Geomechanical monitoring of CO₂ storage reservoirs with microseismicity

Bettina P. Goertz-Allmann, Volker Oye, Steven J. Gibbons, and Robert Bauer

A NORSAR, Gunnar Randers vei 15, 2007 Kjeller, Norway
b Illinois State Geological Survey, 615 E. Peabody, Champaign, IL 61820, U.S.

Abstract

We analyse microseismicity induced during the Decatur, Illinois, carbon capture and storage (CCS) demonstration project. More than 10,000 microseismic events were detected during the injection of 1 Mio metric tons of CO₂ over the course of 3 years. The seismicity occurs in distinct clusters and shows little to no correlation to the progressing CO₂ front. For geomechanical reservoir characterization and seal integrity assessment, we need very high depth resolution for the event locations, such that events can be unambiguously attributed to specific formations. We therefore compare event locations using different sensor distributions (borehole and surface sensors) and/or including additional phase arrival information besides direct phases. Analysis of Brune-type stress drop of induced microseismic events with two different estimation methods exhibits signs of pore pressure diffusion processes within individual clusters, although the overall seismicity across clusters does not correlate with pressure gradient or CO₂ front. An observed distance dependence of stress drop from the nucleation point within a cluster suggests a local pressure gradient over the extent of the cluster.

Keywords: CO₂ storage; induced microseismicity; source parameters; geomechanics; Decatur

1. Introduction

Microseismic monitoring is an increasingly important part of the geophysical monitoring toolbox for efficient surveillance of large-scale sequestration projects. Apart from assessing the long-term seismic hazard within and
around the storage formation and providing mitigation in form of a real-time warning system, microseismic event data can be used for reservoir characterization and optimization of injection operations. The latter typically requires a good number of events in order to obtain reservoir information of sufficient certainty for quantitative use. Achieving this is likely to require seismological monitoring networks with sufficient detectability and recordings over long time intervals. In such cases, the microseismic cloud, and especially the individual event source parameters, can provide geomechanical constraints for history-matching of reservoir models.

We analyse locations and source parameters of induced seismicity at the Decatur, Illinois, carbon capture and storage (CCS) demonstration site [1]. Within three years, from November 2011 to the end of 2014, one million metric tons of CO₂ were injected into the Mount Simon Sandstone formation at about 2 km depth. The injection pressure of < 1 MPa was comparatively low. The seismic monitoring network consists of borehole sensors within the injection well (CCS1) and in a dedicated monitoring well (GMW, see Fig. 1). The deep borehole networks provide excellent detection capability but limited azimuthal coverage. The United States Geological Survey (USGS) also began monitoring microseismicity with a 13-station seismic network at the surface and in three shallow wells (150 m depth). The detection threshold of the surface network is inferior to that of the deep borehole network by about 1-2 magnitude orders, but the azimuthal coverage is significantly better [2]. However, these sensors were not in operation before July 2013, and hence only cover the second half of the injection period.

![Microseismic event locations and nearby receivers](image)

**Fig. 1.** Microseismic event locations (circles) together with nearby receivers are shown in map view and two depth-sections. Circle size is scaled by moment magnitude. Color in map view denotes the event time. The event cluster marked in green is analyzed in more detail. The arrow indicates the direction of the maximum principal stress $S_{\text{max}}$. Lower Mt. Simon, Argenta, and Precambrian interfaces are shown in light brown, brown and grey, respectively. The injection zone within the Lower Mt. Simon is indicated by the yellow squares.
More than 10,000 microseismic events were detected over three years, most of them by the deep borehole sensors. The event catalog [3] revealed significant clustering of seismicity with no direct correlation with the expected progression of the CO₂ front (Fig. 1). A subset of the detected events was located by applying a traveltime- and polarization-based algorithm using P- and S-wave arrival times on the deepest three-component (3C) sensor, and using a 1D anisotropic velocity model [3]. These locations are shown in Figure 1 and selected events are used later for source parameter analysis.

2. Microseismic event locations

Obtaining high quality microseismic event locations can be a challenging task. Data can suffer from poor signal quality (i.e., low signal-to-noise and low detectability at surface recordings), complicated phase identification due to strong heterogeneities (e.g., [4]), or too few data recordings and/or with insufficient azimuthal coverage [e.g., 5]. In addition, uncertainties in the velocity model are often large. In our analysis, we compare the effect on uncertainties of event locations when using (i) only first arrivals in downhole sensors, (ii) additional phase arrivals in downhole sensors, and (iii) using both downhole and surface sensors.

2.1. Surface & downhole recording

Due to signal-to-noise conditions and the small magnitudes inherent in microseismicity, downhole systems typically detect significantly more events with better timing precision than surface networks. This is owed to the fact that downhole sensors are closer to the target source, and placed in a typically much quieter environment. However, downhole systems are often limited in their observational aperture, which is important for accurate lateral locations. Additional phase arrivals can greatly improve event depth resolution. This is illustrated in Figure 2, where we plot ray paths to all receivers for two events at slightly different depths. If the event is in the reservoir (Fig. 2 left), head and direct arrivals at the deepest borehole sensor travel along separate paths. The head wave travels through the faster basement and arrives earlier with weak amplitude, while the direct arrival travels through the sediment and causes a clear secondary phase arrival in the waveform (see inset of Fig. 2 left). For a slightly deeper event from the basement, these ray paths do not separate, and the corresponding phases arrive at the same time (Fig. 2 right). This is even better observed in the spectrograms shown in Figure 3: The left column shows a reservoir event recorded in two deep borehole stations. The first arriving head wave contains the highest frequencies whereas the secondary phase comprises the direct arrival. The right column of Figure 3 shows a basement event and we observe that there is no distinction of two separate P phases since both head and direct arrival arrive at the same time.

Fig. 2. Theoretical ray diagrams between source (white circle) and receivers (red triangles) for two different events: in the reservoir (left) and in the basement (right). Rays for head waves are shown as black dashed lines and for direct waves in magenta solid lines. The background colour denotes P-wave velocity. The insets show real waveforms of similar representative events where one event (red trace) was used as master event.
Fig. 3. Comparison of spectrogram and waveform for two example events recorded at the two deepest downhole sensors within CCS1. The left column shows records of a reservoir event and the right column a basement event. Dashed vertical lines mark different phase (i.e. P-head wave, P-wave, and S-wave arrivals, respectively).

Fig. 4. Comparison of records from three selected stations. Top: the deepest borehole sensor within the injection well (PS3_1). Middle: record from a surface sensor (DEC13). Bottom: record from a 150 m deep borehole sensor (DEC02). All three components per station are shown (red=Z, blue=N, green=E). Dashed vertical lines on PS3_1 denote head wave arrivals. Direct phase arrivals for P and S are shown by solid vertical lines.

Figure 4 shows waveforms for one example event recorded at a deep borehole sensor (PS3_1), a shallow 150 m deep sensor (DEC02) and a surface sensor (DEC13). Note that head wave arrivals can only be distinguished for the deepest sensor (compare with Fig. 2). Surface and shallow borehole (150 m deep) recordings are generally of lower frequency and less detail. To investigate location error, we utilize a gridsearch approach using traveltimes and polarization to locate the event. Figure 5 compares location residuals for first arrivals from downhole sensors only (top), downhole sensors including additional phase arrivals (middle), and combined downhole and surface sensors (bottom), respectively. Jointly using surface and downhole data shows the least spread of the residuals, but requires careful weighting of the input data to mitigate uncertainties in the near-surface velocity structure. We have placed a stronger weight on phase arrival times of the downhole sensors. Using only borehole sensors, the event location shows fairly large lateral residuals with poor azimuthal coverage (top part of Fig. 5). The depth resolution is improved if head and direct phases are included. In this case, the event location is constrained better to the lower
reservoir formation (middle section of Fig. 5). Adding surface sensors, which allows us to obtain a wider azimuthal distribution, does clearly have the largest effect on the location residual. The location is now constrained much better laterally with no clear effect on depth resolution (lower part of Fig. 5). A detailed horizontally layered (1D) isotropic velocity model is used for locating events (see Fig. 2). An anisotropic model, as used by [3] for locating events, may result in a lateral shift, but has no effect on the individual location residual.

Fig. 5. Location residual plots for borehole sensor first arrivals only (top), borehole sensors with additional phase arrivals (middle), and combined borehole and surface sensors (bottom). Map view and cross-sections are shown. The stars mark the best-fitting event locations. Station locations are indicated by small triangles in the cross sections on the side.
Overall, using a combination of borehole and surface sensors to increase azimuthal coverage, together with additional waveform information for better depth resolution can significantly reduce the relative location uncertainty. However, at Decatur only few events are sufficiently well recorded at the surface. The main challenge remains the uncertainty in the velocity model, in particular the near-surface velocity heterogeneities. Therefore, absolute event location uncertainties remain large. However, methods such as waveform cross-correlation can be applied to refine relative event locations. An unambiguous assignment of a microseismic event to a specific formation of sometimes only a few 10’s of meters of thickness (e.g., storage formation or cap rock) requires an analysis of the full waveform also including later phases, and not just the first arrival. This is shown by [6] where sediment and basement events can be differentiated and an event migration pattern in individual clusters is resolved.

Further improvement can be achieved by using a Bayesian location algorithm [7, 8]. Given multiple events located close to each other, a Bayesian location identifies consistent patterns in the traveltime residuals within clusters and corrects implicitly for what appears to be a consistent bias at a given station [9]. This is well suited for the inversion of data from joint borehole and surface networks: due to their better azimuthal coverage, surface networks can provide better constraints on the location. However, they typically record far fewer events due to their inferior detection threshold. The Bayesian location can use these few well-constrained event locations to resolve consistent bias and thus improve the location uncertainty for the whole cluster. One of the very promising aspects of using a Bayesian location is the estimation of traveltime corrections with respect to the used velocity model. It can therefore help to identify any shortcomings in applied velocity models.

3. Source parameter analysis

The analysis of stress drop [10] and b-value [11] can provide additional information about the injection process. Systematic spatio-temporal variations of these parameters have been observed during injection operations [e.g., 12, 13] and can give an insight into the type of fracturing (new tensile fracture opening versus pre-existing fractures) and the related in-situ pore pressure and stress regime. Previously, pore-pressure could be linked to b-value [13] and stress drop [12, 14]. It is suggested that the decrease in b-value/ increase in stress drop with increasing distance from the injection point (as, e.g., observed for the Basel EGS induced seismicity) is likely caused by a decrease in pore-pressure. Previous studies suggest that hydrofracture-, mining-, and reservoir-induced earthquakes have lower average stress drop than natural seismicity due to possibly different tectonic setting, the amount of fluid present, or shallower hypocentral depths [see e.g., 15, 16]. Other studies find no difference between stress drops of natural and induced seismicity [17]. Since uncertainties in stress drop estimates are usually large, we need to focus on relative variations computed with the same methodology. In this study we focus on lateral stress drop variations at Decatur.

3.1. Spectral stacking

We compute displacement spectra using the multi-taper spectral estimation method [e.g., 18] over a 0.3 s long time window around the S arrival, starting 0.05 s before the available phase pick on all deep borehole sensors. Only spectra with a signal-to-noise (STN) above two, measured between 15 and 250 Hz, are included in the analysis. The displacement source spectrum can be described by [10]:

\[ A(f) = \frac{\Omega_0}{1+\left(\frac{f}{f_c}\right)^{2n}} \]  

where \( \Omega_0 \) is the average low-frequency level which is proportional to the moment magnitude \( M_w \). The spectral fall off rate \( n=2 \). \( f_c \) is the corner frequency and can be related to the stress drop by assuming a circular rupture [19] as follows:

\[ \Delta \sigma = \frac{7}{16} M_0 \left(\frac{f_c}{k \beta}\right)^3, \]  

where \( \beta = 2500 \text{ m/s} \) is the shear-wave velocity and \( k=0.37 \) for S-wave spectra.
We separate spectral source effects from attenuation effects along the travel path and site effects near the receiver by applying an iterative least-squares stacking approach [20, 21]. The stacking approach exploits the redundancy contained in recording many events by an array of receivers. As a result, we obtain an average source spectrum per event that has been corrected for path and receiver effects. The smaller magnitudes especially may still suffer from noise and we therefore excluded events with a moment magnitude smaller than -0.8. In addition, we require a minimum number of five traces per event.

The spectral stacking approach can only resolve relative variations between all source spectra and therefore an empirical correction to absolute spectra is necessary before the corner frequency can be estimated. A constant stress drop source model of 0.5 MPa is assumed. The reader is referred to [20] for further details about the approach. By fitting the data to a pre-defined stress drop model, we basically enforce a mean stress drop of this value for the entire dataset. The chosen value of 0.5 MPa is rather low, but allows us to resolve a corner frequency for a larger subset of events. Note that the focus of this study lies in resolving relative spatial source parameter variations, which can be estimated robustly. Absolute values depend solely on the chosen stress drop model and should neither be interpreted nor compared to other studies. Figure 6 shows some example source spectral fits for individual events. All events with large rms-misfit and corner frequency estimates above 400 Hz (red in Fig. 6) are excluded from further analysis.

Lateral variations in stress drop are observed (Fig. 7). Although individual stress drop estimates are highly scattered and usually have large uncertainties, we observe a systematic pattern within the microseismic cloud. The most prominent feature is an apparent distance dependence of stress drop from the injection point (Fig. 7 and 8). The apparent decrease of stress drop from the injection point is likely a result of insufficient attenuation correction. This arises from the fact that the azimuthal distribution of receivers is very limited (we are only using borehole stations close to the injection well for this analysis), and the resulting lack of crossing wave paths results in insufficient
separation of source- and propagation path effects. We therefore concentrate on lateral source parameter variations within a selected event cluster (Fig. 9). The cluster forms a linear feature with an extent of about 800 m in the direction of the maximum horizontal stress. All events within this cluster have a similar distance of about 1 km to the injection point and are therefore subject to similar path effects. Lower stress drop estimates are observed at the Southwestern side increasing towards the Northeast (Fig. 9 and 10). It is interesting to note that the Southwestern point is also the cluster nucleation point. This is in agreement with an observed change in b-value along the same direction from high b-values to lower b-values at the Northeast [6]. Although the overall spatio-temporal event evolution (clustered events) cannot be explained by a simple diffusive process, we seem to observe distance dependence within this cluster that could point to pore pressure diffusion processes at the cluster level. The change of stress drop with distance from the cluster nucleation point (Fig. 10) together with the change in b-value observed by [6] could suggest some fluid percolation process, where a localized constriction is broken up and pressure can diffuse into the next lateral compartment.

![Fig. 7. (a) Lateral stress drop variations of microseismic events (circles). Circle size is proportional to moment magnitude. The injection well (CCS1) is shown by the black square, the geophysical monitoring well (GMW) is shown by the black triangle, and some USGS stations are shown by the black diamonds.](image)

![Fig. 8. Stress drop versus distance to the injection point. Circle size is proportional to moment magnitude.](image)
Fig. 9. Lateral stress drop variations within a selected event cluster (colored circles) in map view (left), and two depth sections (middle and right). Other events are shown as grey circles. Circle size is proportional to moment magnitude. The injection well (CCS1) is shown as a square, the geophysical monitoring well (GMW) is shown as a triangle, and some USGS stations are shown as diamonds. The squares in the depth sections mark the deepest borehole sensors (PS3s) within CCS1.

Fig. 10. Distance dependence of stress drop to the Southwestern starting point for all events within the selected cluster. The mean stress drop in 50 m bins is shown by the red squares together with respective standard deviations (vertical bar). Circle size is proportional to moment magnitude.

It is interesting to note that the increase on stress drop happens over the first 300 m from the SW nucleation point of the cluster, and then flattens out. This is in principle similar to what was observed by [12] in Basel. However, we can unfortunately not compare these results to pore pressure because the closest reliable in-situ pressure measurement is in the monitoring borehole about 1 km away.

3.2. Multiple empirical Green’s functions

To further investigate and verify the observed change in Brune stress drops within the selected event cluster, we employ the multiple empirical Green’s function method (MEGF, see, e.g. [22]. The purpose of the MEGF method is to decouple the strong dependencies of path-, sensor-, and attenuation effects, when estimating seismic moment and corner frequency from a group of co-located microearthquakes. In the MEGF approach, we compute the spectral ratios of all possible event pair combinations for all available traces that exceed a given signal-to-noise ratio (in this case larger than 2). Each spectral ratio is then independent of the above path-, sensor-, and attenuation effects, as we assume the events to be collocated and, hence, only source radiation would differ between selected events. The accurate computation of seismic moment and corner frequency can then be used to compute stress drops [16]. Theoretical spectral ratios are fitted to the observed data in the frequency range from 15 to 200 Hz. Outside of these
frequencies, the noise was dominating and these parts were disregarded. We compare corner frequencies and seismic moments of subclusters of events within the above selected event cluster in Figure 11. A subcluster in this context is a subset of events within the cluster that have highly correlating waveform. The subcluster that is located closer to the southwest cluster nucleation point shows relatively smaller corner frequencies, compared to the subcluster that is further to the Northeast and away from the onset of event activity. This is consistent with the previous observation presented in Figure 10, and obtained via the spectral stacking method.

Fig. 11. Comparison of corner frequency and seismic moment for a subcluster close the cluster nucleation point (Southwest, blue) and a subcluster furthest away (Northeast, red).

4. Conclusions

Using a combination of borehole and surface sensors to increase azimuthal coverage, together with additional waveform information for better depth resolution can significantly reduce the relative location uncertainty of microseismic events. The main challenge remains the uncertainty in the velocity model. Absolute event location uncertainties often remain large. At Decatur only few events are sufficiently well recorded at the surface with strong unknown near-surface velocity heterogeneities. Advanced methods, such as waveform cross-correlation or Bayesian location methods, can be applied to refine relative event locations within only a few 10’s of meters of thickness (e.g., storage formation or cap rock). Furthermore, a detailed analysis of Brune-type stress drop, applying two different methods, reveals significant lateral variations. Although the overall spatio-temporal event evolution (clustered events) cannot be explained by a simple diffusive process, we seem to observe a distance dependence of stress drop within a microseismic cluster that could point to pore-pressure diffusion processes at the cluster level and suggests some fluid percolation process.

Acknowledgements

We thank A. Wüstefeld and K. Iranpour for providing MStudio/ray-tracing visualization. This work was supported as part of the Center for Geologic Storage of CO₂ (GSCO₂) an Energy Frontier Research Center funded by the U.S. Department of Energy, Office of Science, Basic Energy Sciences under Award # DE-SC0C12504.

References


