2-D ELASTIC FULL WAVEFORM INVERSION FOR A TRANSMISSION EXPERIMENT IN CRYSTALLINE ROCK – A SYNTHETIC STUDY

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ABSTRACT

The demand for a high resolution image obtained by seismic surveys rises for most geotechnical applications. We test the applicability of an elastic full waveform inversion (FWI) approach for a crystalline host rock below ground. We performed seismic measurements with multicomponent receivers using transmission geometry in the GFZ Underground Lab in the research and education mine "Reiche Zeche" of the TU Bergakademie Freiberg. A 2-D FWI is applied to gain a high-resolution velocity model by using the waveform of the first arrival P-wave. The expected resolution potential of the FWI is tested with synthetic data. A preprocessing flow was developed, tested and applied on the synthetic data. As the true model, a random media velocity model which mathematically represents the crystalline rock was used. The starting model was obtained from a first arrival travel time tomography. After 250 iterations, the inversion converged successfully. Measured and synthetic waveforms as well as the true and the inverted velocity model match almost perfectly. Also, the velocity model was reconstructed very well. The unresolved structures are in the range of the smallest wavelength. The developed preprocessing sequence used in the inversion is a very well suitable strategy.

INTRODUCTION

For developments below ground, such as tunnels or caverns, the knowledge of the geological surrounding is important on the one hand to minimize the investment risks, and on the other hand to maximize the safety, e.g., for the local workers. In science and industry, it is common practice to determine the velocity distribution by performing seismic measurements and by using common methods for the interpretation such as a tomography of the first breaks. However, the resolution of the subsurface structures are limited to an order of the wavelength because of the high-frequency assumption of the ray theory. Because underground structures are often built in crystalline rock environment with strong heterogeneities which causes significant scattering of seismic waves the resolution of travel time based methods are usually not sufficient to detect small scale structures, e.g., fractures or small cavities. In contrast to these methods, the FWI uses the full content of the recorded wavefield, i.e., it also takes phases and amplitudes of both direct and scattered waves into account. We aim to perform a Full Waveform Inversion (FWI) to gain a high resolution P-wave velocity model of a crystalline host rock within the research and education mine "Reiche Zeche" of the TU Bergakademie Freiberg, Germany. We want to investigate the potential of multi-parameter reconstruction to have a better identification of weak rock. We introduce an elastic FWI with future application to field data.

We perform synthetic tests using the almost identical geometry as in the real measurement to reconstruct the properties of a crystalline rock which is a Freiberger Greygneiss with heterogeneities at different length scales. In this report, we first give some insights on the field data acquired in Freiberg. Then, we introduce a preprocessing flow which is applied to the data. Afterwards, the preprocessing is applied to a synthetic

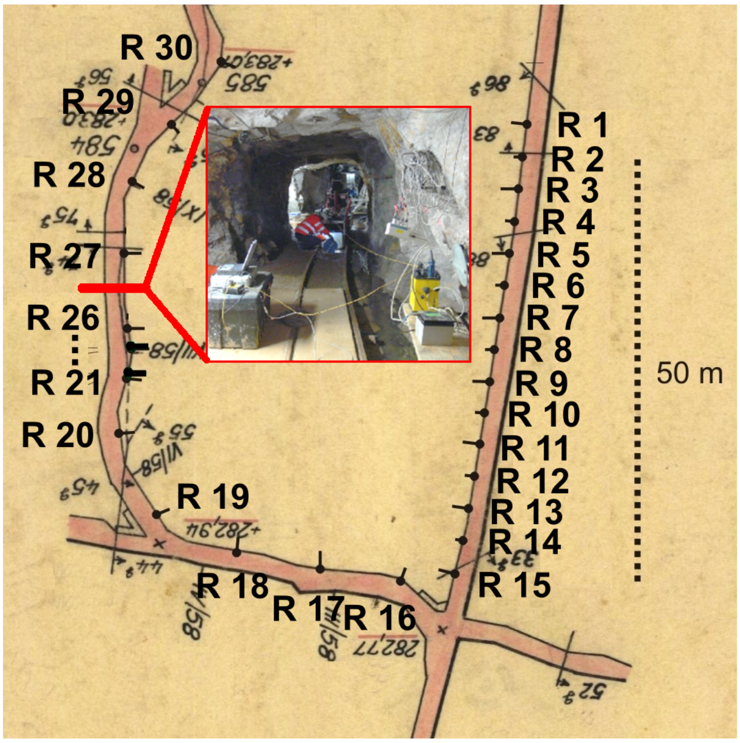

Figure 1: Section of an old underground map from the year 1962 showing the measuring area modified after Koop (1962): The block is surrounded by three galleries in a depth of 150 m. 30 3-C receiver anchors are mounted within the block (R 1 to R 30). The photo shows the inside of the left gallery (Photo: A. Jurczyk, GFZ).

data set based on the measurement geometry in Freiberg. The synthetic data is then inverted using an elastic 2-D FWI code.

FIELD AND DATA DESCRIPTION

The survey area is an old silver mine in Freiberg, Germany. Figure 1 shows a top view of a map from the year 1962. A block of crystalline rock (Freiberger Greygneiss) of about 50 m x 80 m is surrounded by three galleries which form an almost flat 2-D plane. No prominent geological discontinuities are expected, except of inhomogeneities which are common for gneiss.

The German Research Centre for Geosciences (GFZ), Potsdam, Germany, has set up a laboratory and the necessary infrastructure to perform seismic measurements below ground. 30 holes were drilled with different depth in which 30 3-component (3-C) receivers were deployed (Figure 1, R 1 to R 30). The receiver distance is approximately 4-9 m. In December 2009, seismic data of 76 shots spaced almost equidistantly (source spacing about 2 m) around the block, were recorded. The source was a magnetostrictive vibrator which generates sweeps from 300 to 3000 Hz (Borm and Giese, 2003). The source- and receiver positions are in a transmission geometry and located in an almost ideal plane. The advantages of using a transmission geometry is that the data that can be compared to crosswell data. This has been shown by Zhou et al. (1997).

Correlograms of a receiver gather for the first measurement of the sweep signal and the recorded data of receiver 21 are shown in Figure 2. The orientation of the x- and y-component is illustrated in the small sketch in Figure 2 b). Please note that the source locations follow the shape of the block, i.e., they are not in-line. P-waves are highlighted by the red color in Figure 2 and are clearly identifiable. Especially in Figure 2 a), the arrival of the direct P-wave can be easily seen, in particular for the shots 1 to 28. Since the tunnel surface waves, highlighted in green, have a significantly higher amplitude on the x-component, P-waves are not clearly visible in the shot position range 55 to 79. Tunnel surface waves are usually Rayleigh waves propagating along a tunnel wall (Jetschny et al., 2010). Because of the different polarization, the S-waves are most dominant on the y-component, highlighted in blue.

FULL WAVEFORM INVERSION

Description of the methodology

The forwards modeling, which is an essential part for any inversion, is done with a parallel time domain elastic Finite Difference code (Bohlen, 2002). The inversion is performed using a 2-D elastic FWI code by Köhn (2011) based on the adjoint method applied in time domain (Tarantola (1984); Mora (1987)). Because of the almost plane geometry of sources and multi-component receivers for the measurement, a 2-D elastic FWI method appears to be suitable. A 3-D inversion would require much more computational resources.

Preprocessing of the data

The seismograms at small offsets are dominated by the tunnel surface waves Figure 2 a, green colour). Jetschny et al. (2010) showed that the propagation characteristic of surface waves at a plane interface significantly differs from surface waves within a tunnel environment (tunnel surface waves or TS-waves). Using the 2-D forward modeling code, the interface between rock and air can be treated as such a plane surface. This implies, that we cannot invert for the TS-wave. Therefore, we decided to invert for only the waveform of the first arriving P-wave as demonstrated by Sheng et al. (2006). Nevertheless, we use both receiver components for the inversion. Commonly, unwanted surface waves are filtered by a f-k filter. However, the filtering may produce artifacts which may influence the inversion. Instead, we first manually review each trace in the time domain in order to isolate the P-wave phase. We neglect traces where both the P-wave and the surface wave are not clearly separated in time. For all other traces, we apply a time window to mute further phases except of the P-wave. Additionally, traces are normalized to their maximum (Figure 3 b). This processing scheme has been successfully applied to a field data set of Ketzin (Zhang et al., 2012).
Now, we first test the full waveform inversion of P-wave phases on synthetic data using the actual measurement geometry used for the real data acquisition.

SYNTHETIC EXAMPLE

Test and starting model

The host rock encountered in the GFZ Underground Lab is a crystalline rock. In order to test to which degree whether our approach can resolve small structures, a random media model with small and large scale velocity perturbations is used. Such a random media volume is commonly used to simulate crystalline rock properties. Therefore, we created a random velocity model with an exponential autocorrelation function with a correlation length of 10 m and a Gaussian distribution of $\sigma = 5$ %. Previous studies of the field data have shown that the average v_p velocity is roughly in the range of 6000 $\frac{\mathrm{m}}{\mathrm{s}}$. We used this value as the mean v_p for the velocity model.
The v_s velocity model is calculated by the $v_s \approx \frac{v_p}{\sqrt{3}}$ criteria. The density model is kept constant at 2550 $\frac{\mathrm{kg}}{\mathrm{m}^3}$ which is typical for crystalline host rocks. The model is illustrated in Figure 4 a) and has been discretized on a 468 x 664 Cartesian grid.
Due to the high non-linearity of the inversion problem and in order to gain robustness, the FWI requires a sufficiently accurate velocity model as a starting model. Pratt and Goulty (1991) showed that a starting model gained from a travel time tomography can be sufficient for FWI. In order to obtain such a good starting model, we performed a FD forward modeling for each shot location and manually picked the first arrivals in the corresponding receiver gathers. These first arrival times are then inverted using a standard travel time tomography on the basis of the simultaneous iterative reconstruction technique (SIRT). The starting model gained from the travel time inversion is shown in Figure 4 b).
From this Figure, it is obvious that the derived velocities of the starting model are generally too high. In

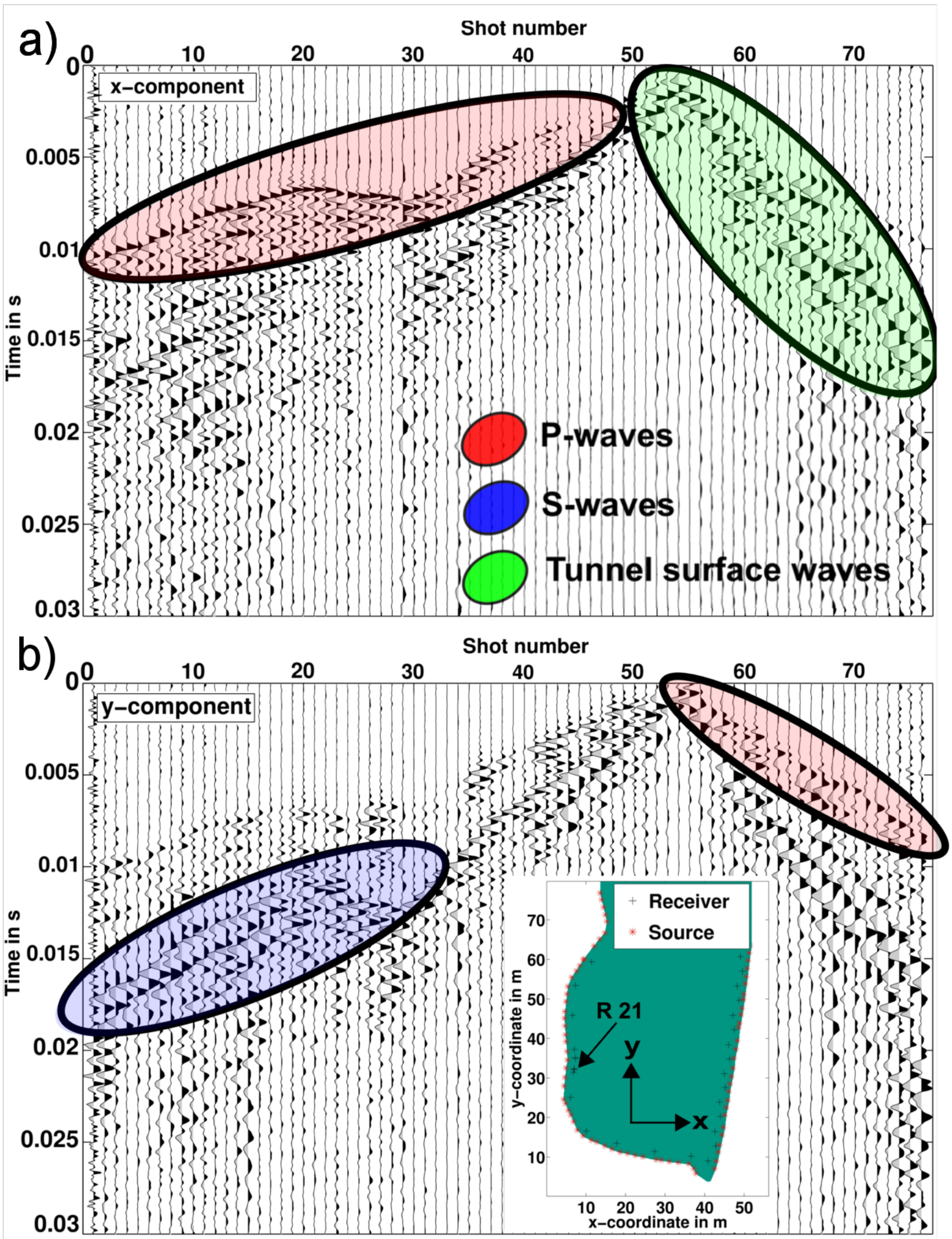

Figure 2: Field data: common receiver gather (R 21) of vibroseis data: Traces are normalized to their maximum amplitude. In **a)**, the x-components and in **b)**, the y-component is shown. Different wave types are highlighted with different colors.

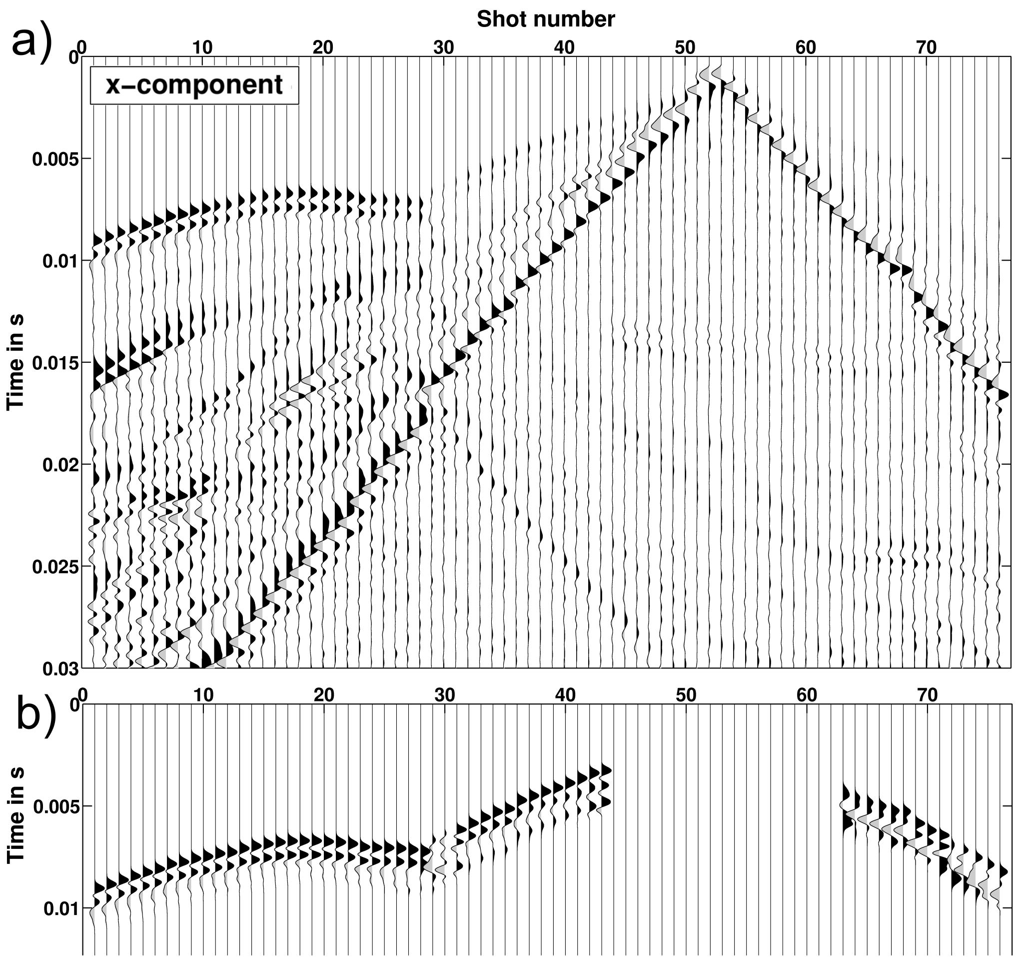

Figure 3: Synthetic common receiver gather of receiver 21: The geometry of the sources has been adapted from the field measurement. **a)**, unprocessed seismograms. **b)**, preprocessed data of a).

contrast, the velocities at the rock-air interfaces are too low. This can be explained by the very small source receiver distances and the resulting near field effects. Due to a lack of ray coverage at the top of the model, an additional gradient to the initial constant velocity has been interpolated and applied to the starting model to avoid an artificial interface. If we compare the structure of the true model and the starting model, the size ranges from about 5 to 10 m and corresponds to the chosen correlation length of 10 m of the true model. For a real data experiment, we do not know the true model. Hence, we will use this starting model as it is.

Data example

The synthetic data example gained from FD forward modeling using the true model (Figure 4 a) and the same geometry as used in the field measurements is shown in Figure 3 a). Again, all traces are normalized to their maximum amplitude. In comparison to Figure 2 a), it can be seen that the general behavior of the first arrival waves are equal. In the seismogram, the shots close to the receiver are dominated by the surface waves. For the shots in a transmission geometry, the first arrival is the P-wave and later followed by the S-wave. As described in the previous section, we intended to only use the waveform of the P-wave phases. In Figure 3 b), the receiver gather is shown after the preprocessing. The near offset traces, where P-wave and surface wave are not clearly separated, are muted. After muting of further phases, the P-wave can also be seen for the shot range 62 to 76 in Figure 3 b).

Inversion parameters

For the forward modeling, we have chosen a half period of a sin^3 with a frequency of 1000 Hz as the source wavelet. The frequency content is up to 2000 Hz, nevertheless, more than 75% of the energy has a frequency below 1000 Hz. The frequency content resembles the source which was used for the field measurement. With a velocity of 6000 $\frac{\mathrm{m}}{\mathrm{s}}$, the smallest wavelength is, thus, about 3 m. The forward modeling of each of the 76 shots required approximately 3000 timesteps.
The misfit definition for the inversion is the L2-norm. Additionally, we applied a preconditioning of the gradient around the sources to prevent source artifacts. Besides, no frequency selection has been applied, i.e., all frequencies are used from the very first inversion. By trace killing, only 70 % of the traces are used. The density is constant and will not be inverted. After 250 iterations, the misfit function did not decrease further for several iterations and the waveforms of the seismograms matches very well. The 250 iterations on a grid consisting of 310752 grid points and considering 3000 time steps take about 24 hours on 96 cores.

RESULT OF THE FWI

In Figure 5, seismograms of receiver R 5 and R 28 are shown. The blue traces are the true seismograms corresponding to the true model (Figure 4 a), the black seismograms are the starting seismograms corresponding to the starting model (Figure 4 b). The seismograms differ in the traveltime as well as in the waveform. After 250 iterations the seismograms of the inverted model are shown in dashed red. It is obvious, that the true seismograms and the seismograms after 250 iterations match quite perfectly. The FWI can handle both, the discrepancy in traveltime and in waveform.
Figure 4 c) shows the FWI-result for the v_p velocity model. First of all, it is noticeable that the inversion corrects for the high velocity of the starting model. 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 2 to 3 m and is in the range of the smallest wavelength. Nevertheless, some artifacts at the receiver positions and high velocity contrasts, close to the gallery walls can be observed. In the upper most part, the inversion did not update the model because of zero ray coverage.
To conclude, the FWI was able to resolve the small scale structures. These is the base for inverting for the field data.

CONCLUSIONS AND OUTLOOK

Based on an experiment conducted on a crystalline rock block in the research and education mine “Reiche Zeche“, synthetic tests demonstrate the applicability of the FWI inversion. The actual measurement was

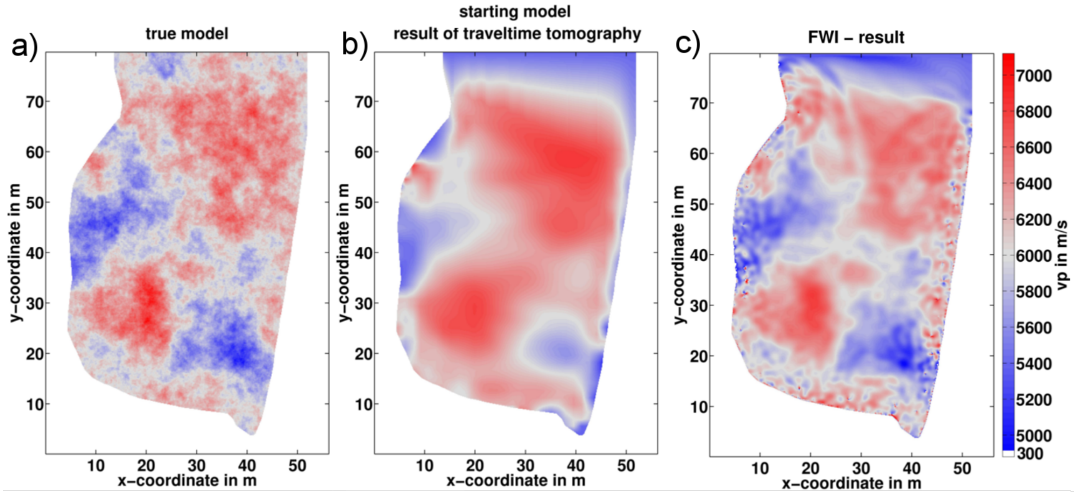

Figure 4: Result of synthetic study: **a)** True model: A Random media model with a velocity perturbation which simulates a crystalline rock. **b)** Starting model: The starting v_p velocity model for the inversion which was derived from a first arrival travel time tomography of the true model. **c)** FWI-result: Inverted v_p velocity model after 250 iterations using the first arrival P-waves only.

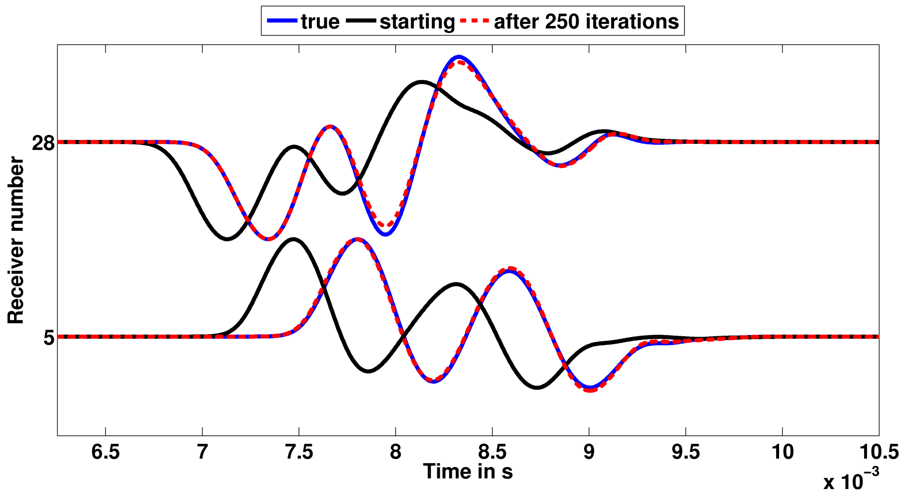

Figure 5: Comparison of true (blue), starting (black) and inverted seismograms (red dashed) for receiver 5 and 28 of shot 40.

performed with a high frequency magnetostrictive vibrator source at 76 points and 30 3-C receivers, all located in one plane in a gallery along three sides of the block. The preprocessing flow for inverting the real data with an elastic FWI was tested on a random velocity perturbation model. Since we use a 2-D source and receiver geometry and, thus, the inversion is sensitive to an almost 2-D plane within the rock, it is feasible to first employ a 2-D elastic inversion code. However, the accurate modeling of tunnel surface waves is than impossible. Instead, we focus on the inversion of only the P-wave phases by applying trace killing of near offset receivers, time windowing, muting of further phases and trace normalization. The inversion results are very promising. Starting from a model gained by a first arrival travel time tomography, most of the structures of the true model have been recovered down to structure sizes of about the shortest wavelength considered in the forward modeling. Therefore, our preprocessing approach by only using the first P-wave arrival is suitable to get a high-resolution image after the inversion.
For the final inversion of the field data, two more features are needed to be considered in the inversion process. First, the data has to be transformed by a 3-D to 2-D transformation, and second an inversion for the source time function is inevitable. Later on, we will also include S-wave arrivals in the FWI scheme.

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