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FAST-TRACK PAPER

Estimating the crust permeability from fluid-injection-induced seismic emission at the KTB site

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SUMMARY

During the hydraulic-fracturing experiment in the German Continental Deep Drilling Borehole (KTB) in December 1994, microseismic activity was induced. Here we develop a technique for estimating permeability using the spatio-temporal distribution of the fluid-injection-induced seismic emission. The values we have obtained for the KTB experiment (0.25×10^{-16} to 1.0×10^{-16} m^2) are in a very good agreement with the previous hydraulic-type permeability estimates from KTB deep-observatory studies. In addition, our estimates of the hydraulic diffusivity support the previously calculated value for the upper crust, which is of the order of 1 m^2 s^{-1}. However, this estimate now relates to the depth range 7.5–9 km.

Key words: crust, fault zones, fluid dynamics, permeability, poroelasticity, seismicity.

1 INTRODUCTION

In December 1994 a hydraulic-fracturing experiment was carried out in the German Continental Deep Drilling Borehole in the depth interval 9030–9100 m (Engeser 1996). About 200 m^3 of KBr/KCl brine was injected for approximately 40 hours. The fluid injection induced almost 400 microearthquakes in a spatial domain extending to 500–700 m from the borehole in a lateral direction in the depth range 7.5–9 km (Zoback & Harjes 1997). Here we use the spatio-temporal distribution of these seismic events in order to estimate the average permeability of the crystalline crust at the KTB site.

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 estimates can vary by orders of magnitude, even for adjacent locations. Moreover, permeability measurements are scale-dependent. Thus, large-spatial-scale measurements of the crust permeability (such as seismicity-based estimates) cannot be replaced by laboratory measurements. Some seismicity-based estimates of the crust permeability (or hydraulic diffusivity) are known from the literature (e.g. Ohtake 1974; Fletcher & Sykes 1977; Simpson, Leith & Scholz 1988). For instance, when analysing the evolution of reservoir-induced seismicity, Talwani & Acree (1985) found the hydraulic diffusivity, D, of the crust to range from 0.5 to 50 m^2 s^{-1}. Summarizing different seismicity-based studies, Scholz (1990) suggested a narrower range, 1–10 m^2 s^{-1}.

The physical background of such estimates is as follows. It is assumed that the state of stress in the crust is close to a critical one, that is the crust is in a failure equilibrium. Therefore, small perturbations of this state can lead to induced microseismicity. 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 microseismicity relative to the beginning of the injection is interpreted as the 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, the following rough estimate is usually used:

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

In the case of time-harmonic pore-pressure perturbations, an estimate of the diffusivity can be obtained directly from phase-shift information (analogous to tidal-tilt analysis; see e.g. Westerhaus 1996). In such a case, no information about Δp is required and a well-constrained estimate of D is possible (see also our discussion in Section 5). 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 physical concept described above to estimate the permeability at the KTB site from the injection-induced seismicity. We propose an approach to interpret the pore-pressure diffusion which leads to an improvement of eq. (1).

2 PORE-PRESSURE DIFFUSION

We approximate the real configuration of the fluid injection in KTB 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 deformation are described by the Biot (1962) equations. In the general case, these equations predict the existence of two compressional and one shear wave 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 several 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: $N = MP_\mathrm{d}/H$; $\alpha = 1 - K_\mathrm{d}/K_\mathrm{g}$; $M = (\phi/K_\mathrm{f} + (\alpha - \phi)/K_\mathrm{g})^{-1}$; $H = P_\mathrm{d} + \alpha^2 M$; $P_\mathrm{d} = K_\mathrm{d} + 4/3\mu_\mathrm{d}$. Here $K_{\mathrm{f,d,g}}$ are the 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 eq. (2) satisfying this boundary condition in the case of a time-harmonic perturbation $p_0(t) = p_0 \exp(-i\omega t)$ reads as follows:

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

where ω is the angular frequency and r is the distance from the injection point to the point where the solution is sought. From eq. (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 estimate 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, the triggering of 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 triggering of the event 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 a 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 (that is 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

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

3 HYDRAULIC DIFFUSIVITY AT THE KTB SITE

During the KTB hydraulic-fracturing experiment, approximately 400 microseismic events were induced (Harjes 1995; Zoback & Harjes 1997; Buesselberg, Harjes & Knapmeyer 1995). About 90 events with magnitudes larger than -1.5 were located, where the largest event had a magnitude of 1.2. Fig. 1 shows the spatio–temporal distribution (that is distance r versus time t) of all located events.

Three curves satisfying the equation

$$r = \sqrt{4\pi D t} \tag{7}$$

are plotted in Fig. 1 for three different values of the hydraulic diffusivity: $D = 0.5$, 1 and 2 $\mathrm{m^2\,s^{-1}}$. For given values of the diffusivity and time t, eq. (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 Fig. 1 are the triggering fronts for the given values of the diffusivity. An arbitrary point

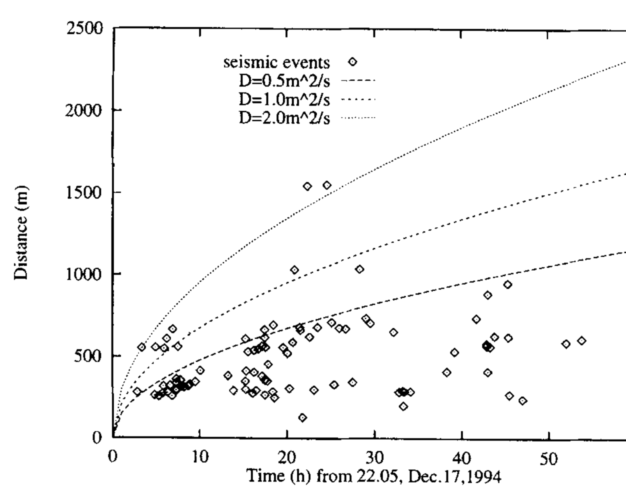

Figure 1. Distances of the events from the centre of the injection interval versus their occurrence times.

triggering is possible after such a front has arrived, but 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 eq. (7), and the corresponding curve (7) will be an upper bound of the multitude of points in Fig. 1.

We see that the estimate $D=0.5$ m^2 s^{-1} 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^{-1}. They lead to the occurrence of a few earlier and more distant events (at a time of about 22 hr, approximately 1500 m away 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.

Finally, our estimates of the hydraulic diffusivity are in excellent agreement with the values mentioned above given by Scholz (1990).

4 ESTIMATING THE PERMEABILITY

In order to calculate the permeability we turn to eq. (3) and to the definitions of the poroelastic moduli given below eq. (3). 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:

$$N=\left[\frac{\phi}{K_{\mathrm{f}}}+\frac{\alpha}{K_{\mathrm{g}}}\right]^{-1}. \tag{8}$$

Note that generally the first term in brackets in eq. (8) cannot be neglected because usually $K_{\mathrm{f}} \ll K_{\mathrm{g}}$, especially in the case of partial gas saturation of the fluid.

To estimate N we used the following data from the laboratory, and measurements from logs at the corresponding depth intervals in the KTB (Bram & Draxler 1995; Emmermann *et al.* 1995). For the grain material we assume the density and P- and S-wave velocities of an amphibolite gneiss composite: 3000 kg m^{-3}, 6500 m^2 s^{-1} and 3800 m^2 s^{-1}, respectively. Additionally, we used the values of these quantities obtained *in situ* by the log measurements: 2900 kg m^{-3}, 5900 m s^{-1}, and 3500 m s^{-1}, 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 estimates $\phi=0.003$ and $K_{\mathrm{f}}=2.3\times10^9$ Pa, we obtain $K_{\mathrm{g}}=7.0\times10^{10}$ Pa, $K_{\mathrm{d}}=5.0\times10^{10}$ Pa. Thus, we arrive at the following value: $N \approx 2\times10^{13}$ Pa.

Assuming the value $\eta=10^{-3}$ Pa s we obtain the following estimate: $k=0.25\times10^{-16}$ to 1.0×10^{-16} m^2. These values are in excellent agreement with the upper limits of the former permeability estimates from the hydraulic experiments at the KTB (Huenges *et al.* 1997).

5 DISCUSSION

The seismic-based estimate of the crust permeability is subject to several uncertainties. We consider our estimations mainly as an order-of-magnitude calculation. Furthermore, our estimations correspond rather to the upper limits of the average-permeability range for two reasons. First, this follows from the physical principles of the method developed here (see eq. 6). Second, during hydraulic-fracturing experiments the increased pore pressure enhances the permeability in the region adjacent to the borehole (Huenges *et al.* 1997).

It is interesting to note that the large-scale hydraulic effects are possibly related to more significant pore-pressure variations than those necessary for triggering seismic emission. Thus, estimates based on hydraulic experiments probably provide lower limits on the permeability. For example, the test on the hydraulic communication between the KTB main borehole and the pilot borehole at a depth of about 4000 m provided an estimate $D=0.12$ m^2 s^{-1} (Kessels & Kück 1995). Similar values can be obtained from fluid-level observations in the pilot hole after the injection in the main hole (Huenges *et al.* 1997). The corresponding values of permeability are of the order of 10^{-17} m^2. However, it is important to note that the permeability estimates are much more uncertain than the estimates of hydraulic diffusivity.

The issue of what critical value of the pore-pressure perturbation Δp is required to trigger microseismicity is interesting and important. We think that Δp is a strongly fluctuating quasi-randomly distributed characteristic of the medium. One can try to find limits for the minimum Δp considering reservoir-induced seismicity. For instance, in the recent paper of Ferreira *et al.* (1995) an example of a reservoir was shown, where the seasonal microseismicity has been induced by water-level fluctuations of magnitude less than 5 m. This corresponds to a perturbation of the pore pressure that is less than or of the order of 5×10^4 Pa. On the other hand, using eq. (4) along with data on the injected fluid volume, porosity and the overpressure developed (approximately 50 MPa) we arrive at an estimate of 3×10^3 Pa for the most distant events from Fig. 1. However, because the triggering of earthquakes is a non-linear process, the minimum Δp may depend on many factors (such as frequency, stress, etc.; see e.g. Gomberg, Blanpied & Beeler 1997), and one can expect even smaller its values.

In the paper by Ferreira *et al.* (1995) mentioned above, a 3 month time-lag ($\Delta t=0.25$ yr) between seasonal water-level change ($\omega=2\pi$ yr^{-1} is the characteristic frequency) and a change in the level of seismicity at a depth 3 km $\le L \le$ 4 km was observed. Using the formula for the relaxation velocity of a time-harmonic perturbation (given below eq. 4) we obtain the following estimate: $D=L^2/2\omega\Delta t^2 \approx 0.4-0.6$ m^2 s^{-1}. Again, this estimate is very close to our results for D. Ferreira *et al.* (1995) used approximation (1), which usually tends to overestimate D. They obtained values of 1.2–2.1 m^2 s^{-1}.

6 CONCLUSIONS

We have developed a technique for permeability estimation using the seismic emission induced by borehole fluid injection. The values we have obtained for KTB are in very good agreement with the previous permeability estimates from the KTB deep observatory. In addition, our estimates of the hydraulic diffusivity support the previously calculated value for the upper crust, which is of the order of 1 m^2 s^{-1}. Furthermore, this new estimate relates to the greater depth of 7.5–9 km. Our approach indirectly supports the hypothesis that the state of stress in the crust is close to a critical one, that is the crust is in a failure equilibrium.

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