Imaging of passive seismic sources by reverse modelling

D. Gajewski and E. Tessmer, University of Hamburg, Germany

Summary

The passive seismic method for reservoir monitoring and characterisation and monitoring of hydraulically induced fractures is developing into a main stream technology in the oil business. Current techniques rely on the fact that the recorded event is detectable at most of the stations of the recording array. Weak events, not visible in the individual seismogram of the array, are missed out. We present a new approach, where no picking of events in the seismograms of the recording array is required. The observed wavefield of the array is reversed in time and then considered as the boundary value for the reverse modelling. Assuming the correct velocity model, the reversely modelled wavefield focusses on the hypocenter of the seismic event. The origin time of the event is given by the time where maximum focussing is observed. The spatial extent of the focus resembles the resolution power of the recorded wavefield and the acquisition.

Introduction

The problem of earthquake location is one of the most basic problems in seismology. Although numerous applications exist worldwide, the inherent non-linearity prevents earthquake location and tomography from being a standardised routine tool. The earthquake location problem is stated as follows (Pujol, 2004): Given a set of arrival times and a velocity model, determine the origin time and the coordinates of the hypocenter of the event. This definition inherently assumes that the arrivals of an event are visible on a certain number of the recorded seismograms of the observing array. This also means, that the arrival has to be identified in the seismogram prior to the actual localisation of the event. This not only requires the correct identification of the onset of the arrivals but also the proper correlation of the individual phases among the different stations of the array.

The determination of the excitation time and hypocenter of a seismic source is traditionally performed by minimising the difference between the observed and predicted arrival times of some seismic phases. In these techniques it is assumed that the event is visible on at least a few stations of the recording array. In this paper we follow a different approach. Similar to Kao and Shan (2004) we will not assume that the event is visible on the individual seismogram of the array and we therefore do not pick arrival times. The information present in all recordings of the whole array is exploited. In our approach a reverse modelling technique is used to propagate the emitted seismic energy back to its origin. The wave propagation for this situation is basically a one way process and stacks all energy at the source position, if the correct velocity model is used. Kao and Shan (2004) developed the Source-Scanning-Algorithm where they exploit the stacking advantage by computing brightness functions for trial locations and origin times.

Method

The backward propagation of the recorded wavefield in the current version of our localisation method is based on acoustic numerical seismic modelling. The Fourier method, a pseudo-spectral method (Kosloff and Baysal, 1982), is applied. However, finite-differences algorithms or elastic modelling could also be used. The main idea is that, in contrast to seismic forward modelling initiated by a (highly) localised source, seismograms reversed in time are used as initial conditions at the receiver locations.

The modelling scheme is used to propagate the wavefield, which is fed into the numerical model at the receiver stations, backwards in time. This can be done since the wave equation has no inherent preferred direction in time. It is expected that after propagating the wavefield backwards in time, all energy will be focussed at the source location, which will lead to large amplitudes at this position. These large amplitudes can be easily detected by scanning the image at every time step. A continuation of the process of backward propagation beyond maximum focussing will spread the wavefield away from the source.

The propagation in time is performed iteratively by a time-stepping scheme step by step with small time increments Δt. At each time step the entire grid representing the subsurface pressure field is scanned for its maximum value and the spatial position thereof is stored. The location of the maximum pressure amplitude over all time is the source position and its time is the excitation time of the source we are searching for. The excitation time is obtained from the absolute time of the first sample fed into the back propagation and the number of time steps needed to focus the event.

Assuming the recorded wavefield stems from a point source and the subsurface velocity model is known, ideally the wavefield should focus perfectly. The spa-

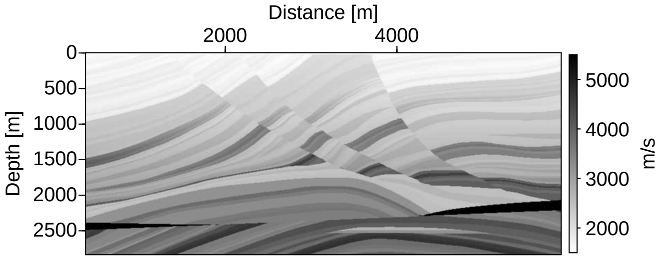

Fig. 1: Velocities for a portion of the Marmousi model.

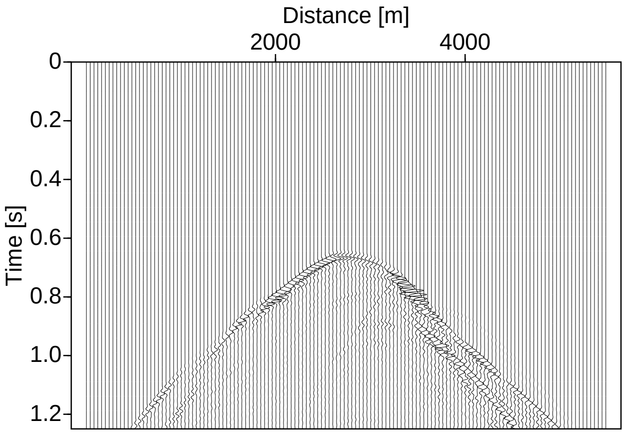

Fig. 2: Seismograms from a source at (2800, 1520) m for the Marmousi model.

tial extent of the focus reflects the resolution power of the experiment. This depends on the bandwidth of the recorded seismograms and the acquisition geometry. Thus, the localisation uncertainty is automatically obtained with this method. However, the receiver locations are not perfectly sampled around the source position (spatial undersampling). Therefore, much information for reconstructing the total wavefield at the source is missing. Consequently, we cannot expect that the wavefield perfectly collapses at the source location. The numerical examples show that this method works surprisingly well even for erroneous velocities and low fold arrays.

Numerical Case Studies

All times mentioned in the following text are relative to the exact source time, which is 0 s (zero time); i.e., negative numbers correspond to times prior to the real excitation time and positive numbers indicate times after the real excitation time. The performance of the localisation methodology described above is demonstrated using a part of the Marmousi model (Versteeg and Grau, 1991) (see Fig. 1) for which synthetic seismograms were computed by the Fourier method utilising a code developed by Tessmer (1990). For this model the velocities vary between 1500 m/s and 5500 m/s. The chosen grid spacing was 8 m. A Ricker-wavelet with a maximum frequency of 90 Hz where the dominant frequency is 45 Hz was used. The position of the explosive source is at (2800, 1520) m. 5000 time steps with a time increment of 0.25 ms were calculated. The calculations are done on a numerical grid of 693 x 385 nodes. Seismograms for the above mentioned source position are shown in Fig. 2. Only every fifth trace is displayed.

Despite the fact that the Marmousi velocity model is rather complex, the reverse modelling of the seismograms for the localisation yields the exact source position and the correct excitation time. Snapshots of the wavefront are shown in Fig. 3. The excellent localisation can be explained by the good coverage of the recording array. However, if we use only 6 receivers, i.e., a receiver spacing of 480 m the reverse modelling of the corresponding seismograms for the localisation still yields an almost accurate source position. The vertical coordinate is missed by only 8 m. The excitation time error is 2 ms. It is important to note, that for correct reverse modelling the geometrical spreading losses should be removed for sparse receiver arrays, which is not correctly handled by the current reverse modelling approach. We use time-dependent weights for a crude correction of spreading losses. If such corrections are not applied the largest amplitudes are observed directly at the receivers for sparse receiver spacing. A more detailed consideration of this issue is given in the discussion section below.

If the reverse modelling is performed with a wrong velocity model the source location and excitation time is also wrong. For this test the velocities are chosen 10% higher than the correct velocities in the entire subsurface model. The scan for maximum amplitudes yields source coordinates of (2800, 1496) m. This is correct in horizontal direction. The vertical position, however, is about 2% incorrect in depth (1496 m instead of 1520 m). Also, the determined excitation time is incorrect by 76 ms. Another example of an incorrect velocity model would be a smoothed version of the exact velocities. Such models are typically obtained from a tomography study. Performing the reverse modelling with a 100 times applied 3-point smoothing operator almost repeats the results for the unsmoothed velocities. The obtained errors are negligible for a localisation.

In order to examine the effect of noise on the localisation procedure a signal-to-noise (SNR) ratio of 0.5 is chosen. For such a SNR the seismic event cannot be identified in the individual seismogram of the section, i.e., it is impossible to pick the event. After reverse propagation of the wavefield, the estimated position of the source is wrong by a third of the dominant wavelength, i.e., below the resolution limit of the data. The timing error is 3 ms. In view of the fact that the events cannot be visually identified in the seismograms, this is a remarkably good result which is explained by the stacking of energy from the large number of receivers. In the corresponding snapshots showing the wavefront collapsing at the source location the focal area is clearly visible despite a higher noise level in the image. Numerical examples on 3-D models support the above given conclusions.

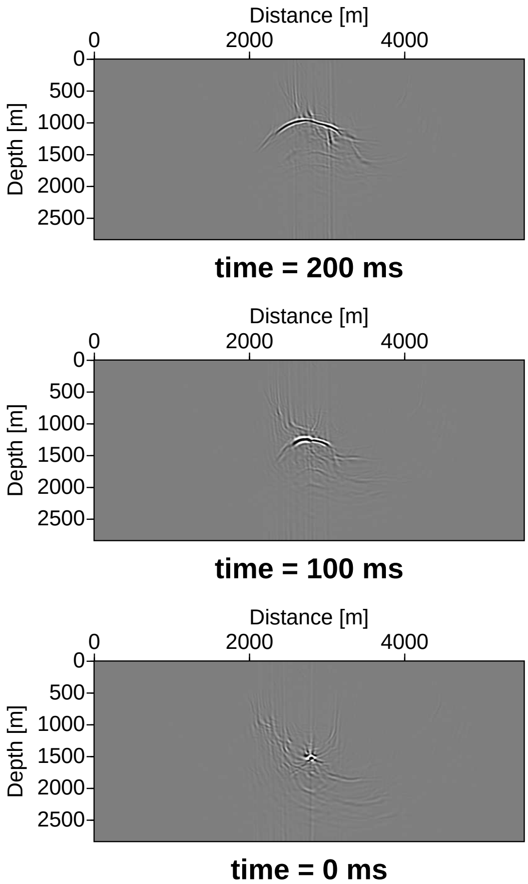

Fig. 3: Snapshots of the reversely modelled data for the Marmousi velocity model. The maximum amplitude is observed at the exact position and correct excitation time.

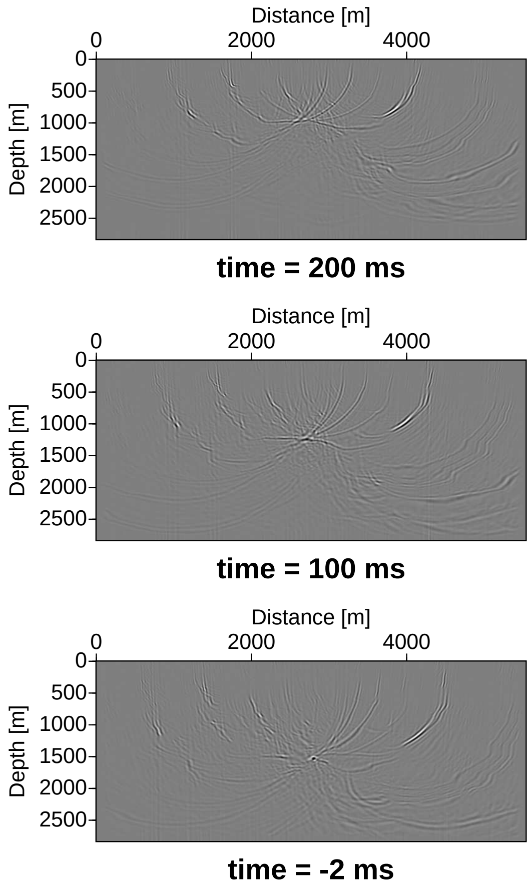

Fig. 4: Snapshots of the reversely modelled data for the Marmousi velocity model with 60× fewer receivers compared to Fig. 3. The maximum amplitude is found at -2 ms, i.e., 2 ms too early.

Discussion and Conclusions

We have presented a new technique for the localisation of seismic events and the estimation of their excitation times where no picking of seismic events is required. The examples show that the method works extremely well if a sufficiently high coverage with receivers is present. The algorithm still yields very good estimates even if the data are very noisy, here S/N = 0.5. This makes the method attractive especially for weak onsets which are not detectable on single traces. As one would expect, the quality of the localisation and excitation time estimate degrades if the subsurface velocities are not exactly known. A smoothing of the velocities does not degrade the localisation. The numerical experiments have shown that the algorithm works very well even for source radiation not visible in the individual seismogram of the array if a sufficient number of receivers is available (stacking advantage).

Even in rather complex subsurface models the localisation works surprisingly well. A sparse receiver coverage, however, may lead to wrong localisations if no spreading corrections are applied. The reversely propagated wavefield interferes constructively at the source position and the highest amplitudes should be observed at this point. If we use only very few receivers, i.e., a heavily undersampled wavefield, we observe the highest amplitudes at the receiver locations. The wavefields back propagated from the receivers experience spreading losses in the reverse modelling and consequently display lower amplitudes than at the receiver position. A possible solution is to restrict the search area (or volume) to locations which are sufficiently far from the receivers and where constructive interference of the wavefield is already present. This, however, requires a priori information on the possible source location. This can be exploited when observing aftershocks, hydraulically induced seismicity or seismicity caused by production and enhanced oil recovery.

The physically more appealing solution is to account correctly for the removal of geometrical spreading during back propagation. The spreading losses collected during the propagation from the source to the receivers need to be removed in the reverse modelling in order to obtain the correct radiated amplitude. In the current localisation method based on the wave equation this is crudely implemented by a simple time scaling of the images at every time step. Another option is the removal of geometrical spreading using a wavefield continuation method for back propagation which is based on high frequency asymptotics. The removal of geometrical spreading in reflection data is known from true amplitude migration (Schleicher et al., 1993; Tygel et al., 1992).

The potential to update the velocities based on the focussing of the event are not investigated yet. The experience with migration velocity analysis, which is also based on the focussing of energy at a common depth point, is motivating to progress into this direction. Having the source location and the excitation time estimated by the method described above a mapping of reflectors below the source location can be performed in a pre-stack reverse-time migration manner or with another pre-stack depth migration algorithm.

An extension of the localisation procedure to three dimensions is straight forward and was verified by numerical examples. The computational effort for the downward continuation using the wave equation increases considerably in this case. Downward continuation techniques based on high frequency asymptotics may help with this issue.

The method used in this paper can be applied almost in real time if a fast downward continuation method is used since no picking of arrivals is required. The data recorded by the array are stored in blocks. Each block represents a certain time, e.g., 10 s. This wavefield is then reversed in time and processed according to the method described above. The results can be displayed on a screen and inspected for focussed energy. Activity in the subsurface can be visualised almost instantaneously as bright spots on the screen. The new approach might be also particularly valuable for areas with very weak seismic activity or poor recording conditions (small SNR), for production monitoring in oil and gas reservoirs, for monitoring of hydrofracs and seismic swarm events.

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