# FULL WAVEFORM INVERSION IN CRYSTALLINE HOST ROCK: ANALYSIS OF RECEIVER COUPLING

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# ABSTRACT

In order to explore the application potential of full waveform inversion (FWI) in crystalline host rocks we acquired vibroseis seismic data in the GFZ-Underground-Laboratory of the research and education mine "Reiche Zeche" of the TU Bergakademie Freiberg. 76 shot positions and 30 3-C receivers were mounted around three galleries. In the elastic FWI only the first arrival P-waves are considered. A synthetic FWI test was successfully performed for a random medium velocity model describing the local gneiss. The used geometry was the same as in the field measurement. The starting model was a smoothed version of a travel time tomography of the true model. The preprocessing of the data mainly comprises a windowing of the first arrival P-wave. After 75 iterations the inversion converged. The inverted model is very similar to the true model.

In the field data inversion the source signal and the receivers coupling is unknown. In order to analyze changes of the wave field related to source and receiver locations a new semblance analysis is developed. The analysis revealed that wave field changes are mainly associated with the receiver positions. One pronounced receiver related effect is a ghost reflection at the tunnel surface which varies significantly with the depth of the receiver that alternate from 1 - 2 m. A synthetic forward modeling study using three different scenarios revealed that the best agreement between synthetic and observed data is obtained if the tunnel wall is not considered, i.e. no ghost reflections are modeled. Therefore a new deconvolution strategy was developed to correct receiver side effects and to remove ghost reflections. In the deconvolution the desired response is the simulated data without tunnel surface. After deconvolution the coherence between all receivers is improved and ghost reflections are reduced. It remains to be demonstrated that the residuals between synthetic and field data are caused by the velocity model and not by further source or receiver related effects.

## INTRODUCTION

In this work we explore the potential of the time-domain Full Waveform Inversion (FWI) in crystalline host rock under almost laboratory conditions. We use the infrastructure at the GFZ-Underground-Lab and performed multi-component vibroseis measurements in crystalline rock (Freiberger Greygneiss) within the research and education mine "Reiche Zeche" of the TU Bergakademie Freiberg, Germany. The FWI is an iterative minimization method to reduce the difference between computed and observed wavefields. FWI has the potential to use the full information content of multi-component seismograms and leads to a significantly higher resolution of seismic velocity models compared to traveltime based methods. The challenges in the GFZ underground lab are (1) the pre-processing and incorporation of the vibroseis source characteristics into the FWI workflow, (2) strong small-scale medium heterogeneities within the crystalline rock mass close to the receivers and (3) ghost reflections. The inversion scheme, the pre-processing of the data and synthetic inversion results are shortly described in the first part. In the second and third part, the real data is presented and solutions for the occurring problems are proposed.

# THE FWI METHOD AND SYNTHETIC CASE STUDY

FWI aims to minimize the misfit between modeled and observed seismograms. This can be implemented efficiently using a conjugate gradient approach (Tarantola, 1984; Mora, 1987). The adjoint-state method enables to calculate the gradient of the misfit function as zero-lag cross-correlation between a forward propagated wavefield and a wavefield propagated back from the receivers into the medium. We use the time-domain 2-D elastic FWI code developed by Köhn (2011), which is available under the terms of GNU GPL at www.opentoast.de. For the inversion of crystalline host rock vibroseis data this code was further enhanced by implementing different misfit definitions and especially a source wavelet inversion. The source wavelets are estimated by a stabilized deconvolution of the recorded data with the simulated data for the current subsurface model.

To explore the possible resolution of FWI in the Freiberger underground environment, the FWI workflow is first tested with synthetic data. This simulated scenario is very close to the real measurements in the GFZ-Underground-Lab. 76 shot position with an average spacing of 2 m and 30 receiver positions around the three galleries were simulated (Figure 1). The receivers GP1 - GP15 along the "Richtstrecke" are drilled alternately 1 m respectively 2 m deep in the host rock. Because of the almost plane transmission geometry of sources and receivers, the 2-D elastic FWI method was chosen. To get a better understanding of the involved wavefields we show trace-normalized correlograms of a common receiver gather (CRG) for Receiver 1 (Figure 1, GP1) in Figure 2. The dominant wave type recorded close to the source is the tunnel-surface wave. For the shots on the opposite site of the Freiberger Gneiss the first arrivals are build mainly by the directly propagating P-wave. For the Shots SP50 - SP55 the direct S-waves exhibits increased signal amplitudes caused by the specific radiation characteristic of the vibroseis source. Due to the normalization of traces to their maximum amplitude this leads to apparently smaller signal amplitudes of the direct P-wave.

Using the 2-D elastic forward modeling performed during the FWI we cannot simulate the tunnel surface wave properly (Jetschny et al., 2010). We thus restrict the FWI to the directly transmitted P-waves highlighted in Figure 2. All other traces are not inverted. We apply a time window around the directly traveled P-wave to damp later phases. Additionally, traces are normalized to their maximum amplitude before going into the FWI (Sheng et al., 2006). The Freiberger Greygneiss exhibits heterogeneities on all scales and can be approximated by a random medium model. We choose a correlation length of 10 m and a Gaussian velocity distribution with a standard deviation of 5% and an average P-wave velocity of 6000  $\frac{\text{m}}{\text{s}}$  (Kneib, 1995). The S-wave velocity model is calculated via  $v_s \approx \frac{v_p}{\sqrt{3}}$ . The density model is kept constant at 2550  $\frac{\text{kg}}{\text{m}^3}$ . The starting model is obtained by travel time tomography of the transmitted P-waves. The model update was only allowed within the gneiss block.

In this synthetic study we used a Klauder wavelet as the source time function extracted from the field data and filtered it with a butterworth lowpassfilter of 750 Hz. The required 74 iterations of the FWI took 12 hours on 96 cores. In Figure 3, we compare the true model, the starting model obtained by a traveltime tomography and the FWI result. In the inverted model the large scale structures could also be reconstructed very well. Even inside these structures many detailed velocity distributions are similar to the true model. The typical structure size of objects which cannot be reconstructed by the FWI is approximately 4 to 5 m and lies in the range of the smallest wavelength. Nevertheless, some artifacts at the receiver positions and high velocity contrasts close to the air-rock interfaces can be observed. In the upper most part, the inversion did not update the model because of zero ray coverage.

Altogether we can conclude that the FWI was able to resolve the small scale structures with the expected resolution..

# CHALLENGES FACED BY FIELD DATA

#### Semblance analysis

In crystalline host rock a significant influence of small scale heterogenities in the vicinity of both the source and the receiver is often observed. We refer to these effects as source and receiver coupling. In the synthetic example we assumed a known source signal and a perfect coupling of sources and receivers. To explore the relative contributions of source and receiver effects we compare waveforms in the CSG and CRG. As a measure of coherence between seismograms we use the semblance.



**Figure 1:** Top view of the GFZ-ULab: 30 3-C receiver (GP1 - GP30, inverted triangles) and 76 shot positions (SP1 - SP76, stars) are located along the three galleries. The encircled source locations (SP18 - SP20) and receiver location (GP10) get lost due to excavation of incline and chamber which was built after the meausurement (Krauss, 2013).



**Figure 2:** Correlograms of a common receiver gather (CRG) for Receiver 1 (Figure 1, GP1). The traces are normalized. The dominant wave type is the tunnel-surface wave for shot numbers below 28 (highlighted by an dashed circle). Shot numbers above 36 show the transmitted P-wave as first arrival (highlighted with a doted circle).



**Figure 3:** Result of the synthetic case study. Comparison of P-wave velocity models: a) True model: A Random media model with a velocity perturbation which simulates a crystalline host rock. b) Starting model: The model was derived from a first arrival travel time tomography of the true model. c) FWI-result: Inverted P-wave velocity model after 75 iterations using the first arrival P-waves only and a bandpass filtered Klauder wavelet with a corner frequency of 750 Hz.

In order to calibrate the semlance analysis we also test our new seblance analysis with synthetic data. Three different scenarios are considered with the same velocity model we used for the synthetic study. In the first scenario we use a 2-D model without any tunnel. The second scenario consists of a 2-D model with plane tunnel surfaces perpendicular to the propagation plane. In the third scenario we simulate a 3-D model with cylindrical tunnel surfaces.

For the field data and for the three scenarios we consider the semblance analysis for the first arrival Pwave. The same preprozessing as described above has been applied. For the semblance analysis the energy of each trace d is normalized, so that effects of the radiation pattern and the offset-dependent amplitude decrease is removed.For two equal seismograms the maxium value of the semblance is one. A semblance based cross-correlation is performed for all possible combinations of sources and receivers. The maximum value of the correlograms are summed and normalized as follows::

$$SEM = \frac{\sum_{n=1}^{n_r} [max(\frac{d_i \oplus d_j}{ns})]_n}{n_r} \tag{1}$$

where  $n_r$  is the number of possible combinations,  $\hat{d}_i$  respectively  $\hat{d}_j$  is the energy normalized trace, ns is the number of samples,  $\oplus$  denotes the cross correlation and  $i, j = 1, ..., t_{max}$  the number of traces for each CRG or CSG.

In Figure 4, the semblance for the field data for the CSG and the CRG are shown. The two most important areas of the plots are in the left upper and right lower part. For these traces, most of the shots and receivers are close to each other and on the same side of the investigated block. Also, in these parts the total number of correlations is much higher for nearby shots and receivers. For some shot and receiver combinations, no semblance could be calculated due to muted traces.

For the CSG (Figure 4a) some white crosses are apparent, e.g., for Shots 34 or 74. Obviously these shots differs most. In general, the overall reproducibility of the vibroseis signal seems to be very consistent. For the CRG in Figure 4b also white colored crosses like in the CSG are observable, e.g., Receiver GP7, GP26 or GP29. Another more crucial effect is visible: For the Receiver GP1 - GP15 in the CRG, the pattern



**Figure 4:** Semblance values for a) different shot pairs and b) different receiver pairs for the field data. The colorbar is clipped at the lowest value. For some shot and receiver pairs no correlograms could be calculated (yellow parts) because of muted traces. In the title the SEM value (Equation (1)) and the maximum possible value is shown.

is like a checkerboard. Just the waveforms of every second receiver fits to each other. This checkerboard pattern corresponds to the depth of two adjacent receivers varying from 1 - 2 m or vice versa. This might be caused by ghost reflections at the tunnel wall as the distance to the tunnel surface between neighbouring receivers changes by  $\pm 1$  m. All other receivers area in the same depth of 2 m, except of Receiver 25. For those receivers, the geometry and the velocity pertubation within the host rock is responsible for the different waveforms.

We thus investigate the influence of the ghost effect for the three different synthetic scenarios and compare the results with the field data for Receivers 1 to 15. In Figure 5 the semblance values for the three above described scenarios are shown. In scenario 1 (2-D model without tunnel, Figure 5a) no ghost reflections from the tunnel air interface is simulated. In scenario 2 (2-D model with plane tunnel surfaces perpendicular to the acquisition plane, Figure 5b) scenario 3 (3-D model with rectangular tunnel surface, Figure 5c) the checkerboard is very distinct. For a 3-D simulation with a finite tunnel structure the checkerboard is also obvious but not that dominant. Nevertheless, the best agreement with the observations might be achievied if we assume no tunnel in the forward calculations of the wavefields. This can further be emphasized by the accumulated semblance value. For the 2-D model without tunnel the value results in 581/711. The field data gives 571/711.

The comparison of the semblance plots of the field data (Figure 4) and the synthetic scenarious (Figure 5) revealed that the best agreement is achieved if the tunnel is not considered in our forward modelling. In other words, the best way to account for the ghost reflection is to try to remove them in the field data for example by deconvolution. Although the actual velocity model is not the same for the synthetic scenarios, the influence of the velocity distribution is neglectable in contrast to the different receiver depth.

The effect of the ghost reflection is also visible in the waveforms. In Figure 6 two seismograms of the field data and the synthetic data for scenario 1 and 2 are shown (Figure 1, GP4 and GP5). The data is preprozessed as described befor and the amplitude is normalized for each trace. The source was directly on the opposite site of the gallery in a transmission geometry (Figure 1, SP60). The two receivers are drilled in the host rock with different depths: Receiver 4 is drilled with a depth of 1 m and Receiver 5 with a depth of 2 m.

First of all, the field data and the synthetic data matche quit well. Modeled with the tunnel model, in



**Figure 5:** Semblance values for three different synthetic scenarios. a) For a 2-D model without a tunnel. There is no tunnel-air interface. b) For a 2-D model with a tunnel. The tunnel wall act like an infinite interface. c) For a 3-D model with finite tunnel height. In the title the summed up values and the maximum possible values are shown.

the black seismogram of Receiver 5 it is noticeable that the first peak is splitted in two peaks in contrast to Receiver 4. This is typical for the ghost reflection. Depending on the considered receiver, the two peaks can also be plurred into each other. Both, in the field data as well as the red seismograms, modeled without a tunnel, the double peack is not observable in both Receivers.

In the next step we describe a deconvolution approach to remove possible receiver coupling effects including the ghost reflection in the field data.



**Figure 6:** Three seismograms for two different Receivers (GP4 and GP5) for one Shot (SP60). GP4 is drilled with 1 m depth and GP5 with 2 m depth. For Receiver 5 and the 2-D model with tunnel (black seismogram), the ghost reflection can cleary be seen. This is not the case for Receiver 4. Also for the field data (grey seismogram) as well as for the synthetic data modeled without a tunnel (red seismograms) for both receivers no ghost reflection is visible.

# **Receiver site effects**

To reduce the different couplings of the sources, Pratt (1999) introduced an iterative source time function inversion which is similar to a deconvolution.

In the synthetic example we assumed a known source signal and a perfect coupling of sources and receivers. We assumed no unknown small-scale heterogeneities (fractures, tunnel surface topography) in

the vicinity of the sources and receivers. As we can see in Figure 4, unfortunately these assumption are invalid for the real measurements. Especially at the receivers we observe a site response which differs significantly for some receivers location. Previous studies and test measurements revealed that the receiver coupling is likely caused by small-scale heterogeneities such as fractures or local tunnel wall topography in the immediate vicinity of the receivers. During the drilling of the borehole, small-scale fractures around the anchor and the geophone at their tips may be generated. The effect of the small-scale heterogeneities close to the receivers on the recorded transmitted P-waveforms can again be studied in Figure 7a. This time we have a look hat the seismograms. The site effect is mainly observed on the receiver site, i.e., in the CSG. The CRG shows relatively coherent P-wave forms for each shot indicating a good reproducibility of the vibroseis source for different shot locations (Figure 7b).



**Figure 7:** a) Exemplary common shot gather (CSG) for Shot 56 and b) common receiver gather (CRG) for Receiver 3. c) Same CSG as in a) after the correction of the receiver coupling.

From Figure 7 it is obvious that an inversion and correction of the receiver coupling is required for FWI. The approach is similar to iterative source time function deconvolution introduced by Pratt (1999) for common shot gathers which is generally required in field data applications. Here we apply the deconvolution approach to common receiver gathers to remove the effects of small-scale heterogeneities close to the receivers (Maurer et al., 2012). In our approach we minimize the least squares misfit function for each frequency  $\omega$  in the Fourier-domain

$$s(\omega) = \frac{\vec{u}^t \cdot \vec{d^*}}{\alpha + \vec{u}^t \cdot \vec{d^*}} \tag{2}$$

where  $\vec{u}$  denotes the synthetic data and the vector  $\vec{d}$  the field data of a CRG. The superscript t is the

transpose operator and the superscript \* means the complex conjugated.  $\alpha$  is a water-level parameter which is necessary for a stable deconvolution to avoid small values in the denominator. The gained filter functions for the receivers are then convolved with the field data (Figure 7c) in the time domain. In contrast to a source time function inversion, the field data is fitted to the synthetic data because the synthetics are free of any receiver effects. In our case, we choose the synthetic data which was modeled without any tunnel. In Figure 7c the same CSG as in 7a is shown after the receiver coupling correction. The coherence of the waveforms has been improved and the receiver coupling effect is eliminated.

A semblance analysis is repeated with the seismograms in Figure 7c. We expect a better coherence especially for Receiver 7 which clearly sticks out in Figure 4a. Figure 8 shows that the checkerboard pattern was significantly reduced. The coherence between the receivers is more distinct and no receiver is highlighted.



**Figure 8:** Semblance values for different receiver pairs for the field data after the receiver coupling correction. By a comparison with Figure 4b, the coherence is much higher.

### SUMMARY

For measurements and studies in the underground a block of crystalline rock (Freiberger Greygneiss) in the GFZ-Underground-Laboratory within the research and education mine "Reiche Zeche" of the TU Bergakademie Freiberg, Germany, was chosen. With the local infrastructure seismic measurements with 30 multi-component receivers and 76 vibroseis sweeps as sources were performed. With synthetic test small scale structures of a randomly distributed velocity mdoel could be reconstructed very well. The used source wavelet, a Klauder wavelet, was extracted from the field data.

To invert for the field data we first have to identify effects in the data we are not able to model correctly. For this, a semblance analysis was performed comparing the waveforms of the field data and three different modeled scenarios. For synthetic velocity models which includes a tunnel, both in 2-D as well for 3-D models, the dominate effect for the first arrival is the ghost reflection at the tunnel wall. Another characteristic of the field data is that the recorded waveforms differ significantly from some receivers, even if they are very close to each other. This is caused by individual unknown medium heterogeneities at the receivers locations. By a receiver coupling correction deconvolution, the various couplings of the receiver could be successfully corrected. A greater coherence is visible in the wave fields and the data is now processed to be inverted with the FWI. It remains to be demonstrated that the residuals between synthetic and field data are caused by the velocity model and not by further source or receiver related effects.

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