Controlled source seismology: Seismic reflection

Amplitudes reflected and transmitted

The amplitude of the reflected, transmitted and converted phases can be calculated as a function of the incidence angle using Zoeppritz’s equations.

Reflection and transmission coefficients for a specific impedance contrast. The critical angle is 30 degrees.
Amplitudes reflected and transmitted

Simple case: Normal incidence

Reflection coefficient

\[
R_C = \frac{A_R}{A_i} = \frac{\rho_2 V_2 - \rho_1 V_1}{\rho_2 V_2 + \rho_1 V_1}
\]

Transmission coefficient

\[
T_C = \frac{A_T}{A_i} = 1 - R_C = \frac{2 \rho_1 V_1}{\rho_2 V_2 + \rho_1 V_1}
\]

These coefficients are determined from the product of velocity and density – the impedance of the material.

\[ R_C \text{ usually small} \quad \text{– typically 1% of energy is reflected.} \]

→ We must stack many arrivals to enhance reflections

Shot gathers

direct arrival

reflection hyperbolae become fatter with depth (i.e. velocity)

offset

surface waves “ground roll” i.e. noise
Common depth point gathers

To enhance signal to noise we use more than one shot

Fold of coverage = \( \frac{\text{number of receiver}}{2 \times \text{shot spacing}} \)

where shot spacing is in units of receiver spacing

Typical fold values: 24, 48 or 96

Normal move out (NMO)

The hyperbolae shaped reflection is referred to at the normal move out

\[
T_x = \frac{2SR}{V_1} = \frac{2}{V_1} \sqrt{h_1^2 + \left(\frac{x}{2}\right)^2}
\]

or

\[
T_x^2 = T_0^2 + \frac{x^2}{V_1^2} \quad \Rightarrow \text{a hyperbolae}
\]

The importance of NMO

- Having determined the layer velocity, we can use the predicted quadratic shape to identify reflectors
- Then correct (shift traces) and stack to enhance signal to noise

\[
\Delta T_{\text{NMO}} = T_x - T_0 \approx \frac{x^2}{2T_0 V_1^2}
\]
Normal move out (NMO)

From the traveltime equation we can see that the hyperbolae become fatter/flatter with increasing velocity

\[ T_x^2 = T_0^2 + \frac{x^2}{V_1^2} \]

Subtract the correction from the common depth point gather

\[ \Delta T_{NMO} \approx \frac{x^2}{2T_0V_1^2} \]

Stacking velocity

In order to stack the waveforms we need to know the velocity. We find the velocity by trial and error:

- For each velocity we calculate the hyperbolae and stack the waveforms
- The correct velocity will stack the reflections on top of one another
- So, we choose the velocity which produces the most power in the stack

\[ \Delta t_{NMO} = \frac{x^2}{2\alpha_1^2t_0} \]
Stacking velocity

Multiple layer case

A stack of multiple horizontal layers is a more realistic approximation to the Earth

• Can trace rays through the stack using Snell’s Law (the ray parameter)
• For near-normal incidence the moveout continues to be a hyperbolae
• The shape of the hyperbolae is related to the time-weighted rms velocity above the reflector
  ➔ Generate a velocity spectrum
  ➔ Pick stacking velocities

The interval velocity can be determined from the rms velocities layer by layer starting at the top

Multiples: When two velocities produce a high amplitude stack one is probably a multiple

Multiples

Due to multiple bounce paths in the section
  ➔ Looks like repeated structure

Diminished by stacking, but it is better to remove them prior to stacking:

Removed with deconvolution
  • easily identified with an autocorrelation
  • removed using cross-correlation of the autocorrelation with the waveform
Velocity determination

The interval velocities cannot be determined very well from stacking velocities.

When the depth of the reflector is much greater than the horizontal offset the NMO correction is insensitive to velocity i.e. the width of the velocity spectrum is broad meaning a range of velocities will work.

To estimate velocities of deep layer we must use complementary refraction surveys.

Example

Two layer model: \( \alpha_1 = 6 \text{ km/s}, z = 20 \text{ km} \)

Equation of the reflection hyperbolae:

\[
 t = \frac{2}{\alpha_1} \sqrt{z^2 + \frac{x^2}{4}} = \frac{1}{3} \sqrt{400 + \frac{x^2}{4}}
\]

Normal move out correction:

\[
 \Delta t_{NMO} = \frac{x^2}{2 \alpha_1^2 t_0} = \frac{x^2}{480}
\]

For a 5 km offset:

- \( \alpha_1 = 6 \text{ km/s} \) then 0.052 sec – correct value
- \( \alpha_1 = 5.5 \text{ km/s} \) then 0.062 sec
- \( \alpha_1 = 6.5 \text{ km/s} \) then 0.044 sec

Are these significant differences?
What can we do to improve velocity resolution?
"Non-horizontal" "non-infinite" layers

Diffraction

Huygens' principle: Every point on a wavefront acts as a point source generating spherical waves.

At the end of an interface diffracted waves radiate out in all directions.

Diffractions also occur from any irregularities that are comparable in scale to the wavelength of the signal.

Typical wavelengths?

Diffraction hyperbolae

A to B: normal incidence reflections
B to C: diffractions from P

Equation of the diffraction hyperbolae:

\[ t = \frac{2}{\alpha_1} \sqrt{z_1^2 + x^2} \]

Note that a diffraction hyperbolae has a greater curvature than a reflection hyperbolae.

Recall the equation for a reflection hyperbolae:

\[ t = \frac{2}{\alpha_1} \sqrt{z_1^2 + \frac{v}{2} x^2} \]
Dipping layers

For horizontal reflectors the reflection point is vertically below the source/receiver.

For dipping layers the reflection comes from a point up dip.

Therefore, a traveltime section will always show a reduced dip:

$$\tan \delta' = \sin \delta$$

Migration

The process of trying to move reflections back to their point of origin.

Intended to deal with:
- Dipping interfaces
- Curved interfaces
- Diffractions
- Reflections from the 3rd dimension

The various migration methods all try to recognize diffractions by their hyperbolic shape and project the energy back to the origin point.

- Migration collapses diffraction hyperbolae to a point, and corrects dips.
A non-migrated section

Fig. 4.47 Finger record from the northern Aegean Sea, Greece, across a zone of active faults extending up to the sea bed. The sea floor is underlain by a layered sequence of Holocene muds and silts that can be traced to a depth of about 50 m. Note the diffraction patterns associated with the edges of the individual fault blocks.

Anticlines and synclines

Anticlines broaden
Syndines produce a bow-tie
Anticlines and synclines

(a) Pre-migration stack

(b) Migrated stack
Resolution of structure

Consider a vertical step in an interface
To be detectable the step must cause an
delay of \( \frac{1}{4}\) to \( \frac{1}{2}\) a wavelength
This means the step \( h \) must be \( \frac{1}{8}\) to \( \frac{1}{4}\) the wavelength (two way traveltime)

**Example:**
20 Hz, \( \alpha = 4.8 \text{ km/s} \) then \( \lambda = 240 \text{ m} \)
Therefore need an offset greater than 30 m

Shorter wavelength signal (higher frequencies) have better resolution.

What is the problem with very high frequency sources?

When you have been mapping faults in the field what were the vertical offsets?
Fresnel Zone

Tells us about the horizontal resolution on the surface of a reflector

First Fresnel Zone

The area of a reflector that returns energy to the receiver within half a cycle of the first reflection

The width of the first Fresnel zone, \( w \):

\[
\left( d + \frac{\lambda}{4} \right)^2 = d^2 + \left( \frac{w}{2} \right)^2
\]

\[
w^2 = 2d\lambda + \frac{\lambda^2}{4}
\]

If an interface is smaller than the first Fresnel zone it appears as an point diffractor, if it is larger it appears as an interface

Example:

30 Hz signal, 2 km depth where \( \alpha = 3 \text{ km/s} \) then \( \lambda = 0.1 \text{ km} \) and the width of the first Fresnel zone is 0.63 km