider only body waves.

Factors affecting the amplitude of seismic waves

Many factors affect the amplitude of seismic waves and some of them are shown in Fig. 1. Many of these do not involve the subsurface of the Earth. In order to make meaningful interpretation of amplitude variations, the effects of irrelevant factors need to be removed.

The energy density of the seismic wave decreases as it travels down in the subsurface. The decrease of energy density is inversely proportional to the square of the distance over which the wave has travelled, assuming a constant velocity. This phenomenon is called as spherical divergence. Because velocity increases with depth, ray path curvature causes the energy density to fall off even more rapidly. The rate of amplitude change for multiples and primaries differ because the curvature depends upon the velocity variations.

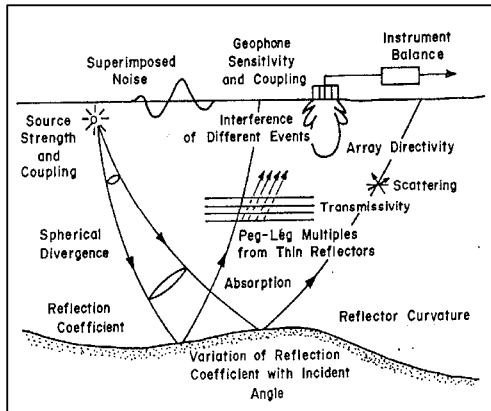


Fig. 1. Factors which affect amplitude of seismic wave.

Absorption is another factor, which affects amplitude. The loss of energy in the Earth due to absorption is described in various ways viz., i) by a quantity called 'Q' (the amount of energy in a seismic wavelet compared to the amount of energy dissipated in one cycle); ii) by an absorption coefficient, which is an exponential decay factor, and iii) by logarithmic decrement, a measure of the change of amplitude between two successive cycles. The loss of energy by absorption increases with frequency.

The shape of a reflector affects the energy density. The curvature of the reflector acts to focus or defocus the energy. Velocity situations such as a gas cap can have similar effects, acting as a lens to distort reflections from underneath them.

Seismic reflection Method

The seismic reflection method is one of the best geophysical methods because it produces the best images of the subsurface. These data resolve mappable features such as faults, folds and lithologic boundaries measured in the 10's of meters, and image them laterally for 100's of kilometers and to depths of 50 km or more.

Geometry of reflection paths

When there is a series of interfaces separating individual formations having different velocities and the distances between the interfaces are large compared with the seismic wavelengths employed, a separate reflection should theoretically be observed from each interface, although this does not always occur for a number of reasons.

The time required for the waves to travel from a near surface source to the reflectors and back to receivers on the surface are used, along with all available information to determine the structure of the reflecting surface. This process forms the geometrical basis for the reduction used with the reflection method.

Reflection from horizontal surface

Let us assume a horizontal reflecting interface at a depth z below the Earth's surface. The seismic velocity above the interface is V_0 and that below it is V_1 (Fig. 2a). The path of the reflected wave generated at the shot point and received by a detector at a distance x (also known as offset) away from it consists of two segments which have travelled from the surface to the reflecting interface and back to the surface. If T is the total time, the total path length L is related to x and z by the formula:

$$L = 2\sqrt{z^2 + (x/2)^2} = V_0 T \dots \dots \dots (1)$$

$$V_0^2 T^2 = 4z^2 + x^2$$

$$V_0^2 T^2 / 4z^2 - x^2 / 4z^2 = 1$$

Therefore, it is evident from the above equation that the relation of T to x is hyperbolic for a horizontal reflector. The equation (1) can also be written as:

$$T = (2/V_0)\sqrt{z^2 + (x/2)^2}$$

$$T^2 = 4/V_0^2 (z^2 + x^2/4)$$

$$z^2 = T^2 V_0^2 / 4 - x^2 / 4$$

$$z = [\sqrt{(V_0 T)^2 - x^2}] / 2$$

Fig. 2b shows the relation between travel time and horizontal distance for a reflected ray. It can be shown that the curve is actually a symmetrical one, as it holds for negative and positive value of x . The portion corresponding to the negative values is not shown in the figure. The axis of symmetry of hyperbola representing the relationship is the line $x=0$.

After squaring, equation (1) can be written as :

$$T^2 = x^2/V_0^2 + 4z^2/V_0^2$$

The plot of T^2 against x^2 is a straight line, which has a slope of $1/V_0^2$ and an intercept of $4z^2/V_0^2$. The Fig. 2b (inset) shows the linearity between T^2 and x^2 . The velocity V_0 can readily be determined from the slope of

the line thus obtained. Similarly the equation for travel time for the reflection from a bed (Fig. 3b), dipping at an angle ϕ can be derived.

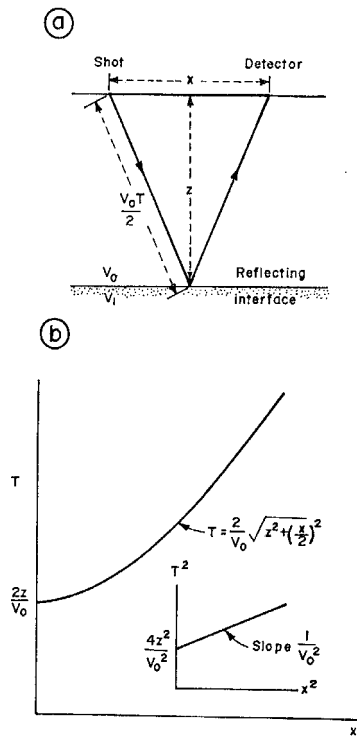


Fig. 2. (a) Reflected wave from single interface. V_0 is the constant velocity of P-wave in a media between source and reflecting surface.

(b) Travel-time curve for P-wave, reflected from a horizontal surface at depth Z . Inset graph shows linear relation between T^2 and X^2 .

As the inclination of the down going ray decreases (the angle with the vertical increases), the down travelling ray eventually approaches the boundary at the critical angle, $\sin^{-1}(V_0/V_1)$. At angles smaller than the critical, a large proportion of the energy in the wave is transmitted (refracted) downward into the layer below the interface. At critical-angle incidence, the refracted wave travels horizontally along the boundary at a speed of the underlying medium. At any greater angle of incidence, there is total reflection, and the wave does not penetrate into the lower material at all. Reflection continues to take place at angles greater than the critical angle, but it is evident from Fig. 3a that the ray refracted horizontally along the interface is returned to the surface at the critical angle, will reach a distant point on the surface before the reflected wave reaches the same point. The direct waves that have taken a horizontal path along the

Earth's surface will be the first event to arrive up to the distance x_{cross} , which is shown in Fig. 3a.

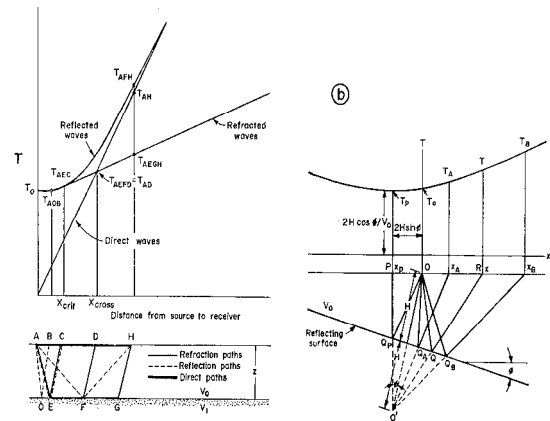


Fig. 3. (a) Time-distance relation for reflected, (b) Ray-path trajectories and refracted and direct waves. time-distance relation for dipping bed.

Geometry of refraction paths

Mechanism for transmission of refracted waves

Refraction ray paths are not always as easy to predict as reflection paths. In a layered Earth, rays refracted along the top of high speed layers travel down from the source along slant paths, approach them at the critical angle, and return to the surface along a critical angle path rather than along some other path such as the normal incidence.

Let us consider a hypothetical subsurface consisting of two media, each with uniform elastic properties and the upper horizon separated from the lower one by a horizontal interface at depth z (Fig. 4). Longitudinal velocity of seismic waves in the upper layer is V_0 and the lower layer is V_1 with $V_1 > V_0$. A seismic wave is generated at point S on the surface and the energy travels out from it in the hemispherical wave fronts.

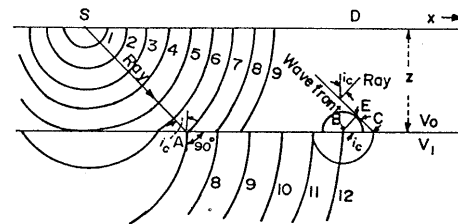


Fig. 4 Mechanism for transmission of refracted waves in two-layered Earth.

A receiving instrument is located at point D at a distance x from S . If x is small, the first wave to arrive at D will be the one that travels horizontally at a speed V_0 .

At greater distance, the wave that took an indirect path, travelling down to, along and up from V_1 layer arrive first because the time gained in travel through the higher speed material makes up for the longer path.

When the spherical wave fronts from S strike the interface where the velocity changes, the energy will be reflected into the lower medium according to Snell's law. The processes are demonstrated in Fig. 4 for the time corresponding to the wave front 7. At point A on wave front 7, the tangent to the sphere in the lower medium becomes perpendicular to the boundary. The ray passing through this point now begins to travel along the boundary with the speed of the lower medium.

Thus, by definition, the ray SA strikes the interface at the critical angle. To the right of A, the wave fronts below the boundary travel faster than those above.

The material on the upper side of the interface is subjected to oscillating stress from below which, as the wave travels, generates new disturbances along the boundary. These disturbances themselves spread out spherically in the medium with a speed of V_0 . The wave originating at point B in the lower medium will travel a distance BC during the time in which the one spreading out in the upper medium will attain a radius of BE. The resultant wave front above the interface will follow the line CE, which makes the angle i_c with the boundary. It can be seen from Fig. 4 that:

$$\begin{aligned} \sin i_c &= BE / BC \\ &= V_0 t / V_1 t \\ &= V_0 / V_1 \end{aligned}$$

The angle which the wave-front makes with the horizontal is the same as that which the ray perpendicular to it makes with the vertical, so that the ray will return to the surface at the critical angle $[\sin^{-1}(V_0/V_1)]$ with a line perpendicular to the interface.

The simplest and most useful way to represent refraction data is to plot the first-arrival time T vs the shot-detector distance x similar to that of reflection data. In case of the sub-surface consisting of discrete homogeneous layers, as in Fig. 5a, this type of plot consists of linear segments.

Time distance relation for refraction paths

Two layer case - Let us determine the time-distance relations for the case (Fig. 5a) of two media with respective speeds of V_0 and V_1 , separated by a horizontal discontinuity at depth z .

The direct wave travels from shot to detector near the Earth's surface at a speed of V_0 , so that $T = x/V_0$. This is represented on the plot of T versus x as a straight line, which passes through the origin and has a slope of $1/V_0$. The wave refracted along the interface at depth z ,

reaching it and leaving it at the critical angle i_c , takes a path consisting of three legs, AB, BC and CD. To determine the time in terms of horizontal distance traveled, the following relations are considered:

$$\begin{aligned} \sin i_c &= V_0/V_1 \\ \cos i_c &= \sqrt{1-V_0^2/V_1^2} \\ \tan i_c &= \sin i_c / \cos i_c \\ &= V_0/\sqrt{V_1^2-V_0^2} \end{aligned}$$

The total time (T) along the refraction path ABCD is given as :

$$T = T_{AB} + T_{BC} + T_{CD}$$

The above equation can also be written as :

$$\begin{aligned} T &= [z / (V_0 \cos i_c)] + [(x - 2z \tan i_c) / V_1] + [z / (V_0 \cos i_c)] \\ &= (2z / V_0 \cos i_c) - (2z \sin i_c / V_1 \cos i_c) + (x / V_1) \end{aligned}$$

This can be transformed as :

$$\begin{aligned} T &= [2z / (V_0 \cos i_c) (1 - \sin^2 i_c)] + [x/V_1] \\ &= [x / V_1] + [2z \cos i_c / V_0] \\ &= [x / V_1] + [2z \sqrt{1 - (V_0/V_1)^2} / V_0] \end{aligned}$$

Therefore,

$$T = [x / V_1] + [2z \sqrt{V_1^2 - V_0^2} / V_0 V_1]$$

On a plot of T vs x , the equation of straight line which has a slope of $1/V_1$ and intercepts the T axis ($x = 0$) at a time (T_i) is given :

$$T_i = 2z \sqrt{[V_1^2 - V_0^2] / V_0 V_1} \dots \dots \dots (2)$$

where T_i is known as the intercept time.

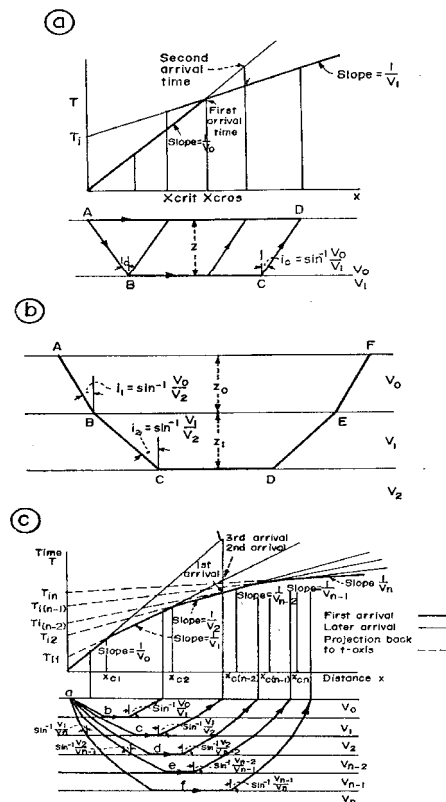


Fig. 5. (a) Travel-time curve for refracted wave for a layer

separated from its substratum horizontal interface. X_{crit} is the critical distance and X_{cross} is the crossover distance.

(b) Refracted ray paths for two layers separated by two horizontal interfaces.

(c) Ray paths, time-distance curve and critical distances for multilayered horizontal interfaces. X_{c1} , X_{c2} etc. are crossover distances for successively dipping interfaces.

Crossover distance - At a distance x_{cross} (Fig. 5a), the two linear segments cross. At a distance less than this, the direct wave travelling along the top of the V_0 layer reaches the detector first. At greater distances, the wave refracted by the interface arrives before the direct wave. For this reason, x_{cross} is called the *crossover distance*.

Depth calculation - The depth (z) to the interface can be calculated from the intercept time by using the equation (2) or from the crossover distance. This equation can be solved for z to obtain

$$z = (T_i / 2) V_0 V_1 / \sqrt{V_1^2 - V_0^2},$$

where T_i can be determined graphically (Fig. 5a) or numerically from the relation

$$T_i = T - x / V_1.$$

Three layer case - For three formations with velocities V_0 , V_1 and V_2 ($V_2 > V_1 > V_0$), the treatment is similar but somewhat complicated. Fig. 5b shows the wave paths. The ray corresponding to the least travel time makes an angle $i_1 = \sin^{-1}(V_0/V_2)$ with the vertical in the upper most layer and the angle $i_2 = \sin^{-1}(V_1/V_2)$ with the vertical in the second layer and i_2 being the critical angle for the lower interface. The time along each of the two slant paths AB and EF through the uppermost layer can be calculated as described above. Similarly, the time along each of the legs BC and DE crossing the middle layer can be calculated. The time for the segment of path CD at the top of the V_2 layer is CD/V_2 . The expression for the total travel time from A to F can be given by

$$T = T_{AB} + T_{BC} + T_{CD} + T_{DE} + T_{EF}$$

The intercept time (T_i) and depth to the horizon (d) can be determined by solving the above.

Multi-layer case - The time-depth relations derived for the two layer case can be readily extrapolated to apply to a large number of layers as long as the speed in each layer is higher than the one above it. This is illustrated by Fig. 5c which shows ray-paths and time-distance plots for six layer case, the lower most designated for generality as the n^{th} layer. Each segment on the plot represents first arrival from the top of one of these sub-surface layers, the horizontal ray-paths corresponding to the respective segments being designated by the letters b to f. Deeper the layer, the greater is the shot-detector distance at which arrivals from the former become the first to be observed. In other words, the crossover distance for each layer

increases with the depth. The slope of each segment is simply the reciprocal of the speed in the layer if the wave has travelled horizontally along the same. The intercept time of each segment depends on the depth of the interface at the bottom of the corresponding wave path as well as on the depths of all those interfaces that lie above it in the section.

Low speed layer

To calculate depths for additional layers using the refraction technique discussed above, it is essential that deeper layers have successively higher speed. If any bed in the sequence has a lower speed than the one above it, it will not be detectable by refraction shooting at all. This is because the ray entering into such a bed from the top layer is always deflected in the downward direction, as shown for the interface between V_0 and V_1 layers in Fig. 6a and thus can never travel horizontally through the layer. Consequently, there will be no segment of inverse slope V_1 on the time-distance curve. The presence of such an undetected low speed layer will result in the computation of erroneous depths to all interfaces below it if only observed segments are used in the calculations.

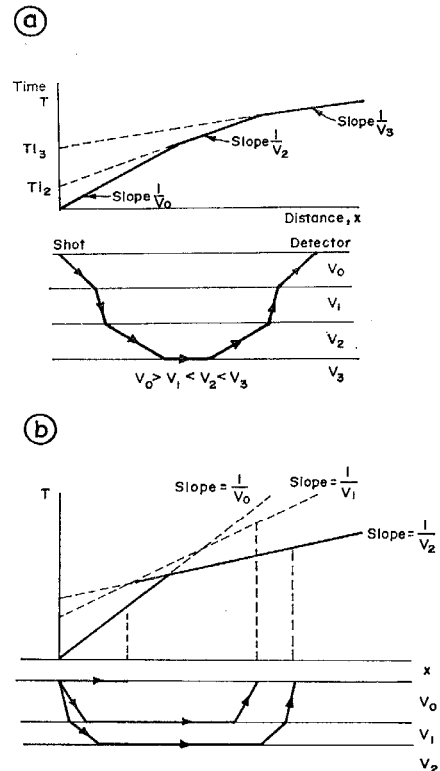


Fig. 6. (a) Ray paths and time-distance curve where low speed layer V_1 lies below higher speed layer V_0 . $V_3 > V_2 > V_1$. (b) The blind zone in refraction shooting. It is not possible to observe a first arrival from the top of V_2 layer because arrivals from either V_3 or V_1 are earlier to all receiving distances.

Blind zone

A similar kind of error may occur if the thickness of the layer with a speed intermediate between that of the layer overlying and the one below is small and/or the velocity contrast between this and the layer that underlies is inadequate. If only first arrivals can be observed on the records, the refracted waves from the intermediate layer will not be discernible because they will always reach the surface at a later time than from either a shallower or a deeper bed, depending on the shot-detector distance. Fig. 6b shows the relation of the

time-distance segments from the respective layers when such zone, called as a blind zone, presents as an intermediate layer. Therefore, it is important to take such zones into account for proper accuracy in shallow refraction investigations, particularly for engineering purposes. If a test borehole in the survey area reveals a layer undetectable by refraction, its effect can be taken into account in calculating depths to high-speed beds.

Multiple fold profiling

The multiple fold profiling also known as common-depth-point (CDP) technique is extensively used in attenuating several kinds of noise. Basically, signals associated with a given reflection point but recorded at a number of different shot and receiver positions are composited together after removal of normal move out. Fig. 7a illustrates the recording arrangement when six signals are composited. Fig. 7b shows the successive shot positions for a set of single-end shooting spreads giving six-fold multiple coverage. Reflection takes place at a point when the shot is at position 1 and the receiver at position 21 and continues when the shot is at position 2 and receiver at 17. Such a reflection also occurs with shots and the receivers at four other combinations of positions shown in Fig. 7b. At the bottom of Fig. 7b, the dots show reflection points which correspond to all recorded wave paths. It can be seen that there are six such paths for each subsurface position. Adding the proper signals for coincident subsurface reflection positions as indicated on the diagram and plotting the output traces thus obtained on a record section is referred to as STACKING. Following are some useful thumb rules to calculate different parameters in multiple fold profiling:

$$\text{Max. Fold Coverage} = [\text{No. of Channel}] / [2 * (\text{Shot Interval}) / (\text{Group Interval})]$$

$$\text{CDP per Shot} = [\text{No. of Channel} / \text{Max. Fold Coverage}]$$

$$\text{First CDP} = [\text{No. of Channel} + (\text{Near Offset} / \text{Group Interval})]$$

$$\text{Last CDP} = [\text{Total Shots} * (\text{No. of CDPs per Shot}) + (\text{First Full Fold CDP} - 1)]$$

$$\text{First Full Fold CDP} = [\text{First CDP} + (\text{No. of Channel} - \text{CDP per Shot})]$$

$$\text{Last Full Fold CDP} = [\text{Last CDP} + (\text{No. of Channel} - \text{CDP per Shot})]$$

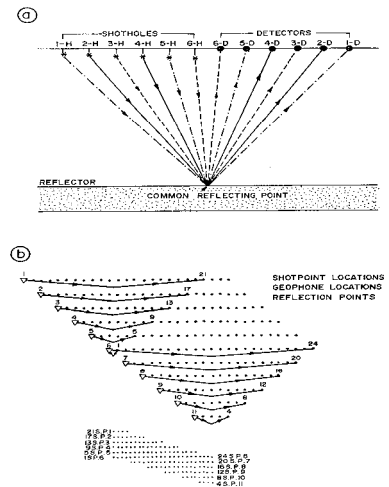


Fig. 7. (a) Ray paths for reflections from a single point in six fold profiling. (b) Shot and receiver combinations giving six-fold multiplicity. Patterns at bottom show reflecting points for successive spreads.

The recording of offshore seismic data is complicated by the fact that all of the recording equipment must be encased in an oil-filled cable or streamer (about 10 cm in diameter) that is towed behind the vessel. The hydrophones are connected together in groups and may be placed at every meter or so, in a long streamer. The front-end of the streamer is connected to the vessel by a complex system of floats and elastic stretch-sections, which are designed to eliminate any noise reaching the streamer from the vessel. The end of the streamer, farthest from the vessel, is connected by similar stretch-sections to a tail-buoy. This buoy may contain its own GPS receiver and radar reflector so that its position can be established.

Although the oil used within the streamer is designed to give the streamer neutral-buoyancy, changes in tides and currents can affect the depth of the streamer. Mechanical depth controllers (known as birds) are placed at intervals along the streamer which adjust the angle of their "wings" to correct for undesirable changes in depth. Compasses are also used within the streamer itself to check the angle of the streamer at several positions along its length.

Ocean Bottom Seismometer

The seismometer (or geophone) is a detector that is placed in direct contact with the earth to convert very small motions of the earth into electrical signals, which are recorded digitally. The Ocean Bottom Seismometers

(OBS) are normally designed to record the earth motion under oceans and lakes from air-gun seismic sources. The air gun array is towed behind a ship usually at a distance of about 50-60 m and about 25 m below sea surface. The air gun shooting is carried out either at fixed time interval or at a fixed distance interval at a ship's speed of about 4-5 knots. The seismic waves, generated by air gun at the sea, travel at different paths as mentioned earlier. Each OBS then receives several seismic waves for each shots arriving at different times. Each one of these arrivals is called a ray path. By combining the time of the arrival, and the magnitude and phase of ray paths at many different locations and from several OBS, a model of the earth's structure can be created.

The sensors used in the OBS consist of one vertical seismometer, two horizontal seismometers, and one hydrophone (Fig. 8). The seismometers are gimbal-mounted to ensure that the sensors are level to operate efficiently. The horizontal seismometers are mounted 90 degrees to the vertical and 90 degrees to each other. The hydrophone provides information that is similar to the vertical seismometer, and under certain conditions can have a better signal/noise ratio. The horizontal geophones, mounted in the OBS, record S-wave (shear wave) information. This method of imaging the earth is called wide-angle seismic reflection and refraction because the trajectories of the ray paths are generally at wide-angle from the vertical. The OBS typically provide information about the structure of the Earth down to about 30-40 kilometers.

Seismic refraction work carried out at sea is faster and cheaper than a similar work carried out on land, because a non-explosion source can be towed behind the ship and fired repetitively while the ship is moving. On land, holes have to be drilled and explosives lowered into them. Land refraction work consists, therefore, of few shots and many seismometers, which require many people and large amount of time to install and move around. Marine work uses fewer OBS and many more shots.

The OBS normally consists of an aluminum sphere which contains sensors, electronics, enough alkaline batteries to last few days on the ocean bottom, and an acoustic release. The two sphere halves are put together with an O-ring and a metal clamp to hold the halves together. A slight vacuum is placed on the sphere to better ensure a seal. The sphere by itself float, therefore an anchor is needed to sink the instrument to the bottom. The anchor can be a flat metal plate or cemented slab. The instrument is designed in such a way that it can be deployed and recovered easily. All that is needed (for deployment and recovery) is enough deck space to hold the instruments and their anchors and a boom capable of lifting an OBS off the deck and lower it into the sea. The

OBS is bolted to the anchor and then dropped (gently) over the side.

Prior to deployment, the data logger is programmed with the number of sensors to record, the sample rate for data acquisition, and the start time. At the designated time, information from the selected sensors is recorded on the data logger's which is usually a high capacity hard disk. This information is recorded continuously for all selected sensors until either the hard disk is filled or the OBS is recovered and the data collection stopped.

The OBS is recovered with an acoustic release. The ship is positioned at the deployment location, and an electronics unit on the ship transmits a pulse at a specific frequency. The OBS, sitting on the ocean bottom, detects this pulse and replies with its own pulse at a different frequency. The time from the moment that the ship sent the first pulse to the time the ship receives the return pulse from the OBS is used to determine the distance between the ship and the OBS. Once the OBS's location is confirmed, a coded transmitted pulse is sent from the surface to initiate the release. The OBS then is released from its anchor and enables a double pulse return (to indicate that it has released). Once the OBS is reached the sea surface, the instrument is retrieved and washed with fresh water. The next deployment requires downloading the data onto another computer, replacing the batteries, testing the system, and programming in new parameters.

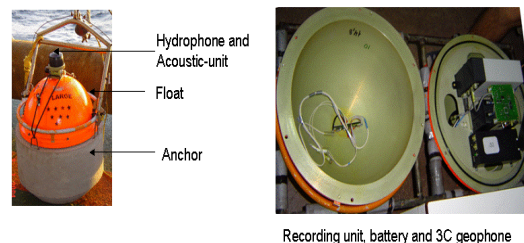


Fig. 8. Ocean Bottom Seismometer.

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