SEISMIC DATA PROCESSING
(Velocity analysis & Migration)

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And above all, I am thankful to the Almighty for his grace.

Declaration

This is to certify that the dissertation report on “**Seismic Data Processing(Velocity analysis & Migration)**” submitted to the University of Petroleum & Energy studies, Delhi by **Krishn Kumar**, in partial fulfillment of the requirement for the award of the degree of **Masters of Technology (Petro Informatics)**, is a bonafide work carried out by him under my supervision and guidance.

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Abstract

By seismic data processing we mean any computational technique that attempts to remove noise or wave propagation effects in order to create an image of the subsurface. Once this image is created there are different methods to mine data for further information. Example include attribute analysis, velocity analysis, migration etc. Seismic reflection is the preferred technique for subsurface stratigraphic and structural mapping and it is routinely used for hydrocarbon exploration and characterizing the structure of the earth's upper crust because of the time and skill typically required for processing seismic reflection data.

It will solve problem related to velocity analysis and migration using software PROMAX. Velocity analysis enhances residual normal moveout correction by properly taking velocity in account. Seismic migration is collapsing diffraction events on un-migrated records to points, thereby moving or migrating reflection events to their proper locations creating true image of structures within the earth.

Table of Content

Chapter 1: Introduction

The search for hydrocarbons is called prospecting or exploration of oil and gas or hydrocarbons. Since there is not any way to be absolutely sure where new oil and natural gas reserves are, we have to collect clues that indicate what lies deep beneath the earth's surface. We look at two types of clues

- Surface
- Sub-surface

1) **Collecting Surface Clues---** One way to gather clues is to look at surface data like rock properties, surface structure and oil and natural gas seeps.

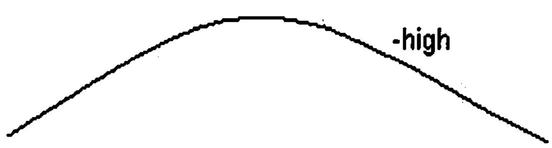
2) **Collecting Sub-Surface Clues----** We get much more useful information by looking at geological structures and rock properties below the surface. Advanced exploration technology even helps us locate new reserves at existing sites that were thought to be depleted. We typically use three ways to look below the surface:

- Gravity Surveys.
- Geomagnetic surveys.
- Seismic surveys

1.1Gravity surveys

Gravity methods depend upon the relative density of the ore deposit and surrounding wall rock. Measurements can only be made at fixed stations on the ground, and complicated corrections are required for station position and topographic conditions. The typical ore deposit is not dense enough, is too small and irregular, and occurs in a deformed structural environment, making clearly defined gravity anomalies difficult to discern and interpret. The method has been very successful in exploring for large deposits of petroleum, natural gas, sulfur, and salt. Limited application has been reported in exploration for barite. Gravity surveys can provide useful information where other methods do not work. For example, gravity may be used to map bedrock topography under a landfill, where seismic refraction is limited. Gravity can also be used to map lateral lithologic changes, and faults.

Constraints: Gravity surveys are relatively slow and expensive. Detect ability varies with target size, depth and density contrast. Interpretation of data often requires control data from drilling, outcrops, or other sources. Detailed surface topographic survey data is also required.



Method: Gravity is the attraction between masses. The strength of this force is a result of the mass and distance separating the objects. A gravimeter is used to measure the earth's gravitational attraction at various points over the area of interest. Gravity anomalies are due to differences in density of underlying materials. Gravity anomalies are extremely small relative to the total field and are usually measured in micro-Gals (one micro-Gal is about 1 billionth of the earth's total gravitational field). The equipment used in a gravity survey is extremely delicate and precise. Data interpretation is time consuming even with the use of sophisticated computer programs.

1.2Geomagnetic surveys

Certain minerals distort the earth's field, and where sufficiently large concentrations of such minerals occur, variations can be measured by magnetometers mounted in aircraft, in ground vehicles, or positioned at stations on the ground. Magnetite iron ores have been found in many areas of the world using the airborne magnetometer. Magnetic nickel ore, and asbestos-bearing serpentine associated with certain magnetic intrusive rocks have been found, using the magnetometer. Some geophysicists propose the use of the magnetometer to detect gold placer deposits, because of their common association with black sands largely consisting of the mineral magnetite. Magnetometer surveys are rapid and efficient. Magnetometers can be used to detect buried ferrous metal objects (tanks or drums) or bedrock features with contrasting magnetite content. Detection depends on the amount of magnetic material present and its distance from the sensor. A single steel drum can be detected at burial depths up to 15 or 20 feet. Burial depth can be estimated from magnetometer data collected using the gradient method.

Constraints: Utilities, power lines, buildings, and metallic debris can cause interference. Solar magnetic storms may cause fluctuations in readings. The size and depth of objects affect detect-ability.

Method: When a ferrous material is placed within a magnetic field such as the earth's, it develops an induced magnetic field. The induced field is superimposed on the earth's field at that location creating a magnetic anomaly. A magnetometer survey for hydrogeologic and engineering applications is conducted on foot, by one operator. The survey can be along single lines or along a series of parallel traverses with readings taken every 5 to 50 feet. Spacing of traverses and readings depends on the width of the expected anomaly. For instance, tank searches may be conducted at a 5-foot spacing while geologic mapping may be conducted at a 50-foot spacing. In the gradient method, the total field is measured simultaneously at two elevations by using two sensors on a staff separated by a fixed distance. The difference in magnetic intensity between the two sensors divided by the distance between them is the vertical gradient. This technique reduces interference from solar magnetic storms and regional magnetic changes. This technique is particularly useful for locating small, shallow objects and is also useful for estimating burial depth of objects.

1.3 Seismic surveys

The method depends upon the velocities of acoustical energy in earth materials, and has been enormously successful in searching for petroleum, natural gas, and sulfur, where the large deposits may be located by simply determining attitude of the enclosing strata. In engineering and hydrogeology, seismic refraction has many applications. Often, bedrock structure and topography control contaminant migration. Seismic refraction is a valuable tool for mapping bedrock troughs and fractures. It is usually more cost-effective and gives better coverage than drilling alone.

Constraints: Layer velocity (density) must increase with depth. Layers must be of sufficient thickness to be detectable. Data collected directly over loose fill (landfills) or in the presence of excessive cultural noise will result in sub-standard results. Single narrow fractures are too small to be detected.

Method: The seismic refraction method utilizes sound waves. Sound travels at different velocities though different materials and is refracted at layer interfaces.

A seismic wave is usually generated by a small explosive charge, a shotgun shell, or a sledge hammer. The wave's travel time from the sound source to refracting layers, along those layers and back to detectors (called geophones) is precisely measured. From the time-distance relationships, subsurface layer velocities and thicknesses can be calculated.

Fracture zones can often be detected because they usually have a lower seismic velocity than solid bedrock. The velocity of sound through water saturated material is about 5000 feet per second while the velocity though crystalline bedrock generally ranges from 12,000 to 18,000 feet per second. By calculating the velocity of sound along the bedrock surface, low velocity zones, which may represent fractures, can be delineated.

Some systems utilizes up to 24 geophones at one time. The geophones can be linearly spaced any distance apart, but most often are spaced 10 to 50 feet apart. In general, the greater the expected bedrock depth, the greater the geophone spacing. Shorter spacing and sometimes radial patterns are used in fracture zone detection studies.

Both 12- and 24 channel seismographs are used in the field. The 24-channel instrument has an internal computer. Seismic field data are stored on 3.5 inch disks for later computer analysis. Final results are provided to the client in a full report which includes tables of subsurface depths and elevations, and cross-section profiles. Results are also available as bedrock contour maps and on computer disks.

Seismic Reflection

- graphically depicts subsurface stratigraphy and bedrock profiles
- can often differentiate unconsolidated units such as sand, clay, or gravel
- marine surveys

Seismic reflection is useful for graphically profiling subsurface stratigraphy. It is used to map clay and sand lenses and bedrock troughs. Applications of this technique include determination of depth to bedrock, aquifer location studies, and mapping of overburden stratigraphy.

Constraints: Reflection surveys are highly site specific. A shallow ground water table is required. Reflection surveys are useful for exploration depths of 50 feet to several hundred feet.

Method: Seismic reflection is a geophysical technique in which acoustic waves, reflected directly from underground surfaces with density contrasts, are used to map soil and bedrock stratigraphy. Successful application depends upon the ability of the ground to transmit high frequency seismic energy (saturated clays, for example, transmit high frequency energy quite well). The method overcomes some of the potential problems encountered in refraction surveys (such as the assumption that subsurface layer velocities increase with depth and that layers are thick enough to be detectable). The equipment used is nearly identical to that used in a seismic refraction survey. The field technique, however, differs and ground coverage is usually slower than with a refraction survey.

1.4 Seismic Data Processing Steps

Seismic data Processing is a mix of digital filtering theory and practical application of digital techniques to assemble and enhance image of subsurface geology.

Main tasks of seismic data processing are to:

- Correct for recording strategies and ray-path geometries
- Take advantage of the information that recording techniques provide
- Enhancement of S/N ratio.
- Provide the clearest possible image of the surface

Seismic-data processing is composed of basically five types of corrections and adjustments:

- Time
- Amplitude
- Frequency-phase content
- Data compression (Stacking)
- Data positioning (migration)

Time adjustment fall into two categories: Static and dynamic. Static time correction shift a whole trace. The correction is constant over time. Dynamic time correction (normal moveout) are a function of both time and offset and convert the times of the reflection into coincidence with those that would have been recorded at zero offset.

Amplitude adjustments correct the amplitude decay with time due to spherical divergence and energy dissipation in the earth. There are two broad types of amplitude gain programs: structural amplitude gaining or automatic gain control (AGC) ,and relative true amplitude gain correction. The first scales amplitude to be nearly alike and is generally chosen for structural mapping purposes. Relative amplitude information so that amplitude anomalies associated with fancies changes, porosity variations, and gaseous hydrocarbons are preserved.

The frequency –phase content of data is manipulated to enhance signal and attenuate noise. Deconvolution is the inverse filtering technique used to compress an oscillatory source waveform, often seen on marine data , into as near a spike (unit–impulsefunction) as possible.

The data-compression technique generally used is the common midpoint (CMP) stack .It sums all offset of a CMP gather into one trace. Forty–eight to 96 fold stack are common . Convention 2-D seismic data initially exits in a 3-D space: the three axes are time, offset, and a coordinate x along the line of survey. Stacking compresses the offset axis onto the zero offset, which lies on the midpoint axis.

The data-positioning adjustment is migration. Migration moves energy from its CMP position to its proper spatial location. In the presence of dip, the CMP location is not the true subsurface location of the reflection. Migration collapses diffractions to foci. Migration techniques have been developed for application pre-stack, post-stack or a combination of both.

Chapter 2:Seismic Reflection Methods

2.1 What is a seismic section

The seismic reflection method works by bouncing sound waves off boundaries between different types of rock . The reflections recorded are plotted as dark lines on a seismic section. A seismic section resembles a geological cross-section, but it still needs to be interpreted. One major difference between a geological cross-section and a seismic section is that the vertical axis is in time, rather than depth. In the earth's crust, seismic waves travel typically at about 6000 m/s so that 1 second of two-way travel time corresponds to about 3 km of depth. Another difference is that the reflections are plotted halfway between the source and the receiver. These are referred to as un-migrated data. The process that moves the reflections in their correct spatial position is referred to as migration, and the resulting seismic section is referred to as a migrated section. Interpreters like to use both, and both un-migrated and migrated data are presented in this project.

2.2 Why the seismic reflection method

The science of LITHOPROBE is spearheaded by the seismic reflection method because it is the geophysical technique which produces the best images of the subsurface. These data resolve map able features such as faults, folds and lithologic boundaries measured in the 10's of meters, and image them laterally for 100's of kilometers and to depths of 50 km or more. Seismic reflection profiling is the principal method by which the petroleum industry explores for hydrocarbon-trapping structures in sedimentary basins.

2.3 Seismic data acquisition

The method works by bouncing sound waves off boundaries between different types of rock . As opposed to earthquake seismology, where the location and time of the source is an unknown that needs to be solved for, seismic reflection profiling uses a controlled source to generate seismic waves. On land, LITHOPROBE has been using large truck-mounted vibrators as a source (the "Vibroseis" method), and occasionally dynamite is used. At sea, large arrays of airguns, which rapidly eject compressed air, are deployed. The reflected signals are recorded by geophones, or hydrophones at sea, which resemble ordinary microphones.

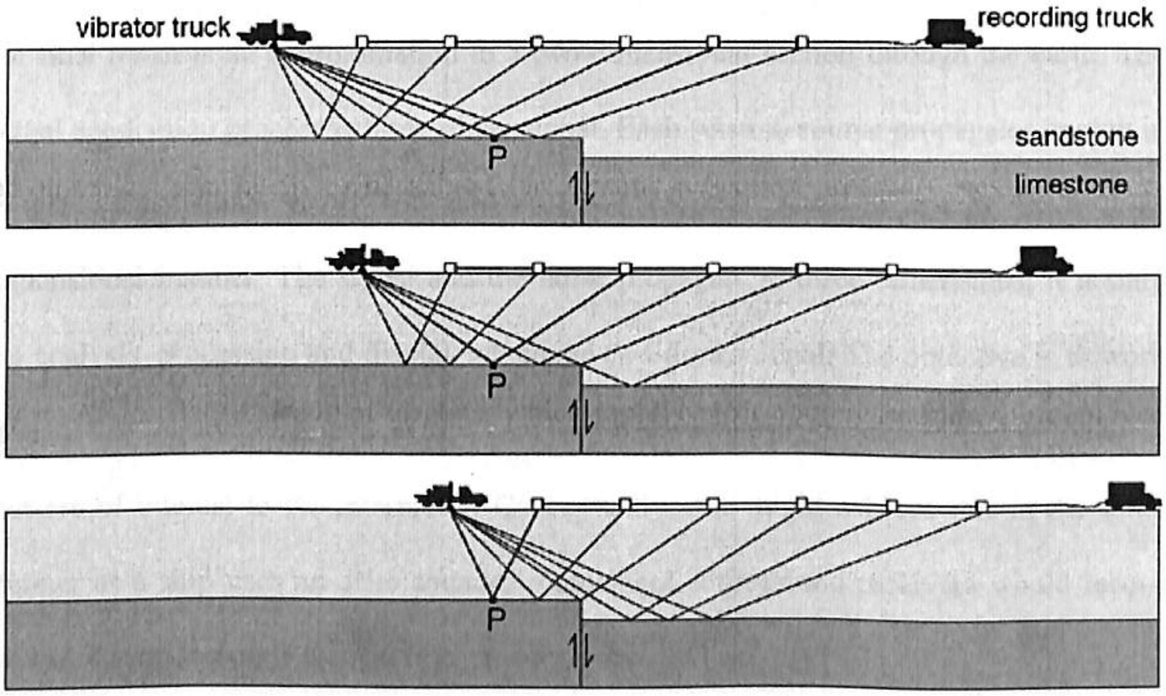


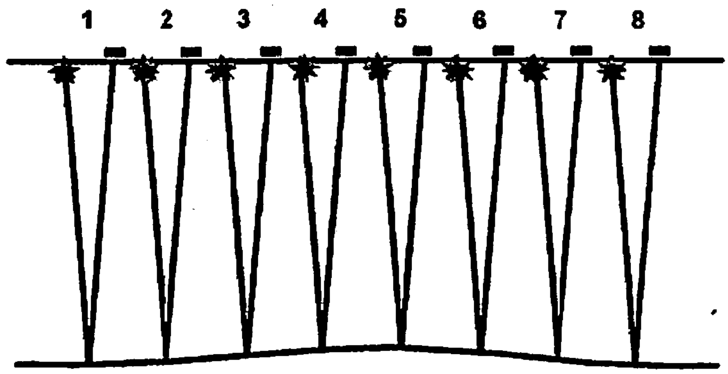
Figure2.1 Seismic data acquisition.

During a seismic survey, a cable with receivers attached to it at regular intervals is laid out along a road or towed behind a ship. The source moves along the seismic line and generates seismic waves at regular intervals such that points in the subsurface, such as point P in Figure, are sampled more than once by rays impinging on that point at different angles. As a shot goes off, signals are recorded from each geophone along the cable for a certain amount of time, producing a series of seismic traces. The seismic traces for each shot (called a shot gather) are saved on magnetic tape in the recording truck.

2.4 Different seismic acquisition methods.

2-D Seismic Method

Most seismic work (and particularly reconnaissance work) is conducted along lines. Such work is often called 2-D (two-dimensional, since the final output approximates to a two-dimensional section through the earth). However, it is important to remember that only the final result is an approximation to a two-dimensional section through the earth. The actual acquisition process is three-dimensional. Each seismic source propagates energy in a three dimensional earth and each seismic receiver receives energy in a three dimensional manner. The signal and the noise propagate in three dimensions; it is only the analysis, processing and display which are two-dimensional. The objective is to work along a line to produce a vertical section which best represents the parameters of the earth that are of interest to the interpreter. Conceptually, one might achieve this in the same manner as a ship uses an echo sounder coincident sources and receivers would record data at regular intervals along a line, usually a straight line.



(Figure 2.2 : *Basic single channel recording of reflection times along a 2-D line*);

3-D Seismic Method

The previous section stated that the 2-D method is an attempt to produce a vertical cross-section within a three dimensional world. Increasingly, data is collected over a grid of sources and receivers. These two dimensions at the surface, together with the estimations of depth in the subsurface lead to this technique to be termed 'three-dimensional seismic'. This technique frees the geophysicist from the assumption that the subsurface is flat in a direction perpendicular to the seismic line. It has many other advantages including improved resolution and better signal to noise potential. This presentation will introduce most concepts using a two-dimensional approach, since this leads to an easier understanding; however, the concepts can readily be extended to three dimensions. The presentation will introduce the concepts of the three-dimensional seismic method, but more detailed explanations will be found in other presentation.

2-D or 3-D?

A properly designed 3-D survey will always provide a better image than a 2-D survey. Both ambient and coherent noise is better attenuated and out of plane reflections are properly positioned. Factors which sway the decision towards 3-D are :

- structural complexity,
- extensive faulting,
- need for improved spatial resolution, and
- high cost of exploration failure.

Frequently a 2-D survey does not solve the exploration problem and will be repeated with closer line spacing. The repeated surveys will then be followed by a 3-D survey. These repeats are expensive and involve lengthy delays. In most circumstances, especially marine, the economics favor an early acquisition of 3-D data .

Multi-Component Seismic

Conventional seismic generates both Pressure and Shear waves, but only the vertical component of these waves is measured on land, while only the pressure component is measured at sea. Multi-component seismic (MCS) uses detectors orientated in three directions at each receiver point. MCS might better be called vector seismic, as it attempts to measure the magnitude and direction of particle motion in three dimensions. This allows the simultaneous recording of both Pressure and Shear waves. The main advantages of MCS are the improved ability to predict fluid type, lithology and fractures, permeability, and also sometimes to provide a better structural image.

Time Lapse (4D) Seismic

The production of hydrocarbons from a reservoir may in some cases affect the seismic image. A simple example of this is the upward movement of the oil-water contact as the oil is produced. Recording seismic data before and during production may provide valuable information by showing differences between the images; for example, it may identify un-swept pockets of oil. Before recording a time lapse survey, it is important to model the reservoir to determine whether the production will create a measurable effect on the seismic image. It is also necessary to examine the cost benefits of such a survey. The effects of production are likely to be very subtle. Ideally, the acquisition parameters should be identical for all surveys in the time lapse experiment, but changes in equipment and techniques make this unlikely. For this reason, it is important to consider the potential need for a later time lapse survey during the design of the very first survey over a field.

2.5 Multiple Coverage Assumptions

It is important to remember the assumptions made in the multiple coverage method. The common mid-point and common-depth-point are only the same when there is no lateral variation in velocity and when there is no dip. The NMO correction is only correct when the above conditions are met, and also when there is no depth variation of the velocity. For the most effective cancellation of coherent noise, this noise must be identically generated by each source. Seismic processors have ways to mitigate the first two conditions, but the last is often overlooked.

Chapter 3: Seismic Imaging

This chapter will review the fundamental ideas and concepts of seismic imaging. Theoretical topics like waves, convolution, correlation, and Fourier transforms are reviewed briefly. This chapter discusses also how these concepts can be user to obtain a seismic image (conventional earth model) and how seismic data is acquired. Seismic acquisition is explained by producing a synthetic data set from the introduced layered earth model.

3.1 Waves

Seismic imaging is accomplished by creating seismic energy and analyzing the returning signal that traveled through the earth. Seismic energy propagates through the earth in the form of waves. A wave is "a disturbance that is propagated through the body or on a surface of a medium without involving net movement of material. Waves are usually characterized by periodicity".

Waves that are used for seismic imaging fall in the class of body waves, namely P-waves and S-waves. The particle motion of a P-wave, or primary wave is parallel to its direction of propagation, whereas it is orthogonal for an S-waves, or secondary wave. P-and S-waves are also called compressional and shear waves, respectively.

In seismic reflection imaging, only P-waves are normally encountered, S-waves play a minor roll however, they gained popularity during the past years and they are now occasionally used in 3-component (vertical, in-line, and transverse) seismic. This project will focus on conventional reflection seismic based on P-waves, but it should be pointed out that there is no such thing as a pure P-wave: while traveling through the earth, P-waves are converted into S-waves and vise versa.

Important differences between P- and S- waves are their travel velocities and their ability to travel through fluids:

- P-waves are faster than S-waves. When the seismic science started out with the examination of earthquakes it was observed that the compressional waves arrive always earlier than the shear waves and therefore they were called primary waves.
- P-waves can travel through fluids, whereas S-waves can-not.

The second type of waves are surface waves. These are waves that travel at boundaries between two media with different physical properties even below the surface. Example of surface waves are Love and Rayleigh waves. In seismic imaging they are undesired, due to their high-energy content unavoidable, hence they contribute as "noise". Mathematical description of waves can be complex, however for the context of this project it is sufficient to use the following simple concepts:

- Wavefronts

A wavefront is a surface of equal phase and traveltime. In an isotropic medium it is also a surface of equal distance from the original disturbance, in which case the surface would be sphere (example: stone drops in water). At a large distance from the wave origin, the surface can be locally approximated by a plane. Wavefronts for heterogeneous medium are shown in figure 3.1

- Huygen's Principle

This principle states that every point of an advancing wavefront can be considered as a source of secondary wave. The envelope of all the wavefronts of the secondary waves is then again the primary wavefront at a later time. Thus, the primary wavefront is tangent to the secondary wavefront(figure 3.1)

- Rays

A line, which is everywhere perpendicular to a wavefront, is known as a ray or raypath. If a wavefront is distorted by a non-uniform velocity distribution, the ray is bent so that it is always perpendicular to the instantaneous wavefront.

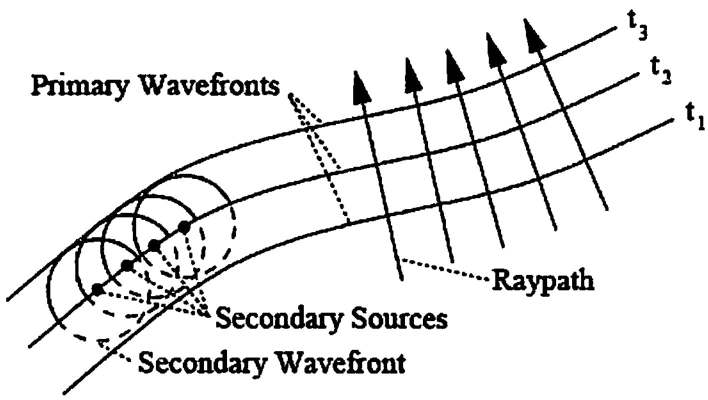


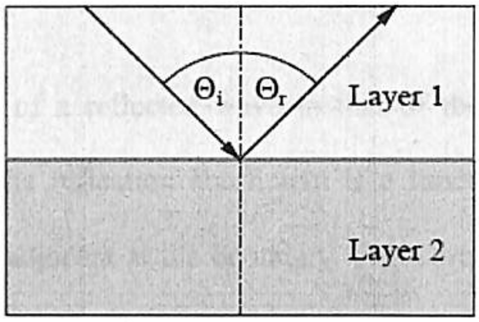
Figure 3.1: *Wavefronts and rays. A primary wavefront is shown at three different times. Raypaths are perpendicular to the wavefronts. Secondary sources and wavefronts can be used to construct a primary wavefront at a later time (Huygen's principle).*

Rays provide a very powerful concept to follow the travelpath of waves, much simpler than the complex construction of wavefronts using Huygen's principle.

- Law of Reflection

The behavior of a reflected ray at an interface of two mediums with different characteristics can be described by the following equation:

$$\theta i = \theta r \text{ Law of Reflection}$$



Law of reflection 3.2 An illustration showing a reflection across a planer boundary.

Snell's Law

The behavior of a refracted ray at an interface of two mediums with different characteristic can be described by the following equation:

$$\frac{V_1}{V_2} = \frac{\sin\Theta_i}{\sin\Theta_r}$$

Where V1 is the wave speed in media 1, V2 the wave speed in media 2, θi the angle of incidence and θr the angle of the refracted wave (Figure 3.3).

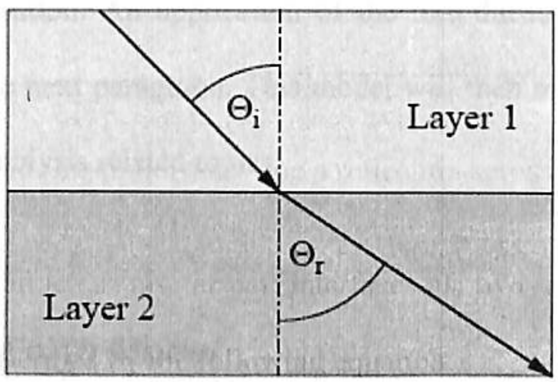


Figure 3.3 : Shell's law. An illustration showing refraction across a planar boundary.

The above stated equation can be directly derived from Huygen's principle! The angel θi, for which θr, becomes 90 degrees is called critical angle: the ray which belongs to this angle is called critically refracted ray.

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

- Reflectivity

The ratio of the amplitude of a reflected wave to that of the incident wave is called reflection coefficient R. The reflection coefficient is a function of the velocities and densities of the two media, adjacent at the boundary. For a wave traveling in medium 1 and reflected from the boundary to medium 2, is given by:

where ρ is the density and V the velocity of the indexed media. A positive reflection coefficient R implies no phase reversal of the wave, a negative one does. The reflection coefficient R in this form, however, is only valid for a normally incident ray (θ_1=0). In the more general case, an incident P- wave will have a reflected S-wave. The amplitudes of the reflected P-and S-waves, can be obtained from Knott or Zoeppritz's equations.

- **Wavelets**

  A wavelet is a seismic pulse, usually consisting of only a few cycles. A wavelet has a finite duration. An application of the introduced concepts is the layered earth model, the next paragraph. This model will then motivate the discussion of further signal analysis related topics.

3.2 The Layered Earth Model

A layered earth model is a simple model of the. Each layer in the model represents a layer of rock or sediments with similar characteristics. A unique velocity can be assigned to each layer. Hence, Snell's law of reflection are sufficient to determine raypath geometry within this model.

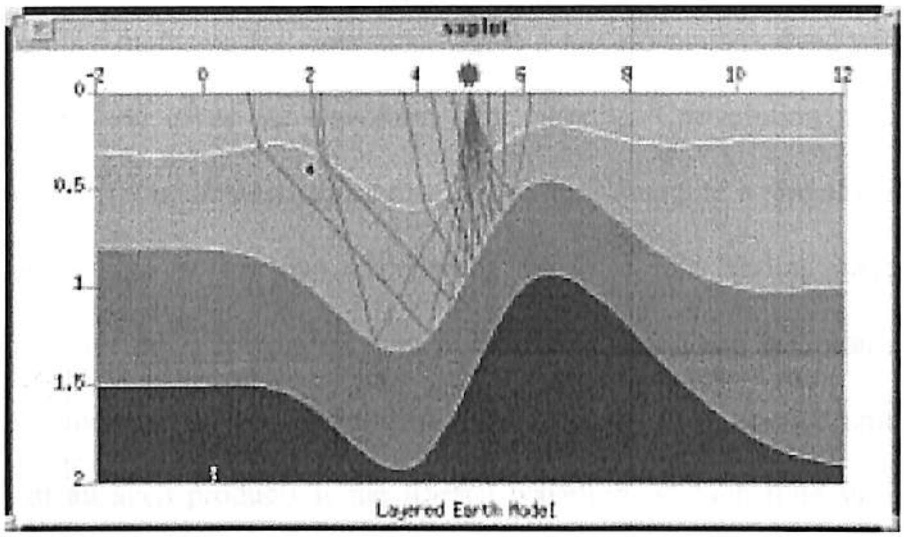


Figure 3.4: Layered earth model. Each layer has a constant velocity assigned Raypath geometry can be constructed using Snell's law and the law of reflection.

The rays shown in this figure are examples for refracted and reflected rays that return to the surface after being reflected at the second interface of Figure 3.4. The idea of seismic reflection imaging is to use these reflected rays to form a picture of the sub-surface. To see how this task can be performed, it will be necessary to discuss a few more theoretical concepts: convolution, correlation and Fourier transforms.

The model shown in Figure 3.4, with its three continuous reflectors and its small isolated reflecting body between the first and second reflector (at x=2, t=0.5), was generated by Seismic Unix with the script model . The model is a "sloth" (slowness2) model, where velocities in different layers are entered by:

$$sloth = \frac{1}{V^2} \quad , \text{V in [km/s]}$$

Layers are specified by their edges, and the "sloth" within each layer is calculated by triangulation from the edges. In the model presented here, the velocity (or " sloth") for each layer is constant.

3.3 Convolution and Correlation

Convolution is a mathematical operation, where a waveform convolved with the impulse response of a filter gives the waveform after filtering. Convolution is also known as linear filtering. To understand this process, one may imagine a waveform, reversed in time, sliding along the impulse response of the filter. The sliding increment can be infinitesimal: however, for digital data it must be at least the sampling interval. The waveform is multiplied with the impulse response of the filter at each time step and the total sum of all such products is the filtered waveform at each time step (principle of linear superposition).

An example of convolution is apparent from the above layered earth model (Figure 3.4). Each boundary between two layers has a reflection coefficient R. A reflectivity series is a time series in which these reflection coefficients are displayed versus time. Yet, the reflectivity series can be considered as filter for a wavelet that travels along the ray-path shown above through the model. The wavelet is convolved with the reflectivity series and the result is a seismogram (Figure 3.5).It is obvious that the seismogram resembles the reflectivity series. The smaller the wavelet is and the more that energy is concentrated in the wavelet, the better will be the similarity between the seismogram and the reflectivity series – a perfect result would be achieved by sending an impulse through the earth instead of a wavelet. Unfortunately, there is no such seismic source that could produce an impulse. Another way to transform the seismogram back to the reflectivity series is inverse filtering. A z transform can be used to calculate the inverse wavelet for a given source wavelet. This is also known as de convolution. However, the source wavelet, generated by the seismic source, is in general unknown and the inverse filter can only be estimated or predicted. More about " predictive de convolution". A filter operation, which is similar to convolution, is correlation. It was stated earlier that convolution passes the wavelet, reversed in time, through the filter. Correlation, on the other hand, passes the original wavelet through the filter without reversal. The result of this can be seen in Figure 3.6. The correlation of a waveform with another waveform is called cross correlation. An important application of cross-correlation is autocorrelation, the cross correlation of a wavelet with itself. An auto-correlation is an indicator of how much a wave resembles itself after a certain time, or how repetitive it may be (Figure 3.6.).

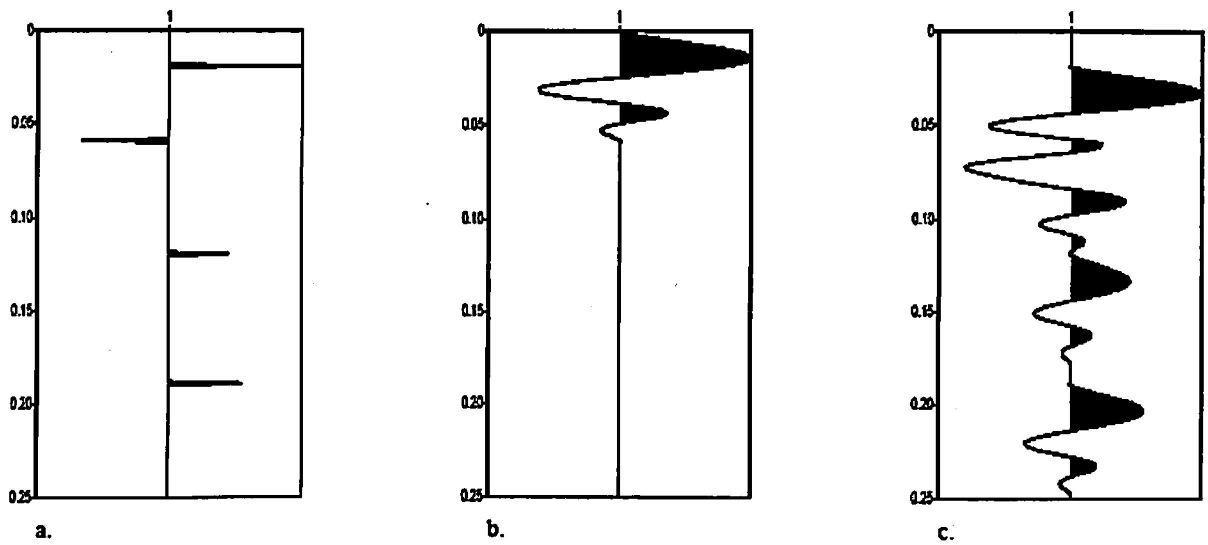
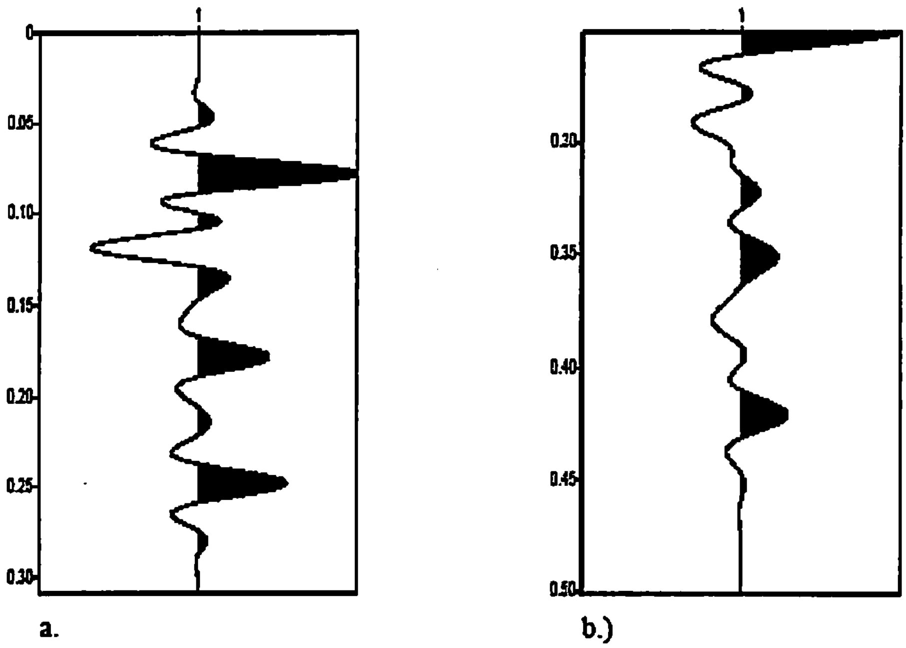


Figure 3.5: *Convolutional earth model.* (a) *Reflectivity series in time.* (b) *Source wavelet to be convolved with the reflectivity series.* (c) *Convolved signal derived from (a) and (b).*



ɟure 3.6: (a) *Crosscorrelation of the source signal and the convolved data in Figure 3.5 (b) and (c) respectively.* (b) *Autocorrelation of the convolved data in Figure 3.5 (c).*

The output of a linear filter process is always longer in time than the input. This can be easily verified with the model of the sliding wavelet. In case of an autocorrelation, the output is symmetrical and therefore not often displayed in its full length. Convolution correlation and the following Fourier transforms are comprehensive mathematical concepts –impossible to discuss sufficiently in this report.

3.4 Wavelets and Fourier Transforms

Every waveform can be represented by sum of sinusoids with different frequencies, known as the Fourier series. The summed sine waves may have different amplitudes and different phase shifts. These amplitudes and phase shifts can be displayed as function of frequency as shown in Figures 3.7 and 3.8. This representation of the wave is called frequency domain representation and is fully equivalent to the usual time domain representation (time –amplitude).The Fourier transforms converts waves from the time domain to the frequency domain. Most processing algorithms rely on the frequency domain representation of the signals. One can imagine, for example how easily frequency filtering can be done in the frequency domain. Fourier transforms can be also used to analyze signals. Figures 3.7 and 3.8 are two examples of such an analysis:

- Figure 3.7 shows an impulse or spike, containing all frequencies in the same amount: the phase spectrum shows a linear function, wrapping to ^every time it exceeds $+\pi$.
- Figure 3.8 presents an analysis of a linear–phase wavelet: containing a band of spectral frequencies, while having a linear phase function. Like an impulse, this wavelet is symmetric in time. If the peak of the wavelet occurred at time zero, it would have zero phase over all frequencies and would then be called a zero phase wavelet.

It was stated that the similarity between seismogram and reflectivity series is better if more energy is concentrated in the wavelet sent in the ground. Wavelets that fulfill this requirement the best and can be used in seismic data processing are minimum-phase or zero phase wavelets are symmetric, concentrating the energy mostly in the middle of the wavelet at time zero-phase wavelets for interpretation.

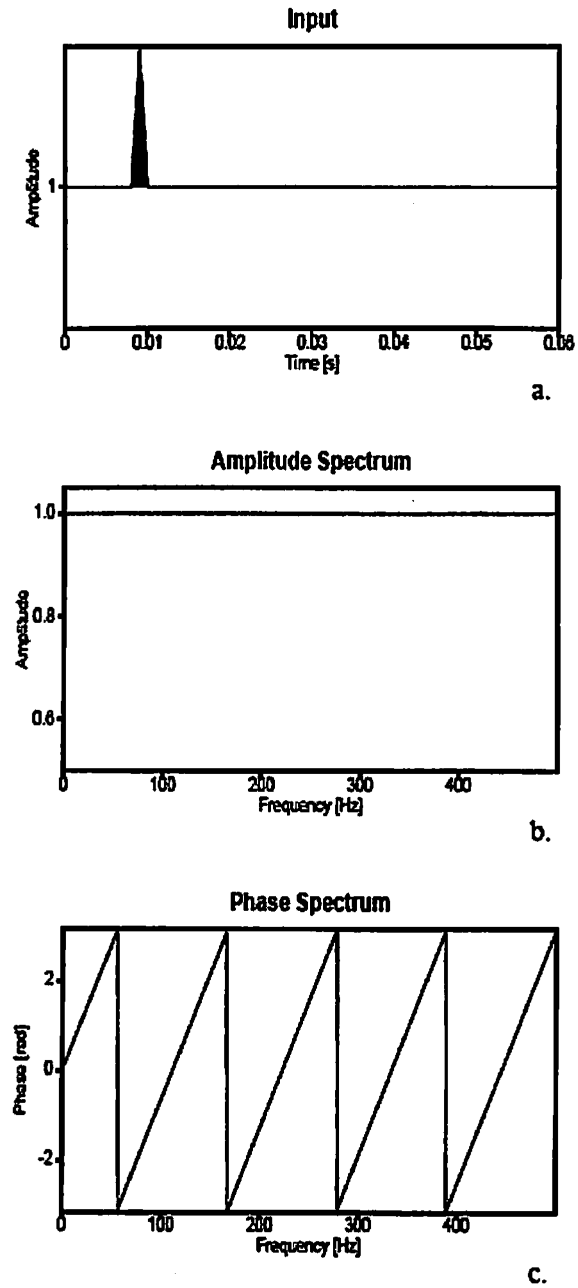


Figure 3.7: *Fourier analysis of an impulse.* (a) *Time domain representation.* (b) *Amplitude spectrum.* (c) *Phase spectrum.*

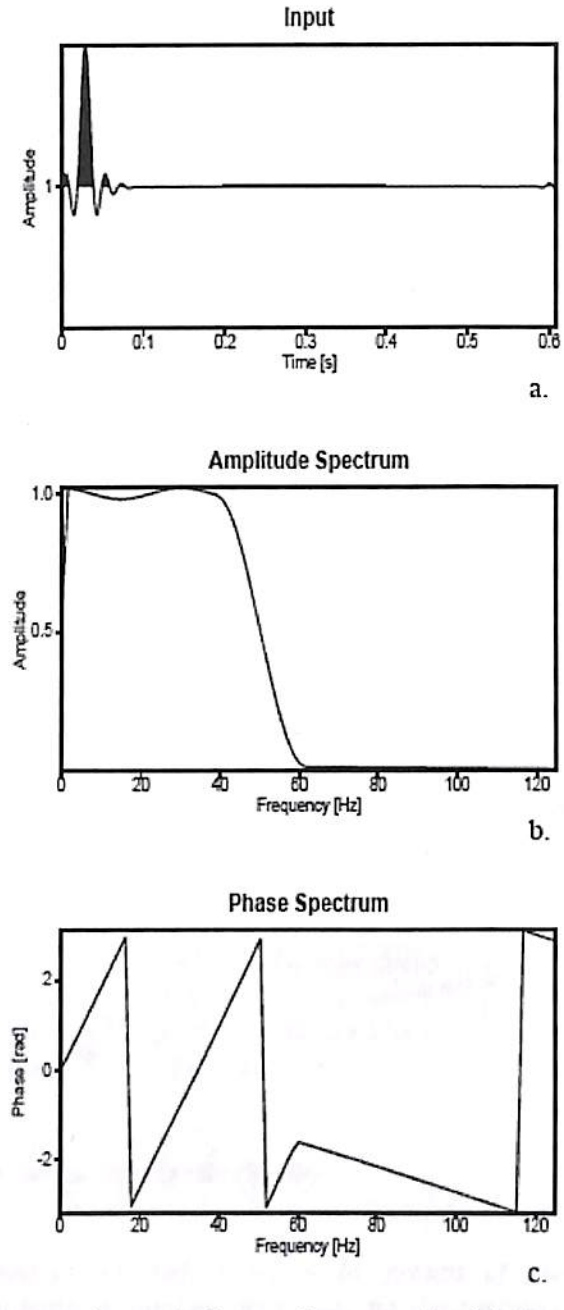


Figure 3.8: *Fourier analysis of a linear-phase wavelet.* (a) *Time domain representation.* (b) *Amplitude spectrum.* (c) *Phase spectrum.*

3.5 Seismic Acquisition

First a survey must be designed by defining the location of the shotpoints and receivers. Next, groups of geophones or hydrophones are placed at receiver locations. The phones within one group are laid out in an array designed to eliminate surface waves(figure 3.9). After the phones are connected to the recording equipment, the source can be fired at the source location. The signals received by the geophones or hydrophones are then recorded to tape. The number of channels (each group is a channel) of data collected is dependent on the recording equipment. The "recording equipment" that was used for generating data from the layered earth model is only capable of recording one channel. The survey layout is 200 shots, equally spaced at 50m, 60 traces are recorded per each shot location; receiver spacing is also 50m located midway. This gives a dataset of 12000 traces.

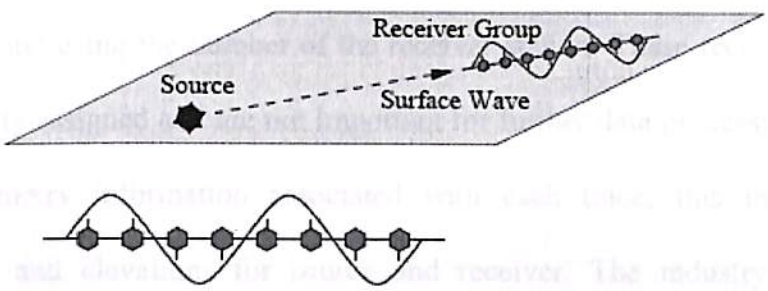


Figure 3.9: *Canceling of surface waves by groups of geophones. Signals of geophones within an array add up destructively when caused by a surface wave, while they add up constructively when caused by a reflected wave from the subsurface.*

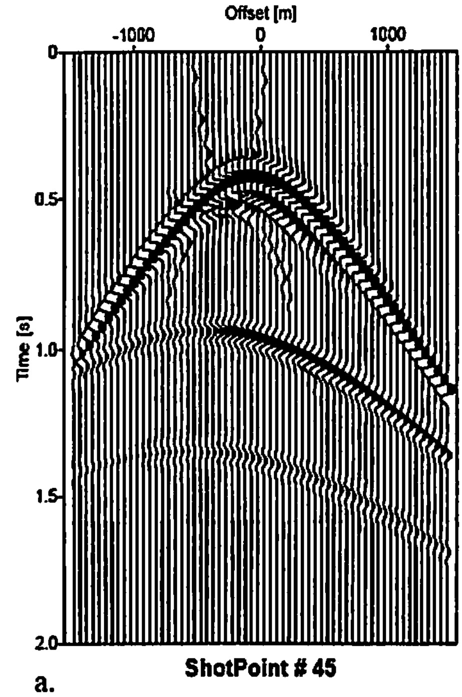
Chapter 4: Processing Synthetic Data

The synthetic data that was produced in the last chapter will be used in this chapter to introduce basic processing steps. Basic processing steps refers to all the manipulation of seismic data that is necessary to produce a migrated time section from shot gathers. The shot gathers are assumed to have geometry information associated in the trace header(SEG-Y format). The data presented in this chapter is free of noise and hence, simple to process. The significant problem of noise and how to improve the signal to noise ratio is discussed in the next chapter.

4.1 Shot Gathers

Seismic data is usually recorded in a demultiplexed SEG-D format. In SEG-D format all seismic traces are sorted into ensembles, called field files. Each field file contains all the traces that were recorded from one shot-point and is therefore called shot gather. The sorting is done using the number of the receiver station. These receiver station numbers are arbitrarily assigned and are not important for further data processing. More important is the geometry information associated with each trace; this includes, geographic coordinates and elevations for source and receiver. The industry standard for data containing geometry information is SEG-Y.

The synthetic data generated in last chapter was in SEG-Y format. The data file contains the location for both the source and receiver for each trace written in the sx(source) and gx(ground station) header locations. There is also a header location called offset, containing the horizontal distance between shotpoint and receiver. A flat surface will be assumed, hence, there is no need to have elevation assigned to receivers and sources.



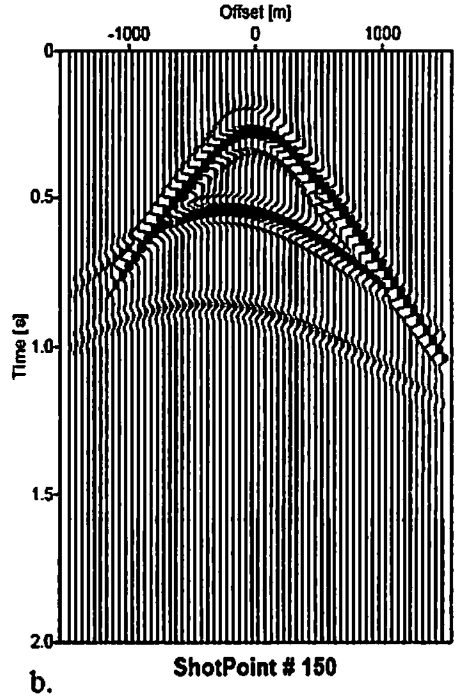


Figure 4.1 : (a) Shot gather from shotpoint #45 (b) Shot gather from shotpoint #150.

Two of the 200 shot gathers in the dataset are shown in figure 4.1. It is obvious that the arrivals from specific reflector do not follow straight lines. Additional horizontal traveltime for traces recorded at far offsets delays the arrivals for these traces. The resulting shape of reflectors in a shot gather is hyperbolic. By developing the equation for traveltime with a simple geometric construction, this can be easily verified.

4.2 Common-Midpoints (CMP)

Traces in a shot gather, as seen in figure 4.1, are caused by reflections at different location on the reflector. Yet, the traces of all shot gathers can be resorted so that all traces that have been reflected at one particular subsurface point are grouped together. These new groups, or ensembles, are called common-midpoint (CMP) gathers; the presumed reflection point is called common-midpoint (CMP). A CMP is the geometric midpoint between source and receiver and is the location of the true common-reflection point for a horizontal reflector and a horizontal recording datum. Figure 4.2 gives an example for a flat reflector imaged with 5 sources and 5 receivers- sources and receivers and equally spaced (not all ray paths are shown). In the presence of dipping reflector, more advanced processing can take advantage of sorting to presumed true reflection points at depth, which may not always lie at the common-midpoint. All traces in a CMP gather carry information about one point of the reflector, which for now is assumed to be at the location of the CMP(flat reflector). The more traces are contained in a CMP gather, the better will be the obtained information about the reflector. The number of traces in a CMP gather is termed "fold". The maximum fold shown in figure 4.2 is 5, the minimum fold is 1. Traces within a CMP gather differ in the way they traveled through the earth; all traces have different offsets and therefore different traveltime. CMP gather look similar to shot gather, although they usually contain fewer traces. The lower trace count is a result of a higher amount of CMPs – in our example the number of shotpoints in 200, whereas the number of CMPs is 458. Two of these 458 CMP gathers are displayed in figure 4.3.

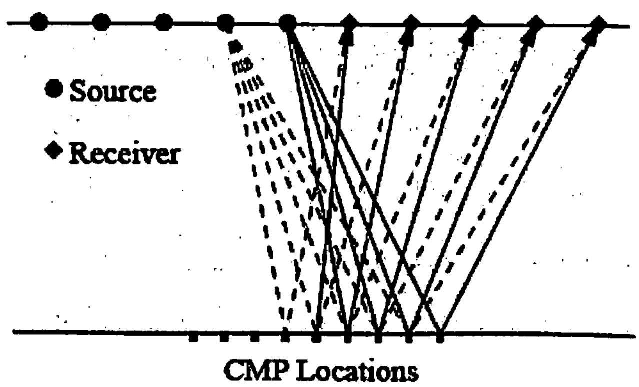


Figure 4.2: Common-midpoints are the geometric midpoints between source and receiver.

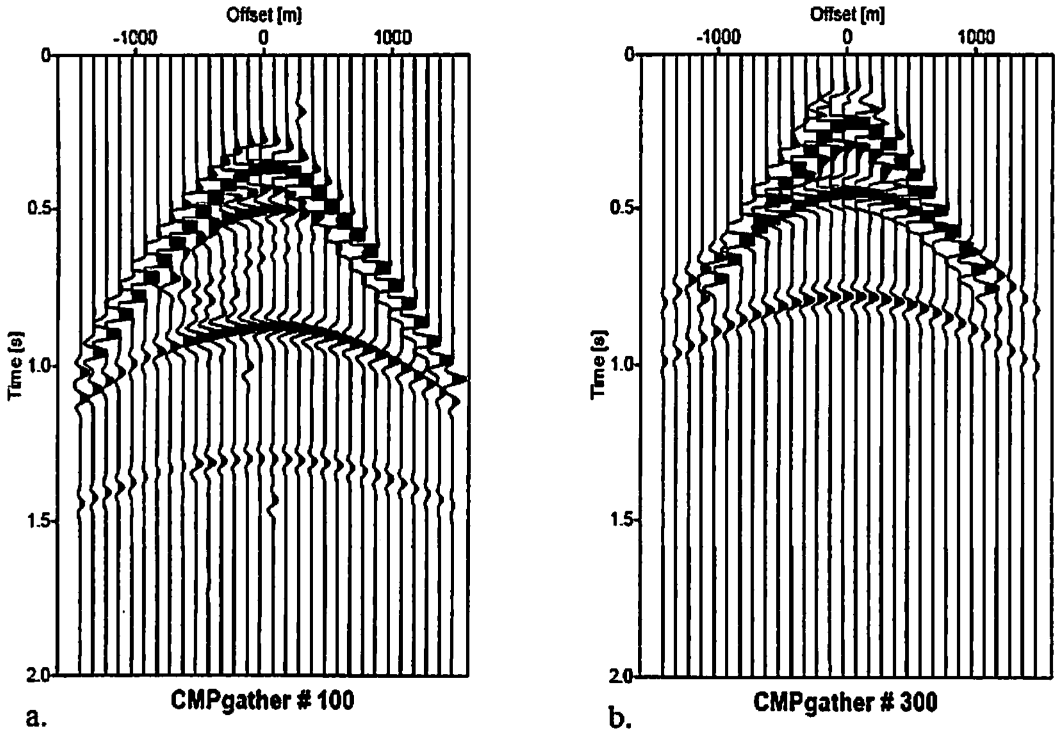


Figure 4.3: (a) *CMP gather #100.* (b) *CMP gather #300.*

The data in CMP gathers, is divided into two parts. First, the CMP number for each trace is calculated and assigned to the header "CDP", which stands for common-depth-point, a term often used improperly, and equivalent to CMP. The use of the term "common-depth point" should be avoided because there is no common point at depth if reflectors dip. Common-midpoints on the other hand are always defined by the geometric midpoint between source and receiver. After assigning the headers, the traces are sorted, using the newly assigned CMP numbers. As mentioned above, the second sort key is the offset.

4.3 Velocity Analysis

We will now sum different traces, also called stacked, to improve the signal-to noise ratio and to decrease the amount of data which will be processed to obtain an image of the subsurface. Before the stacking, a certain correction is applied on the different traces by carrying out a velocity analysis.

A good velocity model is the basis for :

• Stacking (Improvement of S/N-Ratio)

• Appropriate conversion from travel-time into depth

• Geometrical Correction (Migration)

Methods for Velocity analysis.

The aim of the velocity analysis is to find the velocity, that flattens a reflection hyperbola, which returns the best result when stacking is applied. This velocity is not always the real RMS velocity. Therefore, a distinction is made between:

• v_{stack}: the velocity that returns the best stacking result.

• v_{rms}: the actual RMS-velocity of a layer.

For a horizontal layer and small offsets, both velocities are similar. When the reflectors are dipping then v_{stack} is not equal to the actual velocity, but equal to the velocity that results in a similar reflection hyperbola.

There are different ways to determine the velocity:

• (t^2 -x^2)-Analysis.

• Constant velocity panels (CVP).

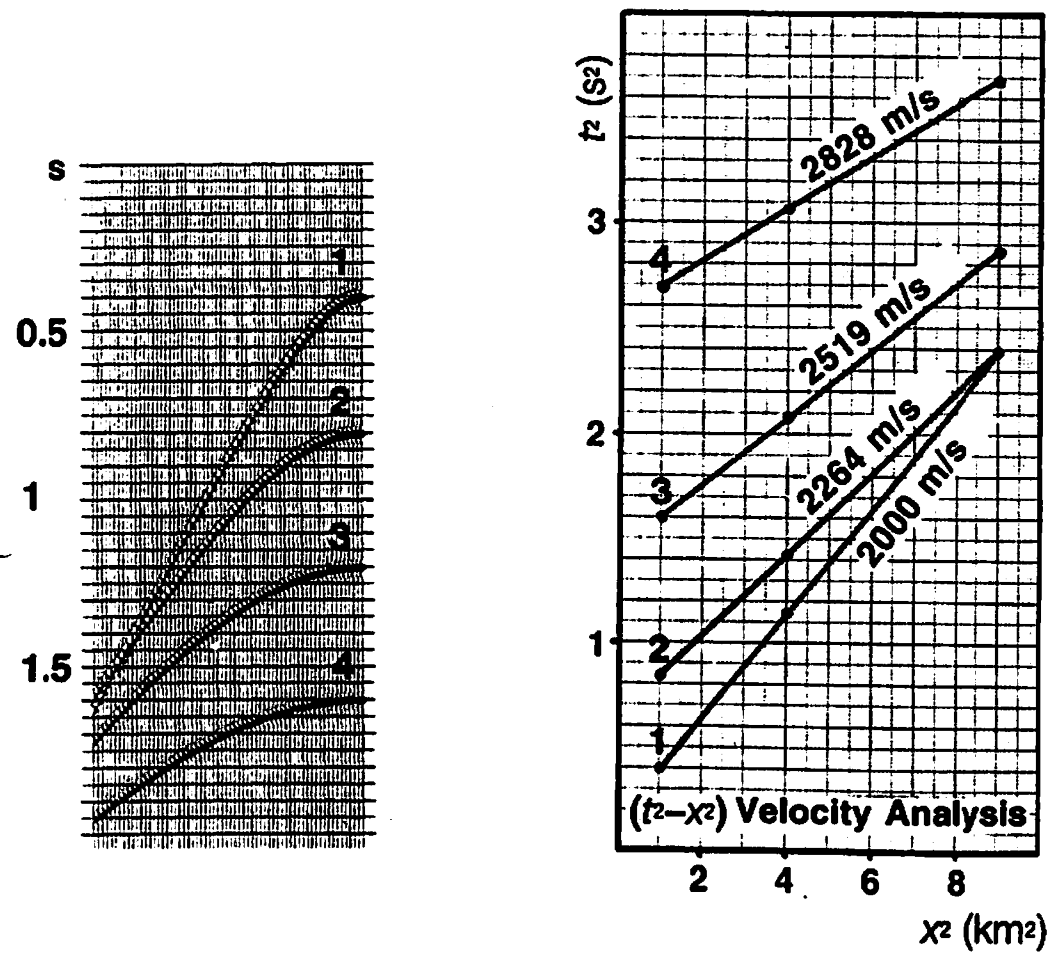
• Constant velocity stacks (CVS).

• Analysis of velocity spectra.

For all methods, selected CMP gathers are used.

(t^2-x^2)-Analysis

The (t^2-x^2)-Analysis is based on the fact, that the Moveout-expression for the square of t and x result in a linear event. When different values for x and t are plotted, the slope can be used to determine v^2, the square root returns the proper velocity.



Example of a t2-x2-Analysis.

CVP - "Constant velocity panels"

The NMO-correction is applied for a CMP using different constant velocities. The results of the different velocities are compared and the velocity that results in a flattening of the hyperbolas is the velocity for a certain reflector.

CVS - "Constant velocity stacks"

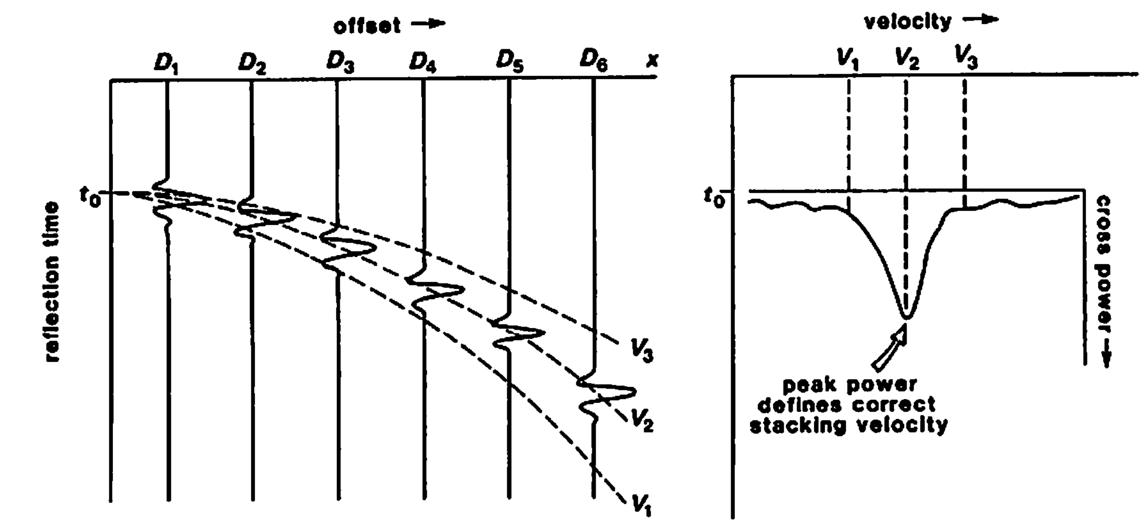
Similar to the CVP-method the data is NMO-corrected. This is carried out for several CMP gathers and the NMO-corrected data is stacked and displayed as a panel for each different stacking velocity. Stacking velocities are picked directly from the constant velocity stack panel by choosing the velocity that yields the best stack response at a selected event. CVP and CVS both have the disadvantage that the velocity is approximated as good as the distance between two test velocities. Both methods can be used for quality control and for analysis of noisy data.

Velocity-Spectrum

The velocity spectrum is obtained when the stacking results for a range of velocities are plotted in a panel for each velocity side by side on a plane of velocity versus two-way travel-time. This can be plotted as traces or as iso-amplitudes. This method is commonly used by interactive software to determine the velocities.

Different possible methods can be used to determine a velocity spectrum:

- amplitude of stacking
- normalized amplitude of stacking
- Semblance



Factors influencing velocity estimates

The accuracy of the velocity analysis is influenced by different factors:

- Depth of the Reflectors
- Moveout of the Reflection
- Spread length
- Bandwidth of the data
- S/N-Ratio
- Static Corrections
- Dip of the Reflector
- Number of traces

By a combination of CMP's that lie close together (Super gather), the accuracy is increased when a small number of traces per CMP are available (low coverage).

Errors due to dipping layers and insufficient static corrections can be reduced (DMO discussed later on).

There are several types of velocities defined in seismic reflection work: interval velocity, RMS velocity, normal moveout velocity, and stacking velocity are only the most important ones. Each of them is significant, although the stacking velocity is of highest importance in the context of velocity analysis for imaging. CMP gather contain traces from the same subsurface reflection point, but with different travelpath. These different travelpaths obviously cause different traveltimes. Using both pieces of information, time and distance, it is possible to estimate the velocity of the medium. However the travel distance is not really known, but the offset is. This turns out to be sufficient to solve the problem. Stacking velocities are found (picked) by running a velocity analysis on a CMP gather. The idea behind such a velocity analysis is to find out the velocities that flatten out the hyperbola on the CMP gathers and increase the power of the overall stack. Stack power is the energy contained in an event after adding all traces of a CMP gather together. Velocities can thus be picked for each event on the CMP gathers. There are three such events in the example presented here. Additionally a velocity for the top and the bottom of the time window should be chosen. For marine data, the velocity of the top would equal the water velocity. Velocities are not picked for each CMP. Velocities are picked using a few CMPs and then interpolating across the remaining CMPs. In general, this method works fine with only a few processing tools needing a more elaborated velocity field. A discussion of several tool to pick velocities is presented on the next page. All of these tools can, and should be, used. Relying on just one tool, like the following semblance plot, is not sufficient and should be avoided. The script that is presented at the end of this paragraph is an interactive tool to pick velocities. It provides all the tools below and it provides the opportunity to practice velocity picking and explore the meaning of over and under correction: if the picked velocity is too high, the event will be concave downward, if it is too low, it will be concave upward instead of a straight line. The first case is known as under correction, the second as over correction.

- **Semblance plot**

Semblance is a measure of multi channel coherence. To obtain a semblance trace for a specific velocity, all traces of a CMP gather, other than the zero offset trace, are corrected for their "late arrival" assuming the given velocity. The correction for the late arrival is an upward shift to correct for the additional horizontal travel time, also known as normal move out correction (NMO). The shifted traces are summed together and the semblance is calculated by:

$$\text{Semblance} = \frac{(\text{Amplitude})^2 \text{ of the Stacked Trace}}{\text{Sum of the } (\text{Amplitude})^2 \text{ of the Unstacked Traces}}$$

The higher the semblance, the better aligned are the event on the traces for the applied velocity. A semblance plot displays the calculated semblance over time and velocity. A high semblance should be picked for each event. Figure 4.4 shows a CMP with its semblance plot. Because synthetic data without noise was used for this semblance plot, both the numerator and the denominator in the semblance equation approaches zero producing artifacts of high semblance at the top and bottom line of the gather where there are no events.

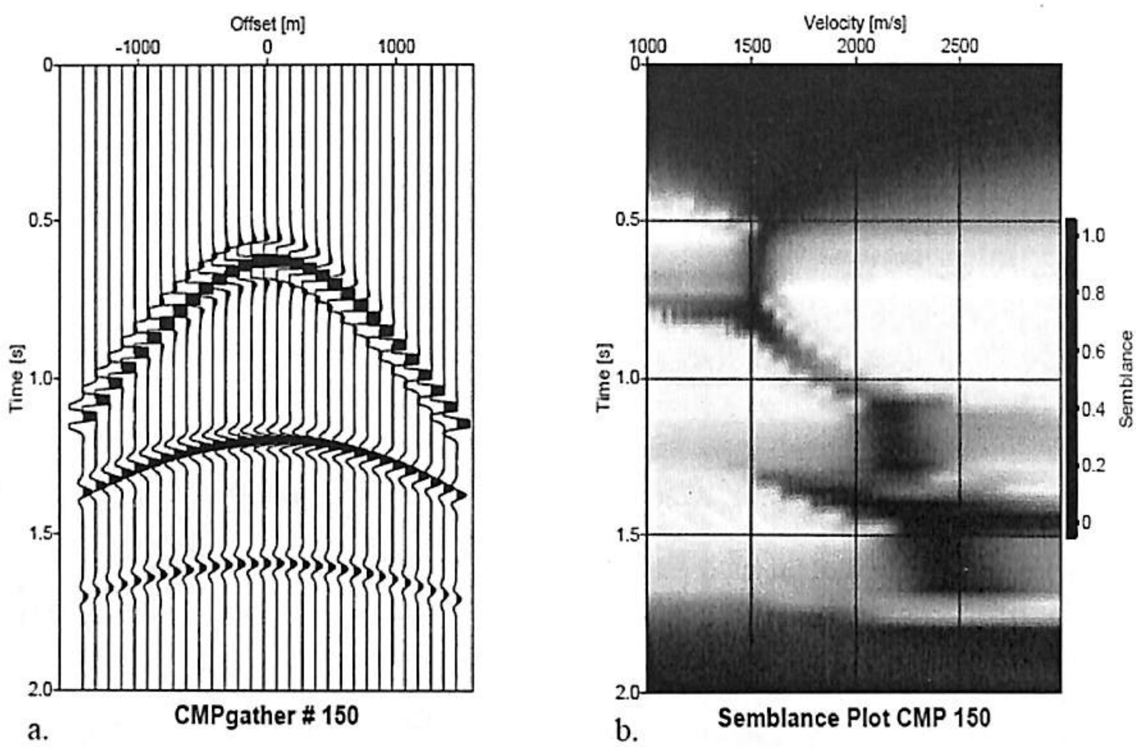


Figure 4.4: (a) *CMP gather #150 and* (b) *semblance for CMP gather #150.*

- **Constant velocity Stack**

A constant velocity stack (CVS) shows different panels of several stacked CMPs after NMO correction. Each panel is created by using one constant velocity over the entire time range. This velocity is used for an NMO correction of the chosen CMP and several neighboring CMPs. The traces in each CMP gather are added up (stacked) and the resulting traces are displayed next to each other in a panel. The panels give a preview of what the final stacked section would look like. The continuity the events show and the higher their amplitude is, the better will be the final stacked result. Due to the fact each panel represents only one velocity, the best picks over time will not fall within one single panel. This can be different for multi velocity stacks (MVS). Each panel of a MVS represents a predefined velocity function over time. MVS is commonly used in the industry today, but the idea behind CVS is not much different and it is easier to create. ACMP and its CVS are displayed in figure 4.5

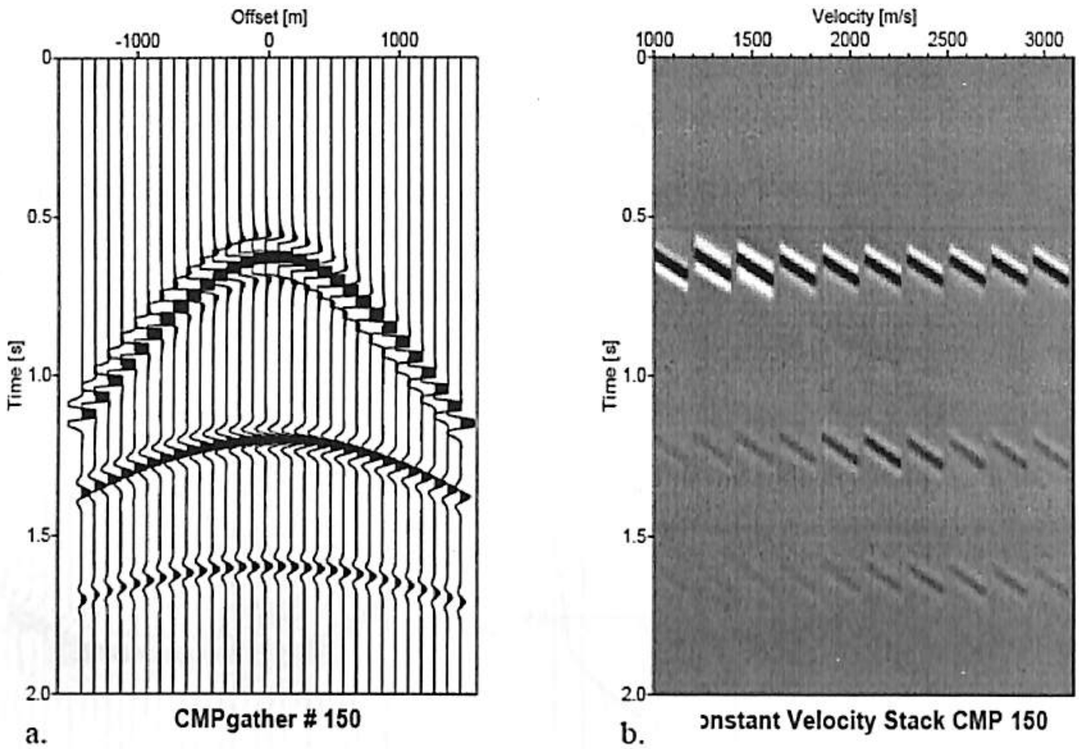
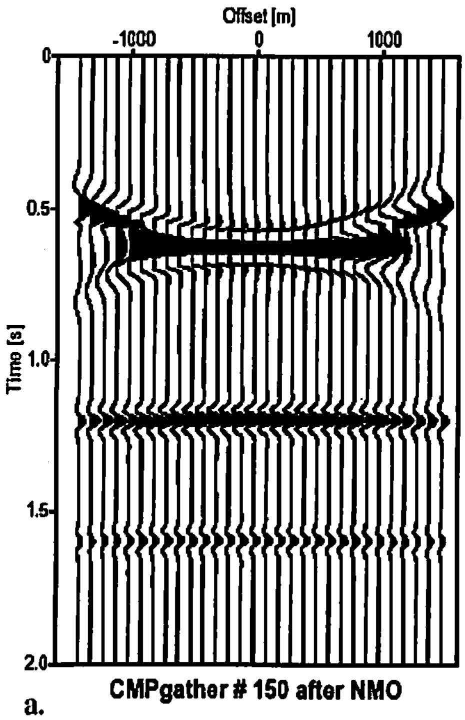


Figure 4.5: (a) *CMP gather #150.* (b) *Constant velocity stack for (a). Notice that the third panel in the CVS shows the highest amplitude for the first event, the sixth panel for the second event, and the eighth for the third event.*

Quality Control (QC) after picking

The semblance plot and the constant velocity stack are tools to decide what velocities are reasonable- they provide a possibility for a first velocity pick. However, a QC of the picked velocities afterwards is essential. Tools for velocity QC are, for example, displays of the velocity profile, velocity field, interval velocities, and CMPs after NMO correction. The velocity analysis in this project support two QC tools: a velocity profile and a NMO corrected CMP after picking(Figure 4.6). The velocity profile should be smooth and it should show increasing velocities with depth. Velocity inversions are possible, although not very likely and when observed, the picks should be double-checked. The corrected CMP should exhibit flat reflectors. Like mentioned above, an upward curved reflector means the picked velocity is too low; a downward curved reflector is the result of a too high velocity pick.The idea of velocity QC is that the feedback from it allows the processing geophysicist to chose better picks, which then will have to be QCed again and so on. This idea is also implemented in the script at the end of this chapter.



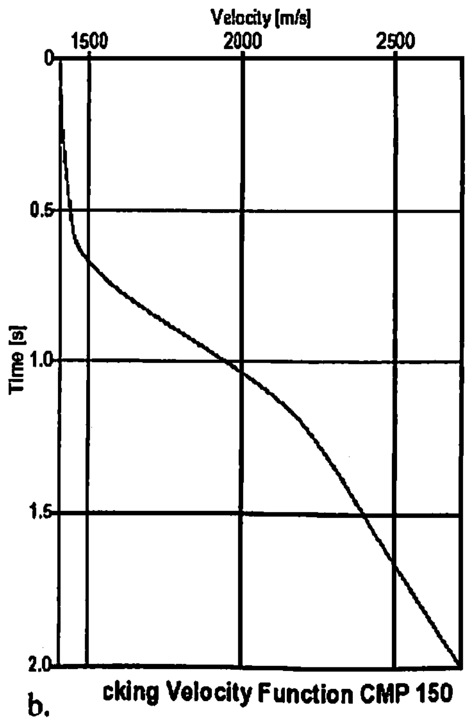


Figure 4.6: *Velocity QC:* (a) *Picked Velocity applied for NMO correction; events should be flat.* (b) *Picked velocity function over time.*

4.4 Normal Moveout Correction

Normal moveout (NMO) applies given velocity profile to traces in a CMP gather and corrects traces with a non-zero offset for their additional traveltime from source to receiver. NMO is the processing step that was also briefly mentioned while discussing velocity analysis in the last section. Yet, the correction is not only a shift in the traces, as implied in the last section, it also causes a stretch in the traces, called NMO stretch.

The NMO stretch is caused by two facts:

- The travel distance from source to receiver is not a linear function in depth(non-zero-offset case).
- The velocity profile is in general a non-constant function.

Considering these circumstances, it is obvious that the greater the offset and the shallower the time, the greater the upwards shift of the trace has to be. However, a differences in the shift of a trace for different time causes a stretch – the NMO stretch (Figure 4.7). The NMO stretch is most severe for shallow times and far offsets where it distorts the waveform and changes the frequency of the signal (Figure 4.8). This low amplitude is not desired and there is no reasonable treatment other than muting it out.

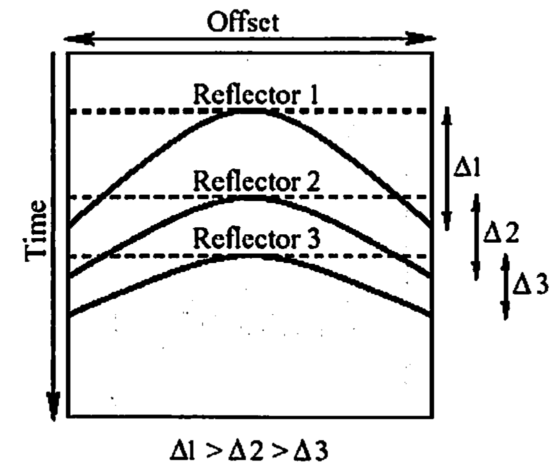


Figure 4.7: *NMO Stretch. Shifts Δi, necessary to flatten out events, vary with time. Time variant shifts cause the NMO stretch.*

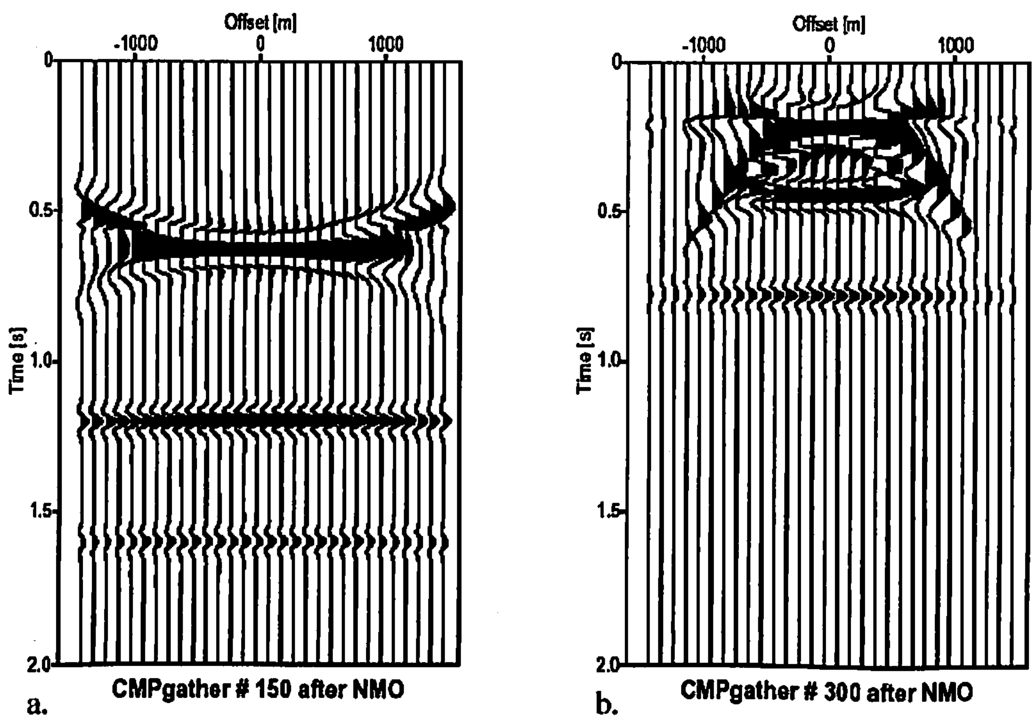


Figure 4.8: *NMO correction of the earlier presented CMP gathers (Figure 4.3).* (a) *CMP gather #150.* (b) *CMP gather 300.*

4.5 Mute

Seismic processing often requires editing of data. This can be scaling traces, killing traces, or deleting parts of a trace. Because this chapter uses relatively perfect synthetic data, only a small amount of editing is necessary. A mute zeros out all information in a specified window. A taper is used between the muted section and the original data to avoid boundary effects, particularly ringing in subsequent processing steps. The mute that is used here is a front mute applied at the top of each trace, starting at zero time and ending at a specified time behind the NMO stretch. Also often seen in seismic data processing is a tail mute at the trace end. The corrected CMP gathers with a mute applied can be seen in figure 4.9. The same CMP location as in figure 4.8 were chosen to demonstrate the effect of the mute.

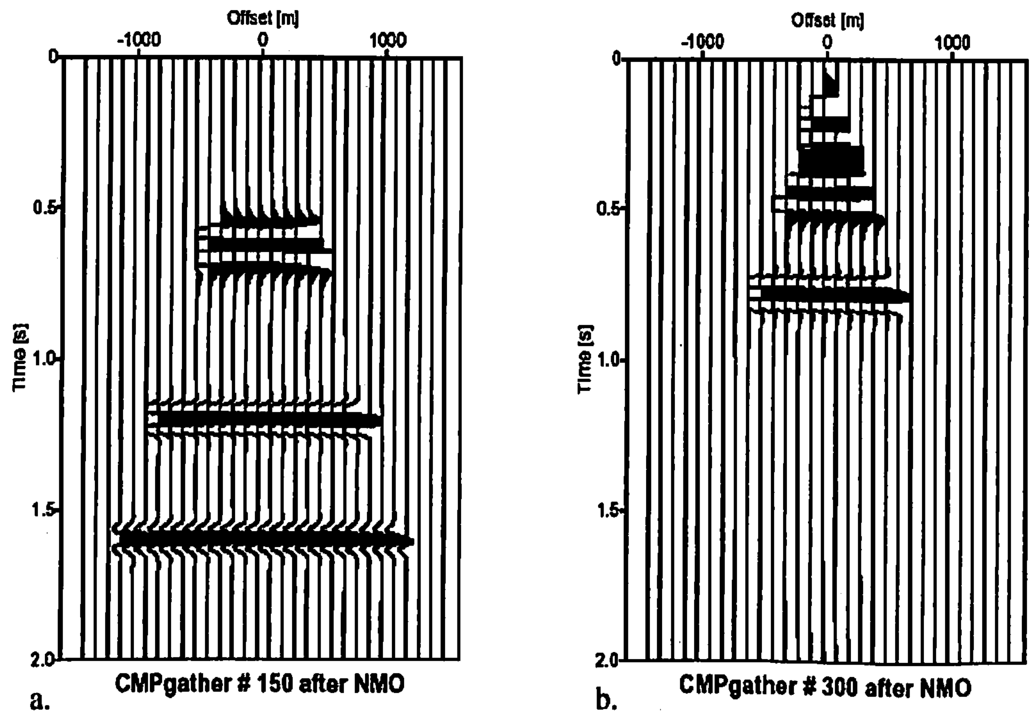


Figure 4.9: *NMO corrected CMP gathers after stretch mute.* (a) *CMP gather #150.* (b) *CMP gather #300.*

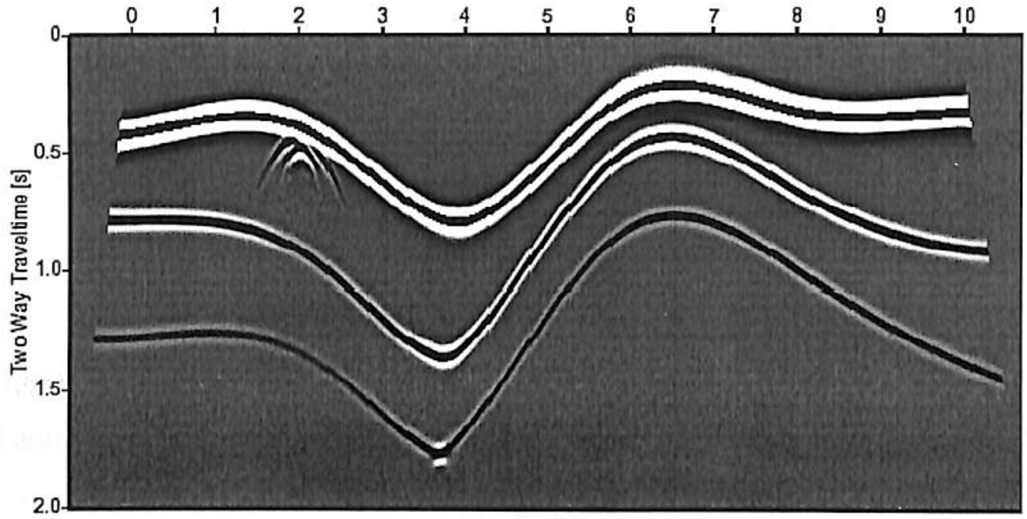
4.6 Stack

At this point in the processing stream, the traces in the CMP gathers can be summed together. All events should be flat lines at this stage and the events on the traces should add up constructively. The summation of the traces is called stacking. The amplitude of a stacked trace is divided by the number of input traces, so there is no increase of amplitude in this operation. However, assuming there would be noise in the data, the signal to noise ratio would increase through stacking. This is due to the fact that noise is often random and does not random and does not add up constructively as the signal does. After all CMP ensembles have been stacked, i.e. each reduces to a signal trace, the resulting traces at each CMP can be displayed next to each other. Such a display is called a stacked section or stack.

The stacked section of the given model (figure 4.10) resembles the original model well, but there are some deficiencies:

- The dips of the reflector are not right.
- Anticlines appear too wide, synclines too narrow.
- The small dense object in the model diffracts energy and is not imaged very well.

These observations can be abridged to the statement: the reflection points are not at the right place. A tool to bring reflectors back to the place where they belong is migration.



Stacked Data

4.7 Migration

The zero offset stack section that was created in section 4.6 imaged incorrect dips of events. The reason for this is that reflections from dipping reflectors occurs at points that are not the CMP locations, as assumed in a stacked section. Post-stack time migration corrects for this incorrect assumption by moving the reflectors to their true position (figure 4.11). Other possible processing techniques to correct for dip are pre-stack dip-moveout processing and prestack and post-stack depth migration.

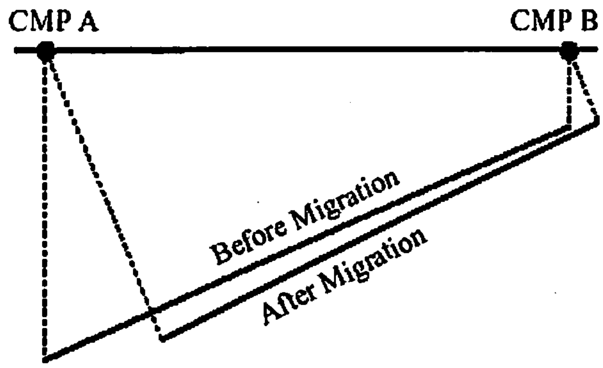


Figure 4.11: *A dipping reflector before and after migration. Migration increases the dip of reflectors.*

Migration also bundles diffracted energy and removes other artifacts in the stacked section such as "bow-ties" caused by synclines. A small bow-tie can be seen below the deepest syncline in the stacked section in figure 4.10. Migration has, like almost any other processing step, some side effects: Migration adds noise to the data. This can be seen in figure 4.12. Migration "smiles" within the data are an indicator for over migration, a migration with too high velocities. Migration tests with different velocities must be made to obtain the best migration result, although deciding about the best result is always a subjective task. Migration smiles at the end of reflectors are unavoidable edge effects, caused by the limited aperture at points where the fold drops to zero because of the mute applied.

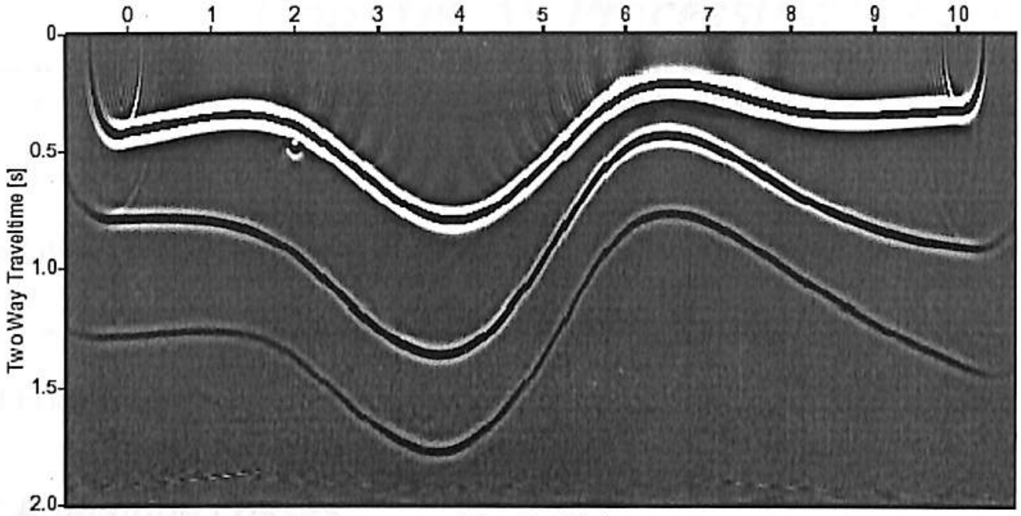


Figure 4.12: *Migrated section of the stack shown in Figure 4.10.*

Migration moves dipping events in the up-dip direction and collapses diffractions, thus enabling us to delineate faults while retaining horizontal events in their positions. The goal of migration is to made the stacked section appears similar to the geological cross-section in depth along a seismic traverse. The migrated section, however, commonly is displayed in time. One reason for this is that velocity estimation based on seismic and other data always is limited in accuracy. Therefore, depth conversion is not completed accurate. Another reason is that interpreters prefer to evaluate the validity of migrated sections by comparing them to the un-migrated data. Therefore, it is preferable to have both section displayed in time. The migration process that produces a migrated time section is called time migration.

Chapter 5: Processing Real Data

Processing real seismic data requires some additional processing tools besides these discussed in Chapter 4. The additional processing address noise and other imperfections in real data, which was not seen in the synthetic data, which was not seen in the synthetic data discussed previously. All shot gathers used in this chapter are taken from a worldwide assortment of common-shot gathers .

5.1 Amplitude Losses

An imperfection of real data is a server loss of amplitude with time. An amplitude loss Could even be observed in the synthetic dataset presented earlier, a result of the model's capability of simulating spherical divergence and partial reflections.

Spherical divergence is a loss of energy, and hence, amplitude due to the fact that the area of a wave-front increases with time. The energy of the wave spreads over the entire wave-front, a spherical partial reflection in homogenous media, and therefore the energy per unit area decreases. Partial reflection also decreases the energy and amplitude of a wave: when a wave is separated into a reflected and a refracted wave, obviously the energy has to be divided too.

The amplitudes of real data are additionally attenuated by other effects such as absorption and scattering. Absorption can be explained as loss of energy due to friction in the earth, A propagating wave forces particles and fluids to move against each other. Scattering is related to spherical divergence: then convert the energy partly to heat. Scattering is related to spherical divergence: the amplitude loss is due to an increasing wave-front area as a wave is scattered by an obstacle.

Processing of synthetic data is possible without any amplitude corrections. Looking at a real shot gather and its unbalanced amplitude function (Figure 5.1) however, shows the importance of amplitude corrections or gain control in real data. Different tools for gain control and amplitude balancing will be discussed in the next paragraphs. All of these tools can be tested with an interactive script shown in Table 5.1 (b), which should be started by using script gain * in Table 5.1 (a). This script allows the combination of several of the gain functions and the viewing of the results on the screen.

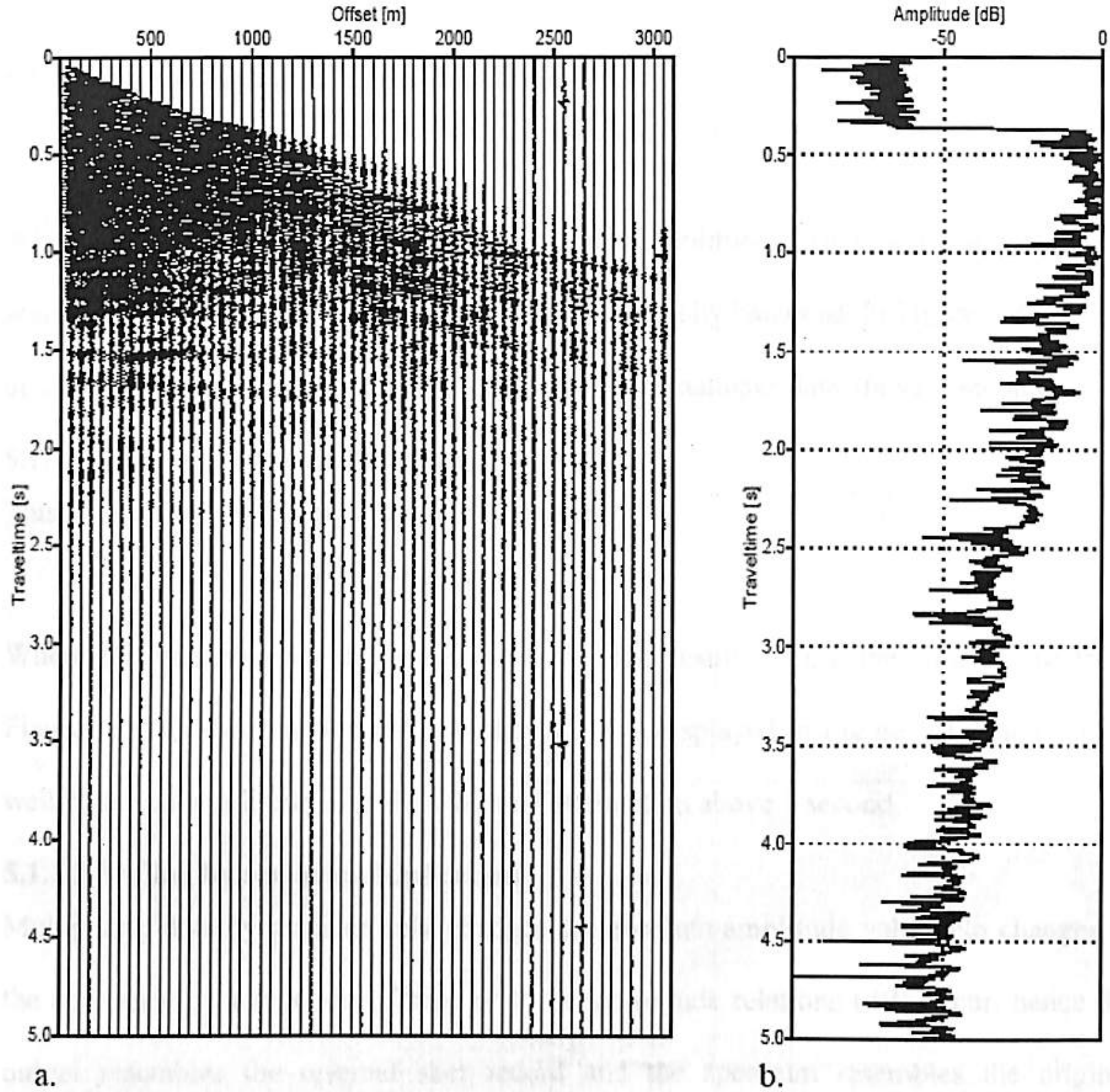


Figure 5.1 : (a) Shot gather [YILMAZ 39] without gain correction. (b) Amplitude envelop trace for a single trace of (a), offset=1000ft.

5.1.1 Data independent Amplitude Corrections

The following amplitude corrections are based on data independent scaling functions. Applying the inverse function will remove the applied scaling and the original data can be restored. This type of scaling functions discussed later are often used at a later processing stage.

5.1.1.1 Multiplication by a Power of Time

Multiplication by a power of time is a gain function of the form

$$A^*=A^*t\ X$$

Where A* is the new amplitude, A the original amplitude, t time and x a scalar. The scalar x should be chosen so that the amplitude is equally balanced. In Figure 5.2, a scalar of 2.8 was chosen. Note that this gain function also attenuates data above 1 second.

5.1.1.2 Exponential Gain Function

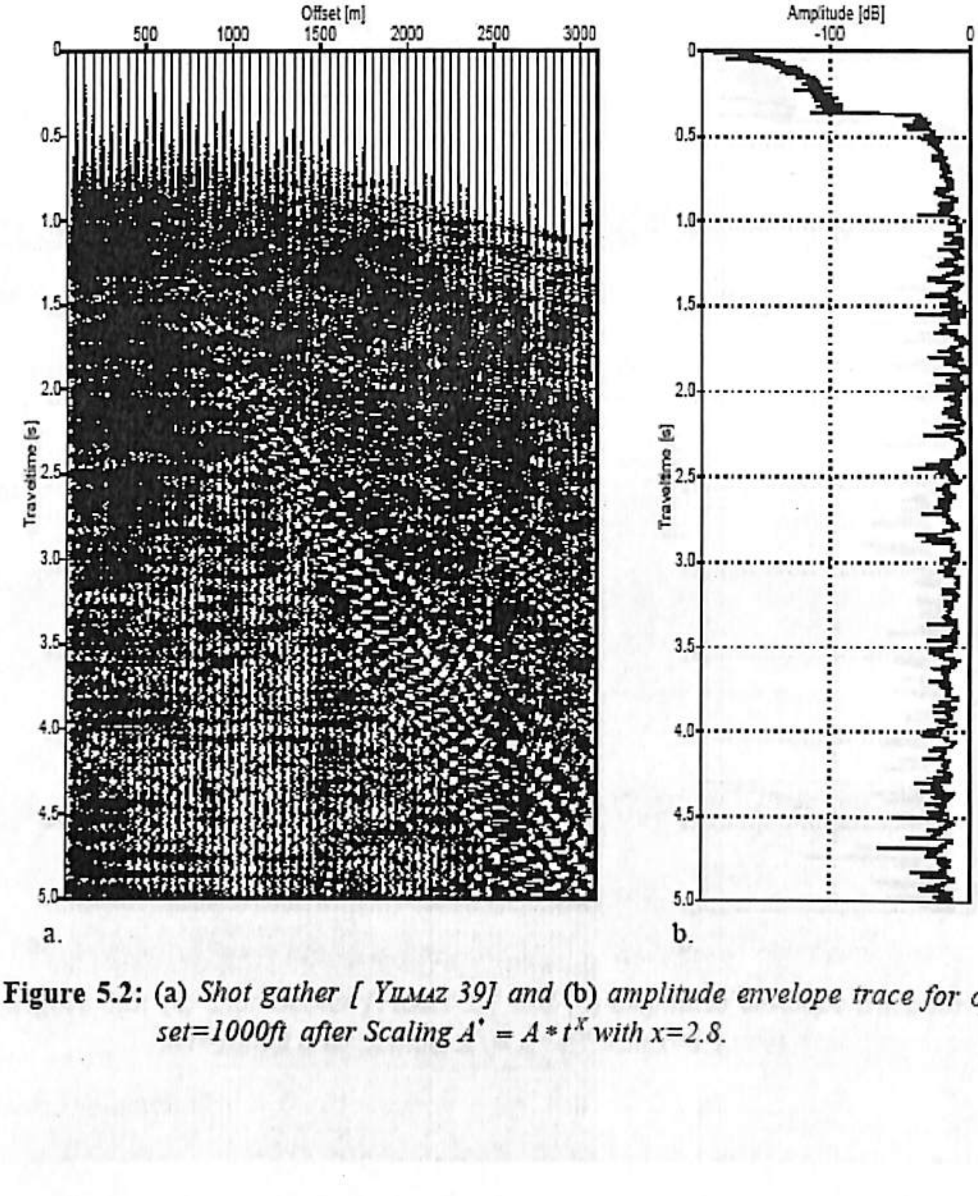
This exponential – type gain function is given by

$$A^* = A^*\ e\ x\text{-}1$$

Where the variables are the same as above. The result scaling the shot record from Figure5.1 with this function and a scalar of 1.1 is displayed in Figure 5.3. The result is well-balanced amplitude over time with no attenuation above 1 second.

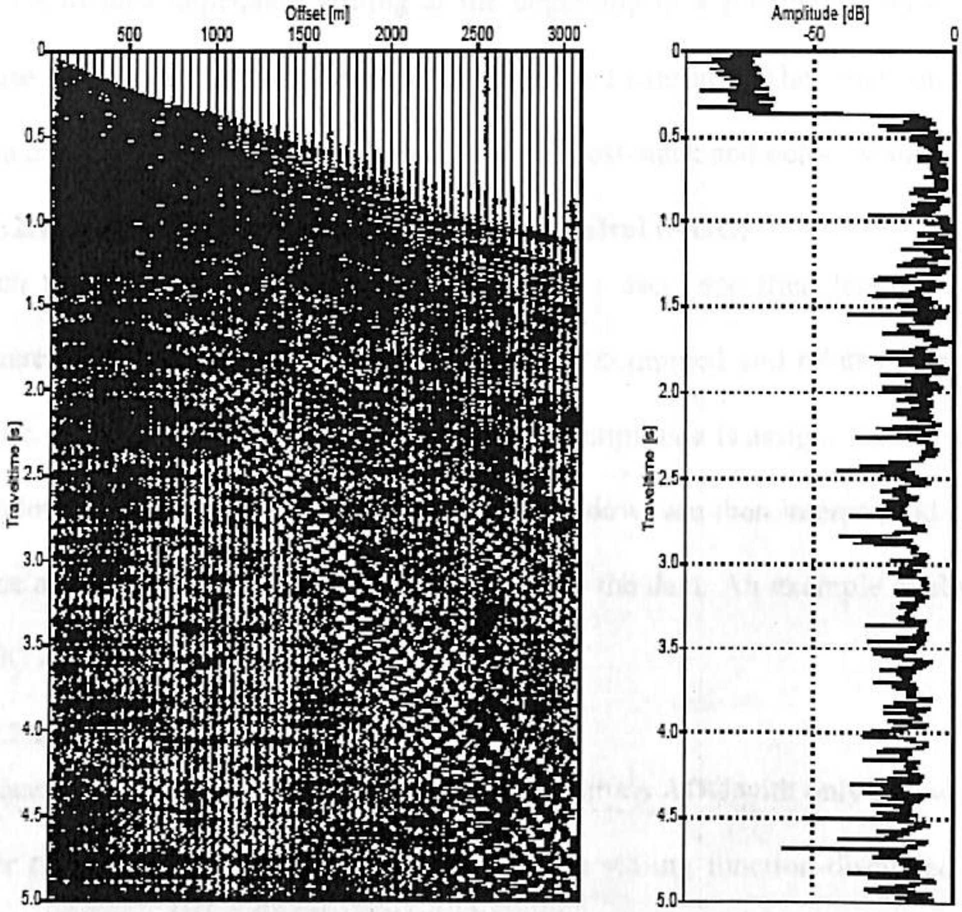
5.1.1.3 Scaling by scalar multiplication

Multiplying data by a scalar only changes the absolute amplitude value. No changes to the amplitude as a function of time or to the amplitude relations (dB) occur, hence the output resembles the original shot record and the spectrum resembles the original amplitude spectrum.



a. b.

Figure 5.2: (a) *Shot gather [YILMAZ 39] and* (b) *amplitude envelope trace for offset=1000ft after Scaling* $A' = A * t^X$ *with x=2.8.*



a. b.

Figure 5.3: (a) *Shot Gather [Yilmaz 39]* and (b) *amplitude envelope trace for offset=1000ft after Scaling* $A^* = A * e^{x*t}$ *with x=1.1.*

5.1.2. Data dependent Amplitude Corrections

Data dependent amplitude corrections use amplitude information from the input data in order to scale it. The advantage of this is a well-balanced amplitude spectrum after scaling, however, the disadvantage is that the scaling cannot later be removed. A poor choice of data dependent scaling at the beginning of a processing flow cans therefore cause serious problems and should be considered carefully. The most common places of data dependent scaling in a processing flow is post-stack and before visualizing data.

5.1.2.1. RMS Amplitude Automatic Gain Control (AGC)

Each trace is subdivided into windows with a user- specified length. The root-mean square (RMS) amplitude over each window is computed and related to a desired RMS value. The scaling factor to obtain the desired amplitude is assigned to the center of each window. These values in the centers of the windows are then interpolated over the entire trace and the resulting function is used to scale the data. An example of RMS amplitude AGC is shown in Figure 5.4

5.1.2.2 Trace Balancing by RMS Values

Trace balancing by RMS value is a RMS amplitude AGC with only one window applied over the entire trace. Trace balancing, like the scaling function discussed in Paragraph 5.1.1.3, does not change the amplitude as a function of time. This can be an important consideration for true amplitude processing .

5.1.2.3 Clipping

Clipping reduces all amplitudes above a user-specified clipping factor (percentage of the maximum amplitude) to the value of the amplitude specified by this factor. Clipping is rarely used to correct amplitude in a processing stream, but it is almost always used to scale data before displaying it-all plots in this report were clipped.

In SU, the parameter précis often used to determine the percentage of data not to be clipped in a display.

I

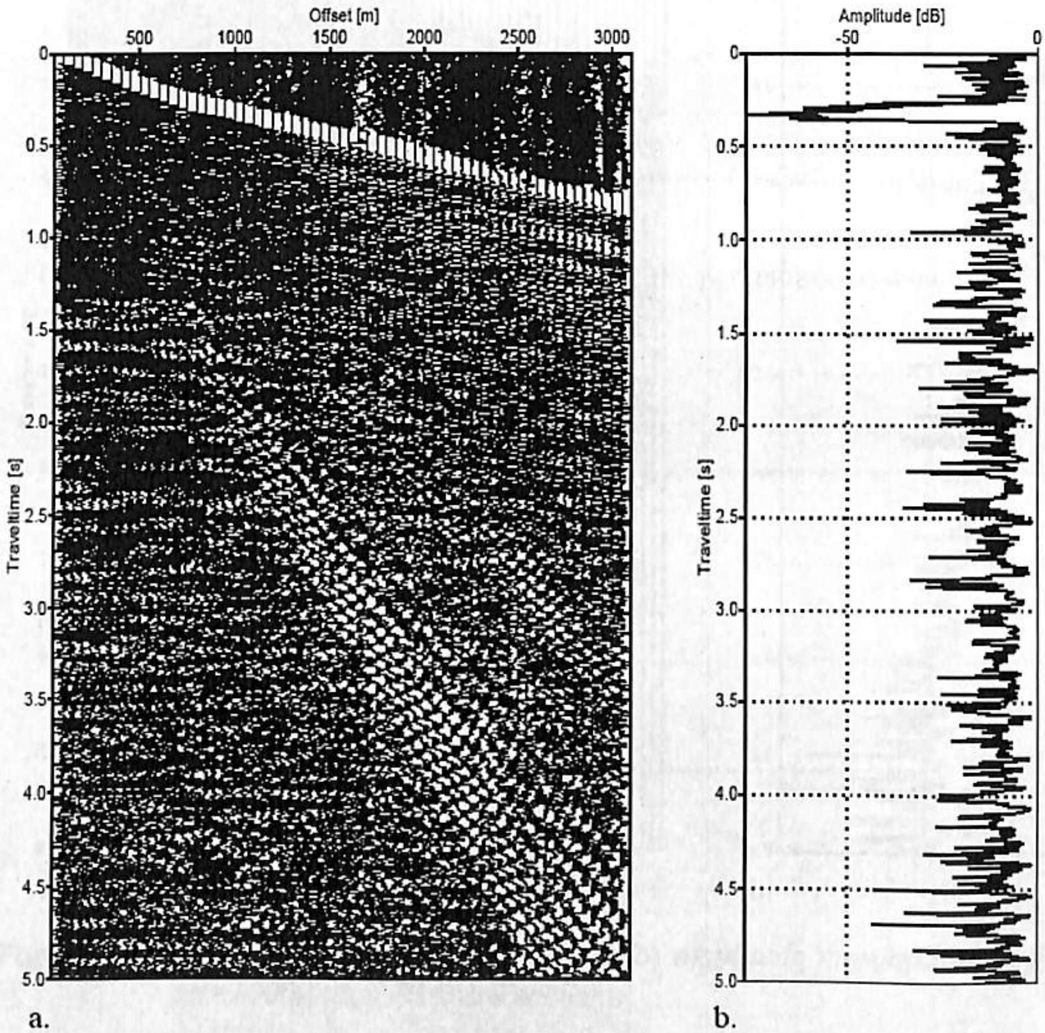


Figure 5.4: (a) *Shot Gather [YILMAZ 39] and* (b) *amplitude envelope trace for offset=1000ft after AGC; windowlength=200ms.*

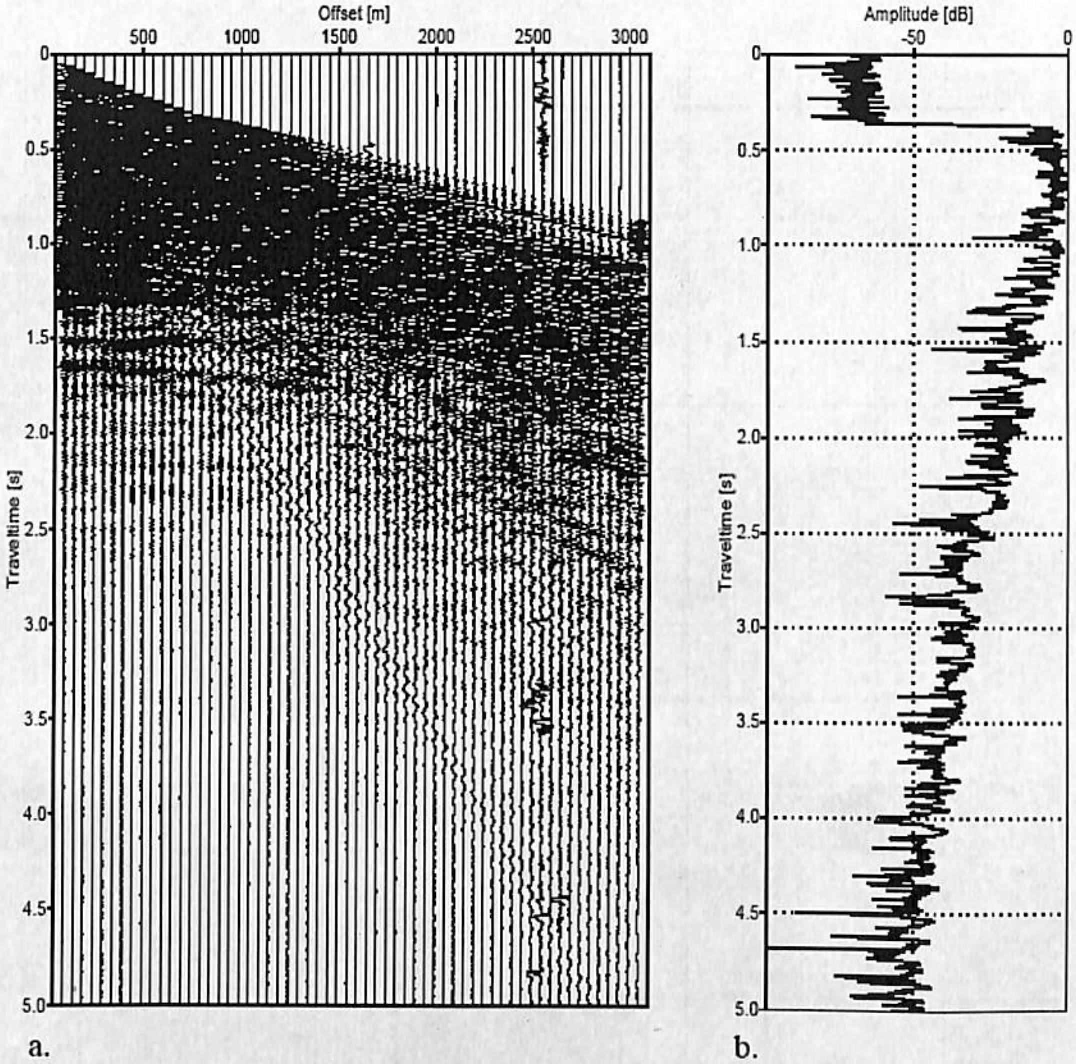


Figure 5.5: (a) *Shot Gather [YILMAZ 39]* and (b) *amplitude envelope trace for offset=1000ft after RMS Balancing.*

5.2 Noise

Synthetic, noise free data were used to study basic processing steps in Chapter 4. However, noise free data never recorded during real seismic acquisition. Table 5.1 gives examples of common observed types of noise that can contribute to real seismic data.

Coherent Noise	Ambient Noise
Ground Roll	Recording Equipment
Direct Wave	Bad Geophone Coupling
Reverberation	Spikes
Ship Noise	Weather / Wind
Rig Noise	Swell Noise
Rig Diffraction	Vehicles
Power Lines	Animals

Table 5.2: *Classification of noise.*

The characteristic of coherent noise is regularity on a trace-to-trace basis. A rig diffraction, for example, is visible on each trace, and it is possible to predict how the noise on the next trace should appear. Ambient noise, on the other hand, is random and unpredictable. There are different processing tools available for both classes of noise. These tools will be discussed in the Sections 5.2.1. to 5.2.4. One tool to reduce ambient noise in seismic data was already discussed in Chapter 4: stacking is the most effective tool at the geophysicist's disposal to cancel out random noise.

Data processing can never eliminate all noise in seismic data. The objective of data processing is to increase the signal to noise ration(S/N) as much as possible. The tools discussed later will help to achieve this objective.

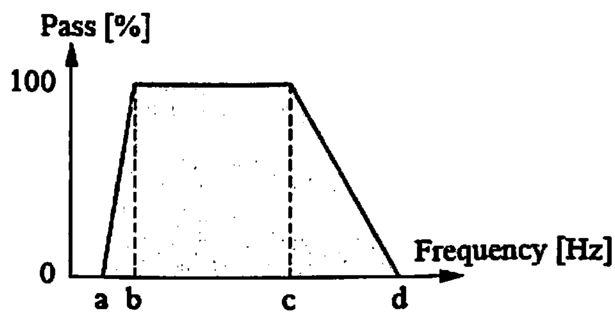
5.2.1 Frequency Filtering

The purpose of frequency filtering is to remove certain (noisy) frequency components of seismic data and to pass the remaining data through the filter unchanged. Surface waves (ground roll), for example, are usually observed as low frequency and large amplitude events and can be filtered out with frequency filtering.

Frequency filtering is performed in the frequency domain. A Fourier transform is done necessary before filtering and an inverse transform after. The user simply has to specify the cut-off frequency and the slope of the taper between full-reject and full-pass. The taper should be designed to avoid boundary effects (ringing). The slope of the taper at low frequencies may be steeper than the slope at high frequencies.

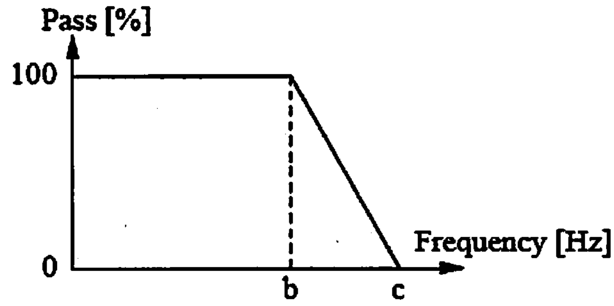
The specification of cut-off frequencies and taper in Seismic Unix is done by specifying four monotonically increasing frequencies and the desired percentage of the amplitude to pass at these frequencies. This setup provides the flexibility to design various filters:

Band-pass filter- The amplitudes associated with the frequencies a,b,c,d are 0,1,1,0. An inverse filter can be designed by choosing the amplitude as 1,0,0,1.



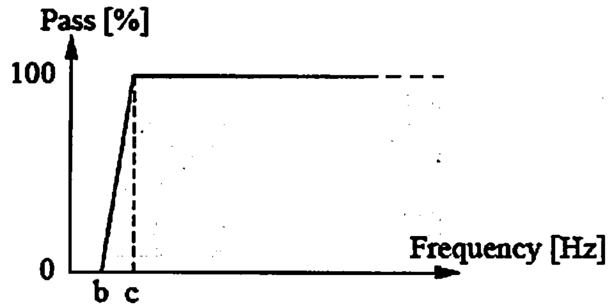
Band-pass Filter

Low-pass/ High-cut filter – The amplitude series for the frequencies a,b,c,d is 1,1,0,0, where frequencies a and d are arbitrary. The taper is between frequencies b and c.

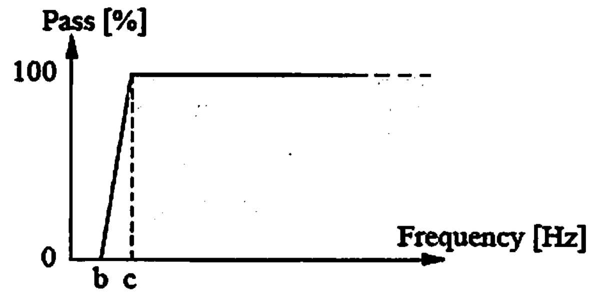


Low-Pass Filter.

High-pass / Low-cut filter- The amplitude series for a high-pass filter is 0,0,1,1, where the frequencies a and d are arbitrary.



High-Pass Filter



High-Pass Filter.

The most common used frequency filter in seismic processing is the band-pass filter. Seismic energy is generally recorded in a certain range of frequencies. The low end of this range marks the ground roll, which is undesired. The high end, above which only noise can be observed, depends on various factors to include : the type of source that was used, the penetration depth of the waves, and the properties of the ground. The seismic data, as well as the frequency can be compared before and after filtering. Also part of the filter test is the output of the inverse filter, or a rejected data. Most ground roll energy could be removed with this filter, but not all of it. A higher low-cut frequency (at the low end of the band) would remove the ground roll completely; however, this would also remove some seismic data. f-k filtering, is used to remove the remaining noise instead of a stronger frequency filter.

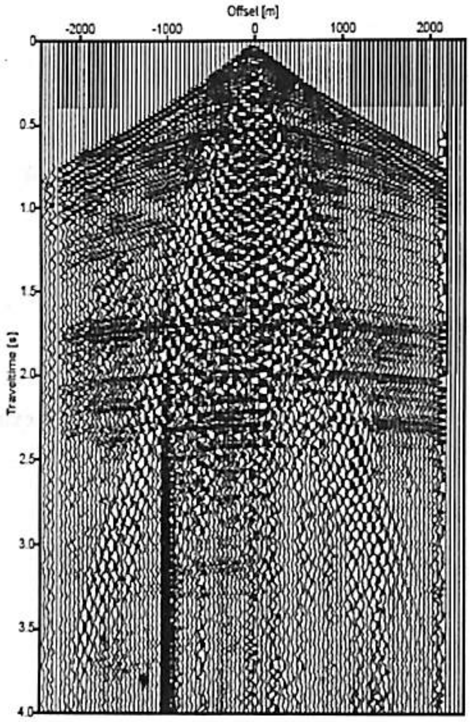


Figure 5.9: *Shot gather [YILMAZ 25] before filtering.*

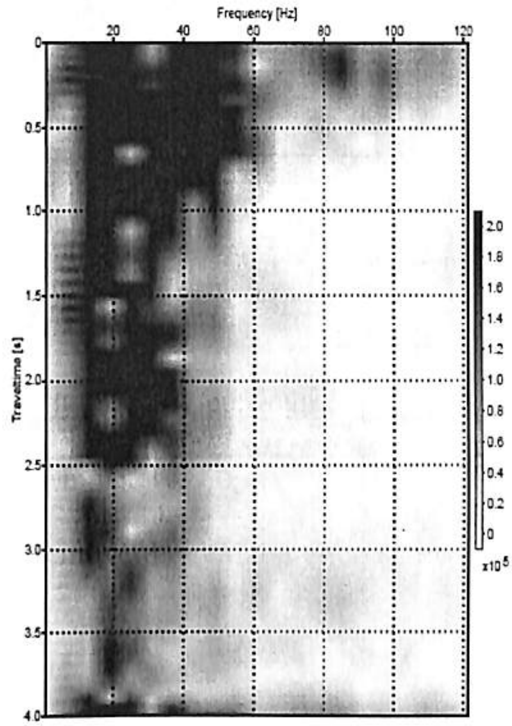


Figure 5.10: *Frequency spectrum before filtering.*

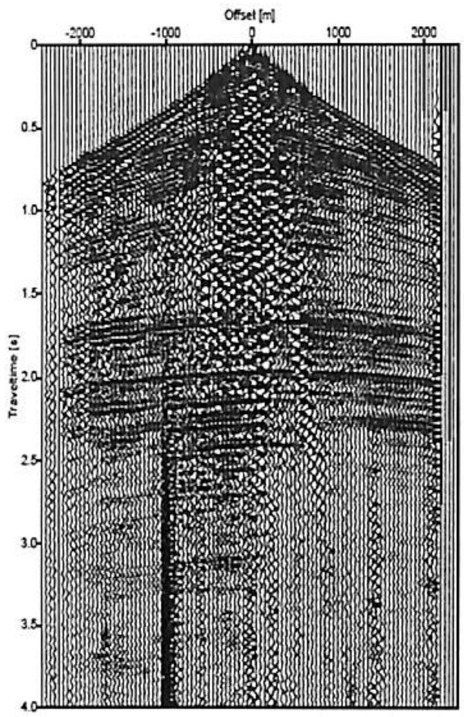


Figure 5.11: *Shot gather [YILMAZ 25] after band-pass filtering (10,12,60,75 Hz).*

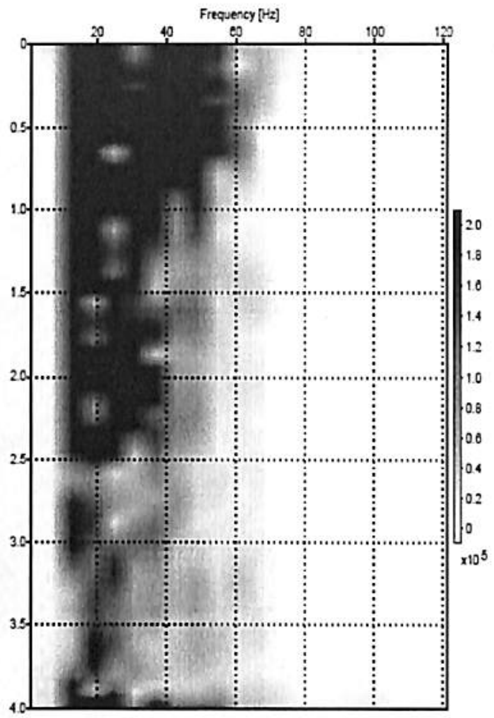


Figure 5.12: *Frequency spectrum after band-pass filtering (10,12,60,75 Hz).*

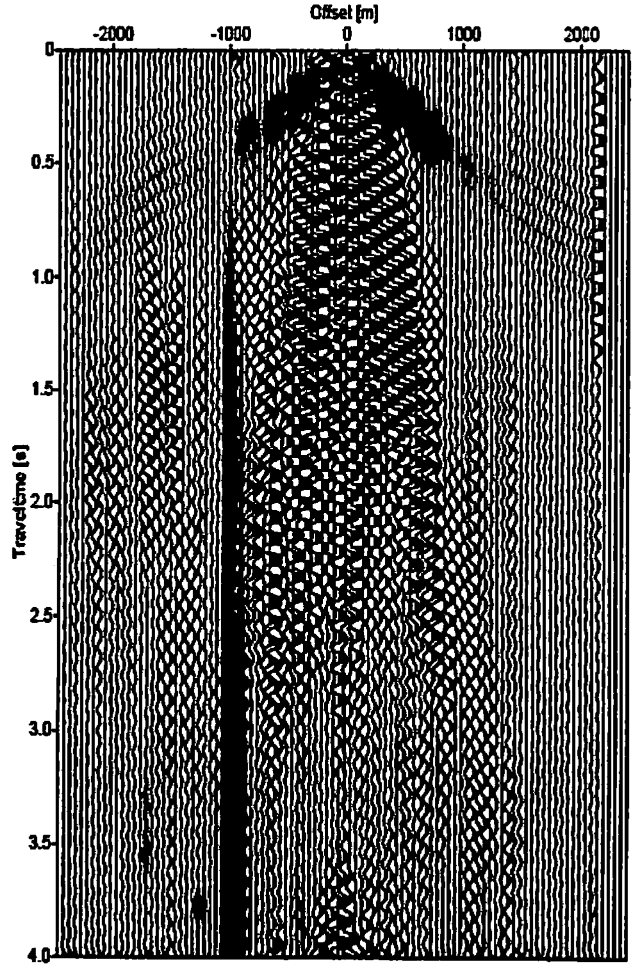


Figure 5.13: *Rejected data. Applying the inverse filter shows that not only noise was rejected: reflections between 1.5s and 2.0s were partly filtered.*

5.2.2 f-k Filtering

Filtering in the frequency-wave number (f-k) domain is also known as dip or velocity filtering: All seismic energy that originates from the same source with the same propagation velocity belongs to one dipping event in the time domain pre-stack gathers. The representation of such a dipping event in the f-k domain is a straight line through the origin- assuming the event contains the full frequency spectrum. Filtering in the f-k domain allows therefore to filter out certain dips and hence, velocities. A 2D Fourier transform is necessary to transform data into the f-k domain. The first Fourier transform converts the time-axis into the frequency-axis; the second transform them converts the x-axis into wavenumber. The wavelength λ of the spatial wave gives the wavenumber k through $k=1/\lambda$. Just as frequency (cycles/second) is the inverse of period, wavenumber ($meter^{-1}$ or cycles/meter) is the inverse of wavelength. Figures 5.14 and 5.15 explain 2D Fourier transforms for two dipping events. The higher frequency of the second x-t representation in each figure was chosen to be twice the lower frequency. The shown frequencies are only two of many outputs from the first Fourier transform. The spatial wave along the x-axis illustrated how the x-axis is converted to wavenumber: trace amplitudes in the x-t domain at an arbitrary time (dashed and dotted line) are taken to interpolate a spatial sine wave through them. As it can be seen in figures 5.14 and 5.15, slopes of steep dipping (slow) events in the time domain are low in the f-k domain; slopes of flat events (fast) in the time domain are high. Dips in different directions are distinct by the sign of the wavenumber. For the filter tests presented in figure 5.16 to 5.20, the filter slopes were chosen as – 0.35, -0.25, 0.25, 0.35. The ground roll is removed effectively by this filter as it can be observed in figure 5.18. The f-k spectrum after filtering (figure 5.19) shows the missing part in the spectrum: flat slopes at low frequencies were filtered out (figure 5.20). This represents the energy from surface waves – low frequency at low propagation speed.

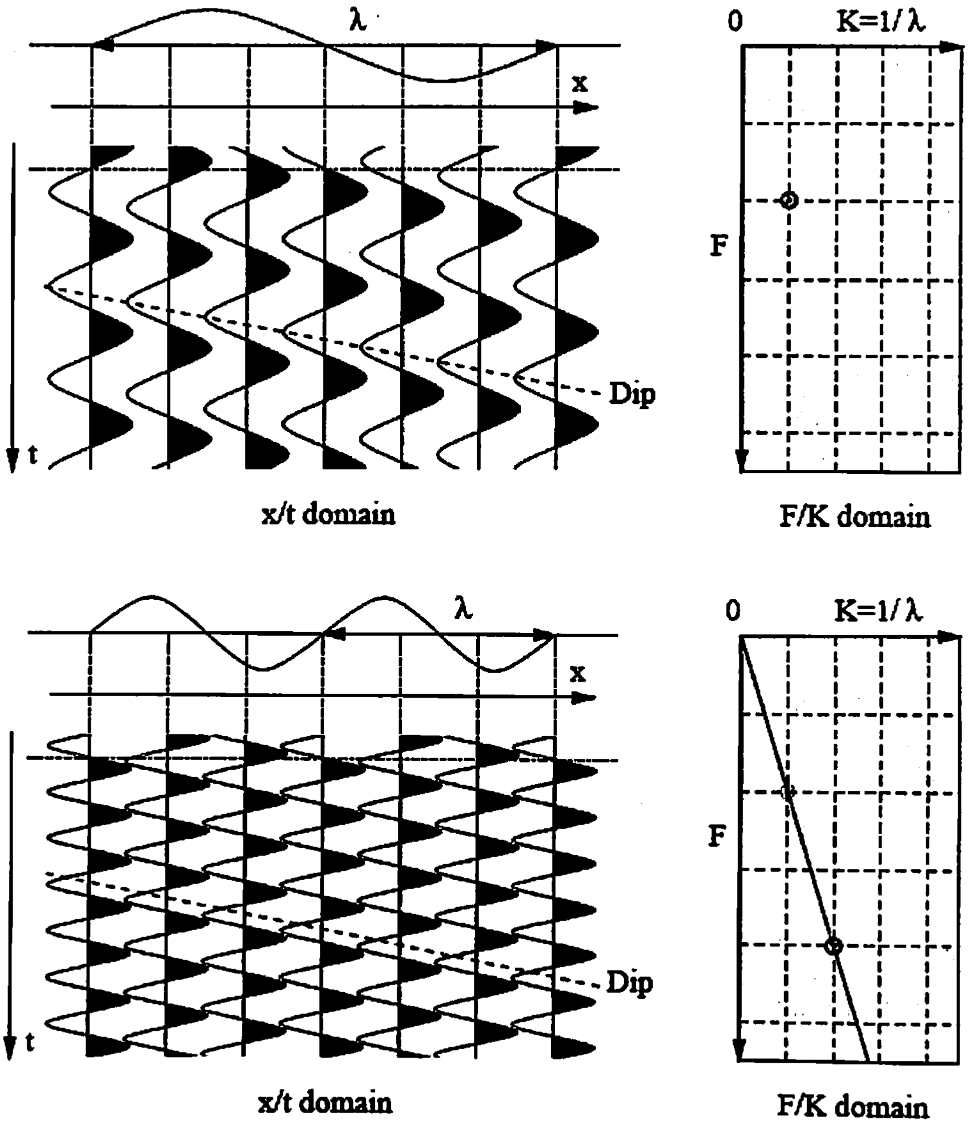


Figure 5.14: *2D Fourier transform of a dipping event. The first Fourier transform decomposes the original seismic traces in sine waves of different temporal frequencies (Chapter 3). Two specific frequencies of the frequency spectrum are shown in this figure. A second Fourier transform is used to create a spectrum of spatial sine waves. An f-k plot shows the amplitudes of these waves over frequency and wavenumber. In this example the amplitude is one; the wavenumber for the second frequency is twice the wavenumber for the first one.*

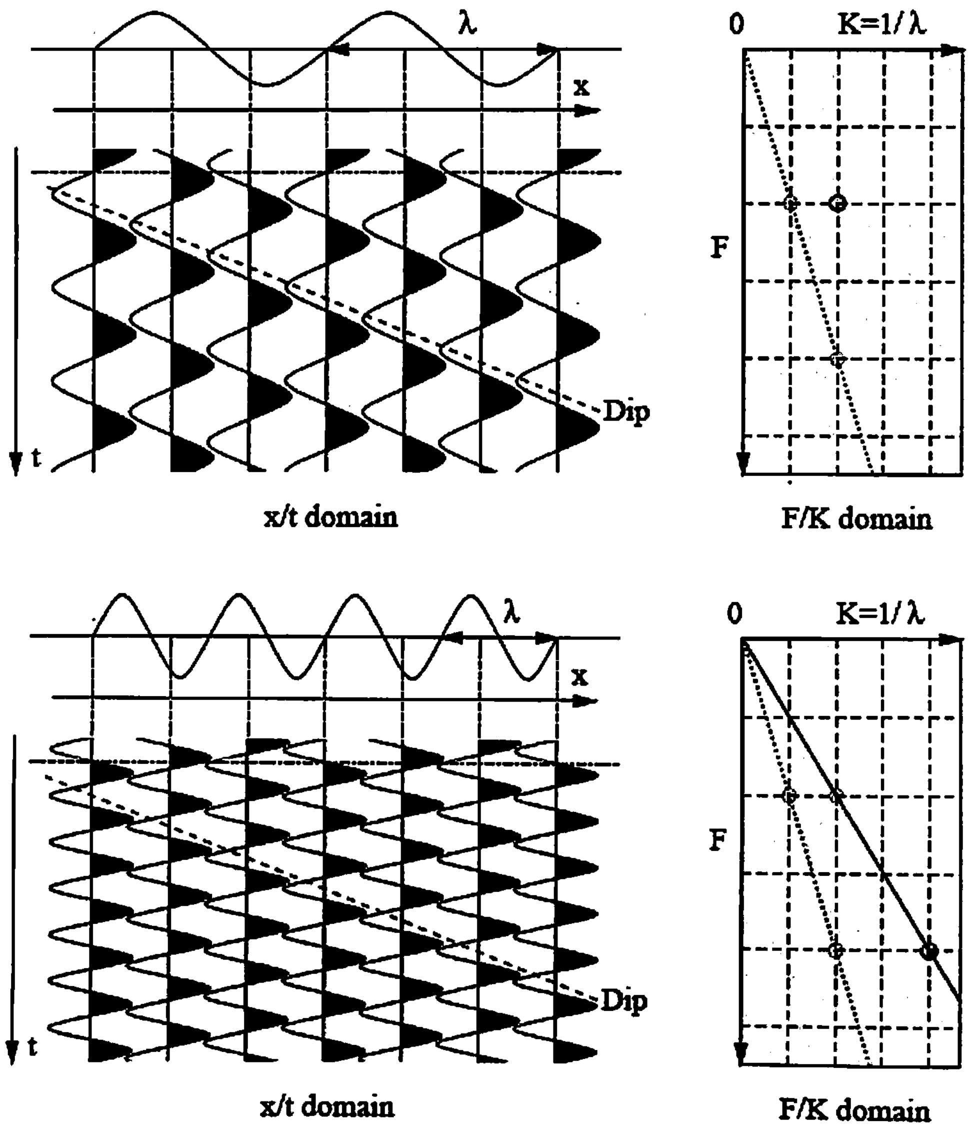


Figure 5.15: *2D Fourier transform for a slower, steeper-dipping event than seen in Figure 5.14. Result is a smaller wavelength of the spatial wave for the same frequencies as analyzed in Figure 5.14. The f-k representation of the steeper dipping event is a line with a smaller slope than seen in Figure 5.14.*

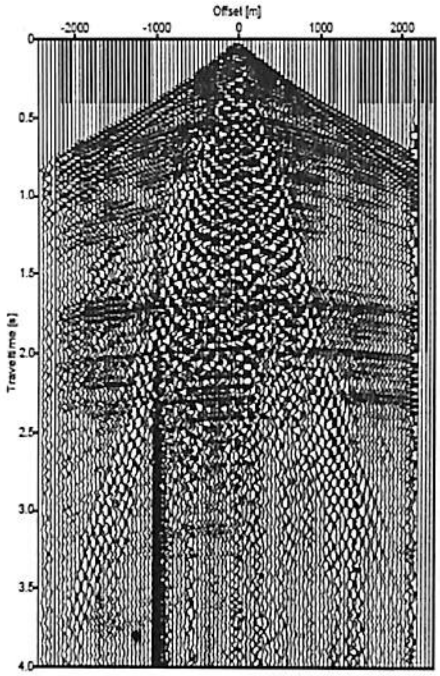


Figure 5.16: *Shot gather [YILMAZ 25] before filtering.*

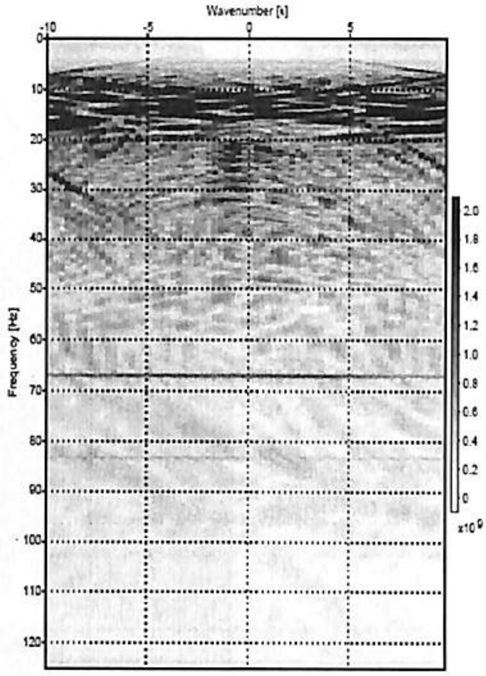


Figure 5.17: *f-k Spectrum before filtering.*

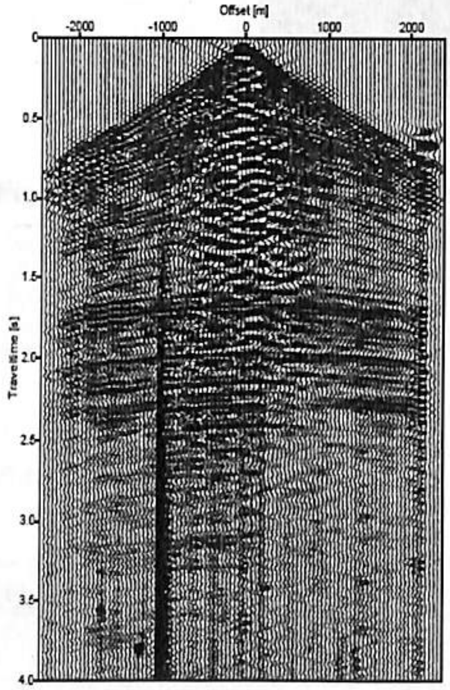


Figure 5.18: *Shot gather [YILMAZ 25] after f-k filtering (-0.35, -0.25, 0.25, 0.35 s/km).*

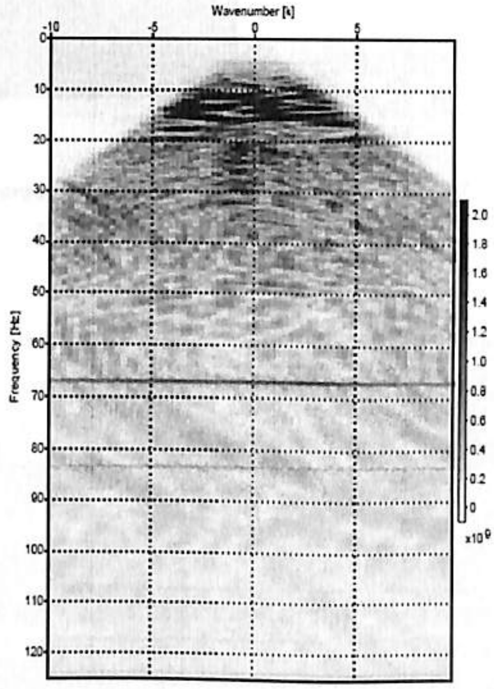


Figure 5.19: *f-k Spectrum after filtering (-0.35, -0.25, 0.25, 0.35 s/km).*

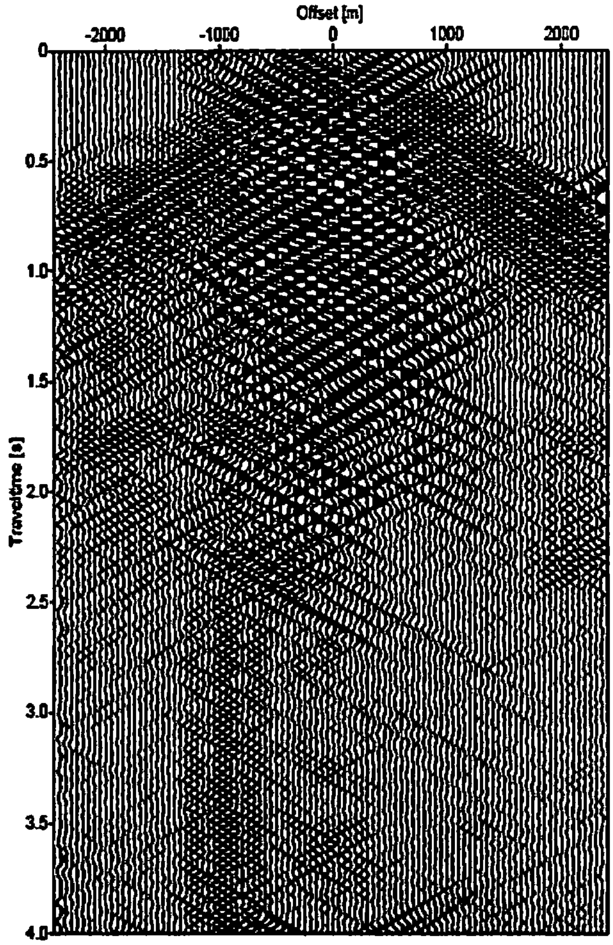


Figure 5.20: *Rejected data. Applying the inverse filter shows the dips of rejected events.*

5.2.3 De-convolution

De-convolution, or abridged “decon” , is the inverse process to convolution. When the convolution earth model was discussed earlier in Chapter 3, it was the reflectivity series. A de-convolution that uses a know input wavelet is called deterministic de-convolution. Deterministic de-convolution can transform the source wavelet (if known) back into a spike, but it can not correct for any changes in the shape of the source wavelet. The source wavelet, however, always changes during propagation through the earth- the resulting wavelet is called a system wavelet. Statistical de-convolution, also known as Wiener predictive error filtering, estimates this system wavelet from an autocorrelation analysis of seismic input trace. Statistical de-convolution needs therefore no known or assumed input wavelet and can be used on any seismic data.

Two types of statistical de-convolution are popular: spiking de-convolution and predictive de-convolution, where the first is a special case of the second. To understand the differences between these two de-convolution modes, the objectives and input variables of statistical de-convolution have to be clarified:

- The objective of statistical de-convolution are to enhance the resolution and to remove noise, in the form of multiples, from seismic data. Higher resolution is obtained by transforming the system wavelet back to an impulse that contains higher frequencies. Transforming the system wavelet to an impulse, and hence the seismic trace to the reflectivity series, also removes reverberations or multiples. Multiples occur when a wave is reflected at the same reflector twice or more often after bouncing back from a reflector above. Strong multiples are observed in marine data, where waves easily bounce back from the water surface.
- Required input for statistical de-convolution is the system wavelet, which is estimated from an autocorrelation (figure 5.21). The gap length should be measured from the beginning of the autocorrelation to the second zero crossing; this gap length approximates the length of the system wavelet. The operator length should be chosen in a way that the first reverberated system wavelet is included in the operator. This makes an effective attenuation of multiples possible.

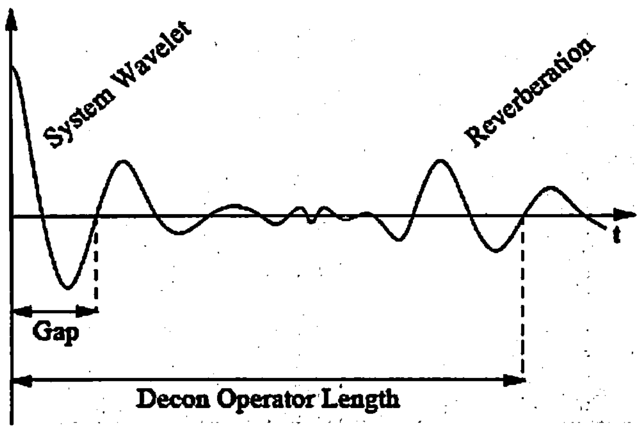


Figure 5.21: *Autocorrelation Analysis for Deconvolution Parameters.*

Spiking de-convolution is predictive de-convolution with a default gap length equal the sampling interval. Spiking de-convolution maximizes the resolution (frequency content) but is less effective than a well-designed predictive de-convolution in attenuating multiples. Spiking de-convolution is also more sensitive to the presence of noise in data than predictive de-convolution.

Predictive de-convolution was used for the plots shown in figure 5.22 to 5.27. The original input shot gather had a gain function, a NMO correction with constant velocity, and a front mute applied to it before de-convolution. The preprocessed shot gather (Figure 5.22) shows strong multiple systems in intervals of about 0.5 seconds. After applying a NMO correction with one constant velocity, these multiples occurs as flat events over the entire time range. The autocorrelation and the frequency spectrum of this shot gather are shown in figures 5.24 and 5.26 respectively. Using the 2^{nd} zero crossing in the autocorrelation, a gap length of 0.028 seconds was chosen. The operator length used is 0.500 seconds to address the strong multiple systems. The shot gather after deconvolution (figure 5.23) still shows multiple systems, although they were attenuated. The resolution of the shot gather was improved; the frequency content was increased (figure 5.27). The autocorrelation after deconvolution (figure 5.25) shows that the deconvolution operator improved all but four traces. Looking at figure 5.24, these inconsistent traces can also be spotted before deconvolution.

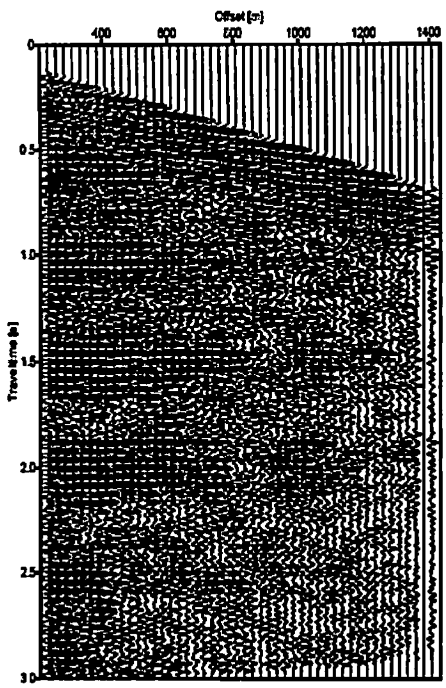


Figure 5.22: *Shot gather [YILMAZ 16] before deconvolution.*

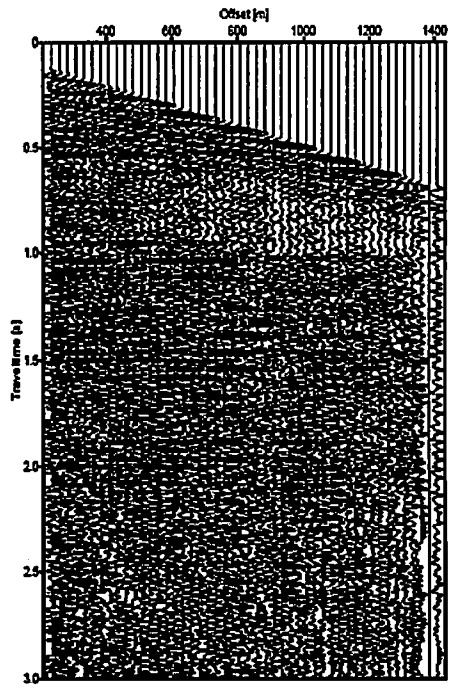


Figure 5.23: *Shot gather [YILMAZ 16] after deconvolution (Gap Length 24ms, OperatorLength 500ms).*

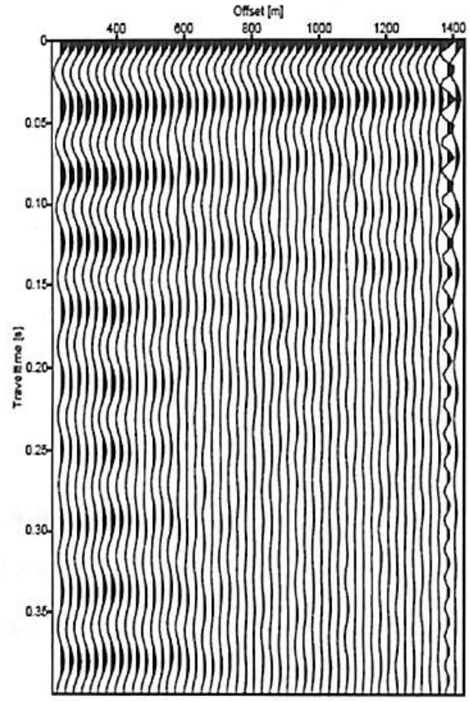


Figure 5.24: *Autocorrelation analysis before deconvolution.*

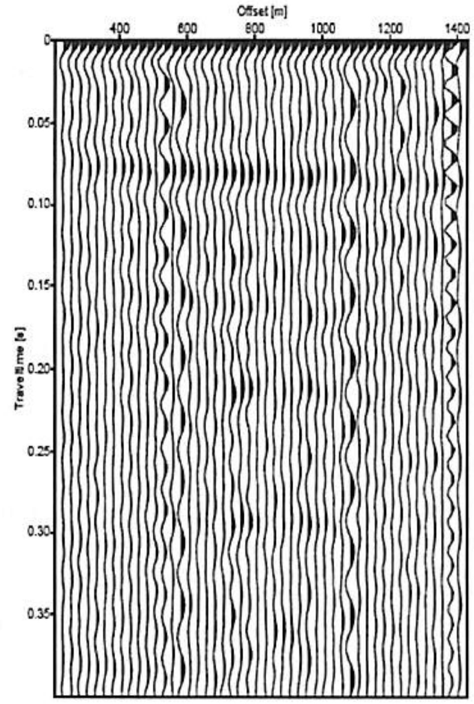


Figure 5.25: *Autocorrelation analysis after deconvolution.*

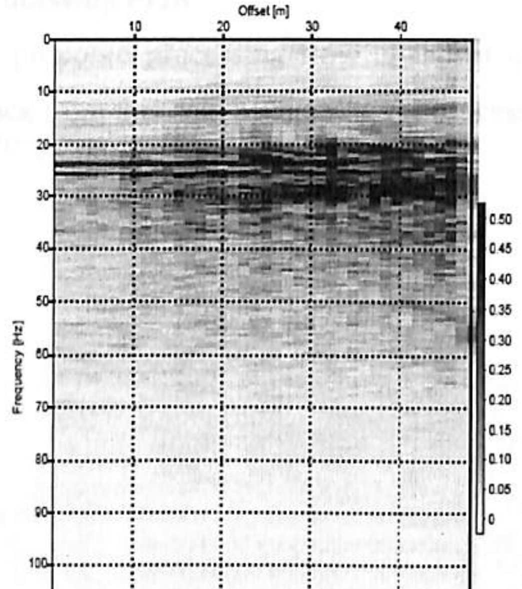


Figure 5.26: *f-x Spectrum before deconvolution.*

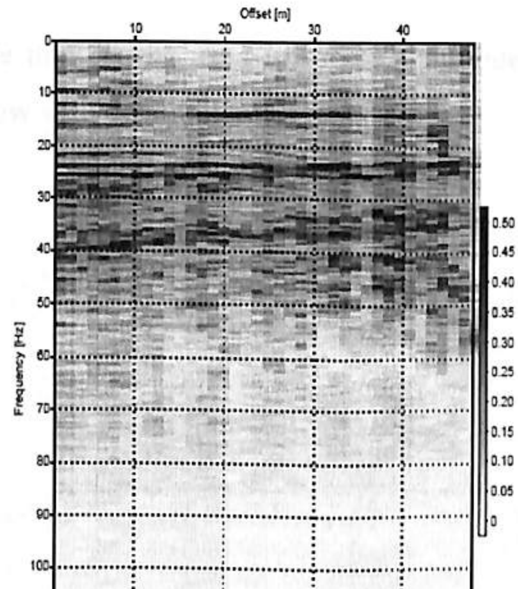


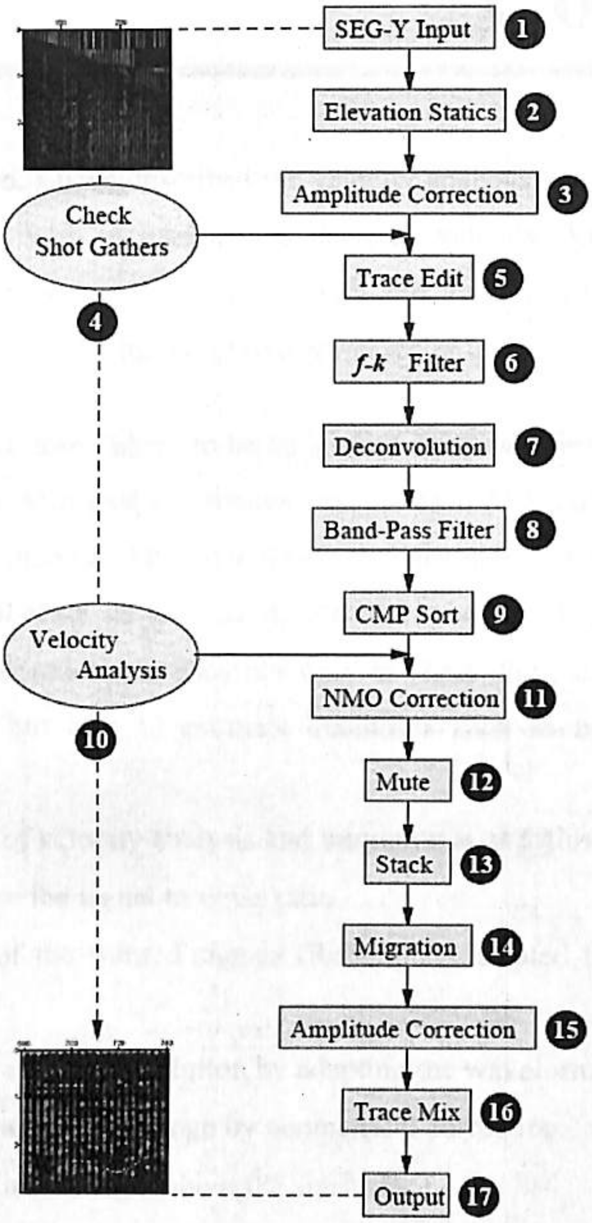
Figure 5.27: *f-x Spectrum after deconvolution.*

5.2.4 Edits, Mutes and De-spike

Bad seismic traces can affect processing routines that rely on input of more than one trace at a time (multi channel processing). An example for this can be seen in figure 5.18, where one bad trace (offset 2150m) affects other traces during f-k filtering. In order to avoid effects like this, bad traces should be edited before any processing. Editing seismic traces usually means "killing" them, or setting all amplitude values on the traces to zero, although, editing seismic traces can also imply reversing polarity (scale by -1), or muting only a part of it. Mutes are mainly used for first arrivals and NMO stretches. Bad traces due to spikes do not necessarily have to be killed. Processing tools such as de-spike can be used to remove spikes and make them suitable for further (multi channel) processing.

Processing Flow

A proposed processing flow is shown in figure that can be used to create a migrated stack from this data by running the processing flow with default parameter.



Proposed Processing Flow

Conclusion

In this dissertation, I have described the velocity analysis and migration with the help of Promax software, with increasing and decrease velocity. Velocity picking is done to enhancing or suppressing the observed high and low velocity zones. The increase and decrease in velocity make the event overcorrected or under corrected accordingly.

Migration is much more likely to be an intermediate step, feeding information into other seismic processes. Migrated amplitudes are used for AVO analysis and for other forms of seismic attribute analysis. Migrated structural information is used in coherency analysis to delineate small-scale as well as large-scale structure. The velocities obtained from prestack depth migration are used not only to check the plausibility of the underlying geologic model, but also to estimate quantities such as hydrostatic pressure in the subsurface.

The main benefit of velocity analysis and migration is as follows :

- To improve the signal to noise ratio
- Isolation of the wanted signals (Reflections isolated from multiples and surface waves)
- To obtain a higher resolution by adapting the waveform of the signals
- To obtain a realistic image by geometrical correction
- To obtain information about the subsurface

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