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

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ABSTRACT

In this paper we present a new approach for estimating the hydraulic parameters of rocks using an analysis of the spatial distribution of microseismic events induced by fluid injection. Such a monitoring of microseismicity can be used to characterize rocks in terms of hydraulic properties. Knowledge about these properties is important especially for an optimized developement of geothermal or hydrocarbon reservoirs. In fluid-saturated rocks, pore pressure changes are controlled by a diffussive process of pressure relaxation. Our approach is based on the hypothesis that the propagation of triggering of injection-induced microseismic events can also be described by a diffusive process of pore pressure relaxation. Due to this diffusive process a number of diffusion-typical signatures may be observed in spatial distributions of induced events. Such signatures are, e.g., pore pressure and the number of occurred events versus the distance from the injection point. We derived an appropriate relation between the number of triggered microseismicity to the distances from the injection point using probability considerations of event occurring. This spatial distribution of the density of microseismicity provides the possibility of large-scale in-situ estimations of hydraulic diffusivity. In this context large spatial scale means of the order of kilometers depending on the size of the seismically active region. Applications of the method to numerical data and to real data of the Soultz-sous-Forêts geothermal Hot Dry Rock site (France), the Fenton Hill Hot Dry Rock site (New Mexico, USA) and the Carthage Cotton Valley gas field (East Texas, USA) are demonstrated.

INTRODUCTION

Injections of borehole fluids into surrounding rocks are used for developments of hydrocarbon or geothermic reservoirs. Such injections often induce small-magnitude earthquakes (see e.g., Zoback and Harjes, 1997; Fehler et al., 1998; Audigane et al., 2002). The nature of such a seismic activity is still under discussion (see e.g., Trifu, 2002). In this paper we show that the probability of the induced earthquakes is remarkably well described by the relaxation law of pressure perturbations due to fluids filling the pore space in rocks. This strongly supports the hypothesis that the triggering of induced seismicity is controlled by the pore pressure relaxation. This fact opens additional new possibilities to characterize hydraulic properties of rocks on a kilometer-scale with high precision.

One widespread hypothesis explaining the phenomenon of the hydraulic 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 preexisting 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 preexisting, 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 pores. Thus, if the hypothesis described above is correctly explaining at least one of 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 are known (see e.g., Audigane et al., 2002; Shapiro et al., 2003).

Here we report evidence of a completely different nature supporting this hypothesis. This evidence additionally illuminates the physics of the fluid-induced microseismicity. Moreover, it opens a new way to estimating hydraulic properties of natural rocks at large spatial scales with high precision.

STATISTIC MODEL OF SEISMICITY TRIGGERING

We start with a simple model which is completely consistent with the above described pore-pressure relaxation hypothesis (PRH). We consider a point-like source of a fluid injection in an infinite porous continuum. Due to a 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 is randomly distributed in space. If in a given point $\boldsymbol{r}$ of the medium at a given time t the pore pressure $p(t, \boldsymbol{r})$ exceeds $C(\boldsymbol{r})$ then this point will be considered as a hypocenter of an earthquake occurred at time t. For simplicity, we assume that no earthquake will be possible at this point again. Then, the probability of an earthquake occurrence at a given point at a given time will be equal to $W(C(\boldsymbol{r}) \leq p(t, \boldsymbol{r}))$, which is the probability of the critical pressure to be smaller than the pore pressure $p(t, \boldsymbol{r})$. If the pore pressure perturbation caused by the fluid injection is a non-decreasing function (what is the case for realistic, step-function-like, borehole injection pressures) then this probability will be equal to

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

where $f(C)$ is the distribution function (probability density function) of the critical pressure. The pore pressure $p(t, \boldsymbol{r})$ is a solution of a diffusion equation describing the process of the pore pressure relaxation.

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 $W = p(t, \boldsymbol{r})/A$. Therefore, in such a simple model the event probability is only 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 source of a constant pore pressure perturbation q switched on at the time 0. The solution of the diffusion equation then has the following form (see Carslaw and Jaeger (1973), chapter 10.2.(2))

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

Here $erfc(x)$ is the complimentary Gaussian error function. For infinite observation time this equation reduces to $p = q/(4\pi Dr)$. Quantity D is an important hydraulic parameter of the rock called hydraulic diffusivity. It is proportional to the Darcy permeability of rocks. 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 will consider a synthetic example.

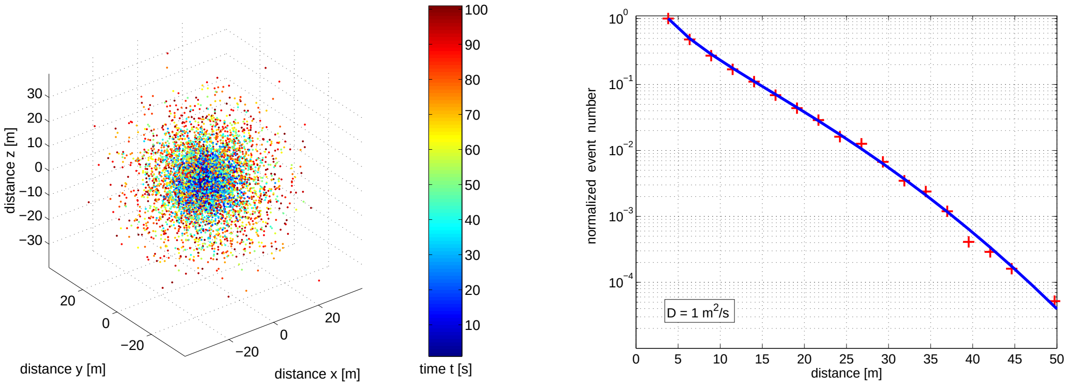

Figure 1: A synthetic microseismicity in a hydraulically isotropic model with the hydraulic diffusivity of $1m^2/s$. At the left: A view of the cloud of seismic events. The colors correspond to the event occurrence time in seconds. At the right: Spatial density of the microseismic cloud versus distance from the injection source (crosses). The line denotes the theoretical distribution given by equation (1).

A synthetic cloud of microseismicity obtained by the modeling consistent with the theory described above (such a modeling was proposed and described by Rothert and Shapiro (2003) is shown at the left of Fig. 1. The following processing can be proposed to compute the occurring probability of seismic events. We count the number of events in concentric spherical shells with the center at the injection point. 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 median radius of the second spherical shell. 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). By fitting the analytical function to the spatial density of events it is possible to estimate the value of the scalar hydraulic diffusivity. A very good agreement of the analytical curve (using D=$1m^2/s$ for calculation) and the event probability in the synthetic microseismic cloud (at the right of Fig. 1) is not surprising. The numerical modeling has been performed in a complete agreement with the PRH.

Before such a comparison can be done for real data the following important complication must be taken into account. Hydraulic properties of natural rocks are usually anisotropic. Their hydraulic diffusivity is a second rank tensor. If the PRH is valid then the geometry of microseismic clouds should be controlled by this tensor (see Fig. 2, left). In order to compare the spatial distribution of the event density with the analytical solution (1), we have to transform a microseismic cloud obtained under conditions of a pore pressure relaxation in an anisotropic medium into a microseismic cloud which would be obtained in an isotropic medium. This can be done by scaling the original event cloud along principal axes and in relation of the inverse square roots of the principal components of the hydraulic diffusivity tensor. Such a 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 scaling of coordinates described above (see also Carslaw and Jaeger, 1973). 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 substantiation of this scaling for microseismic clouds and how to find the relationship between principal components of the hydraulic diffusivity tensor for real situations is described in (Shapiro et al., 1999, 2003).

After scaling, 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 Fig. 2 such a comparison is shown for the synthetic microseismicity from the left of Fig. 2.

As an example of a real microseismicity we consider a data set of microseismic events that were induced

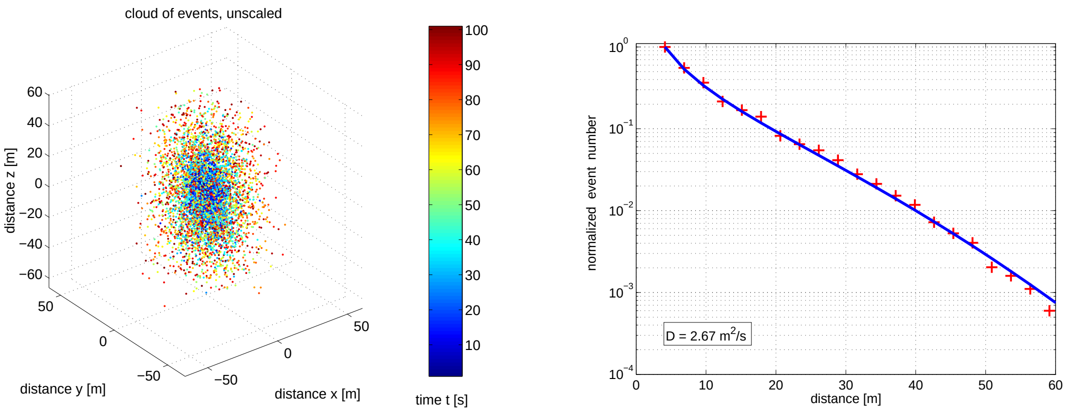

Figure 2: A synthetic microseismicity in a hydraulically anisotropic model. Principal components of the hydraulic diffusivity are $1m^2/s$, $2m^2/s$, and $5m^2/s$. Notations are as in Fig. 1. At the left a view of the cloud of seismic events is shown. At the right the spatial density of the microseismic cloud is displayed versus distance from the injection source.

during the 1993 Hot Dry Rock (HDR) experiments performed at the Soultz-sous-Forêts (France) in the upper Rhine valley (Dyer et al., 1994). 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 Fig. 3.

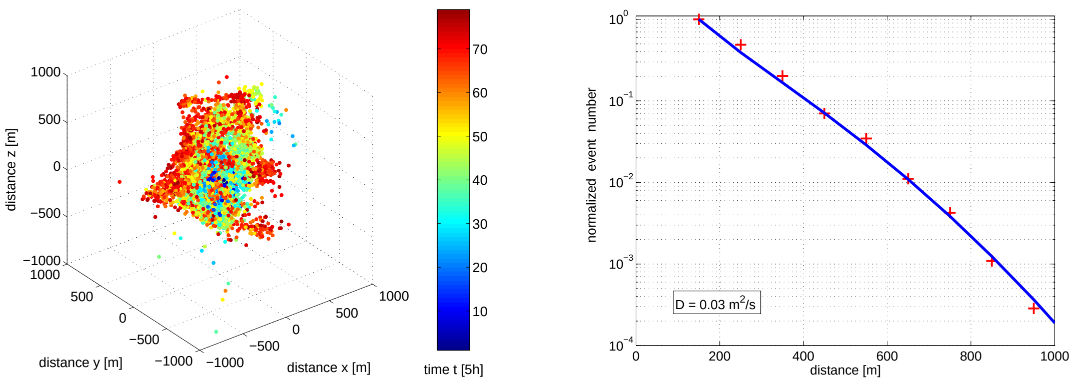

Figure 3: Micro earthquakes of the Soultz-sous-Forêts experiment of 1993. At the left: A view of the cloud of seismic events. The colors correspond to the events occurrence time. At the right: Spatial density of the microseismic cloud versus distance from the injection source (crosses). The line denotes the theoretical distribution given by equation (1).

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 Fig. 3 right. This fit provides an estimate of $D = 0.030m^2/s$. 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 given in (Shapiro et al., 1999) to be equal to $D = 0.023m^2/s$. The estimate of the apparent hydraulic permeability from the same location based on borehole data and reported in (Jung et al., 1996) also can be used to compute the hydraulic diffusivity. This approximately yields $D = 0.022 m^2/s$. The microseismicity-based method of estimating hydraulic diffusivity used in (Shapiro et al., 1999) provides 'order of' estimates only. The borehole-based estimates characterize the rocks in the vicinity of boreholes. Thus, we expect that the event-probability-based approach described above is more precise for characterizing rocks on a kilometer-scale than those proposed in (Shapiro et al., 1999).

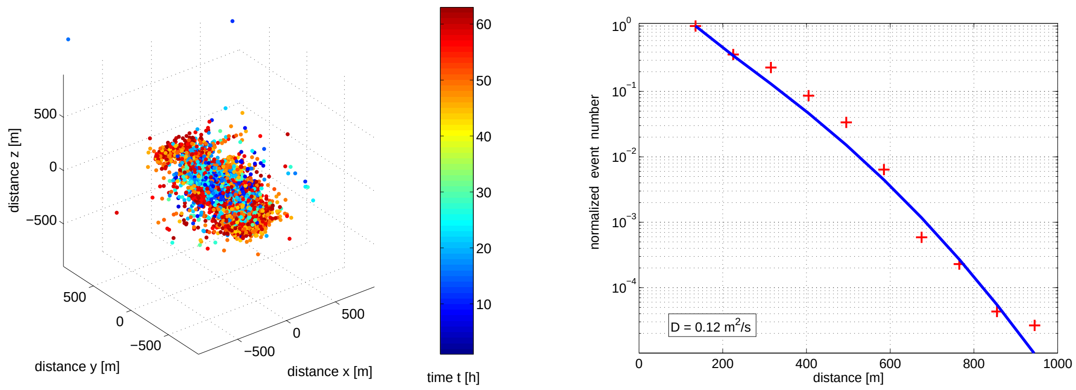

Figure 4: Micro earthquakes located in December 1983 during the hydraulic injection into crystalline rock at Fenton Hill, a HDR site in the USA (for details and further references see Fehler et al. (1998)). Notations as at Fig.1. At the left: A view of the cloud of seismic events. At the right: Spatial density of the microseismic cloud versus distance from the injection source.

Our new approach was also applied onto data of the Fenton Hill Hot Dry Rock site (New Mexico, USA) and the Carthage Cotton Valley gas field (East Texas, USA).

In 1983 the data set of Fenton Hill was obtained during the 'massive hydraulic fracture experiment' (MHF). About 21,600 m^3 of water was injected into crystalline rocks in a depth of about 3,460 meters for about 62 hours (Fehler et al., 1998). For this time interval the data set contains 9355 events. The best-fit analytical function results an estimate of $D = 0.12 m^2/s$ and is shown in Fig. 4.

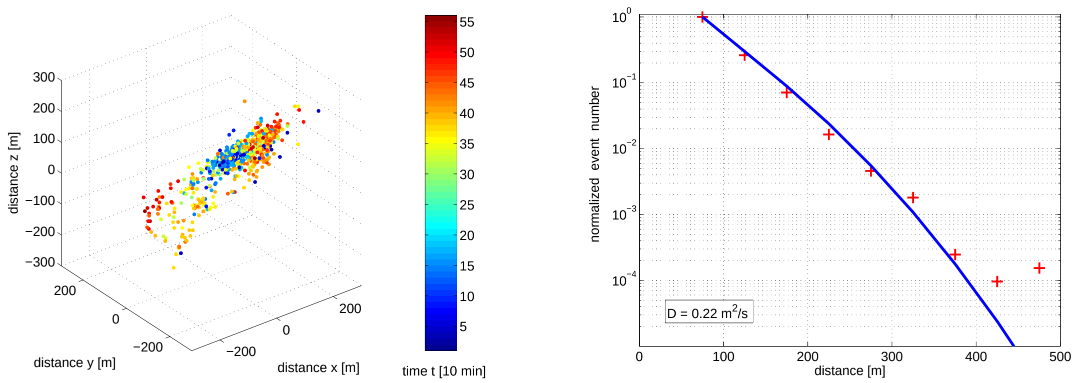

Figure 5: Micro earthquakes induced in the Cotton Valley Field, East Texas during a fluid injection (hydraulic fracturing) experiment. Notations are the same as on Fig. 1. At the left: A view of the cloud of seismic events. At the right: Spatial density of the microseismic cloud versus distance from the injection source.

The data of Cotton Valley was obtained by a hydraulic stimulation of sediment. In May 1997 hydraulic fracture treatments are performed and a consortium of operators and service companies have seismically monitored this treatments. For about 7 hours a total fluid volume of more than 1,100 m^3 was injected into sandstones at the depth of 2,756-2,838 meters. The observed 994 microseismic events occurred in the depth range from 2,750 to 2,850 meters (Urbancic et al., 1999). Our estimation of $D = 0.22\ \frac{m}{s^2}$ (see Fig. 5) represents an average value for the hydraulic diffusivity of the seismically active volume of the Carthage Cotton Valley field.

Our experience with available fluid injection induced microseismic clouds shows that equation (1) predicts probability of seismic events very well (see Fig. 3, Fig. 4 and Fig. 5). This is a strong indication in favor of the PRH.

CONCLUSIONS

The process of pore pressure relaxation is at least one of the dominant triggering mechanisms of the fluid-injection-induced microseismicity. According to this hypothesis, spatio-temporal distributions of microseismicity are controlled by the hydraulic diffusivity of rocks as well as by distribution and the degree of rock criticality. The criticality of rocks can be described by rather simple statistical models as proposed in this paper. Moreover, as we have demonstrated, the spatial distribution of the density of earthquakes provides the possibility to estimate the hydraulic diffusivity on a kilometer scale with high precision.

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