SEISMIC METHODS

Introduction:

Seismic prospecting is one of the most important techniques of geophysical prospecting. The importance and preference of seismic method is due to its high accuracy, high resolution and great penetration. “Extensive application of Seismic methods is in oil and gas prospecting but they are also employed for site investigations in building large scale structures and other civil engineering projects like determination of depth to bedrock (with respect to construction of large buildings, dams and high ways), delineation of sand and gravel deposits (ground water search) and also in detection of water- bearing fracture zones, etc”. *(Parasnis, 1997)*

“The basic technique of seismic exploration involves the generation of waves and measuring the time required for the waves to travel from the source to a series of geophones planted on the surface of earth after refraction, reflection and diffraction from subsurface horizons. The travel time depends upon the physical properties of rock and the attitude of the beds”. *(Telford et al, 1976)*

“The travel time then enables the position of the discontinuities to be deduced”

(Parasnis, 1997)

Thus, the objective of seismic exploration is to deduce information about the rock, especially about the thickness of the beds, characteristics of the material from the observed arrival time and to a lesser extent from variations in amplitude, frequency, and phase and wave shape.

Seismic Method

The seismic method of geophysical exploration utilizes the fact that elastic waves travel with different velocities in different rocks. The seismic method has three principal applications:

- Engineering seismology.
- Earthquake seismology.
- Exploration seismology. *(Yilmaz,2001)*

Engineering Seismology

The seismic method applied to near-surface studies is known as engineering seismology. *(Yilmaz,2001)*

Earthquake Seismology

The seismic method applied to the crustal and earthquake studies is known as earthquake seismology. *(Yilmaz,2001)*

Exploration Seimology

The seismic method applied to the exploration and development of oil and gas fields is known as exploration seismology. *(Yilmaz,2001)*

In exploration seismology, the two main techniques of seismic methods for exploration are:

- Seismic Reflection Method.
- Seismic Refraction Method.

Seismic Reflection Method

In seismic reflection surveying the travel times of reflected arrivals from subsurface interface between mediums of different acoustic impedance are measured.

Reflection:

It is the process by which came back into the same medium from which it was generated after falling/striking at the boundary between the two mediums, as shown in the fig.2:

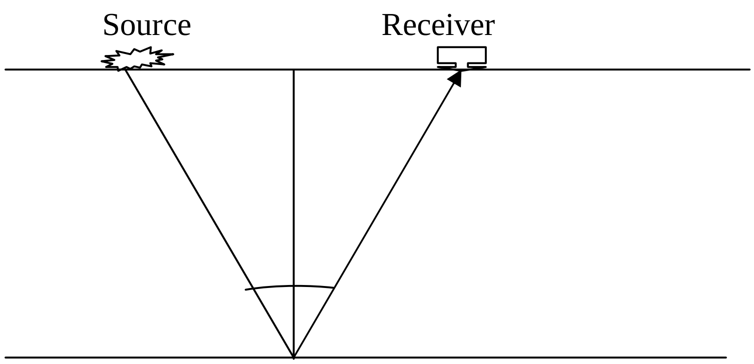


Fig 2: Reflection from an interface between two mediums.

Fundamental Law Of Reflection

This law states that "the angle of incident (between the ray and the normal to the interface) is equal to the angle of reflection (between the reflected ray and the interface)".

Mathematically:

$$\mathbf{i = r}$$

Where,

'*i*' is angle of incident.

'*r*' is angle of reflection.

Seismic Refraction Method

The seismic refraction method utilizes seismic energy that returns to the surface after traveling through the ground along refraction ray paths. This method is normally used to locate refractions from separate layers of different seismic velocities.

(Kearey et al, 2002)

Refraction

Process in which when a wave moves from one medium to other medium, both having different acoustic impedance, then wave changes its path (that is either moves towards or away from the normal to the interface at the point of intercept) as shown in the fig. 3.

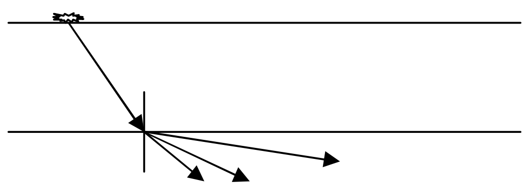


Fig 3: Refraction from one medium to other medium.

Fundamental Law Of Refraction

A wave traversing a boundary between two media of velocity V1 and V2 is such that $\left(\frac{\sin i}{V_1}\right)=\left(\frac{\sin r}{V_2}\right)$

Where
i = incident angle.
r = refraction angle.

Critical refraction:

Every wave from a source in the upper layer when reaches the boundary at different angles of incidence then it continues in the lower layer according to Snell's Law. Now a certain angle of incidence for which angle of refraction is 90° is called as critical angle and refraction at this stage is called as critical refraction as shown in fig. 4.

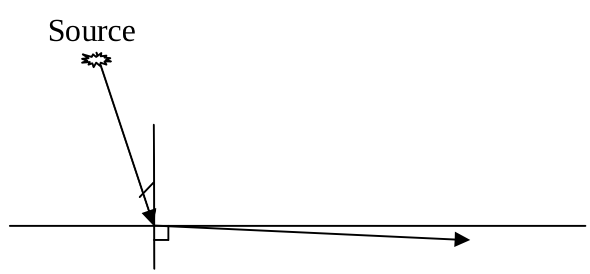


Fig.4: Critical refraction between two mediums.

For critical refraction, Snell's Law attains the form as follow: $Sin(i)_c = \left(\frac{V_1}{V_2}\right)$

Where i_c= Critical angle. *(Robinson &Coruh,1988)*

Diffraction:

Laws of reflection and refraction apply till that the interface is continuous and approximately planar. At abrupt discontinuities in interfaces, or structures whose radius of curvature is shorter than the wavelength of incident wave, the laws of reflection and refraction no longer apply. These discontinuities give rise to a radial scattering of incident seismic energy. This radial scattering is called as diffraction. Diffracted phases are commonly observed in seismic recording and sometime are difficult to discriminate from reflected and refracted phases. *(Kearey et al, 2002)*

Theory Of Elasticity

The seismic method utilizes the propagation of waves through the earth, since this propagation depends upon the elastic properties of the rock. The size of a solid body can be changed by applying forces to the external surface of the body. These external forces are opposed by internal forces, which resist the changes in size and shape. As a result, the body tends to return to its original condition when the external forces are removed. Similarly, a fluid resist changes in size (volume) but not changes in shape. This property of resisting changes in size or shape and of returning to the undeformed condition when the external forces are removed is called elasticity.

Stress:

Stress is defined as force per unit area. Thus, when a force is applied to a body, stress is the ratio of the force to the area, over which the force is applied.

Mathematically:

Stress=F/A (Newton)

Mainly there are two types of stress:

(1) Normal Stress
(2) Tangential Stress *(Heiland,1968)*

so stress can be resolved into normal and tangential stress:

- ***Normal Stress:***

  When force applied is normal to surface of the body then stress is called as normal stress.

- ***Tangential Stress:***

  When force applied is tangential to surface of the body then stress is called as tangential stress.

Strain:

The change in size and shape of the body when external force is applied. These changes are called Strain. It has four types:

(1) Longitudinal Strain.
(2) Transverse Strain.
(3) Shear Strain.
(4) Dilation.

- ***Longitudional Strain:***

  If strain is in length of body, then strain is called as longitudinal strain

- ***Transverse Strain:***

  If strain is in width of body, then strain is called as transverse strain.

- ***Shear Strain:***

  Angular distortion in the body as the result of shear stress.

- ***Dilatation:***

  It is defined as change in the volume per unit volume. If the body has original volume "V" and change is "ΔV", then dilatation, mathematically is given as,

$$\textbf{Dilatation} = \frac{\textbf{Change in Volume } (\Delta V)}{\textbf{Original Volume } (V)}$$ (*Telford et al, 1976)*

Dilation is also called as "Cubical Dilation", and can be defined as "the ratio of change in volume to the original volume before deformation". *(Dobrin,1976)*

Relation between stress and strain

This relation is given by Hook's Law.

- **Hooke's Law**

According to this law, "Stress is the directly proportional to strain provided the elastic limit of the body is not exceeded. This limiting value depends upon the nature of rock body

Mathematically: ***Stress α Strain***

Elastic Module

The linear relationship between stress and strain in the elastic filed is specified for any material by its various elastic modules, each of which expresses the ratio of a particular type of stress to the strain and provides a measure of rigidity. There are certain types of elastic modules as given below.

(1) Poisson's ratio.
(2) Young's modulus.
(3) Shear modulus.
(4) Bulk modulus.
(5) Axial modulus.

Poisson's Ratio:

"The ratio of lateral strain to longitudinal strain" is known as Poisson's Ratio. It is denoted by "σ".

Mathematically $$\textbf{Poisson's Ratio } (\sigma) = \frac{\textbf{Lateral Strain}}{\textbf{Longitudinal Strain}}$$

Young's Modulus:

It is defined as the" ratio between longitudinal stress and longitudinal strain". It is also called stretch modulus. It is denoted by "E".

Mathematically: $$\textbf{Young's Modulus } E = \frac{\textbf{Longitudinal Stress } F/A}{\textbf{Longitudinal Strain } \Delta L/L}$$ *(Kearey et al, 2002)*

Shear Modolus:

The shear modulus is defined as "the ratio of shearing stress "τ" to the resulted shear strain "tanθ".

It is denoted by "μ".

Mathematically: $$\textbf{Shear Modulus } \mu = \frac{\textbf{Shearing Stress}(\tau)}{\textbf{Shearing Strain } (\tan\theta)}$$

(Kearey et al, 2002)

It is also called as *rigidity modulus*. *(Dobrin,1976)*

For liquids and gases, shear modulus **(μ)** =0

(Robinson &Coruh,1988)

Bulk Modulus:

It is defined as "the ratio between hydrostatic pressures "p" applied to a cubic element, the resultant volume strain being the change of volume "ΔV" divided by the original volume "V" ". It is denoted by "K".

Mathematically: **Bulk modulus K = $\frac{\text{Volume Stress P}}{\text{Volume Strain } \Delta V/V}$** *(Kearey et al, 2002)*

Axial Modulus:

It is defined as "the ratio as longitudinal stress to the longitudinal strain in the case when there is no lateral strain, that is when the material is constrained to deform uniaxially". It is denoted by "Ψ".

Mathematically: **Ψ = $\frac{\text{Longitudinal Stress\_(F/L)}}{\text{Longitudional strain(uniaxial) } (\Delta L/L)}$** *(Kearey et al, 2002)*

- **Relationship Between Elastic Moduli:**

  The all four modules can be interrelated in the following way.

$$K = \frac{E}{3(1-2\sigma)}$$

$$\mu = \frac{E}{2(1+\sigma)}$$ *(Dobrin,1976)*

Seismic Waves

ave is a progressive disturbance propagated from point to point in a medium or space without progress or advance by the points themselves.The theory of elasticity reveals that the energy propagated through the earth in the different form of seismic waves. Seismic waves are parcels of elastic strain energy that propagate outwards from a seismic source such as an earthquake or an explosion. *(Kearey et al, 2002)*

Laws Governing The Seismic Waves

There are three fundamental laws that govern the seismic wave propagation.

- Huygen's principle.
- Fermat's principle.
- Snell's law.

Huygen's Principal:

According to this principal, "every point on a wave front is a source of new wave that travels away from it in all directions". *(Robinson & Coruh,1988)*

Fermat's Principal:

It states that "elastic waves travel between two points along the paths requiring the least time". *(Robinson & Coruh,1988)*

Snell's Law:

According to this law "direction of refracted of reflected waves traveling away from a boundary depends upon the direction of the incident waves and the speed of the waves".

Mathematically, $\left(\frac{Sin(i)}{Sin(r)}\right) = \left(\frac{V_i}{V_r}\right)$

Where Vi: Speed of incident wave.
Vr:Speed of refracted wave.
i:Angle of incident wave.
r: Angle of refracted wave. *(Robinson & Coruh,1988)*

Types Of Seismic Waves

Mainly there are two types of seismic waves.

- Body Waves.
- Surface Waves.

Body Waves:

These waves can propagate through the internal volume of an elastic solid. There are two types of body waves.

- P-Waves.
- S-Waves.

Ø ***P-Waves/Compressional Waves/Longitudional Waves:***

These waves propagate by compressional and dilational uniaxial strains in the direction of wave propagation e.g. sound waves. *(Kearey et al, 2002)*

Here motion of the particles of the medium is parallel to the direction of the propagation of wave. The wave path consists of a sequence of alternating zone of compression and refraction. Speed of p-waves is given by:

$$V_p = \frac{(k + 4/3\mu)}{\rho}$$

Where, K =Bulk Modulus.
μ =Shear Modulus.
ρ =Density. *(Robinson & Coruh,1988)*

It is to be noted that V α $1/\rho$. Also that it is well known fact that with depth, "ρ" increases. So V should decrease with depth but actually it is not so. It is because of fact that V α k, which also increases with depth. Here increase in k is much more rapid with respect to the increase in density, that's why V increases with depth.

Ø ***S-Waves/Shear Waves/Transverse Waves:***

Particles of medium moves at right angle to the direction of wave propagation. Theses are slower than P-Waves. Speed of s-waves is given by:

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

Where μ =Shear Modulus.
ρ= Density.

It is well known that for liquids and solids, μ =0, so s-waves can't travel through fluids. S-waves are of two types.

- SV waves.
- SH waves.

i. SV waves:

SV waves are those s- waves in which perpendicular motion of particles to wave propagation motion is in a vertical plane. *(Parasnis,1997)*

ii. ***SH waves:***

SH waves are those s- waves in which perpendicular motion of particles to wave propagation motion is in a horizontal plane. *(Parasnis,1997)*

Ø ***Relationship Between Vp And Vs:***

Both Vp and Vs are interrelated as

$$\frac{V_p}{V_s} = \frac{K}{\mu + 4/3}$$ *(Robinson & Coruh,1988)*

For most consolidated rocks Vp / Vs is between 1.5 -2.0. *(Dobrin,1976)*

Ø ***Characteristics Of Body Waves:***

a) These travel with low speed through layers close to the earth's surface as well in weathered layers. *(Robinson & Coruh,C,1988)*

b) Frequency of body waves in exploration vary from 15Hz to 100 Hz. *(Parasnis,1997)*

Surface Waves

These waves travel along the surface of a medium or closer to the border between two different mediums. In a bounded elastic solid, surface waves can propagate along the boundary of the solid. The amplitude of these waves is greatest at the surface; at depth amplitude of these waves is only a fraction of what is at the surface. *Khan, 1988)* Frequency of surface waves is less than 15Hz. *(Parasnis.D.S, 1997)*

There are two types of surface waves.

- Rayleigh Waves.
- Love Waves.

i. ***Rayleigh Waves:***

These are also called as *"ground roll"* .These waves travel along the free surface of a solid material. The particle motion is always in a vertical plane, elliptical and retrograde with respect to the direction of propagation.

These are the low velocity, low frequency surface waves which often obscure reflections on seismic record obtained in oil exploration. *Dobrin,1976)*

Rayleigh waves involve shear strain. *(Kearey et al, 2002)*

ii. ***Stoneley Waves:***

These are analogous to rayleigh waves. These are also the surface waves which propagate under the special conditions along the interface between two media. *(Parasnis,1997)*

iii. ***Love Waves:***

These are surface waves which are observed only when there is a low- speed layer overlying a higher speed substratum. The wave motion of these waves is horizontal and transverse. Speed of love waves is equal to shear wave speed for following cases.

(1) In upper medium for very short wavelength.
(2) In lower medium for very long wavelength.

These are seldom recorded in seismic prospecting because particle motion is always horizontal. These are similar to SH waves. *(Dobrin,1976)*

Wave Guides, Guided Waves And Head Waves

Along with the reflected, refracted and diffracted there are certain terminologies for waves described as below:

Wave Guides:

These are the low velocity layers in which p and s waves undergo repeated reflections from the base and upper surface of this low velocity layer when these waves are incident on the base of this low velocity layer at angle greater than the critical angle. These layers are closer to the surface. *(Robinson&Coruh,1988)*

Guided Waves:

These are waves confined to a particular wave guide (low velocity layer) such as water layer, coal seam, etc. Velocity of these waves depends upon the wavelength and layer thickness, as well as on density and elastic constants. *(Parasnis,1997)*

Head Waves:

Waves entering or leaving a velocity layer at the critical angle, are called as head waves. *(Khan.Z.R,1988)*

Wave's Conversion

As a wave reaches the boundary between two substances having velocities, it divides up into waves that reflect from the boundary or refract across the boundary. So an incident wave is converted into reflected and refracted waves.

Now an incident wave can be

- P-wave.
- SV- wave.
- SH-wave.

When incident wave is P-wave then it is reflected and refracted as P-wave and S-wave. When incident wave is SV-wave then it is reflected and refracted as P-wave and SV-wave. When incident wave is SH-wave then it is reflected and refracted as SH-wave. *(Robinson.E.S&Coruh,1988)*

Attenuation of Seismic Waves

Attenuation is simply the fall of energy of a wave with increase in distance from source. The energy of a wave in a given medium is directly proportional to the square of its amplitude. *(Dobrin,1976)*

all of energy can be due to

(1) Spherical Divergence.
(2) Absorption.

Spherical divergence is spread out of a wave from its source. As the source wavelet travels farther and farther from the source, its amplitude of vibration grows smaller. It is because of the fact that the continuously expanding spherical wave as expands, the same amount of energy, once received from source at time of onset of wave, has to be distributed over the larger area. Change in amplitude of a wave with distance due to spherical divergence is given as $\mathbf{H=H0/X}$

Where H: amplitude at distance x.
H0: initial amplitude of the wave as it left the source.
X: distance from source. *(Robinson&Coruh,1988)*

Absorption is simply the capture of energy of the wave by particles of the medium through which it propagates. It happens when particles of the medium start to vibrate, due to wave propagation, particles start to collide and so to rub each other. Due to this friction, some of the wave energy is converted into the heat energy. In this way, energy of wave propagating through the medium is decreased and amplitude decreases. Change in amplitude due to this absorption is given as $\mathbf{H=H0*e^{-\alpha x}}$

Where

α: absorption coefficient.

x: distance from source *(Robinson &Coruh,1988)*

Reflection And Transmission Coefficient:

These coefficients compare the amplitudes of an incident wave and a reflected or refracted wave.

Reflection coefficients are those which compare the amplitude of incident wave and reflected wave. If Ai is the amplitude of incident wave and Ar is amplitude of reflected wave then reflection coefficient R is given as follows:

$$\mathbf{R=Ai/Ar}$$

Value of reflection coefficient varies from -1 to+1.*(Khan,1988)*

For R=0, there will be no reflection, wave will be transmitted.*(Kearey et al,2002*

Transmission coefficient is those which compare the amplitude of incident wave and refracted wave. If Ai is the amplitude of incident wave and At is the amplitude of transmitted wave, then transmission coefficient T is given as follows:

$$\mathbf{T=Ai/At}$$

Value of transmission coefficient vary from 0 to2. *(Khan,1988)*

SEISMIC `REFLECTION METHOD

Introduction

This method was developed in 1930's as a tool for oil and gas exploration in sedimentary basins. It is based on reflection of incident waves from an interface. Reflection occurs as horizontal strata commonly contain abrupt changes in seismic velocity and density. That is strata are not homogeneous with respect to the acoustic impedance. Metamorphic and igneous rocks are commonly acoustically homogeneous as compared to sedimentary rocks. These also differ in the sense that these are structurally complex; that is stratification if present is not horizontal.

Reflection method is more preferable as it shows finer details of structure and stratigraphy. In reflection method, those reflections are most useful that arrive nearly vertically, at a distance less than the critical distance. It is because at large distances, direct and refracted waves often interfere with reflections. That's why reflection spread lengths approximate the depth of deepest reflector of interest. *(Lillie,1999)*

Importance Of Seismic Reflection Surveying

It is important because of the following facts.

It gives data which are directly interpretable in terms of geologic structures.

Results are nearly of the same precision for a large range of depths.*(Al-Sadi,1980)*

The Noise Test And Experimental Shooting

Before commencing a seismic reflection survey it is customary to carry out certain shooting procedure for the purpose of determining of dominant noise in the area. One importance of this activity is that it can be used in determination of optimum shooting parameters (source size, no of holes, hole depth and shot at or between the pickets) *(Lillie,1999)*

The Noise Test:

Noise test involve the use of a special short spread for the determination of the apparent wavelength of the coherent noise which is dominant in the area. The spread consists of multigeophone recorders which are interspaced by some small distance.

For better noise reception, the geophones connected to one recorder may be laid down in the form of linear arrays perpendicular to the spread. Here spread itself may be made L-shaped to detect the broad side noise, if any .In both ways we can perform two operations:

١. To move the geophone-spread along the profile for a fix shot location.
٢. To fix the geophone-spread and to move the shot point along the profile relative to the fixed spread. *(Lillie,1999)*

Experimental Shooting:

Experimental shooting is sometimes used to decide the shooting pattern and the trace configuration as an aid /even replacement to the noise test. *(Lillie,1999)*

Control Of Noise In Field:

In order to control the noises in the field following tools are used:

- Source size.
- Source depth.
- Electronic filtering.
- Source arrays.
- Receiver arrays.
- Common depth point/midpoint method. *(Robinson&Coruh,1988)*

Three parameters source size, source depth and electronic filtering are decided on the basis of the noise test and experimental shooting surveys performed prior to the reflection survey.

i. ***Source Size:***

It is not an effective tool, because doing this reflection amplitudes are also diminished as well as undesired ground rolls. *(Robinson & Coruh,1988)*

ii. ***Source Depth:***

As ground rolls travel along the low-velocity weathered zone, so increasing the source depth up to weathered zone, we can overcome the problem of ground rolls. This technique can't be applied in vibroseis seismic survey. (*Robinson &Coruh,1988)*

iii. ***Electronic Filtering:***

It is used in the case in which the ground roll frequency differs from that of reflected waves. A high pass filter is useful in this case. *(Robinson &Coruh,1988)*

iv. ***Source Arrays:***

In order to overcome the problem that ground rolls have frequency same as that of the reflections source array is used. Using the source arrays, a certain number of charges are detonated simultaneously along a line at different positions.

The horizontally traveling surface waves from these sources interfere destructively with one another. At the same time, the wave traveling downward along reflection paths have nearly coincident wave fronts. They interfere constructively to produce a stronger reflected signal.Here wavelength of ground rolls, obtained from noise test and experimental shooting survey, is to be encountered for the selection of source spacing. *Robinson &Coruh,1988)*

v. Receiver Arrays:

In order to overcome the problem that ground rolls have frequency same as that of the reflections source array is used. In receiver arrays several geophones were attached to one take out point. As Raleigh waves spread out horizontally from the source, distorting the land surface much like the ripples. For geophones placed on the surface, some of them will move up while some move downward.

Although each geophone will record high-amplitude ground motion, yet the output from those rose up act to cancel the output from those have moved down. The collective output reaching the recording channel is reduced.

For a sure suppression of ground rolls, we have to place several geophones within a distance equal to the wavelength of ground rolls obtained from noise study of the area.

(Robinson &Coruh,1988)

vi. Common Midpoint Method:

It is most commonly used seismic reflection survey technique. In this method, seismic traces from different shots have a common point on the surface, midway between source and receiver pairs. It is used to enhance signal to noise ratio. Signal to noise ratio is enhanced because we have repetition of reflections from a same point on the reflector. So noise being with property that it changes its phase from trace to trace can be easily identified and removed from the data. *(Lillie,1999)*

Horizental Reflector,Reflection Arrival Time And Layer Thickness:

Let's consider a single horizontal layer in which the seismic waves velocity is "VI" and its thickness is "H", which lies above another material in which the wave velocity is "V2". In the reflection method, this boundary is called as reflector.

The wave emitted from source "S"and reflected from reflector to receiver "R"at a distance "X" from source is shown in the figure 5 given below.

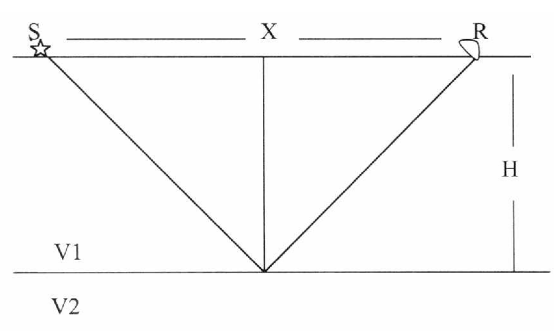


Fig. 5 : Reflection from a horizontal reflector .

As in this case the angle of incidence is equal to the angle of reflection. Due to this the down going and up going path lengths are also equal. Let "r" be the path length from source to reflector. So total path length is 2r. The travel time tx along the whole path in medium of velocity V1 is \
$t_x = 2r/V_1$ and $r = [(X/2)^2 + (H)^2]^{1/2}$

Where 'X' is the offset distance between the source and the receiver and H is the depth to the reflector. Hence $t_x = 2 [(X/2)^2 + (H)^2]^{1/2} / V_1$

Squaring both sides $t_x^2 = X^2 / V_1^2 + 4H^2 / V_1^2$

Dividing by 4H2 and rearranging, we get $t_x^2 / (4H^2 / V_1^2) – X^2 * 4H^2 = 1$

As "H" and "V1" are the constant properties of structure, so above equation is an equation of hyperbola, which is symmetric about X=0.

This result explains why the reflection arrival times develop a hyperbolic curve on t-x graph. So travel time t_x varies with receiver distance X according to a hyperbolic curve. If the source and receiver are at the same location; X=0, then $t_x = t_o$ to=2H/V_1

Where "to" is called as zero off-set time, it is the travel time of the reflected waves along a vertical path just below the source. Now we can express layer thickness "H" in term of "to" and "VI" as follows $H = V_1 * t_o/2$ Substituting this value in

$t_x^2 / (4H^2 / V_1^2) – X^2/4H^2 = 1$, We get $t_x^2 / t_0^2 – (X^2 / t_0^2 V_1^2) = 1$

The above relationship again shows that the reflection arrival time varies hyperbolically with distance. *(Robinson & Coruh, 1988)*

Normal Move-Out Time

The reflected waves travel time "tx" is the sum of zero off-set time "to" and "Δt"; the additional time taken by the reflected wave because the receiver is off-set a distance "X" from the source. The time increment "Δt" is called as normal move-out (NMO) time. So we can express travel time as $t_x = t_O + \Delta t$

That is more convenient form of travel time for analyzing reflection data. Normal move-out time can be derived by using the value of to=2H/V1 in equation given below $t_x = 2 [(X/2)^2 + (H)^2]^{1/2} / V_1$ So we get $t_x = [t_o^2 + (X/V_1)^2]^{1/2}$

Now, on solving this equation by binomial expansion method we get following relation for normal move-out time. $\Delta t = X^2 / 2t_o^2 V_1^2$

This relation shows that NMO decreases with increase of depth as in this case "to" increases. The normal move-out time is subtracted from travel time.NMO is also called as Dynamic correction. After applying the dynamic correction, the reflections from the reflector will align up along a straight line. *(Robinson &Coruh,1988)*

Dipping Reflector

According to the basic law of stratigraphy,deposition occurs in a horizontal way. A reflector is not horizontal usually due to the tectonic forces. A dipping reflector is shown in fig.6 . The time-distance equation for a dipping reflector having dip "θ" is given as, $t_x = (4z^2+x^2+4x\ z\ \sin\theta)^{1/2} / V_1$ *(Robinson &Coruh,1988)*

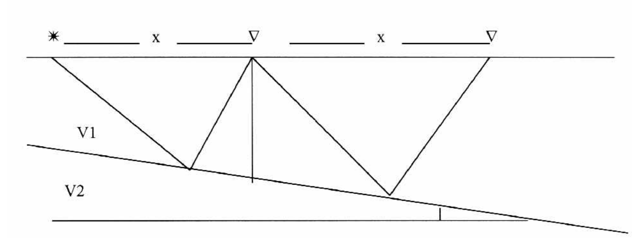


Fig. 6: Reflection from a dipping reflector

Identification Of Reflections On Shot Record

Reflections on the shot record are identified by their hyperbolic travel times. If reflecting interface is horizontal, then apex of the reflection hyperbola is at zero offset. If reflecting interface is a dipping interface, then the reflection hyperbola is skewed in the up dip direction. *(Yilmaz,2001)*

CDP Shooting Or Multifold Reflection Profiling

In seismic reflection survey, it is difficult to recognize the weak reflected pulses on a seismogram. A very common practice used to enhance these weak pulses is multifold reflection surveying.

These multifold reflections are obtained by combining many reflections from the same point on the reflector that where recorded separately with different source-receiver offsets. The different source-receiver combinations have reflection paths reaching the same point on the reflector. This point of common reflection is called the "Common depth point". Because of the different source-receiver offsets, all the reflection travel times will be different.

The common depth point coverage is important due to complicated structures and complex velocity distributions in the subsurface. CDP shooting provides redundant reflection data by use of multiple source-receiver arrangements per trace.

CDP Stack:

The process of adding together the adjusted traces is called the stacking. The CDP traces are combined together to obtain enhanced primary reflections. Common depth point stacking is effectively used for attenuation for multiple reflections. Stacking attenuates or even totally suppresses the long path multiples that have a significantly different move out from primary reflections.

Applying the normal move-out adjustment, we obtain corrected seismogram, such that reflections on each trace should appear at identical times. Other pulses that tend to obscure the weak reflections are not adjusted to identical times. Therefore by adding all the traces together to form a single composite trace, we hope to obtain a large pulse from the sum of the reflections at the identical times, the other pulses would act to cancel one another, Now we have obtained a multifold composite trace. The name multifold come from the fact that we have a multiple traces combination and that making the normal move-out adjustment we are in a sense, "folding" the down going and up coming parts of each path into a vertical path. We can assign a fold number that tells how many individual traces were used to obtain the composite.

The fold of the stacking refers to the number of traces in the CDP gather and can be expressed as a percentage. Mathematically, it is given by

$$\text{No of Folds} = N * \triangle X/2 * \triangle S$$

Where

N: the number of recording channels along a spread.

$\triangle$X: the geophone interval.

$\triangle$ S: the source interval. *(Robinson & Coruh,1988)*

SEISMIC DATA ACQUISITION

Introduction

Fundamental purpose of seismic data acquisition is to record the ground motion caused by a known source in a known location. The initial stage of reflection survey involves field trials in the area to determine the most suitable combination of source, offset, recording range, array geometry and detector spacing to produce good seismic data. The quality of field data can be achieved if field parameters are selected appropriately.

Recording shooting parameters selection is carried out in the field by conducting different experiments. The recording parameters selection depends upon four variables. These variables include source size, source depth, no .of holes and group base. By changing these variables, the best combination is selected for recording and shooting parameters.

Recording Parameters

The recording parameters include,

- Group interval
- Group base
- No of channels
- No of geophones in a group
- Sample rate
- Record length
- Coverage (folds)
- Near offset and far offset

i. Geophone Interval:

It is the spacing between two consecutive geophones.

ii. Group Interval:

The interval between the mid points of two consecutive groups.

iii. Group Base:

It is the total length of various groups that collectively feed a single channel. Group base would act in the way for the suppression of noise Let

Geophone base = 70m
Number of geophones in groups = 24

Geophone interval = 70/23m
= 3.04m

iv. Sample Rate:

In digital recording the time during which discrete sample is recorded.

v. Record Length:

The total length of time for recording on shot is called record length.

vi. Coverage:

The data will be of (n) fold coverage if the seismic rays strike the same subsurface point (n) times and then detected to the receivers.

Here 4 fold coverage is shown:

```
* ▼ ▼ ▼ ▼ ▼ ▼ ▼ ▼
    * ▼ ▼ ▼ ▼ ▼ ▼ ▼ ▼
        * ▼ ▼ ▼ ▼ ▼ ▼ ▼ ▼
            * ▼ ▼ ▼ ▼ ▼ ▼ ▼ ▼
                * ▼ ▼ ▼ ▼ ▼ ▼ ▼ ▼
```

vii. Near Offset And Far Offset:

Due to direct arrivals 2 or 3 channels are kept dead during recording , the minimum distance at which first active channel exists is called near offset, and the distance at which last active channel exists is called far offset.

Shooting Parameters

Shooting parameters include

١. Source size
٢. No of holes
٣. Holes depth
٤. Shot at or between picket

Energy Sources

Energy sources are designed to generate vibrations in the ground. There are different sources of energy, having different level of frequencies. Many considerations (lithology, are accessibility & cost) govern the selection of suitable seismic source for a particular survey. Sources of energy are given as:

- Dynamite
- Vibroseis
- Geograph (dropping of weight)
- Dinoseis (b exploding gas mixture)
- Geoflux (by exploding cord burned in the ground)

i. Dynamite:

This is the most common energy-source used in seismic prospecting. Normally it is exploded inside a drilled hole at a depth ranging from a few meters to several tens of meters. The deeper the charge the less intensive the generated surface waves are. Since the weathered surface layer absorbs high frequency components, it is always advisable to place the charge below the base of this weathered zone. As a field procedure the charge is usually sealed with water or mud to increase charge-coupling with the surrounding medium. The amount of charge per shot point depends on whether it is a single hole or a pattern (multi-hole) shooting. As a rough estimate, the charge weight is 10 -50kg of dynamite.

ii. Geograph:

The Geograph method (also known as the *Thumper*) involves dropping a weight of about 3 tons from a height of about 3 m onto the ground. Since the generated energy is weak in comparison with dynamite shooting, *vertical stacking* is carried out. Thus, the records of 30 to 50 drops made at the one location are stacked together for signal enhancement. The effective generated energy depends on the strength of the impact and the nature of the surface material. Compared with the other type of energy sources, the Geograph is a high frequency source, very safe, fast, in operation and cheap. The drawback of the method is the development of strong surface waves, because of the filtering effect of the surface layer, the high frequency components are severely attenuated. This lowers the resolution capability of the method.

iii. Dinoseis:

Instead of the freely falling weight used in the Geograph method, the weight in this method is power- driven against the ground surface. The impact energy is generated by exploding a gas mixture (propane and oxygen) contained in a chamber, the bottom of which is a moveable plate. The generated pressure-impact on the plate provides the necessary seismic energy. The general characters of the Dinoseis technique are much the same as those of the Geograph. The method surfers from the strong development of surface waves and from poor reflection resolution. As is the case with the Geograph, vertical stacking is carried out in the field the purpose of enhancing the signal-to-noise ratio.

Geoflex:

The seismic source consists of an explosive cord which is buried in the ground at a shallow depth. It is laid down by a hydraulically-operated plough which is especially designed for the purpose. The main advantages of this method are the speed with which the explosive cord is laid down in the field and certain attenuation of the generated surface waves. However, it required specially favourable field conditions such as soft ground which is penetrable by the plough and open country. It is not adviseable to use Geoflex in areas where the surface layer is too thick and unconsolidated, since in such circumstances; the transmitted energy will be much weakened.

iv. Vibroseis:

This method is based on the use of a mechanical vibrator which is hydraulically or electrically driven to exert a force of an oscillating magnitude. The source function (called the sweep) consists of force pulses generated at a frequency which is increased linearly with time from about 6Hz to 60 Hz during a time span of say 6 --16 seconds. Thus, the seismic source is a defined function as regards its amplitude and phase spectra. Apart from the linear taper imposed at the beginning and at the end of the applied sweep, the amplitude is constant during the sweep lapse- time. In other words, the sweep has a white amplitude spectrum over the non-ramped frequencies.

Since the source is active during a time interval of a defined length (sweep duration), each reflection has a record length which is equal to sweep duration. Thus, the obtained record represents the superposition resultant of all the wave trains which correspond to the individual reflections. To extract the actual reflection time, the recorded signal is cross-correlated the applied weep.

The main problem met with in Vibroseis work is the low energy-level of the source. Improvement of the signal-to-noise ratio is effected through vertical stacking which is normally done in the field.

Comparison Between Dynamite And Vibroseis

- Dynamite contains a wide range of frequencies, while Vibroseis contains limited frequencies.
- Dynamite gives a shot impulse, while Vibroseis gives a sweep.
- Dynamite is uncontrollable, while Vibroseis is controllable.
- Dynamite is costly,i.s. great cost and time to drill the hole, while Vibroseis is cheaper.
- Dynamite can,t be used in urban areas, while Vibroseis can be used in urban areas.
- Dynamite is dangerous for health, safety and transportation point of view, while Vibrates is not

Instruments Of Seismic Surveying

Seismic surveying consists of placing some receivers at different locations and then using them to detect vibrations produced by an energy source. The receivers convert the mechanical vibrations into electric current that is transmitted to a recorder, the recorder is designed to preserve, the information in a form that can be displayed and analyzed. Seismic surveying can be done on land or at sea, We will displayed and looking at the instruments that are available for seismic surveying.

These include the receivers, the cable, or other

Type Of Shooting

- Symmetric shooting
- Asymmetric shooting
- End shooting
- Roll in / Roll out shooting

i. Symmetric Shooting:

This type of shooting has equal numbers of channels on both sides of the source.

ii. Asymmetric Shooting:

No of channels are not equal on both sides of source.

iii. End Shooting:

In this type of shooting the channels lie on only one side of the source. This method of shooting is used in multifold profiling.

iv. Roll In / Roll Out Shooting:

If we have 240 channels and want to use roll in and roll out shooting method then left side of the shot point have no channel while on right side there will be 120 active channels for first shot, 121 active channels for second shot, 122 active channels for third shot and so on. This process is called roll in shooting. For shot point no 120 there will be equal no. of channels (i.e.120) on both sides of the shot point. At this stage it is symmetric spread. Later on at shot point no. 121 the roll out process starts.The shooting crew submits a daily report of shooting. For example Line no, shot point, No. of holes, Depth of Explosive and No of detonators. **09 321-25R 120m 7kg3, 322-IOR 121m 7kg 3 323-17R 121m 7kg3**

v. ***Selection of Shooting And Recording Parameters:***

Four variables are used in the selection of these parameters. These variables include:

- No of holes
- Hole depth
- Charge size
- Group base

The source and receiver are selected by following method

- By changing charge size, keeping no. of holes and hole depth constant.
- By changing no. of holes, keeping charge size and holes depth constant.
- By changing depths of hole, keeping charge size and no. of holes constant.
- By changing group base, keeping charge size, no. of holes and holes depth constant.

In 2D reflection survey the source and receiver will be on same line. Group of geophones are arranged in such a way that they give best response. The arrangement of source and receiver is called spread. Spread. is designed according to the target depth and fold coverage of the interested area.

The pattern in which the geophones are placed for different shots along a seismic line is called geophone layouts.

Recording Instruments

i. ***Geophone:***

Geophone is a device used in seismic data acquisition to detect ground vibration. Geophone consists of a magnetized mass fixed to the container, surrounded by a wire coil suspended on spring. When the ground vibrates, the coil moves back and forth around the magnet and induces electric current, which is proportional to the motion of ground.

Before taking the geophones to the field each group is tested in the geophone laboratory by SMT 100 and others methods for following test.

- Polarity (normal / reverse)
- Leakage/ signal leakage
- Shunt resistance
- Proper connection / cut of wire
- Continuity of geophone
- Resistance of the group
- Natural frequency of geophone
- Damping of geophone
- Sensitivity of geophone

The geophone signal is transmitted to recording system by means of seismic cable. A geophone is connected to the cable on a point called" takeout".

§ ***Geophone String:***

Normally there are 12 geophones is single group. There may be one more geophone groups at the station. These 12 geophones in geophone group are connected in such a way that each six geophones are connected in series and then these two pair of six geophones are in parallel.

Resistance of single group can be calculated as:

Resistance of six geophones in series = 272.72*6 = 1636.36

Total resistance of geophone group 1636.36/2818.18 Geophone spacing depends on the noise characteristics of the area.

Type Of Arrays

i. ***Surface Waves:***

This type of array has in line geometry in such a way that geophones are uniformly spaced and has same weight. The geophones are laid out in a straight line.

ii. ***Weighted Array***

When sensitivity of each individual geophone is made different, it is called weighted array.

iii. ***Anal Array:***

Geophones are laid out over an area around the line.

Shot And Geophone Arrays

In the seismic reflection profiling we use group of geophones rather than individual geophones at each station for the cancellation of noise. Each group consists of 12 geophones. The output that goes in to amplifier represents the average ground motion over the group. The basic principle is to design groups so that waves traveling vertically or nearly vertically are reinforced while those traveling horizontally are reduced.

Consider and example of a group of four geophones that covers horizontal distance equal to a wavelength of the surface wave to be cancelled. With this arrangement, at any time the horizontal wave will cause upward motion in two detectors of each group and downward motion in other two. If all four are connected in series, the net signal from this group will be zero because of cancellation as shown in the figure below.

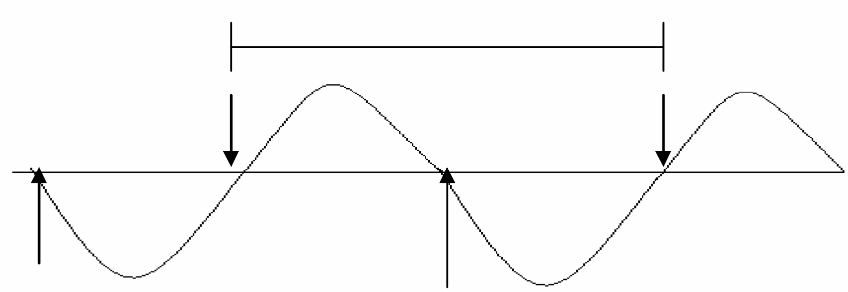


Fig. Eliminating effects of ground roll by use of multiple geophones in series

The low velocity, low frequency surface wave with high amplitude are called ground roll, as these waves have low frequencies, than those of reflections, low cut filters are introduced in to the amplifier circuits to eliminate ground roll. Therefore proper grouping of geophones will reinforce reflected events and cancel horizontally traveling noise.

Determination Of Geophone Spacing

To determine the geophone spacing for a given no. of phones per trace in an area the noise test is necessary. These tests are designed to find the nature of the noise, whether coherent or incoherent. And if the noise is coherent then the range of wavelength of the noise is determined. A series of records is made with geophones planted in the area. This type of data set is often called “Noise profile”.

Noise Profile

The noise profile gives us noise pattern of the seismic area. While recording noise profile, at each station bunched geophones are used. Noise profile shot should be on seismic line. Shots can be from 5-10 and no of channels may be from 24-48. Noise profile should be in the direction of line.

Types Of Seismic Noises

All types of disturbances that interfere with the signal of interest are called noises and the interested pulse is called signal. Noise is dived in to types Coherent noise.

- **Incoherent noise**
- **Random Noise**

Ø ***Coherent Noise:***

The noise that shows a regular pattern in the seismic monitor. This is also called source generated noise. This type of noise is often confused with the signal. Sources of coherent noise includes

- Multiple reflections
- Refracted waves
- Surface waves
- Gonad roll
- Direct waves

Suppression Of Coherent Noise:

This type of noise can be suppressed by the ideal field procedure i.e. by selecting proper geophone spacing. To remove ground roll the charge should be buried at appropriate depth.

Ø ***Incoherent Noise (Random Noise):***

Random noise shows no systematic pattern on the seismic monitor. This noise is common when the shot point overlies or close to gravel, boulders & all of which can cause scattering of waves.

Sources Of Random Noise :

- Wind noise
- Rain noise
- Traffic noise
- Small movements with in earth
- Bad geophone noise

Suppression Of Incoherent Noise In The Field:

The basic tools available for controlling noise in the field are

- Source depth
- Receivers arrays
- Multiple converge
- Electronic filtering
- Better contact of geophone to earth
- Burry geophone to avoid noise
- Avoid power line

Spread Design and its Types

The layout of geophones on the surface that gives output for single source is called spread. Spread is designed with respect to shooting procedure

There are a no of configurations used in reflection profiling,

End Spread:

In this spread the shot point is located on one side of the spread & geophones are arranged in a straight line as shown in the graph part Figure.

In line offset spread

It is a modified form of end spread but the shot is located some distance away from the first geophone as shown in the graph part figure.

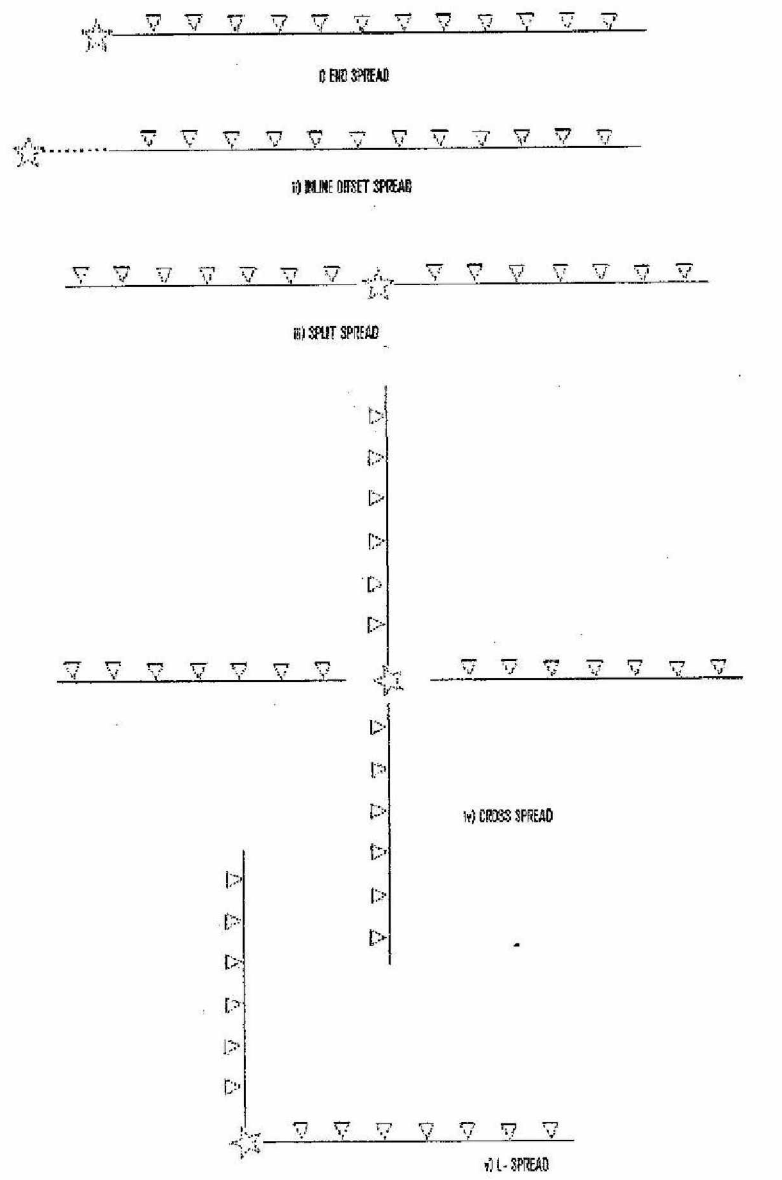


Fig 5.1: Different Configurations used in the field

Split spread. The geophones are arranged on both sides of central shot point as shown in the graph part Figure.

Cross Spread:

The geophones are arranged in cross shape around the central shot point as shown in figure.

L-Spread:

The geophones are arranged around a shot point as L shape as shown in the graph part figure.

Seismic Data Recording

The conventional seismic survey procedure is to monitor ground motions at a large number of surface locations, thus multi-channel recording systems are employed with upto several hundred separate recording channels. The signals are received with a large number of geophones (geophone groups) spread along a line extending from the shot point for a distance of thousands of feet. Each group transmits its data to the recording instruments on one information channel, by means of the seismic cable.

Analogue Recording:

The seismic signals, as generated and detected, are in analogue or continuous from and they must be amplified and recorded in reproducible form. The analogue recording is constrained by its small dynamic range. In order to limitize the signal in the small available dynamic range of the analogue recording system, negative feed back amplification is employed (Telford et al, 1990), which at one hand amplifies the weak signals and at other hand suppresses the strong signals. There are two methods available for the purpose, one of which is automatic gain control (AGC).AGC measures the average output signal level over a short interval and adjusts the gain to keep the output level almost constant regardless of the input level. Time variable gain is the other method used that suppresses the gain anitial level and enhances the gain at later part of the recording when seismic signals have low amplitude.

Digital Recording:

The digital recording represents the signal by series of numbers denoting geophone output at regular intervals of time. The digitization of signal involves loss of frequencies higher than half the sampling frequency, known as aliasing. An anti aliasing filter is, therefore, applied before digitization to suppress these higher frequencies.

The analogue output of the seismic amplifiers is digitized by an A/D (analogue to digital) converter by means of multiplexing. All seismic signals recorded digitally are registered in the form of binary numbers, each representing a sample value for an individual channel. The required switching rate of A/D converter is determined by the required digital sampling interval and number of channels to be multiplexed.

The multiplexed digital data are recorded in a standard tape format. This formatting process involves distribution of the bits of each binary word on a defined number of information tracks.

Multi-channel Reflection Surveying And CDP Profiling

Applying the basic requirement of a multi-channel reflection survey is to obtain recording of reflected pulses at several offset distances from a shot point. Reflection profiling is normally carried out along profile lines with shot point and its associated spread of detectors being moved progressively along the line to build up lateral coverage of the underlying geological section. The two most common shot detector configurations are the split spread (or centre shooting) and the single ended or end on shooting). The length of reflector sampled by any detector spread is half there spread length. The common depth point coverage is important due to complicated earth structures and complex velocity distributions. CDP shooting provides redundant reflection data by use of multiple source receiver arrangements per trace, this equals to multi offset coverage of the same reflection point. The usual technique for CDP survey is CMP gathering in which shot points and receivers location have a common mid point (CMP) below which the common depth point is assumed to lie. In theory, the above assumption is true only for planar flat lying reflectors with no horizontal velocity variations. The effect of difference in source receiver offsets is remover by normal move out (NMO) correction. This correction also convinces with information regarding both horizontal and vertical velocity variations. After applying NMO correction, traces of a common mid point gathers are summed algebraically, this process is known as stacking.The fold of the stacking refers to the number of traces in the CDP gather and can be expressed as a percentage. Mathematically, it is given by **No of Folds (N) = 2n**

Where N is the number of channels along a spread and n is the number of channels the profile is moved ahead, also known as move-up rate. Stacking attenuates or even totally suppresses the long path multiples that have a significantly different move out from primary reflections, thus when the later are stacked in phase, the former will be added out of phase and will be partially cancelled.

Display Of Seismic Data:

CDP profiling data from two dimensional surveys are conventionally displayed as seismic sections, in which the individual stacked seismograms are plotted side by side, with their time axes arranged vertically. There are different methods of visualization of traces e.g. variable area, variable density, wiggly line etc *(Rehman, 1996)*

SEISMIC DATA PROCESSING

Introduction

Seismic data are recorded in the field on magnetic tape and later this information is transmitted in to a seismic section. All of the intervening steps comprise the data processing phase of seismic exploration. Data processing is an approach by which the raw data recorded in the field is enhanced to the extent that it can be used for the geological interpretation. *(AL.Sadi,1980)*

Data processing is to convert the information recorded in the field into a form that mostly facilitates geological interpretation. *(Dobrin,1976)*

Factors Cotroling The Data Processing Results

Seismic data processing strategies and results are strongly affected by field acquisition parameters. Additionally, surface conditions have a significant impact on the quality of data collected in the field. Lack of seismic reflected events on seismic section is not the result of a subsurface void of reflectors. Rather it is caused by low signal-to-noise ratio resulting from energy scattering and absorption in the medium of propagation.

Surface conditions have an influence on how much energy from a source can penetrate into the subsurface. Besides surface conditions, environmental conditions and demographic restrictions can have significant impact on field data quality.

Other factors that can influence the quality of data are weather condition and condition of recording equipment. In addition to field acquisition parameters, seismic data processing results also depend on the technique used in processing

Processing algorithms are designed for and applied to either single channel time series, individually, or multi-channel time series.

Processing In General

Data processing is a sequence of operation, which are carried out according to the pre-defined program to extract useful information from a set of raw data. As an input-output system, processing may be schematically shown as:

INPUT				**OUTPUT**
OBSERVATIONAL DATA	→	**PROCESSING SYSTEM**	→	USEFUL INFORMATION

(AL.Sadi,1980)

Primary Stages Of Seismic Data Processing

There are three primary stages in Seismic Data Process in, each is aimed at improving seismic resolution(ability to separate two reflection events that are very close together). These three stages in their usual order are:

- Deconvolution.
- Stacking.
- Migration. *(Yilmaz,2001)*

Beside of these primary steps, other processing techniques may be considered as secondary as they help to improve the effectiveness of the primary processes. *(Yilmaz,2001)*

Processing Sequence

Briefly, we can describe the data processing procedure in the following main five categories:

- Data Reduction.
- Geometric Corrections.
- Data Analysis and Parameter Optimization.
- Data Refinement.
- Data Presentation.

Data Reduction

Data reduction is done by certain processing operations as discussed below:

- Demultiplexing.
- Correlation.
- Header generation.
- Display.
- Editing and muting.
- Amplitude adjustment.

Demutiplexing

Field reflection seismic data are now recorded on digital magnetic tape in multiplexed form using a certain type of format. These data first are demultiplexed. Demutiplexing is the process by which we transfer the data from time sequential form to trace sequential form. At this stage data is also converted into a convenient format.

(Yilmaz,2001)

The digital seismic data is recorded on magnetic tape by the recorder in the following way:

A11A21A31………Ai1Ai+1,1…………An1
A12A22,A32………Ai2Ai+1,2………….. An2
A13A23,A33……. .Ai3Ai+1,3……. ….An3
..……………………………………………
……………………………………………
A1mA2mA3m……Aim,AI+1,m…… Anm.

(Robinson &Coruh, 1988)

After that data has been demultiplexed , it is stored on tape in a convenient format in the following way:

A11, A12, A13,……….,A1m
A21, A22, A23,……….,A2m
……………………………
…………………………....
…………………………....
……………………………
An1, An2, An3,……….,Anm

Now we have obtained the data in such a form that it can be used for further processing.

Correlation

Correlation is simply the measurement of similarity or time alignment of two traces. Since correlation is a convolution without reversing the moving array, a similar frequency domain operation also applies to correlation.

There are two types of correlation.

- Cross Correlation.
- Autocorrelation. (*Yilmaz, 2001)*

i. Cross Correlation:

Cross correlation measures how much two time series resemble each other. It is not commutative; output depends upon which array is fixed and which array is moved.

As a measure of similarity, cross correlation is widely used at various stages of data processing .For instance traces in a CMP gather are cross correlated with a pilot trace to compute residual static's shift. It is the fundamental basis for computing velocity spectra. *(Yilmaz,2001)*

ii. Auto Correlation:

Cross correlation of a time series with itself is known as auto correlation. It is a symmetric function. Therefore only one side of the auto correlation needs to be computed. *(Yilmaz,2001)*

(a) Vibroseis Correlation:

This involves cross correlation of a *sweep* signal(input) with the recorded vibroseis trace.The sweep is a frequency-modulated vibroseis source signal input to the ground. There are two types of sweep.

- Up Sweep.
- Down Sweep.

UP SWEEP:

When frequency of the vibroseis source signal increases with time

DOWN SWEEP:

When frequency of the vibroseis source signal decreases with time.
(Yilmaz,2001)

(b) Importance Of Vibroseis Correlation:

For vibroseis source, we have a sweep (a train of waves) rather than a short pulse/source wavelet whereas most seismic impulsive sources generate a very short pulse which can be used directly to examine subsurface structure Vibroseis sweep lasts for several seconds depending upon the sweep time. So in case of vibroseis source all reflected and refracted signals on a vibroseis seismogram overlap one another extensively. Even after demultiplexing of the vibroseis seismogram it is impossible to recognize the reflections. So vibroseis correlation procedure is applied. Vibroseis correlation enables us to extract from each of the long overlapping sweep signals on vibroseis seismogram, a short wavelet much like those obtained with seismic impulsive source.

(Robinson. & Coruh, 1988)

(c) Procedure For Vibroseis Correlation:

In vibroseis correlation, we cross multiply, point by point the digitized values of the vibroseis trace with the digitized values of the input sweep. Then we sum up these products to obtain first correlated value for the correlated seismogram trace. Now we advance one point on the vibroseis trace and repeat the process. We plot these correlated values to obtain correlated seismogram. This seismogram represents a short pulse or wavelet rather than the long sweep/wave train, centered at arrival time for each wave indicated on the original trace. The particular type of wavelet produced by the vibroseis correlation is called “Klauder Wavelet”. These Klauder wavelets are then used to develop the correlogram.

(Robinson & Coruh, 1988)

Header Generation:

After that all of the sample from a given field trace are assembled into an array, a large amount of archival information is placed in a reserved block called a trace header, which is located on the just ahead of the data samples .Trace header information may include location and elevation of source and receiver, field record number, trace number etc .A real header block is also placed at the head of each reel, for recording line number, reel number etc.

Display

At the end of the processing sequence, and sometime at the intermediate stages, it is necessary to reconvert the digitized data to analog format so that they can be displayed and then examined visually. There are several methods of plotting digital data in analog form. The most common method is to pass the digital trace through a sample and hold device, which produce an output voltage proportional to each sample and which holds the voltage constant until the next sample arrives. This procedure produces a “stair step” type of analog output, which can then be smoothened into a continuous trace by their passage through a low pass filter.

Editing And Muting

Due to certain unfavorable field-conditions, some of the recorded data are not useful. Part of a trace, the whole trace, and occasionally the whole shot point record may come out very weak or over-ridden by abnormally high energy events. There may be noisy traces, traces with transient glitches, some dead traces and mono-frequency signals on the record Data editing involves complete removal of all undesired data before proceeding further in data processing .This is done by setting to zero ,the all undesired trace-samples. Some time in the field the electrical connections of some traces are inverted by mistake. As the result, the peak-trough sense of such traces comes out reversed with respect to the rest of the record. This is called as polarity reversal. Rectification of polarity reversal of such traces is an important part of data editing .Otherwise if it could not be done these traces must be zeroed. After doing this all the contributing traces per each CDP are gathered together. Each trace in one CDP is identified by it s shot point and receiver numbers .The CDP-gathers may be displayed as such for direct inspection and checking of edited data .

Muting:

Trace- muting is a special type of data editing. This term is applied for process of zeroing the undesired part (or parts) of a trace. In order to avoid stacking non-reflection events (such as first arrivals and refraction arrivals) with reflection, the first part of the trace is normally muted before carrying out the stacking process .This is occasionally referred to as first break suppression . *(Al-Sadi,1980)*

Amplitude Adjustments

Amplitude adjustment is done to recover the true information present in the data .True information means data is irrespective of effect produced due to wave propagation through the subsurface. It is well known fact that a seismic wave is attenuated as it travels in a non-perfectly elastic medium. Along with this effect, signal is further modified by recording station it .So reflection amplitude recorded in the field is the end-result of the interaction of the following main factors:

- Spherical divergence.
- Inelastic attenuation.
- The net- gain imposed by the recording station.

A grain recovery function is applied on the data to correct for the amplitude effects of wave front (spherical) divergence. This amounts to applying a geometric spreading function, which depend upon travel time, and an average primary velocity function, which is associated with primary reflections in a particular survey area . Gain is applied to seismic data for spherical spreading correction. *(Yilmaz , 2001)*

Often AGC (automatic gain control) is applied to raise the level of the weak signals. AGC attempts to make amplitudes similar for all off sets, for all time and for all mid points. A typical method of calculating the median or average amplitude with in sliding windows down the trace , then to calculate the multiples needed to equalize the median value in all the window. *(Dobrin, 1988)*

In interpretation of seismic section, variations in amplitudes of reflections can be the important factors .Lateral amplitude variations, from trace to trace, within a reflection event (bright spots) may be the direct indications of the presence of hydrocarbons. Vertical amplitude variations, from event to event, may be helpful in identifying and correlating reflecting horizons. *(Khan, 1989)*

Geometric Corrections

In order to compensate for the geometric effects, we have to apply certain corrections on the recorded data .These corrections are called as geometric corrections

.These corrections are applied on the traces gathered during trace editing and muting .The geometric corrections are:

- § Static correction.
- § Dynamic correction.

Static Correction

Static correction compensates the effect of weathered layer and elevation effect due to unleveled surface .So static correction is of two types:

- o Elevation correction.
- o Weathering correction.

For land data, elevation corrections are applied at the stage of development of field geometry to reduce the travel times to a common datum level .This level may be flat or floating along the line. *(Yilmaz,2001)*

Dynamic Correction

Dynamic correction compensates the effect of offset of receiver from the source .It is also related to the shape of the subsurface interfaces .It is also of two types:

- o Normal move out correction (NMO).
- o Dip move out correction.

Normal move out correction is related more to the non-dipping interfaces. On the other hand dip move out correction is related to the dipping reflectors. It accounts for the effect of dip of the subsurface interface along with the effect of offset distance of receivers. *(Robinson & Coruh, 1988)*

Dip-move out correction is applied to data following the normal-move out correction using flat-event velocities. *(Yilmaz,2001*)

Figure 7 is the diagrammatical representation of the concept of static and dynamic corrections.

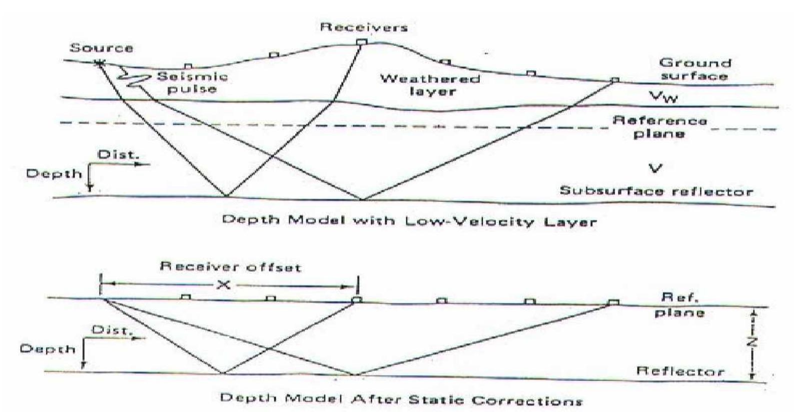


Fig. 7: Diagrammatic representation of static and dynamic corrections.

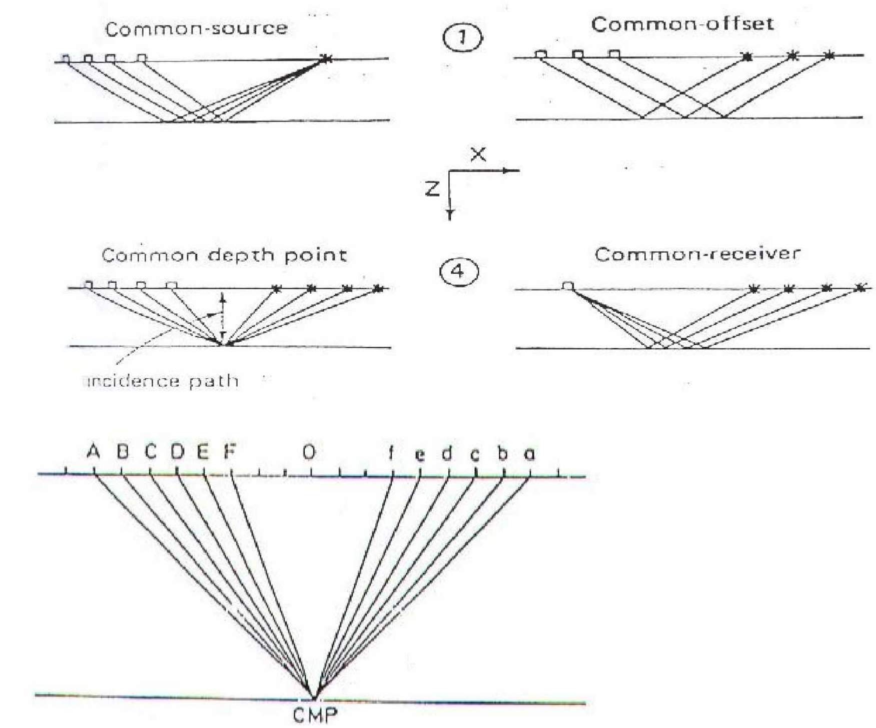
Trace Gathering

Traces are routinely gathered into groups having some common elements. The types of gathers usually made are:

- o Common Source Point Gather.
- o Common Depth Point Gather.
- o Common Receiver Point Gather.

- Common Offset Gather.
- Common Mid Point Gather.

The concept of various types of Trace Gathers is shown in the figure 8 as fallow:



CMP Gathers

Fig.8: Diagrammatic representation of different trace gathers

Data Analysis And Parameter Optimization

Three steps involved in this procedure. These three are:

- Filtering.
- Deconvolution.
- Velocity Analysis.

Filtering:

A filter is a system, which discriminates against some of its input. Seismic data always contain some signal information, which we want to preserve. Everything else is called noise, and we want to remove it. These systems, which are generally called filters work either by convolution in the time domain or by spectral shaping in the frequency domain to remove these undesired informations.

The most common types of filters used are as follows:

- Low pass frequency filter.
- High pass frequency filter.
- Band Pass frequency filter.
- Notch filter.
- Inverse filter.
- Velocity filter.

i. ***<u>Seismic Noise:</u>***

Noise is the undesired information contained on a seismic record which one does not wish to use and are to be filtered .There are two types of seismic noises encountered in seismic survey are.

- Coherent Noise.
- Incoherent Noise.

ii. ***<u>Coherent Noise:</u>***

It displays some regular patterns on seismogram. It is seismic energy which aligns from trace to trace or record to record on seismic record .Often, it is very similar to the signal and usually more difficult to overcome than the incoherent/random noise. By examining the patterns of coherent noise, we can devise field procedures to reduce it. It has various sources as discussed below:

- Ground Rolls.
- Guided Waves.
- Multiples.
- Noise Due to the Air Wave.
- Cable Noise.
- Side-Scattered Noise.
- Noise Due to Power Lines. *(Yilmaz, 2001)*

a. ***<u>Ground Rolls:</u>***

It is the vertical component of surface waves .It is recognized by its low frequency, strong amplitude and low velocity. These are eliminated by receiver arrays. *(Yilmaz,2001)*

b. ***<u>Guided Waves:</u>***

Guided waves make the early arrivals. These are generated because of strong velocity contrast between the water layer and the substratum, trapped within the water layer and the substratum and guided laterally through the water layer. These are attenuated by CMP stacking. (Yilmaz,2001)

c. ***<u>Multiples:</u>***

These are secondary reflections with interbeded ray paths. Guided waves contain multiple energy. These can be attenuated with CMP stack method.

(Yilmaz,2001)

d. ***<u>Noise Due To The Air Wave:</u>***

This noise can cause a serious problem while shooting with surface charges such as geoflux. The only effective way to remove this noise is to zero out the shot gathers along a narrow corridor containing this energy (muting).

e. ***<u>Cable Noise:</u>***

It is linear and low in amplitude and frequency. It appears as late arrivals on shot record. *(Yilmaz, 2001)*

f. ***<u>Side-Scattered Noise:</u>***

It occurs at water bottom, where there is no flat, smooth topography. They can be on or off the vertical plane of the recording cable. *(Yilmaz, 2001)*

g. ***<u>Noise Due To Power Lines:</u>***

This noise appears in the form of mono frequency wave, with a frequency about 50-60 Hz. Filter used to remove/suppress these noises is called as *Notch Filter. (Yilmaz, 2001)*

iii. Incoherent Noise:-

It is also known as Random Noise. It is the seismic energy that does not align up from trace to trace or record to record on seismic record It displays no systematic pattern. This noise is uncorrectable. We can overcome random noises by recording more than one trace from the same location. It has various sources such as given below:

- Wind Noise.
- Transient/Small Movements in vicinity of seismic cable.
- Water Flow Noise.
- Electrical Noise from Instrument.
- Bad Geophone Noise (improperly plotted).
- Short Wavelength Propagation Noise. *(Yilmaz,2001)*

Deconvolution

It is the process by which the wavelet associated with the significant reflections is compressed and reverberatory energy that trails behind each reflection is largely attenuated .It is a filtering process designed to improve resolution and suppress multiple reflections. Deconvolution can be considered either in the time domain or in the frequency domain. In the time domain the object is to convert each wavelet with its reverberations and multiples, into a single spike. If we know the shape of the wavelet, we can design an operator which, when convolved with the seismic trace, with convert each wavelet into a single spike.

i. Key Parameters For Deconvolution:

- Type of Deconvolution-spiking or prediction.
- Portion of trace to be the source of the autocorrelation.
- Length of filter operator.
- White noise factor.

There are two types of deconvolution .

- Spiking Deconvolution .
- Predictive Deconvolution .

a. Spiking Decovolution:

It assumes the input is minimum phase and all frequency are to be leveled in the spectrum. The effect of this type of filter is to concentrate the energy of the pulse as near as possible to the front of the wavelet, i.e. to turn the wavelet into as near a spike as possible.

b. Predictive Deconvolution:

Predictive Deconvolution uses the autocorrelation of a trace to ascertain the periodicities within the data. The geophysicist determines from the autocorrelation(s) the necessary operator length, usually a few hundred millisecond that will span significant reverberation caused energy on the autocorrelation, and a gap, or time delay after the zero log value. The filter predicts reverberations and multiples. The predicted trace is subtracted from the observed trace to give the prediction error, which should be the trace with the predicted reverberations and multiples.

ii. Effect Of Deconvolution:

Because of deconvolution, the wavelet associated with the significant reflections is compressed and reverberatory energy that trails behind each reflection is largely attenuated by deconvolution *(Yilmaz,2001)*

Velocity Analysis:

Velocity analysis is performed on selected CMP or CDP gathers. The out put from one type of velocity analysis is a table of numbers as a function of velocity vs. two way zero off set time also called as velocity spectrum. Numbers present in the table represent some measure of signal coherency along the hyperbolic trajectories governed by velocity, off set, and travel time. The curve in each spectrum represents the velocity function based on picked maximum coherency values associated with the primary reflections. The pairs of numbers along each curve denote the time_ velocity values for each pick. These velocity time pairs are picked from these spectra based on maximum coherency peaks to form velocity functions at analysis locations. *(Yilmaz,2001)*

- ***Condition On Velocity Analysis:***

In areas with complex structures, velocity spectra (defined above) often fail to provide sufficient accuracy in velocity picks. In that case, the data on staked with a range of constant velocities (called as constant velocity analysis), and the constant velocity stacks themselves are used in picking velocities. *(Yilmaz,2001)*

Data Refinement

The processes described till now are used to make data free of the factors that decreases its quality. Also these processes are used to reformat the data and to diagnose its characteristics. Data refinement consists of the following two main stages.

- Stacking.
- Migration.

Along with these two processes, there is another procedure occasionally used in data refinement and is called as Residual Statics.

Stacking

Stacking is simply the process of adding up together the traces present in certain gathers, obtained during the seismic data acquisition .It is applied only when the all necessary corrections have been applied. The result of stacking is the corrected gather. In the "corrected gather" the traces have been gathered into the depth order. Both the static and dynamics corrections have been applied to it and the traces have been muted. All that remains is to stack the data. Stacking result in a single stacked trace as an out put for each depth point present in gathers.

One or other of two considerations is the basis for selecting the seismogram traces that will be stacked. Common offset stacking is done with traces that have the same source-receiver offsets, all of which are centered on the same point.

Stacking is a data compression of one to two orders of magnitude. The signal-to-random noise ratio is increased through an N fold stack by N. After stacking, the data are displayed at the surface location of the midpoint between source and receiver. When all adjustments to the data have transformed the offset data into time and phase coincidence with the zero offset traces, the common midpoint CMP and CDP are both widely often interchangeably. With dipping reflectors, the CMP after conventional processing is not the CDP. The correct positioning of reflection point will be by migration. *(Dobrin, 1988)*

Migration

The process of shifting the reflection points to the positions that correctly image the reflector and remove diffraction images, so that we may get an accurate picture of underground layers. If the reflector is flat, the reflection point will be located directly beneath the shot/receiver station, and the record section displays the event in its true position, plotted in time rather than depth. However, if the reflector is not flat, the reflection point will not lie directly beneath the shot/receiver position, and the true position of the reflector will differ from its apparent position. So migration is a tool used in seismic processing to get an accurate picture of the subsurface layer. It involves geometric repositioning of recorded signals to show a boundary or other structure, where it is being hit by the seismic wave rather than where it is picked up Now, not only the position but the dip angle can incorrectly imaged by vertically plotting.

i. ***Important Features Of Migration:***

Following are the important features of migration.

- § Migration steepens the reflectors as the dip angle of the reflector in the geologic section is greater than in the time section.
- § Migration shortens the reflectors as the length of the reflector on the geologic section is shorter than in the time section; thus,
- § Migration moves reflectors in the up dip direction
- § When migration is applied in case of the undulating reflector the crests become narrower and troughs become broad.

ii. ***Types Of Migration:***

With respect to the stage when migration is applied on the seismic data during processing, there are two important methods of migration.

- o Pre-Stack Migration.
- o Post-Stack Migration.

a) ***Pre-Stack Migration:***

Pre-stack migration is essentially when seismic data is adjusted before the stacking sequence occurs. The popular form of pre-stack migration is depth migration (PDM). PDM requires the user to know more about velocities of the layers. Once the user inputs these into the data with velocity analysis methods, there will be some error in the image. This error is caused by dipping reflectors or diffractions. The PDM will adjust the picture according to the velocities given.

When It Is Applied? Pre-stack migration is often applied only when the layers being observed have complicated velocity profiles, or when the structures are just too complex to see with post-stack migration.

Advantages And Disadvantages:

Pre-stack is an important tool in modeling salt diapirs because of their complexity and this has immediate benefits if the resolution can pick up any hydrocarbons trapped by the diapir. Overall, pre-stack migration, depth and time, is a valuable tool in better imaging seismic data, but it is limited by the amount of time and money required to conduct a pre-stack migration.Most of the pre-stack migration will be run when post-stacking has failed to resolve the layers or structures. However, with advances in computers, pre-stack migration will eventually become more economical.

b) ***Post-Stack Migration:***

Post stack migration is the process of migration in which the data is stacked after it has been migrated. This process is for many reasons, mainly because of its reasonable cost compared to pre-stack migration.

Basic Idea:

As in pre-stack migration, post stack migration is based on the idea that all data elements represent either primary reflections or diffractions. This is done by using an operation involving the rearrangement of seismic information so that reflections and diffractions are plotted at their true locations.

The reason that migration is needed is due to the fact that variable velocities and dipping horizons cause the data to record surface positions different from their sub-surface positions. The stacking is accomplished by making a composite record by combining traces from different records. Filtering is involved with stacking because of timing errors or wave-shape difference among the data being stacked.

Advantages And Disadvantages:

A disadvantage of using post stack migration compared to pre-stack migration is that it does not give as clear results as pre-stack

Post stack usually gives good results though, when the dip is small and where events with different dips do not interfere on the migrated section. Its cost is less than the pre-stack migration section.

Pre-stack migration is essentially when seismic data is adjusted before the stacking sequence occurs. The popular form of pre-stack migration is depth migration (PDM).

iii. Time And Depth Migration:

Now on the basis of the form of data on which migration is applied, there are two types of migration.

A process which collapses diffractions and moves dipping events toward the true position but leaves the migrated image with a *time axis* which must be depth converted at a later stage. Time migration assumes that the diffraction shape is hyperbolic and ignores ray bending at velocity boundaries.

The true Earth coordinates are of course in depth, not time. Even so, interpreters often need data in time coordinates, because the standard interpretation systems, log synthetics, and seismic-attribute techniques work with time and frequency, not depth and wavelength.

A time-migration can also be easily compared to the input stack section since they have the same vertical axis. Time migration is very fast and is robust to errors (sometimes up to 10%) in the velocity model. Further, errors in the shallow velocity model do not affect imaging of deeper structures.

Judging the "correctness" of a time migration is a rather arbitrary process but lateral velocity variations will cause positioning errors as indicated by *image rays*. An image ray, as shown in the adjacent figure for a single diffraction point, is normal to the recording surface and will show the lateral positioning error due to time migration. Unlike rays which indicate the direction of propagation, the image ray has no physical meaning. If an interpreter is at all worried about the positioning of data prior to drilling a well then a depth migration should always be performed. Surprises may result - particularly in 3D where structures can change dramatically in depth.

Time migration, following the tradition of NMO and stack, uses an *imaging* velocity field, i.e., one that best focuses the migrated image at each output location. This velocity field is free to change from point to point, so that time migration, in essence, performs a constant-velocity migration at each image point, where the constant changes from point to point.. *"The goal of time migration is to produce an image, not a geologically valid velocity field!"*

Depth migration, in contrast, uses an *interval* velocity field, i.e., a model of the Earth's subsurface. The interval velocities used are averages of the actual Earth velocities, where the average is taken over some characteristic distance such as a wavelength. This allows depth migration to model seismic wave behavior within the Earth much more accurately than time migration can. In particular, it allows us to use depth migration, especially depth migration before stack, as a velocity estimation tool.

Time migration will tolerate a certain amount of error in the velocity model. The commonest way to build a velocity model for time migration is to take the picked stacking velocities (assuming they are equal to *Vrms*) and smooth them sufficiently that the interval velocity model (from *DIX* conversion) is smooth and approximately resembles the geological structure. Almost always in modern processing DMO or pre-stack time migration will have been applied to attempt to remove the dip-dependence of stacking velocities. For time migration to be worthwhile the velocity model can vary smoothly in depth and very slowly laterally.

iv. Hazards In Migration Process:

١. Conventional migration is performed in the plane of the record section. The record section displays a two dimensional slice of the three dimensional earth. If the seismic line is not aligned perpendicular to strike, the reflections cannot be properly migrated in the record section plane.

٢. All migration methods are based on simplified models of the real earth. They involve assumptions about ray paths which cannot be verified. If the assumptions are poor the results will be poor. All of the routine migration methods in general use involve one very important assumption that over burden velocity layer are horizontal. If the real earth departs from this model, standard migration methods fail.

SEISMIC VELOCITIES

Introduction

Velocity as a seismic parameter plays an important rule in almost all the steps involved in seismic processing and interpretation. The accuracy of data reduction, processing and interpretation depends mainly on the correctness of velocity measurements. *(Al Sadi,1980)*

Seismic velocities are an important parameter in wave propagation. Its variation, along with density variation, decides the efficiency of reflection and transmission at any interface. A wave incident on a boundary is divided into reflected and refracted P-wave and S-wave. The amplitude of the seismic wave is changed by reflection and refraction. Reflection Coefficient is the ratio of the amplitude of the reflected wave to that of the incident wave. Seismic velocities are generally needed for the inspection of lithology and physical nature of rocks and also for calculation of dip and depth of interfaces. The seismic interpretation has been described as the process of solving for a velocity distribution from data measured in terms of time, which must be presented in geologic terms. The velocity varies both with lateral position and with depth. In many areas, seismic velocity data can be used to identify lithology in discrete formation within the geologic section. *(Dobrin & Savit 1988)*

Velocity as a seismic parameter plays an important role in almost the whole range of activities involved in seismic prospecting. The accuracy of data reduction, processing and interpretation of seismic data depends mainly on the correction of velocity measurements.*(Al-Sadi, 1980).*

Effect Of Physical Properties Of Rocks On Seismic Velocities

In terms of litho logy, whenever there is a change in grain size and mineralogical composition of rock, velocity behavior changes. An increase in grain size will result in the increase of velocity. The seismic velocities in rocks are affected by several factors : *(Telford et al, 1976).*

- Porosity
- Density
- Age of the rock
- Overburden pressure or Depth of burial
- Fluid content in the pores
- Litho logy or Mineralogical composition of rock
- Temperature
- Faust (1951) also found that in older rocks velocity is high, because for a long time they present under pressures, cementation and other factors that might increase its velocity.
- Faust (1953) suggested an empirical formula for velocity in terms of depth "Z" and the formation resistivity "R".

$$V=2*103(Z\ R)^{1/6}$$

Where,
V= Velocity in ft/sec
Z=Depth in ft
R=Resistivity in ohm-ft

Porosity

In the case of porosity higher will be the velocity and vice versa. The relationship between porosity (Φ) and velocity (V) is given by the following expression,

$$1/V = \Phi / V_f + (1+\Phi) / V_m$$

Where,
V_f = Velocity of the Pore Fluid
V_m = Velocity in Rock Matrix
V = Velocity in saturated rock
Φ = Fractional porosity

Density

The relationship of density and elasticity with the seismic velocity is given as:

$$(\text{Velocity})^2 = \text{Effective elasticity} / \text{Density}$$

From this relation, the velocity is directly proportional to elasticity and inversely proportional to density. So it is expected that the denser rocks would have low velocity, however the reserve is true in nature. The reason is that as the material becomes more compact its elasticity increases in such a way that it reduces the effect introduced by increased density.

Age Of Rock

An older rock might be expected to have a higher velocity, have been subjected for a long time to pressures, cementation, and other factors, which might increase its velocity.

Overburden Pressure Or Depth Of Burial

A quantitative relationship between velocity, depth and age of the rock for the shale and sandstone section, which is given as:

$$V=K\,(Z\,T)^{1/6}$$

Where,
V = Velocity in feet per second
Z = Depth in feet
T = Age in years
K = Constant

Fluid Content In The Pores

In actual rocks the pores spaces are filled with a fluid. This fluid may be in form of air or any other gas or different liquid solutions. These fluids can affect the seismic velocities. Seismic velocities will be low in case of gases and different solutions present in the pore spaces of rock.

Lithological And Mineralogical Composition Of Rock

Lithological and mineral composition of rocks also affects velocity of seismic waves. As described earlier, that average velocities for igneous rocks is higher than that for other types and they show a narrower range of variation than sedimentary and metamorphic rocks. Mineralogy of rock surely causes variations in seismic velocities. For example the velocity of a same wave will be different for sands, silts and clays etc.

Temperature

Seismic velocities decrease slightly with increase in temperature.

Different Velocities In Earth Materials

LITHOLOGY	Vp (km/sec)
Unconsolidated Materials Sand (Dry)	0.2-1.0
Sand (Water saturated)	1.5-2.5
Clay	1.0-2.5
Sedimentary Rocks Sandstones	2.0-6.0
Lime stones	2.0-6.0
Dolomites	2.5-6.5

(Kearey & Brooks, 1991)

Types Of Seismic Velocities

In seismic exploration and interpretation, there are certain types of seismic velocities. Each of these velocities conveys its own type and degree of information.

Average Velocity V_{avg}

This is simply the depth (Z) of a reflecting surface below a datum divided by the observed one way reflection time (t) from the datum to the surface so that

$$V_{avg} = Z / t$$

If Z represents the sum of the thickness of layers
Z1, Z2, Z3,………….., Zn. The average velocity is defined as:

$$V_{avg} = \frac{Z_1 + Z_2 + Z_3 + ...+ Z_n}{t_1 + t_2 + t_3 + ...+ t_n}$$

$$n = \frac{\sum Z_k}{\sum_1^n t_k}$$

The average velocity is used for time to depth conversions and for migration. The average velocity is used for conversion because it is the true verticals velocity in the ground. *(Robinson & Coruh, 1988).*

Interval Velocity V_{int}

Seismic wave velocity measured over a depth interval.

$$V_{\text{int}} = \Delta Z / \Delta t$$

Where,
$\Delta z = z_2 - z_1$
$\Delta t = t_2 - t_1$

Root Mean Square Velocity V_{rms}

If the seismic section consists of horizontal layers with respective interval velocities V1, V2, V3…………., Vn and one way interval times are t1, t2, t3,.tn, then Vrms for n layers model is obtained from the relation.

$$V^2_{rms} = \frac{V^2_1 t_1 + V^2_2 t_2 + V^2_3 t_3 + ...+ V^2_n t_n}{t_1 + t_2 + t_3 + ...+ t_n}$$

$$V^2_{rms} = \frac{\sum_1^n V^2_k t_k}{\sum_1^n t_k}$$

The Rms velocity gives a better result, when single layer case is used for Δt calculation. In fact, RMS velocity differs from the average velocity more and more as the layering becomes complex.*(Robinson & Coruh, 1988).*

Stacking Velocity V_{st}

Stacking velocity is used for NMO correction and also to evaluate depth to reflection. The relation of this velocity is

$$T^2 = \frac{T_0^2 + X^2}{V^2_{st}}$$

Where,

X =Offset distance

T_0 =Travel time

T =Travel time

V_{st} =Stacking Velocity

Stacking velocity is almost greater than average velocity. The relation between RMS and Stacking velocity in general is

$V_{avg} \leq V_{rms} \leq V_{st}$ *(Dobrin & Savit, 1988)*

Dix Velocity Formula

Dix (1955) showed that if there are two horizontal reflectors having time of T1 and T2 with respective RMS velocities V1 and V2, then the interval velocity Vint 12 between the reflectors is obtained from the relation.

$$V^2_{int}12 = \frac{V^2_2 T_2 - V^2_1 T_1}{T_2 - T_1}$$

This formula has been widely used to determine interval velocities between two uniformly dipping reflectors from RMS velocities.

Variation In Seismic Velocities

There are two types of variations in seismic velocities

i. Lateral Variations In Seismic Velocities:

These variations are supposed because of slow changes in density and elastic properties due to changes in litho logy or physical properties.

ii. Vertical Variations In Seismic Velocities:

These variations are due to litho logical changes of layering and increasing pressure due to increasing depth. Normally seismic velocities increase with the increase in depth. *(Robinson & Coruh, 1988)*

Methods For Velocity Determination

Velocity measurements are made at the beginning of a survey of an area. When the velocities have been determined, they are used throughout the area. There are two methods used for velocity determination.

i. Well Shooting

This is the most direct method of determination the velocity of seismic pulses in a well and measure the travel time nears the surface down to various points in the well. *(Dix, 1952).*

ii. Continuous Velocity Logs (Cvl)

It is a record of velocity variation through the borehole section, velocity measured directly in boreholes using a sonic probe, which emits high frequency pulses and measure the travel time of the pulses through a small vertical interval of well rock.

(Kearey & Brooks, 1991)

Importance Of Seismic Velocities

The Seismic Velocities may be used to establish the following:

- True depth
- Stacking of seismic data
- Migration of seismic data
- Possible lithology determination
- Possible porosity estimate