Influence of pore-pressure on the event-size distribution of induced earthquakes

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During an Enhanced Geothermal System (EGS) experiment, fluid is injected at high pressure into crystalline rock, to enhance its permeability and thus create a reservoir from which geothermal heat can be extracted. The fracturing of the basement caused by these high pore-pressures is associated with microseismicity. However, the relationship between the magnitudes of these induced seismic events and the applied fluid injection rates, and thus pore-pressure, is unknown. Here we show how pore-pressure can be linked to the seismic frequency–magnitude distribution, described by its slope, the $b$-value. We evaluate the dataset of an EGS in Basel, Switzerland and compare the observed event-size distribution with the outcome of a minimalistic model of pore-pressure evolution that relates event-sizes to the differential stress $\sigma_D$. We observe that the decrease of $b$-values with increasing distance of the injection point is likely caused by a decrease in pore-pressure. This leads to an increase of the probability of a large magnitude event with distance and time.


1. Introduction

Enhanced Geothermal Systems (EGS) represent a promising alternative for clean energy. For such systems, two boreholes are drilled to depths below 3 km into the basement, between which a fluid, typically water, is circulated to extract the heat at temperatures well above 100°C. The fluid is pumped at high pressure into the first borehole, to increase the permeability of the rock in a process called reservoir stimulation. This process can be accompanied by micro-earthquakes. Their hypocenters reveal critical information about the ongoing evolution of the reservoir. Few events eventually happen to be large enough to be felt at the surface, creating nuisance for the population or even damage to the building stock. The seismic hazard and risk associated with the induced seismicity is in some cases, such as for the EGS project beneath the city of Basel (Switzerland), estimated as too high to be acceptable for society [Baish et al., 2009]. Induced seismicity is currently the largest obstacle to the widespread installation of EGS systems near urban centers. Enhancing our understanding of the physical processes that control the generation of induced seismicity is therefore an urgent scientific challenge.

The basic principles of pore-pressure induced seismicity are well understood: the increase in pore-pressure decreases the normal stress on the rock volume, resulting in a lower effective stress [Pearson, 1981]. The failure releases pre-existing tectonic stresses with the consequence of nucleating sudden slip – in other words, micro-earthquakes. However, the current statistical, physical and numerical models of induced seismicity cannot sufficiently explain the critical parameters that ultimately control seismic hazard and risk, i.e., the scaling of earthquake sizes, the temporal occurrence of events, the amount of induced seismicity and the maximum possible magnitude. To investigate the physical processes controlling these parameters, we analyse the seismicity related to the EGS experiment conducted in Basel in December 2006.

The seismicity induced during the experiment in Basel was closely monitored by a six-sensor borehole array (see Figure 1a), which recorded over 11,000 events of which over 3,500 could be located [Häring et al., 2008]. The located events range from moment magnitudes $M_w$ 0.1 to 3.2, with three events above $M_w$ 3. Today, seismicity at the site is still slightly above the assumed long-term background, its decay being indistinguishable from a typical aftershock sequence [Bachmann et al., 2011].

2. Earthquake-Size Distribution

The cumulative number of earthquakes, $N$, in a given volume generally follows a power law distribution and can be expressed as $\log N = a - bM$ [Gutenberg and Richter, 1942], where $a$ and $b$ are constants that describe the productivity and the relative size distribution, respectively. Higher $b$-values indicate more small events relative to larger events and vice versa. Equation (1) and slight modifications thereof are used in essentially all seismic hazard studies [e.g., Giardini et al., 2004] as it allows extrapolation from the observed smaller events to the infrequent larger ones. Studies of micro-earthquakes on faults [e.g., Schorlemmer and Wiemer, 2005] have shown that the $b$-value, when mapped with high quality data at high resolution, varies in the Earth’s crust over distances of a few kilometers or less. These studies, combined with the analysis of regional and global focal mechanism data [Gulia and Wiemer, 2010; Schorlemmer et al., 2005] as well as laboratory work [Amirano, 2003] indicate that the $b$-value is inversely proportional to the differential stress $\sigma_D$ and thus may qualitatively be used as a stress meter at depth in the Earth’s crust, where generally no direct measurements are possible. In subduction zones [van Stiphout et al., 2009] and volcanic systems [Wiemer and McNutt, 1997], it has been argued that high $b$-values are also related to the presence of fluids. In this study we
fluids away from the injection point can be illustrated as the evolution of the observed seismicity as a function of time and distance (Figure 2a). We find that the $b$-value substantially decreases from the co-injection period ($b_{co} = 1.57 \pm 0.06$) to the post-injection period ($b_{post} = 1.14 \pm 0.06$) [Bachmann et al., 2011] (inset to Figure 2a). This indicates an increase of the seismic hazard for the post-injection phase relative to the co-injection phase. Further systematic behavior of the seismicity is documented by analysis in one dimension, the distance to the casing shoe, separately for the co- and post-injection period (Figure 2c). While a relation between distance and $b$-value is not obvious for the shortest distances during the injection, we find a systematic decay of $b$-values in both periods beyond a distance of 200 meters that also holds when considering the uncertainty in the data.

[7] We introduce a new focus-centered mapping technique to determine the three-dimensional distribution of the $b$-values. The local magnitude of completeness $M_c$ is determined using the closest 150 events in space and is calculated with the maximum curvature method [Woessner and Wiemer, 2005]. $M_c$ defines the catalogue threshold for determination of the $b$-value, and ranges from $M_c = 0.7$ - closest to the deepest borehole sensor, to $M_c = 1$ furthest away from the sensor (Figure 1a). To determine the $b$-value with the maximum likelihood method [Utsu, 1999], we require at least 25 events with $M \geq M_c$ in each sample. The resulting $b$-values range from 0.8 to 3.5 (Figure 1b) with a mean standard error of $\sigma(b) = 0.2$. We observe a very systematic behavior: highest $b$-values are observed close to the casing shoe, forming a toroidal (doughnut-shaped) area, and lowest values are located at the outermost edges of the seismicity cloud. However, the variability is high within the event cloud. We also observe lower values close to the borehole and larger values further out, but the general behavior is constant. The temporal component of this variability is shown by analyzing the co- and post-injection period separately (Figures 3a and 3b, respectively); high $b$-values are observed near the injection point and occur during the injection, while areas with lower $b$-values are found further away from the borehole and develop after shut-in. Large magnitude events (LME; defined in our context as events felt at the surface, i.e., events with $M_c \geq 2.5$) are located in regions with below average $b$-values ($b \leq 1.3$). We evaluate this by calculating the causal $b$-value for each event, where we determine $b$-values based on the closest 150 events in space, preceding the event. This method implies that we miss the earliest LME, if they are within the first 150 events. By comparing the distribution against a random correlation of $b$-value and magnitude, we find that it is significant at the 99.95% level to observe five LME in regions with below average $b$-value.

[8] Based on all these observations, the first conclusion we draw is that $b$-values are spatially highly variable in the induced seismicity sequence of the EGS in Basel, decreasing systematically with time and distance from the casing shoe. The current state of knowledge on $b$-values of induced sequences is that they are generally higher than normal tectonic events [Wyss, 1972; Shapiro et al., 2011]. Our analysis unravels a much more complex yet systematic pattern; values range from typical tectonic $b$-values at the edges of the seismicity cloud to extremely high values near the injection point. Our detailed mapping of the spatial distribution of the $b$-value can be used to estimate the probability of occurrence

Figure 1. Overview. (a) Overview of the experiment showing the depth (a.s.l.) of the 3560 located events (circles) and the location of the seismic stations (triangles). The red plane marks the top of the crystalline basement, within which all events occurred. The colorscale indicates the recording completeness ranging from $M_c = 0.7$ to 1. (b) Close-up of the events with the overall $b$-value distribution based on all events. While values range from 0.8 to 3.5, the colorbar is limited from 1 to 2 for a clearer visibility.
3. Geomechanical Model

To develop a geomechanical understanding of the observed systematic behavior of the $b$-values with space and time, we simulate the induced seismicity cloud in space and time under the hypothesis that high $b$-values near the injection point are a response to the pore-pressure perturbation. A similar model was previously introduced by Rothert and Shapiro [2003], however here we include the earthquake sizes in the model to be able to analyze $b$-value distributions.

We randomly distribute potential failure points (we call them seed faults), representing pre-stressed faults, in a three-dimensional space centered around the injection point. Each seed fault is assigned a minimum and maximum principal stress $\sigma_3$ and $\sigma_1$, based on a background stress regime according to Häring et al. [2008] and a Gaussian perturbation of 10% (K. F. Evans, personal communication, 2011). Using a constant cohesion (7 MPa) and coefficient of friction (0.85 for intact rock), we can analytically calculate the Mohr-Coulomb diagrams and thus a failure criterion for each seed fault. In case of an unstable seed fault, we randomly reassign the principal stresses until we acquire the desired number of stable seed faults. Here we use 300,000 seed faults in a volume of 1 km$^3$. We introduce a time-dependent point pressure source at the injection point. The pressure-time function is linearly increasing over 6 days, a simplified version of the actual injection pressure time series at Basel [Dinske et al., 2010]. We propagate the pore-pressure through the model space based on linear diffusion in a hydraulically isotropic medium with an effective diffusivity of 0.05 m$^2$/s. Dinske et al. [2010] introduced an analytical solution to the diffusion equation [Wang, 2000] for the case of a linearly increasing source time function. We use their analytical solution assuming an effective source radius of 70 m, in order to reach realistic values of the pore-pressure perturbation. The naturally occurring pore-pressure variation in the reservoir due to tides is assumed to be on the order of 2000 Pa, which has been measured at 1 km depth in the middle of the Furka tunnel, Switzerland (Evans, personal communication, 2011). Therefore we assume that the
triggering front of the induced seismicity should closely follow this isobar.

Figure 4a depicts the model schematically. The seed faults are represented by the points closest to failure and are color-coded according to their respective differential stress, defined as $s_D = \sigma_1 - \sigma_3$. The pore-pressure evolution with time reduces the normal stresses shifting the Mohr circles towards the failure envelope. An event is induced once the Mohr circle touches the failure envelope. For simplicity, we here assume that all seed faults are optimally oriented and each seed fault can only be triggered once. To assign a magnitude to an event, we exploit the observation from laboratory and natural earthquake analysis that $b$-values are inversely related to $s_D$ [Amitrano, 2003; Schorlemmer et al., 2005]. As scaling law, we use a simple linear scaling between the observed ranges of $b$-values (0.8 to 3.5) and $s_D$ values of our model (20 to 150 MPa). The magnitudes are then randomly drawn from a power-law distribution with the assigned $b$-value for each failure occurrence. Using a linear diffusion model with homogeneous and isotropic hydraulic diffusivity is a strong simplification for two reasons: (i) the opening of fractures potentially modifies the diffusivity, creating non-linearity [Hummel and Müller, 2009] and (ii) actual pore-pressure propagation can be highly anisotropic, occurring preferentially along zones of weakness [Evans et al., 2005]. However, the model serves as a first-order approximation and can explain some features of the observed seismicity well [Goertz-Allmann et al., 2011].

We run a total of 100 simulations, whereas results of only one simulation are usually presented. The evolution of the seismicity as a function of time and distance of the simulated seismicity (Figure 2b) resembles the one we find from the data (Figure 2a). The same holds for the evolution of the $b$-value with distance; $b$-values of the simulation decay with distance (Figure 2d), however we do not reproduce the complexity of the $b$-values for the nearest 200 m to the

Figure 3. The $b$-value distribution. Cross section along the N-S axis, with $b$-value distribution based on (a) co-injection and (b) post-injection events, stars in Figure 3b indicate events with $M_w \geq 2.5$. (c) Examples of frequency–magnitude distributions for two sampled volumes with a high $b$-value of 2.93 ± 0.40 (red) and a low $b$-value of 0.83 ± 0.1 (blue). The potential for larger events such as $M_w$ 3.5, estimated solely by extrapolating the FMDs, is up to 10,000 times higher in the blue volume. (d–f) Equivalent plots to Figures 3a–3c with the simulated events. The sample volumes in Figure 3f show $b$-values of 2.31 ± 0.30 (red) and 1.00 ± 0.10 (blue). The colorbar of the $b$-values is limited from 1 to 2 in all subplots. All distances are relative to the injection point at [7.59; 47.59; 4.37] for the observed events and at [0; 0; 0] for the simulated events.
injection point. Likewise, the spatial distribution of the \( b \)-values (Figures 3d and 3e) resembles the one we find from the data. While we reproduce the general patterns like the highest values close to the injection point and the lower values during the post-injection period, we do not replicate the finer details.

13 To understand the geomechanical reasons for the good match between model and observation, we show the state-of-stress of all seed faults that were triggered in one simulation of the model (Figure 4a). Out of the 300,000 seed faults, 791 ruptured. We divide these into events triggered with pore-pressure changes \( \Delta P \) above (red square) and below (green square) the mean value of 2.74 MPa and determine the corresponding FMDs of the two subsets (inset of Figure 4b). The corresponding \( b \)-values are \( b_{\text{high}P} = 1.45 \pm 0.105 \) (red) and \( b_{\text{low}P} = 1.11 \pm 0.05 \) (green) (indicated by the FMD inset).

4. Discussion

14 Our results suggest the following conceptual model for the scaling of induced seismicity. Near the borehole (0–200 m), the pore-pressure rapidly rises once the stimulation starts, and numerous events are induced. These failures sample a wide range of differential stresses, including very small \( \sigma_D \) that were originally far away from the failure criterion but still ruptured due to the high pressure perturbation. It is these small \( \sigma_D \) values that cause a systematic bias towards higher \( b \)-values, such that events close to the injection point exhibit a higher \( b \)-value, if we assume that \( b \)-values are inversely related to \( \sigma_D \). As the pore-pressure front progresses, seed faults experience only a moderate increase in pore-pressure. The differential stresses of these faults follow more closely a typical tectonic earthquake–size distribution. Once the pumping is stopped, pore-pressures near the well decrease rapidly, and this area is then essentially devoid of seismicity [Parotidis et al., 2004]. At distances exceeding 200 m, pore-pressures continue to increase gradually, although absolute values remain low. We suggest that static and dynamic stress changes caused by the numerous earthquakes themselves will contribute to inducing (or in some cases

Figure 4. Model. (a) Mohr-Coulomb plot for the 3D model. Circles denote the points closest to failure for all 300,000 seed faults. Increasing the pore pressure (\( \Delta P \)) shifts the potential seed to the left; an event is triggered if a Mohr-Coulomb circle touches the failure envelope. This is indicated with the half circle (dashed black line) representing the mean stress state. The arrow indicates the minimum pore pressure perturbation needed to trigger this seed fault. The ranges of the minimum and maximum principal stress \( \sigma_3 \) and \( \sigma_1 \) are indicated with a solid and dashed green line, respectively. (b) Result of one simulation of the model. A total of 791 seed faults were triggered. The \( b \)-values for events triggered with high \( \Delta P \) (red rectangle) and low \( \Delta P \) (green rectangle) are significantly different, \( b_{\text{high}P} = 1.45 \pm 0.105 \) (red) and \( b_{\text{low}P} = 1.11 \pm 0.05 \) (green) (indicated by the FMD inset).

Figure 5. Probability for large magnitude events. Occurrence probability of M 4 event, varying with (a) time and (b) radial distance from the injection point for a varying \( b \)-value (white) and a constant \( b \)-value (gray). The errorbar denotes the standard deviation based on 100 simulations of the geomechanical model. The dashed line in Figure 5a marks the shut-in time and in Figure 5b the location of the largest observed Basel event.
inhibiting) events in these regions. These stress changes also contribute in the vicinity of the injection, but there they are negligible when compared to the pore-pressure. The observed exponential decay of seismicity [Bachmann et al., 2011] with typical aftershock parameters is then a combination of the continued gradual expansion of the pore-pressure front and the gradual decay of aftershock activity caused by static and dynamic stress changes.

[15] Based on the simple geomechanical simulation, we can also address the question of how large future events will be and how likely they will be. Therefore, we evaluate the synthetic seismicity cloud in time and space for a and b-values within specific time- and distance bins. For each time- or distance bin, we can estimate the probability p of an event exceeding a certain magnitude M as $p = 1 - e^{-x}$, where $x = 1/(a - bM)$ [Wiemer, 2000]. The probability of an event exceeding M 4 is shown in Figure 5. We choose time bins of $10^5$ s, moving at $10^4$ s intervals, and distance bins of 100 m, moving across distance in 10 m increments. We compare the probability based on our synthetic event cloud with varying b-values (white) with the probability based on an event cloud synthesized from a constant b-value (1.2, gray). We find that the probability for LME increases with time during the injection and then decays after the shut-in. A varying b-value causes a substantial increase of the mean probability for the time period right after shut-in relative to before shut-in (Figure 5a), and shifts the maximum of the probability to further distances from the injection point (Figure 5b). These calculations are consistent with observations of LME during hydraulic stimulations in geothermal systems not only in Basel [Deichmann and Giardini, 2009], but also elsewhere [Baisch et al., 2010]. Our observations and the stochastic model approach presented can form the basis for and improved probabilistic real time forecasting system of induced seismicity in future EGS experiments.

[16] The observed, highly systematic distance and time evolution of the b-values in Basel calls in our opinion for a fundamental mechanism as an explanation. The link between differential stress and b-value is a fundamental mechanism, established across a large range of scales, from the analysis of individual rock samples in the laboratory to global seismicity [Amitrano, 2003; Schorlemmer et al., 2005]. Adding this link as the only independent ingredient to our minimalist model of induced earthquakes is sufficient to explain the observed b-values evolution in space and time. Furthermore, the ability to explain the size distribution of induced events is strong additional evidence that differential stress is one of the governing parameters that influences b-values for a wide range of tectonic settings.

[17] We propose that our conceptual model, developed for the well-recorded induced seismicity in the natural laboratory of the Basel EGS, is universally able to explain a wide range of observations of the earthquake-size distribution. We speculate that the same mechanism of pore-pressure changes is responsible for other high b-values anomalies in the Earth’s crust: in aftershock zones near the areas of highest slip [Wiemer and Katsumata, 1999]; in volcanic regions near magma chambers [Wiemer and McNutt, 1997] and in subduction zones [van Stiphout et al., 2009]. It is worthwhile noting that in the immediate vicinity of the rupture plane of larger earthquakes, static and dynamic stress changes are expected to be large enough to trigger events with a wider than usual range of differential stresses, thus leading to higher b-values with in the aftershock sequence close to the rupture fault plane.

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