3 The Seismic Velocity

3.1 Introduction

The wave-propagation velocity plays a fundamental role in the theory and application of seismic waves. It enters in the wave-motion equation as well as in the wave changes occurring at interfaces. Seismic-energy partitioning at interfaces (reflection and transmission coefficients) is governed by variations of the acoustic impedance which is, in turn, dependent on the wave velocity. For a given seismic wave-type, velocity of any type of seismic waves is function of density and elastic constants of the medium in which the wave is moving. Further, we have for a given medium; each wave type has its own propagation velocity. Thus, for example, P-waves move faster than S-waves which, in turn, move faster than Rayleigh waves. In the common earth material where Poisson's ratio is about (1/4), P-wave is nearly (1.7) times as fast as S-wave and (1.9) times as fast as Rayleigh wave in the same medium. That is why we see in a typical earthquake seismogram, that P-waves arrive first, followed by S-waves and then followed by Rayleigh waves (Fig. 3.1).

3.2 Factors Influencing Seismic Velocity

Since velocity of a seismic wave is basically a function of the physical properties (density and elasticity) of the traversed medium, it is naturally affected by these properties. The main factors affecting the propagation velocity of seismic waves are: rock lithology, elastic coefficients, bulk density, and fluid contents. These factors are here-below briefly presented.

3.2.1 Lithology Effect

In nature, rocks differ widely in their chemical and physical properties. These properties (density, porosity, fluid saturation, crystallization, mineral content, rock texture) make up the lithological type of rocks. This means that the seismic velocity, which is a function of the overall physical properties, depends on the lithological nature of the medium. Igneous rocks, for example, are normally characterized by their high seismic velocity compared with the sedimentary rocks.

It is very common that a rock type is characterized by a wide range of velocities rather than by one single value. This is because of variation of properties found within the one defined rock-type. A sandstone rock, for example, may have a velocity value anywhere in the range of approximately (2–6) km/s. The corresponding range for limestone is (3–7) km/s. Usually, there are overlaps in the velocity ranges for the different lithologies. For this reason it is difficult to identify the type of lithology on basis of velocity criterion alone. The following Table 3.1 contains ranges of values of P-wave velocity, bulk density, and

H.N. Alsadi, *Seismic Hydrocarbon Exploration*, Advances in Oil and Gas Exploration & Production, DOI 10.1007/978-3-319-40436-3_3

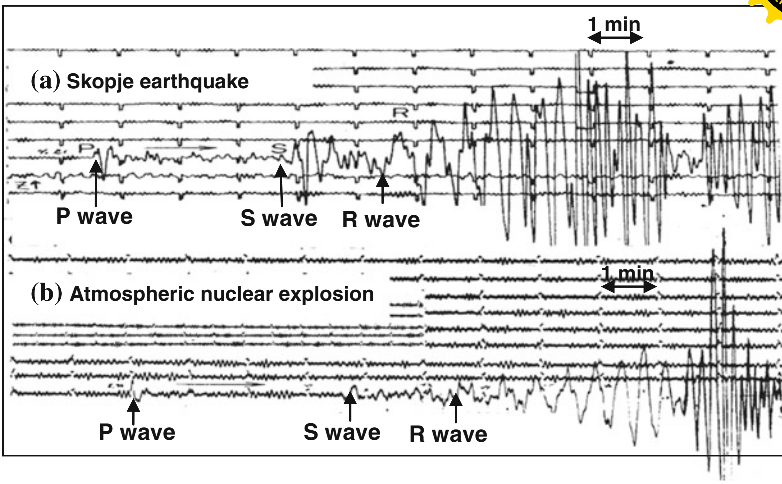

Fig. 3.1 Long period vertical-component seismograms recorded at Uppsala (Sweden). **a** Skopje earthquake, July 26, 1963, mag. 6.0. **b** An atmospheric nuclear explosion, mag. 5.4. Arrivals of *P*, *S*, and Rayleigh (*R*) waves are as shown (Båth 1973)

acoustic impedance for the most commonly known rock-types (Al-Sadi 1980, p. 70).

3.2.2 Elasticity and Density Effects

The wave motion equation of seismic waves includes the velocity factor present in its mathematical expressions. For a solid homogeneous medium, the propagation velocity for P- and S-waves ($\mathbf{v_p}$ and $\mathbf{v_s}$) are functions of elastic constants of the medium (Lame's elastic constants, $\boldsymbol{\lambda}$ and $\boldsymbol{\mu}$) as well as its bulk density ($\boldsymbol{\rho}$). These functions are:

$$\mathbf{v_p} = [(\boldsymbol{\lambda} + 2\boldsymbol{\mu})/\boldsymbol{\rho}]^{1/2}$$

and,

$$\mathbf{v_s} = [\boldsymbol{\mu}/\boldsymbol{\rho}]^{1/2}$$

Although the mathematical expressions of the velocity-density relations show that velocity is inversely proportional to the square root of density, it is a common observation that velocity appears to be increasing as density increases (Nafe and Drake 1963; Gardner et al. 1974). The explanation for this discrepancy, is that as the material becomes more compact (that is as density increases) its elastic coefficients increase in such a way that it offsets the effect introduced by the density increase. An empirical relationship between P-wave velocity (v) and bulk density ($\boldsymbol{\rho}$) for the common sedimentary rocks (sandstone, shale, limestone, anhydrite …) is given as follows (Gardner et al. 1974):

$$\boldsymbol{\rho}(\mathbf{v}) = \mathbf{k}\mathbf{v}^{1/4}$$

where (**k**) is equal to 0.31 when (**v**) is in m/s and equal to 0.23 when (**v**) is in ft/s.

Density values for common rock types are found in Table 3.1 presented above.

3.2.3 Porosity and Saturation Fluid Effects

Porosity (defined as the pore volume per unit volume) has an effective role in seismic velocities because of its direct relation to the bulk density. In a porous rock, velocity is affected by porosity as well as on the type of interstitial fluid. In general there is an inverse relationship between porosity (**Ø**) and velocity (**v**) of rocks. Velocity of a fluid-saturated rock is given by the following empirical formula which is normally called the time-average equation (Wyllie et al. 1958):

the effect of ($\mathbf{P_p}$) in compressing the porous rocks is in opposite direction to that created by ($\mathbf{P_o}$), the compaction net effect (compaction pressure, $\mathbf{P_c}$) is proportional to the pressure difference ($\mathbf{P_o} - \mathbf{P_p}$). That is:

$$\mathbf{P_c} = \mathbf{P_o} - \mathbf{P_p}$$

As rock compaction process brings about a corresponding change in the physical properties, it is expected that seismic velocity (or interval transit time) would vary with the applied compaction pressure. Velocity increases with increasing compaction pressure and decreases with increase of interstitial pore pressure. This velocity-pressure relationship forms the basis used in predicting subsurface pressure conditions from seismic velocity data. Use of seismic velocity as an indicator of pressure conditions was used by several workers like Pennebaker (1968), Reynolds (1970), Louden et al. (1971), and Aud (1974). Departure of the measured velocity of a given geological section from the normal-compaction trend serves as indicator of abnormal-pressure zone.

3.2.6 Other Velocity-Affecting Factors

In addition to the above mentioned factors, there are other less important factors which can cause small effects on seismic velocity. Examples of such factors are pore shape, anisotropy, temperature, and wave-frequency. It is found that velocity decreases slightly with increase of temperature. It is found that, for an increase of 100 °C, velocity decreases by about 5 % and that the velocity drop-rate is considerably more when the rocks are saturated with heavy crude oil or tar. Velocity in water saturated rocks experiences sudden increase as temperature is lowered to the freezing point (Sheriff and Geldart 1995, pp. 120–121).

Velocity is practically unchanged over a broad frequency range. Small change is observed in case of body-wave dispersion (velocity variation with frequency). Low-frequency components are "faster" because of frequency-dependant absorption phenomenon. In case of anisotropic medium, velocity varies with changes in propagation direction. The main factors affecting velocity can be summarized as follows:

(i) Lithology
In general, igneous and metamorphic rocks have larger velocity values than sedimentary rocks. Sub-types of rock lithologies are characterized by wide ranges of velocities which are overlapping with each other. For example, the velocity-range for sandstone (2–5) km/s is overlapping with that of shale (1.5–4.5) km/s, and so on.

(ii) Elasticity and Density
Theoretically, velocity is directly proportional to elastic constants and inversely with density. However, because of the greater effect of elasticity, it is observed that velocity has an apparent direct proportionality with density changes.

(iii) Porosity and Saturation Fluid
Velocity is inversely proportional to porosity, and the change in velocity depends on the type of pore fluid. Velocity is lowered when the pores are filled with water or oil, and more lowered when the pores are filled with gas.

(iv) Depth and Geological Age
In general, velocity increases as both of the overburden depth and geological age increase. A power-law relationship is connecting the velocity with the depth and geological age. The power index for sand-shale sequence is found to be around the value of (1/6–1/4).

(v) Overburden Pressure
Due to compaction caused by the overburden weight, elastic constants as well as density get increased. There is a direct relation between velocity and the net compaction pressure which is equal to the difference between pore pressure and overburden pressure (geostatic pressure as it is sometimes called). Velocity increases with increasing compaction pressure and decreases with increase of interstitial pore pressure.

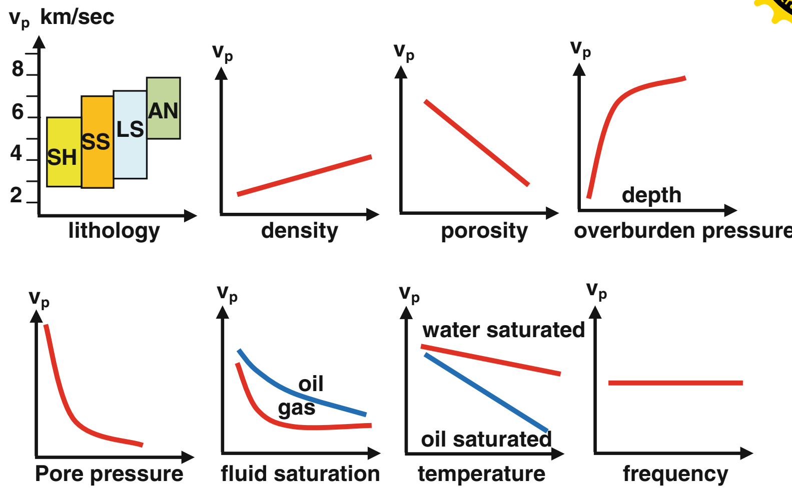

Fig. 3.2 Main factors affecting seismic P-wave velocity (v_p). The symbols *SH*, *SS*, *LS*, and *AN* denote shale, sandstone, limestone, and anhydrite respectively

(vi) Other Minor Factors

Pore shape: Velocity decreases when pores are elongate.

Anisotropy: Velocity varies with changes in propagation direction.

Temperature: Velocity slightly decreases with increase of temperature.

Frequency: Velocity practically unchanged with frequency. With normal dispersion, low-frequency components are "faster" because of frequency-dependant absorption phenomenon.

Main factors affecting velocity are shown in Fig. 3.2.

3.3 The Velocity Function

Seismic propagation velocity, as we have seen from the previous discussion, depends on a number of variables, like type of lithology, geological age, density, porosity, pressure, and depth. In seismic reflection exploration work, velocity is usually presented as mathematical functions. The velocity function is taken to describe its variation with reflection travel-distance (or, more often, with travel-time).

Since the real earth is not homogeneous (typically layered medium), seismic waves move with different velocities in different parts of its travel-path. In this case, it becomes very useful to express the propagation velocity (**V**) as a mathematical function of the travel-distance, **V**(**x**) or function of travel-time, **V**(**t**). For practical applications, velocity is, more often, expressed as function of vertical reflection time, that is, as function of the two-way (reflection) vertical time **V**(**T**). In a homogeneous, uniformly changing, or layered medium, the distance-versus-time functions can be linear, curved, or segmented respectively. The corresponding V(**T**) functions can be plotted as curved or straight lines. Typical velocity functions are shown in Fig. 3.3.

Due to the geological complexity of the earth, it is not, in practice, possible to express seismic velocity-variation in the form of an explicit mathematical function. However, under certain conditions, a geological medium is approximated by a defined set of specifications that can allow derivation of a mathematical function describing the variation of velocity along the travel-path. One of these models is one in which the instantaneous velocity (**V**) increases linearly with depth (**h**). This model is expressed by the linear function; $\mathbf{V(h) = V_0 + kh}$, where ($\mathbf{V_0}$) is the velocity value at the surface ($\mathbf{h = 0}$) and (**k**) is the velocity-depth gradient. The significance of this relation is that it gives a close approximation to the actual velocity variation observed in many sedimentary-basin areas (Dobrin 1960, p. 77). Derivation of the ray-path geometry (which turns out to be of circular form) and travel-time functions, are given by several authors (see for

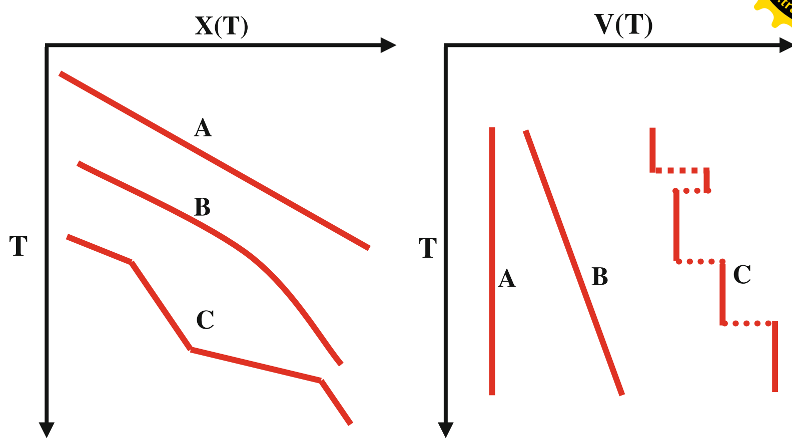

Fig. 3.3 Sketches of velocity functions for three cases: **a** homogeneous medium, *V*(*T*) is line of zero slope, **b** uniformly changing velocity, *V*(*T*) is line of constant slope, **c** layered medium, curve is made up of zero-slope linear segments

example, Nettleton 1940, pp. 257–261 or Sheriff and Geldart 1995, pp. 93–94).

3.4 Types of Velocity Functions

Unlike electromagnetic waves which are all of one type (e.g. wave of visible light which is of S-wave type), seismic waves are of many types, each of which has its own velocity-value which is dependent upon the density and elastic constants, as mentioned above. Because of the inhomogeneous nature of the crustal part of the Earth (which is typically made up of layered rock media), several types of velocity functions are needed to express velocity variation as function of travelled distance. Five velocity-types are in common use in seismic exploration. These are:

(i) Instantaneous Velocity (V)
(ii) Interval Velocity (IV)
(iii) Average Velocity (AV)
(iv) Root Mean Square (RMS) Velocity (RV)
(v) Stacking Velocity (SV)

3.4.1 Instantaneous Velocity

In cases where the subsurface geology is of variable nature, the seismic velocity which is governed by the physical properties of the medium will accordingly be changing with changes of the location in the travel-path of the moving wave front. This means that at any point of the travel-path, the velocity can be measured or computed at that point. This is the (instantaneous velocity) which can be defined to be the velocity of wave motion at a given point in a medium traversed by an advancing seismic wave.

Mathematically, the instantaneous velocity (**v**) at a point is given by the slope of the tangent to the distance-versus-time curve at that point (Fig. 3.4).

The instantaneous velocity, (**v**) of a wave moving along a distance (**x**) is defined as:

$$\mathbf{v} = \mathbf{dx}/\mathbf{dt}$$

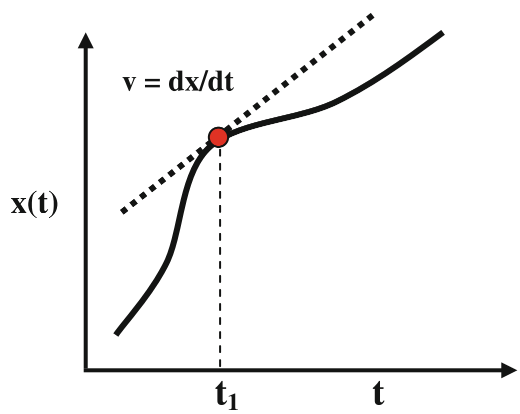

Fig. 3.4 Definition of the instantaneous velocity (v), given by the slope of the tangent to the x–t curve, $v = dx/dt$ at $t = t_1$

The closest example to the instantaneous velocity is the velocity function in which the interval transit time (which is equal to the reciprocal of velocity) is measured across each depth-interval of one meter down a drill hole. The resulting record (the sonic log) gives detailed velocity-information at points spaced by 1 m interval. This is a practical approximation of the instantaneous velocity-variation with depth.

3.4.2 Interval Velocity

The interval velocity (**IV**) is defined to be the average velocity over an interval of the travel-path. It is usually measured or computed for individual geological layers. Thus, the interval velocity, of a geological formation of thicknesses (**Δh**) and interval transit time (**Δt**), is given by **IV = Δh/Δt**. For a multi-layer geological section, the interval-velocity function is constructed by computing the interval velocity for each layer of that section (Fig. 3.5).

3.4.3 Average Velocity

As the name implies, the average velocity (**AV**) is obtained from dividing the total distance travelled by the wave by the time spent in covering that distance. In seismic exploration work, the average velocity is computed for a set of layers, usually starting from the datum plane down to the required reflector level.

For a vertical travel-path, the average velocity of a geological section made up of (**n**) layers will be given by dividing the total thickness $\mathbf{h_n}$ ($= \mathbf{\Delta h_1 + \Delta h_2 + \Delta h_3 + \cdots + \Delta h_n}$) by the travel-time $\mathbf{t_n}$ ($= \mathbf{\Delta t_1 + \Delta t_2 + \Delta t_3 + \cdots + \Delta t_n}$), that is:

$$\mathbf{AV_n = h_n / t_n}$$

or,

$$\mathbf{AV_n} = \frac{(\mathbf{\Delta h_1 + \Delta h_2 + \Delta h_3 + \cdots + \Delta h_n})}{(\mathbf{\Delta t_1 + \Delta t_2 + \Delta t_3 + \cdots + \Delta t_n})}$$

An example of computation is shown in Fig. 3.6.

The average velocity (**AV**) can be calculated for a multi-layer section, given the interval velocities (**IV**) for each of the section layers. This can be done by the following formula:

$$\boxed{\mathbf{AV_n = (\sum IV_n \,.\, \Delta t_n) / \sum \Delta t_n}}$$

It is also possible to derive the interval velocity of a certain layer in a pack of layers, given the average velocity data. Thus, in order to compute the interval velocity ($\mathbf{IV_n}$) of the **nth** layer, we need to know the two average velocities, ($\mathbf{AV_n}$) and ($\mathbf{VA_{n-1}}$) for the section from the datum level down to the base and down to the top

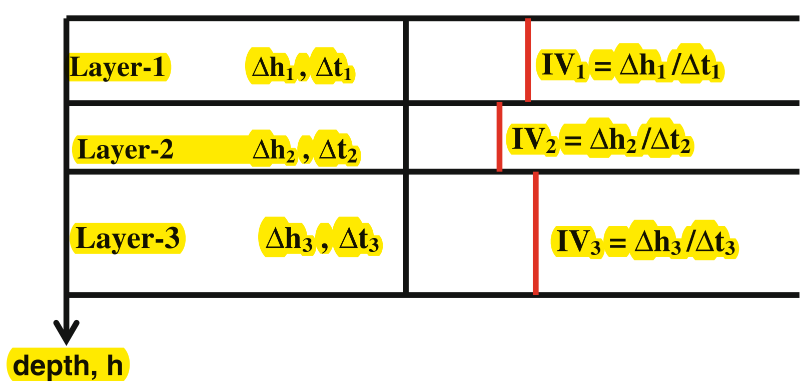

Fig. 3.5 Interval velocity function of a three-layer geological section. IV_1 at depth Δh_1, IV_2 at depth $\Delta h_1 + \Delta h_2$, and IV_3 at depth $\Delta h_1 + \Delta h_2 + \Delta h_3$

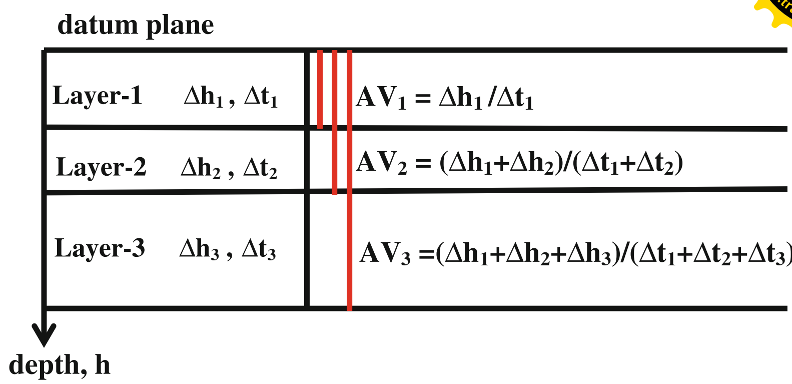

Fig. 3.6 Average velocity (AV_1, AV_2, AV_3) function of one, two, and three-layers of thicknesses (Δh_1, Δh_2, Δh_3). The corresponding interval travel times are (Δt_1, Δt_2, Δt_3)

of that layer respectively. Using the corresponding travel times, ($\mathbf{T_n}$) and ($\mathbf{T_{n-1}}$), the required formula is:

$$\mathbf{IV_n = (AV_n \cdot T_n - AV_{n-1} \cdot T_{n-1}) / (T_n - T_{n-1})}$$

3.4.4 Root Mean Square Velocity

Root mean square (RMS) velocity (**RV**) is defined as the square root of the average of the squares of the weighted interval-velocities, where the weighting factors are the layer thicknesses or the interval transit times. In seismic exploration, the RMS velocity, like other types of velocities, is usually computed for vertical ray paths of waves traversing a multi-layer medium (medium made up of horizontal layers of different interval velocities). A three-layer model is shown in the following Fig. 3.7.

The root mean square velocity (**RV**) can be calculated from the interval velocity data (**IV**), down to the **nth** layer, by use of the following formula:

$$\mathbf{RV_n = [(\Sigma IV^2_n \cdot \Delta t_n) / \Sigma \Delta t_n]^{1/2}}$$

From the RMS velocity (**RV**), it is possible to derive the interval velocity (**IV**), given the two RMS velocities for the section, from the datum level down to the base ($\mathbf{RV_n}$) and down to the top ($\mathbf{RV_{n-1}}$) of that layer. Using the

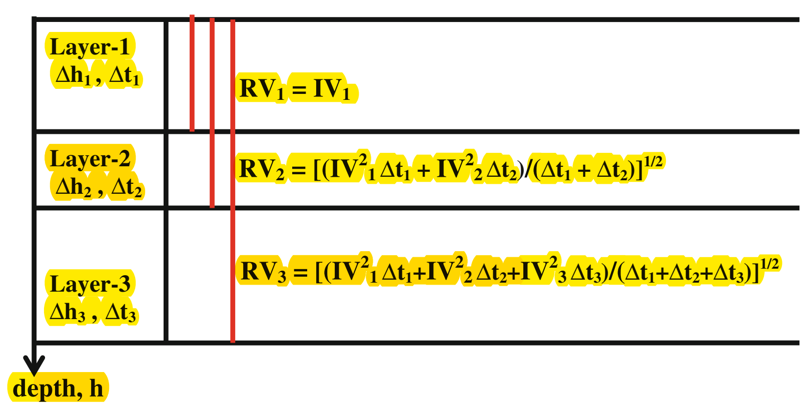

Fig. 3.7 Root mean square (RMS) velocity (RV_1, RV_2, RV_3) function of one, two, and three-layers of thicknesses (Δh_1, Δh_2, Δh_3). The corresponding interval travel times are (Δt_1, Δt_2, Δt_3)

corresponding travel times, ($\mathbf{T_n}$) and ($\mathbf{T_{n-1}}$), the required formula is (Dix 1955):

$$\boxed{IV_n^2 = [(RV_n^2 \cdot T_n - RV_{n-1}^2 \cdot T_{n-1})/(T_n - T_{n-1})}$$

For a given geological section, the RMS velocity is typically a few percents larger than the corresponding average velocity (Sheriff 1973, p. 228).

3.4.5 Stacking Velocity

Stacking velocity is the main velocity variable that enters in the NMO-correction formula (explained in Sect. 4.3). It is applied to remove the time-contribution of the receiver-to-source distance (called, offset) from the total reflection travel-time. This is clarified in the following (Fig. 3.8).

Stacking velocity is determined by velocity analysis technique whereby the time contribution of offset is removed before stacking of the CMP gather-traces. This is one of the fundamental processing steps (called NMO correction) which bring about coherency of the reflection arrivals which, on stacking, give enhanced reflection event. It is called stacking velocity because of its role in enhancing the stacked reflection signal. Since its direct role is in the NMO correction, the term (NMO velocity) will be more appropriate than the commonly applied term (stacking velocity). The CMP concept is explained in Sect. 4.4.

Role of the stacking velocity in enhancing the reflection signal is schematically shown in Fig. 3.9.

In dipping parallel reflectors (dip angle, ($\boldsymbol{\theta}$)), with parallel velocity layering, the stacking velocity (calle it $\mathbf{V_S}$) is equal to RMS velocity ($\mathbf{V_{RMS}}$) divided by cosine of the dip angle ($\boldsymbol{\theta}$) that is (Sheriff and Geldart 1995, p. 134):

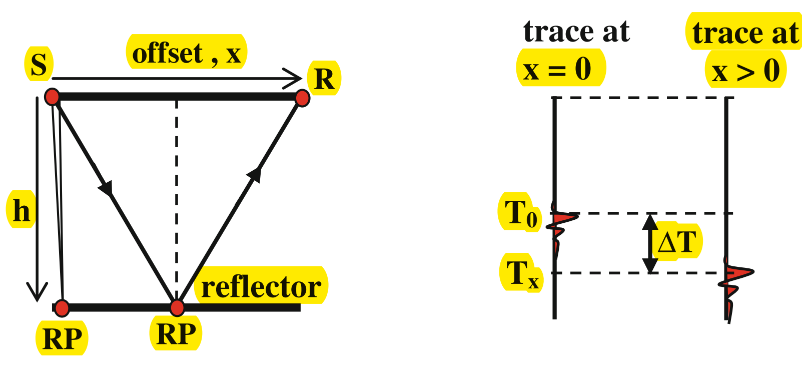

Fig. 3.8 Definition of the stacking velocity. It enters in the *NMO* correction formula ($\Delta T = T_x - T_0$), where T_0 and T_x are the reflection travel-times for zero-offset and x-offset respectively

$$T_0 = 2h/V$$

$$T_x = (4h^2 + x^2)^{1/2} / V$$

$$\Delta T = T_x - T_0 = [(x/V)^2 + (T_0)^2]^{1/2} - T_0$$

V is the stacking velocity

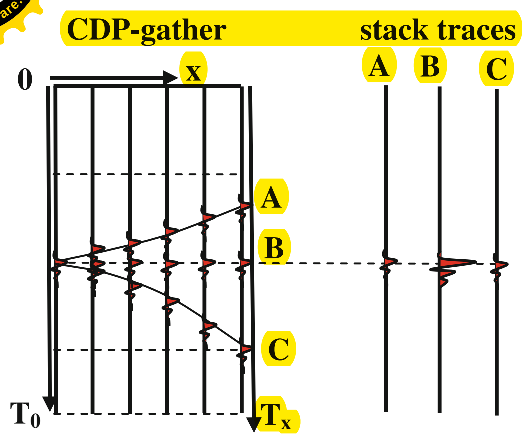

Fig. 3.9 Role of the stacking velocity in the *NMO*-correction of a *CDP* trace gather. In this model, three stacking velocities were applied (cases: *A*, *B*, and *C*). The optimum velocity (case, *B*) gave the highest staked signal

$$\mathbf{V_S = V_{RMS}/\cos\theta}$$

This relation between the stacking velocity and RMS velocity can be applied by interpreters in their interpretation work of seismic reflection data and in velocity-changes studies.

Stacking velocity is sometimes called RMS velocity, because stacking velocity is the nearest in value to RMS velocity in a multi-layer medium. Like RMS velocity, stacking velocity is slightly greater than the average velocity.

3.4.6 The Apparent Velocity

The Apparent Velocity (**V_a**) is defined as the propagation velocity measured in the direction other than the true propagation direction. In seismic exploration, the apparent velocity is often dealt with in connection with the movement of plane wave-front advancing along a path inclined with respect to the ground surface. A plane wave approaching a ground surface with true velocity (**V**), along a ray-path making an angle (**θ**) with the normal to the surface will have an apparent velocity (**V_a**) of its motion along the surface (Fig. 3.10).

During a time interval (**Δt**), the wave front moves a distance of (**VΔt**) with the true velocity (**V**), while, at the same time, the moved distance at the surface (**Δx**) is covered by the apparent velocity (**V_a**), that is:

$$\mathbf{V_a = \Delta x/\Delta t}$$

hence,

$$\mathbf{V_a = V/\sin\theta}$$

This relation shows that the apparent velocity is always greater than the true velocity by a factor depending on the angle of approach (**θ**, in this example). The apparent velocity approaches infinity when (**θ** = 0), that is when the wave path direction is perpendicular to the surface plane. In any case the apparent velocity is always greater than the true velocity with which the wave is approaching the horizontal surface plane.

Another point of interest which is related to this subject is the apparent wavelength. Suppose that the measurements were taken for time of one period (**Δt = τ**), distance measured over the surface corresponding to time of one period will be

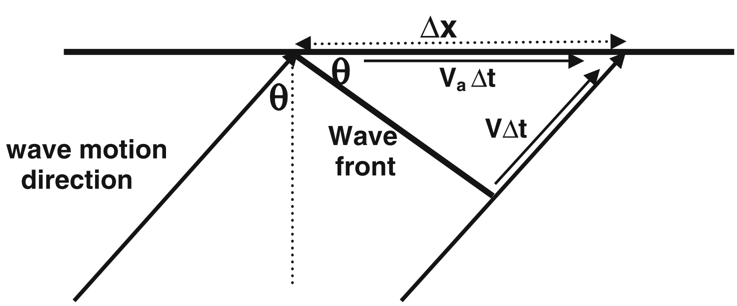

Fig. 3.10 Ray path geometry of a plane wave approaching the surface plane at an approach angle (*θ*). Velocity in the ray-path direction (*V*), and velocity of wave front measured on the horizontal plane surface is the apparent velocity (V_a)

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the apparent wavelength ($\Delta x = \lambda_a$), hence (using the relationship; $\lambda = V\tau$):

$$\lambda_a = \lambda / \sin\theta = 2\pi / k_a$$

where (k_a) is the corresponding apparent wave number.

3.4.7 The Group and Phase Velocities

The group velocity and phase velocity are distinguished in cases of dispersive waves. This phenomenon has practically no applications in the field of seismic-exploration, since the body waves, which are used in seismic exploration show no significant dispersion. Wave dispersion is common with surface waves under certain conditions. Wave dispersion occurs as result of variation of velocity with frequency, and in this case, two velocities are distinguished; the group velocity and phase velocity.

The group velocity is defined to be the velocity with which the seismic energy-packet (represented by the wave-train envelope) travels. The phase velocity, on the other hand, is the velocity of a certain frequency component of the moving wave. As usual, the two velocities differ in value and consequently, a wave-peak, or a wave-trough appears to move within the wave-train. Detailed account on this subject is found in the geophysical literature as in Richter (1958, pp. 243–244), Telford et al. (1990, pp. 153–154), and Sheriff and Geldart (1995, pp. 60–62).

3.4.8 Representations of the Velocity Functions

The velocity types that play important roles in seismic exploration are: interval velocity, average velocity, RMS velocity, and stacking velocity. In reflection seismology these velocities are usually plotted as functions of reflection travel time, as shown in Fig. 3.11.

As far as velocity types are concerned, there are other types of velocities of less importance to the seismic exploration applications. Most important of these are the apparent velocity, group velocity, phase velocity.

3.5 Velocity Determination Methods

Seismic velocity plays an important role in all activities involved in seismic exploration surveying. Its importance stems from the fact that the end product of any seismic survey is a time-image (and not depth image) of the subsurface geology. Thus, to convert the time-image data into depth domain, velocity must be made available. In seismic reflection surveying, velocity computations, aim at finding the seismic velocity (average- or interval-velocity) expressed as function of depth. There are several ways to compute the seismic velocity-function. Velocity computations are made either by methods based on borehole data or by methods based on analysis of seismic data. The two groups of velocity determination methods are:

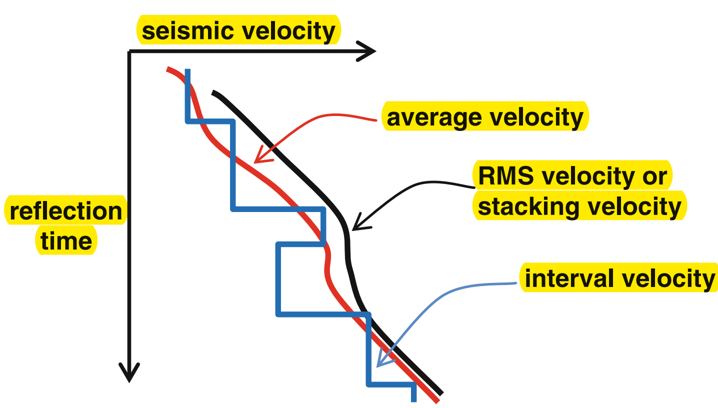

Fig. 3.11 Representation of the main types of velocity functions used in seismic reflection exploration

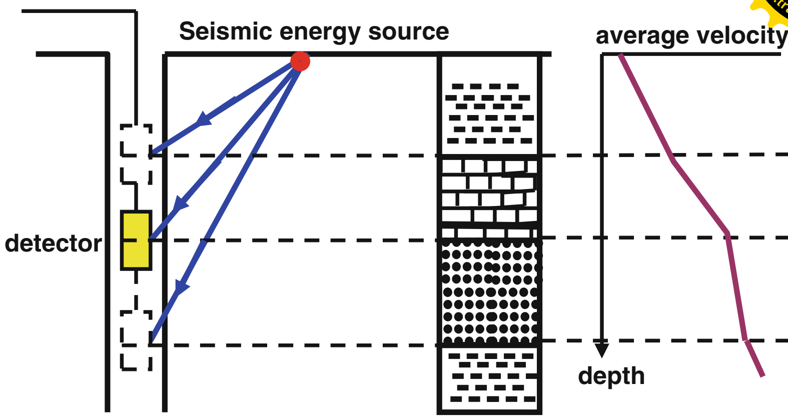

Fig. 3.12 Method of well velocity surveying. The detector is placed at a geological-formation boundary. The curve is representing the computed average velocity as function of depth

Borehole-Based Methods

(i) Well Velocity Surveying
(ii) Up-hole Velocity Survey
(iii) Continuous Velocity Survey

Seismic Data Analysis Methods

(iv) ($X^2 - T^2$ Method)
(v) ($T - \Delta T$) Method
(vi) Velocity Analysis
(vii) Seismic Inversion

3.5.1 The Well Velocity Survey

The normal well velocity surveying proceeds by generating seismic waves from a seismic energy source located on the surface near the well head. The directly arriving wave is recorded by a detector placed at a certain depth inside the well (usually at a boundary of a geological formation). The shooting and recording process is repeated at all of the geological-boundaries penetrated by the drill-hole. For more detailed surveying, recordings are made at additional intermediate detector-positions of smaller spacing. From the source-to-detector travel time, corrected to the vertical path, the average velocities are calculated and plotted against depth (Fig. 3.12).

The source is either dynamite charge fired in a shallow drill-hole or air-gun operated in a mud pit. The detector is normally a specially-designed geophone provided with a lever that makes the geophone to be well-pressed against the borehole wall. It may be a hydrophone-type detector suspended inside the well which is filled with the drilling fluid.

The average velocity ($\mathbf{V_{av}}$) is computed from the slant travel time (**T**) of the direct wave recorded by the detector placed at depth (**h**) using the formula (Fig. 3.13):

$$\mathbf{V_{av} = h/T_s \cos\theta}$$

The interval velocity of a certain geological formation is derived by dividing the formation thickness by the interval transit time of that formation. Another way of computing of the interval velocity is calculated by the mathematical relationship connecting the interval velocity with the average velocity.

The computed velocity function is very important for the interpretation process. It is used for calibrating the sonic log and check the integrated time of the sonic log, hence the name (check-shot surveying) which is sometimes used as another name for the well shooting method.

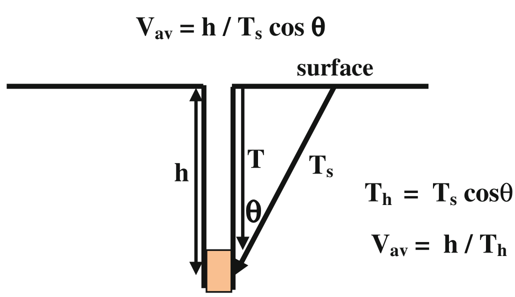

Fig. 3.13 Computation principle of the average velocity

... the analysis is explained in more details in Sect. 10.7.1.

3.5.7 Seismic Velocity Inversion

The amplitude of a reflection signal (**A**) is dependent on the reflection coefficient (**R**) which is, in turn, depending on the acoustic impedance (**Z**) which is equal to density (**ρ**) multiplied by velocity (**V**). This relationship (**Z = ρV**) shows that velocity can be determined from amplitude measurements, given the seismic impedance and density data. The basic principle, upon which the seismic inversion is based, will be presented in a simplified way, as follows:

For a plane seismic wave perpendicularly incident at an interface separating two layers of acoustic impedances ($\mathbf{Z_1}$ and $\mathbf{Z_2}$), the reflection coefficient (**R**) is given by:

$$\begin{aligned}\mathbf{R} &= (\mathbf{Z_2}-\mathbf{Z_1})/(\mathbf{Z_2}+\mathbf{Z_1}) \\ &= (\boldsymbol{\rho}_2\mathbf{V}_2-\boldsymbol{\rho}_1\mathbf{V}_1)/(\boldsymbol{\rho}_2\mathbf{V}_2+\boldsymbol{\rho}_1\mathbf{V}_1)\end{aligned}$$

Assuming the amplitude of the incident wave is unity (1), and the reflected amplitude is (**A**), the reflection coefficient (**R**), by definition, becomes equal to (**A**) and hence, we can use the following equivalent form:

$$\begin{aligned}\mathbf{A} &= (\mathbf{Z_2}-\mathbf{Z_1})/(\mathbf{Z_2}+\mathbf{Z_1}) \\ &= (\boldsymbol{\rho}_2\mathbf{V}_2-\boldsymbol{\rho}_1\mathbf{V}_1)/(\boldsymbol{\rho}_2\mathbf{V}_2+\boldsymbol{\rho}_1\mathbf{V}_1)\end{aligned}$$

giving:

$$\mathbf{Z_2} = [(\mathbf{1}+\mathbf{A})/(\mathbf{1}-\mathbf{A})]\cdot\mathbf{Z_1}$$

Since density variation is very small, compared with velocity variation, that is putting ($\boldsymbol{\rho}_1 = \boldsymbol{\rho}_2$), the direct (**A-V**) relationship can be obtained which is:

$$\mathbf{V_2} = [(\mathbf{1}+\mathbf{A})/(\mathbf{1}-\mathbf{A})]\cdot\mathbf{V_1}$$

This is an inverse problem in which the acoustic impedance (expressed by velocity) can be obtained from amplitude data. If the velocity ($\mathbf{V_1}$) of the surface layer of a layered medium is known then using the inversion formula, $\mathbf{V_2} = [(\mathbf{1} + \mathbf{A})/(\mathbf{1} - \mathbf{A})]\cdot\mathbf{V_1}$, the velocity ($\mathbf{V_3}$) of the neighboring deeper layer is computed. By repeating the computation, velocities of the rest of layers are sequentially determined.

Normally this approach is applied in transforming a seismic stack section into acoustic impedance section, or into what is called pseudo-impedance when density is ignored. The computations are normally carried out by software especially designed for this purpose.

3.6 Uses of the Seismic Velocity Data

All types of the seismic velocity have important role in the seismic exploration activities (data acquisition, processing, and interpretation). Velocity enters in the travel-time functions of all body and surface waves (direct, reflected, refracted, and diffracted waves). To start with, velocity governs reflection and transmission coefficients. In processing of seismic data, velocity forms an important factor in travel-time corrections like

Table 3.2 Fields of application and precision assessment of various types of velocity (after Al-Chalabi 1979)

Velocity	Main uses	Precision requirements
Stacking velocity	Stacking of seismic sections	Modest to low
	Migration processing	Modest to low
	Estimation of RMS velocity	Dependant on situation
RMS velocity	Migration-velocity estimation	Generally modest
	Interval-velocity estimation	Dependant on situation
	Average-velocity estimation	Dependant on situation
Interval velocity	Lithologic and stratigraphic studies	High to modest
	Interpretation works	Modest to low
	Abnormal pressure detection	High to modest
	Ray tracing computations	Dependant on situation
	Migration processing	Generally modest
	Average-velocity estimation	Dependant on situation
Average velocity	Depth conversion	Generally modest
	Interpretation works	Modest to low

Precision requirements: high = 0.1–1.0 %, modest = 1–5 %, low >5 %

static and dynamic (NMO) corrections. Amplitude compensations (as in geometrical spreading and inelastic absorption) and seismic migration depend on velocity. In interpretation activities, velocity has a fundamental role in time-to-depth conversion and in mapping structural and stratigraphic features.

Summary of fields of application and precision assessment of the various velocity-types are given in the following Table 3.2.