

## Characterization of hydraulic properties of rocks using probability of fluid-induced micro-earthquakes

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### ABSTRACT

*The use of borehole fluid injections is typical for exploration and development of hydrocarbon or geothermal reservoirs. Such injections often induce small magnitude earthquakes. The nature of processes leading to triggering of such microseismicity is still not completely understood. Here we consider induced microseismicity using as examples two case studies of geothermal reservoirs in crystalline rocks and one case study of a tight gas sandstone reservoir. In all three cases we found that the probability of induced earthquakes occurring is very well described by the relaxation law of pressure perturbation in fluids filling the pore space in rocks. This strongly supports the hypothesis of seismicity triggered by pore-pressure. Moreover, this opens additional possibilities of using passive seismic monitoring to characterize hydraulic properties of rocks on the reservoir scale with high precision.*

### INTRODUCTION

Injections of borehole fluids into surrounding rocks are used for developments of hydrocarbon or geothermal reservoirs. Such injections often induce small-magnitude earthquakes (see e.g., Zoback and Harjes, 1997; Fehler et al., 1998; Audigane et al., 2002; Rutledge and Phillips, 2003). The nature of such seismic activity is still under discussion (see e.g., Trifu, 2002). Possible mechanisms of induced microseismicity are, for example, hydraulic fracturing or slip initiation by redistribution of elastic stress. Another widespread hypothesis explaining the phenomenon of hydraulically-induced microseismicity (e.g., Nur and Booker, 1972; Pearson, 1981; Shapiro et al., 2002) is that the tectonic stress in the earth's crust at some locations is close to a critical stress causing brittle failure of rocks, for example, by sliding along pre-existing cracks. Increasing fluid pressure in a reservoir causes pressure in the connected pore space of rocks to increase (the pore space includes pores, cracks, vicinities of grain contacts, and all other possible voids in rocks). This leads to an increase of the pore pressure at the critical locations as well. Such an increase in the pore pressure consequently causes a decrease of the effective normal stress, usually acting compressionally on arbitrary internal rock surfaces. This leads to sliding along pre-existing, favorably-oriented, subcritical cracks.

The change of pore pressure in space and time is controlled by the diffusion process of pressure relaxation in fluids saturating the pores. Thus, if the hypothesis described above is correctly explaining at least one of the dominant mechanisms triggering fluid-induced microseismicity, then a number of diffusion-typical signatures should be observed in the spatio-temporal distributions of the earthquakes. Several of these signatures related to the temporal evolution of microseismicity clouds and supporting the hypothesis above are known. These are, for example a parabola-like forward triggering front and a back front of seismicity. Such fronts are visible on plots of event distances from the injection source versus their occurrence times (see e.g., Audigane et al., 2002; Shapiro et al., 2003; Parotidis et al., 2004). Another diffusion-related signature is a convergence of seismicity clouds to ellipsoid-similar figures by normalization of event coordinates by square roots of their occurrence times (see Shapiro et al., 2003). These signatures are related to the shape and the size of seismicity clouds. In contrast to them, we report here evidence of a completely

different nature supporting the hypothesis described above. This evidence is related to the spatial density of seismic events.

In the following we propose a statistical model describing the probability of triggering of micro-earthquakes. Then we briefly describe how the data of microseismic monitoring should be processed in order to obtain useful information from the spatial density of the event distribution. Finally, using three case studies we show that the theoretically-predicted feature of the spatial density of events can be very well observed. Such a feature additionally illuminates the physics of the fluid-induced microseismicity. Moreover, it opens a new way to estimate hydraulic properties of natural rocks at large spatial scales with high precision.

### STATISTICAL MODEL OF SEISMICITY TRIGGERING

We start with a theoretical model which is completely consistent with the above described pore-pressure relaxation hypothesis (PRH). The assumptions of this model are the following: we consider a point pressure source in an infinite, homogeneous, permeable, porous continuum. We assume that the hydraulic diffusivity of this continuum is independent of position and time. Due to fluid injection and the consequent process of pressure relaxation, the pore pressure  $p$  will change throughout the pore space. We assume that a critical value  $C$  of the pore pressure necessary for occurrence of a seismic event is randomly distributed in space. The statistical properties of  $C$  are assumed to be independent of position (i.e.,  $C(\mathbf{r})$  is a statistically homogeneous random field). If at a given point  $\mathbf{r}$  of the medium pore pressure  $p(t, \mathbf{r})$  increases with time, and at time  $t_0$  it becomes equal to  $C(\mathbf{r})$  then this point will be considered as a hypocenter of an earthquake occurring at time  $t_0$ . For simplicity, we assume that no earthquake will be possible at this point again. Then the probability of an earthquake occurring at a given point at a given time will be equal to  $W(C(\mathbf{r}) \leq p(t, \mathbf{r}))$ , which is the probability of the critical pressure to be smaller than the pore pressure  $p(t, \mathbf{r})$ . If the pore pressure perturbation caused by the fluid injection is a non-decreasing function (which is the case for realistic, step-function-like, borehole injection pressures) then this probability will be equal to

$$W = \int_0^{p(t, \mathbf{r})} f(C) dC,$$

where  $f(C)$  is the probability density function of the critical pressure. The pore pressure  $p(t, \mathbf{r})$  is the solution of the diffusion equation describing the process of pore pressure relaxation. According to this theoretical model the seismic criticality of rocks is defined by the quantity  $C$ . If  $C$  is high we speak about a stable domain of the medium. If  $C$  is low we mean a seismic critical domain. Our experience with microseismic data shows that usually  $C$  is of the order of  $10^2 - 10^5$  Pa.

The simplest possible distribution of the critical pore pressure is a uniform distribution  $f = 1/A$ , where  $A$  is a normalizing constant having the following physical meaning:  $A = C_{max} - C_{min}$ , with  $C_{max}$  and  $C_{min}$  standing for maximum and minimum possible critical pressures. In this case  $W = p(t, \mathbf{r})/A$  and the event probability is proportional to the pore pressure perturbation. A question arises: is such a distribution of micro-earthquakes observed in reality? A positive answer to this question would be of significant importance for our understanding of the physics of microseismicity triggering as well as for useful applications of this phenomenon. Our observations provide such a positive answer.

### PROCESSING OF PASSIVE MONITORING DATA

Below we will consider several examples of borehole fluid injections. The pore pressure relaxation in a porous medium surrounding an injection source is described by the differential equation of diffusion. In many situations realistic conditions of borehole fluid injections can be approximated by a point pore pressure source of constant strength  $q$  (this quantity has physical units of power) switched on at the time  $t = 0$ . The solution of the diffusion equation then has the following form (see Carslaw and Jaeger (1973), chapter 10.2.(2))

$$p = \frac{q}{4\pi Dr} \operatorname{erfc} \left( \frac{r}{\sqrt{4Dt}} \right). \quad (1)$$

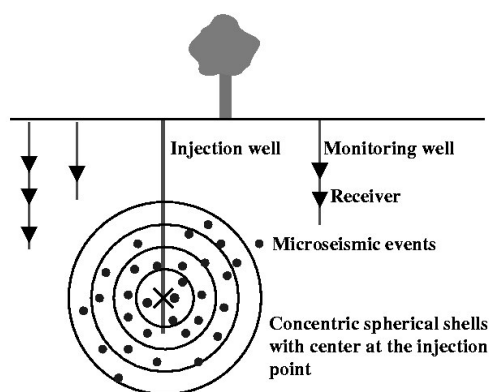
Here  $\operatorname{erfc}(x)$  is the complementary Gaussian error function and  $t$  denotes observation time. The quantity  $D$

is an important hydraulic parameter of the rock called hydraulic diffusivity. It is proportional to the Darcy permeability  $K$  of rocks (Shapiro et al., 1999, 2002):

$$D = \frac{NK}{\eta}, \quad (2)$$

where  $\eta$  is the pore-fluid dynamic viscosity and  $N$  is a poroelastic modulus related to porosity, bulk modulus of fluid and elastic moduli of the grain material and drained rock skeleton (see also Detournay and Cheng, 1993; Rice and Cleary, 1976; Van Der Kamp and Gale, 1983).

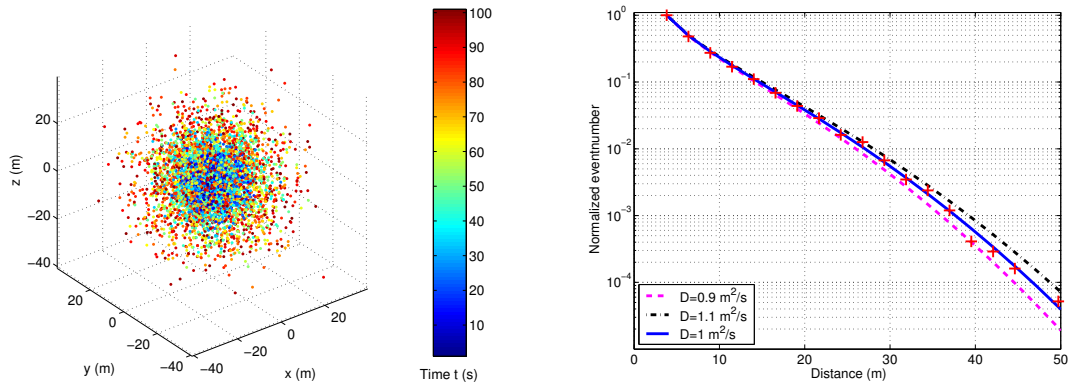
The event probability is proportional to the volumetric density of microseismic events. Therefore a comparison of the volumetric event density of microseismicity clouds induced by fluid injections with equation (1) is required. Before we do this for a microseismic data set obtained during a real borehole fluid injection, we consider a synthetic example.



**Figure 1:** A sketch of possible acquisition and processing geometry for passive monitoring data. Multicomponent receivers can be located in several boreholes (Soutz case and Cotton Valley case) or in a single monitoring borehole. Receivers also can be placed on the Earth surface (not shown here). The cross denotes the injection point. All other notations are explained on the figure.

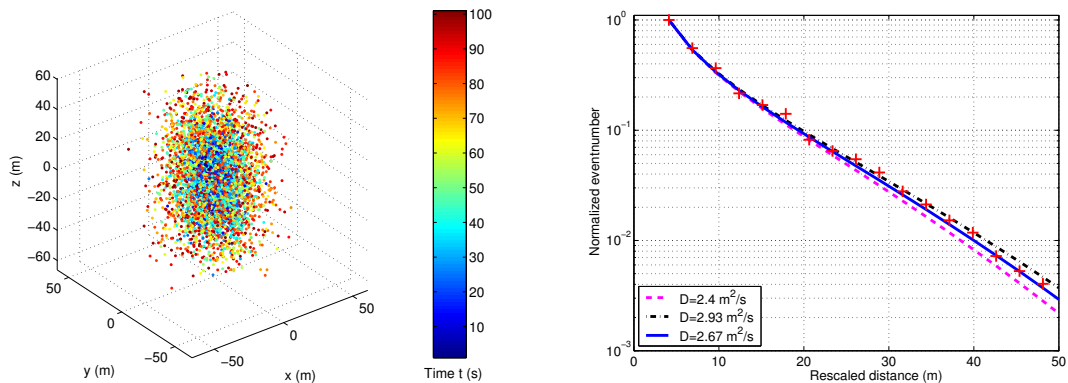
Figure 1 shows a sketch of a passive monitoring data acquisition. A synthetic cloud of microseismicity was obtained in a way consistent with the theory described above (such modeling was proposed and described by Rothert and Shapiro (2003)). This cloud is shown at the left of Figure 2. The following processing can be proposed to compute the probabilities of seismic events occurring. We count the number of events in concentric spherical shells with the center at the injection point (see Figure 1). This number of events is assigned to the mean radius of the corresponding shell. By normalizing event numbers to the shell volumes we obtain the event density. We normalize the event densities by the event density in the second spherical shell in order to work with non-dimensional quantities and to eliminate insignificant proportionality factors. (The first shell, which is actually a sphere around the injection point is not considered, because of the singularity of the diffusion equation solution).

To compare these numerical results with predictions of equation (1), the analytical function  $p(r, t)$  is also normalized by its value at the mean radius of the second spherical shell. Note that  $p(r, t)$  is a function only of the radial distance from the injection point. The time in the Gaussian error function is the total period of the injection (which is also equal to the total period of observation, i.e., event counting). A very good agreement of the analytical curve calculated with a hydraulic diffusivity of  $D = 1\text{m}^2/\text{s}$  and the event probability in the synthetic microseismic cloud (at the right of Figure 2) is not surprising. The numerical modeling is completely consistent with the PRH. The dashed line in Figure 2 was obtained using a 10 per cent smaller hydraulic diffusivity and the dashed-dotted line using a 10 per cent larger  $D$ . It should be noted that the observation time  $t$  (i.e., event counting time) can be smaller than the total injection time. If



**Figure 2:** Synthetic microseismicity in a hydraulically isotropic model with the hydraulic diffusivity of  $1\text{m}^2/\text{s}$ . At the left: a view of the cloud of seismic events. The colors correspond to the event occurrence time in seconds. At the right: spatial density of the microseismic cloud versus distance from the injection source (crosses). The lines denote the theoretical distribution given by equation (1).

$t$  is too large the normalized function  $p(r, t)$  will become insensitive to  $D$ . On the other hand, if  $t$  is too small the statistics of events can become insufficient for the processing. In all synthetic examples and case studies given in this paper the choice of  $t$  equal to the injection time is completely adequate.



**Figure 3:** Synthetic microseismicity in a hydraulically anisotropic model. Principal components of the hydraulic diffusivity are  $1\text{m}^2/\text{s}$ ,  $2\text{m}^2/\text{s}$ , and  $5\text{m}^2/\text{s}$ . Notations are as in Figure 2. At the left: A view of the cloud of seismic events is shown. At the right the spatial density of the microseismic cloud is displayed versus rescaled distance from the injection source. The theoretical curve corresponds to the arithmetic average of the principal components of the diffusivity tensor.

In the case of real data one more important complication must be taken into account. Hydraulic properties of natural rocks are usually anisotropic. Their hydraulic diffusivity is a second rank tensor (Shapiro et al., 1999, 2002, 2003). If the PRH is valid then the geometry of the microseismic cloud should be controlled by this tensor (see Figure 3, left). In order to compare the spatial distribution of the event density with the analytical solution (1), we have to transform the microseismic cloud obtained under pore pressure relaxation in an anisotropic medium into a microseismic cloud which would be obtained in an isotropic medium. For this a coordinate rotation is necessary from the original coordinate system (usually a geographical and depth coordinate system of event locations) to a coordinate system orientated along the principal axes of the diffusivity tensor. In addition, after this rotation, the new coordinates of events

must be scaled in relation to the square roots of the corresponding principal values of the diffusivity tensor. Validity of this scaling procedure is a consequence of the equivalence of the diffusion equation in an isotropic medium to the diffusion equation in an anisotropic medium by the transformation of coordinates described above (see also Carslaw and Jaeger, 1973, pages 38 - 43). The hydraulic diffusivity in the resulting isotropic diffusion equation is equal to the arithmetic average of the principal components of the diffusivity tensor. A detailed description of this transformation for microseismic clouds and how to find the relationship between principal components of the hydraulic diffusivity tensor for real situations using microseismic clouds is given in Shapiro et al. (1999, 2003). Note that the algorithms of inversion for a diffusivity tensor described in these publications are able to provide orientations and relations between the tensor's principal components which are quite exact. However, their absolute values (i.e., the magnitude of the arithmetic average of the principal components) are estimated less exactly. This disadvantage can be compensated by using the event probability analysis we describe here.

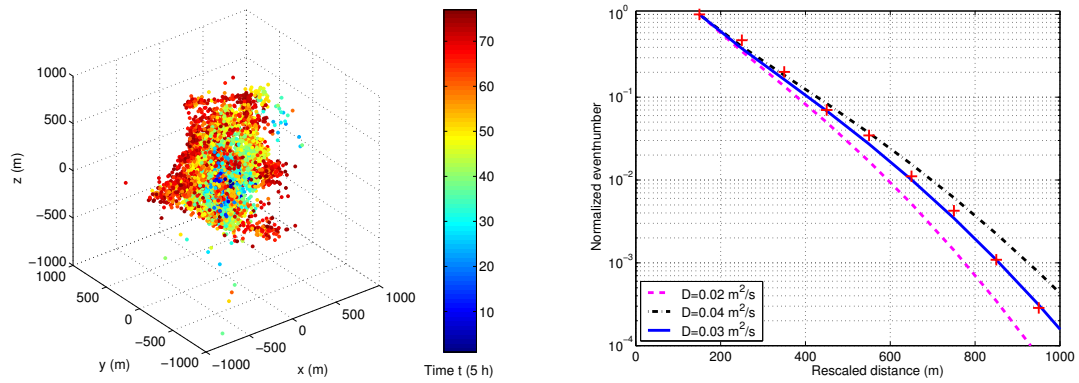
After the transformation of the seismicity cloud, a comparison between predictions of equation (1) and the event density is possible. It is completely analogous to the one in an isotropic medium. At the right of Figure 3 such a comparison is shown for the synthetic microseismicity from the left of Figure 3.

### CASE STUDIES

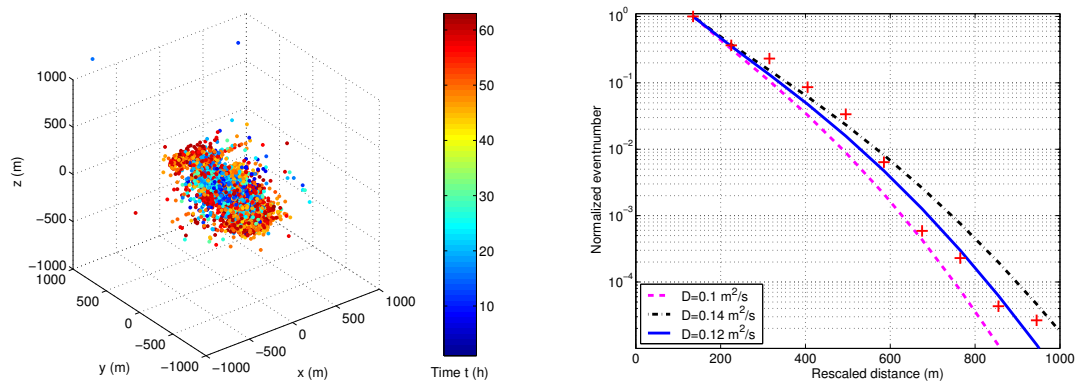
As an example of real microseismicity we consider a data set of microseismic events that were induced during the 1993 Hot Dry Rock (HDR) experiments performed at the Soultz-sous-Forêts (France) in the upper Rhine valley (Jones et al., 1995). About 25,300 m<sup>3</sup> of water were injected into the crystalline rock in a depth of approximately 2,920 m. The absolute location error of events was of the order of 20 - 80 m. The data set contains about 9150 events which were induced during 379 hours of injection. The cloud of micro-earthquakes is displayed at the left of Figure 4. The tensor of hydraulic diffusivity in Soultz is characterized by a significant anisotropy with an approximate relation of principal components as 7 : 19 : 52 (Shapiro et al., 2003). After scaling the microseismic cloud the event density can be compared with the distribution predicted by equation (1). The average of the principal components of the hydraulic diffusivity  $D$  can then be fitted to match the data. The best-fit curve is shown in Figure 4 right. This fit provides an estimate of  $D = 0.03 \text{ m}^2/\text{s}$ . An excellent agreement of the observed event density with the theoretical curve given by equation (1) is evident.

Independent methods of estimating hydraulic diffusivity at the same location yield similar values: the average principal component of the diffusivity tensor can be computed from results (again based on microseismicity analysis but using completely different features, i.e., fitting of the triggering front) given in Shapiro et al. (1999) to be equal to  $D = 0.023 \text{ m}^2/\text{s}$ . The estimate of the apparent hydraulic permeability from the same location based on a borehole injection/flow test and reported in Jung et al. (1996) also can be used to compute the hydraulic diffusivity. This approximately yields  $D = 0.022 \text{ m}^2/\text{s}$ . The microseismicity-based method of estimating hydraulic diffusivity used in Shapiro et al. (1999) provides order of estimates only. The error estimates from this publication are  $\pm 50$  per cent. The borehole-based estimates characterize the rocks in the vicinity of boreholes. In turn, two additional curves (Figure 4 at the right) corresponding to  $D = 0.03 \pm 0.006 \text{ m}^2/\text{s}$  show that the uncertainties in the average diffusivity are smaller than  $\pm 20$  per cent. Therefore, the event probability based approach is more precise for characterizing rocks on a kilometer-scale than those proposed in Shapiro et al. (1999).

We also applied the new approach to data obtained during fluid-injection experiments at the Fenton Hill Hot Dry Rock site (New Mexico, USA). In 1983 for about 62 hours more than 21,600 m<sup>3</sup> of water were injected into the crystalline rock in a depth of about 3,460 meters (Fehler et al., 1998). During this time interval about 9355 events were recorded and located. The accuracy of the location was better than 100 m (House, 1987). The tensor of diffusivity is characterized by an anisotropy with an approximate relation of the principal components as 2 : 3 : 6 (Shapiro et al., 2003). The best-fit of the analytical function (1) gives an estimate of  $D = 0.12 \text{ m}^2/\text{s}$  and is shown in Figure 5. Fitting of the triggering front and the back front of seismicity (see Parotidis et al., 2004) provide estimates of  $D = 0.14 \text{ m}^2/\text{s}$ . A good agreement of these values is evident. Additionally Figure 5 (at the right) shows two curves with  $D = 0.1 \text{ m}^2/\text{s}$  and  $D = 0.14 \text{ m}^2/\text{s}$  which correspond to  $\pm 20$  per cent deviation. Three data points in the distance range of 300 - 500 m are located even outside this deviation interval. We explain this by an influence of a strong hydraulic heterogeneity of the medium.



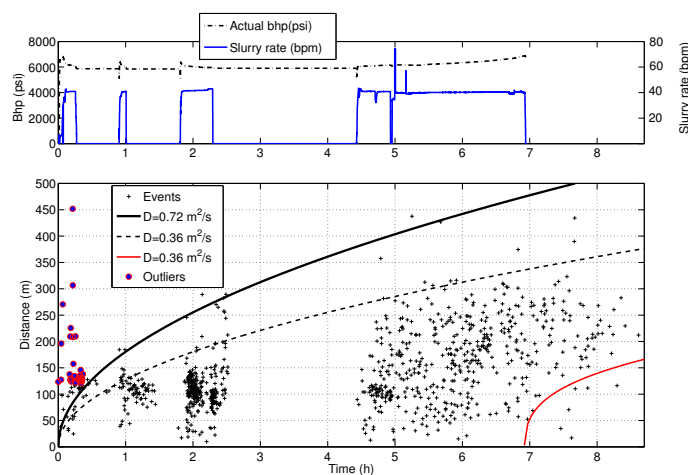
**Figure 4:** Micro-earthquakes of the Soutz-sous-Forêts experiment of 1993. At the left: a view of the cloud of seismic events. The colors correspond to the event occurrence time. At the right: spatial density of the microseismic cloud versus rescaled distance from the injection source (crosses). The line denotes the theoretical distribution given by equation (1).



**Figure 5:** Micro-earthquakes located in December 1983 during the hydraulic injection into crystalline rock at Fenton Hill, a HDR site in the USA (for details and further references see Fehler et al. (1998)). Notations are the same as in Figure 2. At the left: a view of the cloud of seismic events. At the right: spatial density of the microseismic cloud versus rescaled distance from the injection source.

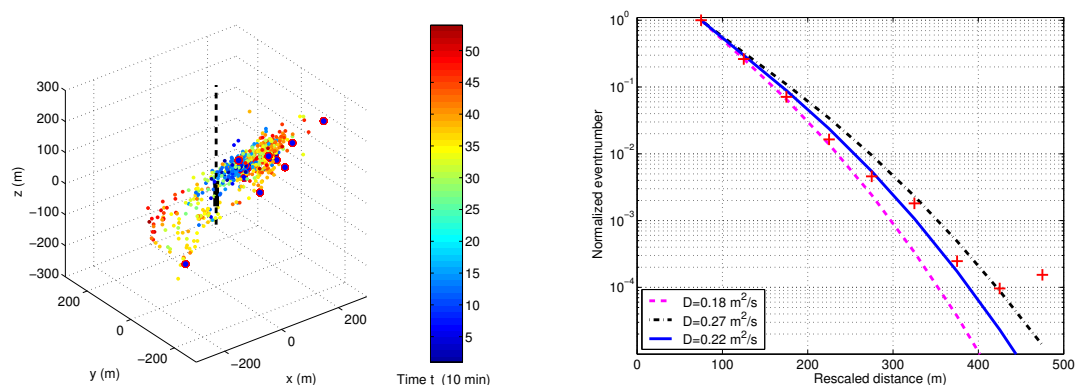
Especial attention is required by the third example from the Carthage Cotton Valley gas field (East Texas, USA). In contrast to the previous two case studies concerning crystalline rocks, the Cotton Valley data set was obtained by a hydraulic fracturing of sediments. On May 14th, 1997 a hydraulic fracture treatment was performed and for about 7 hours a total fluid volume of more than 1,100 m<sup>3</sup> was injected into sandstones at the depth of 2,756-2,838 meters (Rutledge et al., 2004). The localized 994 microseismic events occurred in the depth range from 2,750 to 2,850 meters (Urbancic et al., 1999). The event location errors ranged from 14 m near the treatment well to 24 m at the wings of the eastern and western edges of the fracture zones.

In order to discuss this case study in more detail we show here the distance versus time plot for these micro-earthquakes (Figure 6). Two parabolic curves corresponding to the triggering front  $r = \sqrt{4\pi Dt}$  (see Shapiro et al., 2002, for the theory of the triggering front) with two diffusivity estimates  $D = 0.36$  m<sup>2</sup>/s and  $D = 0.72$  m<sup>2</sup>/s are shown here. The majority of events is located below a parabolic envelope with a diffusivity ranging from  $D = 0.36$  m<sup>2</sup>/s to  $D = 0.72$  m<sup>2</sup>/s. A few events occurred relatively far away from the injection source at early times (see the large circles at Figure 6 and Figure 7 at the left). We propose that they were caused by hydraulic fracturing rather than by a diffusive process of pore-pressure relaxation. Probably, a long thin crack filled with fluid was quickly formed during the first hour of the hydraulic fracturing. This crack has influenced the effective hydraulic properties of the medium. An analysis of the shape of the square-root-occurrence-time normalized microseismicity cloud of the Cotton Valley (Rentsch, 2003) shown that diffusivity tensor is characterized by a very strong anisotropy. The approximate relation of principal components is 1 : 3 : 46. After the transformation of coordinates we can analyse event probability (see Figure 7, right). Again we see that the majority of events was triggered by a diffusive process. Between 400 m and 500 m distance from the injection source we observe more events than predicted by equation (1). These outliers correspond to the events which probably occurred during hydraulic fracturing at the very beginning of the injection. The estimation of  $D = 0.22$  m<sup>2</sup>/s  $\pm$  20 per cent represents an arithmetic average of the principal components of the effective hydraulic diffusivity tensor characterizing the seismically active volume of the Carthage Cotton Valley field. The value  $D = 0.72$  m<sup>2</sup>/s is close to the maximum principal value of the diffusivity tensor.



**Figure 6:** Top: actual bottom hole pressure (psi) and slurry rate (bpm) versus time. Bottom: event distance from the injection point versus occurrence time for the Cotton Valley microseismicity. Two black lines show locations of the triggering front for  $D = 0.36$  m<sup>2</sup>/s and  $D = 0.72$  m<sup>2</sup>/s. The red line represents the back front for  $D = 0.36$  m<sup>2</sup>/s. The Cotton Valley diffusivity is characterized by strong anisotropy (Rentsch, 2003; Rindschwentner, 2001). We propose that events denoted by circles above the solid parabolic envelope were triggered by the hydraulic fracturing. Data are courtesy of T. Urbancic (ESG).

The resulting diffusivity estimate seems to characterize the medium after the hydraulic fracturing, because it is mainly based on the statistics of events occurring after the outliers.



**Figure 7:** Micro-earthquakes induced in the Cotton Valley Field, East Texas during a fluid injection (hydraulic fracturing) experiment. Notations are the same as in Figure 2. At the left: a view of the cloud of seismic events. The large circles correspond to events which are assumed to occur during fracturing. These are the same events as those shown by circles at Figure 6. The solid black line represents the injection interval and the dashed line the interpolated treatment well. At the right: spatial density of the microseismic cloud versus rescaled distance from the injection source.

The diffusivity anisotropy reflects already the influence of the hydraulically initiated fracture as well as the influence of the lamination of the sedimentary rocks. This well explains strong anisotropy of the diffusivity and its relatively high values as well as the observation that the largest principal component is approximately horizontal.

Figures 6 and 7 document application of three independent approaches (parabolic envelope, backfront and events density) to the Votton valley data set. They all give results of the hydraulic diffusivity in the same order of magnitude. The estimation of the hydraulic diffusivity using the parabolic envelope is mainly based on later events (4.5 - 9 hours after the injection start) by fitting the outer parabolic boundary of the spatio-temporal distribution. Most events are located below the parabolic envelope on an  $r - t$  plot. A second signature, the back front, also fits the data after the injection stop (see the red curve on Figure 6). The seismicity back front describes the propagation of the pore-pressure maximum for a boxcar pressure source of the duration of 6.92 hours. Behind the back front the triggering of seismicity is very unprobable, corresponding to our model. For a detailed description of the back front signature we refer to Parotidis et al. (2004). The third signature, the event density, considers all events between the injection start and the injection stop. This signature also fits the observed event number per unit volume up to 450 meters distance from the injection source (Figure 7). According to these three signatures it is a reasonable assumption that one of the dominant triggering mechanism at Cotton Valley is the pore-pressure diffusion. The estimated diffusivity reflects the effective hydraulic diffusivity of the medium. This effective hydraulic diffusivity represents a complex average (upscaling) over the intact rock, the hydraulic fractures and the hydraulic heterogeneities.

This case study firstly shows that a plot like one shown in Figure 6 can help to identify events which are probably caused by different triggering mechanisms, for example, hydraulic fracturing and pore-pressure diffusion. One can expect that the diffusion triggered events are located mainly below a parabolic envelope of the triggering front. Events which we can identify as hydraulic fracturing events are outliers - early far events located above the triggering front. This, however, does not exclude that some events triggered by hydraulic fracturing can be located below the parabolic envelope.

The case study of the Carthage Cotton Valley gas field corresponds to the most complex geological and technical conditions. It does not comply with our model assumptions in a number of ways. The hydraulic diffusivity is a function of space because of heterogeneity of the medium (e.g., layering). It is also a function of time because of creating a hydraulic fracture. The last process increased the spatial heterogeneity of the diffusivity. Moreover, the fluid entering the hydraulic fracture was varied significantly



during the period of injection from treated water to a crosslinked gel and to a mixture of the gel and proppant sand. Finally, Figure 6 shows that the pressure regime was not exactly a non-decreasing function. Such complexity of the Cotton Valley experiment makes meaningful a very detailed analysis of the relationship between the time-dependent seismicity rate and pressure and flow rate curves. This analysis is beyond of the scope of our study whose main goal was to introduce a new quantitatively interpretable signature of induced microseismicity, the probability of events. However, already our analysis of the gross statistics of the microseismicity shows that even for this complex case study the seismicity statistic can be described by a diffusion triggering process. The corresponding diffusivity tensor is a rather complex (effective) average of the diffusivity tensor of the propped hydraulic fracture and the heterogeneous diffusivity of the surrounding rocks. It effectively characterizes several processes: the diffusion of fluid-pressure in the propped fracture, the leak-off pressure diffusion from the hydraulic fracture into the matrix and natural crack system of the surrounding rocks and the diffusion of pressure in the rocks. This effective diffusivity should be significantly larger than the diffusivity of the intact rocks. However, it can be proposed as an orientation value for properties of the hydraulic system of the stimulated reservoir after the hydraulic fracturing.

## CONCLUSIONS

Observations show that clouds of fluid-injection-induced microseismicity require hours or days to reach a size of several hundred meters. This is definitely too slow a process to be described just by the elastic stress equilibration which has a characteristic time of elastic wave propagation, i.e., seconds or less. Our analysis provides additional evidence in favor of the hypothesis that the process of pore pressure relaxation is at least one of the dominant triggering mechanisms of fluid-induced microseismicity. According to this hypothesis, spatio-temporal distributions of microseismicity are controlled by the hydraulic diffusivity of rocks as well as by the distribution and degree of rock criticality. We define the rock criticality as the minimum pore pressure required at a given location to trigger a seismic event. The criticality of rocks can be described by rather simple statistical models like a uniform probability density function. We propose a simple statistical model of the fluid-induced seismicity. Using this model and analyzing three case studies we have shown that the probability of induced seismic events is well described by the law of pore pressure relaxation (diffusion, see eq. (1) and Figure 4, Figure 5 and Figure 7). Using this observation we show that the spatial distribution of the density of earthquakes provides a possibility to estimate the hydraulic diffusivity on a kilometer scale with a high precision. The case study of the Carthage Cotton Valley gas field indicates a way of distinguishing seismic events probably triggered by the process of hydraulic fracturing from seismic events triggered by pore pressure relaxation. For this an  $r - t$  plot of the induced seismicity (like Figure 6) can be analyzed. Early events which are located clearly above the parabolic envelope of diffusion-triggered earthquakes are proposed to be triggered by hydraulic fracturing. Such a hypothesis could be of a high practical importance and requires further detailed studies.

The assumptions of our theoretical model are summarized in the section "Statistical model of seismicity triggering". We consider them as simplifying rather than restrictive assumptions. Clearly, in reality, hydraulic properties are heterogeneously distributed in rocks. Also the statistics of seismic criticality can be much more complex than assumed here. We think, however, that the model proposed here correctly reproduces the "large-scale-picture" of the phenomenon providing us with accurate, scaled-up, effective hydraulic characteristics of rocks. The case studies support the validity of such a conclusion.

It is clear also that explaining the overall statistics of seismicity is not enough to explain triggering mechanisms of all events. Already the Cotton Valley example indicates at least two important mechanisms at work: hydraulic fracturing and pore-pressure diffusion. Hydraulic heterogeneities and/or heterogeneities of the criticality of the medium can also be of importance. These heterogeneities may exist before injection started. However, they may also be caused by the injection (Rutledge and Phillips, 2002). The corresponding events may be triggered by non-linear diffusion or a process of tensile opening and propagation of small cracks (Dvorkin and Nur, 1992). Such events can also be found below the envelope of  $r - t$  plots. Coupled processes like thermo-poroelastic stress relaxation may also be of significance. Finally, other signatures of seismicity, such as strength of events or fault plane solutions, which are now not included in our model, may be included in our future analyses.

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