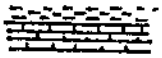
KEY CONCEPTS

Reflection seismic uses sound waves to investigate the subsurface. Acoustic impedance is the rock property that governs reflections

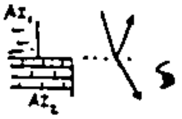
Acoustic impedance = interval velocity x density.

Reflections arise at boundaries across which acoustic impedance (AI) changes, eg.



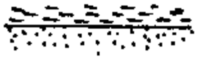
Shale $AI_1 = 2.5 g/cm^3 \times 3000 m/s$

Limestone $AI_2 = 2.6 g/cm^3 \times 4500 m/s$



Reflection
and refraction

No reflection occurs if the acoustic impedance does not change, even if lithology changes.



$AI_1 = 2.3 g/cm^3 \times 2300 m/s = 5290$ units

$AI_2 = 2.5 g/cm^3 \times 2116 m/s \quad 5290$ units

No reflection
Only refraction

Ray path bending (refraction) responds to velocity change: the greater the velocity change so the greater the refraction

Reflection Strength

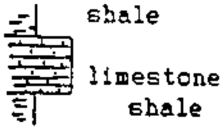
The greater the difference in the acoustic impedance, the stronger the reflection.

The size of the change is defined by the reflection coefficient (RC).

$$RC = \frac{AI_2 - AI_1}{AI_1 + AI_2}$$

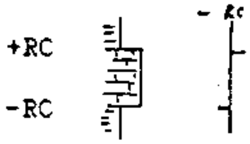
RCs are +ve (positive) if AI increases, eg

RCs are -ve (negative) if AI decreases, eg

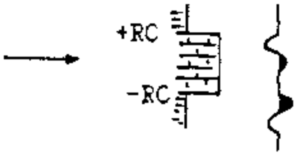


Reflections from positive RCs are compressional. Reflections from negative RCs are refractional.

The ideal seismic response to an acoustic impedance change would be a spike,



but life is more complicated and we get a wave consisting of troughs and peaks

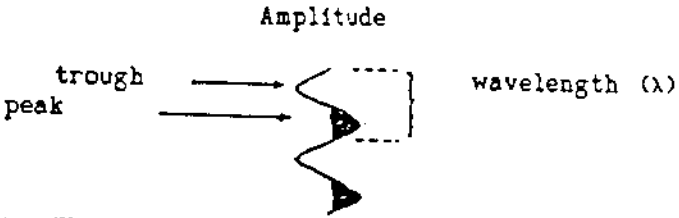


Some basic definitions

Frequency = number of times a wavelet repeats per second,

$\nu = \frac{1s}{\lambda(ms)}$ measured in Hertz,

eg. a wavelet with a duration of 50ms (.05s) has a frequency of 25 Hertz (Hz)



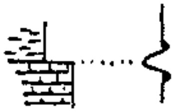
$$\text{Wavelength (m or ft)} = \frac{\text{velocity}}{\text{frequency}}$$

Waveforms are complicated but pragmatic interpreters can, as a first assumption, hope that the wavelet and response on the seismic section is either

MINIMUM PHASE

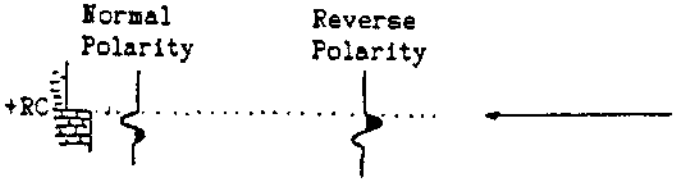
Maximum amplitude in first peak or trough. Wavelet starts at the AI boundary

ZERO PHASE



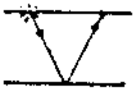
Symmetric wavelet with maximum amplitude at the AI boundary

Polarity is the way in which seismic data is recorded and displayed. SEG normal polarity data has compressional waves (from +RC boundaries) recorded as a negative number on tape and displayed as a trough (white). A negative reflection has the same shape but is reversed - every peak a trough, every trough a peak.

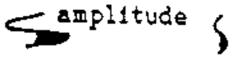


example of normal and reverse polarity minimum phase wavelet

WHAT WE MEASURE - the basics (we can do much more than this)

1.  travel time down to a reflector and back to the surface = two-way time (TWT) which is related to depth by:

$$\text{time} = \frac{2 \times \text{depth}}{\text{velocity}}$$

2. 

- the relative amplitude gives a measure of the acoustic impedance change

3.  reflection polarity which (for interference free reflections) can show if the boundary has a positive or negative reflection coefficient

4. Interval velocity - in favourable circumstances.

All these 4 parameters can carry a lithological and geological message!

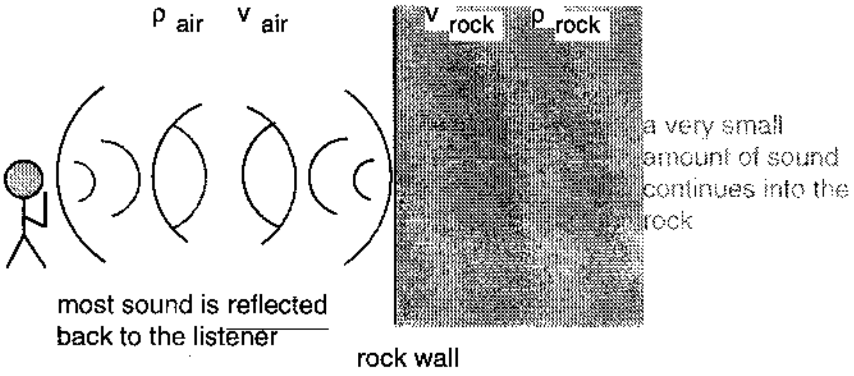
LECTURE 3 — DESCRIPTIVE GEOMETRY: SEISMIC REFLECTION

3.1 Echo Sounding

Geology presents us with a basic problem. Because rocks are opaque, it is very difficult to see through them and thus it is difficult to know what is the three-dimensional geometry of structures.

This problem can be overcome by using a remote sensing technique known as seismic reflection. This is a geophysical method which is exactly analogous to echo sounding and it is widely used in the petroleum industry. Also several major advances in tectonics have come from recent application of the seismic reflection in academic studies. I'm not going to teach you geophysics, but every modern structural geologist needs to know something about seismic reflection profiling.

Lets examine the simple case of making an echo first to see what the important parameters are.



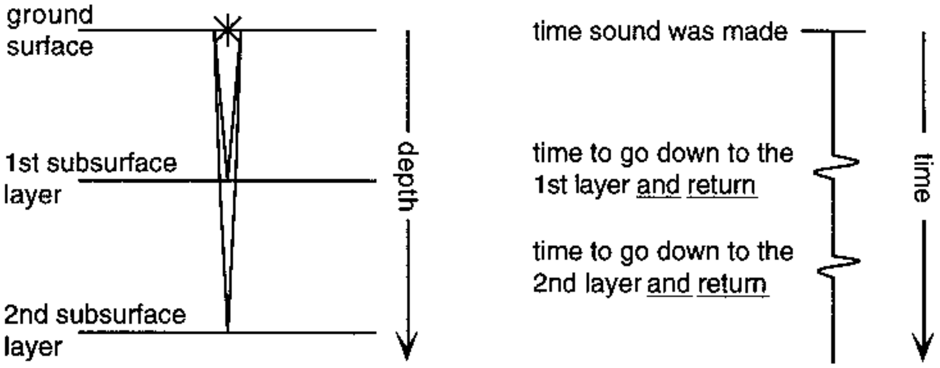
Why do you get a reflection or an echo? You get an echo because the densities and sound velocities of air and rock are very different. If they had the same density and velocity, there would be no echo. More specifically

$$velocity = \mathbf{V} = \sqrt{\frac{E}{\rho}}$$ (E = Young's modulus)

and

$$\text{reflection coefficient} = R = \frac{\text{amplitude of reflected wave}}{\text{amplitude of incident wave}} = \frac{\rho_2 V_2 - \rho_1 V_1}{\rho_2 V_2 + \rho_1 V_1}$$

In seismic reflection profiling, what do you actually measure?

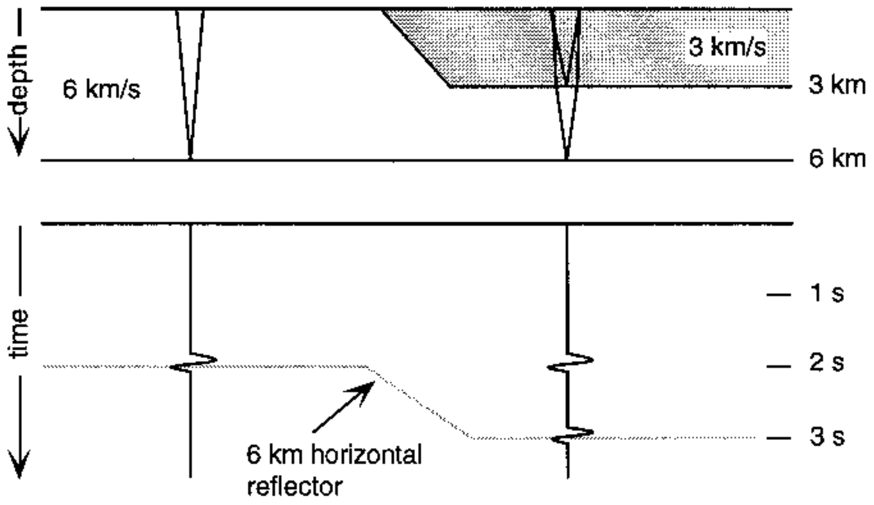


The above illustration highlights three important things about seismic reflection profiling:

1. Measure time, not depth,
2. The time recorded is round trip or two-way time, and
3. To get the depth, we must know the velocity of the rocks.

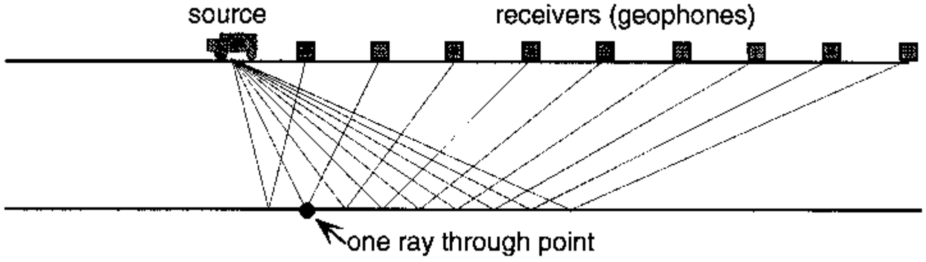
Velocities of rocks in the crust range between about 2.5 km/s and 6.8 km/s. Most sedimentary rocks have velocities of less than 6 km/s. These are velocities of P-waves or compressional waves, not shear waves. Most seismic reflection surveys measure P- not S-waves.

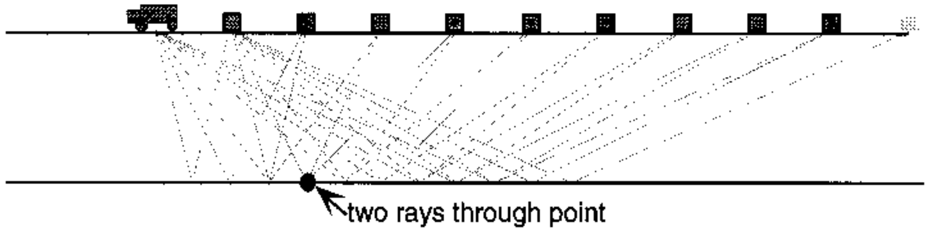
Seismic reflection profiles resemble geologic cross-sections, but they are not. They are distorted because rocks have different velocities. The following diagram illustrates this point.

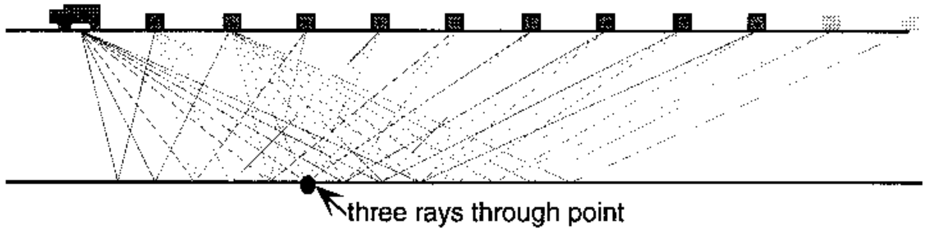


3.2 Common Depth Point (CDP) Method

In the real earth, the reflectivity at most interfaces is very small, $R \approx 0.01$, and the reflected energy is proportional to R^2. Thus, at most interfaces ~99.99% of the energy is transmitted and 0.01% is reflected. This means that your recording system has to be able to detect very faint signals coming back from the subsurface.







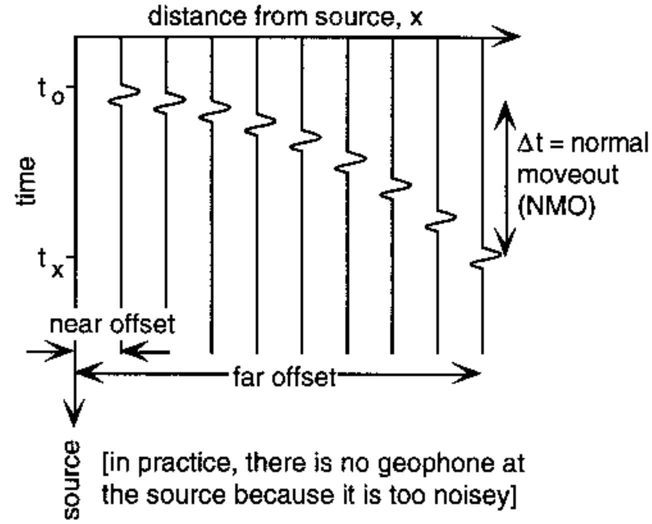
The black dot, and each point on the reflector with a ray going through it, is a common depth point. Notice that there are twice as many CDPs as there are stations on the ground (where the geophones are). That is, there is a CDP directly underneath each station and a CDP half way between each station (hence the name "common midpoint")

Also, in a complete survey, the number of traces through each midpoint will be equal to one half the total number of active stations at any one time. [This does not include the ends of the lines where there are fewer traces, and it also assumes that the source moves up only one station at a time.] The number of active stations is determined by the number of channels in the recording system. Most modern seismic reflection surveys use at least 96 (and sometimes -- but not often -- as many as 1024 channels), so that the number of traces through any one CDP will be 48.

This number is the data redundancy, of the fold of the data. For example, 24 fold or 2400% means that each depth point was sampled 24 times. Sampling fold in a seismic line is the same thing as the "over-sampling" which you see advertised in compact disk players.

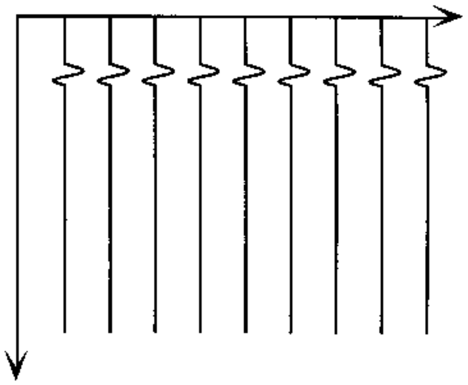
Before the seismic reflection profile can be displayed, there are several intermediate steps. First, all of the traced through the same CDP have to be gathered together. Then you have to determine a set of velocities, known as stacking or NMO velocities, which will correct for the fact that each ray through a CDP has a path of a different length. These velocities should line up all of the individual "blips" corresponding to a single reflector on adjacent traces

CDP Gather



[in practice, there is no geophone at the source because it is too noisey]

CDP Gather with NMO



the NMO velocity is whatever velocity that lines up all the traces in a CDP gather. It is not the same as the rock velocity

The relation between the horizontal offset, x, and the time at which a reflector appears at that offset, tx, is:

$$t_x^2 = t_0^2 + \frac{x^2}{V_{stacking}^2}$$

or

$$\Delta t = t_x - t_0 = \left(t_0^2 + \frac{x^2}{V_{stacking}^2} \right)^{\frac{1}{2}} - t_0$$

<u>If</u> you have a very simple situation in which all of your reflections are flat and there are only

vertical velocity variations (i.e. velocities do not change laterally), then you can calculate the rock interval velocities from the stacking velocities using the Dix equation:

$$V_{i12} = \left(\frac{V_{st2}^2 t_2 - V_{st1}^2 t_1}{t_2 - t_1} \right)^{\frac{1}{2}}$$

where V_{i12} is the interval velocity of the layer between reflections 1 and 2, V_{st1} is the stacking velocity of reflection 1, t_1 is the two way time of reflection 1, etc. The interval velocity is important because, to convert from two-way time to depth, we must know the interval, not the stacking, velocity.

Once the correction for normal moveout is made, we can add all of the traces together, or stack them. This is what produces the familiar seismic reflection profiles.

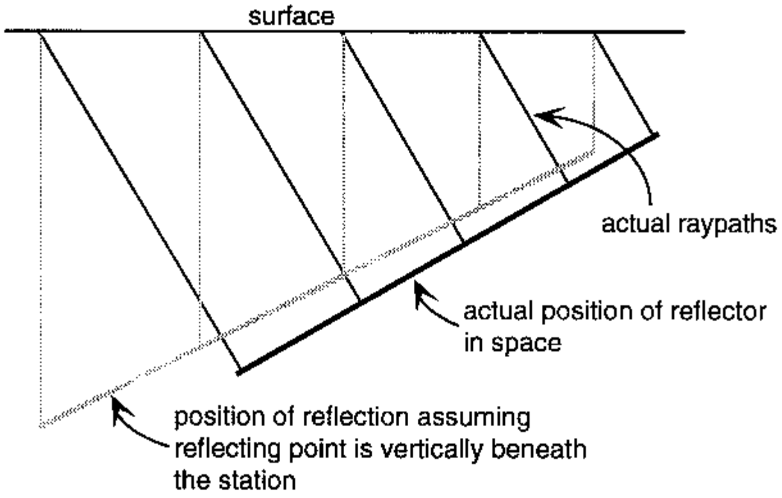
Processing seismic data like this is simple enough, but there are huge amounts of data involved. For example a typical COCORP profile is 20 s long, has a 4 ms digital sampling rate (the time interval between numbers recorded), and is 48 fold. In a hundred station long line, then, we have

$$\frac{(200 \text{ CDPs})(48 \text{ sums})(20 \text{ s})}{\frac{0.004 \text{ s}}{\text{data sample}}} = 48 \times 10^6 \text{ data samples}.$$

For this reason, the seismic reflection processing industry is one of the largest users of computers in the world!

3.3 Migration

The effect of this type of processing is to make it look like the source and receiver coincide (e.g. having 48 vertical traces directly beneath the station). Thus, all reflections are plotted as if they were vertically beneath the surface. This assumption is fine for flat layers, but produces an additional distortion for dipping layers, as illustrated below.



Note that the affect of this distortion is that all dipping reflections are displaced down-dip and have a shallower dip than the reflector that produced them. The magnitude of this distortion is a function of the dip of the reflector and the velocity of the rocks.

The process of migration corrects this distortion, but it depends on well-determined velocities and on the assumption that all reflections are in the plane of the section (see "sideswipe", below). A migrated section can commonly be identified because it has broad "migration smiles" at the bottom and edges. Smiles within the main body of the section probably mean that it has been "over-migrated."

3.4 Resolution of Seismic Reflection Data

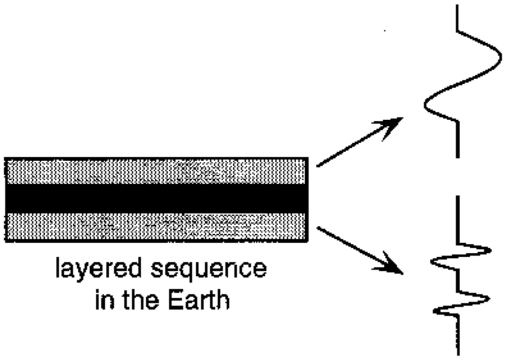
The ability of a seismic reflection survey to resolve features in both horizontal and vertical directions is a function of wavelength:

$$\lambda = \text{velocity} / \text{frequency}.$$

Wavelength increases with depth in the Earth because velocity increases and frequency decreases. Thus, seismic reflection surveys lose resolution with increasing depth in the Earth.

3.4.1 Vertical Resolution

Generally, the smallest (thinnest) resolvable features are 1/4 to 1/8 the dominant wavelength:

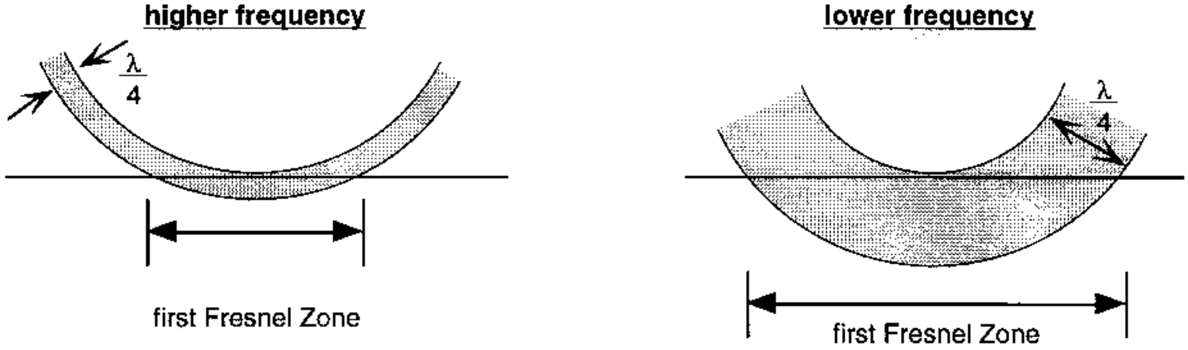


At low frequencies (long wavelengths) these three beds will be "smeared out" into one long wave form

At higher frequencies (shorter wavelengths) the three beds will be distinguishable on the seismic section

3.4.2 Horizontal Resolution

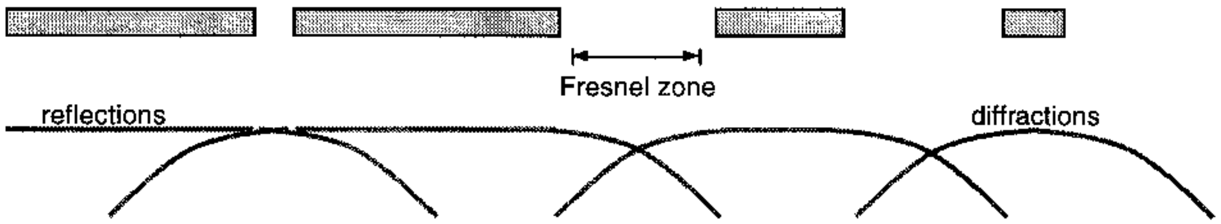
The horizontal resolution of seismic reflection data depends on the Fresnel Zone, a concept which should be familiar to those who have taken optics. The minimum resolvable horizontal dimensions are equal to the first Fresnel zone, which is defined below.



Because frequency decreases with depth in the crust, seismic reflection profiles will have greater horizontal resolution at shallower levels.

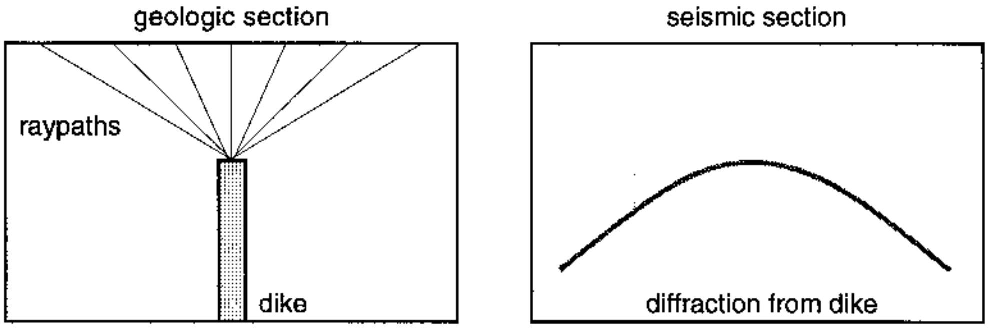
At 1.5 km depth with typical frequencies, the first Fresnel Zone is ~300 m. At 30 km depth, it is about 3 km in width.

Consider a discontinuous sandstone body. The segments which are longer than the first Fresnel Zone will appear as reflections, whereas those which are shorter will act like point sources. Point sources and breaks in the sandstone will generate diffractions, which have hyperbolic curvature:

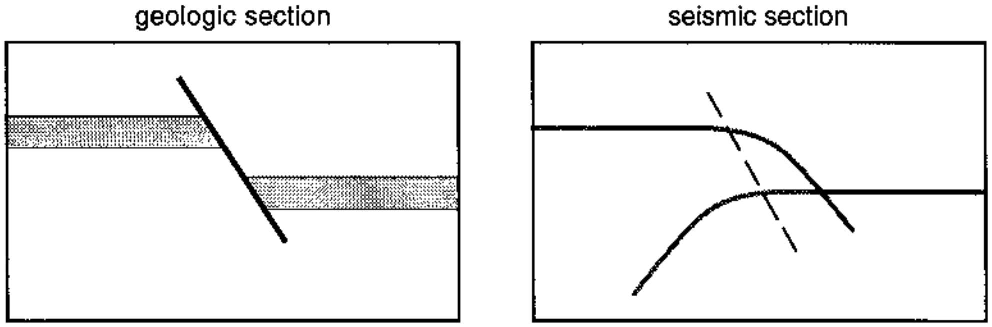


3.5 Diffractions

Diffractions may look superficially like an anticline but they are not. They are extremely useful, *especially because seismic reflection techniques are biased toward gently dipping layers and do not image directly steeply dipping or vertical features.* Diffractions help you to identify such features. For example, a vertical dike would not show up directly as a reflection but you could determine its presence by correctly identifying and interpreting the diffractions from it:



High-angle faults are seldom imaged directly on seismic reflection profiles, but they, too, can be located by finding the diffractions from the truncated beds:



The shape and curvature of a diffraction is dependent on the velocity. At faster velocities, diffractions become broader and more open. Thus at great depths in the crust, diffractions may be very hard to distinguish from gently dipping reflections.

3.6 Artifacts

The seismic reflection technique produces a number of artifacts -- misleading features which are easily misinterpreted as real geology -- which can fool a novice interpreted. A few of the more common "pitfalls" are briefly listed below.

3.6.1 Velocity Pullup/pulldown

We have already talked about this artifact when we discussed the distortion due to the fact that seismic profiles are plotted with the vertical dimension in time, not depth. When you have laterally varying velocities, deep horizontal reflectors will be pulled up where they are overlain locally by a high velocity body and will be pushed down by a low velocity body (as in the example on page 2).

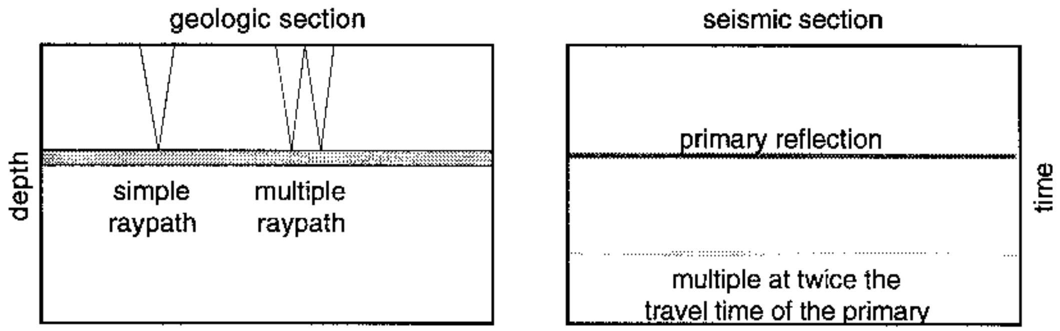
3.6.2 Multiples

Where there are very reflective interfaces, you can get multiple reflections, or multiples, from those interfaces. The effective reflectivity of multiples is the product of the reflectivity of each reflecting interface. For simple multiples (see below) then,

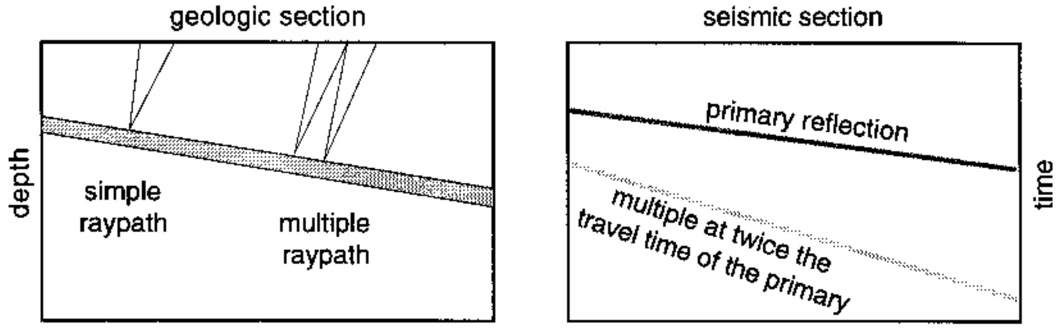
$$R_{\text{multiple}} = R^2_{\text{primary}}.$$

If the primary reflector has a reflection coefficient of 0.01 then the first multiple will have an effective reflection coefficient of 0.0001. In other words, multiples are generally only a problem for highly reflective interfaces, such as the water bottom in the case of a marine survey or particularly prominent reflectors in sedimentary basins (e.g. the sediment-basement interface).

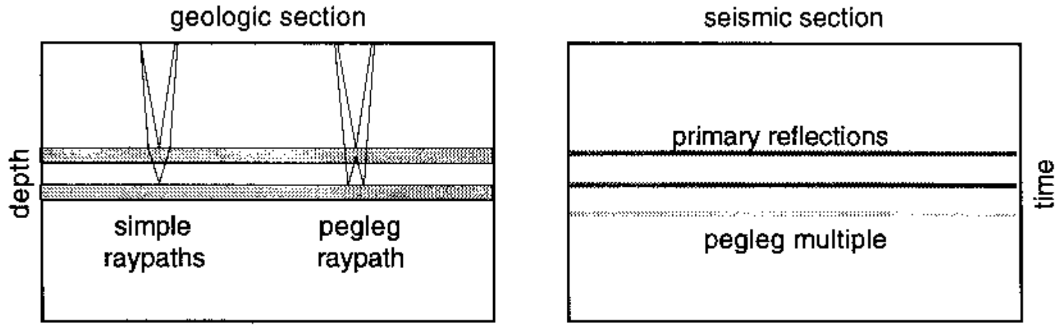
Multiple from a flat layer:



Multiple from a dipping layer (note that the multiple has twice the dip of the primary):

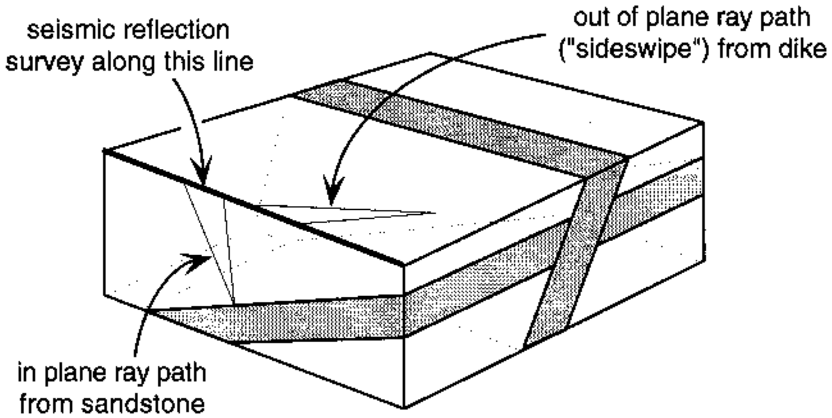


Pegleg multiples:



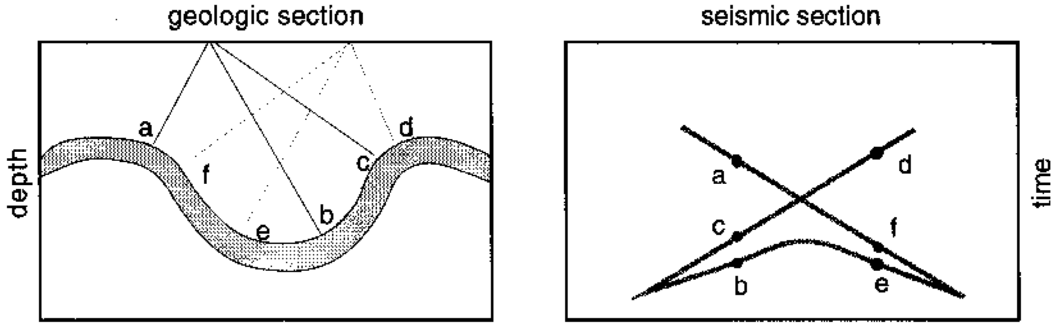
3.6.3 Sideswipe

In seismic reflection profiling, we assume that all the energy that returns to the geophones comes from within the vertical plane directly beneath the line of the profile. Geology is inherently three-dimensional so this need not be true. Even though geophones record only vertical motions, a strong reflecting interface which is out-of-the-plane can produce a reflection on a profile, as in the case illustrated below.



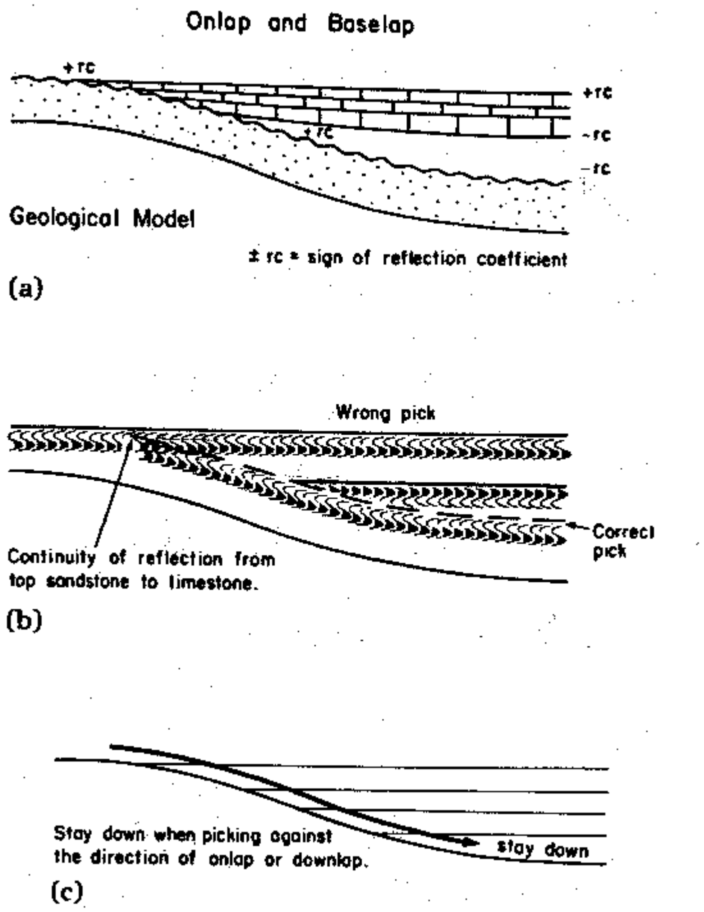
Reflections from out of the plane is called sideswipe. Such reflections will cross other reflections and will not migrate out of the way. (Furthermore they will migrate incorrectly because in migration, we assume that there has been no sideswipe!) The main way of detecting sideswipe is by running a sufficient number of cross-lines and tying reflections from line to line. Sideswipe is particularly severe where seismic lines run *parallel* to the structural grain.

3.6.4 Buried Focus



Tight synclines at depth can act like concave mirrors to produce an inverted image quite unlike the actual structure. Although the geological structure is a syncline, on the seismic profile it looks like an anticline. Many an unhappy petroleum geologist has drilled a buried focus hoping to find an anticlinal trap! The likelihood of observing a buried focus increases with depth because more and more open structures will produce the focus. A good migration will correct for buried focus.

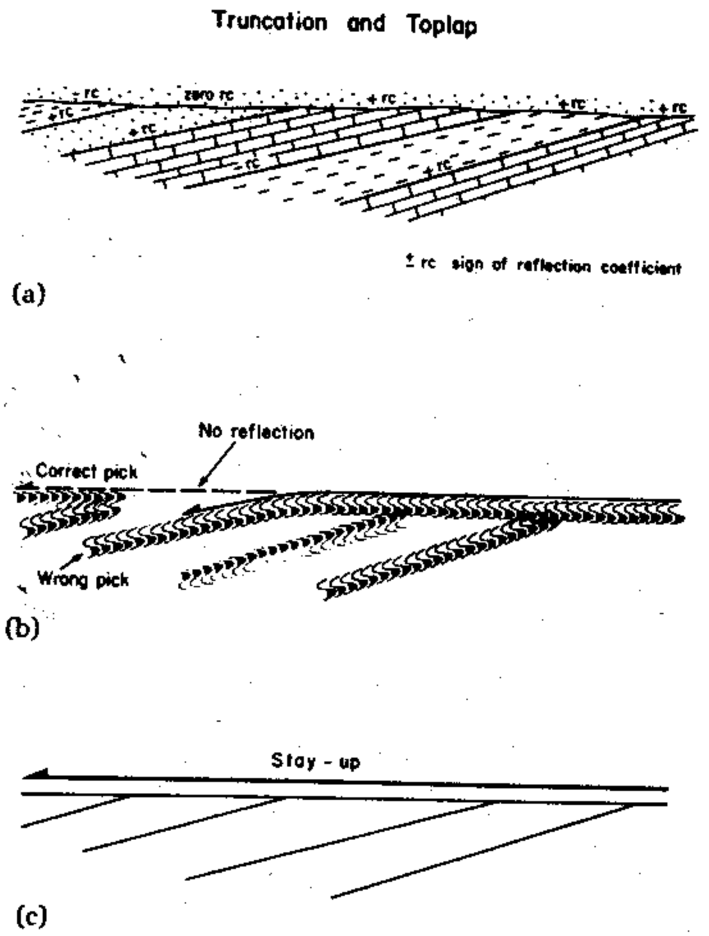
Onlap



Picking criterion—onlap and downlap. (a) *Geological model: A sandstone of intermediate acoustic impedance is onlapped by shales of low acoustic impedance, and limestone of high acoustic impedance. The reflection coefficient signs are indicated on the diagram.* (b) *Seismic expression: The top sand reflection, the sequence boundary defined by the onlap, changes polarity due to the varying reflection coefficients between the sandstone, limestone, and shale. The apparent continuity between the top sand and top limestone reflections is a potential trap for the unwary interpreter.* (c) *A general rule when following an onlapped sequence boundary is to stay down when picking against the onlap direction.*

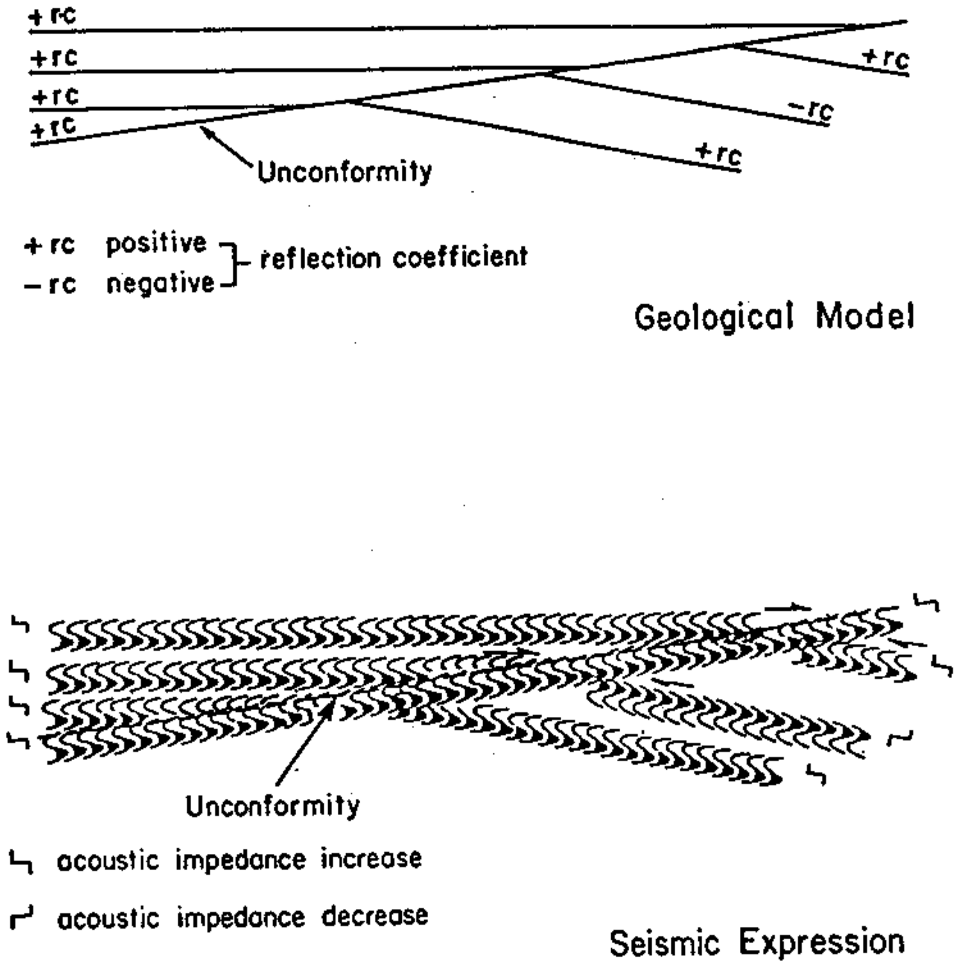
Stay-down

Truncation



Picking criterion—toplap and truncation. (a) *Geological model: An interbedded sequence subcrops an unconformity overlain by a sand. Signs of the reflection coefficient are indicated. There is no acoustic-impedance contrast between the sands.* (b) *Seismic expression: The unconformity has a positive reflection coefficient to the right, no reflection where it is subcropped by sand, and a negative reflection coefficient to the left. A potential interpretation pitfall would be to take the unconformity pick along the top limestone reflection.* (c) *A general rule when following a surface in the direction of truncation or toplap direction is to stay high.*

Stay-up



F 6 Idealised reflection relationships at a sequence boundary for a minimum phase wavelet (local interference effects are not shown). (a) Geological model showing reflection coefficients of the reflectors. The unconformity is assumed to have a large positive reflection coefficient compared with the truncated and onlapped reflectors. (b) Seismic expression: the minimum-phase reflection from the unconformity interferes with the truncated reflections, which terminate against the follow-cycle of the unconformity reflection. The onlap reflections, however, continue onto the unconformity surface and their terminations can be used to locate accurately the sequence boundary.