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Estimating the Permeability from Fluid-Injection Induced Seismic Emission

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ABSTRACT

This paper is a first result of the cooperative research between GOCAD and WIT Consortia. We develop a new technique for estimating permeability of rocks using the spatio-temporal distribution of the borehole-fluid-injection induced seismic emission. We show here how the technique works in the two different fractured zones in crystalline rocks: German KTB- and French Soultz experiments.

INTRODUCTION

During hydraulic-fracturing experiments or other experiments with borehole-fluid injections a micro-seismic activity can be observed. This was the case in the German Continental-Deep-Drilling Borehole at a depth interval of $9030-9100m$(Engeser, 1996). About $200m^3$ of KBr/KCl brine were injected approximately 40 hours. The fluid injection induced almost 400 micro earthquakes in a spatial domain extending to $500-700m$ from the borehole in a lateral direction in the depth range $7.5-9km$ (Zoback and Harjes, 1997).

Very similar situation was observed in France, in the Soultz-sous-Forets area, located in the Rhine Graben (Alsace). This area is considered as an important laboratory for development of low-enthalpy geothermal energy (Hot Dry Rock). Previous works have demonstrated that the underlying granitic basement at Soultz contains a large proportion of sub-vertical fractures favourably oriented for acceptance of artificial hydraulic circulations. In September 1993, a high flowrate stimulation performed at 1850-3400m depth in GPK1 well generated a large seismic cloud. More than 9395 seismic events have been recorded by BRGM and CSMA (Camborn School of Mines, UK) during this period. The center of the injection zone have been estimated at a depth of 2922 m. Most of the seismic events are oriented N-S and directed mainly towards the North (by about 500m) and upwards (by about 400m, though downward growth was also observed) (Gerard et al., 1994)

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In this paper we show how to use the spatio-temporal distribution of such seismic events in order to estimate the average permeability of rocks. It is well known that the permeability is a highly fluctuating parameter of rocks strongly influenced by the presence of cracks and other heterogeneities of the pore space. Its estimations can vary by orders of magnitude even for adjacent locations. Moreover, permeability measurements are scale dependent. Thus, the large spatial scale measurements of the reservoir permeability (like seismicity-based estimations) cannot be replaced by laboratory measurements.

The physical background of such estimations is as follows. It is assumed that the state of stress in the rock is close to a critical one, i.e., the crust is in a failure equilibrium. Therefore, small perturbations of this state can lead to induced micro seismicity. An increase of the pore pressure caused by fluid injection changes the effective normal stress as well as the friction coefficients of the rock mass. Thus, the temporal onset of the micro seismicity relative to the beginning of the injection is interpreted as a time delay Δt necessary for the pore-pressure diffusion to cause a sufficiently large perturbation Δp of the pressure at a given distance L to trigger seismic events. The necessary value of Δp is a strongly fluctuating quantity and, therefore, usually the following rough estimation is used (e.g., Ohtake, 1974; Fletcher and Sykes, 1977; Simpson et al, 1988, Talwani and Acree, 1985)

$$D \approx L^2/\Delta t, \tag{1}$$

where D is the hydraulic diffusivity.

In the case of time-harmonic pore-pressure perturbations, an estimation of the diffusivity can be obtained directly from phase-shift information (analogous to tidal-tilt analysis: see, e.g., M. Westerhaus, 1996). In such a case, no information about Δp is required and a well-constrained estimation of D is possible (see also our discussion in section 6). However, in hydraulic-fracturing experiments, the pore pressure perturbation is not time-harmonic. Moreover, to first approximation, it is equal to a step function.

In this paper we follow the above described physical concept to estimate the permeability from the injection-induced seismicity. We propose an approach to interpret the pore-pressure diffusion which leads to a new technic of the permeability estimation.

PORE-PRESSURE DIFFUSION

We approximate the real configuration of the fluid injection by a point source of the pore pressure in an infinite homogeneous isotropic poroelastic saturated medium. In this case, the diffusion of the pore pressure can be considered in terms of the mechanics of poroelastic media.

The linear dynamics of poroelastic deformations is described by the Biot (1962) equations. In the general case, these equations predict the existence of two compressional and one shear waves in the system whereas the shear wave in the fluid is neglected. The first compressional and the shear waves are normal seismic P- and S-waves propagating in the medium. The second type of compressional wave is a diffusional wave for frequencies lower than the critical Biot frequency (for the media under consideration the critical frequency is usually of the order of MHz). It corresponds to the process of pore-pressure diffusion.

In the extremely low-frequency range, we obtain the following equation from the Biot system:

$$\frac{\partial p}{\partial t} = D\nabla^2 p. \tag{2}$$

This is the equation of the diffusion of the pore-pressure perturbation p in the rock mass. The hydraulic diffusivity can also be obtained from the Biot system of equations:

$$D = Nk/\eta, \tag{3}$$

where k is the permeability, η is the pore-fluid dynamic viscosity and N is a poroelastic modulus defined as follows: $N = MP_d/H; \alpha = 1 - K_d/K_g; M = (\phi/K_f + (\alpha - \phi)/K_g)^{-1}; H = P_d + \alpha^2 M; P_d = K_d + 4/3\mu_d$. Here $K_{f,d,g}$ are bulk moduli of the fluid, dry frame and grain material respectively; μ_d is the shear modulus of the frame and ϕ is the porosity. We ignore all non-mechanical (e.g., chemical or electro-chemical) interactions between the solid and the fluid.

We consider the following boundary condition: an initial pore-pressure perturbation is given as a function of time $p_0(t)$ (signature of the pore-pressure source) on a small spherical surface of radius a with its centre at the injection point. The injection point is the origin of the spherical coordinate system. The solution of equation (2) satisfying this boundary condition in the case of a time-harmonic perturbation $p_0(t) = p_0 \exp(-i\omega t)$ is:

$$p(r,t) = p_0 e^{-i\omega t}\frac{a}{r}\exp\left[(i-1)(r-a)\sqrt{\frac{\omega}{2D}}\right], \tag{4}$$

where ω is the angular frequency, and r is the distance from the injection point to the point where the solution is sought. From equation (4), we note that the solution is an exponentially attenuating spherical wave. This is the second compressional wave of the Biot theory with an attenuation coefficient equal to $\sqrt{\omega/2D}$ which is the reciprocal diffusion length, and a slowness equal to $1/\sqrt{\omega 2D}$ which is the reciprocal velocity of the relaxation.

Now, an estimation of the diffusivity D can be obtained using the following logic. A realistic injection signal is close to the step function: $p_0(t) = 0$, if $t < 0$, and $p_0(t) = 1$, if $t \geq 0$. However, triggering a seismic event at a time t_0 is due to the rectangular pulse $p_0(t) = 0$, if $t < 0, t > t_0$, and $p_0(t) = 1,$ if $0 \leq t \leq t_0,$ (because the evolution of the injection after the event triggering is not relevant to this event). The dominant frequencies of this signal are in the range of 0 to $\omega_0 = 2\pi/t_0$. Thus, if the event occurred at the distance r_0 then the relaxation times of the pore-pressure perturbation are of the order of $r_0\sqrt{t_0/(4\pi D)}$ and larger. However, we expect that the first triggerings can occur before a substantial relaxation (i.e., a relatively large change of the pore pressure) is reached. Therefore, for the earliest events

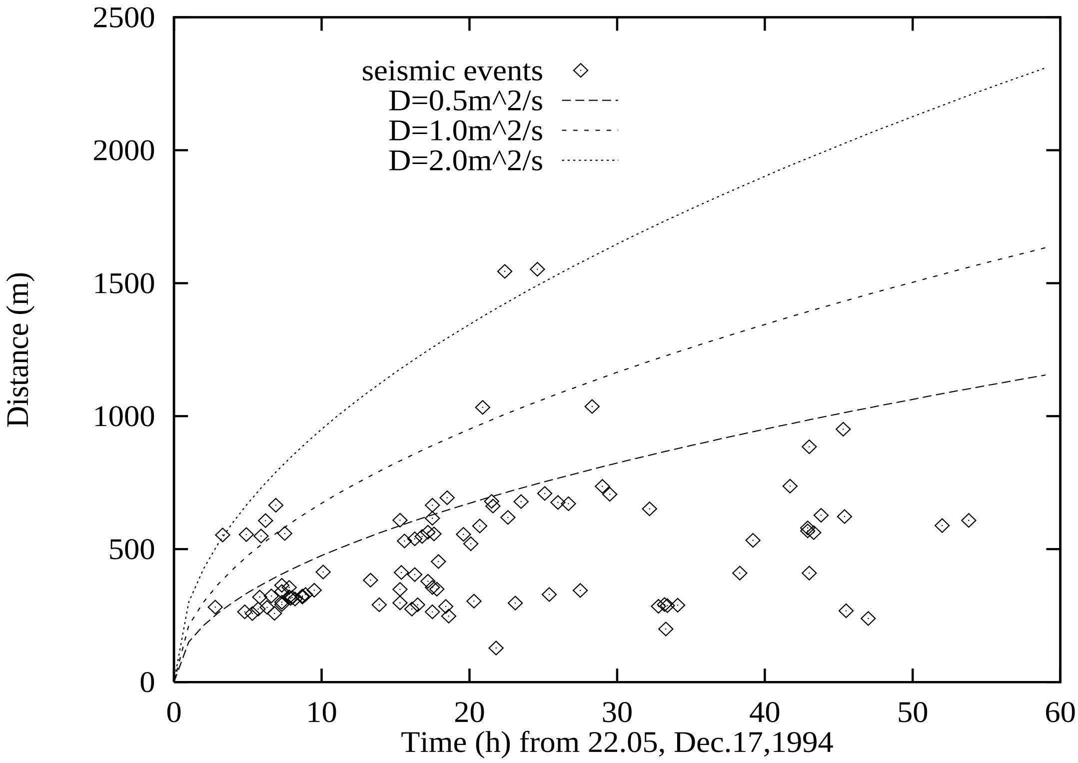
$$t_0 \leq r_0\sqrt{t_0/(4\pi D)} \tag{5}$$

From this inequality we obtain

$$D \leq \frac{r_0^2}{4\pi t_0}. \tag{6}$$

HYDRAULIC DIFFUSIVITY

Let us consider firstly the KTB experiment. During the KTB-hydraulic-fracturing experiment, approximately 400 micro seismic events were induced (Harjes, 1995; Zoback and Harjes, 1997; Buesselberg et al., 1995). There were located about 90 events with magnitudes larger than -1.5 where the largest event had a magnitude of 1.2. Figure 1 shows the spatio-temporal distribution (i.e., distance r versus time t) of all located events.



In the same Figure three curves satisfying the equation

$$r = \sqrt{4\pi Dt} \tag{7}$$

are plotted for three different values of the hydraulic diffusivity: $D = 0.5, 1$ and $2m^2/s$. For a given values of the diffusivity and time t equation (7) provides distances from the injection point to the outer boundary of the region, where a substantial pore-pressure relaxation has been reached. Therefore, the curves shown in Figure 1 are the triggering fronts for the given values of the diffusivity. In an arbitrary point triggering is possible after such a front has arrived, however, it is unlikely before. Thus, if the value of D has been correctly selected, the distance r for the majority of earthquakes must be smaller than the values given by equation (7) and the corresponding curve (7) will be an upper bound of the multitude of points in Figure 1.

We see that the estimation $D = 0.5 m^2/s$ is in good agreement with the majority of the events. It is also clear that there exist some zones which are probably fault zones with larger values of diffusivity, close to $D = 2 m^2/s$. They lead to the occurrence of some few earlier and more distant events (at a time of about 22h, approximately 1500m apart from the injection point, and with magnitudes in the range -0.8 to -0.5). Of course, an accidental generation of such events due to a remote triggering is also possible.

The next two figures show similar results for the Soultz experiment. The first Figure shows the multitude of the events (more than 9400 events have been localised) in the distance-time plot.

Our estimation of the hydraulic diffusivity in this case is $0.04 m^2/s$. This correspond to the theoretical curve shown in the Figure below:

ESTIMATING THE PERMEABILITY

In order to calculate the permeability we turn to equation (3) and to the definitions of the poroelastic moduli given below. In the case of low-porosity crystalline rocks terms of order α^2 can be neglected in comparison with terms of order 1 and α (for instance, in the situation considered here, $\alpha \approx 0.3$). In addition, terms of order ϕ can be neglected in comparison with terms of order α (in our case $\phi \approx 0.003$). Thus, the following approximation of the poroelastic modulus N is valid for crystalline rocks with low porosity:

$$N = \left[\frac{\phi}{K_f} + \frac{\alpha}{K_g} \right]^{-1} . \tag{8}$$

Note that generally the first term in the brackets of equation (8) cannot be neglected because usually $K_f \ll K_g$, especially in the case of partial gas saturation of the fluid.

To estimate N we must use any available data from laboratory and logs measurement.

For example, in the case of KTB experiment we used some measured and roughly estimated data as follows. For the grain material we assume the density, the P-wave and the S-wave velocities of an amphibolite-gneiss composite: $3000 kg/m^3, 6500 m/s,$ and $3800 m/s$, respectively. Additionally, we used the values of these quantities obtained in situ by the log measurements: $2900 kg/m^3, 5900 m/s,$ and $3500 m/s$, respectively. Further, we assume that the in-situ-measured bulk modulus of low-porosity crystalline rocks is a good approximation of K_d. For the fluid we assume properties of the water. Using for the porosity and the fluid bulk modulus the estimations $\phi = 0.003$ and $K_f = 2.3 \times 10^9 Pa$, we obtain $K_g = 7.0 \times 10^{10} Pa,\ K_d = 5.0 \times 10^{10} Pa$. Thus, we arrive at the following value: $N \approx 2 \times 10^{13} Pa$.

Assuming the value $\eta = 10^{-3} Pa \times s$ we obtain the following estimation: $k = 0.25 \times 10^{-16}$ to $1.0 \times 10^{-16} m^2$. These values are in excellent agreement with the upper limits of

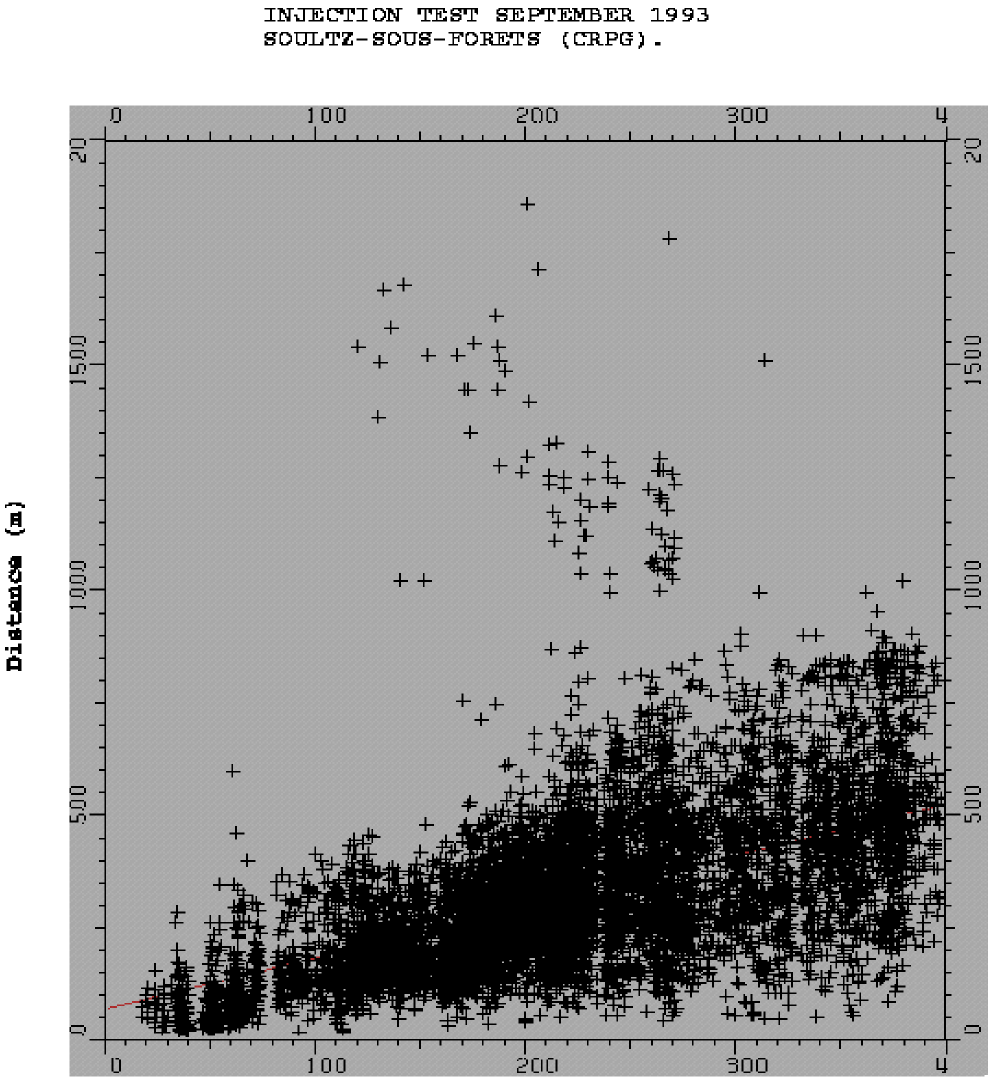


Figure 1. Distances of the events from the center of the injection interval versus their occurrence times.

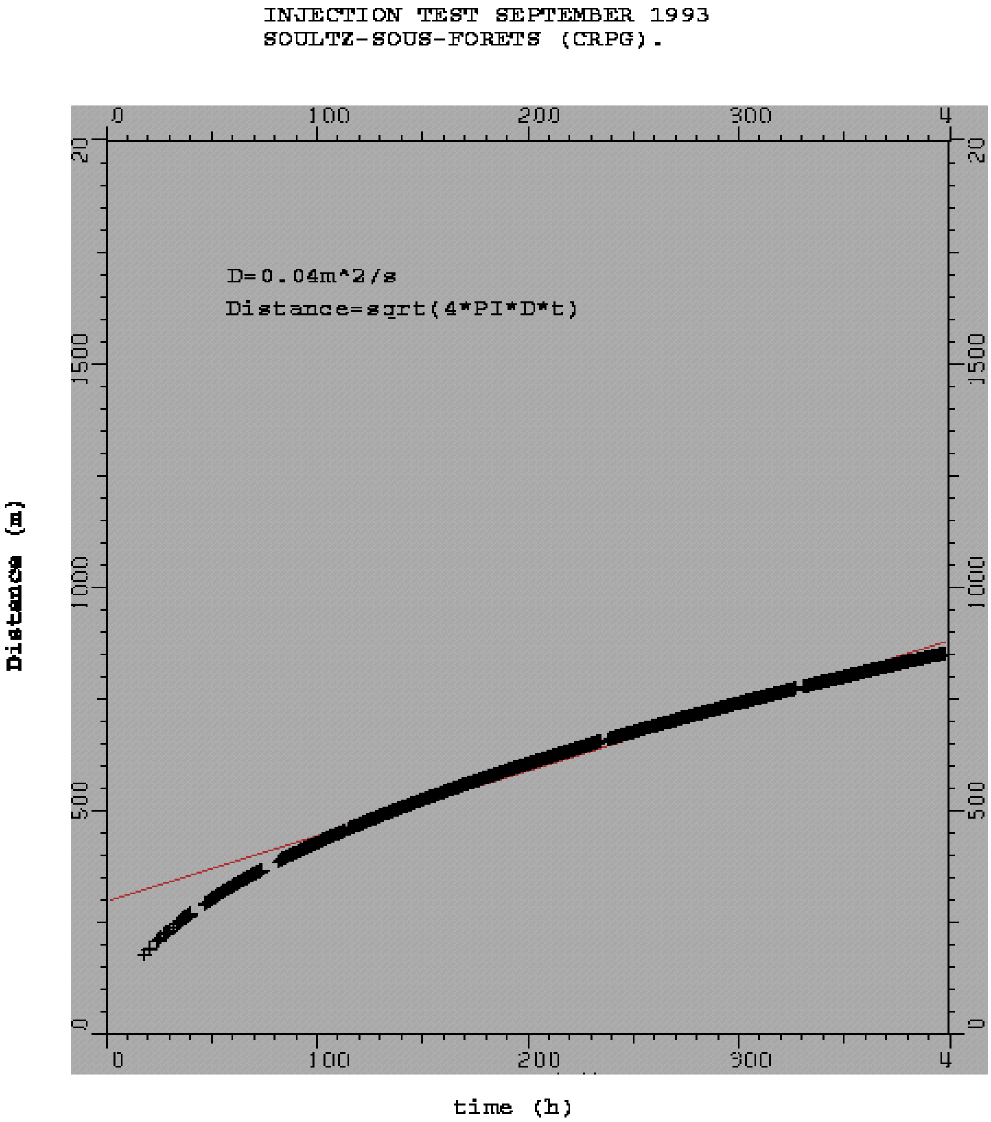


Figure 2. Model of distances with hydraulic diffusivity (D).

the former permeability estimations from the hydraulic experiments at the KTB (Huenges et al, 1997).

Interesting, that the estimate of the permeability for the Soultz case is also close to $1.0 \times 10^{-16} m^2$ and it also coincides with independent hydraulic measurements.

CONCLUSIONS

We have developed a technique for permeability estimation using the seismic emission induced by a borehole-fluid injection. The values we have obtained for KTB and Soultz experiments are in very good agreement with the previous permeability estimations from hydraulic observations. Moreover, our approach can provide directly the mean permeability tensor of reservoirs. Our approach indirectly supports the hypothesis that the state of stress in the crust is close to a critical one, i.e., the crust is in a failure equilibrium.

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PUBLICATIONS

A part of results presented here were published by (Shapiro et al., 1997).