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Chapter 6

Seismic Refraction Tomography at Limburg - a candidate site for Einstein Telescope

6.1 Introduction

Seismic methods for imaging the subsurface can be broadly categorized into passive seismic and active seismic methods. In the last chapter we discussed the application of passive seismic methods for imaging the subsurface. Here we focus on the active seismic measurements conducted at Terziet in Limburg. The passive seismic study gave us 2D group-velocity maps which can be largely attributed to contrasts in shear wave velocity of the medium. In order to get better sensitivity to P-wave velocity changes an active survey was performed. From the passive study conducted prior to this and as discussed in chapter 5 we could infer subsurface information reliably down to 3 Hz which corresponds to about 80 m in depth. Due to the high impedance contrast at depths of 35 to 40 m, the sensitivity of the surface waves was observed to reduce by an order of magnitude to deeper subsurface velocity changes. Hence, we conduct the active survey to be able to image up to depths of 300 m and obtain the P-wave velocity model of the subsurface.

Active seismic surveys are typically carried out by injecting a signal of known strength and phase into the subsurface and then receiving its response at desired locations with the help of geophones. Signals received at the geophones are a complex combination of compressional and shear waves returning back to the surface after interacting with the medium of propagation. Hence the waveform received by the geophones is a superposition of the transmitted and the reflected wavefields in the media. There are several types of analysis possible on the data that are recorded. Reflection and refraction processing use the phase of the recorded signals to infer the P-wave velocity of the different subsurface media. This is similar to performing a tomography of the region. Waveform modeling uses both the phase and the amplitude of the signals to invert for a subsurface velocity model. However, full waveform modeling schemes are computationally expensive and a good starting subsurface model is needed to be able to
converge to a reliable solution. Next, we first discuss the active measurement set-up, the data quality, and put forth the arguments for performing seismic refraction processing. We also discuss the refraction tomography method we use for imaging the subsurface and the results of the data analysis are presented.

6.2 Seismic source

There are two well known methods of injecting a known source signal into the subsurface. The first one is the use of dynamite or other explosives buried in a shot-hole and detonated at a given point in time. While this process is effective in ensuring a strong source signal because the shot-hole can be made deep enough to reduce the energy attenuation by the soft soil, the intrusive nature of the method makes it disadvantageous. The second method uses a vibroseis system to inject signals into the ground. The latter was chosen as the source in our survey. A comparison of both seismic sources can be found in Drimi, 2004 [189]. A vibroseis is a surface source that emits seismic waves by means of a plate which is in contact with the ground. The driving force to the plate can be applied with a hydraulic system, by electro-dynamic means or even with magnetic levitation. The force necessary to hold the baseplate down on the Earth’s surface is given by what is called the reaction mass. For example in case of a hydraulic drive, oil is pumped alternately into the upper and lower chamber of the piston such that the baseplate is coupled to the ground at all times and as shown in Fig. 6.1(a). At all points the driving force on the baseplate is equal and opposite to the weight of the reaction mass. Additionally, a hold-down mass is used by the system to keep the baseplate in contact with the ground. Generally, the mass of the vehicle on which the vibroseis system is mounted is used as the hold-down mass. Although this mass helps hold the baseplate in contact with the ground, its effect on the baseplate output should be minimal. Hence it is vibrationally isolated from the system with an air spring system with a low spring constant. The resonance frequency of the hold down mass should also be lower than the minimum output frequency of the vibroseis. A detailed definition and working of the system can be found in Baeten, 1989 [190].

The signal injected by a vibroseis into the subsurface is called a sweep. The parameters defining a single sweep are the force on the baseplate, duration of sweep, lowest and highest frequency of interest, number of sweeps at one point, listening time and waiting time between sweeps. In the following sections we briefly state the impact of these factors on the recorded data quality.

6.2.1 Signal to noise ratio

Intuitively we can say that the SNR of the recorded signal from a vibroseis depends on the source duration parameters and as well on the receiver geometry. Assuming that the receiver geometry is fixed and the number of channels measuring the signal is limited, following Lanseley, 2009 [191] the SNR scales as

$$SNR = 20 \log_{10}(N_{VIBS} * F_B * \sqrt{N_{sweeps} \times L_{sweep}})$$  \hspace{1cm} (6.1)
where \( N_{\text{VIBS}} \) is the number of vibroseis, \( F_B \) is the force on the baseplate, \( N_{\text{sweeps}} \) is the number of sweeps at one point and \( L_{\text{sweep}} \) is the time duration of one sweep. For our case we have only one vibroseis and hence \( N_{\text{VIBS}} = 1 \). From Eq. (6.1) we see that the SNR scales linearly with number of sweeps and the sweep duration. The mini-vibroseis\(^1\) used in Limburg as shown in Fig. 6.1(b) could deliver a maximum force of 1200 N for P-waves and hence this force magnitude was used as the operating point. However, there are several challenges to employing longer sweep times. Firstly, it is problematic to dwell for long periods at low frequencies as it requires more energy and the displacement of the reaction mass becomes large. Secondly, the influence of unwanted transient noise in the signal during a long sweep can be problematic. During a long single sweep any short duration high amplitude noise will result in a high amplitude time-reversed replica of the source in the cross-correlated signal. The solution to such a problem is injecting many sweeps at one point. A stack of the signal (also known as the diversity-stack) for many such identical sweeps reduces the effect of such noise. Another important factor that plays a role in the selection of the number of sweeps and the sweep duration is the desired productivity of the field survey. Productivity of a survey is defined as the number of shot points covered during a day. Since we had to cover about 600 shot points over a period of nine days, a desired productivity of 70 shot points per day was needed. Hence in our survey we used a sweep length of 30 s and six such sweeps at each point. Another concern for long duration sweeps is an increased ground roll. By dwelling for a long time at the ground resonance frequency we may build up amplitudes and create a stronger ground roll. However, closed loop amplitude control of the sweep fundamental prevented higher ground roll magnitudes from being observed.

\[ \text{Figure 6.1: (a) Schematic view of the generation of the driving force in a hydraulic vibroseis. (b) A picture of the min-vibroseis along with the vehicle used for the measurements at Limburg.} \]

\(^{1}\)https://seismic-mechatronics.com/seismic-sources/lightning/
6.2.2 Frequency band

High-frequency waves give a better resolution for shallow imaging, while low-frequency waves propagate deeper and give useful reflections. The geology at Limburg and the presence of the bedrock at shallow depths with high elastic impedance requires a significant amount of energy at low frequencies for the signals to travel deeper than 40 m (depth to hard rock). A simulation of the acoustic P-wave equation was carried out using a preliminary P-wave velocity model of the region. A layered model of P-wave velocities was taken with the medium extending 600 m in horizontal and 400 m in vertical direction respectively. The first layer of the P-wave velocity model mimics the soft soil with a velocity of 400 m/s and extends up to a depth of 15 m. This is then followed by a 20 m thick layer of P-wave velocity 1800 m/s. Then the hard rock with P-wave velocity of 2600 m/s extends up to 180 m. Finally, the last layer with velocities of 4000 m/s continues into half-space. Two Ricker\(^2\) wavelet sources of peak frequency 40 Hz and 80 Hz were used for the simulation and the shot coordinate for both of them was (300 m, 10 m). Displacements corresponding to each of these sources were calculated at a depth of 10 m and at offsets of one meter.

The direct arrival in time-offset domain is a straight line and is marked as the blue-dotted line in Fig. 6.2(b). Important observations are the separation of shallow reflection events R\(_1\) and R\(_2\) from the first two interfaces which appear as smeared hyperbolas in Fig. 6.2(b) and more distinct in Fig. 6.2(c). On the contrary, the deeper reflections are better visible in the 40 Hz section than in the 80 Hz section. All reflection events represented as hyperbolas in the time-offset domain arrive withing the first 0.4 s. Multiple-reflected events from the reflectors are marked as M\(_1\) and M\(_2\).

Although the peak frequency of the source wavelet is important for greater depth of penetration of the body waves and better resolution, the baseplate force is an important factor too. For example the force imparted by a 10 kg weight-dropper [192] will attenuate very fast so that reflections from layers beneath the bedrock will not be measurable. Typically seismic

\(^2\)Time domain Ricker wavelet with peak frequency \(f_p\) is expressed as \(A(t) = (1 - 2\pi^2 f^2 t^2)e^{-\pi^2 f^2 t^2}\).
6.2 Seismic source

Vibroseis trucks are capable of delivering a baseplate force in the order of $10^4$ N. A maximum force of 1200 N in our case was a severe limitation. In order to account for this, we used a relatively longer sweep duration of 30 s. A linear sweep was used for the survey. The source frequency band was initially desired to be white in the frequency band 5 to 100 Hz to ensure good shallow reflector resolution as well as identification of deeper reflectors. However, due to the relatively smaller hold-down mass of the mini-vibroseis the lowest frequency had to be limited to 7 Hz. Additionally, for smooth operation a 10% time domain cosine taper was used at the start and the end of the sweep. Hence the peak force of 1200 N was only realized in the frequency band 12 to 95 Hz. Figs. 6.3(a) and (b) show the measured time domain signal of a single sweep and the corresponding amplitude spectrum respectively.

![Figure 6.3](image)

**Figure 6.3:** (a) Measured time series of the vibroseis sweep. (b) Amplitude spectrum of the sweep signal.

6.2.3 Duration parameters

The three time duration parameters for vibroseis operation and good quality of data are the sweep time, listening time and the waiting time. We discussed that the sweep time $S$ is based on factors like the total amount of energy that we desire to inject into the subsurface and the productivity of the survey. Listening time $L$ is the extra time period allocated for the geophones to record after a sweep has been completed. Hence, for each stretch of sweep we record the signals for a time period $S + L$. In principle, $L$ is the extra signal length that is necessary when we correlate the sweep of length $S$ with the recorded signals of length $S + L$. Hence the cross-correlated signal is only computed up to a maximum time lag $L$. From simulation results shown in Figs. 6.2(b) and (c) we know that most reflections will arrive within the first 0.4 s of the listening time. The direct arrival are the second fastest waves in a signal and for an offset of 1000 m should arrive within the first 3 s at maximum. The ground roll or the surface waves are the slowest propagating event in the seismogram. From passive seismic studies we already know that they have velocities as low as 150 m/s at 10 Hz. Using this velocity and assuming a maximum offset of 1000 m up to which high frequency surface waves might be visible, the listening time was fixed to $L = 7$ s.

The listening time is followed by a period of waiting time before we can proceed with the
next sweep at the same location. Selection of the waiting time is important for slip-sweep operations [193]. When a vibroseis generates a frequency $f$ because of harmonic distortion it also generates harmonic energy at frequencies $2f, 3f, ..., nf$. Since, we have a linear upsweep signal (meaning that the frequency of the source signal increases with time) the correlation of the sweep signal with that recorded by the geophones removes the harmonic energy due to its presence in the negative time axis of the correlation series. Hence, we did not need any waiting time after the listening period.

### 6.3 Source Receiver Geometry

Unlike passive seismic where a non-uniform sensor distribution is desired for sampling the seismic noise field over a broad frequency range, active seismic surveys are characterized by a regular sensor and source geometry. Sensors are laid out on the field with uniform sensor spacing along straight lines and the same goes for the source points. Distribution of shot points both along the receiver line (inline) and that orthogonal to it (crossline) are preferred. This is because most data processing of active seismic surveys are performed in the Common Midpoint (CMP) domain. For understanding terms like CMP and other terminologies, a few of the fundamental concepts related to active seismic data acquisition is discussed below.

- **Shot gather:**
  Given one shot location, the signals received by all the receivers for that particular shot when stored as a function of source-receiver offset and time is referred to as shot gather. The time domain signal recorded by a particular sensor is referred to as a trace. Hence, if there are $n$ sensors recording one particular shot or the vibroseis sweep, the number of traces for that particular shot gather would be $n$.

- **Common midpoint gather (CMP):**
  Every trace in a shot gather is characterized by the midpoint between the shot point and that particular receiver trace location. If the stratification of the subsurface would comprise flat layers then every such source receiver pair would give us a unique point of reflection for every reflector. For flat reflectors this would be the midpoint of the line joining the source and the receiver, although the depth of the reflection point can vary depending on the depth of the reflector. Traces sorted by this common midpoint form a common midpoint gather. It is necessary to note at this point that a CMP gather is not necessarily a common reflection point gather as in the case of dipping reflectors. The midpoint between the source and receiver is not equal to the point of reflection for a dipping reflector.

- **Foldage:**
  Foldage in active seismic is defined as the number of traces in one CMP gather. This is achieved by binning the survey area into square bins and then numerically computing the number of traces that fall in a particular bin. Sorting active seismic data traces into CMP gathers is a common convention, hence computing the data foldage prior to the survey is useful. Although each trace in a CMP gather corresponds to the same surface midpoint, each of them has a different source-receiver offset. Typical seismic data
processing steps like normal moveout correction (NMO) and dip movement correction (DMO) for a dipping reflector are executed in CMP domain followed by a process of stacking [194]. Hence the number of traces per CMP gather is a measure of the SNR achieved after stacking. Theoretically if the foldage of a CMP bin is $F$, then the SNR of the signals stacked for that CMP bin scales with $\sqrt{F}$.

![Figures 6.4](image-url)

**Figure 6.4:** (a) Receiver line R1 laid out on a map of the area with the red arrow pointing to the location where the borehole was drilled. (b) Source points corresponding to R1 on a map of the area. (c) Receiver layout in cartesian coordinates. (d) Two parallel line of source points separated by a distance of 10 m in cartesian coordinates. (e) Foldage map for receiver line R1 with a bin size of 3 m along x-y directions respectively.

Figs. 6.4(a) and (b) show the receiver layout and the source points corresponding to the first active seismic line R1. A total of 182 wireless geophones were deployed in a straight line with an spacing of 3 m. At some points the receiver interval was doubled owing to some physical constraints in deploying the sensors. A receiver spacing of 3 m was selected in order to avoid spatial aliasing of the ground roll which consists of slowly propagating surface waves. A receiver spacing of 3 m cannot sample surface waves with wavelength less than 9 m which corresponds to high frequency surface waves beyond 20 Hz. Hence we still expect some spatial aliasing of surface waves at high frequency and for small source-receiver offset data. Figure 6.4(c) shows the receiver coordinates on a cartesian plane. The intent was to position the sensors in a straight line such that the $y$-coordinate would always be constant. However, a rms of less than a meter could not be avoided. Corresponding to the receiver line R1, a total of 277 shots were fired using the vibroseis. The shot positions were divided into two parallel lines as can be seen in Figs. 6.4(b) and (d). The lines were separated by about 10 m of crossline offset. The objective behind firing two shot lines was to enhance the foldage of the data and as well as perform imaging using each of the shot lines separately. Shots were fired
at every alternate receiver location with a constant offset of 2.5 m along the crossline. The 2.5 m offset was introduced to avoid shooting exactly over the sensors. Source points farther than 200 m from point A and outside the receiver line were positioned at irregular intervals. These high-offset shot points are very useful for imaging deeper subsurface features. The distance between the farthest source point A in Fig. 6.4(b) and point C in Fig. 6.4(a) was 402 m. Similarly, points B and D were separated by a distance of 165 m. The receiver line R1 was 554 m in length. Ideally with 182 sensors and 3 m spacing it should have been 543 m, but since some sensors had to be placed irregularly and at double the planned receiver spacing at some positions, the length of the receiver line exceeded the ideal one. The maximum source-receiver offset for line R1 was 956 m. Fig. 6.4(e) shows the foldage map of the survey area with a bin size of 3 m in both inline and crossline direction. The red arrow shown in Fig. 6.4(a) shows the point of the intended drilling. From the foldage map, we see that we will collect full fold data in the vicinity of the drilling area.

Similar to the receiver line R1, a line R2 was shot 50 m north and parallel to line R1. A total of 179 sensors was deployed along line R2 with a spacing of 3 m and a total spread length of 548 m. A total of 230 shots were fired along two parallel lines separated by about 10 m and a shot interval of 6 m. The field near the drilling point shown with a red arrow in Fig. 6.5(a) could not be accessed by the vibroseis vehicle and hence a gap in the shot line is observed in Figs. 6.4(d) and 6.5(d). The distance between point E in Fig. 6.5(b) and G in Fig. 6.5(a) was 181 m. Points F and H were 100 m apart. The maximum source-receiver offset equaled 729 m. From both Figs. 6.4(e) and 6.5(e) we see that we get a maximum foldage of about 238

![Figure 6.5:](image)

Figure 6.5: (a) Receiver line R2 laid out on a map of the area with the red arrow pointing to the location where the borehole was drilled. (b) Source points corresponding to R2 on a map of the area. (c) Receiver layout in cartesian coordinates. (d) Two parallel line of source points separated by a distance of 10 m in cartesian coordinates. (e) Foldage map for receiver line R2 with a bin size of 3 m along x-y directions respectively.
which is typically quite satisfactory in seismic data acquisition. The objective behind conducting the seismic measurements along two separate receiver lines was to determine the lateral consistency subsurface images computed along each of the receiver lines.

### 6.4 Seismic data quality

Seismic data was recorded by the two receiver lines R1 and R2 for all the 277 and 230 shot points respectively as shown in Figs. 6.4 and 6.5. To give an impression of the quality of the data we consider two specific shot points, one with an end-on spread (receivers positioned on one side of the shot point) and the other with split-spread (receivers positioned on both sides of the shot point) geometry. Fig. 6.7(a) shows the source and receiver locations overlaid on the topography of the region corresponding to an end-on spread geometry. For clarity in understanding, we plot one of the vertical traces in Fig. 6.7(b) corresponding to a source-receiver offset of 212 m. Fig. 6.6 shows a single trace in the time domain recorded by a geophone which was at a distance of about 212 m from the source. P-waves propagating directly between the source and receiver are observed as the first burst of energy in the trace. These waves arrive at the geophone between 0.1 and 0.2 s after the shot is fired, hence propagating with speeds between 1,000 to 2,000 m/s. The Rayleigh waves which are the second most strong signals observed in the trace propagate the slowest. They arrive at the receiver between 1.2 and 1.5 s and hence propagating with speeds between 100 and 200 m/s.

As seen in Fig. 6.6, the signals recorded by the geophone between 0.3 and 0.5 s suffer from weak SNR. However, this part of the signal is the most important for observing signals that are reflected back to the Earth’s surface from the different subsurface layers. Hence, an automating gain correction (AGC) is applied to the data for better visualization [194]. Fig. 6.7(b) shows the AGC applied shot gather corresponding to a source x-coordinate of 1084 m in the time-distance domain. The signal recorded at each receiver which is also referred to as a trace is depicted at each receiver position in the form of a vertical wiggle. As mentioned

![Figure 6.6: A single time-domain seismic trace recorded for a source-receiver offset of 212 m corresponding to the shot location shown in Fig. 6.7(a). P-waves which travel directly between the source and receiver mark the first event in the trace, followed by guided waves and finally the Rayleigh waves.](image-url)
earlier, the first arrival in the shot gather is the direct P-wave arrival propagating at about 1,000 to 2,000 m/s up to source-receiver offsets of 250 m. A change in the time-offset slope of the direct arrival is observed beyond source-receiver offsets of 250 m and is due to the critically refracted P-wave overtaking the direct arrival. Along the time axis the direct arrival is followed by the arrival of several wavetrains or guides that are parallel to the direct arrival and hence propagating with the same velocity as the direct P-waves. These are the guided waves. Guided waves can be visualized as leaky modes and appear due to multiple reflection between the unconsolidated top soil and the hard rock. Since the geology in the region is characterized by a transition from soil to hard rock at approximate depths of 30 – 40 m, the presence of guided waves is obvious. Other expected features are the surface waves arriving late in the seismograms. Unlike guided waves, these waves are normal mode solutions of the wave equation. Reflection events which appear as hyperbolas in the time-offset domain are not observed or are obscured by the guided wavetrains.

In order to get a deeper insight into the data quality we perform a frequency-wavenumber transform of the data also known as the f-k transform. Fig. 6.8(a) shows the f-k transform of the raw shot gather. Ground roll and direct arrivals appear in the f-k transformed data as events with a linear moveout. Ground roll has a smaller slope in the f-k domain as it has slow propagation speeds. The direct waves are the ones that have a much steeper slope in the f-k domain and they encompass the outer edges of what is called the signal cone. If seismic reflections are present in the data they should be visible within this cone. Since reflections are propagating almost vertically up, they propagate with high apparent velocities. Fig. 6.8(b) shows the masked data in the f-k domain. Here an f-k filter with high velocity cut-off of 2500 m/s and a taper of 500 m/s is applied. The f-k mask is used to remove ground roll and direct arrivals that propagate with less apparent speeds. The part of the data that is retained is what we already stated as the signal cone. After masking the f-k transformed data, they transformed back to the time-offset domain and used for further processing. Figs. 6.9(a) and (b) show the data before and after f-k filtering. No reflections are visible in the shot records after f-k filtering.

Figure 6.7: (a) Source position and the receiver spread overlaid on the topography of the region. The source is located at a x-coordinate of 1084 m and the receiver spread is between 402 and 956 m. (b) AGC applied raw shot gather after corresponding to the source-receiver spread shown in Fig. 6.7(a)
We now carry out a similar analysis on a shot gather with split-spread geometry. Fig. 6.10(a)

![Figure 6.8](image)

**Figure 6.8:** (a) f-k transform of the raw shot gather. (b) Filtered shot gather in the f-k domain.

and (b) show the source-receiver geometry and the shot gather data respectively for split-
spread setting. As observed previously the direct arrival is followed by guided wavetrains
within the first 0.4 s of the data. The ground roll is much more dominant in this time-offset
section than in the last one as the geophones are spread closer to the shot point. The end-on
spread that we showed in the last example had a source-receiver offset of at least 200 m and
hence surface waves were significantly attenuated by the time they reached the first receiver.
Other notable features are the two separate time-offset moveouts of the ground roll. This is
the due to the presence of both the fundamental and the higher order mode of surface waves.
Hence Rayleigh waves propagate with two different velocities corresponding to the funda-
mental mode and the first overtone. Fig. 6.11(a) shows the f-k transform of the shot gather
that was shown in Fig. 6.10(b). Unlike the previous case, aliasing of the ground roll is clearly
observed. We stated previously that for smaller source-receiver offsets or higher wavenumber
range, aliasing of the ground roll cannot be avoided due to their low propagation speeds. The
aliased ground roll is also visible in the signal cone of the f-k transformed data.
Hence inspite of f-k filtering the aliased ground roll is difficult to remove for smaller source-
receiver offsets. Since these aliased events have a negative slope in the f-k domain, they
Figure 6.10: (a) Source position and the receiver spread overlaid on the topography of the region. The source is located at a x-coordinate of 740 m and the receiver spread is between 402 and 956 m. (b) AGC applied raw shot gather corresponding to the source-receiver spread shown in Fig. 6.10a.

Figure 6.11: (a) f-k transform of the raw shot gather which shows the ground roll characterized by a gentle slope in the f-k domain and its aliased parts which enter the signal cone and makes it hard to remove by masking. (b) Filtered shot gather in f-k domain with some aliased ground roll energy still left in the signal cone.

would appear as back scattered events in the time-offset domain. This is shown in Fig. 6.12. However, no dominant reflections are observed. From the solutions of an acoustic P-wave equation in a synthetic model we already know that if there are dominant reflections in the data, they should appear in the first 0.4 s of the data as shown in Fig. 6.2. Hence we can conclude that the reflections are entangled with the guided waves and too small in magnitude to be isolated from the guided waves. Due to this property of the shot records, seismic reflection processing was difficult to accomplish. Instead we use the recorded seismic data for seismic refraction processing. In the next section we describe the methodology of seismic refraction processing and the results we obtained.
6.5 Seismic refraction processing

6.5.1 Fundamentals

The signal injected into the subsurface interacts with the medium and after a complex process of reflection, transmission and refraction is received by the geophones on the surface of the Earth. The first wave to arrive at a receiver is either the direct P-wave or the refracted P-wave. For small source-receiver offsets the direct P-wave propagating straight between the source and the receiver is the first to arrive at the receivers. However, since the velocity of the medium increases as we go deeper, for long offset receivers there is a possibility that ray paths that propagate deeper can arrive before the direct P-wave. This is typically the case for critical refraction. If the Earth is visualized as a stack of horizontal layers with velocity continuously increasing downward then at a certain offset the ray incident on an interface will suffer critical refraction following Snell’s law. At this angle of incidence the ray travels along the interface of the two media while propagating with the velocity of the faster medium. Eventually after a certain offset this refracted wave on the seismogram precedes the direct wave. Fig. 6.13(a) shows the direct and the refracted wave propagation paths in a two-layer medium with P-wave velocity $V_{p1}$ and $V_{p2}$ respectively, such that $V_{p1} > V_{p2}$ and the thickness of the top layer is $h$. Receiver $R_1$ at an offset $x_1$ sees the direct P-wave as the first arrival whereas at receivers $R_2$ and $R_3$ the refracted wave propagating along the interface of layer 1 and 2 arrives first. The offset $x_2$ of receiver $R_2$ is also known as the cross-over distance. This cross-over distance is a function of the critical angle of incidence $\theta_c$ which further depends on the velocity of the two media. The cross-over distance $d_c$ can be expressed as

$$d_c = 2h \sqrt{\frac{V_{p2} + V_{p1}}{V_{p2} - V_{p1}}}.$$  \hfill (6.2)

Hence in case of a layer over a half-space velocity model, given the first arrival travel-times for a set of source-receiver offset data, it is possible to estimate both the depth of the top layer and also the velocities of the two media. The velocities are first estimated using the slopes of the time-offset curve and the depth can be calculated from the cross-over distance $d_c$. However, this treatment is fairly elementary and is a simple representation of the lithology.
of the subsurface which in reality might comprise a number of such layers. In the next section we discuss a generalized theory for dealing with such problems.

Figure 6.13: (a) Schematic diagram of a two-layer model showing the propagating rays for the direct and the refracted arrival between source and receivers. (b) First-arrival traveltime diagram as a function of source-receiver offset.

### 6.5.2 DeltatV method

DeltatV method [195] can be used for obtaining a shallow P-wave velocity model given the first-arrival P-wave traveltimes for a set of known sources-receiver locations. A velocity model is computed and updated iteratively to match the observed first-arrival traveltimes. The data that are collected on the field are in the form of shot records, that correspond to the first-arrival traveltimes for one source and all the receivers. Methods that work in the common shot domain are however subject to much ambiguity and human interpretation. Palmer 1981 [196] proposed a method of obtaining the velocity model by assuming constant velocity layers where the velocity was only allowed to change horizontally. However, it becomes evident that in cases of geological faults, pinch-out and other low velocity anomalies that depict sharp lateral velocity contrasts, the method will fail due to its fundamental assumption of horizontal stratification. One solution to this problem is to work in the CMP domain as is done for the case of seismic reflection processing. All shot-receiver combinations that fall in one CMP gather are sorted according to their offset and then processed. In this way we compute a velocity model for each particular CMP. Once this is accomplished we can combine the information from all the CMPs to obtain a so called 1.5 − D model of the subsurface.

The DeltatV method starts with the assumption that the velocity inside each layer is changing linearly either with a negative or positive slope. The utility of such an assumption is

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3 A pinch-out is a geologic phenomenon of successively wedge-shaped younger rock strata extending progressively further across an erosion surface cut in older rocks.

4 The Earth is represented as a 1D model, but waveform modeling is performed in 2D or 3D.
that rays propagating between source and receiver follow a circular path in a medium with linear velocity change \[197\], and hence making an analytic implementation possible. Such an assumption of velocity change is also reasonable as has been stated in Alekseev, 1973 \[198\]. Using circular ray paths it is possible to model the refracted rays and their traveltimes with simple parametric equations. The velocity \(v(z)\) inside such a layer can be expressed as \(v(z) = V_t + \gamma z\), where \(z\) is the depth, \(V_t\) represents the velocity at the top of the layer \(V_b\) and \(\gamma\) the rate of change of velocity inside the layer. Then following Gebrande, 1985 \[199\] the following relations hold:

\[
d = \frac{2V_b\sqrt{1 - \frac{V_t^2}{V_b^2}}}{\gamma} \quad (6.3)
\]

\[
t = \frac{2}{\gamma} \log\left(1 + \frac{\gamma\sqrt{1 - \frac{V_t^2}{V_b^2}}}{V_t}\right) \quad (6.4)
\]

where \(d\) represents the source-receiver offset, \(t\) is the ray travel time between source and receiver, and \(V_b\) is the velocity at the bottom of the layer. In context of the problem at hand, we know the values of \(d - t\) tuples for each CMP shot gather. The only unknowns are \(V_b\), \(V_t\) and \(\gamma\). One iteration of the DeltaT method determines each of the above parameters for one layer. At every iteration, it starts with first estimating the value of \(V_b\). From the \(d - t\) travel time curve, a candidate \(d - t\) tuple is determined that corresponds to the ray grazing the bottom of that layer. Now, one CMP gather has multiple \(d - t\) tuples and the selection of the appropriate \(d - t\) tuple for a ray grazing the bottom of the layer is based on the determination of points of inflection in the \(d - t\) curve. After the selection of the desired \(d - t\) point on the curve, a linear regression is performed using neighboring points on the curve. The slope of the curve gives an estimate of \(V_b\). The value of \(V_t\) is however defined implicitly by Eqs. (6.3) and (6.4). Hence \(V_t\) must be computed numerically. A Newton-Raphson root search scheme for the implicit function

\[
f(b) = \frac{t\sqrt{V_b^2 - V_t^2}}{d} - \arccos\left(\frac{V_b}{V_t}\right) = 0 \quad (6.5)
\]

is used to obtain the value of \(V_t\). The upper limit on the value of \(V_t\) is given by

\[
V_{t_{max}} = \sqrt{V_b \frac{d}{t}}. \quad (6.6)
\]

The Newton-Raphson method is hence implemented in the closed interval \([0, V_{t_{max}}]\). Candidate rays or \(d - t\) tuples for which \(V_{t_{max}}\) equals or exceeds \(V_b\) are skipped. After the values of both \(V_t\) and \(V_b\) have been determined, we compute the rate of change of velocity in the layer using the relation:

\[
\gamma = \frac{2}{d}\sqrt{V_b^2 - V_t^2}. \quad (6.7)
\]

Finally the layer thickness is obtained from

\[
h = \frac{V_b - V_t}{\gamma}. \quad (6.8)
\]
Fig. 6.14(a) shows a two-layer geological model with the depth of each layer given by \( z_1 \) and \( z_2 \) respectively. The source is represented as \( S \) and the receivers are located at distances of \( d_1 \) and \( d_2 \) from the source. The offsets \( d_1 \) and \( d_2 \) displayed in the figure are such that we get grazing rays in both cases. The ray traversing from the source \( S \) to the receivers as stated earlier are arcs of the circle \( C_1 \) and \( C_2 \) as shown in the figure. The corresponding travelt ime curve for such a hypothetical model is shown in Fig. 6.14(b). At offsets \( d_1 \) and \( d_2 \) the slopes are locally determined by implementing a linear regression among the points neighboring the offsets \( d_1 \) and \( d_2 \). This is used to determine the bottom velocity of that layer. The corresponding velocity model obtained for such a case is shown in Fig. 6.14(c). In the top layer the velocity changes from \( V_0 \) to \( V_1 \) and both values are derived as described earlier.

After the subsurface parameters for the top layer have been determined, the process implements a reduced version of the steps implemented earlier to determine the subsurface parameters for the next layer. The \( d-t \) tuples that were ignored in the last iteration due to their too high offset are now used in the analysis. These rays however have already propagated partly in the first layer and are grazing the bottom of the second layer. This problem is tackled by solving Eq. (6.3) and (6.4) by physically lowering the source-receiver coordinates to the top of second layer and then implementing the methodology as described earlier. The velocity at the top of the second layer \( V_2 \) need not be same as the velocity at teh bottom of the first of layer, as this method allows for jumps in velocity across layer boundaries. This process is repeated for multiple iterations until all the \( d-t \) tuples for the particular CMP have exhausted. The algorithm then moves to the next CMP and repeats the same set of steps as described earlier. Since there is no limitation on the velocity of the top of the succeeding layer, velocity-inversions can also be modeled correctly. However, if the trapped velocity-inversion is very thin and the receiver interval is not small enough, there might be problems resolving such thin layers. A small receiver separation corresponds to a higher ray density and hence the approach is sensitivity to subtle changes in subsurface velocity.

Figure 6.14: (a) An illustration of the ray paths traversed between a source and two receivers for a two-layer subsurface model with linear velocity-depth change. (b) Corresponding \( d-t \) tuples for the two-layer model showing the linear fits at offsets \( d_1 \) and \( d_2 \). (c) Output velocity-depth model.
6.5.3 Implementation and results

Inputs to the DeltatV method are the arrival times of the fastest propagating P-waves at each of the receivers, and the location of the source points. Selecting the first-arrival times from the shot records is accomplished by using a combination of an automatic scheme followed by a manual quality check. Since we have a total of 507 shot records, it was possible to check every first arrival times manually. The first stage in the automatic time selection scheme is preprocessing the data. The seismic source used in our studies has a white amplitude spectrum in the frequency band 12 to 95 Hz. While the ground roll dominates the low frequency content of the data, the first arrival times are concentrated in the frequency band 25 to 95 Hz. This is also observed in the f-k spectrum of the shot gathers shown in Figs. 6.8 and 6.11. Hence the data are first bandpass filtered in the frequency band 25 to 95 Hz. The automatic first-arrival selection scheme is based on computing the modified energy ratio (MER) of each trace in the shot record. This scheme has been successfully used in the geophysics community to detect the arrival-time of P-waves from earthquakes [200]. MER is a concept derived from the energy ratio (ER). ER for the \(i^{th}\) time sample in a seismic trace is computed as

\[
ER(i) = \frac{\sum_{k=i}^{i+L} x(k)^2}{\sum_{k=i-L}^{i} x(k)^2},
\]

(6.9)

where \(x(k)\) is the \(k^{th}\) sample of the seismic trace and \(L\) is a window length neighboring this \(k^{th}\) sample which is defined by the user. The MER for the \(i^{th}\) sample is then given by the relation

\[
MER(i) = |[ER(i).x(i)]|^3.
\]

(6.10)

We notice that the value of MER depends on the choice of the time window length \(L\). As a rule of thumb the value of \(L\) should always be less than half of the peak period of the source wavelet. For our case we selected \(L\) to be 0.01 s which is suitable for a carrier wavelet of peak frequency 50 Hz. The first peak in the MER time series is indexed as the arrival of the fastest wave at each sensor. This process is repeated for all the traces in the shot record and then stored along with their associated shot and receiver locations. It is necessary to store the shot and receiver locations so that we can later sort the trace and the arrival times in order to get the information in the CMP domain. Fig. 6.15(a) shows the shot and receiver locations for a split-spread geometry. The automatic selection of first arrival times and those selected manually are shown in Fig. 6.15(b). In principle, the MER method works accurately when the shot records are not very noisy. However, we do check every shot record manually for any inconsistencies. It is also worth noting that first-arrival times are not selected from the traces that are less than 25 m away from the source point. The first-arrival time resolution in that case is marred by the ground roll from the seismic source which is significantly higher in energy than the direct P-wave and dominates the signal amplitude.

First-arrival selection is carried out for all the 507 shot records. The selections are then divided into two sets corresponding to each of the receiver lines. Fig 6.16(a) shows the source and the receiver locations for line R1 and figure 6.16(b) shows the corresponding first arrival time picks as a function of x-coordinate for the shot records. For clarity in the figures
Figure 6.15: (a) Source position and the receiver spread overlaid on the topography of the region for a split-spread shot record. (b) First-arrival times from automatic and manual selection for the bandpass filtered shot gather.

we plot the first-arrival travel times for every fourth shot record. Similarly Figs. 6.16(c) and (d) show the source and receiver locations corresponding to receiver line R2 and the first-arrival times respectively. The first-arrival times are then sorted in the CMP domain and the

Figure 6.16: (a) Source position and the receiver spread overlaid on the topography of the region for receiver line R1. (b) First-arrival times as a function of the x-coordinate for every fourth shot record along receiver line R1. (c) Source position and the receiver spread overlaid on the topography of the region for receiver line R2. (d) First-arrival times as a function of the x-coordinate for every fourth shot record along receiver line R2.

DeltatV method is applied to the CMP gathers corresponding to each receiver line separately. Fig. 6.17(a) shows the receiver line R1 laid out on the map of the region. The results from the DeltatV method is shown in Fig. 6.17(b). Points A, B, C and D marked in Fig. 6.17(a) are the same as those marked along the offset in the output P-wave velocity model beneath the receiver line R1. An overthrust fault is visible in the velocity-depth section and marked with a dotted line in Fig. 6.17(b). Similarly the results for the second receiver line R2 are shown in Fig. 6.17(c) and (d). Points E, F, G, H in Figs. 6.17(c) and (d) are identical points. These points are marked to precisely point to the location of the fault in the region. Since, we aim to conduct drilling upto depths of 300 m, a-priori information about subsurface faults are
vital. Previously, a drilling campaign was conducted at the site in the year 2017, but it had to be stopped due to snapping of the drill-bit at a depth of 170 m. The drill string could be retrieved down to a depth of 140 m, but below that depth the drill-stem was stuck and hence the drilling had to be stopped. Since we aim to conduct another drilling campaign to reach upto depths of 300 m, the information about the location and characteristics of the subsurface fault is instrumental in aiding the drilling process and also in the selection of the drilling site. Additionally, the P-wave velocity profile in Fig. 6.17(b) indicates a raised bedrock between points C and D. This could also be intuitively concluded from Fig. 6.15(b), where the first-arrival times between 700 and 800 m have a smaller slope in the time-offset domain.

![Figure 6.17](image)

**Figure 6.17:** (a) Receiver line R1 shown on a map of the region. (b) Derived P-wave velocity model beneath receiver line R1 with the dotted line showing the dip of the overthrust fault. (c) Receiver line R2 shown on a map of the region. (d) Derived P-wave velocity model beneath receiver line R2 with the dotted line showing the expected dip of the overthrust fault. However, (b) and (d) does not show the same subsurface structure which may be due to artefacts introduced by the DeltatV method.

### 6.5.4 DeltatV pitfalls

The DeltatV method computes the velocity model beneath every CMP separately, and therefore might lack lateral continuity in the velocity model obtained from consecutive CMPs. While this is one of the advantages for modeling geological faults and pinchouts, the model obtained from the DeltatV method is vulnerable to artefacts in the resulting velocity model.
This is specifically observed in the results in Fig. 6.17(d) at depths between 100 and 150 m and between horizontal offset points $F$ and $G$. A high velocity artefact is observed above the fault plane of the overthrust fault. These artefacts could be due to errors in the travel time picks. Typically the models obtained from the DeltatV method serve as a good starting model for further processing of the data using other refraction processing methods. In the next section we discuss Wavepath Eikonal Tomography (WET) and the results obtained with that method.

### 6.6 Wavepath Eikonal Tomography

General ray tracing schemes do not account for the finite frequency effects of wave propagation. It models the travel times between the source and receiver based on a single ray between them. Such rays are assumed to be propagating with infinite frequency and are sometimes referred to as a mathematical rays. In reality, the propagation characteristics of a ray between a source and the receiver are affected by physical structures present in the vicinity of the mathematical ray. One other factor can be the very bandlimited nature of the source signal. This is where the concept of Fresnel volume gets introduced. Kravtsov and Orlov, 1979 [201] defined the Fresnel volume as the region around the mathematical ray that contributes to the ray’s propagation characteristics. This region around the mathematical ray is referred to as the physical ray. Consequently, a high frequency signal will have a narrow-width Fresnel zone and vice versa for low frequencies. For example in a homogeneous medium the ray path between two points is just a straight line joining the two points and the corresponding Fresnel volume is an ellipsoid with the two points as foci. Fig. 6.18 shows the schematic representation of the Fresnel volume for a ray propagating between points $A$ and $B$. The plane $\Sigma_F$ intersects the ray path between $A$ and $B$ such that the normal to the plane $\Sigma_F$ is the same as the local ray direction at the point of intersection $O$. Point $F$ on the plane is part of the Fresnel volume if it satisfies the following traveltime equation

\[
|t(A, F, f) + t(F, B, f) - t(A, B, f)| \leq \frac{T}{2}, \tag{6.11}
\]

where $f$ and $T$ are the frequency and the time period of the propagating wave respectively. Such Fresnel volumes exist for both direct and reflected rays. While for a homogeneous medium or for a layer over a half-space, computing the shape of the Fresnel volume is simple and analytically possible, for a complex medium of propagation with lateral inhomogeneity the shape of the Fresnel volume is complex and needs to be computed numerically. The advantages of using the Fresnel volume approach to computing the source-receiver traveltimes is because it accounts for some of the effects such as shadow zones and multipathing that are not modeled by mathematical rays [202]. However, the disadvantage of such an approach is the computational cost of the method. A finite difference solution to the Helmholtz equation for every frequency and every source-receiver pair needs to be computed. The WET method is similar to the method proposed by Harlan, 1990 [203] as it forward models the traveltimes between the source and the receiver by solving the Eikonal equation [204] instead of the Helmholtz equation. This makes the method significantly faster while still accounting for the wavepath effects following the propositions of Woodward and Rocca, 1988 [205]. In the next section we discuss the algorithm for performing the WET on our data.
6.6 Wavepath Eikonal Tomography

Figure 6.18: Schematic representation of a Fresnel volume $\Omega_F$ for a source and receiver positioned at A and B respectively.

6.6.1 WET algorithm

Unlike the DeltatV method which works in the CMP domain, the WET method takes into account all the ray paths between all combinations of source and receiver of the data simultaneously. Hence the method starts with an initial 2D or 3D velocity model and forward models the first-arrival traveltimes. After the traveltimes have been forward modeled, the phase residuals between the observed and the modeled travel times are back projected along the modeled ray volumes to update the velocity model at each iteration. The phase misfit function $\epsilon$ is expressed as

$$\epsilon = \frac{1}{2} \sum_s \sum_r \sum_\omega R_{rs} \Delta \phi(x_r, x_s, \omega)^2,$$

where the summations account for all combinations of source-receiver indices and all frequencies of interest $\omega$. Here $\Delta \phi(x_r, x_s, \omega) = \phi_{cal}(x_r, x_s, \omega) - \phi_{obs}(x_r, x_s, \omega)$ is the difference between the calculated and the observed phase, also referred to as the phase residual. The weighing function $R_{rs}(\omega)$ is directly related to the amplitude spectrum of the source wavelet. The introduction of the term $R_{rs}(\omega)$ while computing the phase misfit ensures that the gradient of the reconstructed velocity field is smooth in a way that is consistent with the source spectrum and the path of propagation. Inclusion of this term in the phase misfit function is also pivotal in introducing the concept of Fresnel volumes while computing the theoretical first-arrival traveltimes.

Computation of the phase residuals is followed by an update to the velocity or the slowness model for the next iteration. In order to update the slowness field the next step involves computing the gradient $g(x)$ of the phase misfit function $\epsilon$ with respect to the slowness parameter $s(x)$ which is the inverse of the velocity. The gradient $g(x)$ is then expressed as

$$g(x) = \frac{\partial \epsilon}{\partial s(x)} = \sum_s \sum_r \sum_\omega R_{rs}(\omega) \frac{\partial \phi_{cal}(x_r, x_s, \omega)}{\partial s(x)} \times \Delta \phi(x_r, x_s, \omega).$$

The slowness field at the $(k+1)^{th}$ iteration is then updated using the steepest-gradient-descent method as

$$s(x)^{(k+1)} = s(x)^{(k)} - \alpha^{(k)} g(x),$$

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where $\alpha^{(k)}$ represents the steepest-gradient-descent step length at the $k^{th}$ iteration. The method for computing the partial derivative of the phase misfit function with respect to the slowness parameter in Eq. (6.13) can be found in Schuster, 1993 [206]. The slowness field is updated at each step until the rms travel time error between the modeled and the observed travel times converge. A limit on the maximum number of steepest-descent iteration can be set such that the algorithm stops when reaching the maximum iteration limit. The numerical stability of this method when applied on synthetic and crosswell data has been discussed in detail in Schuster, 1993 [206].

### 6.6.2 WET results

The inputs to the WET algorithm are the first arrival travel times for every source-receiver pair and also an initial 2D P-wave velocity model of the subsurface. The P-wave velocity model of the subsurface computed with the DeltatV method is used as the starting model for the WET algorithm. The physical paths are constructed with a Ricker wavelet of peak frequency 50 Hz, which was chosen based on the average spectrum of the signal received at the receivers. A peak frequency of 50 Hz does not mean that other frequencies do not contribute, but that the maximum contribution is observed around that frequency. Figs. 6.19 (a) and (b) show

![Figure 6.19](a) Receiver line R1 shown in a map of the region. (b) WET P-wave velocity model beneath receiver line R1 with the dotted line showing the dip of the overthrust fault. (c) Receiver line R2 shown in a map of the region. (d) WET P-wave velocity model beneath receiver line R2 with the dotted line showing the expected dip of the overthrust fault.

the receiver line R1 and the output P-wave velocity model from the WET method. Points A,
6.7 Conclusion and Recommendations

B, C, and D marked on Figs. 6.19(a) and (b) are identical points. They work as markers to identify the exact surface location of the fault. An rms traveltime residual of 1.68 ms (an error of less than 1% with respect to the observed traveltimes) is obtained at the end of 20 WET iterations. Similar results are shown in Figs. 6.19(c) and (d) but corresponding to receiver line R2. This receiver line extends further North as compared to R1. The increase in the bedrock depth North of line R1 is more pronounced and can be seen in the P-wave velocity images. However the fault-plane for both the lines shows a similar dip and ensures a lateral continuity between the two velocity profiles. The travel time rms residual for receiver line 2 was higher than for line R1. This is the reason that the image resolution in Fig. 6.19(d) is smaller than for the image shown in Fig. 6.19(b). Figs. 6.20(a) and (b) show the ray count per pixel attained while imaging the subsurface beneath receiver line R1 and R2. This allows to assess how well the subsurface was illuminated and how reliable the tomography results were. For both receiver line R1 and R2 a good ray count between 300 and 350 was achieved upto depths of 150 m.

![Figure 6.20: (a) Ray count per pixel beneath receiver line R1 and (b) R2 showing that the area around the overthrust fault-plane is well illuminated by rays. Points marked as A, B, C, D etc. are the same as the points marked in Fig. 6.19](image)

6.7 Conclusion and Recommendations

We discussed the application of seismic refraction methods for imaging the shallow subsurface. The original objective of the active seismic measurements was to perform reflection processing of the data. However, due to the weak source strength and the particular geology of the region, deep subsurface reflection signals could not be isolated from the guided wave-trains. The high impedance contrast between the soft soil and the hard rock at a shallow depth of 40 m was the reason for the observation of the guided waves. Nevertheless, since the data acquired had large source receiver offsets and the first arrival energy was sufficiently strong to be recorded approximately a kilometer away from the source, we could use the first arrival traveltimes for performing seismic refraction tomography. We explored two methods known as the DeltatV and WET for analyzing the data. The DeltatV method was found suitable for obtaining an initial P-wave velocity model of the subsurface without any manual interven-
tion. Although the results from DeltatV method suffer from artefacts, they still provided an acceptable starting model (close to the actual geology of the region) for the WET method. If the initial model is not close to the actual geology of the region then WET may oversmooth the solution leading to missing out on important subsurface features. In our case we believe that the final P-wave velocity model that we obtain from the WET method is satisfactory and that it produces a clear image of the subsurface overthrust fault. Since the tomography was carried out along two receiver lines parallel to each other we could check for consistencies in the subsurface images obtained for both receiver lines. As the separation of the two receiver lines was only 40 m, we observe that the fault is located at the same physical point along both the receiver lines.

The drilling operation performed in 2017 was between the two lines and exactly above the fault plane. The top 100 m of the subsurface did not cause much problem as the fault is not pronounced at those depths. It is only between 100 and 150 m where the dip of the fault can cause a problem. The overthrust fault observed at the site can be interpreted as one part of the bedrock thrusting itself above the other part. What this leads to is a low velocity intrusion into the high velocity hard rock. As a result one part of the bedrock is raised and the other part lowered on each side of the fault-plane. Since part of the soft soil will always intrude into the harder bedrock in such a situation, there is a chance that the drill bit turns and orients itself along the fault plane such that it drills into the softer soil first. When this rotated drill-bit encounters a hard rock interface, the force it exerts on the hard rock is no more vertical but is along the dip of the fault and this increases the chance of the drill-bit snapping. This occurred with the last drilling operation that failed and we could drill only upto depths of 170 m and obtain subsurface logging information down to a depth of 140 m. The seismic survey results now gives us a clear picture of the subsurface and where we want to drill. It is recommended that the drilling be performed further South where the bedrock is almost horizontal and slightly raised compared to the location where the first drilling operation was executed. The main objective of drilling the borehole is to position a seismometer at a depth of about 250 m. Hence a higher bedrock depth at the point where the seismometer will be placed is also very suitable for attenuation of high-frequency noise originating at the surface.

Caveat of the images obtained using automated seismic refraction methods is that it yields a subsurface velocity model with a continuous velocity gradient within a layer. Although we stated previously that this assumption is not far away from what we observe in reality, lithological interfaces obtained from these images are not sharp. This is the reason geophysicists prefer reflection interpretations to refraction. Reflection processing provides an output P-wave velocity model and as well an image of the subsurface with well defined interfaces. However, the velocity models obtained from refraction data analysis can also also be useful for migrating the seismic reflection data from the time to the depth domain. In the data that we collected at Terziet we cannot decipher the reflections due to high magnitude guided waves arriving at the receivers in the same time interval. We already observed that normal processing schemes which operate in f-k or \( \tau - p \) domain fail to remove such signals from the data. This surely opens up possibilities to explore methods that may allow to isolate the guided waves from the reflections.