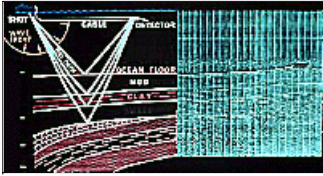


Seismic reflection notes

Introduction



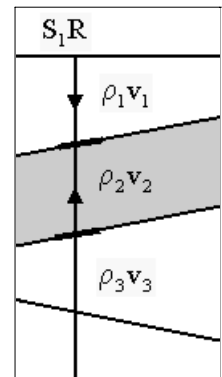
This figure was modified from one on the [Lithoprobe slide set](#). LITHOPROBE is "probing" the "litho" sphere of our continent, the ground we are living on, the basis of our ecosphere. How earth scientists are investigating the unseen depths of the continent, and the often startling discoveries they have made, is described in 80 slides and 35 pages text. [Go to the homepage of Lithoprobe for many links describing this nation-wide project.](#)

NOTE: While the active components of this resource (quiz questions etc) are being written, consider using the summary work sheet to help you take notes and to encourage you to think about what is being learned in this package.

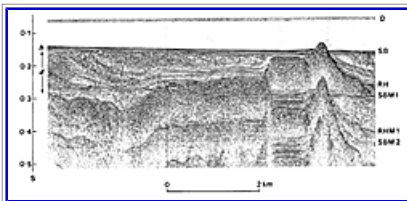
The basic elements of seismology were covered in an earlier course. There we focused upon seismic refraction. Head waves propagated along an interface separating media of different velocities and were recorded as first arrivals at the surface. These arrivals were then analysed to yield information about the velocities and thicknesses of the layers.

In reflection seismology we record seismic pulses that are reflected from boundaries which separate layers that have different acoustic impedances. The acoustic impedance is the product of the velocity and density. Information in the seismogram which comes after the first arrival is important. Reflection seismic data are acquired in the same manner as refraction data but the processing is considerably different. In reflection seismology, seismic records from many sets of shots and receivers are used to generate an ideal seismogram which has reflections that correspond to a vertically travelling wave as shown in the diagram below. The reflections occur at travel times that are determined by the velocity and the length of the travel path in each layer. These seismograms are ideal for interpretation.

The ideal seismograms are acquired at regular distances along the surface and are composited into a seismic section. This generates an image of the substructure that can be used in an identical manner to a radar section. The examples below illustrate how the seismic images can be interpreted in terms of geologic structure

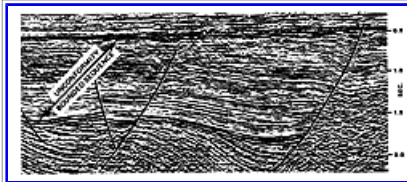


Two Examples



Air gun record from the Gulf of Patras, Greece, showing Holocene hemipelagic (h) and deltaic (d) sediments overlying an irregular erosion surface (rockhead, RH) cut into tectonized Mesozoic and Tertiary rocks of the Hellenide (Alpine) orogenic belt. SB: sea bed reflection; SBM1 and SBM2: first and second multiples of sea bed reflection; RHM1: first multiple of rockhead reflection.

From Kearey, Philip and Michael Brooks, [An Introduction to Geophysical Exploration](#). 2nd ed. Blackwell Science: 1991.



A seismic section from the northern Amadeus basin, central Australia, illustrating a depositional sequence bounded by major unconformities.

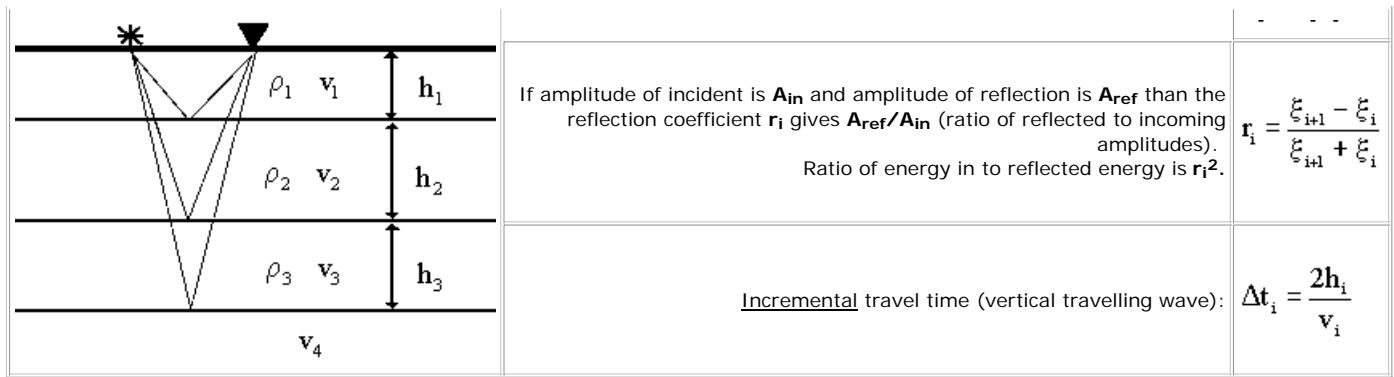
From Kearey, Philip and Michael Brooks, [An Introduction to Geophysical Exploration](#). 2nd ed. Blackwell Science: 1991.

In order to generate the previous images there are numerous operations that need to be applied to the data. Much of the data processing is tied to the hypothesis that the earth's properties vary most strongly in the vertical direction. As background we first review the principles of a reflection seismogram and then consider travel time curves that arise from a horizontally layered earth.

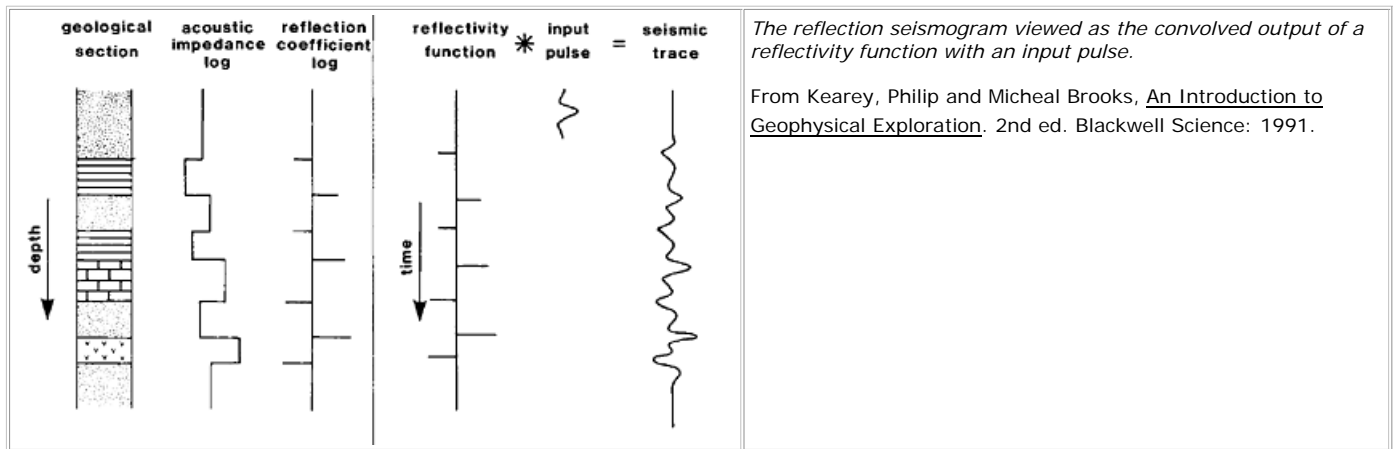
An important concept not covered in these web-notes is the "Optimum-offset" method of seismic reflection data acquisition.

Normal Incidence Reflection Seismogram

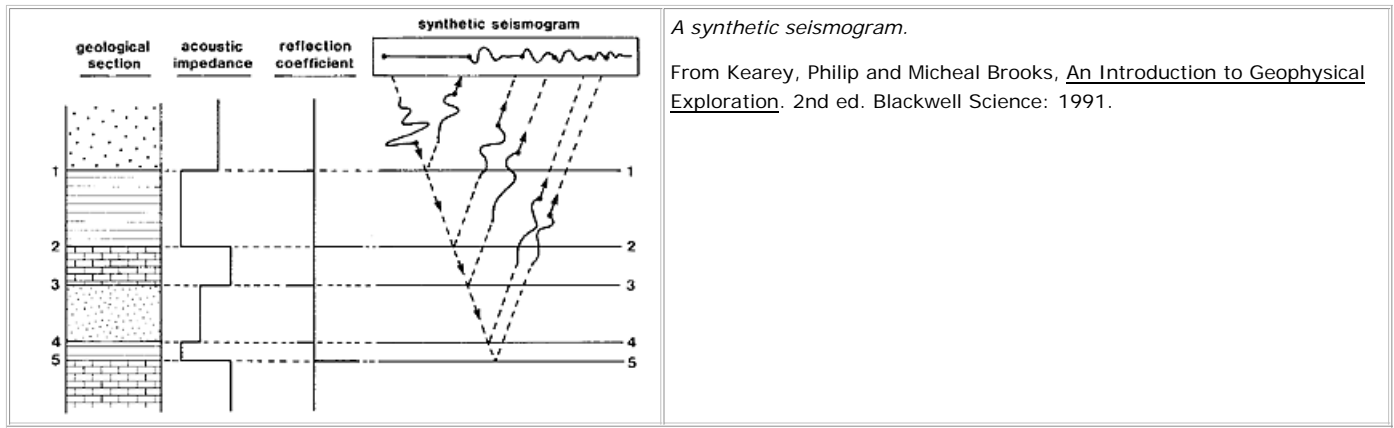
The principles of the normal incidence reflection seismogram are illustrated in the diagrams below. A source and receiver are at the surface of a layered earth whose properties are variable. The reflection and transmission coefficients depend upon the change in acoustic impedance, and thus on both density and velocity. The travel time for the wave to go from the source to the reflecting interface and back to the surface depends only upon the length of the travel path and the velocity of each layer. The travel time formula given below, is for a wave which travels vertically and for which the source and receiver are coincident. (The source and receiver are offset slightly in the diagram for visual clarity). This produces the "normal incidence" seismogram.



The normal incidence seismic trace is obtained by the convolution of a seismic wavelet (input pulse) with the reflectivity function. The amplitude of each spike on the reflectivity function is equal to the value of the reflection coefficient that corresponds to a particular boundary. (In reality this value is further altered by the transmission coefficients). The times for each reflection event are obtained by knowing the layer thickness and velocities. Each impulse on the reflection function generates a scaled replication of the seismic wavelet. The composite of all of the reflection events generates the seismic trace.



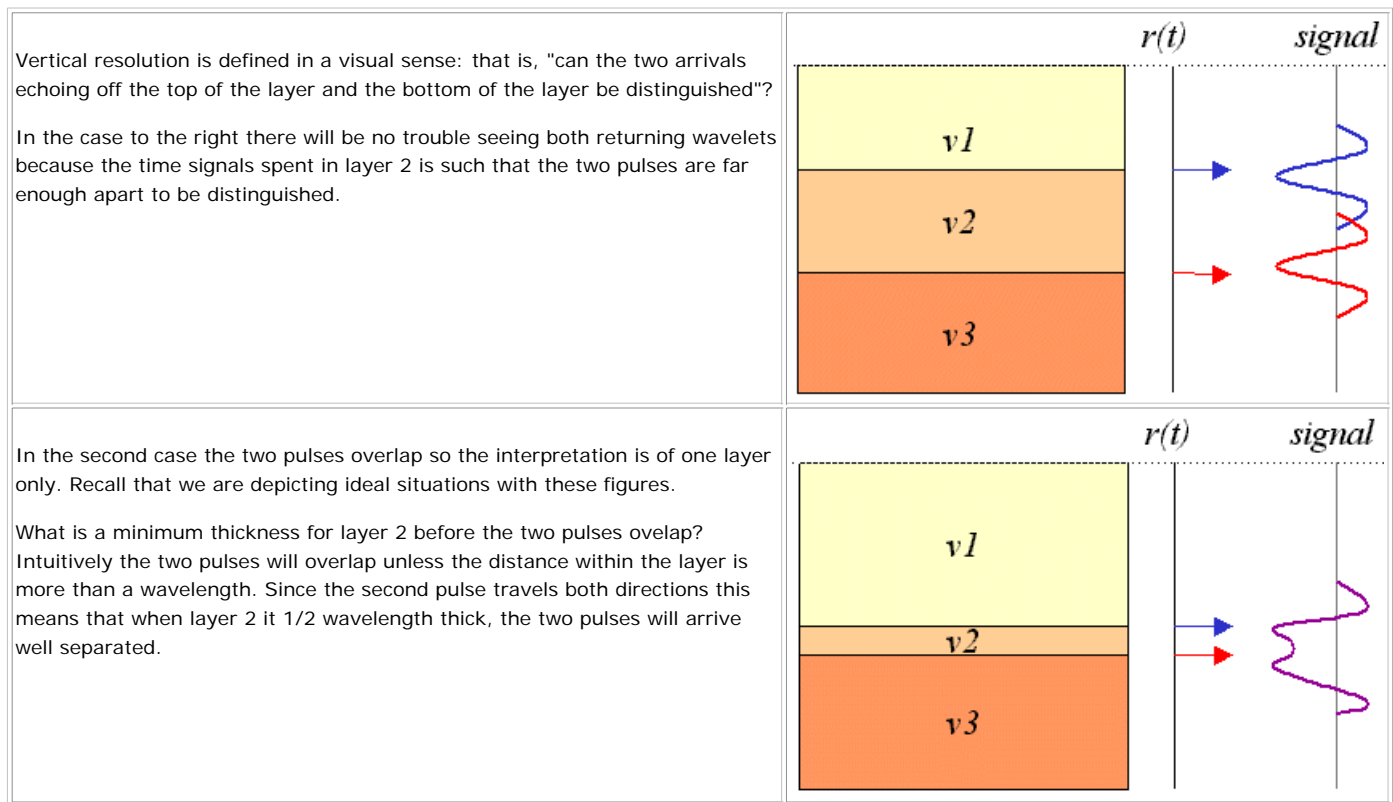
Notice how the negative reflection coefficients change the polarity of the signal recorded at the receiver.



Vertical resolution

The vertical resolution of a seismic data sets describes how thin a layer can be before the reflections from its top and it's bottom become indistinguishable. This depends upon the shape of the seismic source wavelet. The shape, or more particularly the signal wavelength depends upon the frequency and the velocity of the materials. Recall that *wavelength = velocity * time*.





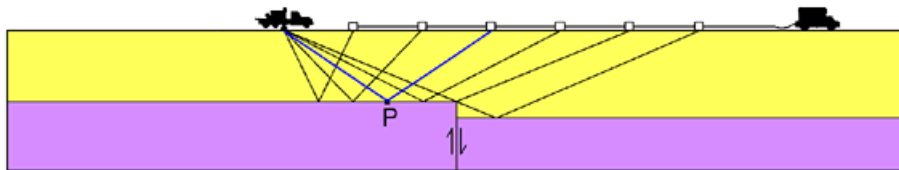
In fact, the *theoretical* minimum thickness is 1/4wavelength. In other words, careful processing can extract the pulses if they overlap by half a wavelength, but this theoretical minimum is never really achieved in practice.

How does this thickness in terms of wavelength translate into real distances? As noted above, wavelength depends upon frequency of the signal and on velocity of the medium. Therefore the vertical resolution of a survey depends upon both the signals used and the materials themselves.

Finally, recall that as energy propagates into the ground, higher frequencies attenuate faster than lower frequencies. Since seismic signals consist of a range of frequencies (i.e. they have a broad spectrum), their spectral character will change as they travel. The result is that resolution of a survey is poorer at greater depths.

Seismic reflection surveys

Seismic data are routinely acquired using a source and multiple receivers. The entire array is progressively moved along.



There are several common geometries for the "instrument" - that is the configuration of sources and receivers.

1. The simplest is a "common offset" array in which the source and receiver distance is always the same. This is the way most GPR surveys are performed. Signals at one receiver are recorded from the shot (or source), then the shot-receiver are moved to a new location and the process is repeated.
2. A variation on this is the "optimum offset" array in which many receivers are recorded for each shot. Then the shot and receiver string are moved to a new location and the process is repeated. This works well if targets are relatively shallow (perhaps less than 100m or so) AND if reflecting horizons are clear and distinct (ie if the acoustic properties vary significantly and sharply across the boundaries).
3. The most common, though most expensive, form of surveying is the "multichannel reflection survey". As for optimum offset surveys, many receivers are used for each shot. The difference is that the survey system is carefully designed so that each reflecting point in the subsurface is sampled more than once. In other words, the objective is to obtain several different echoes (reflections) from identical subsurface points. This type of surveying involves some care in setting up the field work, and some effort in the processing steps. Details will be covered next.

Multichannel Reflection Survey

First consider the source-receiver geometry. The geometry can be "split spread" in which case there is a central shot with receivers on both sides, or a "single-ended spread" in which the receivers are always on one side of the source. Split spreads are common in land surveys; single-ended spreads are common in marine surveys.

Shot-detector configurations used in multichannel seismic reflection profiling.
 (a) Split spread, or straddle spread. (b) Single-ended spread.

Click for larger image

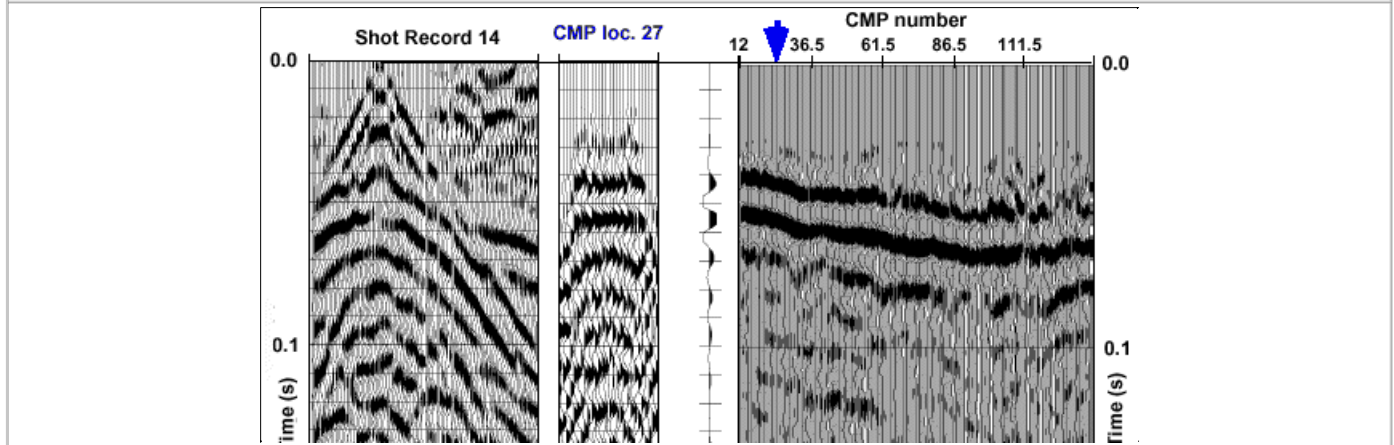
Both from Kearey, Philip and Micheal Brooks, *An Introduction to Geophysical Exploration*. 2nd ed. Blackwell Science: 1991.

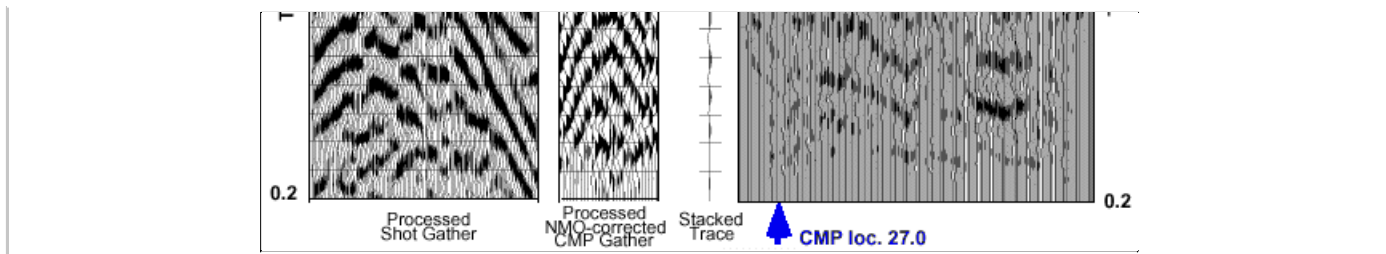
A split spread seismic record is shown above right. The seismic traces all belong to a single source and hence this is referred to as a "Common Source Gather". The first arrivals are direct or critically refracted arrivals. Reflection hyperbolae from numerous boundaries are observed. The strong energy in the triangular central portion is ground roll caused by surface waves. It masks the reflection events.

Fundamental procedure

In order to benefit from gathering several echoes from each reflecting point there are numerous operations that need to be applied to the data. Much of the data processing is tied to the hypothesis that the earth's properties vary most strongly in the vertical direction. The table shown next illustrates the fundamental procedural concept underlying the creation of a final seismic reflection section:

<p>Objective: We want to characterize the earth using echo sounding with this geometry:</p>	<p>(1)</p>
<p>Reason for using many "redundant" echoes - to reduce noise: We need to gather several versions of the experiment and stack:</p>	<p>(2)</p>
<p>Logistics: However, surveying with one shot and many geophones is more cost-effective:</p>	<p>(3)</p>
<p>Therefore: Field work must be arranged as follows. <i>Blue italic text refers to the figure below.</i></p> <ol style="list-style-type: none"> 1. We gather data using the geometry of type (3). <i>Data from one shot into many geophones ("common shot data") are shown below under the label Shot Record 14.</i> 2. Next, sort many of these "common shot data" so that traces appear as if gathered using the geometry of type (2). <i>All traces that reflected under one location are collected into a "common mid point gather", one from each of many common shot data sets. See the panel under CMP loc. 27.</i> 3. Stack these traces to produce one trace which represents measurements obtained using the desired geometry of (1). This is the CMP (common mid point) trace. <i>This is the single trace next to the CMP panel.</i> 4. Then many of these CMP traces are combined into one cross section of the earth's structure. <i>Traces are labelled CMP number, and the one trace shown is identified with arrows.</i> 5. Interpretation in terms of geology is the final step. 	



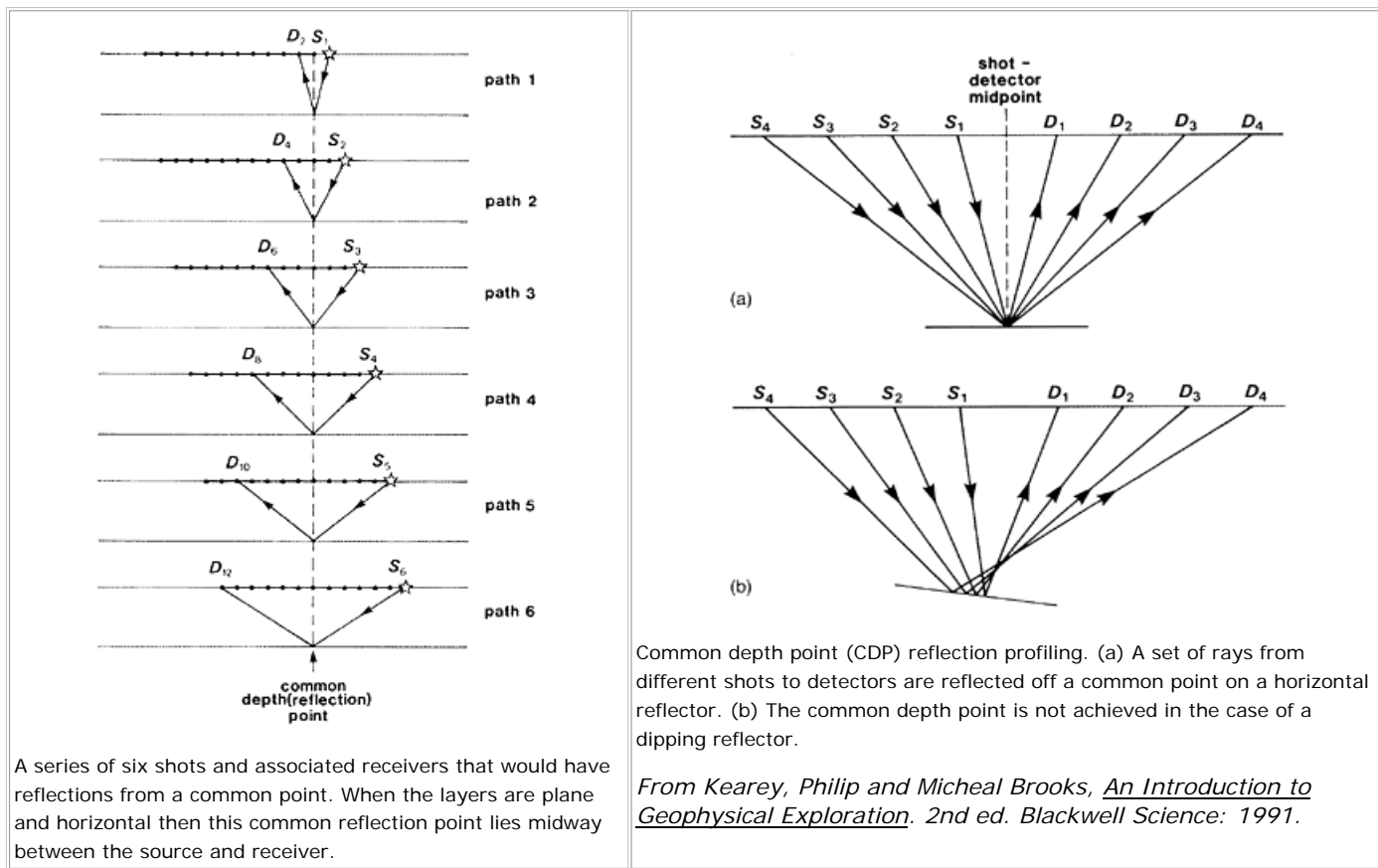


The procedures to accomplish these steps will be explained in subsequent pages of the notes.

Common Reflection Points

Multiple shots and receivers are used in a reflection seismic profiling specifically so that some subsurface points are sampled more than once. Ultimately the goal is to identify all the reflections due to that point on the various seismograms and stack these to get an enhanced signal to noise ratio. The idea is illustrated below left.

The collection of seismic traces that correspond to a particular midpoint is called a Common Midpoint (CMP) gather. In older literature, this collection of traces was referred to as a Common Depth Point (CDP) gather. That is not strictly correct as the diagram below right illustrates.

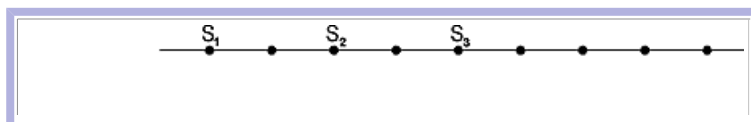


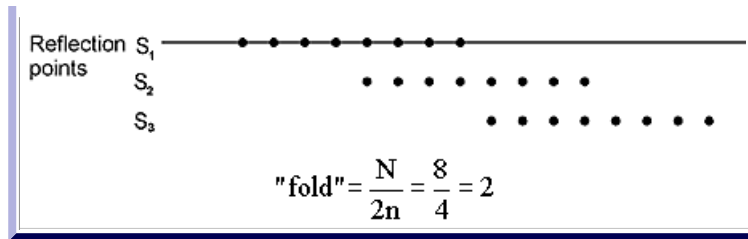
"Fold" in Seismic Reflection Surveying

The *fold* refers to the number of times a particular subsurface point has been sampled. It is equal to the number of traces in the CMP gather and is numerically evaluated by,

$$\text{"fold"} = \frac{N (\text{\# of geophones})}{2n}$$

where n is the moveup rate in units of geophone spacing. For example, the schematic below shows a single ended spread with 8 geophones and moveup rate of n=2.





We see that each point in the subsurface is sampled only twice. Notice that the distance between the reflection points in the subsurface is half the geophone spacing.

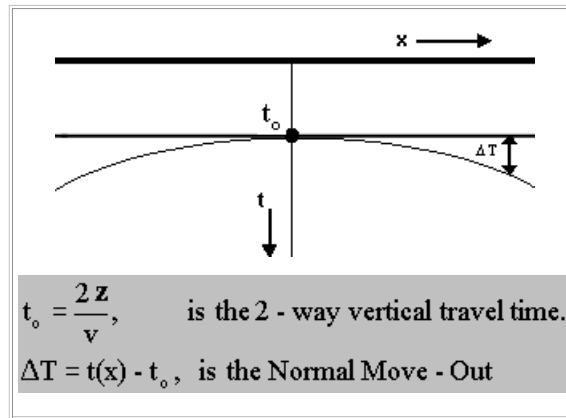
Travel Time Curves for a Single Layer

In the diagram to the right a source S and a set of receivers lie on the surface of the earth. The earth is a single uniform layer overlying a uniform halfspace. A reflection from the interface will occur if there is a change in the acoustic impedance at the boundary.

Let x denote the "offset" or distance from the source to the receiver. The time taken for the seismic energy to travel from the source to the receiver is given by

$$t(x) = \frac{(x^2 + 4z^2)^{1/2}}{v}$$

This is the equation of a hyperbola. In seismic reflection (as in radar) we plot time on the negative vertical axis, and so the seismic section (without the source wavelet) would look like.



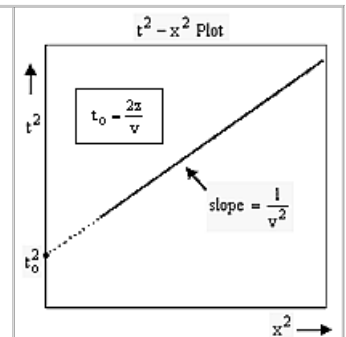
In the above diagram t_0 is the 2-way vertical travel time. It is the minimum time at which a reflection will be recorded. The additional time taken for a signal to reach a receiver at offset x is called the "Normal Moveout" time, ΔT . This value is required for every trace in the common depth point data set in order to shift echoes up so they align for stacking. How is it obtained? First let us find a way of determining velocity and t_0 .

For this simple earth structure the velocity and layer thickness can readily be obtained from the hyperbola. Squaring both sides yields,

$$t^2(x) = t_0^2 + \frac{x^2}{v^2}$$

This is the equation of a straight line when t^2 is plotted against x^2 .

Now, to find ΔT , we must rearrange this hyperbolic equation relating t_0 , x the Tx-Rx offset, t at x or t(x), and the ground's velocity V.



The sequence of algebra shown to the right has only one complicated step - a binomial expansion must be applied to obtain a simple relation without square roots etc.

The approximation is valid so long as the source-receiver separation (or offset) is "small" which means much less than the vertical depth to the reflecting layer (i.e. $x \ll Vt_0$). The result is a simple expression for normal moveout.

Each echo can be shifted up to align with the t_0 position, so long as the trace

$$t^2 = \frac{x^2 + 4z^2}{V^2} = \frac{4z^2}{V^2} \left(\frac{x^2}{4z^2} + 1 \right)$$

$$t = \left(\frac{2z}{V} \right) \sqrt{1 + \left(\frac{x}{2z} \right)^2} = t_0 \sqrt{1 + \left(\frac{x}{Vt_0} \right)^2}$$

apply binomial expansion to get

position (x), the vertical incident travel time (t_0) and the velocity are known. Velocity can be estimated using the slope of the $t^2 - x^2$ plot, or with several other methods, which we will discuss in pages following.

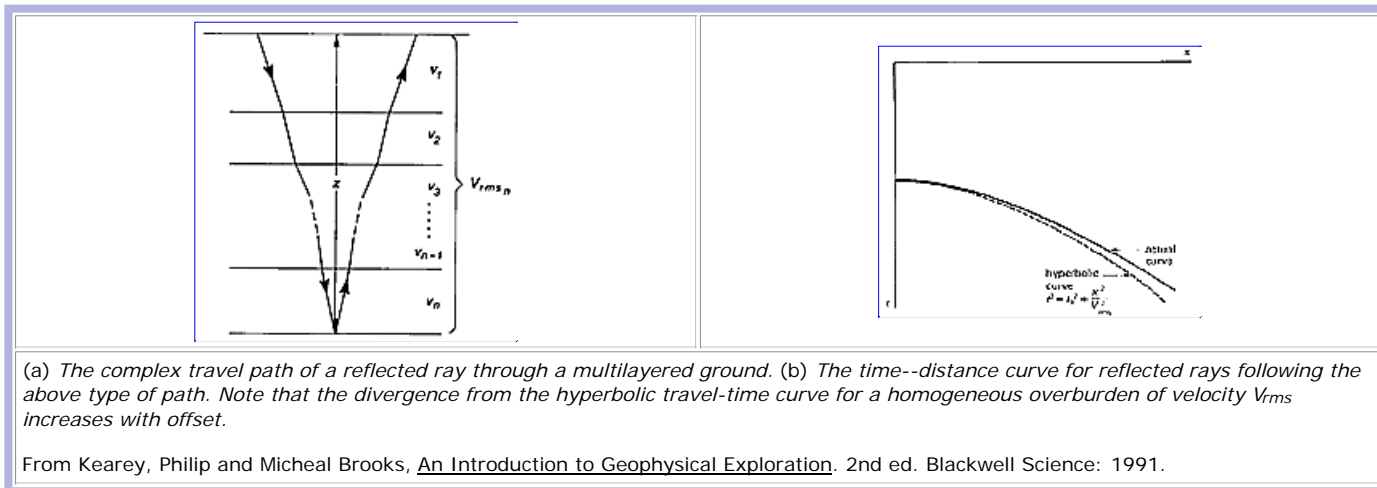
$$t \approx t_0 \left(1 + \frac{1}{2} \left(\frac{x}{Vt_0} \right)^2 \right) \quad \text{if} \quad \frac{x}{Vt_0} \ll 1$$

Now, since the normal moveout is $\Delta T = t_x - t_0$

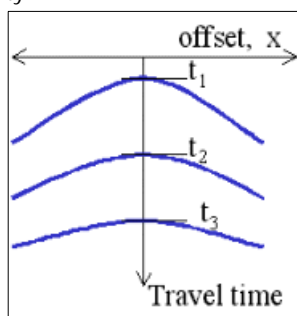
$$\Delta T \approx \left(t_0 + \frac{t_0 x^2}{2V^2 t_0^2} \right) - t_0 \approx \frac{x^2}{2V^2 t_0}$$

Travel Time Curves for Multiple Layers

If there are additional layers then the seismic energy at each interface is refracted according to Snell's Law. The energy no longer travels in a straight line and hence the travel times are affected. It is observed that for small offsets, the travel time curve is still approximately hyperbolic, but the velocity, which controls the shape of the curve, is an "average" velocity determined from the velocities of all the layers above the reflector. The velocity is called the RMS (Root Mean Square) velocity, V_{rms} .



As outlined in the figure above, the reflection curve for small offsets is still like a hyperbola, but the associated velocity is V_{rms} , not a true interval velocity.



For each hyperbola,
$$t_n \approx \frac{(x^2 + 4Z_n^2)^{1/2}}{V_n^{rms}}$$

By fitting hyperbolas to each reflection event one can obtain (t_n, V_n^{rms}) for $n = 1, 2, \dots$. The interval velocity and layer thickness of each layer can be found using the formula below.

$$v_n \approx \left[\frac{(V_n^{rms})^2 t_n - (V_{n-1}^{rms})^2 t_{n-1}}{t_n - t_{n-1}} \right]^{1/2}$$

$$\Delta Z = z_n - z_{n-1} = \left(\frac{t_n - t_{n-1}}{2} \right) v_n$$

These formulae for the interval velocity and thickness of the n^{th} layer are directly obtainable from the definition of V_n^{rms} given above. The RMS velocity for the n^{th} layer is given by,

$$V_n^{RMS^2} = \left[\frac{\sum_{i=1}^n v_i^2 \tau_i}{\sum_i \tau_i} \right]$$

$$\sum_{i=1}^n \dots$$

where v_i is the velocity of the i^{th} layer, and τ_i is the one-way travel time through the i^{th} layer.

"Stacking" Data in CMP Gathers

Imagine recording four seismic traces from one source (top panel, right). If we plot the travel time for a seismic signal as a function of distance between receiver and source we see that time increases (middle panel). The curve through the traces forms a hyperbola (details are in a subsequent page).

For horizontal flat surfaces, the change in travel time for a set of increasing source-receiver spacings of a CMP gather (bottom panel) will be identical to the "common shot gather" (top panel). The travel time curve from the reflector will appear approximately as a hyperbola. Unlike for the common shot gather, in the CMP gather all of the arrivals correspond to the same reflection point.

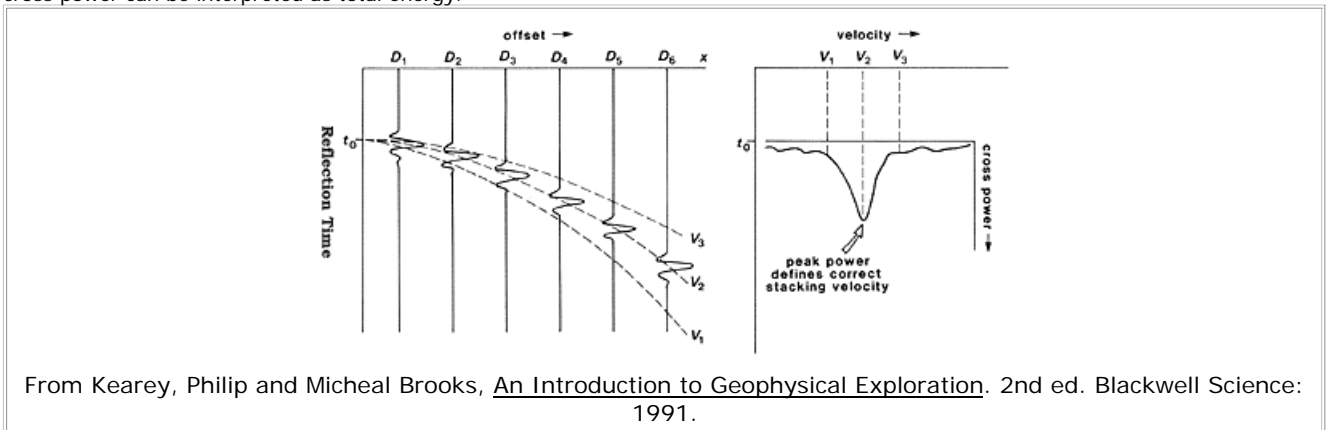
The hyperbolic representation for the travel time curve is exact if the velocity above the reflector is constant, and if the reflector is flat. For layered media we saw that the travel time curve was hyperbolic, but the velocity used should be the RMS velocity. Unfortunately we don't know what this velocity is, so we attempt to estimate it from the data themselves. We proceed as follows.

1. Assume that each reflection event in a CMP gather has a travel time that corresponds to a hyperbola,

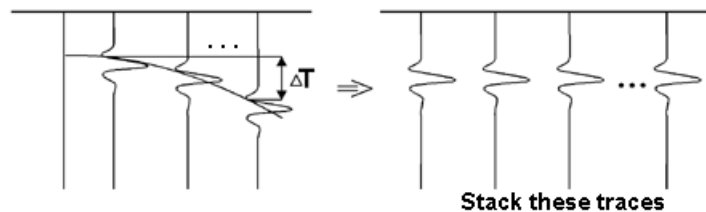
$$t^2(x) = t_0^2 + \frac{x^2}{V_{st}^2}$$

where V_{st} is a "stacking" velocity, or sometimes called the Normal Moveout Velocity, V_{nmo} .

2. For each reflection event hyperbola, perform a velocity analysis to find V_{st} . This is done by first choosing t_0 . Then choose a trial value of velocity v_1 . The associated travel time hyperbola is generated and it forms a trajectory on the CMP gather. Sum the energy of the seismic traces along the trajectory and plot this value on a graph of velocity versus energy. Repeat this procedure for different trial velocities. Choose as V_{st} , the velocity that yields the largest energy. In the diagram below v_2 represents the stacking velocity. The term cross power can be interpreted as total energy.

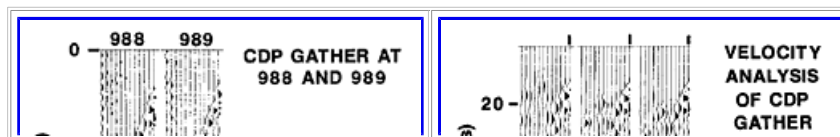


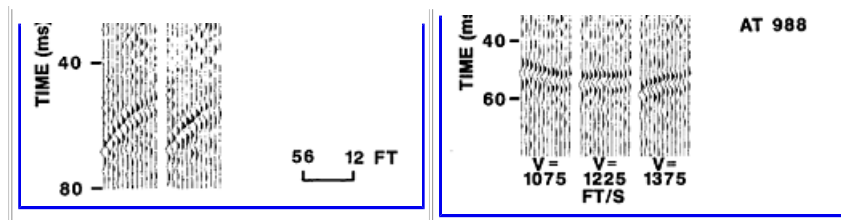
3. Calculate the Normal Moveout Correction: Again, using the hyperbola corresponding to V_{st} , compute the normal moveout for each trace and then adjust the reflection time by the amount ΔT .



4. Finally, stacking the normal moveout corrected traces generates a single trace. Each trace corresponds to a zero-offset trace, that is, the seismic trace that would have been recorded by a receiver that is coincident with the source.

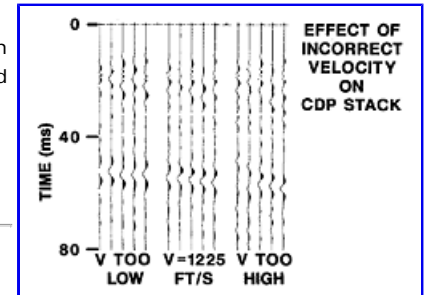
As an example consider the two CMP gathers in the figure below left (click for a larger version). The most prominent seismic wavelet at times between 50 and 70 ms is a bedrock reflection from about 9 m below the surface. The geophone offsets were 3.7 m (12 ft) for the nearest traces and 17 m (56 ft) for the farthest trace with 1.22 m (4 ft) between geophones. The reflection even of interest is at about 55 msec.





In the figure above right, the CMP gather for point 988 has been moveout corrected using different velocities. The moveout correction using a velocity of 1225 ft/s (328 m/s) causes the reflection events to occur at about the same time for all traces. Stacking these signals will produce a high quality reflection signal. Conversely, a velocity of 1075 ft/s is too small and produces too large a correction at far offsets.

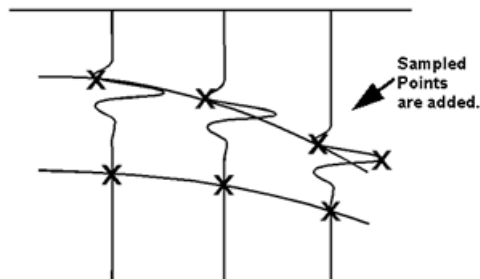
To show what happens if the wrong velocity is chosen for carrying out the normal moveout correction, five 12-fold CMP traces are shown (right) processed with three different velocities. When the velocity is too low, the frequency of the reflection wavelet is lowered and it is therefore depicted too shallow on the seismic section. When the velocity is too high the frequency decreases and the reflection wavelet is depicted at too large a time on the seismic section. The correct velocity gives the correct position for the wavelet and preserves the high frequencies which allow best resolution of small features and thin beds. Correct velocity is about 373 m/s (1225 ft/s).



Practicalities of CMP Processing

For each CMP gather, the velocity analysis is done at many times. A common method of displaying this work can be seen in the example to the right (click the image for a larger version). The *velocity spectrum* defines the stacking velocity as a function of reflection time. The energy (cross-power) function used to define the stacking velocity is calculated for a large number of narrow time windows along a seismic section. The velocity spectrum is typically displayed alongside the draped section. In the example shown here, stacking velocities for all times less than about 3 seconds are reasonably clear. Beyond 3 seconds there are multiple peaks that are caused by overlapping hyperbolic trajectories. The low velocity arrivals are due to multiple reflections in the near surface. They arrive at the same time as the reflection events from deep horizons.

A stacking velocity is defined for each time sample. The normal moveout correction and stacking can be carried out by summing along a different hyperbola for each time on the seismogram. Equivalently, the CMP gather may first be NMO corrected and then stacked. Both approaches are done in practise.



Summary: Essential Elements in CMP Processing

There are many different processing steps that *could* be performed. An example from GS Baker, 1999, is shown in the flow chart image here (click for a larger version.). **However**, the essential steps are summarized in the following short list.

1. Obtain CSP (Common Source Point) gathers.
2. Sort into CMP (Common Midpoint) gathers.
 - Reflection events (coming from approximately the same point in the earth) appear as hyperbolic trajectories. The goal is to stack them into a single trace.
3. For each event, perform a velocity analysis to find the stacking velocity.
4. Perform NMO correction and stack.
 - This yields a single trace corresponding to a coincident source and receiver.
5. Composite the traces into a CMP processed section.

These are the only steps we will be concerned with in these notes. Other steps may be used by experienced contractors and they may be necessary to produce more useful sections for interpretation, but the details are beyond the scope of this set of notes.

Static Corrections

Before carrying out the NMO correction it is usually necessary to perform a static correction, which amounts to moving the entire seismic trace up or down in time.

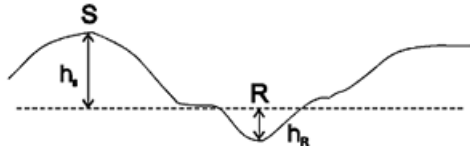
There are two main reasons for applying static corrections.

- i. Put shots and receivers on a flat datum plane.
- ii. Correct for near surface velocity anomalies beneath the source or receiver.

Elevation Statics

Consider the need for (i). Common midpoint shot receiver pairs acquire data on an irregular interface (right figures, top panel). Time differences are caused because of extra travel time associated with elevation of source and receiver. As a result reflections observed on the CMP gather will not have a hyperbolic form and they will not be amenable to normal CMP processing (bottom panel).

The correction procedure involves establishing a datum on which to locate source and receiver, and then adding or subtracting the incremental time. The reference velocity will be that of the upper layer.



The reflections of interest are usually coming from great depth and the upcoming energy is travelling nearly vertical. So the static correction due to elevation expressed as a change in travel time is $\Delta t = \frac{h_s}{v_1} - \frac{h_r}{v_1}$

and this time is subtracted from the record. That is, the whole seismic record is shifted in time by a value Δt .

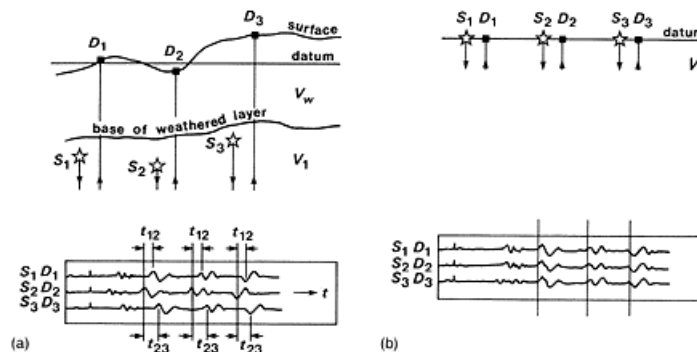
After static correction, the subsurface events will look more like an hyperbola and they will be ready for velocity analysis, NMO and stacking.

Near-surface velocity anomalies

Similar bulk shifts in time will occur if there is anomalous velocity beneath a source or receiver or if the thickness of the weathering layer changes substantially. The amount to be subtracted from the seismic trace time is given by the following formula:

$$\Delta t = \frac{h_a}{v_a} - \frac{h_a}{v_1} = \left(\frac{v_1 - v_a}{v_1 v_a} \right) h_a$$

The first layer is often highly weathered and it has variable thickness and velocity. It is also usually poorly consolidated and therefore is a poor transmitter of seismic energy. In exploration it is common to drill through the weathered zone into the upper layer. Drill measurements establish the thickness at the shot. By having a seismometer at the surface above the shot one can estimate the velocity of the weathered zone. The diagram below shows the differences between reflection events on adjacent seismograms due to the different elevations of shots and detectors and the presence of a weathered layer. After static corrections the seismograms should show better alignment of reflection events.



From Kearey, Philip and Micheal Brooks, An Introduction to Geophysical Exploration. 2nd ed. Blackwell Science: 1991.

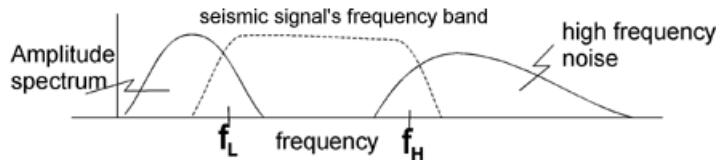
Filtering of Seismic Data

The interpretation of seismic data is made purely on the basis of what is observed in the final processed section. CMP processing greatly enhances the signal to noise ratio and allows coherent reflections to be visible. However, the data that goes into the CMP processing is often contaminated with "noise". The ground roll in the CSP gather shown previously is a good example. Also, the data might be contaminated with wind noise or instrument vibrations. These, and other types of noise, can be partially removed by using various filtering operations.

Three Useful Filtering Operations

1. Frequency Filtering

Wind, instrument, and cultural effects can generate unwanted noise at frequencies outside the seismic band. Often these are high frequency signals. Alternatively, ground roll and ship generated noise are low frequency. If this is the case then the seismic signal might lie in a frequency band that is distinct from the noise. This is illustrated below. The unwanted noise can be removed using bandpass filtering in the frequency domain.



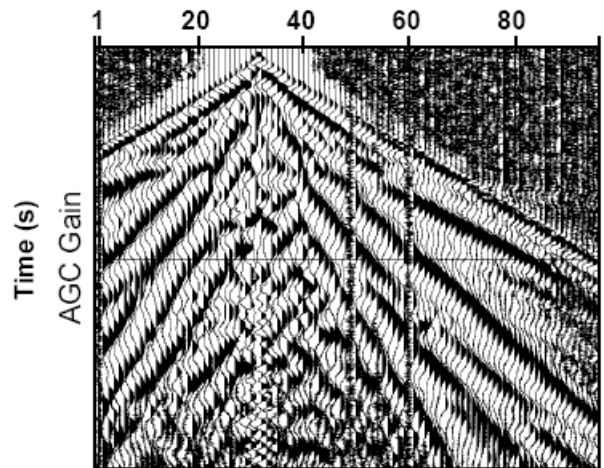
One can apply a frequency filter to remove all signals that have a frequency high than f_H and all frequencies lower than f_L .

The figures to the right illustrate effects of low pass and high pass filters. First examine the figure and try to identify aspects caused by high frequencies. Then try to identify aspects caused by lower frequency components in the signal.

Click buttons below to switch between different versions of the same data displayed after filtering.

| no filter | low pass | high pass |

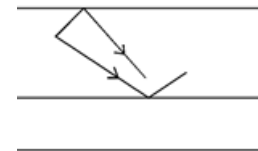
Which of these three could also be described as the result of applying a **low cut** filter, and which is the result of applying a **high cut** filter?



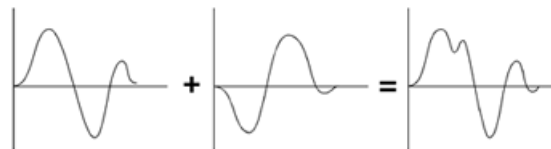
2. Deconvolution

a) Deghosting

In marine work a source is detonated slightly below the water surface. The primary seismic wavelet propagates downward but there is an upward propagating wave that reflects from the water surface and then propagates downward. The latter reflection is referred to as a 'ghost'.



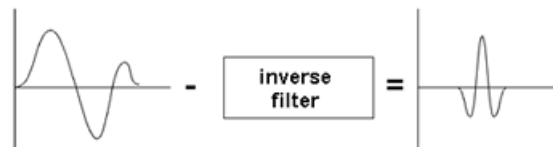
A seismic wavelet is composed of the original wavelet plus a closely spaced reflection. This complicates the wavelet as shown in the figure below.



A deghosting filter can remove the effect of the ghost so each reflection looks like a primary wavelet.

b) Wavelet or Signature Deconvolution

An "inverse filter" is designed so that the original source wavelet is contracted to a narrower and symmetric form. e.g.



This can greatly enhance the vertical resolution.

c) Predictive Deconvolution

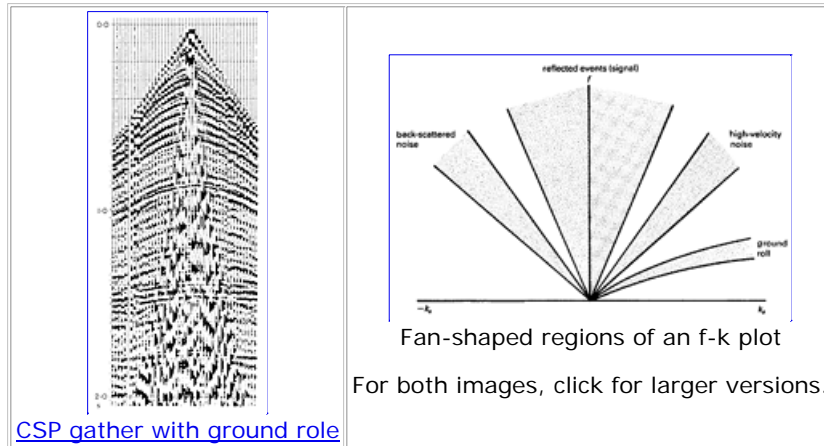
This type of processing can remove some multiples from a seismic section. Reverberation (multiples of the ocean-bottom reflection) in marine surveys is a common example of this type of problem.

3. Velocity or f-k filtering

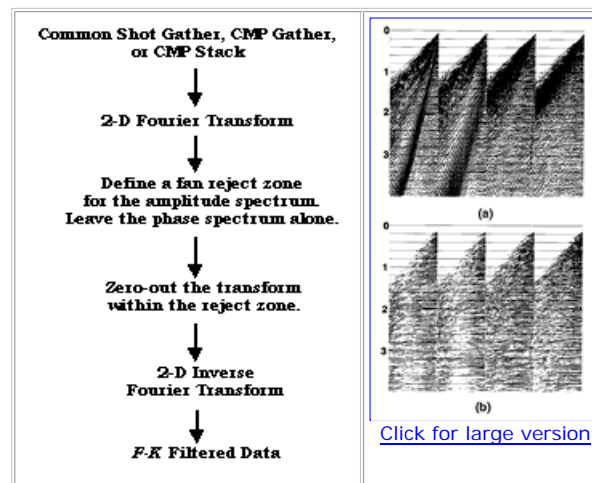
In shot gathers or on final sections we often have events that appear with a specific slope. Notice the ground roll, in the CSP gather below left. These are characterized by the late-arriving, high amplitude, low-frequency events which define a steep triangular-shaped central zone which mask the reflected arrivals.

Ground roll events have a small apparent velocity or equivalently a large dip. They can be isolated by taking a 2-D Fourier

transform. In that domain the ground roll is located in a fan-like region. By zeroing the Fourier transform values in this fan and then inverse Fourier transforming, we remove the ground roll. The diagram below right sketches these fan-shaped regions of an f-k plot for a typical shot gather which contains reflections events and noise.



The procedure for carrying out f-k filtering is provided in the flow chart, below left. The final diagram shows four shallow marine records before and after f-k filtering to remove coherent linear noise. Hyperbolic reflections are observed after removal of the noise.

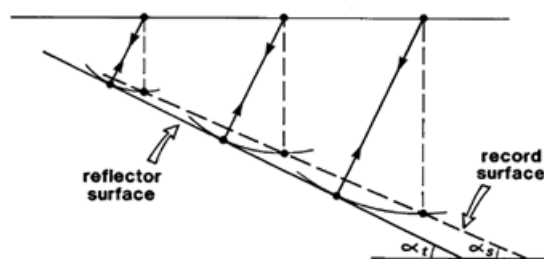


Migration

The ideal goal of seismic processing is to have a section which consists of true reflections which arise from structure directly beneath the source. The image should provide quantitative information about geologic structure. The vertical axis ideally is depth but generally the time-to-depth conversion is not carried out and the axis remains time. A processing step which is important in putting reflectors in their correct location is migration. The need for migration is most evident when layers are not flat lying. Consider a single dipping layer and seismic traces corresponding to a coincident source and receiver.

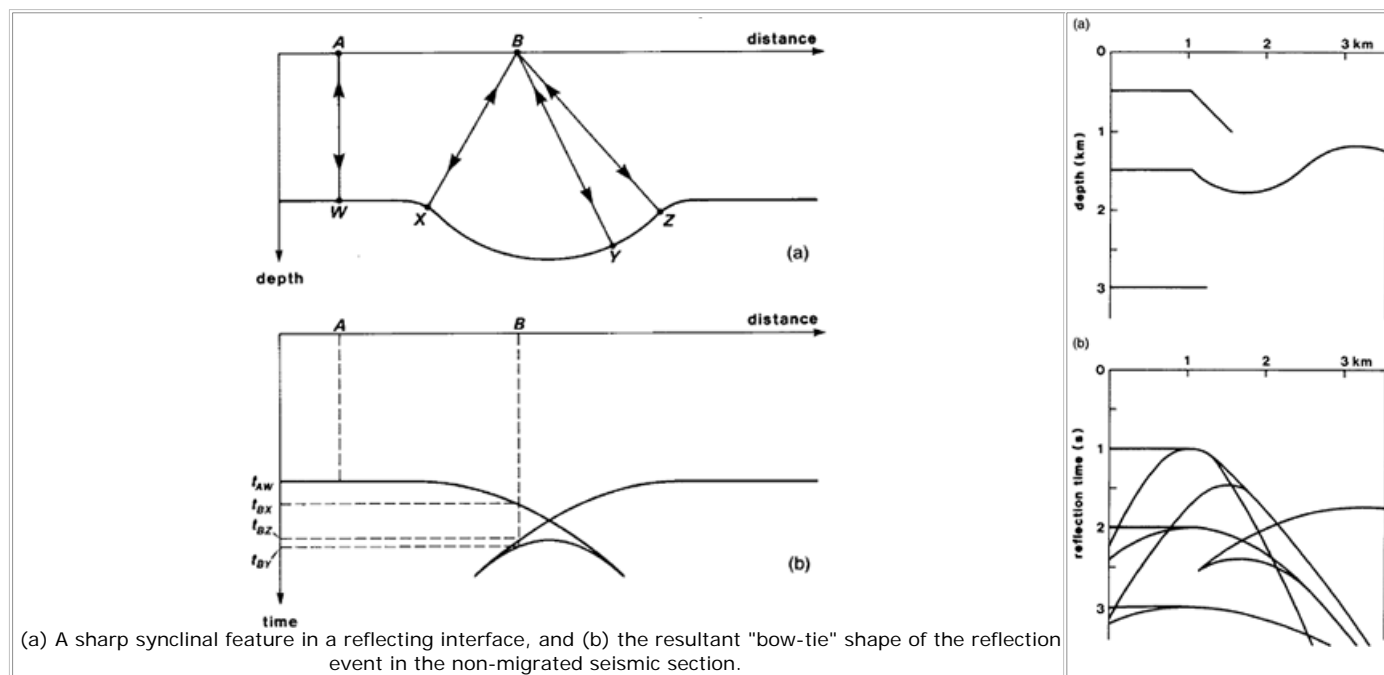
The reflection time is given by $2l/v_i$. But when we plot the section we think of the reflections arriving from directly beneath the shot. The result is that a sloping reflector in the ground will appear as a sloping reflector in the seismic section, but the slopes will be different. The true dip will be greater than that observed on the section. The relationship between the two dips is

$$\sin(\alpha_r) = \tan(\alpha_s)$$



A planar-dipping reflector surface and its associated record surface derived from a non-migrated seismic section.

Other artifacts are visible in non-migrated sections. Synclines will be observed as "bow ties." Their width on the seismic section is narrower than reality and their structural depth is diminished. Conversely, anticlines will appear wider than they really are. Finally a point scatterer will be observed as a hyperbola.

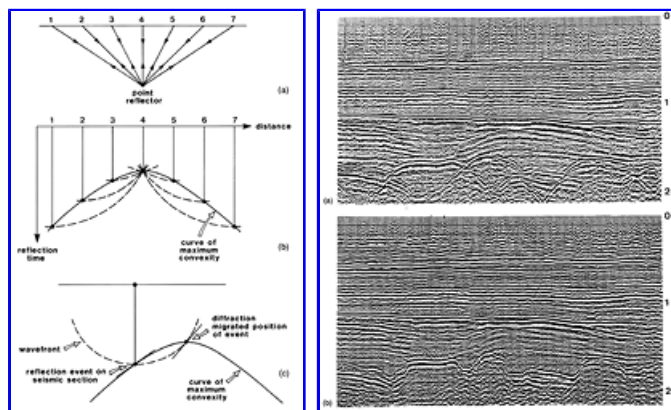


The plots above show several synthetic geologic structures and their associated non-migrated seismic sections.

Migration is the process of reconstructing a seismic section so that reflection events are repositioned under their correct surface location and at a correct vertical reflection travel time. The effects of migration are to:

- collapse a diffraction hyperbola back to a point;
- make dipping structures appear with correct dip angle;
- remove the bow ties from synclines and
- shorten anticlines.

The principles of migration are shown in the figure below left. A single point diffractor will generate a hyperbolic trajectory in a travel time section. Migration collapses the hyperbolic event back to its apex.



Click images for larger versions.

The example above right shows an unmigrated section (top). The section after migration is shown in the bottom panel.

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