MICROTREMOR LOCALIZAION

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ABSTRACT

Passive seismic methods are of keen interest to the petroleum industry. Reservoir monitoring, hydrofrac imaging or mapping enhanced oil recovery are a few applications of the passive seismic method in exploration seismics. In this study we present a passive seismic imaging method based on move out correction and cross-correlation stacking which is applicable to point source and microtremor data. The method does not require picking of events and uses all recorded channels simultaneously which allows to detect very small events. The method has similarities to the diffraction stacking method and differs particularly in the way the time axis is collapsed in the imaging process which is performed by zero lag cross-correlation. The maximum of the image function obtained by this method provides the source location. The moveout of first and most energetic arrivals are considered and the results of most energetic arrivals provided better source locations in complex media. A comparison with the diffraction stacking method for source localization confirmed the potential of new method.

INTRODUCTION

The Earth's ambient noise has been used over the past four decades for sedimentary basin studies, but only in the past ten years geophysicists have begun to apply ambient seismic noise, a technique known as passive seismic, to hydrocarbon exploration. It does not require artificial sources or large staff, i.e., it is low cost, and can be used in sensitive environments where conventional seismic faces problems (Emidio and Nunes, 2010) . Recently the passive seismic method has even been suggested as a direct hydrocarbon indicator (e.g. Emidio and Nunes (2010); Steiner (2009); Lambert et al. (2009); Graf et al. (2007); Saenger et al. (2007). This method uses a simple empirical observation of an increase of the spectral energy of the ambient seismic noise at frequencies between 2 and 4 Hz. This frequency anomaly was observed above several hydrocarbon reservoirs but was absent outside the reservoir area (Figure 1). A physical explanation for this observation is still lacking. We will not discuss the physical relevance of this hypothesis in exploration but rather focus in this contribution on the localization of seismic events. The low frequency events mentioned above are called microtremors and represent a continuous seismic emittance from the source area whereas classical seismic events are of short duration. Microtremor events were also observed in volcanic areas and in several seismological applications. The localization of the origin of microtremor events may be helpful in seismology, volcanology or hydrocarbon studies. The first step in these applications always is the reliable localization of the origin of the microtremor data. Since microtremor data are continuous in time they can not be picked. The processing of these data requires other techniques for the localization than for classical seismic events. Imaging or back projection methods as used for active seismic data have a not yet exploited potential as localization techniques. The collapse of the time axis (steered by an appropriate imaging condition) to obtain a source image, however, requires special consideration.

Gajewski and Tessmer (2005) introduced a localization method based on time reverse modeling, which does not require any picking of events and uses all recordings simultaneously. The advantage of this method is that the focusing of energy in the back projection process allows to image very weak events, which could not be identified in the individual seismogram of the recording network. Gajewski et al. (2007) proposed

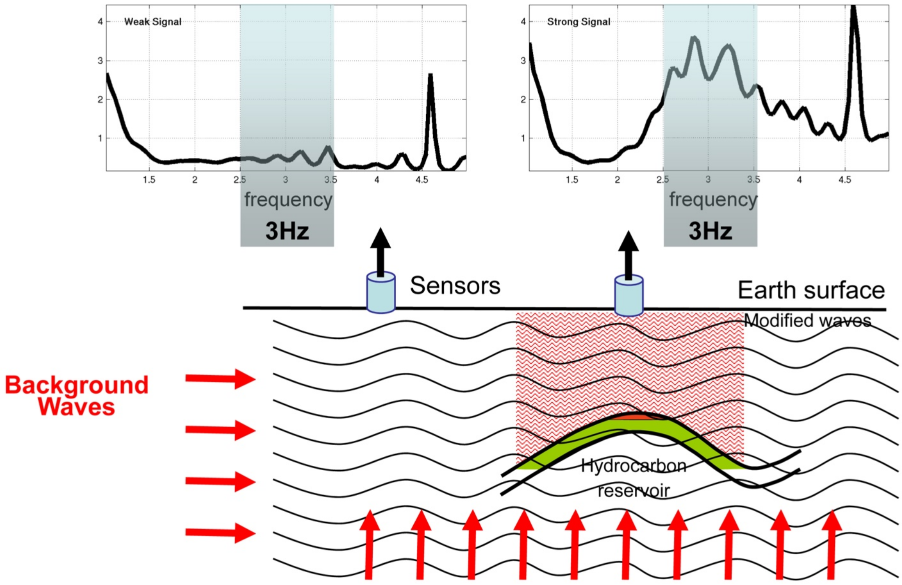

Figure 1: Schematic Passive Technology (a) Spectral Microtremor Analysis, (b) Typical geological structure (Figure taken from Steiner (2009)).

a new technique using a stacking approach to back project passive seismic observations. This method also does not require event picking and uses all channels simultaneously. The subsurface is discretized and each subsurface location is considered to be a potential location of a seismic event. Anikiev et al.(2008, 2009) and Gajewski et al. (2007) have shown, that the stacking method is a suitable tool to locate the origin of seismic events even when the signal to noise S/N ratio is rather poor. Steiner et al. (2008) applied time reverse modeling (TRM) to image microtremor events to the subsurface. The idea of locating the spatial origin of signals with time reverse modeling has been adapted from existing studies by other authors in physics, medicine and seismology. There are various methods in TRM to collapse the time axis. Steiner et al. (2008) used the absolute particle displacement or velocity per grid point throughout the entire time of the reverse modeling. In this imaging condition the highest values correspond either to high amplitude of one wave front or to the positive interference of recorded signals reversely propagated to the corresponding subsurface locations. In this paper, we present a new passive seismic imaging method using move out correction and cross-correlation. The cross-correlation serves as the imaging condition to collapse the time axis of the recording window and is maximum if the two considered traces are coherent. This imaging condition is equally applicable to point source events of short duration and microtremors.

After the introduction we introduce the new localization method based on static moveout shift and cross-correlation. We briefly review the diffraction stacking approach for source localization and present several 2-D numerical examples for point source and microtremor data to verify and characterize the localization methods. For complex media we consider the moveout of first and most energetic arrivals and investigate its influence on the localization.

LOCALIZATION METHOD

Localization using the cross-correlation stacking method

The coherence of two traces follows from how well correlated they are. This can be quantified by the cross-correlation function (Winter and Steinberg, 2008). In signal processing, cross-correlation is a measure of the similarity of two waveforms as a function of a time-lag between them. This feature is exploited in the localization method. The localization procedure is as follows:

A discretized sub surface is considered. Each node of the sub surface grid represents a potential source location. The nodes are called image points in the following. The traveltime of each image point to the

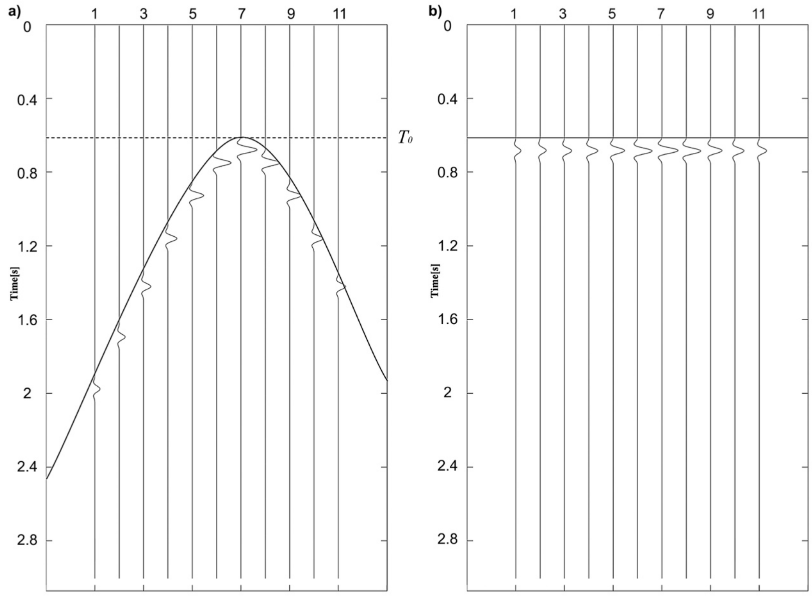

Figure 2: Seismograms for the homogeneous model (a) before and (b) after static moveout shift. For the correct velocity model and the correct location the event in the data is flat after static moveout correction

receiver locations of the recording network is determined. For this step a velocity model is required. This part of the localization procedure is common to the correlation and diffraction stacking approach. The smallest traveltime for each image point represents the apex location of the corresponding traveltime curve and is called zero time T_0. Based on this T_0 the move out of the other receiver locations is determined and applied as a static shift to the input seismograms. For a correct velocity model the seismic event should be flat after applying the static move out shift (see Figure 2)

Now a zero lag cross-correlation is applied which represents the imaging condition to collapse the time axis of the considered time window. Any trace may be considered as the master trace in this process. The cross-correlation results are stacked for all offsets providing the amplitude for the image point under consideration. The procedure is repeated for all image points providing the image function or section. The maximum of the image function indicates the source location. Some more details about the procedure are given in the following paragraphs.

NMO correction As soon as the traveltimes for each image point are available, a static moveout correction is performed to prepare the data for correlation and stacking. As mentioned before, the minimum time of the traveltime curve represents the apex time T_0 where the apex location in a heterogeneous medium generally does not coincide with the lateral position of the image point under consideration. The normal move out, $\Delta T_{NMO}(X)$, is the time difference in the seismic section between the one-way traveltime, $T(X)$, at an offset X from the apex location at some offset X_{apex} with the minimum traveltime T_0:

$$\Delta T_{NMO}(X) = T(X) - T_{X_{apex}} \tag{1}$$

In the data this move out is applied as a static time shift for each trace to align them to a common T_0. For the correct velocity mode and the correct location of the image point, all events would be perfectly aligned within the considered time window of the data.

Cross-correlation In this step we used the well known cross-correlation equation for two signals $x(n)$ and $y(n)$, each one with finite energy, which is defined as

$$r_{xy}(l) = \sum_{n=-\infty}^{+\infty} x(n)y(n-l) \tag{2}$$

and index l is the (time) shift (or lag) parameter. In this study we applied the zero lag cross-correlation. A critical issue is the choice of the master trace in the correlation process. We may consider the trace with the best S/N ratio as master trace which may require additional processing steps. An obvious choice could be to choose each trace as master trace and to stack all correlation results for each image point. If we have N traces in the data, we obtain $N \times N$ correlations. This brute force approach will not have a dependence on the master trace but is computationally involved.

Stacking To increase the signal-to-noise ratio (S/N) and to avoid a particular choice of the master trace the amplitudes of all correlation results for all offsets and master traces are summed for each image point in this step. This result corresponds to the amplitude of the image point. The loop over all image points provides the final image of the localization. For elastic media the procedure may be applied to both horizontal and vertical components of the data. All components are stacked to provide the final amplitude of the image point in this case.

Localization by the diffraction stacking method

The application of a diffracting stacking method to the localization of seismic events was suggested by Anikiev et al. (2007). They assumed seismic events to be caused by point sources. As a first step, the subsurface is discretized. Each node of the grid represents a potential source position the so called image point. This step is the same as for the cross-correlation approach. The stacking function S, along the aperture $2L$, is

$$S(x_i, z_i) = \int_{-L}^{L} u(\xi, \tau_0) d\xi \tag{3}$$

where u describes the input seismograms, τ_0 represents the traveltime from the image point (x_i, z_i) to the receiver located on the surface $(\xi, 0)$. The result is squared and the process is repeated and stacked by time steps Δt for the chosen time window from t_1 to t_2, i.e., through the whole data buffer. The procedure is repeated for every image point (x_i, z_i), i.e.,

$$DS(x_i, z_i) = \int_{t_1}^{t_2} S^2(x_i, z_i) dt \tag{4}$$

Where DS is the amplitude for each image point. This procedure is repeated for all image points and as a result we get a spatial representation of the image function for the chosen time buffer. The source location corresponds to the position of the maximum of the image function (Zhebel (2010) and Anikiev et al. (2007)).

Actually both methods, cross-correlation (CC) and diffraction stacking (DS) are very similar. Both methods consider the moveout but differ in the way the time axis is collapsed, i.e., in the imaging condition. For CC it is the zero-lag cross-correlation, which corresponds to the multiplication of the master trace and considered trace amplitudes and stack over all offsets and all times. For DS it is the sum of the squared stack amplitude along the computed diffraction traveltime for all times in the time window. Various other potential imaging conditions can be applied. For CC we could just consider the maximum correlation result instead of stacking the results for all master traces or we could correlate only neighboring traces and stack over all offsets. For DS we could just consider the maximum amplitude within the time window instead of stacking the squared stacked results for all times. These alternative options will be considered in future investigations. Both methods rely on traveltimes to consider the moveout in the data. We briefly discuss this issue in the following section.

Traveltime Computation

Transit times for seismic waves are computed by a variety of ways. Generally, more complicated media require more expensive and involved schemes to find the transit time. Over the past few decades, the growing need for fast and accurate prediction of high frequency wave properties (most commonly traveltime) in complex 2-D and 3-D media has spawned a prolific number of grid and ray based solvers. Grid based schemes, which usually involve the calculation of traveltimes to all points of a regular grid covering the velocity medium, became increasingly popular in recent times. They are often based on finite difference solutions of the eikonal equation (Vidale (1988); Popovici and A.ethian (2002); Rawlinson et al. (2007)) which are computationally efficient and highly robust. The combination of the latter makes them a viable alternative to ray tracing. The disability to compute later arrival traveltimes and a lower accuracy compared to ray tracing are disadvantages of FD eikonal solvers. Moreover, it is difficult to compute quantities other than traveltimes without first extracting the ray paths.

The determination of travel times in media that vary laterally as well as vertically has traditionally required some form of ray tracing. In this method the trajectories of paths corresponding to wavefront normals are computed between two points. This approach is often highly accurate and efficient, and naturally lends itself to the prediction of various seismic wave properties(Julian and Gubbins (1977) and Rawlinson et al. (2007)). Disadvantages of ray tracing include its complicated implementation. It may fail to converge to a two-point path. Moreover ray theory is applicable in smoothly varying media. This usually requires to smooth the velocity model such that the variations are small in dimension compared to the dominating wave length of the signal (Červený, 2001) .

In this study we used both, ray tracing and FD eikonal solver to compute traveltimes and compare the localization results. The NORSAR-3D ray modeling is used to compute the traveltimes of first and most energetic arrivals. In this program package the ray tracing system is solved using a Taylor expansion method or a 4th order Runge-Kutta method. The Taylor method is faster whereas the 4th order Runge-Kutta method is more accurate.

Synthetic Examples in 2D

Seismic event location is an inherently 3-D problem. Despite this fact we consider 2-D synthetic examples in this section to illustrate the method and to study its performance. The 2-D numerical examples serve as a feasibility study for verification purposes and to discuss advantages and disadvantages. We consider homogeneous and heterogeneous models. We consider point source data of short duration and microtremor events. Since the latter are considerably lower in frequency content than acoustic emissions this issue needs special attention for complex media.

Homogeneous medium Prior to the application of the method to complex models we need to verify the concept of the cross-correlation approach on a generic example were we consider a sparse network of receivers. We assume a 2D homogeneous medium with the size of 9000 m in X-direction and 3000 m in Z-direction. P-wave velocity is 2500 m/s. The total number of receivers is 11 in X direction located at $Z = 0$ with a spacing of 750 m starting from $X = 750$ m. We consider an explosive source at $(5250, 1500)$ m, i.e. the acquisition is not symmetrical with respect to the lateral position of the source. NORSAR-3D is used to compute the traveltimes although analytical computations would be possible for this simple example. Figure 3 shows the image function of the cross-correlation stacking were we consider different traces as master traces in the correlation (1st in Figure 3a, 7th in Figure 3b, 11th in Figure 3c). Figure 3d shows the stack of the correlations where each trace was used as master trace. The maximum of the image function coincides with the real source position. We observe that the maximum of the image function is independent on the choice of the master trace whereas the pattern of noise and the amplitude show such a dependence. The stacking of all results removed this dependence and optimized the image function. Similar to the diffraction stack the localization will work well in the presence of uncorrelated noise which may completely mask the actual signal. The potential to detect these events increases with increasing number of receivers in the observing network. However, the example shows that we can locate the event also in a sparse network. The requirement on the data quality (i.e., S/N ratio) are higher in this case.

Heterogeneous medium Here we consider the velocity model used by Steiner (2009). This model consists of ten sedimentary layers with P-wave velocities between 1200 to 3000 m/s above a crystalline basement with the P-wave velocity of 6000 m/s. A 50 m thick and 2000 m wide area represents a reservoir. The total number of receivers is 11 in X-direction located at $Z = 0$ with spacing of 750 m starting from $X = 750$ m (see Figure 4). S-wave velocities are determined from P-wave velocities by application of a factor of $1/1.4$.

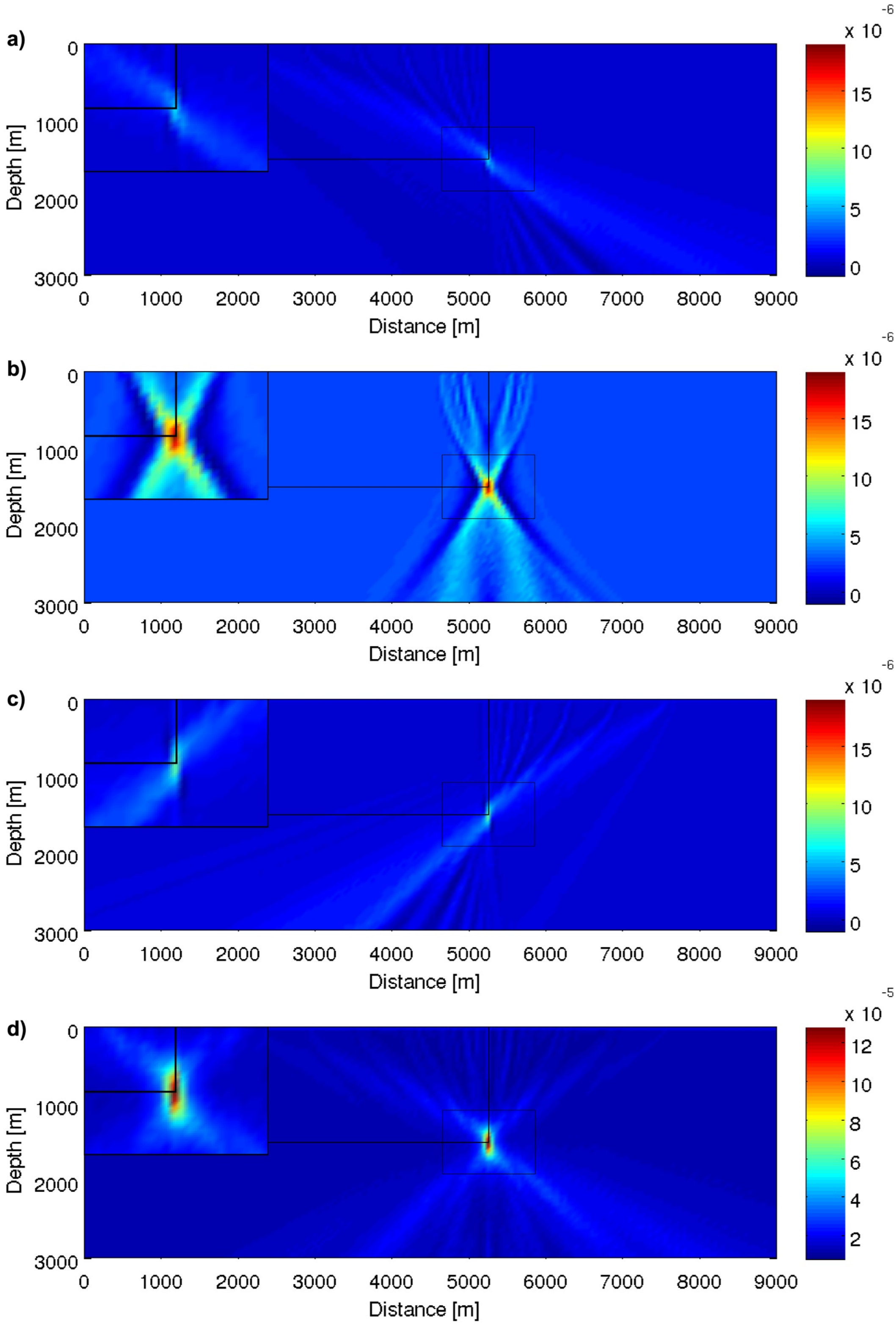

Figure 3: Localization of an explosive source in a homogeneous medium, using cross-correlation stacking obtained by (a) 1st, (b) 7th, (c) 11th trace of seismogram as master trace and (d) stacking all the results using all traces as master trace.

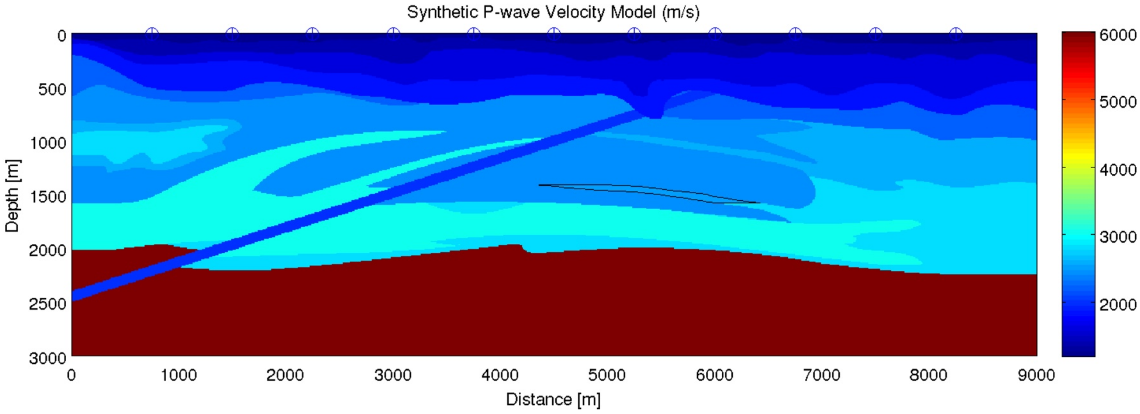

Figure 4: The synthetic velocity model consists of ten sedimentary layers above a basement unit. A thin area representing the reservoir defines the seismic source region of microtremor signals and blue circles represent the position of seismometers. Color legend on the right shows the P wave velocity (Steiner, 2009).

Localization in the unsmoothed velocity model We first consider the model as displayed in Figure 4. This does not allow the application of ray tracing. Therefore a finite difference eikonal solver Vidale (1988) was used to generate traveltimes for each image point to the 11 receiver positions. We first consider high frequency point source data.

Point source data Synthetic data were generated for a point source at $X = 5000$ m and $Z = 1500$ m. For the elastic modeling of wavefields a pseudospectral modeling method (Anikiev et al., 2008) with a time step of 0.5 ms was used. Wavefields up to a total time of 4 s were computed. The vertical components of Seismograms are shown in Figure 5 were two different prevailing frequencies in the signal are considered. A Ricker signal is used as a source time function. A signal with cut off frequency of 60 Hz is considered to represent the spectral content of an acoustic emission observed at the surface and a signal with a cut off frequency of 4.5 Hz representing the spectral content of a microtremor signal. The prevailing frequencies in the considered events are about half the cut off frequency. The traveltime curve of the first arrivals is also displayed in the seismogram sections. The visual appearance of the data is quite different due to the different interferences for the high and low frequency data. The apparent time delay of the low frequency event compared to the high frequency event will not affect the localization process. However, its different move out compared to the high frequency event will influence the localization. Traveltimes computed by FD Eikonal solvers always represent the arrival times of high frequency events. We applied cross-corelation stacking and diffraction stacking using FD Eikonal traveltimes to localize the point source events. Results of the localization of the high frequency (60 Hz cutoff frequency) and low frequency (4.5 Hz cutoff frequency) event for the unsmoothed velocity model are shown in Figure 6 and Figure 7. The localization was applied to both the horizontal and vertical component of the particle velocity separately and then stacked to obtain the image function. The maximum of the image functions is close to the exact source location. However, some high amplitude scatter is also observed. The maximum of image function for the low frequency event is shifted 100 m upward with respect to the actual source location. As expected, the result of the cross-correlation stacking method has a good agreement with the diffraction stacking method. It can be observed that the location process in complex media display are frequency dependence. Whereas for high frequency signals a good localization result was obtained, a location error is observed for the low frequency signal. The computed traveltimes describe the arrival times of high frequency events which might explain the reduced localization performance for the low frequency signal. This conclusion may impact the localization of microtremor events in complex media since these events are of low spectral content.

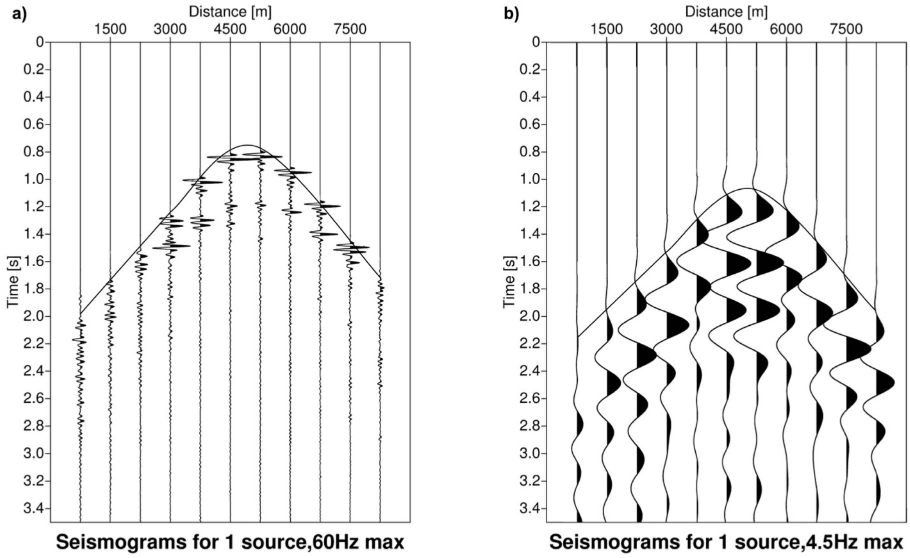

Figure 5: Vertical component seismograms. (a) Ricker wavelet with 60 Hz cut off frequency representing a high frequency signal, (b)Ricker wavelet with 4.5 Hz cut off frequency representing a low frequency signal (Anikiev et al., 2009).

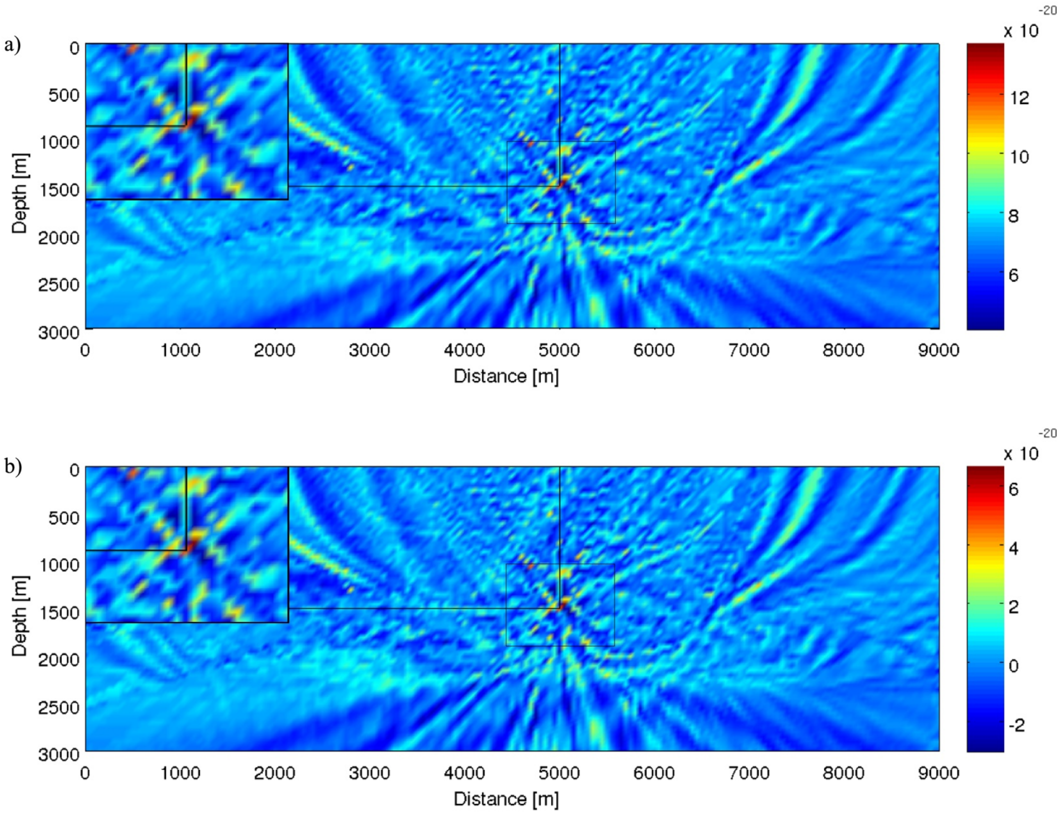

Figure 6: Localization in a complex medium of a point source event with Ricker wavelet of 60 Hz cut off frequency. Stacked results of horizontal and vertical component are shown. (a) Diffraction stacking result (Anikiev et al., 2008) and (b) Cross-correlation stacking result using first arrivals. Color legend on the right shows the value of the amplitude of the image function. The maximum of the image functions is close to the exact source location.

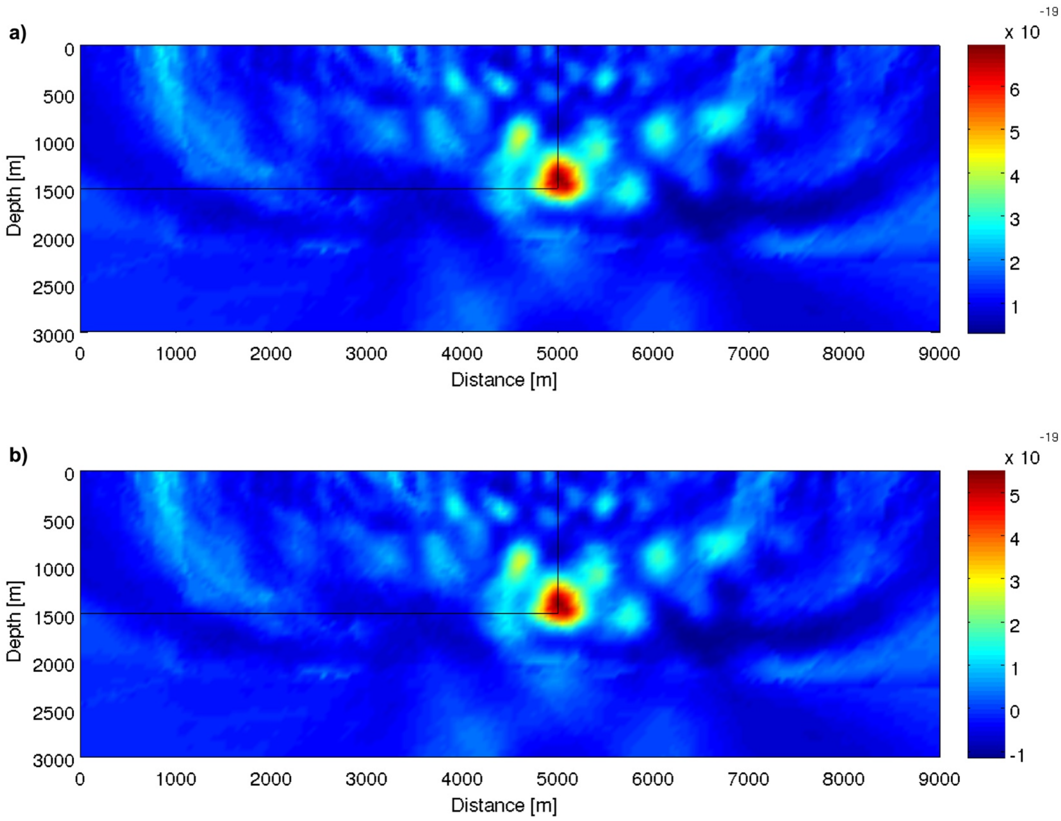

Figure 7: Localization in a complex medium of a point source event with Ricker wavelet of 4.5 Hz cut off frequency. Stacked results of horizontal and vertical component are shown. (a) Diffraction stacking result (Anikiev et al., 2008) and (b) Cross-correlation stacking result using first arrivals.

Microtremor Data Microtremors are low frequency continuous seismic signals of the earth. Microtremor observations above some reservoirs display remarkably similar spectral characteristics, pointing to a common source mechanism, although the depth, fluid content (oil, gas, gas condensate of different compositions and combinations) and reservoir rock type (such as sandstone, carbonates, etc.) for each of the sites are quite different. Synthetic microtremor data used here were generated by Steiner et al. (2008) utilizing two-dimensional finite difference (FD) simulations of elastic wave propagation of random point sources in the reservoir area (indicated by a black curve in Figure 4). Continuous microtremors are simulated by applying 574 low-frequency point sources with a prevailing frequencies ranging between 1.5 Hz and 4.5 Hz . These sources are randomly distributed within the area representing the reservoir, where the source times are randomly distributed during the 180 second numerical simulation. The resulting signal simulates a microtremor signal whose spectra are similar to the natural microtremors. The horizontal and vertical particle velocities are recorded through time with 11 vertical seismometers at the model surface (see Figure 4). The elastic model corresponds to the model shown in Figure 4. The type of the source used in this study corresponds to an explosive source. In Figure 8 we show localization results using diffraction stacking (Figure 8 a) and cross-correlation stacking (Figure 8 b) in the unsmoothed velocity model. White stars in the images indicate the largest amplitude in the image. Again, the image functions of both methods are almost identical and the lateral extension of the source area is well mapped. However, the high amplitude area appears too high compared to the actual depth of the source area. Also the image point with the largest amplitude is observed high above the actual reservoir depth. This could be again an expression of the low frequency content of the data whereas the computed traveltimes correspond to high frequency events. In addition to this fact another complication in complex media may contribute to the localization process: triplications. This can lead to complex interference patterns between first and later arrivals particularly for low frequency signals. Later arrivals can be computed by ray tracing which however requires some smoothing of the complex model.

In Figure 8a first arrival traveltimes computed by a FD Eikonal solver were used to localize the microtremor event. High amplitudes with a lateral extension similar to the reservoir width are observed, however, the high amplitudes occur above the reservoir area.

Localization in the smoothed velocity model

Smoothing the velocity model Since we want to apply ray tracing to compute the first and later arrival traveltimes the velocity model need to be smoothed. A 1-D Hamming window with 200 and 500 m extension was applied to each dimension of the model (see Figure 9). Whereas the smoothing with the 200 m window preserves most of the velocity features of the original model (Figure 4), the result of the 500 m window has lost some of the prominent features.

First and Most energetic Arrivals Many implementations of Kirchhoff migration use first arrival traveltimes. For complex media with wavefront folding later arrivals or most energetic arrivals provide the better image Geoltrain and Brac (1993). The most energetic traveltime is single-valued, but is discontinuous in general and requires the computation of the amplitudes of wavefronts. The most energetic traveltime may differ from the first arrival traveltime in regions where headwaves or triplications appear, in particular near high velocity contrasts (Kim, 2001). In this study we computed traveltimes for first and most energetic arrivals using the NORSAR-3D program package to compare the influence in imaging seismic sources. The first and most energetic arrival traveltime contours for a shot at $X = 5250$ m and $Z = 1500$ m in the smoothed velocity models are shown in Figure 10.

Particularly in the model with 200 m Hamming window we recognize extended areas were later arrivals are the most energetic events. The difference in the first arrival traveltimes for the model with 200 m and 500 m smoothing window is not obvious. For the localization in the following examples we are using the velocity model with the 500 m smoothing window since we want to achieve most stable ray tracing results. We consider first and most energetic arrivals. Before we apply most energetic arrivals to microtremor events we will investigate the influence of later arrivals on high frequency point source data.

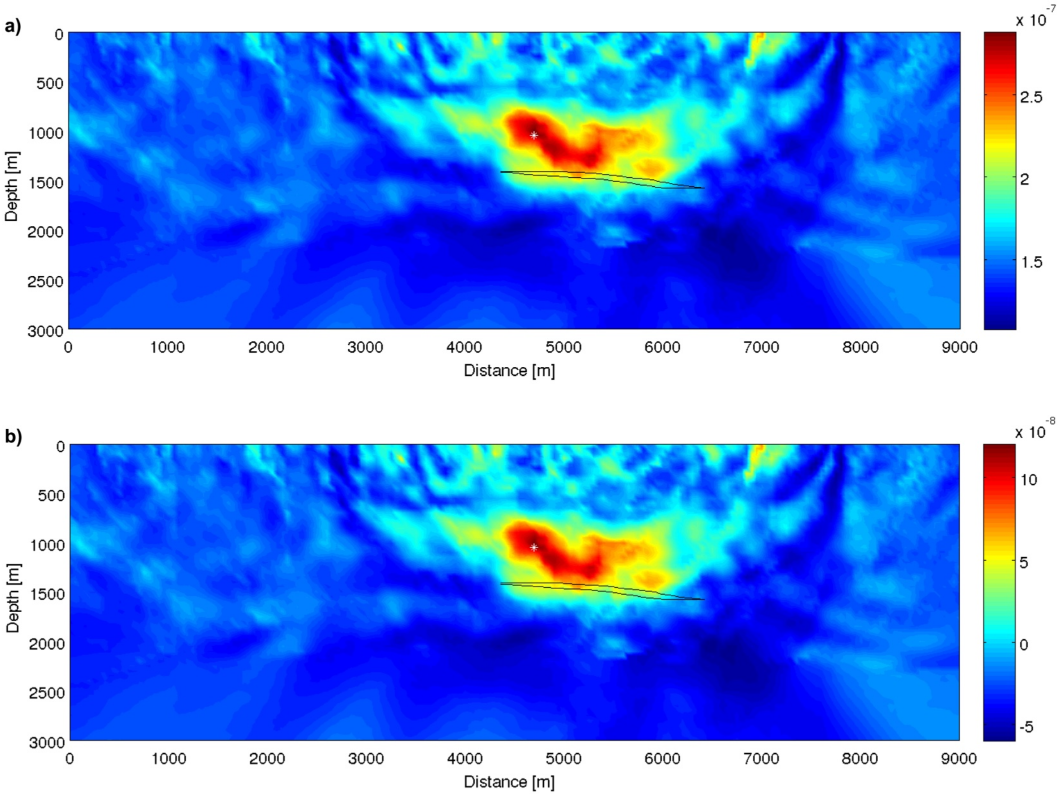

Figure 8: Localization of a synthetic microtremor signal in a complex medium. Stacked results of horizontal and vertical components are shown. (a) Diffraction stacking result (Anikiev et al., 2008) and (c) Cross-correlation stacking result using first arrivals. The white stars indicate the maximum of image function.

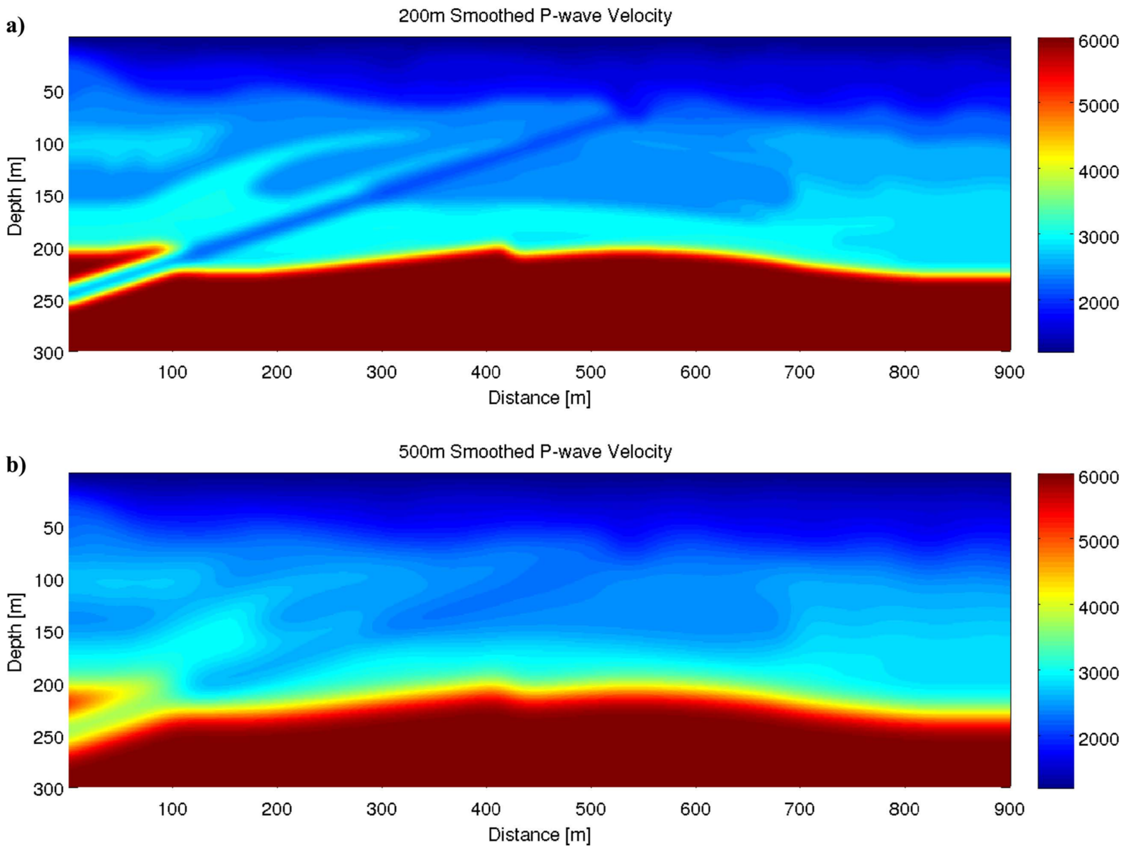

Figure 9: Smoothed velocity models using a Hamming window of (a) 200 m, and (b) 500 m.

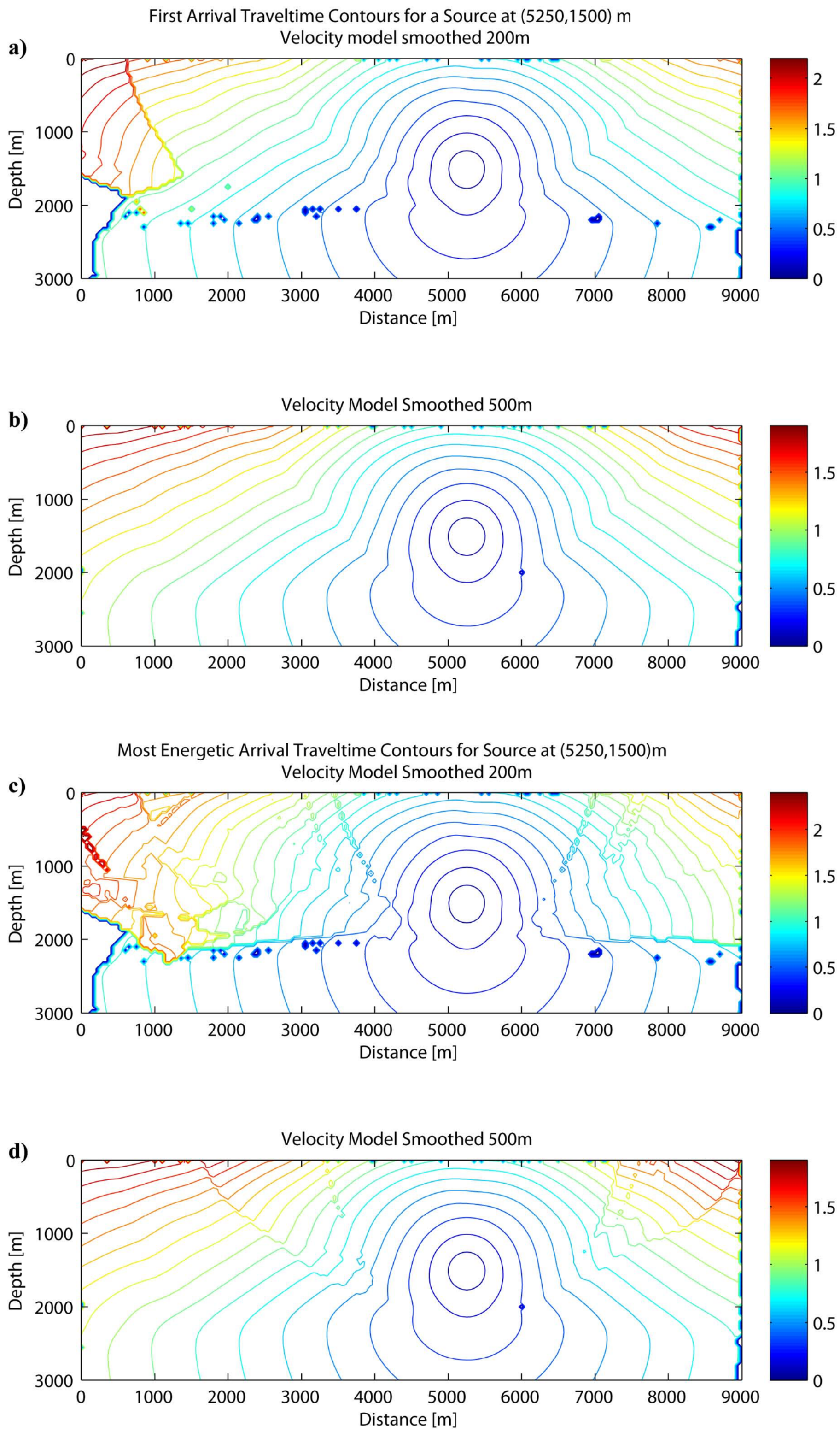

Figure 10: Traveltime contours of first arrivals computed for (a) 200 m, and (b) 500 m smoothed velocity models and most energetic arrivals for (c) 200 m, and (d) 500 m smoothed velocity models. Color legend on the right shows the traveltime value

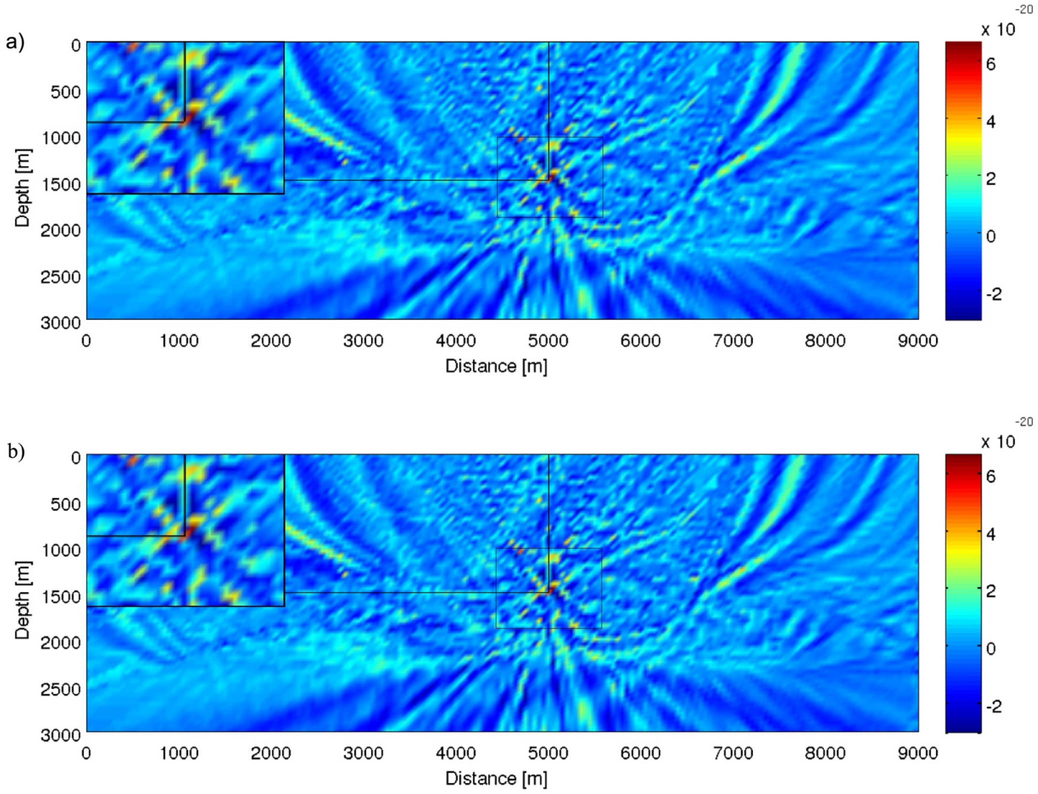

Figure 11: Localization in a complex medium of a point source event with Ricker wavelet of 60 Hz cut off frequency. Stacked results of horizontal and vertical component are shown. (a) Cross-correlation stacking result of first arrivals and (b) most energetic arrivals.

Point source data Results for the localization of a high frequency (60 Hz cutoff frequency) and low frequency (4.5 Hz cutoff frequency) point source event in the smoothed velocity model are shown in Figure 11 and Figure 12. We applied the localization to both the horizontal and vertical component of the particle velocity separately then stacked the results to obtain the final image function. For the high frequency data the first arrival traveltimes match with the move out of the events fairly well (see Figure 5). Therefore we expect the maximum amplitude at the correct location for these data (see Figure 11) which is in fact the case. This is not true for the low frequency signal, since in this case several events are interfering. The first arrivals move out is slightly larger (see Figure 11), which leads to a maximum of the image function at shallower levels. For the low frequency data we observe a difference when using first and most energetic arrival traveltimes. Moreover the imaging result using most energetic arrivals is better focused than the first arrival result. The maximum of image function for first arrivals is shifted about 100 m upward and for the most energetic arrivals is shifted about 70 m downward compared to the real source position (see Figure 12). In the next examples we consider microtremor events.

Microtremor Data We applied the cross-correlation stacking using first and most energetic arrival traveltimes to localize the microtremors. In Figure 13 we show localization results using first and most energetic arrivals. White stars in the images indicate the largest amplitude of the image function.

Figure 13a shows the localization result using first arrival traveltimes computed by raytracing using the model smoothed with a 500 m Hamming window. The cloud of high amplitudes displays a lateral extension similar to the reservoir and its depth is now more close to the actual reservoir depth. The maximum amplitude of the image occurs very close to the reservoir. The localization result for the smoothed model is considerably better than the result for the unsmoothed model. The image using most energetic arrivals displays a lot of noise and a very broad area of high amplitudes with a magnitude similar to the source area.

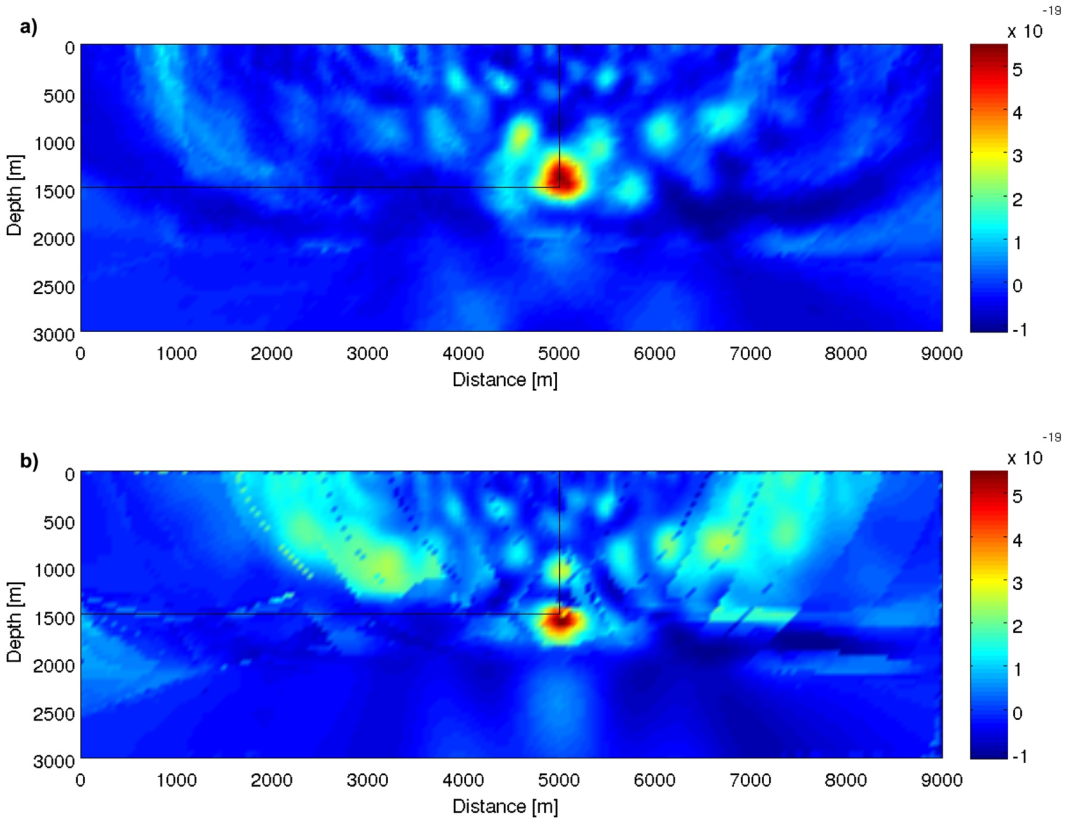

Figure 12: Localization in a complex medium of a point source event with Ricker wavelet of 4.5 Hz cut off frequency. Stacked results of horizontal and vertical component are shown. (a) Cross-correlation stacking result of first arrivals and (b) most energetic arrivals.

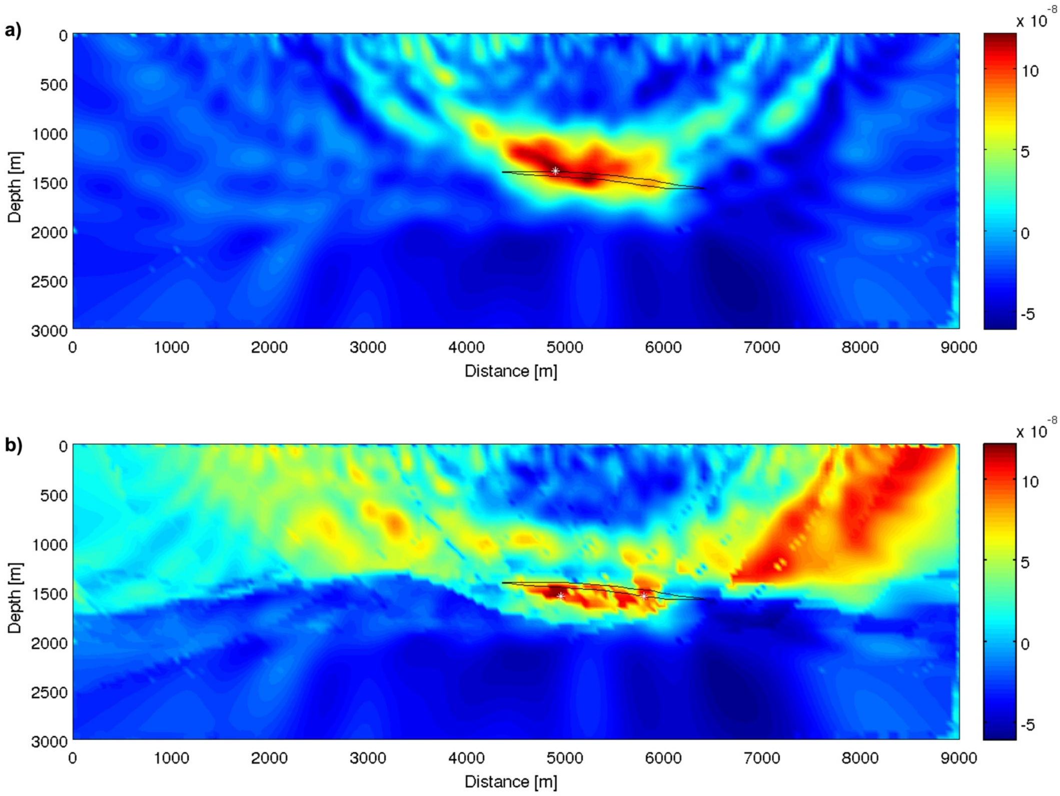

Figure 13: Localizationt of a synthetic microtremor signal in a complex medium. Stacked results of horizontal and vertical components are shown. (a) Cross-correlation stacking result of first arrivals and (b) most energetic arrival. The white stars define the maximum of image function.

DISCUSSION

The cross-correlation (CC) method provides an alternative to the diffraction stacking (DS) method for event localization. Both techniques provide comparable results requiring similar computational demands. Differences in the images obtained by CC or DS are unlikely if the same traveltimes are used in both techniques. Depending on the method used to compute traveltimes a smoothing of the velocity model might be necessary. For the velocity model considered in this study some depth differences in localization results were observed when either first or most energetic arrivals were used. Low frequency events may be more effected by model complexity since the computed traveltimes by ray tracing or FD eikonal solvers are inherently attributed to high frequency events. The Eikonal equation is a result of the high frequency approximation of the equation of motion for heterogeneous media. This is visible in the seismogram sections in Figure 4 where the move out of the high frequency event better matches the moveout of the traveltimes than for the low frequency event. Deviations in moveout directly influence the localization process.

Because of the low frequency content, microtremor events are likely to be more effected by model complexity which is illustrated by the example shown in Figure 9a and 9b. The localization results using the traveltimes obtained for the unsmoothed velocity model displays a shift of the high amplitudes of the image to shallower levels. For the model under consideration smoothing of the input model actually improved the localization result. Simple smoothing of the velocity model can not exactly describe the averaging feature of wave propagation in complex structures where the prevailing wavelength is considerably larger than the spacial model variations. However, for the velocity model considered in this study it helped to better localize the low frequency microtremor events.

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