Chapter 17

Seismic Exploration

Nazmul Haque Mondol

17.1 Introduction

Of all the geophysical exploration methods, seismic surveying is unequivocally the most important, primarily because it is capable of detecting large-scale to small-scale subsurface features. Simply stated, seismic methods involve estimation of the shapes and physical properties of Earth's subsurface layers from the returns of sound waves that are propagated through the Earth. Early wildcatters found oil by drilling natural oil seeps and large folds (anticlines) in exposed rocks. These easy oil prospects were all quickly discovered and drilled, and geologists then turned to seismic surveys to find less obvious oil and gas traps. Seismic technology had been used since the early 1900s to measure water depths and detect icebergs, and by 1924, seismic data were first used in the discovery of a Texan oil field (Milligan 2004). Several introductory and advanced textbooks (e.g., Telford et al. 1990, Sheriff and Geldart 1995, Yilmaz 2001) describe the principles of acquisition, processing and interpretation of seismic data. This chapter reviews the fundamental concepts employed in seismic exploration.

In general, two types of seismic method (reflection and refraction) are common, with reflection seismic the most widely used technique in hydrocarbon exploration. This technique provides an image of the subsurface in two or three dimensions (2D or 3D) (Fig. 17.1). The subsurface seismic images are produced by generating, recording and analysing sound waves that travel through the Earth (such waves are also called seismic waves). The density and velocity changes between rocks reflect the waves back to the surface, and how quickly and strongly the waves are reflected back indicates what lies below. Seismic pulses for exploration surveys are generated in one of three ways, employing an air-gun, vibrator or dynamite. An air-gun source is used for marine acquisition whereas vibrator and dynamite are the common sources for land seismic surveys. The strength of pulses associated with different seismic surveys varies, depending on site-specific factors such as rock types, how deep the survey needs to image and the required source.

17.2 Basic Principles

The first step in seismic exploration is to acquire data, which in most cases is carried out from the surface. To understand the seismic data, a review of the physical principles that govern the movement of seismic waves through layered media is necessary. A seismic source at any point on the Earth generates four types of seismic waves: compressional (P-wave), shear (S-wave), Rayleigh (ground roll) and Love, that travel through the layers (Fig. 17.2). Each layer will have a specific density and velocity. Rayleigh and Love waves are surface waves and propagate approximately parallel to the Earth's surface. Although surface waves penetrate to significant depth in the Earth, these types of waves do not propagate directly through the Earth's interior and have limited significance in oil and gas exploration. On the other hand, P- and S-waves are often called body waves because they propagate

N.H. Mondol (✉)
Department of Geosciences, University of Oslo; Norwegian Geotechnical Institute (NGI), Oslo, Norway
e-mail: nazmul.haque@geo.uio.no

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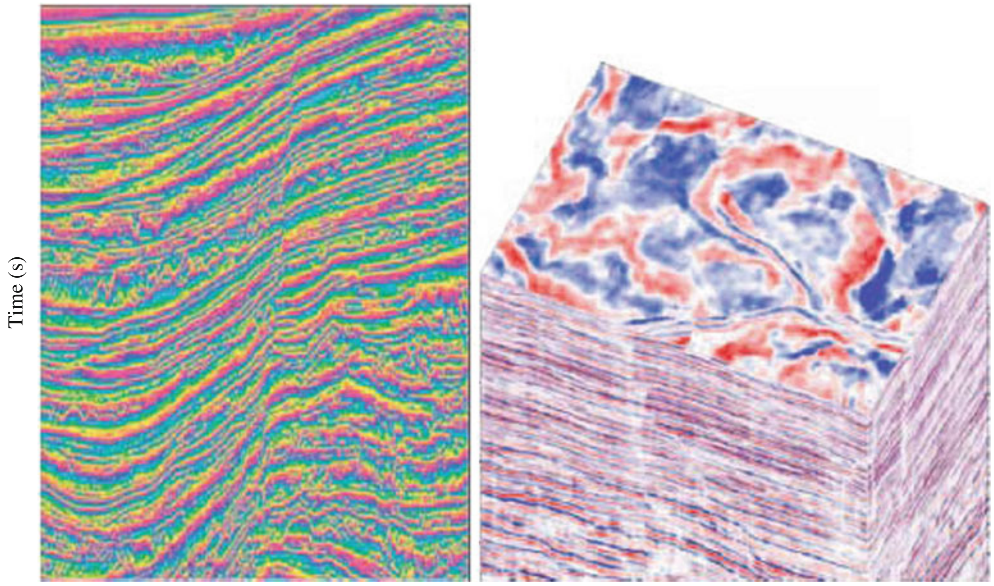

Fig. 17.1 Subsurface imaging by 2D (*left*) and 3D (*right*) seismic reflection data

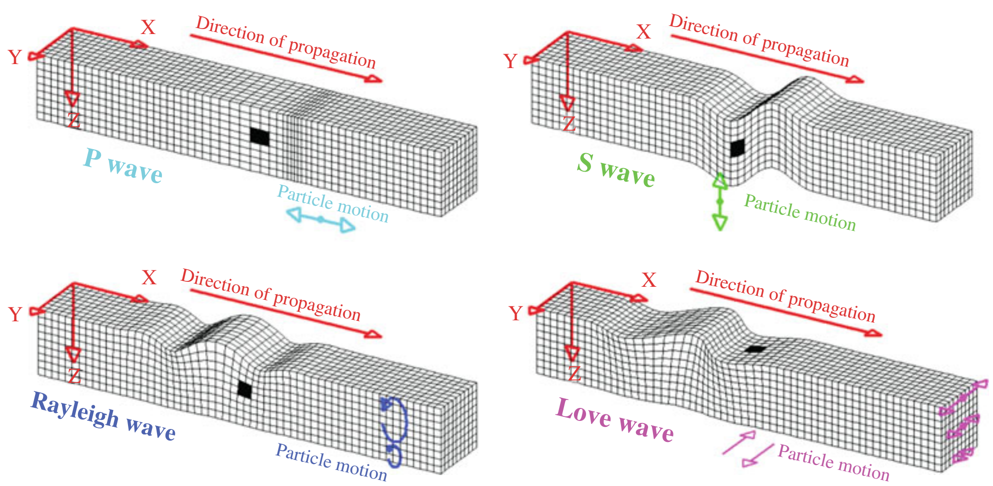

Fig. 17.2 Propagation of body (P and S) and surface (Rayleigh and Love) waves as a function of particle motions. P-waves shake the ground in the direction they are propagating, while S-waves shake perpendicularly to the direction of propagation. Rayleigh waves shake the ground both in the direction of propagation and perpendicular (in a vertical plane) so that the motion is generally elliptical – either prograde or retrograde. Love waves shake the ground perpendicular to the direction of propagation and generally parallel to the Earth's surface. (Source: Braile 2000)

outward in all directions from the source and travel through the interior of the Earth and have great significance in seismic exploration. P-waves move faster than S-waves. The P-wave is a longitudinal wave, the force applied in the direction that the P-wave is travelling. The ground must move in that direction. The ground or Earth is incompressible, so the energy is transferred pretty quickly. In the S-wave, the medium is displaced in a transverse way (up and down – compared to the line of travel), and the medium must move away from the material right next to it to cause the shear and transmit the wave. This takes more time, which is why the S-wave moves more slowly than the P-wave in seismic events. S-waves do not travel through fluids as fluid has no shearing capacity.

The basic principle of seismic survey is to initiate a seismic pulse from a *seismic source* at or near the Earth's surface and record the amplitudes and travel

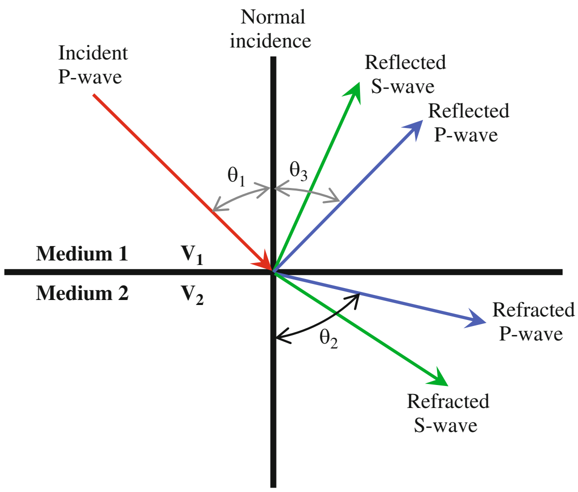

Fig. 17.3 A schematic diagram of reflected and refracted waves generated from an incident P-wave. The angle between the normal to the interface of two media and an incident P-wave is the angle of incidence (θ_1), and is equal to the angle of reflection (θ_3) in isotropic media. The angle of refraction (θ_2) depends on the velocity of the wave in that medium

times of waves returning to the surface after being reflected or refracted from the interface(s) of one or more layers. When a seismic source emits a pulse that propagates through the sedimentary layers, the sound waves travel between the layers with different velocities and will be refracted according to Snell's law:

$$\sin\theta_1/\sin\theta_2 = V_1/V_2 \tag{17.1}$$

where V_1 and V_2 are the velocities of the first and second media, $\sin\theta_1$ and $\sin\theta_2$ are the sines of the incidence and refracted angles, and θ_3 is the reflected angle (Fig. 17.3). Snell's law describes the changes in the direction of a wavefront as it travels in media of different velocities. If the seismic wave is incident at an angle, both reflected and refracted P- and S-waves will be generated at an interface between two media. However, at a fluid-solid interface like the seafloor, S-waves will not exist in the fluid part.

A line or grid of geophones or hydrophones called *seismic receivers* records the reflected and refracted seismic signals. Reflections from the layer interfaces in the subsurface are then measured at receivers (time measurements). If the two layers have different velocities, they will as a rule also have different densities, and part of the acoustic energy will not be refracted, but reflected. How much of the energy is reflected depends on the difference in the *impedance* [P-impedance (Z_p) or S-impedance (Z_s)], which are the

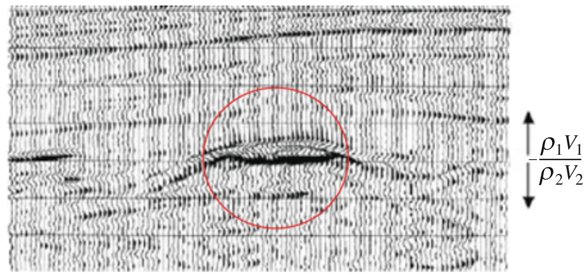

Fig. 17.4 A seismic section showing a large impedance contrast within the target zone (marked by *red circle*)

product of P-wave (V_p) or S-wave (V_s) velocities and density (ρ). The *reflection coefficient* (R) of a normally incident P-wave on a boundary is given by:

$$R = \frac{\rho_2 V_2 - \rho_1 V_1}{\rho_2 V_2 + \rho_1 V_1} \tag{17.2}$$

where ρ_1 and ρ_2 are the densities of the upper and lower layers, V_1 and V_2 are their respective P-wave velocities, and $\rho_1 V_1$ and $\rho_2 V_2$ are the P-impedances of the upper and lower layers respectively. Therefore, anything that causes a large contrast in impedance in the target zone can cause a strong reflection (Fig. 17.4). The possible candidates include changes in *lithology, porosity, pore fluid, degree of saturation* and *diagenesis*. We see that the greater the difference in density and velocity of two layers, the greater the amount of energy which will be reflected. Sandstone will often have significantly different acoustic impedance from shale, and a considerable amount of sound energy will be reflected from the boundary between a sandstone bed and a shale bed. Limestones will tend to have both high velocities and high densities. The result will be even greater contrast in acoustic impedance between limestones and, for example, shales. However, this contrast will always depend on the porosity of the limestone, though even rather porous limestones have relatively high velocities because they are usually well cemented.

17.3 Seismic Sources and Reservoirs

It is essential at this point to understand the principle of seismology. Seismology is based on the transmission of sound waves by the rocks of the crust. Strong earthquakes create pressure waves (natural sources of seismic waves) that are transmitted through the entire Earth and detected by seismographs (receivers) on the

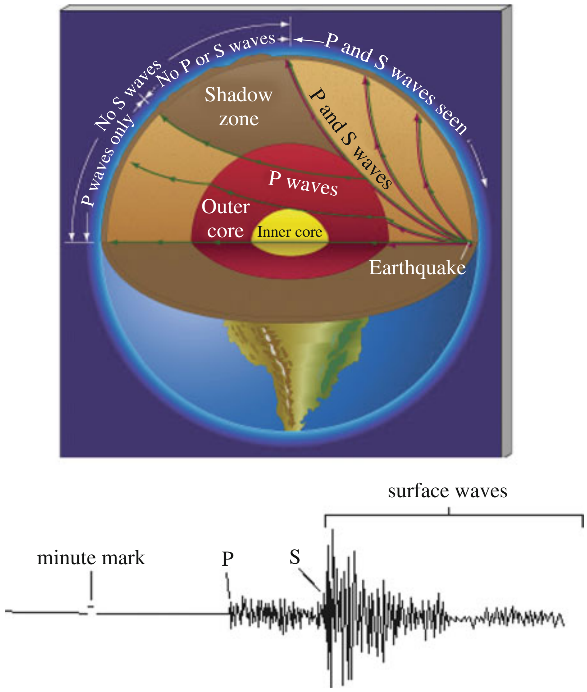

Fig. 17.5 A cross-section of the Earth with earthquake wave paths defined and their shadow zones highlighted (*top*). A typical seismogram (*bottom*) shows the fastest P-waves. The next set of seismic waves is the S-waves. The surface waves (Love and Rayleigh waves) travel a little slower than S-waves (which, in turn, are slower than P-waves) and have lower frequency

other side (Fig. 17.5). Seismic exploration, however, as employed by the petroleum geologist, makes use of artificially generated pulses *(seismic sources)*. The principle is simple; an impulse source sends acoustic energy into the Earth. This energy propagates in many directions and is reflected and refracted when it encounters boundaries between two layers. Sensors (*seismic receivers*) placed on the surface measure the reflected or refracted acoustic energy. These artificial sources are much weaker than the natural seismic source (earthquake) but are more focused towards areas of specific stratigraphic interest.

17.3.1 Seismic Sources

Different seismic sources are usually used in land and marine acquisitions. In marine environments seismic energy is normally generated using arrays of air-guns, whereas in land seismic one often uses explosives or vibrators. An *air-gun* is a device that releases highly compressed air (at typically 2,000–5,000 psi) into the water surrounding the gun (Fig. 17.6b). A *vibrator* is an adjustable mechanical source that delivers vibratory seismic energy into the ground (Fig. 17.6a). A vibrator source sends a controlled-frequency sweep into the ground. The recorded data are then convolved with the original sweep to produce a usable signal. *Dynamite* – a combination of explosive and detonator, is used as a seismic source. The detonator helps to ignite the explosives. When dynamite ignites, a shock wave propagates with a speed of 3,000–10,000 m/s. It provides an impulsive energy that can be converted into ground motion. It is customary to drill a hole to load dynamite and fill it with heavy mud before shooting. Dynamite can generate usable signal strengths and a bandwidth that covers a wide spectrum of seismic energy. It includes a variety of energy sources based on varying explosive output parameters to meet geological and climatic conditions.

17.3.2 Seismic Receivers

Hydrophones and *geophones* serve as receivers for seismic signals. The *hydrophone* is a device designed for use in detecting seismic energy in the form of pressure changes in water during marine seismic acquisition (Fig. 17.7a). It measures pressure variations with the aid of piezoelectric material, which generates a voltage upon deformation. The two piezoelectric elements in one hydrophone are connected and polarised so that voltages due to pressure waves (returning signal) add and voltages due to one-directional acceleration will cancel. In this way the influence of movements due to currents, wave action and so on will be minimised. Hydrophones are combined to form streamers that are towed by seismic vessels or deployed in a borehole. A typical length of a streamer is about 4–6 km where a single receiver section is typically 75 m long and contains 96 hydrophones which are grouped in arrays of a predefined length, mostly 12.5 or 25 m.

The *geophone* is a device used in surface seismic acquisition, both onshore and on the seabed offshore, that detects ground velocity produced by seismic waves and transforms the motion into electrical impulses (Fig. 17.7b). Geophones, unlike hydrophones, detect motion rather than pressure. Conventional seismic surveys on land use one geophone or a group of geophones per receiver location to detect motion in the vertical direction. The three-component

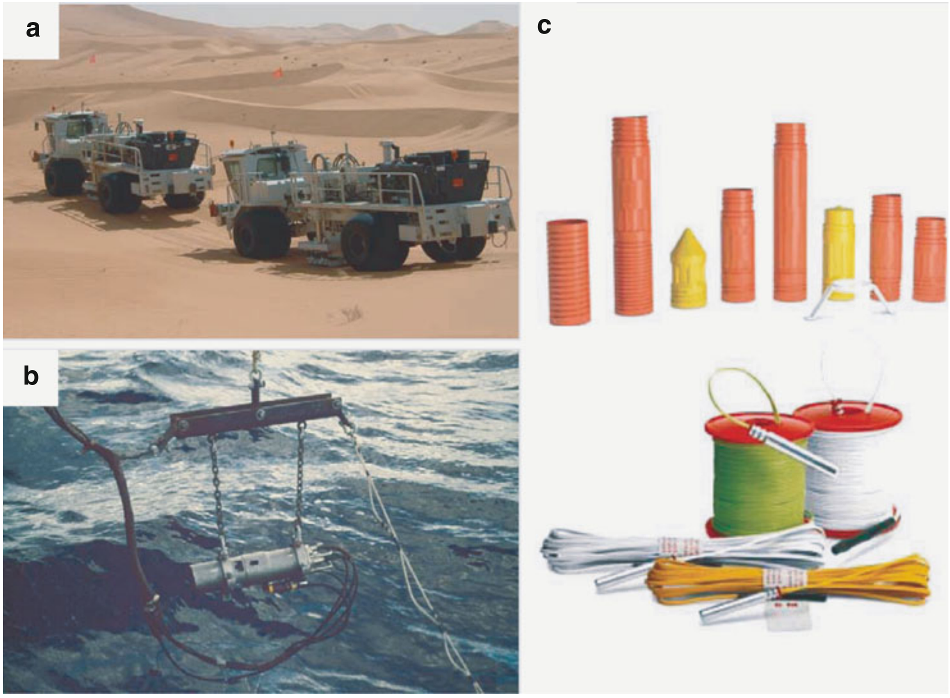

Fig. 17.6 (**a**) Seismic acquisition in a desert where vibrator is the most common source of seismic energy. (**b**) An air-gun before deployment in the water. It releases compressed air into the water during marine seismic survey. (**c**) Explosives (*top*) combined with detonators (*bottom*) form dynamite which is a common source of seismic energy in land acquisition where vibratory trucks can not be used due to rugged topography

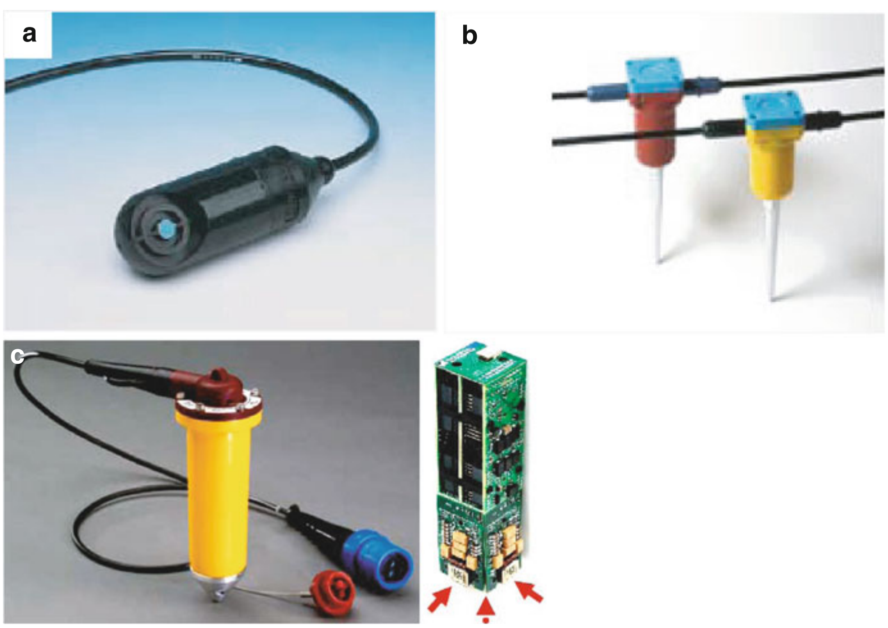

Fig. 17.7 (**a**) Hydrophone, (**b**) geophone and (**c**) multi-component geophone. (Courtesy ION Geophysical)

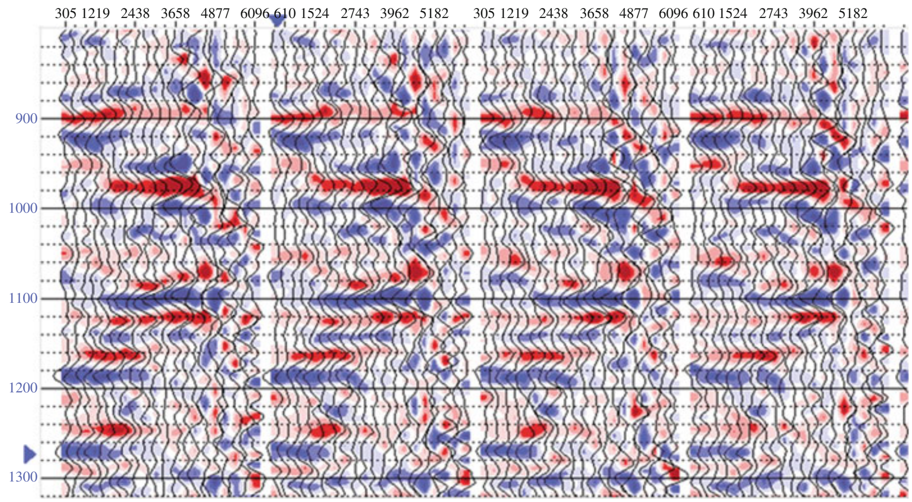

Fig. 17.8 Recorded seismic traces. Each trace consists of one recording corresponding to a single source-receiver pair. Seismic energy recorded at the receivers arrives at different times because of distance of receivers from source. In conventional acquisition, strings of geophones hard-wired together average the individual sensor measurements and deliver one output trace

(3C) geophone is used for direct measurements of shear waves at the seafloor. An essential feature of a seafloor seismic acquisition is the four-component (4C) detector unit, which includes a hydrophone and a three-component (3C) geophone (Fig. 17.7c). The hydrophone and the vertical geophone measure pressure waves, whereas the two extra horizontal geophones measure particle velocity associated with shear wave energy.

17.4 Seismic Acquisition

The seismic survey is an essential part of the whole cycle of petroleum exploration and production. Seismic surveys are carried out on land and in transition zone, shallow marine and marine environments in different ways. The basic principle is an impulse source such as dynamite, air-gun or vibrator that sends acoustic energy into the Earth. This energy propagates in many directions. Downward travelling energy reflects and refracts when it encounters boundaries between two layers with different acoustic properties (Fig. 17.3). Sensors or geophones placed on the surface measure the reflected acoustic energy, converting it into an electrical signal that is displayed as a seismic trace (Fig. 17.8). The typical recorded seismic frequencies are in the range of 5–100 Hz. P-waves are the waves generally studied in conventional seismic data. P-waves incident on an interface at other than normal incidence angle can produce reflected and transmitted S-waves. S-waves travel through the Earth at about half the speed of P-waves and respond differently to fluid-filled rocks, and so can provide different additional information about lithology and fluid content of hydrocarbon-bearing reservoirs.

The recorded *seismic trace* is a *convolution* (*) of the *source signal* and the *reflectivity sequence* of the Earth plus *noise* (Fig. 17.9). A seismic trace can simply be expressed by the Eq. (17.3) where multiples are not considered. Transmission losses and geometric spreading are not included and the frequency-dependent absorptions are also ignored in the equation.

$$S = W * R + \text{Noises} \tag{17.3}$$

where S is the recorded seismic trace, R is the reflectivity and W is the wavelet. A *wavelet* is a kind of mathematical function used to divide a given function into different frequency components and study each component with a resolution that matches its scale. Accurate wavelet estimation is absolutely critical to the success of any seismic inversion. The inferred shape of the seismic wavelet may strongly influence the seismic inversion results and therefore subsequent assessments of the reservoir quality. *Attenuation* (amplitude loss) of seismic waves is an

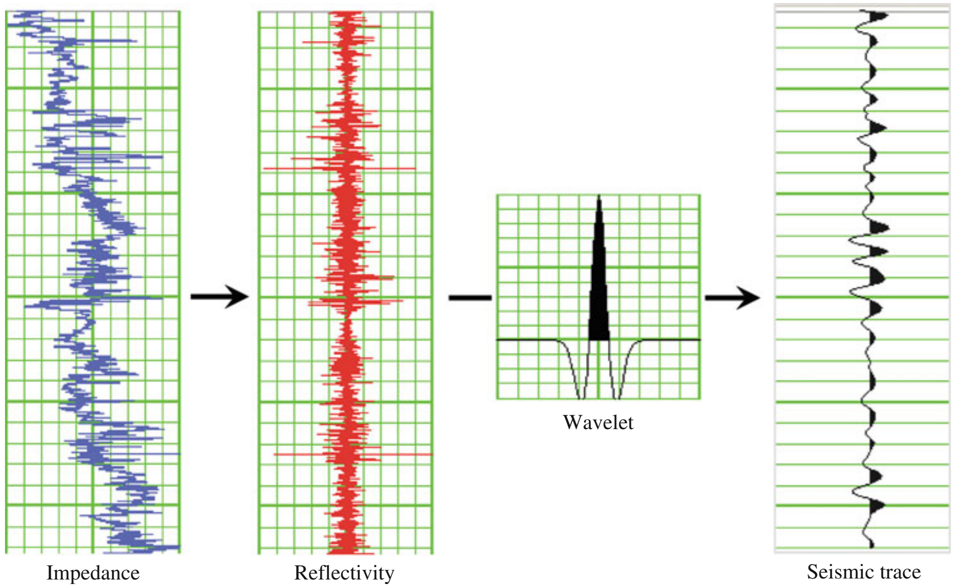

Fig. 17.9 Seismic trace is a result of convolution of a wavelet and the reflectivity series plus noises

important phenomenon and caused by three major factors: (a) Geometric spreading: progressive diminution of amplitude (proportional to the inverse of propagation distance) caused by the increase in wavefront area, (b) Intrinsic attenuation: energy losses due to internal friction and (c) Transmission losses: reduction in wave amplitude due to reflection at interfaces.

17.4.1 Terminology Used in Seismic Acquisition

A *Trace* is a seismic time measurement corresponding to one source-receiver pair, and *offset* is the distance between source and receiver for a given trace. In practice, traces from one *source* are simultaneously recorded at several *receivers*. Then, sources and receivers are moved along the survey line and another set of recordings is made. When a seismic wave travels from a source to a reflector and then back to receiver, the elapsed time is called the *two-way traveltime*. The *common depth point* (CDP) is the halfway point of the path only where the Earth is horizontally layered; it is situated vertically below the *common midpoint* (CMP). A *gather* is a family of traces (e.g. *shot-point gather* is the family of all traces corresponding to the same source firing). *Sorting* of traces by collecting traces that have the same *midpoint* (CMP) is called a *common midpoint gather* (CMP-gather). The number of traces summed or stacked is called a *fold*. For instance, in *24-fold* data, every *stacked trace* represents the average of 24 *traces*. The nominal *fold (F)* is the maximum number of *traces* in a *CMP-gather*. For a standard marine seismic acquisition *folds* (*F*) is given by the formula:

$$F = \frac{N\Delta g}{2\Delta s} \tag{17.4}$$

where N is the number of channels, Δg is the group interval and Δs is the source interval. The rate of repetition of complete wavelengths of seismic waves is referred to as *frequency* (*f*) and is measured in cycles per second or *hertz*.

17.4.2 Marine Acquisition

Marine seismic acquisition is generally accomplished using large ships with one or multiple air-gun arrays for sources (Fig. 17.10). Air-guns are deployed behind the seismic vessel and generate a seismic signal by forcing highly pressurised air into the water at a given interval. Receivers are towed behind the ship in one or several long streamer(s) that are several kilometres long. Marine receivers are composed of piezoelectric

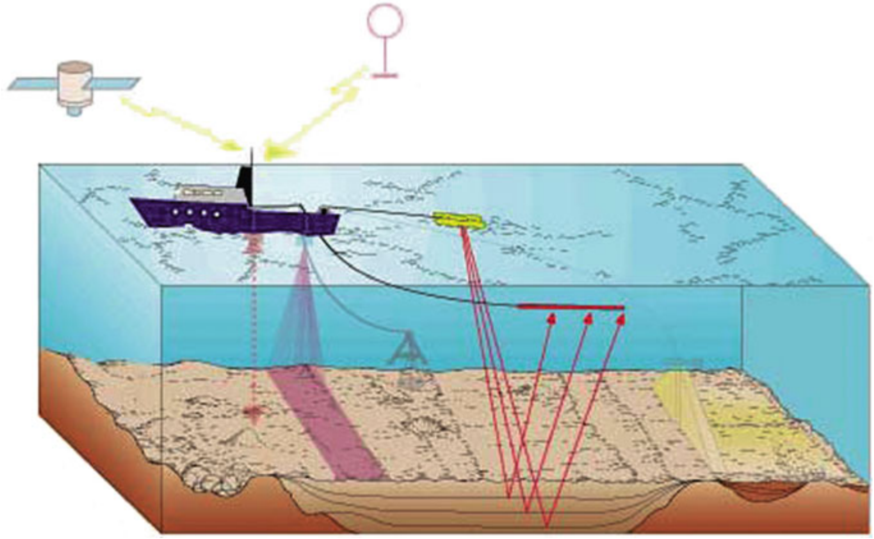

Fig. 17.10 A 2D marine seismic acquisition

hydrophones, which respond to changes in water pressure. In marine acquisition, seismic vessels sail along predetermined patterns of parallel circuits (Fig. 17.11). The length of straight segments is calculated from fold plots, and must include additional length – run in and run out – to allow the cable to straighten after each turn. Marine seismic acquisition is faster than conventional land seismic because it does not require jug hustlers to lay and pick up geophones (Rygg et al. 1992).

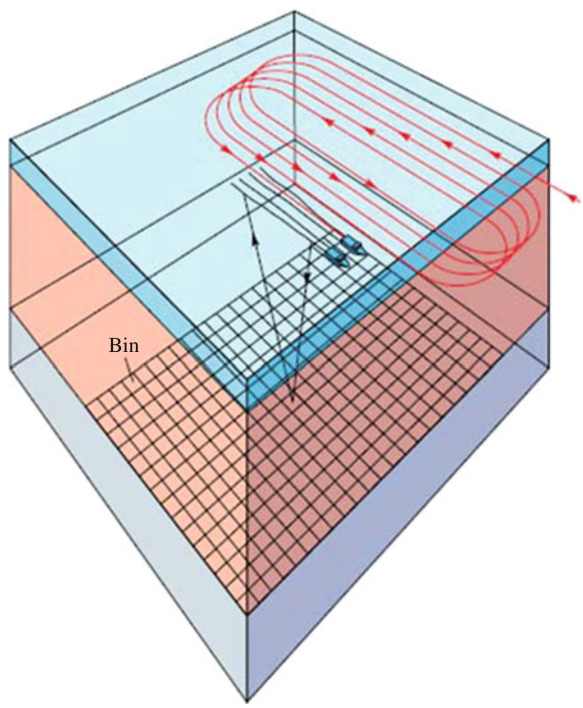

Fig. 17.11 Marine acquisition geometry showing seismic vessels looping in oblong circuits. (Source: Ashton et al. 1994)

The advance to 3D seismic acquisition and imaging of the subsurface, introduced in the 1980s, was perhaps the most important step in seismic exploration (Beckett et al. 1995). The 3D seismic images began to resolve the detailed subsurface structural and stratigraphic conditions that were missing or not discernable from 2D seismic data. With 3D seismic acquisition potential reservoirs are imaged in three dimensions, which allows seismic interpreters to view the data in cross-sections along 360° of azimuth, in depth slices parallel to the ground surface, and along planes that cut arbitrarily through the data volume. Information such as faulting and fracturing, bedding plane direction, the presence of pore fluids, complex geological structure, and detailed stratigraphy are now commonly interpreted from 3D seismic data sets.

In 2D marine data acquisition a single streamer is deployed, whereas in 3D acquisition multiple streamers are towed behind the boat (Fig. 17.12). 3D acquired data can be processed in a more consistent manner and be further manipulated using modern visualisation tools. In 2D data acquisition the data collection occurs along a line of receivers. The

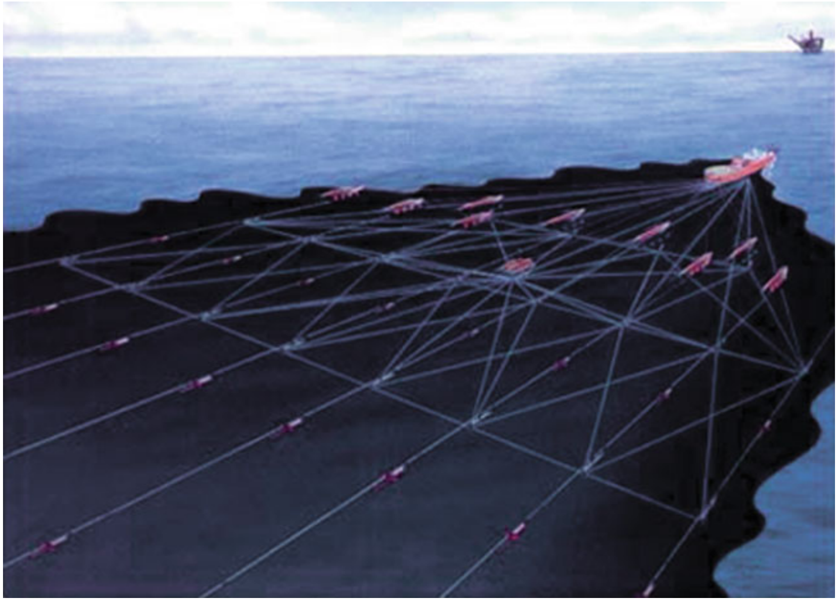

Fig. 17.12 Showing a towed streamer for 3D seismic acquisition. (Courtesy ION Geophysical)

resultant image represents only a section below the line. Unfortunately, this method does not always produce a clear subsurface image. 2D data can often be distorted with diffractions and events produced from offline geologic structures, making accurate interpretation difficult. Because seismic waves travel along expanding wavefronts they have a surface area. A truly representative image of the subsurface is only obtained when the entire wavefield is sampled.

A 3D seismic survey where the vessel is towing many streamers (up to 20) at the same time with multiple arrays of airguns is more capable of accurately imaging reflected waves because it utilises multiple points of observations (Fig. 17.12). In the case of 3D survey we have seen that seismic data are sampled from a range of different angles (azimuth) and source-receiver distances (offsets). After seismic processing the data can then be represented as 3D volume images of the subsurface. In 2D processing, traces are collected into *CMP gathers*, while in 3D, traces are collected into *common-cell gathers (binning)*. To perform 3D binning, a grid is first superimposed on the survey area. This grid consists of cells with dimensions of half the receiver group spacing in the inline direction, equivalent to the CMP spacing in 2D processing, and the line spacing in the crossline direction (Fig. 17.11). In reality, midpoint distributions within a cell are not necessarily uniform since cable shape varies from shot to shot and line to line. Such side drift of the cables is called *feathering*.

Another advanced technique of marine seismic is ocean bottom seismic (OBS) acquisition. It gives the possibility of direct measurement of S-wave data in addition to P-wave data by using ocean bottom cables that have three component geophones (3C) and a hydrophone in addition (thus 4C in total). The 4C cable can be up to 6 km long with 240 stations (i.e. 960 channels since 4C). A typical OBS layout involves 4 or more cables (Fig. 17.13). The optimal choice of acquisition geometry for a 4C survey hinges on both geophysical and financial considerations. Most designs can be classified as either *patch* or *swath*. In *swath* designs, the source lines are parallel to receiver lines, while in *patch* designs, source lines are perpendicular to receiver lines. *Patch* design produces seismic data of a relatively wide range of azimuths, whereas the *swath* design produces data of a limited or narrow range. Moreover, the *swath* design offers a more uniform sampling of offsets with better near-offset coverage. The use of 4C OBS recording has several advantages over conventional towed streamer technology, which includes:

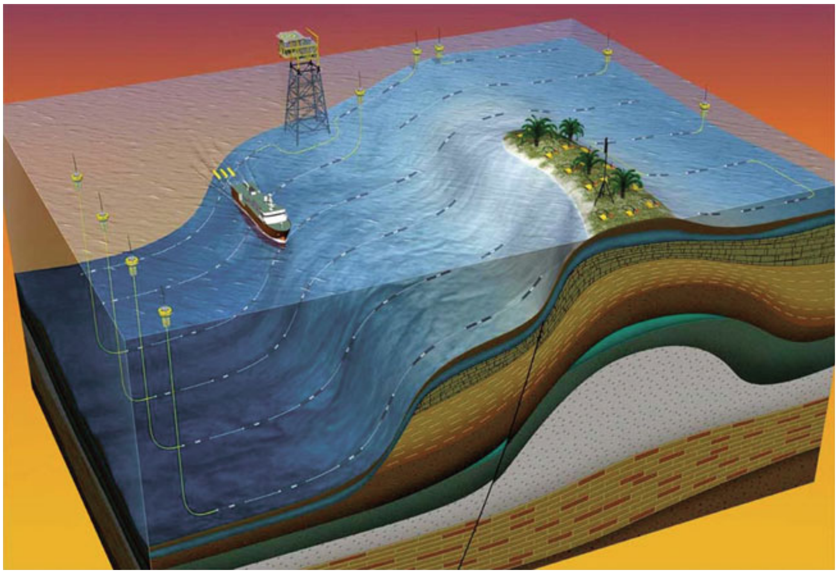

Fig. 17.13 An Ocean Bottom Seismic (OBS) marine acquisition technique. (Courtesy ION Geophysical)

(1) Dual-sensor summation (3C geophone + hydrophone) for the suppression of receiver-side multiples.
(2) Utilising P–S wave conversions for enhanced imaging (Fig. 17.14).
(3) Attenuation of free surface multiples when combined with towed streamer recording.

A comparison of the migrated P–S stack versus the P–P stack is shown in Fig. 17.15. The P–S stack is produced from OBS converted wave data whereas the P–P stack is produced from 3D towed streamer P-wave data. From this comparison it is clear that OBS data can be used to successfully image through a gas chimney.

17.4.3 Land Acquisition

A complication in land acquisition is that, unlike marine data, a seismic line is rarely shot in a straight line because of the presence of natural and man-made obstructions such as lakes, buildings and roads (Fig. 17.16). The shot points and the receivers may be arranged in many ways. Many groups of geophones are commonly used on a line with shot points at the end or in the middle of the receiver array. The shot points are gradually moved along a line of geophones. The variations in ground elevation in land acquisition causes sound waves to reach the recording geophones with different traveltime. The Earth's near-surface layer may also vary greatly in composition, from soft alluvial sediments to hard rocks. This means that the velocity of sound waves transmitted through this surface layer may be highly variable. *Static corrections* – a bulk time shift applied to a seismic trace – are typically used in seismic processing to compensate for these differences in elevations of sources and receivers and near-surface velocity variations (Ongkiehong and Askin 1988).

17.4.4 Vertical Seismic Profile (VSP)

VSP is another technique of seismic acquisition, used for correlation with conventional seismic data (land or marine seismic). The defining characteristic of a VSP (of which there are many types) is that either the energy source, or the receivers (or sometimes both) are in a borehole. In the most common type of VSP, hydrophones, or more often geophones, in the borehole record reflected seismic energy originating from a seismic source at the surface (Fig. 17.17).

The VSPs vary in the well configuration, the number and location of sources and geophones, and how

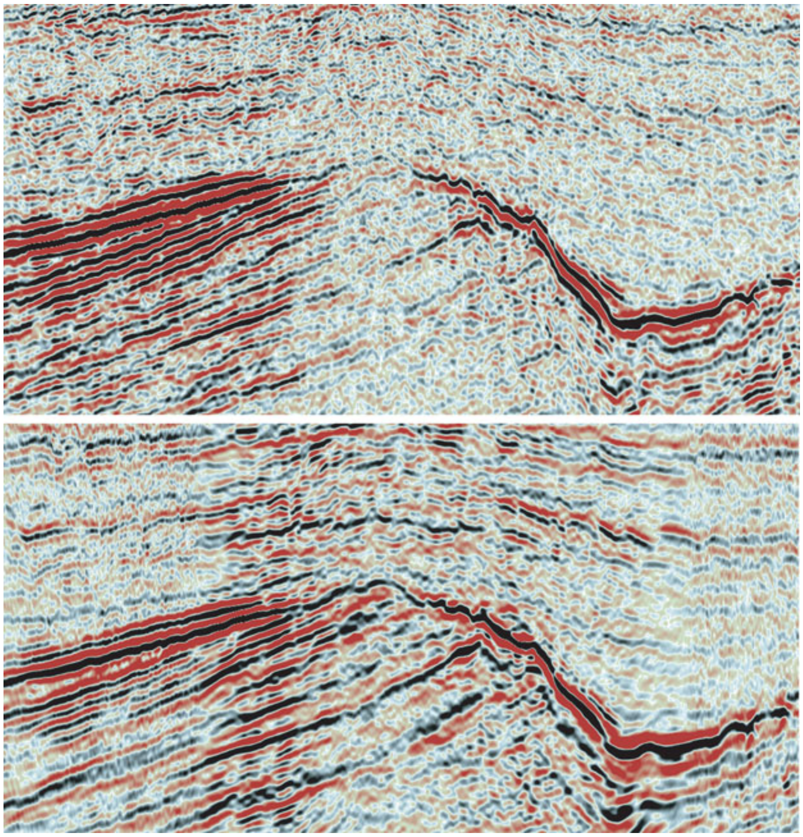

Fig. 17.14 A comparison of seismic data acquired by the towed streamer (*top*) and OBS (*bottom*) techniques. The OBS survey significantly improves the subsurface image. (Source: Thompson et al. 2007)

they are deployed. VSP uses the reflected energy contained in the recorded trace at each receiver position as well as the first direct path from source to receiver. VSPs include the zero-offset VSP, offset VSP, walkaway VSP, walk-above VSP, salt-proximity VSP, shear-wave VSP, and drill-noise or seismic-while-drilling VSP.

17.4.5 4D Seismic

The acquisition of *4D* or *time-lapse seismic* has opened new horizons for monitoring reservoir properties such as fluids, temperature, saturation and pressure changes during the productive life of a field (Aronsen et al. 2004). 4D seismic is based on the analysis of *repeated 3D* seismic data. The differences in seismic attributes over time are caused by changes in pore fluid and pore pressure associated with the drainage of a reservoir under production. Detection of areas with significant changes or with virtually unchanged hydrocarbon-indicating attributes helps to determine new drilling sites in an already existing production field. For this method it is critical that the observed seismic changes can be related to the fluid flow. Differences in data acquisition, survey orientation, processing and data quality can introduce significant noise in a 4D analysis. Hence, such differences must be corrected for as best as possible. Further details of 4D seismic are discussed in Chap. 19. The known applications of 4D seismic can be summarised as:

(1) Monitoring the spatial extent of steam injection used for thermal recovery.
(2) Monitoring the spatial extent of the injected water front used for secondary recovery.
(3) Imaging bypassed oil or gas.
(4) Determining the flow properties of sealing or leaking faults.
(5) Detecting changes in oil-water contact.

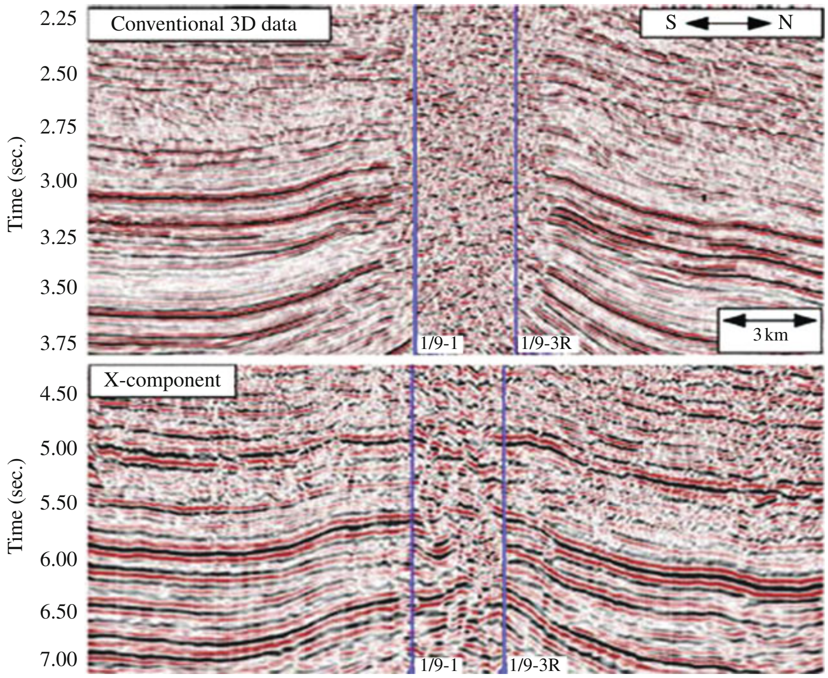

Fig. 17.15 Comparison of P–P stack of conventional 3D streamer data (*top*) and P–S stack of OBS data (*bottom*). Note how the OBS data produces a much better deeper image in the presence of gas versus 3D streamer data. (Source: Granli et al. 1999)

Fig. 17.16 Land seismic acquisition

17.4.6 Permanent Seismic Monitoring

Permanent seismic monitoring is becoming an important tool in the reservoir management toolkit. It is a 4C fiber-optic advanced seismic acquisition technology that is installed permanently on the seabed over a producing field (Fig. 17.18). It reduces acquisition time and cost. Permanent seismic monitoring helps to improve data quality by employing more accurate survey orientation and acquisition geometry (receiver locations) within the repeated 3D seismic surveys compared to conventional OBS 4D survey. Such a method is important in monitoring a reservoir injection process employed to enhance recovery from a producing reservoir.

17.5 Seismic Processing

Seismic technology has achieved amazing feats in exploration and production activities in the past few decades. What we record in the acquisition stage is

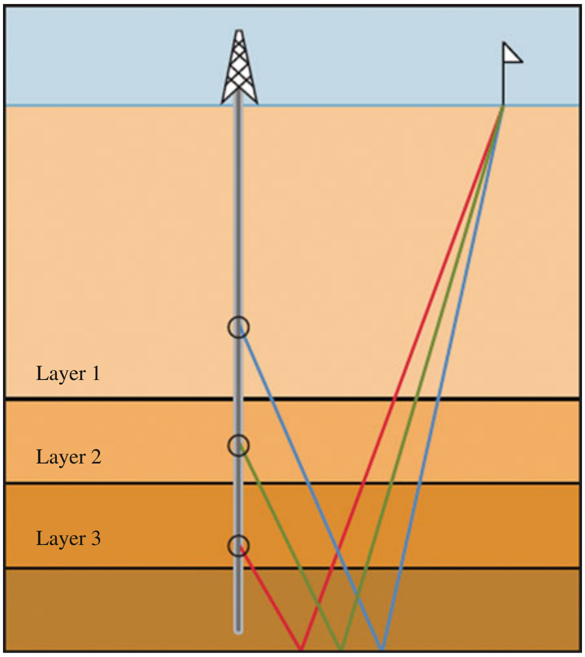

Fig. 17.17 Acquisition of VSP. The downhole geophones record important structural and stratigraphic data generated by a surface energy source

Fig. 17.18 Permanent installation of 4C cables on the sea *bottom* over a producing field. It improves data quality by ensuring more accurate receiver locations within the repeated 3D surveys over a period of time

called *raw seismic data* which contains real signals together with *noise* and *multiples*. This raw data must then be processed by employing advanced methods within signal processing and wave-theory to get better images of the subsurface. The prime objective in the processing stage is to enhance the signal and suppress the coherent and noncoherent noises and multiples (Fig. 17.19). *Coherent noise* is unwanted seismic energy that shows consistent phase from one seismic trace to another. This may consist of waves that travel through the air at very low velocities such as airwaves or air blast, and ground roll that travels through the top of the surface layer, also known as the *weathering layer*. *Multiples* are internal reflections in a layer, which occur when exceptionally large reflection coefficients are present. In marine seismic the water-bottom multiples normally dominate. *Noncoherent energy* is typically nonseismic-generated noise, such as noise from wind, moving vehicles, overhead power line or high-voltage pickup, gas flares and water injection plants.

It has been stated earlier that seismic processing is the alteration of seismic data to suppress noise, enhance signal and migrate seismic events to the appropriate location in space. Seismic processing facilitates better interpretation because subsurface structures and reflection geometries are more apparent. The typical *sampling rate* of seismic acquisition is 2 ms. Digital recording of the incoming wavefield at densely spaced receiver positions ensures that the recorded signal and noise are properly *sampled* and are therefore unaliased. *Aliasing* is the ambiguity that arises because of insufficient sampling. It occurs when the signal is sampled less than twice the *cycle*. The highest *frequency* defined by a sampling interval is termed the *Nyquist frequency* and is equal to the inverse of $2\Delta t$, where Δt is the sampling interval. Frequencies higher than the Nyquist frequency will be folded back. In the noise-free case, aliasing can be avoided by a finer spatial sampling that is at least twice the Nyquist frequency of the *waveform*.

It is important to note that no standard processing sequence exists which can be routinely applied to all types of raw seismic data. The actual sequence will be determined by (a) the purpose of the investigation, (b) extensive testing on selected parts of the dataset and (c) a trade-off between quality and cost. The 2D seismic processing steps typically include static corrections, deconvolution, velocity analysis, normal and dip moveout, stacking and migration.

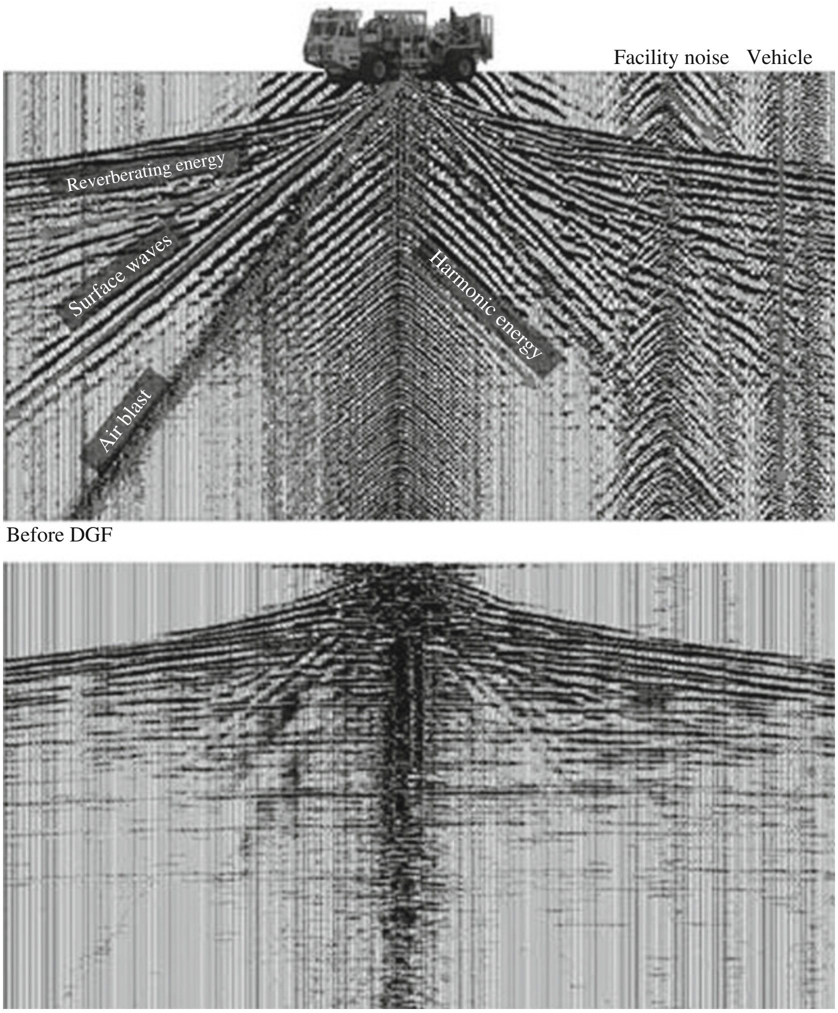

Fig. 17.19 Raw seismic data with coherent and noncoherent noise (*top*). Noise attenuation image after autocorrelation, deconvolution and trace muting (*bottom*). (Courtesy Western Geco)

The following routines are generally applied to raw 2D seismic data in different processing stages:

- True Amplitude Recovery (TAR)
- Autocorrelation
- CMP sorting
- Deconvolution
- Trace Muting
- Velocity picking or velocity analysis
- NMO correction
- DMO correction
- Filtering (F-K and Bandpass filtering)
- Stacking and
- Migration

In the processing stage, bad measurements are edited, datuming applied and corrections of wave-energy decay introduced. The *true amplitude recovery* is applied to increase the amplitude at large travel times. The *autocorrelation* and *deconvolution* are done to compress the wavelet and to attenuate multiples. *Deconvolution* – a technique that can compress the source signature and eliminate multiples – is applied after sorting the data into CMP gathers. The *trace muting* is applied to get rid of unwanted energy. Contributions from the direct waves and possible head waves are removed by trace muting. *NMO correction* and *F-K filtering* are usually applied to attenuate

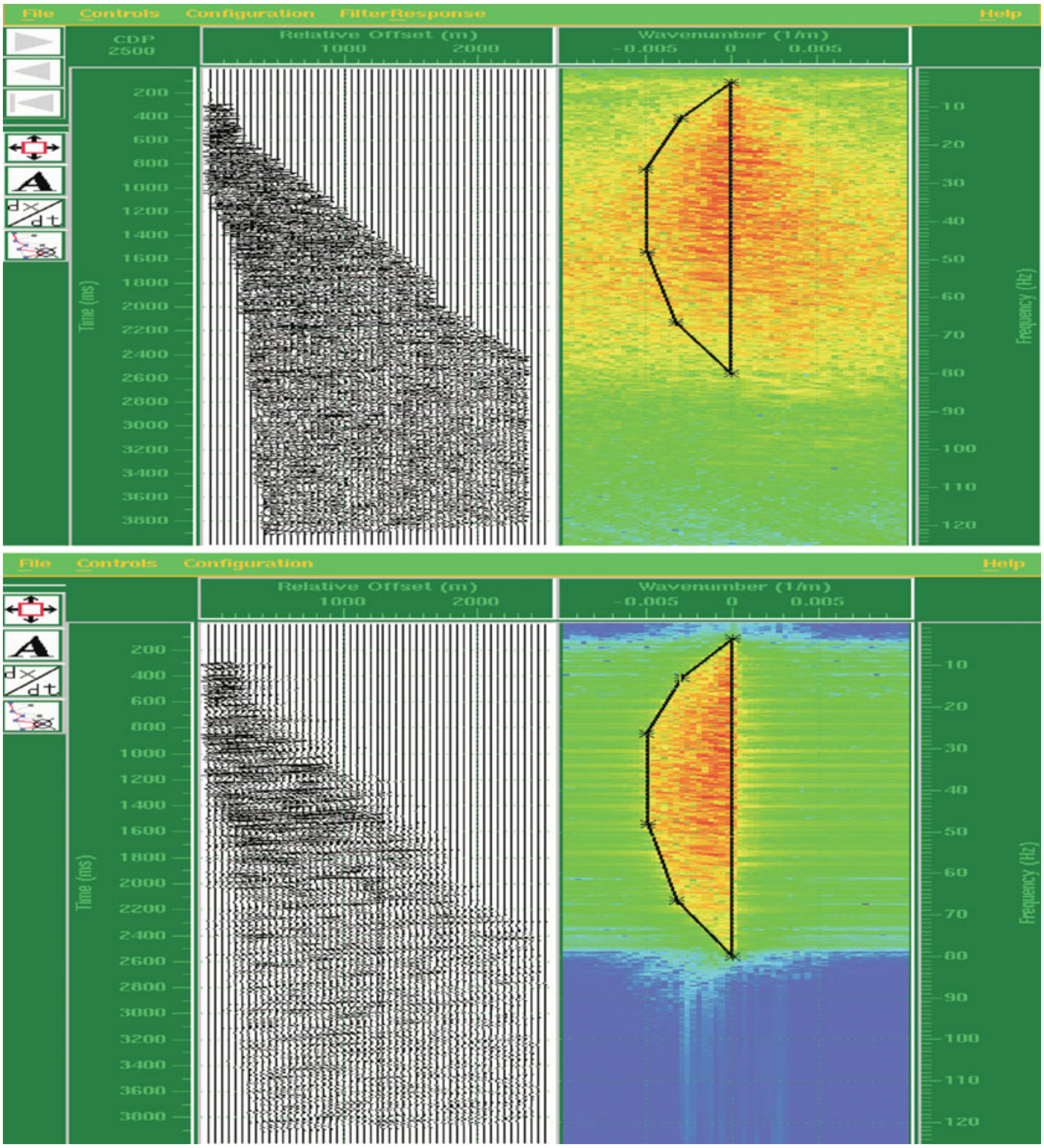

Fig. 17.20 (**a**) A CMP-gather before the F-K filtering: the primaries dipping up and the multiples dipping down in a time-distance display. The F-K domain (*top*, *right*) shows energy distributions of both primary and multiples energy, respectively. (**b**) The same CMP gather after F-K filtering. The F-K filtering accepted only primary energy (within *polygon*) and filtered out multiples energy (*bottom*, *right*)

multiples. Linear coherent noises are also removed by employing F-K filtering (Fig. 17.20).

NMO correction is applied from a space-variant velocity field assuming a horizontal reflector and hyperbolic normal moveout algorithm. The NMO is the difference between the travel time for a certain offset (X) and the vertical (zero-offset) traveltime $T(0)$. Velocities are interpolated for each CDP. Normal moveout is applied according to the following formula:

$$T(X) = \sqrt{\left[T^2(0) + \left(\frac{X}{V}\right)^2\right]} \qquad (17.5)$$

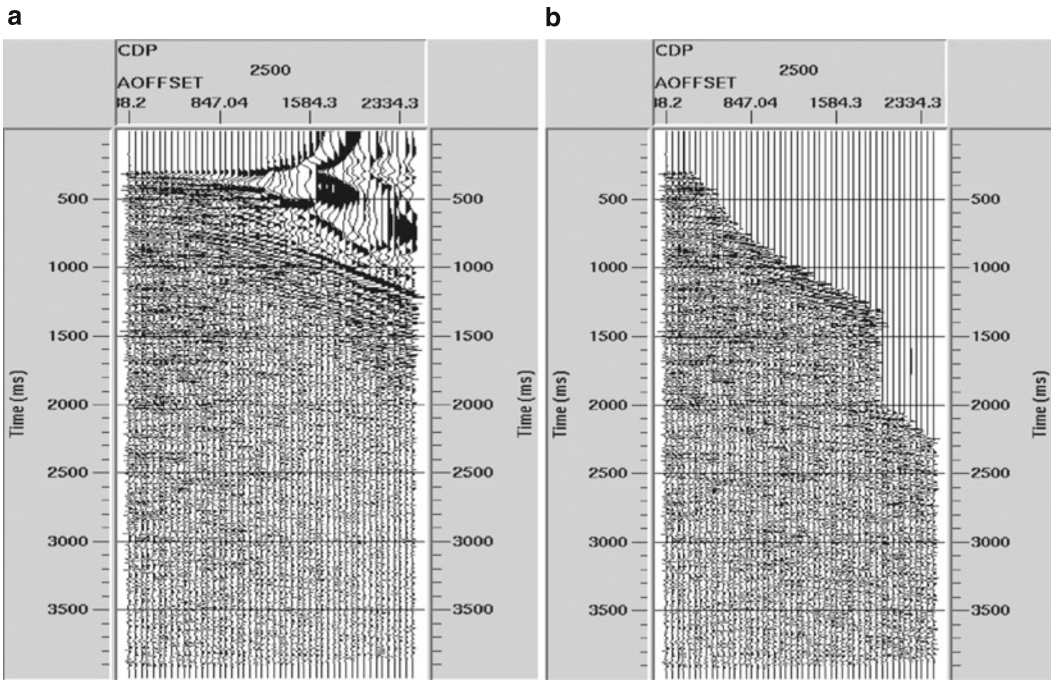

Fig. 17.21 NMO corrected CDP gathers show NMO stretch (**a**) and stretch muting (**b**) at the far offsets. Muting to remove NMO stretch may destroy far offsets information

where $T(X)$ is the two-way travel time for a seismic event, X is the actual source-receiver offset distance, V is the NMO or stacking velocity for this reflection event and $T(0)$ is the two-way travel time for zero offset. A sample-by-sample velocity is built at each of the locations where time-velocity pairs are defined. For any point before the first velocity location or beyond the last location, the first or last velocity function is used. Once the correct velocity function has been interpolated, the exact moveout at each sample is computed based on the actual source-to-receiver offset and velocity at that time sample. *NMO stretch* is a fundamental and long-standing problem in seismic processing. After normal moveout correction the early events are stretched at the far offsets (Fig. 17.21a). If we stack this unmuted gather, the early events suffer a severe loss of high-frequency energy, and thus resolution. This can appreciably reduce the interpretability of the section. There have been many attempts to solve the NMO stretch problem. The most universal is front-end or stretch muting, where samples at the beginning of a trace that have suffered severe NMO stretch are zeroed out (Fig. 17.21b). Stretch muting may leave very little fold at early times, reducing the noise suppression provided by stacking.

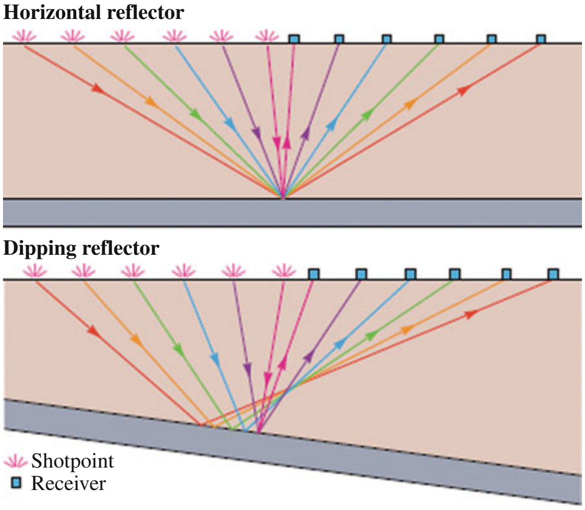

Fig. 17.22 Effect of reflector dip on the reflection point. When the reflector is flat (*top*) the CMP is a common reflection point. When the reflector dips (*bottom*) there is no CMP. A dipping reflector may require changes in survey parameters, because reflections may involve more distant sources and receivers than reflection from a flat layer

In the case of dipping beds, there is no common depth point shared by multiple sources and receivers, so *dip-moveout* (DMO) processing becomes necessary to reduce smearing or inappropriate mixing of data (Fig. 17.22). DMO is not routinely employed; however, it can be useful to solve the problem of conflicting dip.

In practice, a *velocity analysis* has to be carried out, where the main purpose is to determine the velocity distribution as a function of time which will give the most accurate NMO correction. The velocity analysis is used to compute corrections in traveltime that will be applied to all the traces belonging to a CMP-gather. The most common type of velocity analysis is to repeat the procedure of correcting and stacking CMP data for many different velocities and within a discrete time window. By measuring the average absolute value of the data or more precisely the *semblance* within time-windows of different test velocities, and plotting these results in a time versus velocity histogram, it is possible to interpret the velocity information. This type of plot is denoted a *velocity spectrum*. In general, the interpreted velocity function is picked so that it goes through areas with highest values. The basic assumptions in the CMP method velocity analysis is that the geological model corresponds to a slowly varying velocity as a function of depth. In addition, it is assumed that drastic lateral velocity changes do not appear. Then we can interpret high semblance values to be usually related to *primaries* and low semblance values to be related to *multiples* (Fig. 17.23). If the velocity does not increase with depth in some areas it is much more complicated to distinguish between primaries and multiples. If these basic assumptions of velocity analysis are violated both the velocity analysis and the stacked section can be distorted. Such velocity anomalies can be caused by:

- Diffractions (edges, faults)
- Shallow gas pockets
- Multiples
- Complicated reflection sequence (interference)
- Dip
- Reflection from the side (out-of-plane).

Stacking is an important step in seismic processing. Stacking represents summation of *NMO*-corrected traces in a CMP family. The collection of stacked traces forms a seismic section which gives an image (slice) of the subsurface (Fig. 17.24b). The stacking process has two major advantages: (a) it increases the signal-to-noise (S/N) ratio and (b) it amplifies primary energy relative to multiple energy. This second point

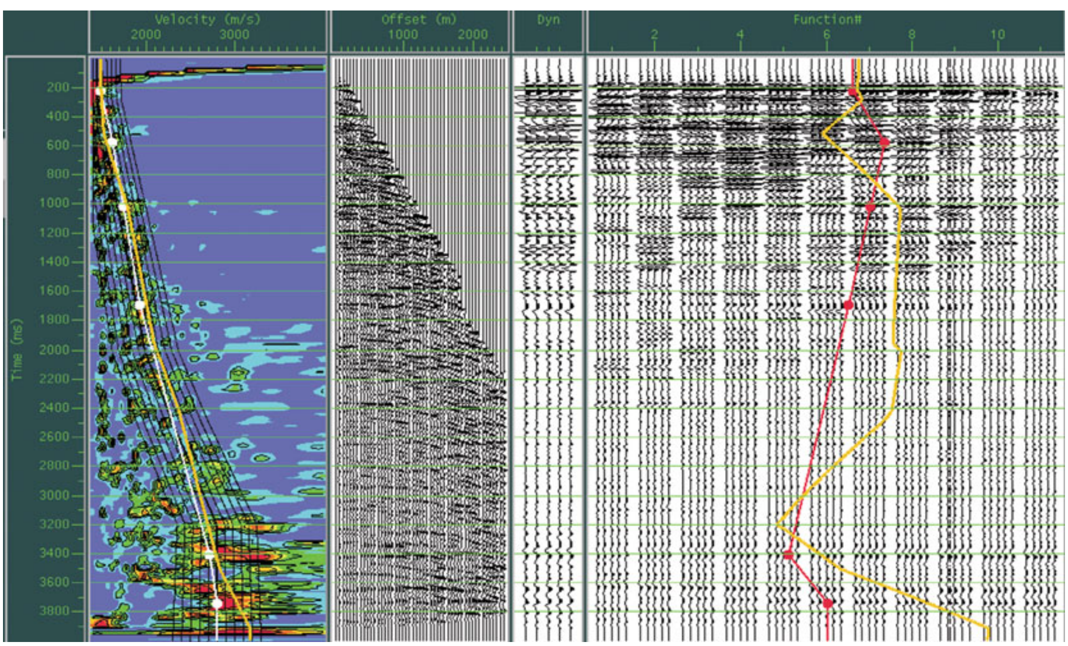

Fig. 17.23 An example of velocity analysis. A velocity spectrum or the semblance histogram (*left*). Plot of velocity analysed CMP data (*middle*) and mini-stacks based on picked velocity function (*right*). By picking maximum values from different panels a time-velocity function can be constructed

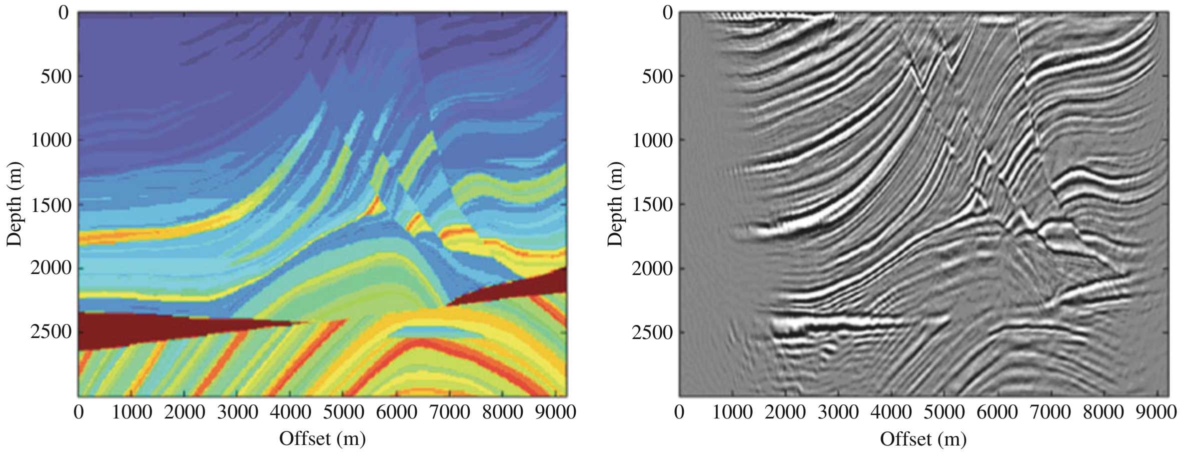

Fig. 17.24 A velocity model (*left*) and a stacked seismic section (*right*). (Source: Versteeg 1994)

depends on a good velocity analysis. In the case of an accurate velocity model (Fig. 17.24a), stacking is the most efficient multiple removal method.

Migration is the process that reverses wave propagation effects to get clear images of the subsurface. Migration processing is extremely important in geologically complex areas. The term *migration* came about because, compared to stack sections, the echoes "migrate" to their true subsurface position. Migration is used for several reasons; the most important one is to move reflectors from seismic "apparent" position to their geological "true" position. Another reason for doing migration is to collapse and focus diffractions. Since we cannot measure the direction from where the signal has been reflected we assume zero offset. It is not correct to assume zero offset where the underlying Earth layer is not plane. Before doing the migration a good velocity model is needed as an input.

The migration procedure consists of two well-defined steps: *downward extrapolation* and *imaging*. Downward extrapolation is a simulation process (based on the one-way wave-equation), where the receivers are moved from the surface down to an arbitrary depth. For each downward extrapolated depth a new seismic section can be formed (*imaging*). From each of these sections only the part close to $t = 0$ is retained. A *migrated section* is now formed by combining these strips. The exploding reflector model or ER-model (Lowenthal et al. 1977) has also been used for migration of seismic data. According to this model, a zero offset seismic section can be produced, where all the seismic reflectors "explode" simultaneously at time zero, and subsequently the data is recorded at the surface. However, the exploding reflector model does not account for reflections for which the corresponding down-going ray path differs from the up-going ray path. In order to produce correct travel times with this concept, all the velocities in the subsurface need to be halved with respect to their actual values. The advantages of seismic migration are:

- Diffractions are collapsed to points
- Deeping reflections move up dip and become steeper
- Triplications (bow ties) associated with synforms are unwrapped
- Crossing reflectors avoided.

Seismic migrations are of four types: Pre-stack time and Pre-stack depth migration and Post-stack time or Post-stack depth migration (Fig. 17.25). In *time migration* the images are displayed in two-way travel times, and wave field extrapolation is done in a time-stepping way (Fig. 17.26a). In *depth migration* the wave-stepping is done with respect to depth, and the images can be represented in a true vertical depth (Fig. 17.26b). Depth migration can handle strong lateral velocity variations. Time migration is most common in practice.

17.6 Seismic Resolution

Seismic resolution is the ability to distinguish separate features; the minimum distance between 2 features so that the two can be defined separately rather than as

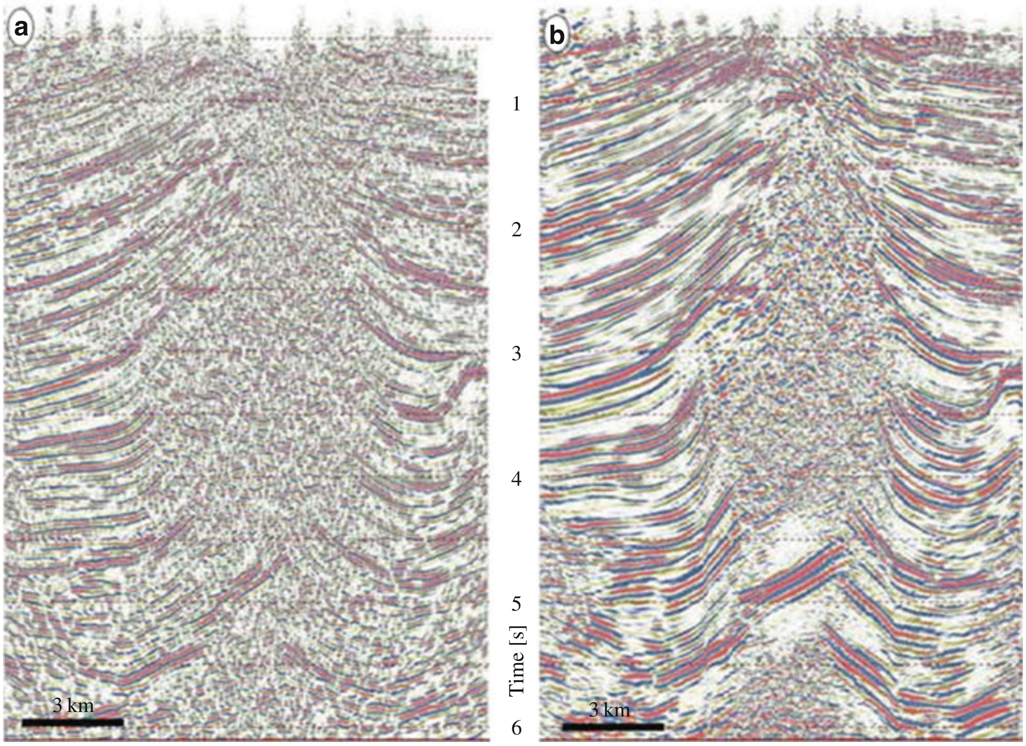

Fig. 17.25 Comparison of time domain images from (**a**) Pre-stack time migration and (**b**) Post-stack time migration

one. The limit of seismic resolution usually makes us wonder, how thin a bed can we see? Normally we think of resolution in the *vertical* sense, but there is also a limit to the *horizontal* width of an object that we can interpret from seismic data.

17.6.1 Horizontal Resolution

The horizontal dimension of seismic resolution is described by the *Fresnel zone*. The Fresnel zone is a frequency and range dependent area of a reflector from which most of the energy of a reflection is returned and arrival times differ by less than half a period from the first break (Fig. 17.27). Waves with such arrival times will interfere constructively and so be detected as a single arrival. Subsurface features smaller than the *Fresnel zone* usually cannot be detected using seismic waves. At spacing greater than one-quarter of the wavelength, the event begins to be resolvable as two separate events. *Migration* can improve lateral resolution by reducing the size of the *Fresnel zone*. For a plane reflecting interface and coincident source and receiver, the Fresnel zone will be circular and with a radius of R_{f} is

$$R_f = \sqrt{\frac{\lambda Z}{2}} \tag{17.6}$$

where λ is the dominant wavelength and Z is the depth down to the target surface. Horizontal resolution depends on the frequency and velocity of seismic waves. If we introduce the centre frequency f_{c} of the pulse (i.e. representing the most energetic part), we have $\lambda \approx V/f_{\mathrm{c}}$, with V being the wave velocity. Hence, we can rewrite the formula for the Fresnel zone as

$$R_f = \sqrt{\frac{VZ}{2f_{\mathrm{c}}}} \tag{17.7}$$

17.6.2 Vertical Resolution

Vertical resolution is the ability to separate two features that are close together. A seismic wave can be considered as a propagating energy pulse. If such a

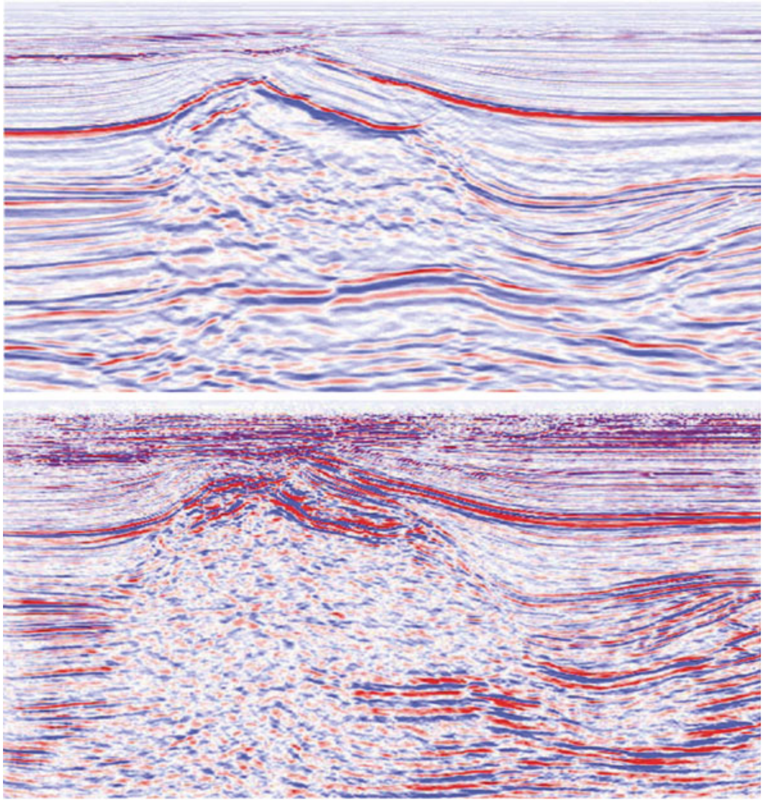

Fig. 17.26 A comparison of Pre-stack time migrated (*top*) and Post-stack depth migrated (*bottom*) images

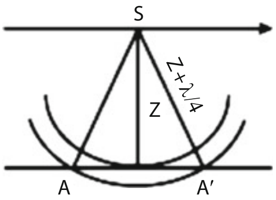

Fig. 17.27 A Fresnel zone in 3D seismic is circular and has diameter A–A´ where S is the source position, Z is the depth down to the target and λ is the wavelength. The size of the Fresnel zone helps to determine the minimum size feature that can be seen in a seismic section

wave is being reflected from the top and the bottom of a bed, the result will depend on the interaction of closely spaced pulses. In order for two nearby reflective interfaces to be distinguished well, they have to be about $\lambda/4$ in thickness which is called the *tuning thickness*. This is also the thickness where interpretation criteria change. For smaller thickness, the limit of visibility is reached and positional uncertainties are introduced.

The typical recorded seismic frequencies are in the range of 5–100 Hz. High frequency and short wavelengths provide better vertical and lateral resolution. One could argue that we could simply increase the power of our source so that high frequencies could travel farther without being attenuated. However, there is a practical limitation in generating high frequencies that can penetrate large depths. The Earth acts as a natural filter removing the higher frequencies more readily than the lower frequencies (absorption effect). This means the deeper the source of reflections, the lower the frequencies we can receive from those depths and therefore the lower resolution we appear to have from great depths (Fig. 17.28). The vertical resolution decreases with the distance travelled (hence depth) by the ray because attenuation preferentially robs the signal of the higher frequency components. *Deconvolution* can improve vertical resolution by producing a broad bandwidth with high frequencies and a relatively compressed wavelet.

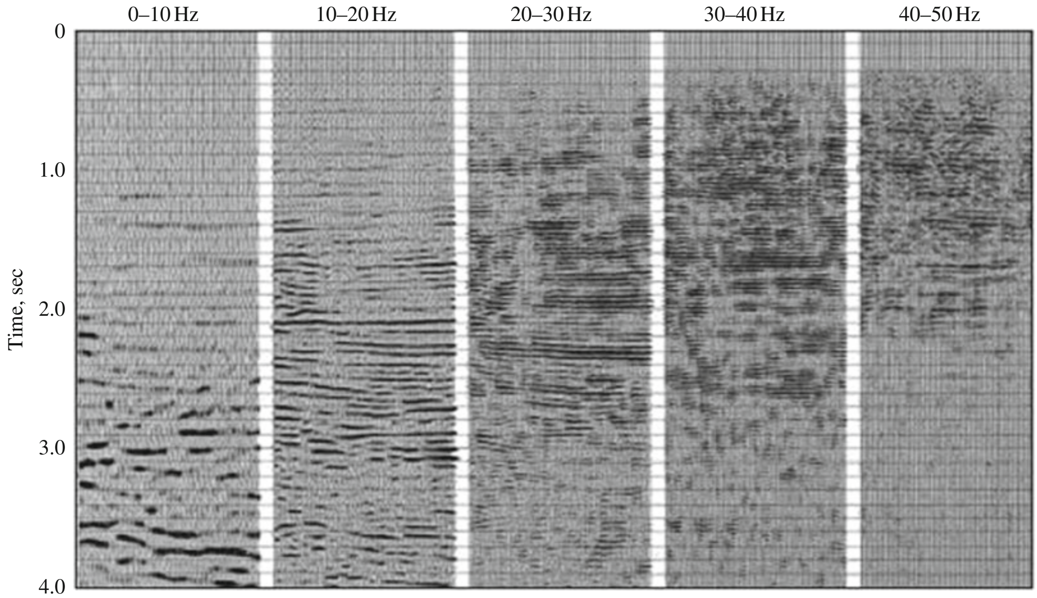

Fig. 17.28 Filtered seismic data showing frequency content variation with depth. Each panel has been filtered to allow a different band of frequencies. As the bandpass rises, the maximum depth of penetration of seismic energy decreases. Lower frequencies (*left*) penetrate deeper. Higher frequencies (*right*) do not penetrate to deeper levels. (Source: Ashton et al. 1994)

As an example, if we introduce the centre frequency f_c of the energy pulse (disturbance) we obtain the following simple relationship between the dominant wavelength (λ), the wave velocity (V) and the centre frequency (f_c):

$$\lambda \cong \frac{V}{f_c} \tag{17.8}$$

The typical values for the dominant wavelength are then (a) $\lambda = 40$ m at shallow depth (upper 300–500 m depth), where $V = 2,000$ m/s and $f = 50$ Hz, (b) $\lambda = 100$ m at intermediate depths (about 3,500 m), where $V = 3,500$ m/s and $f = 35$ Hz and (c) $\lambda = 250$ m at depths of about 5,000 m), where $V = 5,000$ m/s and $f = 20$ Hz. For smaller thicknesses than $\lambda/4$ we rely on the *amplitude* to judge the bed thickness. For thicknesses larger than $\lambda/4$ we can use the *waveform*.

17.7 Seismic Interpretation

Seismic data are studied by geoscientists to interpret the composition, fluid content, extent and geometry of rocks in the subsurface. Interpretation of seismic data will be based on an integrated use of seismic inlines, crosslines, time slices and horizon attributes (Dalley et al. 1989, Hesthammer et al. 2001). The seismic sections or images represent slices through the geological model, which can be input to advanced workstations where the actual interpretation can take place. Seismic data can be used in many ways such as regional mapping, prospect mapping, reservoir delineation, seismic modelling, direct hydrocarbon detection and the monitoring of producing reservoirs. Based on the seismic interpretation one will decide if an area is a possible prospect for hydrocarbon (oil or gas). If the answer is positive, an exploration well will be drilled. The ultimate goal will be the drilling of production wells if the target area proves to be a commercial reservoir. Seismic data contain a mixture of signal and noise. It is therefore crucial to understand the signature of the noise, whether it is systematic or random, dipping or flat-lying, planar or non-planar. It is also necessary to investigate the origin of the noise. The challenge of seismic interpretation is then to fully utilise all the information contained in the seismic data. *Systematic noise* can be related to acquisition procedures, processing artefacts, water-layer multiples, faults, complex stratigraphy and shallow gas. *Random noise* includes natural noise (e.g. wind and wave motion), incoherent seismic interface and imperfect static corrections. Without a sound understanding of these factors as well as knowledge of the limitation of seismic resolution, there is a danger of misinterpreting noise as real features.

In terms of the parameters that are analysed and the interpretation that may be drawn from the analyses, a fourfold hierarchy of seismic interpretation can be achieved. These are *seismic facies analysis*, *seismic structural analysis*, *seismic attribute analysis* and *seismic sequence analysis*. The most important parameters used for interpretation of seismic data are:

Reflection amplitudes: The strength of the reflections. As we discussed above, the proportion of the energy reflected at the boundary between two beds is a function of the difference in the acoustic impedances. If we have an alternating series of different beds, the distance between the bed boundaries in relation to the wavelength of the transmitted seismic signals will play a major part.

Reflector spacing: The distance between the reflectors will indicate the thickness of the bed, but there will be a lower limit to the thickness that can be detected, which will depend on the wavelength.

Interval velocity: The interval velocity of a sequence can provide information about lithology and porosity but this will depend on the stacking velocity and will not be very accurate.

Reflector continuity: The continuity of reflectors will be a function of how continuous the sediment beds are, information which is essential for reconstructing the environment.

Reflector configuration: If we take the compaction effect into account, the shape of the reflecting beds gives us a picture of the sedimentation surface as it was during deposition. The slope of the reflectors, for example, represents the slope of prograding beds in a delta sequence with later differential compaction and tilting superimposed. Erosion boundaries with unconformities will in the same way show the palaeo-topography during erosion.

Instantaneous phase: A seismic trace can be considered an analytical signal where the real part is the recorded seismic signal itself. Mathematically we can compute the complex seismic trace (imaginary parts of the signal) and the instantaneous attributes. The Instantaneous phase is a measure of the continuity of the events on a seismic section. The Instantaneous phase is on a scale of +180° to −180°. The temporal rate of change of the instantaneous phase is the instantaneous frequency.

17.7.1 Seismic Facies Analysis

A seismic profile provides information about the properties of sedimentary rocks. Seismic facies analysis is the description and geological interpretation of seismic reflectors representing sequence boundaries. It includes the analysis of such parameters as the configuration, continuity, amplitude, phase, frequency and interval velocity. These variables give an indication of the lithology and sedimentary environment of the facies (Selley 1998) (Fig. 17.29). Because seismic reflections mainly represent time lines, i.e. sedimentary beds which were deposited contemporaneously, it is also possible to a certain extent to interpret the depositional environment. Large-scale sedimentary features that may be recognised by the configuration of seismic reflectors include prograding deltas, carbonate shelf margins and submarine fans (Sherif 1976).

17.7.2 Structural Analysis

Interpreters can draw horizons and faults on in-lines, cross-lines and arbitrary lines, as well as slices. Horizons can be automatically tracked on vertical seismic displays and horizontal slice displays. Improved tracking algorithms for horizon interpretation, combined with user-interpreted faults and fault polygons, can produce seismic-based interpretation maps (Fig. 17.30). Faults on seismic sections are typically expressed by a loss of reflection amplitude. It must be remembered that the principle we use for calculating the depth to a reflecting boundary assumes that the layering is not too far from horizontal. High-angle faults do not reflect the sound wave back to the receivers to give a signal and faults can therefore not be seen directly on seismic sections. It is the truncations and offset of good reflectors which usually allows us to identify faults. Due to special "edge effects" that we observe near a fault, the reflector terminations which should define the fault are not

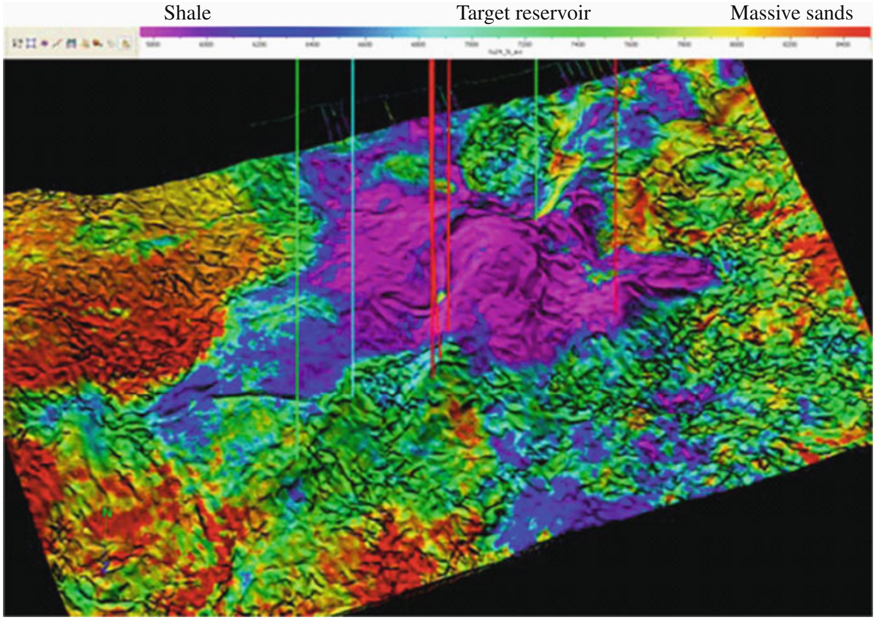

Fig. 17.29 A time slice of a special seismic attribute (S-impedance co-rendered with K_{max} curvature) that is constructed for lithology determination. Fractured, interbedded sandstones (*green*) juxtaposed against the organic-rich shales (*purple*) are the most productive reservoirs in this horizon. (Courtesy ION Geophysical)

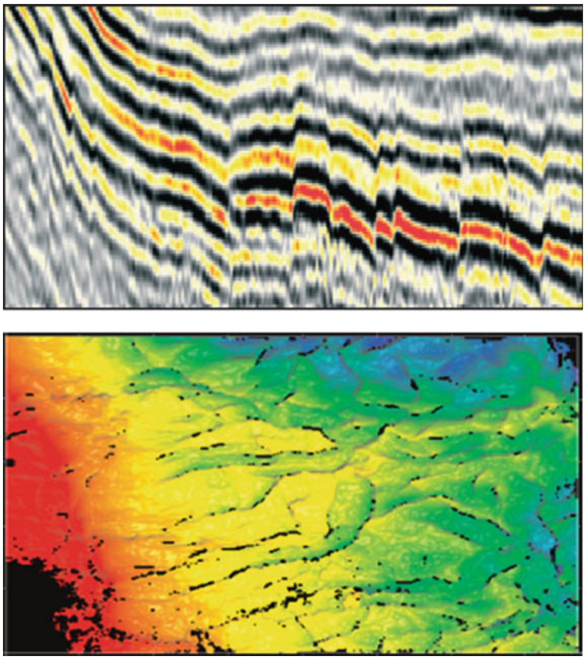

Fig. 17.30 A seismic section showing abundant faulting. Horizons can be interpreted by variable amplitude and pick trough-pick trace shape (*top*). A plan view of a horizon (*bottom*) which can be useful to identify fault patterns from the missing picks and discontinuities in colour. (Source: James 2003)

located in quite their true position on the seismic profile, making it difficult sometimes to define them accurately. Migration of the seismic data goes some way to remedying the problem. Folded beds will only be realistically depicted if the folds are sufficiently gentle that the beds have a low angle of dip.

17.7.3 Seismic Attribute Analysis

A seismic attribute is a quantitative measure of a seismic characteristic of interest. Seismic attribute analysis is concerned with the study of amplitude, polarity, continuity and wave shape. One of the goals of seismic attributes is to somehow capture maximum information by quantifying the amplitude and morphological features seen in the seismic data through a suite of deterministic calculations performed on a computer. The extraction of seismic attributes, such as amplitude envelope, dominant frequency, apparent polarity and instantaneous phase can produce remarkable 3D images of subsurface rock formations (Fig. 17.31). Such analysis may give an indication of the thickness and nature of the upper and lower contacts of a sand body (Selley 1998). Comparison

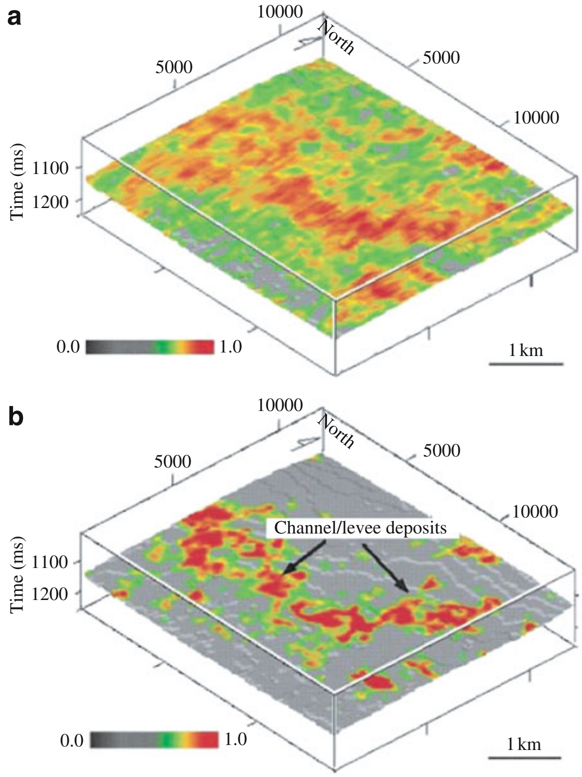

Fig. 17.31 A comparison between average absolute amplitude (**a**) and energy (**b**) in a horizon slice at the same stratigraphic level. Notice that the channel/levee deposits can be recognised, mapped and detected more efficiently from the energy volume than from the amplitude volume. (After Gao 2003)

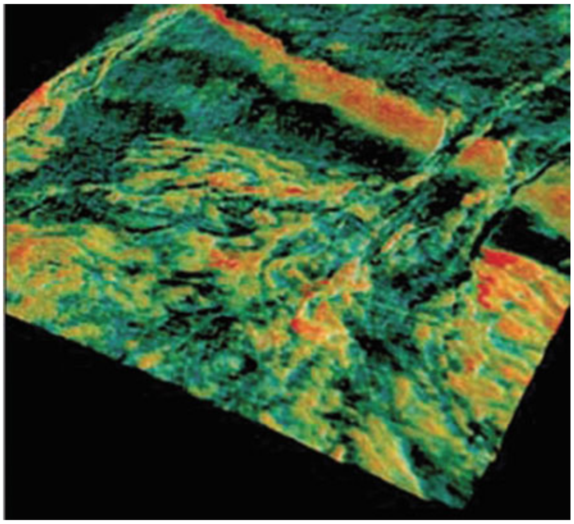

Fig. 17.32 Interpreted 3D seismic volume showing a channel fan complex with a shelf edge. (Source: James 2009)

of observed seismic waves with synthetic traces computed from a geological model give some insight into the depositional environment of the sand, and hence help to predict its geometry and internal reservoir characteristics (Fig. 17.32).

There are now more than 100 distinct seismic attributes calculated from seismic data that can be applied to the interpretation of geological structures, stratigraphy, and rock/pore fluid properties. Taner et al. (1994) divide attributes into two general categories: *geometrical* and *physical*. The objective of *geometrical attributes* is to enhance the visibility of the geometrical characteristics of seismic data; they include dip, azimuth, and continuity. *Physical attributes* have to do with the physical parameters of the subsurface and so relate to lithology. These include amplitude, phase and frequency. Liner et al. (2004) classified attributes into general and specific categories. The general attributes are measures of geometric, kinematic, dynamic or statistical features derived from seismic data. They include reflector amplitude, reflector time, reflector dip and azimuth, complex amplitude and frequency, generalised Hilbert attributes, illumination, edge detection/coherence, AVO and spectral decomposition. *General attributes* are based on either the physical or morphological character of the data tied to lithology or geology and are therefore generally applicable from basin to basin around the world. In contrast, *specific attributes* have a less well-defined basis in physics or geology. While a given specific attribute may be well correlated to a geological feature or to reservoir productivity within a given basin, these correlations do not in general carry over to a different basin.

Over the past decades, we have witnessed attribute developments tracking the breakthroughs in reflector acquisition and mapping, fault identification, bright-spot identification, frequency loss, thin-bed tuning, seismic stratigraphy and geomorphology (Fig. 17.33). More recently, interpreters have used crossplotting to identify clusters of attributes associated with either stratigraphic or hydrocarbon anomalies. Today, very powerful computer workstations capable of integrating large volumes of diverse data and calculating numerous seismic attributes are a routine tool used by seismic interpreters seeking geological and reservoir engineering information from seismic data.

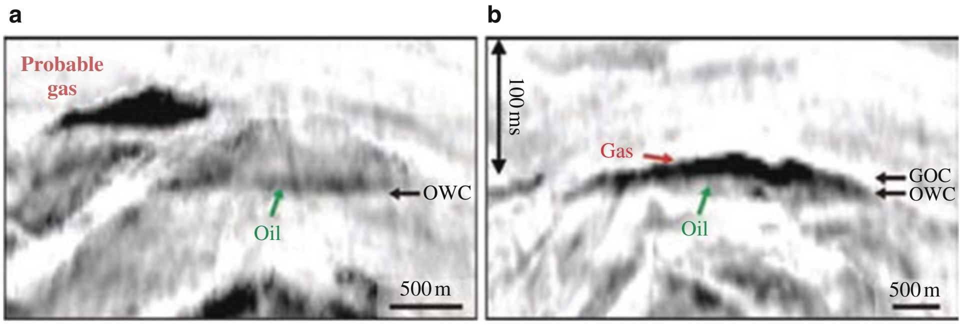

Fig. 17.33 (**a**) A strong soft amplitude anomaly (oil column) at *top* reservoir corresponds to a clear flat event within the reservoir. (**b**) This section shows an acoustically soft body (gas column) within the reservoir with a sharp, flat base. (After Blom and Bacon 2009)

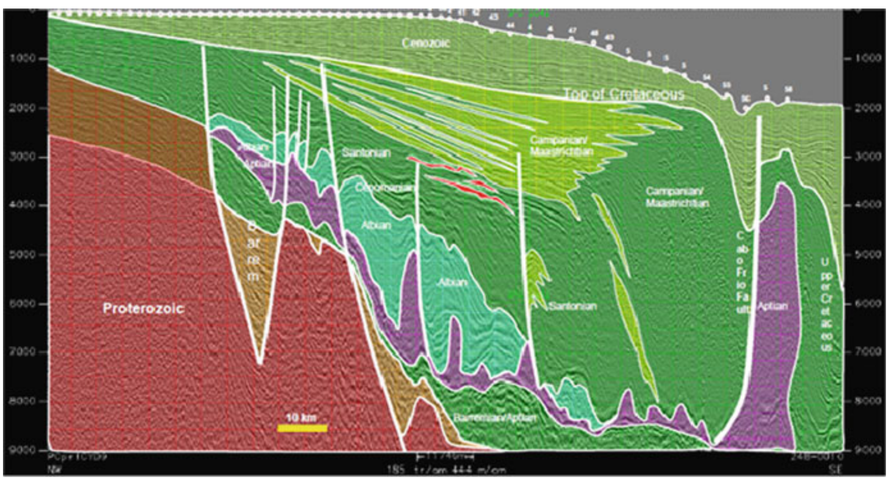

Fig. 17.34 Geological interpretation of a seismic section

17.7.4 Seismic Sequence Analysis

Apart from structural and facies analysis, seismic signatures can be used to interpret the way the basin has been filled in. This is a result of an interaction between the rate of subsidence, rate of deposition and the energy of the depositional environment (Fig. 17.34).

On seismic profiles we can see onlaps onto the land, measure the height range between the lowest and uppermost onlaps, calculate the difference in seismic time, and convert this into approximate thickness. However, we must remember that the thickness of the sediments deposited is due not only to a rise in eustatic sea level, but also to local subsidence of this part of the basin. The weight due to increased water depth will cause further subsidence, and sedimentation will increase the load, resulting in further subsidence to attain isostatic equilibrium. Local tectonic subsidence may produce a relative change in coastal onlap in a seismic profile. *Regressions* are defined as the boundary between land and sea being displaced out into the basin. They may be caused by a fall in sea level which will shift the coastline to further out on the shelf or to the edge of the continental slope, or progradation of a coastline or a delta front. Here unloading of some of the water load and erosion of sediments led to isostatic uplift of the area landward of the coastline, so that the measured regression is greater than the real lowering of the sea level. Here, too, local tectonic movements will play a part. *Transgressions* (geological events during which sea level rises relative to the land and the shoreline moves toward higher ground, resulting in flooding) and *regressions* are not always directly related to sea level changes. When a delta builds out into the sea, there is a local regression on the delta even if the sea level has not fallen. If sedimentation is sufficiently rapid, we can have a local regression on a delta even with a rising sea level. Along a coastline we can in fact have transgressions in some areas and regressions in others, at the same time, depending on the rate of sedimentation or erosion with respect to sea level change. A regression caused by a fall in sea level is called a forced regression. Changes in sea level can be due to:

- Local tectonic movements, for example uplift of a horst or subsidence of a graben structure.
- Plate-tectonic movements which can be of great extent, but are not global.
- Sea level changes. These are global and are called eustatic sea level changes.

During the Quaternary, cyclic changes in sea level of up to 120 m accompanied the growth and decay of continental ice sheets. These changes were rapid and of large magnitude and can be traced throughout much of the world. In the areas which had supported ice sheets, such as Scandinavia, the melting of the ice led to isostatic uplift due to unloading, and this exceeded the rise in sea level so that there was a regression. When seismic stratigraphy was established (Vail et al. 1977), it was assumed that most of the variations in sea level that could be interpreted from seismic records were attributable to eustatic changes and thus could be employed for global correlation. However accumulation of ice on the continents is the only known process capable of producing large and rapid global sea level changes. Such ice ages are known from the Quaternary, the Carboniferous-Permian, late Ordovician and the end of the Precambrian.

Nevertheless, transgressions and regressions can be correlated over greater or lesser distances, depending on the type of tectonic displacement. It is often difficult to distinguish between eustatic sea level changes and ones caused by more local or regional conditions and we therefore prefer to use the term *relative sea level change*. On a delta, one delta lobe may prograde and produce a regressive sequence while a transgression occurs with carbonate sedimentation on an adjacent lobe.

17.8 Integration and Visualisation of Seismic Data

In order to establish a reservoir model, different geophysical measurements such as seismic, well logs and rock physics need to be integrated. Since these measurements are carried out employing very different frequency regimes, there is a problem of scaling. Rock physics models of laboratory measurements use ultrasound frequencies (kHz to MHz) whereas sonic tools (well logging) operate with frequencies in the kHz range. Seismic data fall within the range of hertz (Hz). Assuming the velocity of a sedimentary layer is 3,000 m/s, a seismic frequency of 30 Hz will give 100 m wavelength whereas sonic log frequency of 30 kHz will give 10 cm wavelength and the ultrasound frequency of 300 kHz will give 1 cm wavelength. Therefore, the vertical resolutions of seismic, sonic and ultrasonic frequencies are 25 m, 2.5 cm and 2.5 mm, respectively (Fig. 17.35). The major problem is then how to best incorporate all geophysical data into a reservoir model. Rock physics investigations define how pore-scale variations in properties like mineralogy, fluid content and grain geometry affect the acoustic response of a core sample. As the scale of investigations increases from laboratory to well log to seismic, low frequency measurements (seismic or well logs) average over different rock types and sediments in addition to local pore-scale variations. As a result, spatial heterogeneity and preferential sampling at well log or seismic scales can cause a shift in the rock physics relationship away from that determined in the lab. Therefore, critical understanding and strategies are necessary to incorporate seismic, well logs and rock physics to find links between qualitative geological parameters and quantitative geophysical measurements. Integration strategies and utilisation of different geophysical data are discussed thoroughly in Chap. 18.

As the use of 3D seismic increasingly becomes an integral part of hydrocarbon exploration, visualisation techniques also continue to evolve as software and hardware improves. Visualisation of target datasets of various sizes and formats in a powerful graphical environment is the answer to efficiently understanding data quality problems, anomalies and trends. Such an environment responds dynamically to manipulation

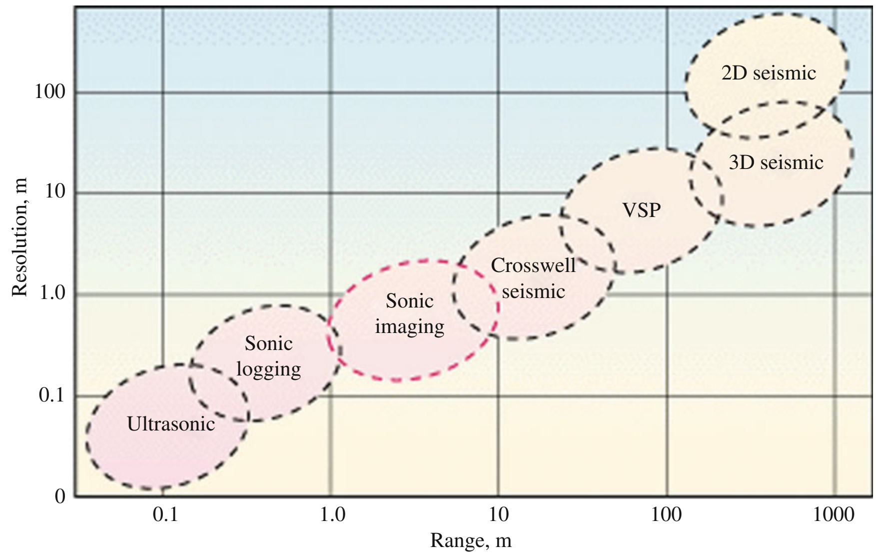

Fig. 17.35 Plot of range versus resolution of various geophysical techniques. (Courtesy: Schlumberger Oilfield Glossary)

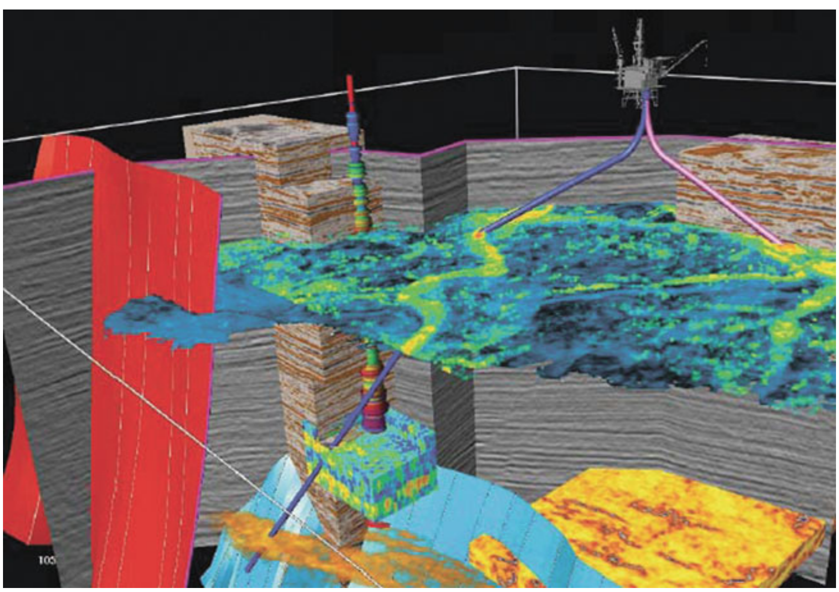

Fig. 17.36 Shows a 3D volume with fault (*red*), horizons (amplitude maps) and a producing well path through a channel. Once a channel is identified, its shape can be extracted by visualising the 3D volume. (Courtesy: Halliburton)

when analysing many datasets on one common canvas, or it can allow travel through time in a 4D visualisation process comparing one survey to another. Advance visualisation techniques allow us to interactively explore regional datasets, integrate live well information into a reservoir model and understand reservoir dynamics in association with a 4D predictive model (Fig. 17.36). 3D visualisation technique is also used to estimate thickness, continuity and lateral extend of a reservoir and its relationship with horizons and faults to see the finest reservoir details. Visualising combinations of complex seismic attributes like amplitude, continuity and AVO (Amplitude Versus Offset) provide a far more accurate picture of hydrocarbon potential, and in less time, than viewing those attributes discretely. By integration and 3D visualisation of massive amounts of data one can add greater value to decision making to differentiate between economic and sub-economic wells in a marginal area. The ability to understand data, gain useful insight, and communicate these with others is greatly enhanced by the use of interactive 3D visualisation. However, the ability to adjust the parameters and the display interactively is crucial to exploring the data and finding the combinations that highlight specific features and relationships.

Further Reading

Aronsen, H.A., Osdal, B., Dahl, T., Eiken, O., Goto, R., Khazanehdari, J., Pickering, S. and Smith, P. 2004. Time will tell: New insights from time-lapse seismic data. Oilfield Review 16(2), 6–15.

Ashton, C.P., Bacon, B., Mann, A., Moldoveanu, N., Déplanté, C., Dickilreson, T.S. and Redekop, G. 1994. 3D seismic survey design. Oilfield Review 6, 19–32.

Beckett, C., Brooks, T., Parker, G., Bjoroy, R., Pajot, D., Taylor, P., Deitz, D., Flaten, T., Jaarvik, L.J., Jack, I., Nunn, K., Strudley, A. and Walker, R. 1995. Reducing 3D seismic turnaround. Oilfield Review 7(1), 23–37.

Blob, F. and Bacon, M. 2009. Application of direct hydrocarbon indicators for exploration in a Permian-Triassic play, offshore the Netherlands. First Break 27, 37–44.

Braile, L.W. 2000. Seismic Waves and the Slinky: A Guide for Teachers. http://www.gg.uwyo.edu/geol2005a/lectures/seismics/WaveDemo.pdf

Dalley, R.M., Gevers, E.C.A., Stampfli, G.M., Davis, D.J., Gastaldi, C.N., Ruijtenberg, P.A. and Vermeet, G.J.O. 1989. Dip and azimuth displays for 3D seismic interpretation. First Break 7, 86–95.

Gao, D. 2003. Volume texture extraction for 3D seismic visualization and interpretation. Geophysics 68, 1294–1302.

Granli, J.R., Arnsten, B., Anders, S. and Hilda, E. 1999. Imaging through gas-filled sediments using marine shear-wave data. Geophysics 64, 668–667.

Hesthammer, J., Landrø, M. and Fossen, F. 2001. Use and abuse of seismic data in reservoir characterisation. Marine and Petroleum Geology 18, 635–655.

James, H. 2003. Has volume interpretation of structure been cracked at last? First Break 21, 38–40.

James, H. 2009. Visualizing 3D features in 3D seismic data. First Break 27, 57–62.

Liner, C., Li, C.-F., Gersztenkorn, A. and Smythe, J. 2004. SPICE: A new general seismic attribute: 72nd Annual International Meeting, SEG, Expanded Abstracts, 433–436.

Loewenthal, D., Lu, L., Roberson, R. and Sherwood, J. 1976. The wave equation applied to migration. Geophysical Prospecting 24, 380–399.

Milligan, M. 2004. What are seismic surveys and how much shaking do they create? Survey Notes, Utah Geological Survey 36(3), 10–11.

Ongkiehong, L. and Askin, H.J. 1988. Towards the universal seismic acquisition technique. First Break 6(2), 46–63.

Rygg, E., Riste, P., Nottvedt, A., Rod, K. and Kristoffersen, Y. 1992. The Snowstreamer - a new device for acquisition of seismic data on land. In: Vorren, T.O., Bergsager, E., Dahl-Stamnes, O.A., Holter, E., Johanses, B., Lie, E. and Lund, T. B. (eds.), Arctic Geology and Petroleum Potential. NPF Special Publication 2, 703–709.

Selley, R.C. 1998. Elements of Petroleum Geology. Academic Press, San Diego, CA, 470 pp.

Sheriff, R.E. 1976. Inferring stratigraphy from seismic data. AAPG Bulletin 60, 528–542.

Sheriff, R.E. and Geldart, L.P. 1995. Exploration Seismology. Cambridge University Press, New York, 592 pp.

Taner, M.T., Schuelke, J.S., O'Doherty, R. and Baysal, E. 1994. Seismic attributes revisited: 64th Annual International Meeting, SEG, Expanded Abstracts, 1104–1106.

Telford, W.M., Geldart, L.P. and Sheriff, R.E. 1990. Applied Geophysics. Cambridge University Press, New York, 770 pp.

Thompson, M., Arntsen, B. and Amundsen, L. 2007. Full azimuth imaging through consistent application of ocean bottom seismic. 77th Annual International Meeting, SEG, Expanded Abstracts, 936–940.

Tucker, P.M. and Yorston, H.J. 1973. Pitfalls in seismic interpretation. Society of Exploration Geophysicists, Monograph Series, No. 2, 50 pp.

Vail, P.R., Mitchum, R.M. and Thomson, S. 1977. Seismic stratigraphy and global changes of sea level, Part 3: Relative changes of sea level from coastal onlap. In: Payton, C.E. (ed.), Seismic Stratigraphy – Applications to Hydrocarbon Exploration. AAPG Memor 26, 63–97.

Versteeg, R. 1994. The Marmousi experience: Velocity model determination on a synthetic complex dataset. The Leading Edge 13, 927–936.

Yilmaz, O. 2001. Seismic Data Analysis: Processing, Inversion and Interpretation of Seismic Data. SEG Press, Tulsa, 2027 pp.