Chapter 2
Seismic Reflection Principles: Basics

Abstract Seismic reflection events are caused by the impedance contrasts at layer interfaces having a minimum width (Fresnel Zone). Seismic studies to represent reliably the subsurface geology, require quality data, which depend on signal-to-noise ratio and resolution, the latter being the ability to image thin geologic features separately. This calls for a seismic broad-bandwidth source consisting of both low and high frequencies that can improve resolution limits to layer thicknesses.

Seismic reflections record attributes such as amplitude, phase, polarity, arrival time and velocity that can be measured or estimated. The attributes define the shape and arrival time of reflection waveforms which depend on rock properties. Estimation of rock properties from seismic waveforms and their vertical and lateral changes in time and space is the essence of seismic interpretation.

Appropriate choice of seismic display modes and plotting scales are also important.

When a seismic wave, generated artificially on surface propagates through the earth and meets interfaces between different kinds of rocks, creates phenomena like reflections, refractions, scattering and diffraction. Of these, the reflection is by far the most significant phenomenon, as it forms the basis of the potent seismic reflection method deployed to image the subsurface for finding hydrocarbons. A compressional wave (P-wave) when incident normal to an interface, causes reflection and transmission also normal to it, but when incident at an angle (inclined), it produces two sets of waves, P-reflected and P-transmitted (refracted) and S-reflected and S-transmitted (Fig. 2.1). We shall limit the discussion to the principles of relatively simpler P-wave reflections, used extensively for measuring rock and fluid properties.

A seismic reflection event to be generated needs necessarily two things, an impedance contrast at the interface of two rock types and a minimum width (Fresnel Zone) of the interface. The reflection amplitude and its continuity depend on the degree of contrast across the interface and its extent and nature. The effectiveness of the reflections to reliably represent the subsurface geology is conditional on the quality of seismic reflection signal, which depends on, (1) the amount of noise recorded in the data and (2) the ability of the seismic wavelet to image the different interfaces separately. The reflection signal quality is thus adjudged by the two

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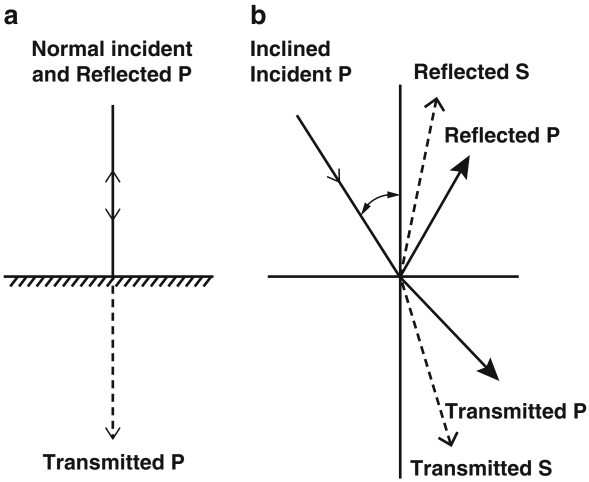

Fig. 2.1 (**a**) A normal incident P-wave on an interface produces one set of waves normal to the interface, the P-reflected and the P-transmitted. (**b**) An inclined incident P-wave, however, produces two sets of waves P-reflected and transmitted and S-reflected and transmitted (refracted)

important factors, the signal-to-noise ratio and the resolving power of the seismic wavelet, briefly discussed below.

Signal-to-Noise Ratio (S/N)

Noise may be defined as all undesired energy, other than the primary reflections from the subsurface strata. It is an inherent part of the seismic recording and processing system present due to ambient (within earth), geological (natural propagation) or geophysical (artifacts during recording and processing) causes. This noise cannot be wished away, but can be effectively reduced by conscious efforts during data acquisition and processing. Noise, though usually unwanted, can be occasionally helpful in interpretation. For example, remnant diffraction noises despite processing may indicate clues to presence of sharp edges, such as faults and other subtle stratigraphic objects. The presence of scattering noise may give an idea about the order of heterogeneity of the reflector, leading to indication of highly tectonised zones, crushed with faults and fractures. Processing of recorded noise as scatters from fractures and fractured zones can also be used as a technique for delineating naturally fractured carbonate and basement fractured reservoirs.

Since noise severely affects seismic clarity in portraying the subsurface image, it is desirable to record good and clean signals with minimum noise. It is a common practice to benchmark the quality of data in terms of a measure of a ratio between signals and noise (S/N). Improved data acquisition techniques including meticulous survey layout plans, field experimentations and strict on-field execution ensure good quality of data. In this context, the common depth point (CDP) seismic data acquisition is a unique technique. A standard technique practiced all over the world, it achieves signal enhancement at the cost of noise, via a summation process of several traces reflected from the same subsurface depth point but with different offsets known as CDP folds. Though summation of higher number of traces in a fixed offset range generally provides better S/N ratio, there may be a limit beyond which it may not be desirable, as adding additional traces (folds) may cost more money without improvement in the seismic images. Also, summation is an integration process, which affects resolution, especially in cases where large far offset traces are included for summing. In areas where the geology promotes good quality seismic reflections, the interpreter may still prefer to look at less-fold CDP data which is likely to offer better resolution and at a lower cost. It may also be noted that data with high S/N ratio does not necessarily assure higher resolution, as the resolution depends on other factors such as source signal frequency, sampling interval and subsurface wave propagation effects, besides noise.

Seismic Resolution

Resolution may be defined as the ability to separate two closely spaced features in depth (time) as well as in space. The resolution we refer to in seismic geophysics is of two types, vertical and horizontal. Vertical (temporal) resolution is the minimum separation in time between two reflections arriving at the surface that allows detection of each reflector separately. Lateral (or spatial) resolution is the minimum lateral distance (spatial) between two close geologic objects that permits each one to be imaged individually. It may be stressed that detection of an event is not the same as resolution which delineates the objects clearly.

Resolution depends on the seismic wavelength with which the subsurface is measured. Wavelength is a fundamental property of a wave which is the distance between successive points of its equal phase (e.g., crest to crest), completing one cycle. It is usually denoted by the symbol λ (see Fig. 2.8) and is defined by the equation $\lambda = v/n$, where *'v'* and *'n'* stand for the velocity and frequency of the wave passing through a medium. Smaller wavelengths provide better resolution whereas wavelengths too large compared to the dimensions of the object, fail to detect it. Since wavelength is a direct function of velocity and inversely that of frequency, seismic resolution happens to be better at shallow depths where seismic wavelength is smaller due to relatively lower velocity and dominant higher frequencies. On the other hand, resolution deteriorates with depth due to longer wavelengths because of increasing velocity and lowering of frequency.

Vertical Resolution

A short sharp zero phase wavelet (high bandwidth) ideally provides the best resolution, as the arrival times of individual reflections from closely spaced reflectors, being of short wavelet durations, do not overlap during recording at the surface. A zero phase wavelet is symmetrical and has maximum amplitude at time zero, chosen as the origin. A zero phase wavelet is an interpreter's desired wavelet which mathematically speaking is a non-causal wavelet. The commonly used seismic sources like dynamite on land and air-guns in marine produce minimum phase or mixed phase wavelets. However, with a Vibroseis source, a zero-phase wavelet, known as *Klauder* wavelet, is realized by a mathematical treatment (autocorrelation) of the known Vibroseis sweep that makes it a preferred choice.

The seismic short source wavelet, further, while traveling within the earth suffers loss of high frequencies due to absorption and gets changed to a long and cyclic ('leggy') wavelet, practically producing a mixed phase wavelet. The large length of the wavelet does not permit enough separation between the arrival times of reflections coming from closely spaced beds and results in overlapping of the individual events, thus losing the ability to resolve the beds separately (Fig. 2.2). It has been demonstrated by synthetic modeling that $\lambda/8$ is generally the limit of bed thickness, below which thinner beds cannot be seen as resolved (Widess 1973). Widess envisaged a wedge model with impedance contrasts remaining the same at its top and bottom but with the signs of the impedance contrasts reversed (Fig. 2.3). However, in many geological situations the impedance contrasts at top and bottom are likely to be different in values and dissimilar in polarities, in which case, the thin bed resolution limit given by Widess model may be different. Nonetheless, practical experi-

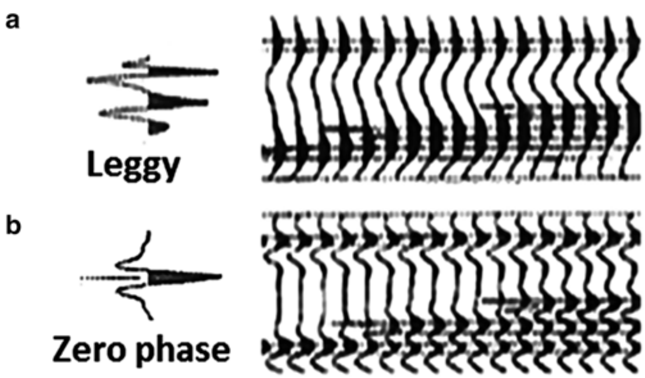

Fig. 2.2 The seismic vertical resolution depends on the type of wavelet embedded in the data. (**a**) A cyclic (leggy) mixed phase wavelet having energy loaded in its later part impedes resolution by causing overlapping of reflections from thin beds. (**b**) The zero phase, short and symmetrical wavelet, has maximum amplitude at zero time with small side lobes and promotes better resolution (Image after Sheriff (1973))

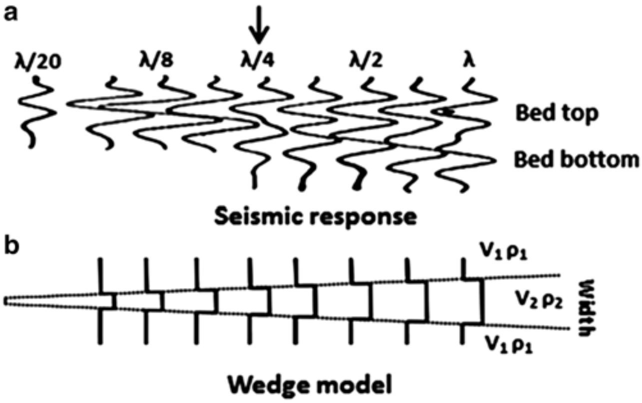

Fig. 2.3 (**a**) The Widess thin-bed wedge model. (**b**) For a bed with thickness greater or equal to the wavelength (λ), the top and bottom reflections are clearly resolvable and this continues till it approaches the quarter wavelength (λ/4). For beds thinner than λ/4, the top and bottom reflections are not seen as distinct, limiting the vertical resolution to quarter wavelength (After Widess 1973)

ence shows that in real-earth situations, where some amount of noise is always present in the data, *λ/4* may be considered a reasonable wavelength as the resolving limit of beds. Broadly, vertical (temporal) resolution varies from 10 to 15 m at shallow depths and from 20 to 30 m at greater depths.

Exploration objectives (reservoirs) are often thin and require improved vertical resolution for proper mapping. Resolution can be enhanced during acquisition by deploying a broad-band wavelet as a source (dynamite) and by recording with smaller sample intervals (temporal, ~2 ms). In addition to the data acquisition efforts, care is taken to retrieve and boost the higher frequencies during processing of the data. The recorded seismic trace is a convolution, a mathematical manner of combining two signals (likened to ~product), of the source wavelet with impedance contrasts present in the subsurface. If the source wavelet can be removed (deconvolved) from the recorded trace through data processing, the impedance contrasts representing geologic rock discontinuities will be left behind. This is the sole aim for seismic investigation, and can be achieved through various data processing techniques known as deconvolution. Deconvolution and zero-phase wavelet are processing steps to increase vertical resolution by suppressing multiples and by compression of the wavelet that is achieved by increasing effective bandwidth.

Lateral (Spatial) Resolution and Fresnel Zone

Huygen's principle stipulates that reflection from a surface consists of a number of diffractions occurring from each point on it and does not come from a single point. Where the reflecting surface is uniform and planar, the diffractions from all points

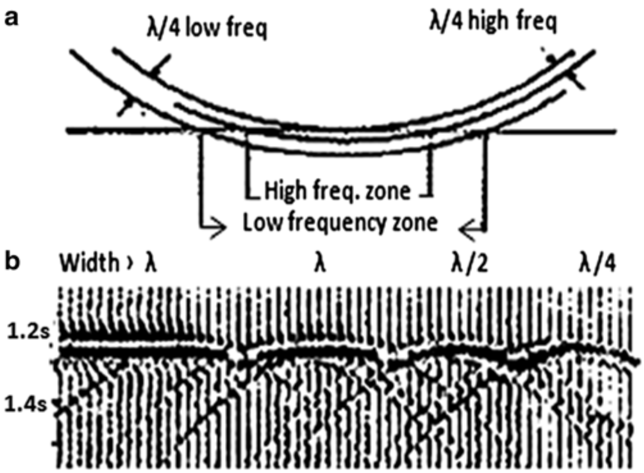

Fig. 2.4 A schematic to illustrate the concept of Fresnel zone. (**a**) Fresnel zone is the effective contact area of a spherical wave with an interface that results in a reflection event. The width of the Fresnel zone is dependent on frequency. (**b**) Synthetic reflection events with variable spatial resolution. Notice the beginning of deterioration in the event at λ/4, which sets it as the limit (After Mickel and Nath 1977)

add constructively to provide a reflection event. If, however, the surface is curved or has less continuity, the diffractions may not add effectively and provide poor reflection.

Seismic waves that originate from a point source are spherical in nature, and when incident on a plane reflector, they sweep through it by producing a succession of contact zones. Nonetheless, the limited planar area, which 'effectively' comes into contact at the interface and collectively contributes to produce a coherent reflection, is called the first *Fresnel zone* (Fig. 2.4a). The seismic wave is a band-limited signal comprising a range of frequencies, and when incident on an interface, each frequency creates its individual area of contact with the interface to cause a reflection. Thus it is important to visualize the phenomenon of reflection as an 'area' concept in two dimensions and as 'volume' concept in three dimensions instead of a single 'point' notion that can have enormous significance in data interpretation and evaluation.

The quality of a reflection depends not only on the area defined by Fresnel zone but also on the type of the reflecting surface. The lateral changes in reflectivity of planar widths less than a Fresnel zone tend to deteriorate the reflection quality. Modeling has demonstrated that interfaces having width less than $\lambda/4$ cannot be viewed clearly and thus defines the limit for spatial resolution (Fig. 2.4b). The Fresnel zone may be considered as a lateral requisite, complimentary to the vertical impedance contrast, responsible for causing a reflection event. Similar to thickness of the bed that determines the temporal resolution, the Fresnel zone width can be considered to serve as a yard stick for defining lateral resolution.

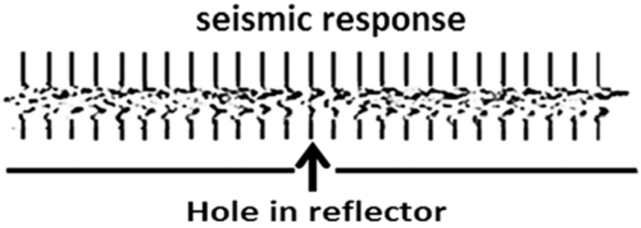

Fig. 2.5 A small discontinuity in the Fresnel zone, such as a small hole in the reflector, has little effect on seismic reflection due to 'wave front healing', a process of diffraction of wave going around the aberration (After Sheriff 1977)

Poor to no reflections, at times, seen associated with fault edges, sharp facies changes, small reefal mounds and erosional unconformities may be examples of inadequate imaging linked to Fresnel's zone width. However, a small discontinuity in the reflecting surface, for instance, a hole cut in the Fresnel zone, will hardly affect the quality of the averaged reflection due to phenomenon of wave front healing (Fig. 2.5), a process by which the waves are diffracted around the discontinuity. This can have important geological implications in that the open fractures and cracks present in rocks may be difficult to be imaged directly by P-wave seismic. Further, Fresnel zones in the subsurface are often not planar but consist of curved surfaces, which is yet another factor that affects quality of reflections. For convex upward surfaces (anticlines), the contact area of the wave with the reflector is small that amounts to loss of amplitudes, where as for concave surfaces (synclines), the contact area being more, provides strong amplitudes. This phenomenon is similar to focusing and defocusing effects of an optical lens.

In reality, it is important to comprehend the Fresnel zone concept through its geometry, the shapes and widths vary greatly depending on several factors, which make it an intricate three-dimensional problem. In a two dimensional case, it may be expressed in terms of a product of seismic wavelength and depth as $R \approx (\lambda \times z/2)^{1/2}$, where R, λ and z represent the Fresnel zone radius, seismic wavelength and depth respectively. The Fresnel zone is small at shallower depths (wave length and distance being small) and increases with depth to the order of hundreds of meters. Since the Fresnel zone width sets the spatial resolution limit, it is important that this be reduced to a minimum to improve spatial resolution and resolve two small geologic objects with clear separation from each other. This is achieved to a large extent by the process of migration that enhances horizontal resolution similar to the role that deconvolution plays in enhancing the vertical resolution.

Migration is a processing technique, which (1) repositions the dipping seismic reflection events to their true geological positions in the subsurface, and (2) collapses the diffractions to improve images and their continuity. Migration eventually preserves true reflection amplitude, creates a more accurate image of the subsurface and more importantly, enhances spatial resolution. For this reason, migration of data is desirable even for data with flat geologic strata. Typically, the Fresnel zone widths, of hundreds of meters in unmigrated data, can be considerably reduced to about 10 m or so by migration. For an effective migration, however, knowledge of proper overburden velocity field and an adequate number of surrounding traces *(aperture)* at the object level is necessary for stacked data. An aperture is the spatial width over which all traces around are considered for migration, and choosing an appropriate aperture is crucial to its effectiveness. Generally, an aperture of twice the Fresnel zone width at the reflection object is adequate (Sun and Bancroft 2001). The migration results suffer gravely near the end of seismic lines as there are no traces recorded and the interpreter should be cautious to consider data in this part.

For better resolution of lateral reflectivity changes of small dimensions, migration may require finer spatial sampling on ground like the temporal sampling used for improving vertical resolution. Take for instance the issue of imaging a small channel of 20 m width, an important geologic object for exploration. Obviously, the channel cannot be resolved with insufficient trace sampling of 25 m though the image with this trace spacing may detect it. The river geometry and more importantly its associated reservoir facies like channel, levee and point bar sands need to be imaged and resolved properly to characterize the reservoir and may necessitate closer trace spacing (subsurface) of no more than 10 m.

Temporal and spatial resolution may be considered somewhat similar in nature and are decided by the wavelength, which is dependent on velocity and frequency of the seismic wave. Both the resolutions depend on velocity but it must be stressed that temporal resolution depends on interval velocity while the spatial resolution is dependent on overburden velocity. As an example, take a limestone bed with an interval velocity of 3200 m/s and an overburden velocity of 2400 m/s for calculating the resolution limits. The vertical and lateral resolution limits for a dominant frequency of 40 Hz, are 20 and 15 m, considering quarter wavelength as the realistic limits of resolution. It is useful for the interpreter to have some idea about resolution limits beforehand; otherwise, he or she may be looking for things that are beyond the capability of the data to offer. It is also interesting to note that the two resolution effects are inter-reliant and improving one tends to better the other (Lindsey 1989).

Interference of Closely Spaced Reflections: Types of Reflectors

We have seen earlier that for beds with thickness, larger than quarter seismic wavelength, reflections from their top and bottom appear as distinct and separate. However, most commonly beds are closely spaced in the subsurface and reflections involving several beds arrive within a time spacing that is less than the length of the seismic pulse. This leads to superposition of the reflections (Fig. 2.6). The ensuing interference can be either constructive or destructive and the resultant composite reflections depend on: (a) number and thickness of the beds (b) magnitude and sign (polarity) of the reflection coefficients and (c) the order of positioning of the individual impedance contrasts. We may consider the behavior of three types of

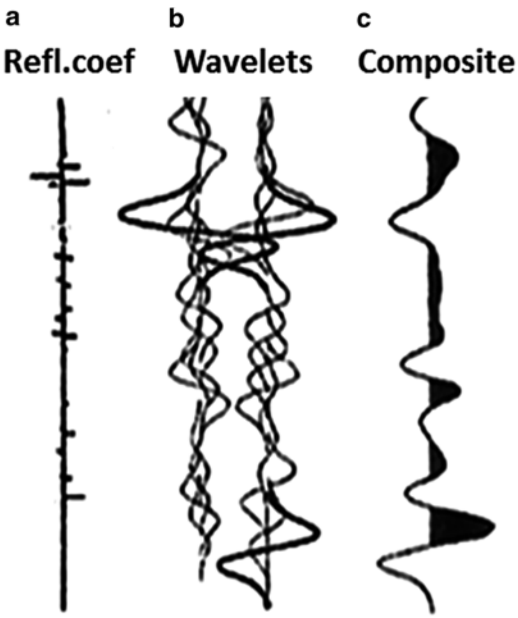

Fig. 2.6 The interference of reflections from closely spaced interfaces. (**a**) Subsurface reflection coefficient series, (**b**) wavelet reflections from the individual beds, and (**c**) the composite reflection caused by superposition of individual reflections from thin beds (Modified after Vail et al. 1977)

reflectors, namely *discrete, transitional* and *complex,* that an interpreter routinely comes across during interpretation (Fig. 2.7).

Discrete Reflectors

Top and bottom of thick beds with sharp impedance contrasts create distinct separate reflections to be recorded and are termed discrete reflectors. The reflections from top and bottom appear well separated with amplitude proportional to reflection coefficients. The onset of the reflection from the interface, either a peak or trough, appears at the right time on record with respect to its subsurface position without any delay.

Transitional Reflectors

A reflector may be termed transitional if there is a gradual gradation of impedance contrasts of one sign, either positive or negative, as in a fining upward channel or coarsening upward bar sand (Anstey 1977). The interference of a succession of reflections of the same signage (polarity) results in a composite reflection of an

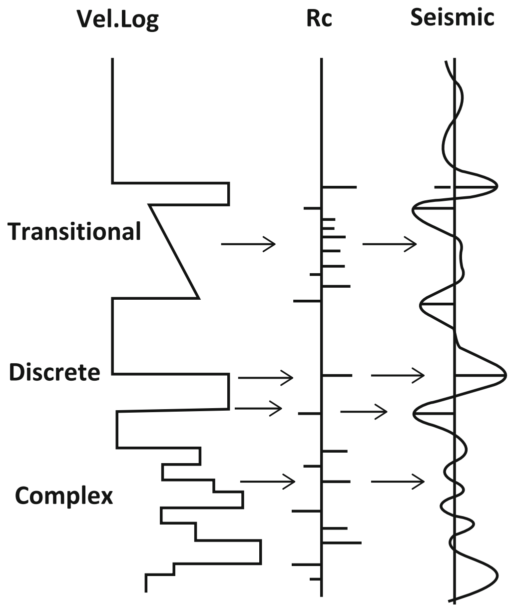

Fig. 2.7 A schematic to show the different types of reflectors. The discrete reflector has thickness that causes top and bottom reflections to be resolvable and with distinctive polarity and exact arrival time. 'Transitional' and 'complex' reflectors are composite events of several closely spaced beds with one or mixed signage respectively. They create reflections with uncertain polarity and delayed arrival time (Modified after Clement 1977)

integrated wave shape. The reflection is generally weak with a low frequency appearance and the event onset is time delayed with respect to the top of the formation.

Complex Reflectors

A complex reflector is a pack of reflectors, spaced closely but with varying magnitudes and polarities of impedance contrasts, which produce a complex reflection. The strength, phase and onset of the reflection are difficult to gauge. Forward seismic modeling may be used as a solution to get an insight to the pattern of a complex reflection.

Innate Attributes of a Reflection Signal

A seismic trace is a log measure of disturbances (particle velocity/ acoustic pressure) of waves reflected from subsurface with time. It records in a waveform the intrinsic attributes of a reflection signal amplitude, phase, frequency, polarity, arrival

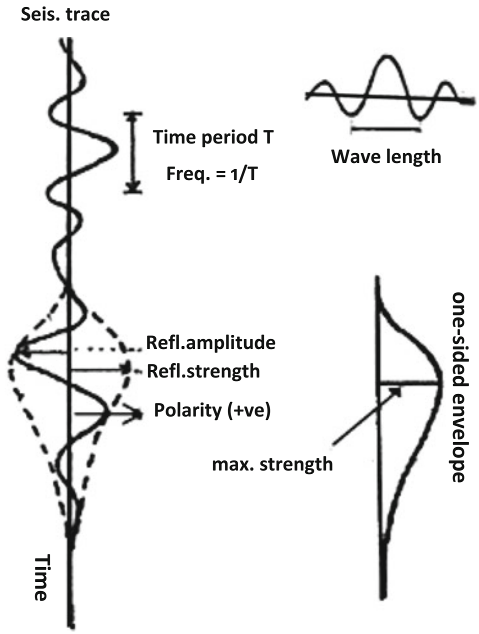

Fig. 2.8 The seismic attributes measurable on a trace, namely the time period, wavelength, reflection amplitude, reflection strength and polarity. Reflection strength is the maximum amplitude of the envelope of a composite reflection, independent of phase (After Anstey 1977)

time and velocity, all of which can be measured or estimated. The reflection attributes (Fig. 2.8) define the shape and arrival time of reflections depending on properties of the rocks. Thus the waveforms carry important geologic information encrypted in them. Estimates of these rock properties from seismic waveforms and their vertical and lateral changes in time and space is the essence of seismic interpretation, which predicts the subsurface structures and stratigraphy required for petroleum exploration. The basic seismic attributes are introduced here; their measurement and application is described in Chap. 10.

Amplitude and Strength

As stated earlier, a seismic wave incident normal to an interface with an impedance contrast produces two waves normal to interface, one reflected back and the other transmitted onward. The amplitude of the reflected wave with respect to that of the incident wave is termed the reflection coefficient (*Rc*) or the reflectivity. Reflectivity depends on the degree of contrast between the impedances on either side and the angle of incidence of the wave. For a normally incident wave, reflectivity *(Rc)* is expressed by the founding equation of seismic reflection method, $Rc = V_2 \rho_2 - V_1 \rho_1 / V_2 \rho_2 + V_1 \rho_1$, where V_1, ρ_1 and V_2, ρ_2 are the velocities and densities of the upper and lower layers respectively. For non-normal (oblique) incidence, there will be, however, two pairs of *'P'*and *'S'* waves (see Fig. 2.1) and the above equation for the normal reflection coefficient gets relatively complicated and assumes a more general form as in Zoeppritz's equations, discussed in Chap. 10.

Amplitudes are measures of particle velocities or pressures and in an ideal case, the maximum value at peak/trough of a pulse wavelet represents the reflection coefficient of an isolated reflector. Where the wavelet is leggy (lengthy and cyclic) and the reflection is of composite nature, as is often the case in nature, it is difficult to choose the appropriate peak/trough for calculation of amplitude to represent reflectivity. In such cases, it is convenient to make use of the reflection strength, which is specified by the maximum amplitude of one side of a symmetrical envelope, centered about the reflection event (Fig. 2.8). Reflection strength is more meaningful as it is independent of phase and relatively less sensitive to the factors affecting amplitude. Reflection strength may have a maximum at a phase other than at peak/trough and may indicate the nature of the composite reflection. Reflection amplitude and its variations are vital clues to predict lithology of formations and their lateral changes, porosities and sometimes pore fluids as in the case of gas reservoirs. However, a crucial limitation of amplitude is its proneness to wide variance due to several other factors that may not be linked to geology.

Phase

Phase may be expressed simply as the time delay with respect to the instant of start of a reflection. Phase is independent of amplitude, indicates the continuity of an event and provides another useful criterion to interpret reflections. In areas of poor reflectivity, where reflection amplitudes are too weak to be manifested and correlated, phase is likely to be helpful in mapping the continuity of the reflection (reflector). Phase mapping is especially sensitive to detection of discontinuities like pinch outs, faults, fractures and angularities as well as unconformities based on 'out of phase' events.

Frequency (Bandwidth)

A seismic wavelet, usually of one to one-and-a-half cycles in the beginning, changes shape progressively during propagation and becomes long and cyclic (leggy) with passage of time. The pulse width of a wavelet on the seismic record in time (time period) provides an estimate of its dominant lowest frequency, and it grows larger with depth during propagation, indicating lowering of frequencies caused due to attenuation. The bandwidth is a measure of the width of a range of frequencies in the wavelet, measured in hertz and is the key to quality of reflection. Bandwidth decides the time duration of the changing wavelet corresponding to depth intervals, reliant on the velocity and controls vertical and lateral seismic resolution. A broad bandwidth consisting of both low and high frequencies is thus considered essential to provide quality seismic images. The lower frequencies in the spectrum help in deeper penetration of energy where as the higher frequencies direct the thin bed resolution. Unfortunately, during propagation of the wave, the earth attenuates the high frequencies and hampers desired resolution at depths.

Because frequency is affected by propagation phenomena like absorption and transmission in the subsurface, its variance can provide at times valuable information on geologic strata and its geometry. Layered beds at shallower depths generally evince reflections with high frequency contents whereas older and harder rocks (for example, Pre-Tertiary rocks) at deeper depths show relatively low- frequency reflections. The experienced seismic interpreter is familiar with the clearly discernible decrease in bandwidth of reflections from the top to bottom of a typical seismic section. Bandwidth, amplitude and phase create the shape and form of a signal, and the individual components can only be measured and analyzed by detailed spectral analysis, discussed later in Chap. 10.

Polarity

The polarity expresses the sign of a reflection coefficient. It is considered positive if the impedance of the rock below is positive (a hard rock underlying a soft rock) and negative, the other way round. Conventionally, on normal processed data (SEG normal polarity convention), peaks and troughs of reflection events represent positive (black) and negative reflection coefficients (white), though an option for plotting reflections with reverse polarity remains with the interpreter. It is important to define the polarity convention clearly in the processed data so as to avoid making basic mistakes in interpreting the geology. Picking of reflection polarity is simple and straight forward in case of discrete reflectors, but is difficult in transitional and complex reflectors where superposition of reflectors leads to distortion of events and provides a composite reflection. Noise in data also acts as a deterrent for clear determination of polarity. Deconvolution and zero phase processing help to some extent in estimating polarity of composite reflection events. Accurate picking of polarity, wherever possible, helps in locating the disposition and nature of the strata in the subsurface.

Arrival Time

Reflections arriving at different times on a record for discrete reflectors indicate the temporal position of rock boundaries encountered in the subsurface. Since the events are recorded in time, accurate velocity function is required to convert the arrival times and determine the depths of the causative reflectors. Despite deploying true velocity, there can still be inexactness in some cases in matching the actual subsurface depth with depth converted from time. For transitional and complex reflectors, the precise time of onset of a reflection is usually recorded as delayed and can pose a problem, *prima facie*. The system of recording and processing of data also behaves like filters and introduces time lags. If the induced delay is not properly taken care of, it may add to the overall delay, which may vary from a few to several milliseconds, depending on type of seismic data (2D/3D). Seismic analysts often find such time shifts in tying a particular reflection phase in different vintages of seismic, especially in 2D data, due to varying recording and processing parameters used and one must be careful before picking and mapping geologic horizons.

Velocity

Velocity is an important seismic attribute, not only to estimate depths of formations, but also to provide vital information on subsurface rock and fluid properties. Basically, velocities are of two kinds, the *overburden* or vertical average velocity, and the *interval* or formation velocity. Vertical velocity is used for conversion of reflection times to depth and the interval velocity for estimating lithology and other rock properties like porosity and fluid contents. The two velocity functions are interrelated; knowledge of one can lead to calculation of the other. The seismic CDP technique permits the calculation of an apparent overburden velocity from multi-trace data processing and is known as the normal move out (NMO) or stack velocity. Stacking velocity is so named, as it is computed mathematically from the normal move out equation which maximizes the effect of summation of traces in a CDP gather. It is a velocity along the direction of the geophones and is affected by factors such as dips of strata and recording spread lengths. Stacking velocities are usually higher (by about 6–10 %) than true vertical average velocity, which can be measured only in a well. Stacking velocities are also referred to as RMS (root mean square) velocities. Where well velocity is not available, the RMS velocity after appropriate correction is used to predict top, bottom and thickness of geologic formations. The lithology and other rock properties can be also inferred from interval velocities (formation velocity) calculated from stack (RMS) velocities.

The velocity used for migration of seismic data, a process that moves the subsurface reflecting points to their true spatial position below the shot point is known as the migration velocity. It is an overburden velocity and applied appropriately, produces relatively clean and accurate seismic images that help predict rock properties better. Generally, it is lower than stack velocity but tends to equal true overburden velocity where migration of data is perfect to provide reliable depth conversions.

Seismic Display

The visualization of seismic data is an integral part of interpretation and as such, it is important that the processed seismic data be displayed in suitable graphic modes and scales. Nonetheless, it depends to a large extent on the objectivity of the interpretation and the perception and creativity of an individual interpreter. Generally, data are displayed in any one of these modes, wiggle trace, variable area, and variable density or in a combination (Fig. 2.9).

- Wiggle trace is a log of reflection amplitudes with time and makes it handy to interpret geologic information from the variability in the waveform shape.
- Variable area (VA) and wiggle displays are wiggles shaded with bias, and make reflection events appear more consistent and convenient for correlation by reflection character.
- Variable density (VD) shows reflection strength, and displayed with color, provides better relative standout and continuity of reflections. VD sections, though more commonly used, do not show the waveform shapes that embed significant geologic information (Fig. 2.10).
- Combinations of variable area and wiggles may be a preferred display for interpreting stratigraphic details.

Though the work stations provide different modes of display, interpreters generally use Variable density sections due to better apparent continuity of reflections and their amplitude stand-outs that can be conveniently used as an attribute display. But

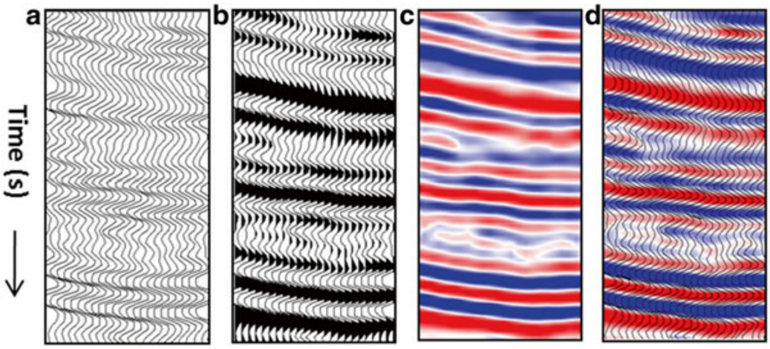

Fig. 2.9 The types of seismic data display modes. (**a**) wiggle, (**b**) wiggle and variable area, (**c**) variable density, (**d**) combination of wiggle and variable density. Note the waveform changes seen clearly in wiggle and variable area display mode (**b**), that carry the crucial geologic information

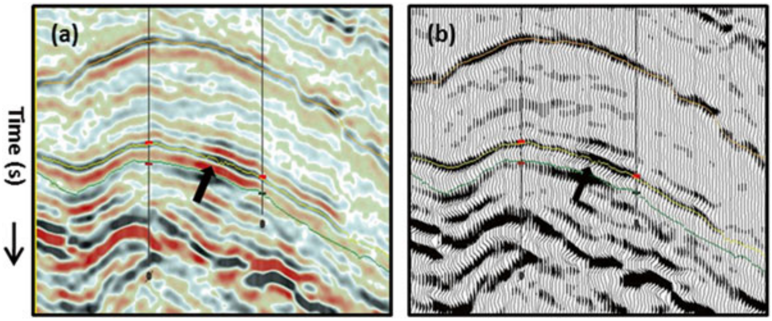

Fig. 2.10 Comparison of seismic (**a**) variable density, and (**b**) wiggle display modes. Reflection stand-outs and continuity seen better in the variable density display, but does not show the changes in waveform that carry important geologic information, whereas wiggle mode clearly shows the variations in waveform shape (trough indicated with an *arrow*) (Image: Courtesy, Hardy Energy, India)

this can be misleading in sedimentary environments of continental to fluvio-deltaic deposits where fast and frequent facies variations are likely to occur causing discontinuous and patchy reflections. Use of Variable density sections for tracking reflection continuity in such cases may be geologically flawed. One may prefer wiggle mode of seismic display that allows inferring geology guided by reflection character relying on the waveform shapes.

Color display is known to increase optical resolution leading to better visual discrimination of features and is used widely. The selection of suitable color and its encoding depends on the artistic attitude of the interpreter but assigning colors in a spectral progression is preferred as it enhances the relative magnitudes well.

Plotting Scales (Vertical and Horizontal)

Plotting scales are extremely important in data display, as reducing or stretching the scales changes visualization of the geologic objectives. The scales are to be suitably chosen depending on the objectivity of the interpretation. Horizontally compressed sections improve perceived continuity of events with gentle dips appearing stronger. Stretched sections, on the other hand, appear to deteriorate reflection continuity with flattening of the dips. Accordingly, faults with small displacement, low dipping progradations, gentle pinch outs and terminations etc., which are subtle but important as exploratory objects, look more conspicuous to be picked on compressed scales. Compressed (squashed) sections are also very useful for interpretation of regional geology for basin evaluation as a long stretch of a profile can be conveniently displayed in one vision frame at a time. Similarly, vertically compressed (half) sections offer the advantage of viewing the entire geological section from the deepest depth to the surface and help in better assessment of geological evolution of a basin. On the other hand, sections stretched in time are often used to magnify details of important targets to be picked for mapping. Each geologic object requires appropriate scales for its clear standout and needs experimenting for choosing the best judicious combination of both the vertical (time) and horizontal (trace) scales along with the mode of display.

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