THE EFFECT OF FAULT ZONE DEVELOPMENT ON INDUCED SEISMICITY

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ABSTRACT

It is widely acknowledged that induced slip on preexisting fractures is responsible for both the induced seismicity and the permeability enhancement in Enhanced Geothermal Systems (EGS). At a given location, the character of the preexisting fractures can be classified along a spectrum from joints or crack-like faults to fully developed, thick faults with cataclasite fault cores. We reviewed data collected at nine projects around the world and found a striking variability in the character of the fractures and faults at each site. There was an excellent correlation between the degree of brittle fault formation and the maximum magnitude of seismic events induced by stimulation. For long term circulation in which injection rate exceeded production rate, the correlation was weaker, but still present. We hypothesize that larger events tended to occur in places with thicker faults due to continuity of slip surfaces. We discuss the interactions between frictional properties, lithology, depth, geological heterogeneity, and seismic hazard. For cases where only very small seismic events were induced, we offer alternative hypotheses to explain the mechanism of their generation. We conclude by discussing our findings in the context of induced seismic hazard analysis in general. Our results suggest that characterization of fault development should receive more emphasis, both in seismic hazard estimation and in reservoir engineering and modeling.

1. INTRODUCTION

1.1 Background

Fluid injection and production induced seismicity has been documented at many sites around the world (Nicholson and Wesson, 1990; McGarr et al., 2002). Induced seismicity occurs when stress is changed on preexisting fractures so that their shear traction exceeds their frictional resistance to slip according to the Coulomb failure criterion (Jaeger et al., 2007):

\[ \tau = \mu (\sigma_n - P) + S_0, \]

where \( \tau \) is shear traction on a fracture, \( \sigma_n \) is normal stress, \( P \) is fluid pressure, \( S_0 \) is the fracture cohesion, and \( \mu \) is the coefficient of friction. When the failure criterion is exceeded on a fracture, slip occurs to relieve shear stress. Alternative mechanisms for very small seismic events are discussed later in Section 4.4.

Slip is seismic when it occurs rapidly enough to generate seismic waves. Seismic slip occurs because of rapid weakening and restrengthening of the friction coefficient. Rate and state friction theory describes how the coefficient of friction can weaken rapidly as a result of tiny asperities on a fracture surface coming out of contact during slip (Dieterich, 1979, 1992, 2007; Dieterich and Kilgore, 1994). The tendency to slip seismically depends on several factors, including the material properties of the materials contacting in the fracture plane, the temperature, and the stress state (Dieterich et al., 1992; Blanpied et al., 1985; Tembe et al., 2010).

From Equation 1, it can be seen that seismicity can be caused by changes in shear traction, normal stress, or fluid pressure. Fluid injection and production can induce stress through several mechanisms. These mechanisms include (1) increase in pore pressure from fluid injection and (2) poroelastic or thermoelastic expansion or contraction of the reservoir due to changes in pressure or temperature (Majer et al., 2007). A third mechanism (3) is that slip induced by mechanisms (1) and (2) can cause stress change. Mechanisms (2) and (3) can act at distances greater than the region of pressure or thermal perturbation because stresses can be transferred through the earth at distances beyond the sources of strain that caused them. Stresses induced by these different processes are heterogeneous and anisotropic, and whether they encourage or inhibit seismicity on a given fracture depends on the fracture's orientation and relative location (Segall, 1989). Stress triggering and stress shadow effects
have been well documented in studies of seismic hazard (Harris, 2000).

Enhanced Geothermal Systems (EGS) are projects where large volumes of fluid are injected into low permeability, typically crystalline rock in order to improve the permeability of the formation for geothermal energy production. Typically, injection is carried out at a fluid pressure less than the least principal stress and the primary mechanism of permeability enhancement is induced slip on preexisting fractures (Tester, 2007).

Induced slip in EGS stimulation usually causes extensive microseismicity. In some cases, projects have been associated with seismic events strong enough to be felt at the surface and which may have caused very minor property damage (Majer et al., 2007; Cladouhos et al., 2010; Majer et al., 2011; Evans et al., 2012). For this reason, estimation of induced seismicity hazard has become an important issue for EGS. Experience with EGS could also help understand induced seismicity as it applies to other settings, such as oil and gas hydraulic fracturing (Holland, 2011; Baisch and Vörös, 2011) or CO₂ sequestration (Cappa and Rutqvist, 2011; Nicol et al., 2011).

Several models have been proposed for use in estimation of induced seismicity hazard. In general, a major shortcoming is that they require injection to calibrate their use. It would be desirable to predict induced seismicity hazard prior to initiating injection or prior to drilling a well.

McGarr (1976) proposed that the seismic moment release during injection should be proportional to the volume of fluid injected, a prediction generally supported by subsequent experience comparing seismicity at different locations (McGarr et al., 2002; Nicol et al., 2011; Evans et al., 2012) and during injection at a single location (Bommer et al., 2006; Baisch and Vörös, 2009, Section 4.2). This effect is also has been observed in some numerical modeling (Baisch et al., 2010; McClure and Horne, 2011a). However, the scatter in the data is very large. The seismic moment release induced at different EGS projects around the world has varied hugely, even among projects involving similar volumes of water (Evans et al., 2012).

Shapiro et al. (2007) and Shapiro and Dinske (2009) combined the assumption that seismic moment release is proportional to fluid injection with an assumed Gutenberg-Richter (GR) magnitude-frequency distribution. But they did not suggest how the parameters of the GR distribution could be determined prior to injection. The GR distribution might be characterized from observations of background seismicity, but it is not necessarily the case that natural seismicity in a region should have the same GR parameters as induced seismicity. Furthermore, in some settings, there may not be enough natural seismicity to characterize the natural GR distribution.

In order to monitor seismicity in real time during injection, Bommer et al. (2006) proposed the traffic light system. They proposed fitting observed seismicity to a GR distribution in real time during injection. Injection would be stopped if the empirical GR distribution predicted that an event larger than a given threshold might occur during injection (Bommer et al., 2006). This approach was further refined in Bachmann et al. (2011), who investigated the use of different statistical models of seismicity. A potential problem with these approaches is that post-injection seismicity tends to follow a different GR distribution than seismicity during injection (Baisch and Vörös, 2009, Section 4.1) with a tendency towards larger events (Häring et al., 2008; Baisch et al., 2010). Numerical modeling has suggested that this behavior is likely due to the continued spreading of fluid after injection stops (Baisch et al., 2006; Baisch et al., 2010; McClure and Horne, 2011a). Bachmann et al. (2011) examined seismicity at the EGS project in Basel, Switzerland and proposed that the post-injection seismicity could be matched with an Omori-Utsu aftershock distribution. Regardless, these methods cannot be used to predict hazard prior to initiating injection.

Triggered seismicity occurs when induced stresses cause a natural earthquake to happen sooner than it would have otherwise. There is an extensive literature on natural triggered seismicity (Freed, 2005; King, 2007; Hill and Prejean, 2007). In EGS seismic hazard analysis, triggered seismicity has been handled with numerical modeling and consideration of the natural rate of seismicity. The stresses induced by geothermal exploitation are calculated on nearby mapped faults, and those stresses are used to estimate the degree to which the occurrence of natural earthquakes may have been accelerated on those faults (Hunt and Morelli, 2006; Vörös and Baisch, 2009).

An effective method to predict seismicity prior to injection could be computational modeling. A typical strategy in EGS shear stimulation modeling is discrete fracture modeling. The standard workflow is to stochastically or deterministically generate a realization of the preexisting fracture network and then numerically simulate the injection (Willis-Richards et al., 1996; Rahman et al., 2002; Ghassemi et al., 2007; Kohl and Mégel, 2007; Bruel, 2007; Baisch et al., 2010; McClure and Horne, 2011a Rachez and Gentier, 2010; Deng et al., 2011). Discrete fracture modeling has the potential to predict seismic hazard directly, by simulating individual
seismic events on faults. However, to simulate seismic events in a predictive way, a model would need to compute both the stresses induced by slip and the evolution of friction during an earthquake, which only a few EGS models have attempted (Baisch et al., 2010; McClure and Horne, 2011a). Some modeling studies have attempted to model EGS seismicity without including evolution of friction (Brue, 2007; Kohl and Mégel, 2007) or as the breaking of bonds between rock particles (Hazzard et al., 2002). Continuum based modeling has been used for modeling of seismicity caused by thermal contraction (Rutqvist and Oldenburg, 2008). Continuum modeling is appropriate for this application because strain cause by thermal contraction is volumetric, and so the stresses induced do not depend on the details of flow or deformation on individual fractures.

One value of modeling is that by stepping through the work flow necessary to parameterize a model, it can become apparent which uncertainties give rise to the significant model behaviors. This study was inspired by the need for better characterization of fracture system geometry and fault frictional behavior in discrete simulation modeling of EGS. Model inputs regarding fractures that need characterization include frictional properties, storativity, shape and spatial extent, patterns of clustering and connectivity, and permeability anisotropy (within individual fractures, as opposed to bulk permeability).

There are countless situations where fault geometry would have a profound effect on model behavior. For example, Ghassemi et al. (2007) investigated the effect of thermoelastic strains in EGS by modeling injection into a single, planar crack. However, in some EGS projects, flow occurs in fault zones with thicknesses of ten meters or more (Genter et al., 2000). The temperature field and thermal stresses that would develop due to injection into a crack would be very different than injection into a thick fault zone. In another example, McClure and Horne (2011b) showed how the pressure transient caused by injection would be affected strongly by the inner geometry of storativity and permeability within a fault.

1.2 Effect of fault development
A geological unit can be categorized according to its degree of fault development (Fetterman and Davatzes, 2011). Faults can form with a variety of modes of deformation from highly localized and brittle to distributed and ductile (Wibberley et al., 2008). Because in this study we are interested in the development of brittle fault zones, we use a categorization of faulting based on the degree of cataclasite formation, from joints or crack-like faults to well developed faults with thick cataclasite core and meter scale damage zones.

These progressive stages reflect the understanding that faults form by the progressive shear and link up smaller faults into larger thoroughlygoing features (Segall and Pollard, 1983; Martel et al., 1988; Cladouhos, 1996; Wibberley et al., 2008; Griffith et al, 2010; Faulknner et al., 2010). It has also been observed, with significant scatter, that fault thickness scales with displacement and spatial extent (Wibberley et al., 2008).

As an example of a well developed fault, Figure 1 is an illustration from Genter et al. (2000) that depicts a typical fault encountered at the EGS project in Soultz, France. Figure 1 is an example, but not the only possible brittle fault geometry. For example, hydrothermal alteration, shown in Figure 1, is not necessary. Also, Figure 1 does not clearly illustrate a damage zone, which is composed of a high density of fracturing decreasing with distance from the fault core (Faulknner et al., 2010).

![Figure 1: Conceptual image of a hydrothermally altered and fractured fault zone at Soultz. From Genter et al., (2000).](image)

Seismic event magnitude scales with the surface area of slip. The seismic moment release during an earthquake is:

$$M_0 = G \int DdA,$$

where $M_0$ is the seismic moment, $G$ is the shear modulus, $D$ is the displacement, and $dA$ is an increment of surface area on the fault (Stein and Wyssession, 2003). From Hanks and Kanamori (1979), the seismic moment magnitude $M_w$ is defined as

$$M_w = \frac{\log_{10} M_0 - 6.06}{1.5},$$

where $M_w$ is the moment magnitude and $M_0$ is defined in units of N-m.
In general, thicker faults have greater spatial extent, and slip over greater spatial extent leads to greater magnitude. Therefore, it would seem reasonable to seek a correlation between fault thickness and induced seismic magnitude. Faults can form in ductile deformation, which could lead to thick, large faults but not a tendency for seismicity. Therefore, we categorize faults not only by thickness, but by their degree of cataclasite formation, which should correlate with the degree to which brittle, strain weakening failure has occurred on the fault.

1.3 Summary of results
We reviewed reports from nine projects around the world that involved injection into crystalline rock (mostly EGS projects) and categorized each according to the degree of brittle fault development (DFD). The projects were Soultz, France, Basel, Switzerland, Rosemanowes, United Kingdom, Bad Urach, Germany, Fjallabacka, Sweden, Ogachi, Japan, the KTB Borehole, Germany, Groβ Schönebeck, Germany, and Cooper Basin, Australia. Some major EGS projects, such as Hijiori, Japan, Fenton Hill, USA, and Desert Peak, USA, were not included because we were not able to find enough information to classify DFD with confidence. We investigated only projects in crystalline rock because induced seismicity has been observed to be much more significant in crystalline than sedimentary rocks (Evans et al., 2012). The subject of lithology is discussed further in Section 4.2.

In the projects reviewed, degree of fault development was very well correlated with the maximum magnitude induced during or soon after stimulation. In two cases, events during long term circulation (with net injection) were significantly greater in magnitude than events associated with stimulation. Including these circulation events, the correlation between maximum magnitude and DFD was weaker but still present. We speculate these differences may be caused by geological heterogeneity, discussed in Section 4.6.

The most likely explanation for the greater magnitudes in places with more developed faults is that larger faults have more spatially continuous slip surfaces that are more conducive to larger magnitudes. This topic is discussed in Section 4.1.

The frictional properties of the materials contacting in the fracture planes affect both the degree of seismicity and the formation of faults. This topic is discussed in Sections 4.2 and 4.3.

At Rosemanowes, there were joints with spatial extent of tens of meters. Because the seismic magnitudes were so small, these larger joints evidently did not slip entirely during individual seismic events. We theorize this may have been caused by either heterogeneity in frictional properties on fractures (discussed in Section 4.2). Another possibility is that the smallest seismic events were caused by mechanisms other than rate and state friction (discussed in the Section 4.4).

We conclude by discussing in Section 4.7 how our results fit into the larger topic of induced seismicity hazard analysis.

We are not aware of a prior study comparing the effect of fault development between EGS projects. These results could be very useful for the estimation of induced seismicity hazard from EGS stimulation. The striking variety in fault development between different locations suggests that degree of fault development is an important factor for not only induced seismicity hazard, but also EGS reservoir engineering and stimulation design in general.

Our results show the value of performing detailed fracture characterization beyond merely mapping fracture orientations and mapping density with depth.

2. METHODOLOGY
For each project considered, we categorized the degree of fault development (DFD) and determined the maximum magnitude induced by stimulation. Our results were based on published reports and personal communication with individuals involved in the projects.

In determining the maximum magnitude, we considered only event that occurred during or shortly after stimulation. In most cases, stimulation related events were the largest events observed at these projects. In cases where maximum magnitudes were observed during long term circulation, the results were reported as a separate data point. In one case, an outlier in the maximum magnitude during stimulation was reported as a separate data point.

Classifications of DFD were based only on direct observation: from wellbore core, surface outcrop, mineshaft, and wellbore imaging logs. Imaging logs are not as useful as direct lithological analysis, but comparisons between core and image logs have shown that it is possible to identify large fault zones from image logs (Genter et al., 1997). Observations from the injector wells were preferred, but in some cases observations from nearby wells, outcrops or mines were used. Because we were concerned with development of brittle deformation, the categories of DFD were focused primarily on the thickness of cataclasite zones. In some cases, a distinctive categorization of DFD could not be made, and the categorization was given in a range.
The categories of DFD were (1) no fractures present, (2) only joints (opening mode fractures) and crack-like faults (thin, planar shear discontinuities) (3) very small faults with at most a few mm thickness of cataclasite, (4) developing faults with a cataclasite zone beginning to develop in a zone with thickness of mm to a few cm, and an altered/damaged zone less than one meter in thickness, and (5) a fully developed core zones thickness of tens of cm and an altered/damaged zone with thickness of meters or more. Figure 1 is a depiction of what a DFD (5) fault might look like.

3. RESULTS

3.1 Summary

In the following sections, we briefly summarize the observations of seismicity and DFD at each project. Most projects were in granite, but the KTB and Bad Urach projects were in gneiss.

The results are summarized in Figure 2, which shows a plot of degree of fault development versus maximum magnitude.

![Figure 2: Maximum magnitude and degree of fault development. Green diamonds show maximum magnitudes during stimulation. Separately tabulated events that are either outliers or associated with long term circulation are shown with blue squares. See Methodology section, for definition of "degree of fault development."](image)

3.2 Soultz

The Soultz project involved several wells that were stimulated in granite at depths around 3.5 km and 5 km during the 1990s and 2000s. Injections involved volumes of tens of thousands of cubic meters of water at flow rates of tens of l/s and wellhead pressures of 5 – 15 MPa (Evans et al., 2012).

The maximum magnitude of the events at 3.5km depth was M = 1.7 during injection and M = 1.9 in the days after injection. During the stimulations of the deeper wells, hundreds of events M > 1.0 were observed, with the largest reaching M = 2.9 (Evans et al., 2012). Circulations tests have been performed at Soultz over the course of months, generally with injection rates less than or equal to production rates, and the largest event associated with those tests was M = 2.3 (Cuenot et al., 2011).

We rated the DFD at Soultz (5). Well core and imaging logs at Soultz showed cataclasite fault cores with thickness of tens of centimeters and damaged/ altered zones with dimensions of ten meters or more (Genter and Traineau, 1996; Genter et al., 2000; Evans et al., 2005).

3.3 Ogachi

The Ogachi project involved several wells that were stimulated at depths from 700m to 1100m during the 1990s (Kitano et al, 2000) in granodiorite (Ito and Kaieda, 2002). Injections for stimulation involved volumes on the order of 3000-10,000 m³ at rates of tens of l/s and wellhead pressures of 18-22 MPa (Kitano et al., 2000).

Kaieda et al. (2010) reported that the maximum magnitude during stimulation at Ogachi was -1.0, except for a single 2.0 event. We tabulated the maximum magnitude as -1.0, and plotted the M = 2.0 event separately because it was such a strong outlier.

Permeable zones prior to stimulation were correlated on wellbore image logs to fractures intersecting the wellbore (Ito and Kaieda, 2002). Detailed investigation of a continuous core of one well from 700 to 1100 m showed that the permeable zones were fractures, not faults (Ito, 2003).

Permeability was associated with open fractures at the edges of dykes, in veins and in individual fractures (Ito, 2003). According to Ito (2003), "faults with gouge greater than 5 mm thick were not found at the Ogachi cores, but faults with slickenside and thin gouge (<2 mm) are abundant ... sporadic minor faults without gouge and up to 3 cm of displacement occasionally displace anhydrite veins and andesite dykes."

Hydrothermal breccia zones were identified at Ogachi (Ito, 2003), but we did not count them in our categorization of fault development. Hydrothermal breccias form from implosion of rock into a void space, perhaps associated with fluid boiling (Sibson, 1986). Because hydrothermal breccia forms from opening mode strain, they would not be expected to form continuous cataclasite slip surfaces required for earthquakes. Hydrothermal breccias can be...
associated with faults, but only at dilational fault jogs (Sibson, 1986). The hydrothermal breccias at Ogachi are not permeable because they are filled with a very fine grained matrix enclosing clasts, nor are they associated with seismicity (Ito, 2003).

On the basis of these observations, we categorize the DFD at Ogachi as (3), containing faults with no more than millimeter scale gouge.

### 3.4 Basel

The Basel project involved a single well that was drilled into granite to a depth around 5 km and stimulated with around 11,500's m$^3$ at rates up to 60 l/s and wellhead pressures of 10 – 30 MPa (Häring et al., 2008). Many seismic events occurred during or soon after stimulation that were greater than M = 2.0, with the largest being M = 3.4 (Häring et al., 2008).

The wellbore was not cored continuously, and there are no any interpretations of the wellbore image logs available in the literature that addressed the topic of fracture zone/fault development. However, on the basis of several sources, we categorized the DFD at Basel as (5) with medium to high confidence.

Häring et al. (2008) reported that there were two "major cataclasite fracture zones" in the open hole interval of the well. Kaeser et al. (2007) performed a petrological study of cuttings and core at selected intervals from the Basel well. From cuttings Kaeser et al. (2007) concluded "only based on cuttings, no clear conclusions on the extent of deformation can be made ... either true ultra-cataclasite (i.e., composed of very fine-grained material with matrix contents > 90%) ... or fragments of the matrix of a protocataclasite or cataclasite s.s. (<50% or 50-90% matrix, respectively, with embedded larger mm-cm-sized rock fragments)" (Kaeser et al., 2007, Section 5.6). There were five zones in the open hole section of the well where cuttings with evidence of cataclasite were observed (Kaeser et al., 2007, Appendix 2), including at 4700m and 4835m, the depths where Häring et al. (2008) reported cataclasite fracture zones.

From these sources (Häring et al., 2008; Kaeser et al., 2007) it is clear that cataclasite fault zones exist in the Basel well, but it is not possible to estimate their thickness. The likely thickness of the faults in Basel can be estimated from comparison with several other wells that were cored nearby in the crystalline basement of northern Switzerland and the southern Black Forest region of Germany (Mazurek et al., 1998). Figure 1-1 in Mazurek et al. (1998) shows the locations of the cored wells and the city of Basel. The cored wells are located from less than 20km away to roughly 100 km from Basel. Mazurek et al. (1998) describes pervasive presence of large, well developed fault zones with decimeter scale cataclasite zones and overall thickness on the order of meters (see Mazurek et al., 1998, Appendix F). On the basis of consistent observations from nearby analog wells, cuttings from the Basel well that indicate cataclasite, and the report of Häring et al. (2008) of "major cataclasite zones," we categorized the DFD at Basel as (5).

### 3.5 Bad Urach

The Bad Urach project involved drilling of a single well to a depth of 4444m into gneiss in three stages between 1977 and 1992. The casing shoe was set at 3320m. The lower part of the casing was perforated within three sections. During the early history of the project, small volume (<100m$^3$) high pressure stimulations were carried out at the true vertical depth (TVD) of the wellbore at that time, around 3334. In 2002, after the well had been deepened, 5600 m$^3$ was injected at rates of tens of l/s and wellhead pressures of 15-34 MPa over the more than 1000 m of open hole at the bottom of the well (Stober, 2011). During the 2002 stimulation, 420 events were detected ranging in magnitude from M = -0.6 to M = 1.8 (Evans et al., 2012). Prior to the 2002 stimulation, there were no observations concerning seismicity at Urach, and so the maximum magnitude during earlier stimulations is not known (Ingrid Stober, personal communication).

The open hole section penetrates successions of biotite-amphibolite gneiss, migmatic gneiss, quartz-diorite gneiss, and biotite-cordierite gneiss. Fractures frequently cut across the wellbore, and leached zones are found on either side of the fractures (Stober, 2011).

We were not able to gather enough data to characterize DFD with a high degree of precision. Here we summarize the available data. Genter (1994) analyzed two cores from Urach, one with length 9m and the other with length 4m. In the 9m long core, K57, cataclasite structures were observed. The thickest cataclasite structure was 5 mm (Albert Genter, personal communication). It is very likely that thicker cataclasite zones were intersected by the well, as the core sampled less than 2% of the open hole section.

Cuttings analyzed from the wellbore contained cataclasite and ultracataclasite. It is believed that larger hydrothermally altered zones were in the range of tens of centimeters and not larger than one meter. From a depth of 2500-3000m, several hundred meters above the injection zone, there was a very disturbed borehole section with alteration zones, fracture zones and cataclastic zones. (Ingrid Stober, personal communication).
It is clear that fractures were present and cataclasite had formed in some faults. Therefore, the DFD should not be categorized as (1) or (2). The thickest cataclasite zone directly observed in the wellbore was 5 mm, which would be on the border between (3) and (4). Because the wellbore was sampled across only a small percentage of its total length, it is very likely that thicker cataclasite zones than 5 mm are present. There is no data to suggest fault zones wider than one meter were found in the wellbore, but it cannot be ruled out. Therefore we assigned Urach a DFD of (3) to (5), with (4) being the most likely.

### 3.6 Rosemanowes

The Rosemanowes project involved the drilling of three wells to depths around 2 km in the Carnmenellis granite in the UK. The project was active from 1978 to 1991. Several stimulations were carried out involving 1000's of m$^3$ of water at flow rates up to 90 l/s and pressures up to 11 MPa (Evans et al., 2012).

The largest event during stimulation was M = 0.16. From 1985 to 1989, various circulation tests were carried out with fluid losses averaging 20% over periods of months, effectively constituting long term injection. In 1987, a M = 2.0 event occurred several kilometers from the injector, and in 1988 a M = 1.8 event occurred near the hypocenter of the first (Evans et al., 2012).

The Rosemanowes project is located in the Cornubian granite batholith in southwest England. Regionally, there is a general trend of NW-SE wrench faulting. Whittle (1991) reported that regional scale mapping suggests that large scale faults likely pass through the Carnmenellis granite (Whittle and McCartney, 1991; Whittle, 1991). Therefore, it is known that in the general region of the Rosemanowes project faulting occurs, at least on the large scale. These regional studies do not give any information about the nature of the deformation at the inferred faults.

On a smaller scale, the fractures of the Carnmenellis granite have been studied extensively through outcrop study, quarries, and underground mapping from mine shafts. In addition, wellbore imaging logs were run in the Rosemanowes wells. Studies of fracturing in the Carnmenellis granite and at the Rosemanowes site have focused overwhelmingly on descriptions of jointing and have not described observation of cataclasite formation (Ghosh, 1934; Heath, 1985; Randall et al., 1990; Whittle, 1991; Pearson et al, 1991; Richards et al., 1991).

Ghosh (1934) described frequent slickensides in the Carnmenellis granite, but not formation of cataclasite zones. Heath (1985) studied fracture permeability in the Carnmenellis granite in several wells up to 700m deep that were drilled and cored roughly 10 km from the Rosemanowes EGS project. Heath (1985) reported that displacements of a few mm could be observed on some joints with development of slickenslides. Heath (1985) observed that permeability was associated with joints, dykes and veins. Major surface and underground mapping studies were performed as part of the Rosemanowes project, and reports on these studies describe only observation of jointing (Randall et al., 1990; Whittle, 1991). Analyses of wellbore imaging logging at the Rosemanowes wells do not mention any indication of fault development (Pearson et al., 1991).

Studies of fracturing within the Carnmenellis granite consistently describe joints, dykes, veins, and very simple crack-like faults. We assigned the DFD at Rosemanowes to be (2).

### 3.7 Fjallbacka

The Fjallbacka project involved drilling, stimulation, and circulation between two wells roughly 500m deep in granite. During the main stimulation in 1986, 400 m$^3$ of viscous gel and water were injected at rates from 20 - 30 l/s (Wallroth et al., 1999) into a 31m long zone near the bottom of the well sealed by an inflatable packer (Eliasson et al., 1990) at injection pressure up to 13 MPa (Evans et al., 2012). It should be noted that this project involved a shallower depth and a lower volume of water than the other projects discussed in this study.

During the 1986 stimulation, 74 events were recorded, with magnitudes ranging from M = -1.3 to M = -0.2 (Evans et al., 2012). During a 1989 circulation experiment, which had 50% recovery of fluid, a seismic event occurred that was felt by project employees. The magnitude is not known.

One of the wells at the Fjallbacka site was cored continuously to 511m. Hydrothermally altered fractured zones were observed, including the development of smectite. There was no report that cataclasite or any development of faults was observed (Eliasson et al., 1990). The assigned the DFD at Fjallbacka (2).

### 3.8 KTB

The KTB project involved a continuously cored pilot hole to 4000m from 1987 to 1989 and a deep borehole to a depth of 9101m from 1990 to 1994. The wells were drilled through paragneisses and metabasites (Hirschmann et al., 1997). Three major injections were carried out. In 1994, about 200 m$^3$ was injected into the main borehole at up to 9 l/s. The water was injected into the 70m open hole section at the bottom of the well (Zoback and Harjes, 1997). In 2000, 4000 m$^3$ of water was injected into
the main borehole at rates up to 1.2 l/s. During the second experiment, seismicity clustered at 3.3 km, 5.4 km, and 6.6 km in addition to the bottom of the wellbore, likely because of casing leaks at the shallower depths (Evans et al., 2012). From 2004-2005, 84,600 m³ water was injected over ten months into the pilot hole open hole section from about 3850-4000m depth (Shapiro et al., 2006).

In the depth range of 3-6 km, the largest seismic event observed during any of the injection experiments was \( M = 0.5 \). At the depth of 9 km, the largest magnitude event observed was \( M = 1.2 \). The \( M = 1.2 \) event occurred during the initial injection test, and in the subsequent injection tests, there was never an event larger than \( M = 0.5 \) (Evans et al., 2012).

The continuously cored pilot hole to 4000m is the best source of information about the character of the fracture network. It is not certain how well those data can be extrapolated to the depth of 9000m of the full hole, which introduces some uncertainty. Our assignment of DFD was based on the cored pilot hole. Various significant fault zones were observed at various depths throughout the both the pilot and deep boreholes (Hirschmann et al., 1997). Zulauf (1992) did a detailed examination of the fracture networks intersecting the pilot hole. Zulauf (1992) inferred various stages of deformation leading to different sets of fractures and faults. In the upper 2000m of the wellbore, cataclasite zones with thickness up to 5 cm were observed. This depth range is far above the injections, and so these observations were not used for the assignment of DFD. Below 2000m, the thickest cataclasite zones were a few millimeters. There were fault zones with thickness of meters. Within the fault zones, there were many thin cataclasite zones spaced from a few centimeters to decimeters apart. These faults do not appear to be associated with purely brittle faulting mechanisms. The deformation is significantly less localized than in faults with layers of cataclasite centimeters or decimeters thick. Instead, deformation is spread across a large number millimeter thickness cataclasite zones. In many cataclasite zones, there were high concentrations of graphite (Zulauf et al., 1990). In some cases, there was evidence of semibrittle deformation, with quartz minerals fracturing, but biotite and muscovite minerals deforming plastically (Zulauf, 1990). Brittle deformation is associated with the strongest strain weakening, which tends to localize failure (Ben-Zion and Sammis, 2003). The rather distributed strain in the fault zones is likely due to ductile failure.

On the basis of the observations from the cored pilot holed, the DFD at KTB is categorized as (3). Meter scale thickness faults were present, but they were associated with ductile or mixed ductile-brittle deformation and contained zones of cataclasisite no thicker than a few millimeters.

### 3.9 Groß Schönebeck

The Groß Schönebeck project involved the stimulation of a well drilled to a depth around 4175m with casing set about 40m above the bottom of the wellbore. The bottom section of the hole was drilled through volcanic andesites, and above that the well was completed with perforations through casing in sandstone. The volcanic section of the wellbore was stimulated with 13,170 m³ at flow rates alternating between around 20 l/s and around 150 l/s. The injection well head pressures alternated between around 30 MPa and around 50 MPa (Zimmermann et al., 2008; Moeck et al., 2009). Subsequently, the sedimentary layer was stimulated at two depths with about 500 m³ of water and 100 tons of proppant. The largest magnitude earthquake detected was -1.0 (Moeck et al., 2009).

The only data available to characterize DFD is wellbore imaging logs and a few meters of core from the volcanic section. Neither the core nor the wellbore imaging logs showed any indication of fault or fracture zones (Günter Zimmermann, personal communication). Therefore we categorized DFD as either (2) or (3).

### 3.10 Cooper Basin

The Cooper Basin project involved the drilling of four wells between depths of 4-5km in granite, starting in 2003. The largest event that has occurred at the Cooper Basin site was \( M = 3.7 \). It occurred during the stimulation of the well Habanero 1 in 2003. During that stimulation, 20,000 m³ was injected at flow rates up to 50 l/s, and injection pressures up to 60 MPa (Asanuma et al., 2005).

Wellbore imaging logs have been run over approximately a 2000m combined in the granite sections of the four wells at the Cooper Basin project. Thick fault zones have been observed in the image logs. In the wellbore imaging logs, the major fault zones contain a core with thickness of a few meters, surrounded by subsidiary fracturing within ten meters of the core (Doone Wyborn, personal communication). This observation is consistent with the standard model of a brittle fault in granite (Figure 1, Genter et al., 2000; Faulkner et al., 2010). Therefore, even though core is not available to confirm, the wellbore imaging logs give strong evidence of large scale faulting. We categorized the DFD as (5).
4. DISCUSSION

4.1 Slip surface continuity

Figure 2 shows that there is an excellent correlation between the degree of fault development and the severity of induced seismicity. The simplest explanation is that thicker, more developed faults have greater spatial extent, and greater surface area of slip during a seismic event causes greater magnitude. Ruptures should find it much easier to propagate across a continuous, large fault than a collection of smaller faults or joints. Any discontinuity in a slip surface should be considered a barrier to rupture propagation. A powerful stress concentration is generated at the rupture front of an earthquake, and the stress concentration helps the earthquake to propagate (Fruend, 1990). Stress concentration weakens with distance from the rupture, and so any discontinuity in the slip surface will tend to create a barrier to propagation. Wesnousky (2006) found that ruptures tended to terminate at fault steps during earthquakes on the San Andreas Fault.

It is conceivable that multiple fractures or faults could slip during a single seismic event, combining their surface area to effectively form a larger surface area than their thickness would suggest possible. The case of many smaller faults organizing into a single large event has been hypothesized to be the mechanism of seismicity at Basel (Häring et al., 2008). Häring et al. (2008) observed strike-slip focal mechanisms oriented in NS-EW planes but seismicity organized into a planar feature oblique to that orientation at NNW-SSE. Häring et al. (2008) also observed several distinct arrivals in the seismic signal, suggesting that the seismic events consisted of slip on several slip surfaces. This mechanism is consistent with outcrop studies of faults granite that describe en echelon arrays of faults connected by tensile steps (Segall and Pollard, 1983; Martel et al., 1988; Griffith et al., 2009). In the brittle regime, faults form complex geometries generally made up of many shear fractures that have linked together (Faulkner et al., 2010). A feedback loop develops between strain weakening and localization of deformation that lead to progressive localization of strain onto fewer, more significant features (Ben-Zion and Sammis, 2003). Therefore, even large fault zones likely do not have perfectly continuous slip planes. But they are likely to be formed of collections of slip planes that have naturally developed over time to align and be spatially clustered in such a way as to be conducive to rupture propagation. These closely aligned slip planes might be considered a single fault. Faults consisting of many slip planes have far greater spatial continuity than distributions of joints or fractures that formed in mechanisms other than brittle faulting.

Continuity of slip surfaces may not be the full explanation. At Rosemanowes, surface outcrops suggested that joints commonly had spatial extents of 80m or more (Whittle, 1990). The seismic moment of a circular crack embedded in a linear elastic material is (Kanamori and Anderson, 1975)

$$M_0 = \frac{16}{7} \Delta \alpha u^3,$$

where $a$ is the radius and $\Delta \sigma$ is the stress drop. Assuming a radius of 40m and a stress drop of 1 MPa, the moment would be $1.5 \times 10^{11}$ N-m, which corresponds (Equation 3) to a magnitude of 1.39. Other than the two events that occurred during long term circulation, no events at Rosemanowes came close to being that large. Based on this analysis, entire joints almost never failed in a single seismic event.

We do not know how to explain the observation that entire large joints apparently never failed during a single seismic event. One possibility is that joint extents from surface outcrop are not representative of joint extents at depth. A second possibility is that reported individual joints were not actually single, continuous fractures. A third is that the natural surface roughness of the joints inhibited rupture propagation. A fourth is that the frictional properties of the joints and their infilling was not favorable for seismic slip, and seismic events were limited to small regions of heterogeneity on the joints where unstable slip was fractionally favorable. A fifth is that seismic slip was entirely unfavorable according to a rate and state framework (discussed in Section 4.2), and other mechanisms are needed to explain the presence of seismicity (discussed in Section 4.4).

4.2 Seismic and aseismic slip

Whether faults tend to slip seismically or aseismically obviously matters for induced seismicity. Seismic slip occurs when friction weakens rapidly on a fault as a consequence of tiny asperities on a fracture surface coming out of contact during slip (Dieterich, 1979, 1992, 2007). According to rate and state friction theory, the friction on a fault is function of sliding velocity $v$, and the average duration of contact for contacting asperities (called state), $\theta$:

$$\mu = f_0 + a \log\left(\frac{v}{v_0}\right) + b \log\left(\frac{\theta v_0}{d_c}\right),$$

where $f_0$, $v_0$, $a$, and $b$ are constants and $d_c$ is the characteristic weakening distance. According to rate and state friction theory, seismic slip should only be
possible if the parameter $a$ is less than the parameter $b$, a condition referred to as velocity weakening. The condition $b > a$, is referred to as velocity strengthening (Dieterich, 2007).

Whether or not rate strengthening or weakening conditions exist depends strongly on composition of the fault gouge and the temperature. The exact composition of a fault gouge would depend on the protolith and the history of mineral alteration. Unfortunately, rate and state experiments have not been carried out in a wide range of gougues and temperatures. The most relevant experiments to faulting in granite were carried out by Blanpied et al. (1995), who performed experiments on wet fractures in granite and found two transitions between velocity strengthening and weakening. They found a transition from strengthening to weakening around 25°C-100°C and a transition from velocity weakening to strengthening around 250°C-300°C. These results are risky to apply in general because the frictional dependence on temperature may vary depending on the specifics of gouge. For example, den Hartog et al. (2012) studied the temperature dependence of friction for simulated fault gouge of illite shale and found rate strengthening behavior for the temperatures up to 250°C. He et al. (2007) studied the temperature dependence of friction in fractures in gabbro and found rate weakening behavior only in the temperature range 200°C to 310°C. Tembe et al. (2010) investigated the frictional behavior of mixtures of quartz, montmorillonite, and illite at room temperature in wet conditions. They found a general trend of increasing rate strengthening and decreasing friction coefficient with increasing clay content in the mixtures.

These results are obviously highly relevant to the issue of induced seismicity, but it is not clear how reliably they could be applied in practice because of the complexity of natural systems and the limited number of the experiments that have been carried out. Nevertheless, the understanding that seismicity depends on the frictional properties of the fault can be instructive. Several general principals can be gleaned from experiments. First, increasing clay content contributes to aseismic slip and a weaker coefficient of friction. Second, temperature plays a role, with seismic slip favored only in certain temperature ranges.

Frictional behavior appeared to play a role in seismicity the KTB borehole, where seismicity was mild. In this case, there was clear evidence of ductile and aseismic slip from core (Zulauf et al., 1990, 1992). Even at Soultz, where there were large cataclasite zones and significant seismicity, evidence suggested considerable aseismic slip may have taken place, at least in the shallower reservoir at 3.5km (Cornet et al., 1997). Some fault zones at Soultz contained significant concentrations of illite (Evans et al., 2005), which has been associated with weaker friction (Tembe et al., 2010) and aseismic deformation (den Hartog et al., 2012). The experience at the shallower Soultz wells suggests that aseismic slip could be possible even in places with thick cataclasite zones. This is especially true because faults could have formed in the past under brittle failure but have subsequently begun to deform aseismically due to alteration related changes in mineralogy, changes in depth, or changes in temperature. The implication is that characterization of mineralogy within faults could be important for induced seismicity hazard estimation.

Evans et al. (2012) noted that induced seismicity is far more likely to be an issue in crystalline than sedimentary rock. We can speculate that this may be a result of a tendency for aseismic, rather than seismic slip, on faults that form in sedimentary rocks. This could be a consequence of the different minerals that are likely to form in gouge from different protoliths.

4.3 Dependence on depth

There was a clear tendency among the projects we studied towards greater seismicity at depth. Evans et al. (2012) also noted a relationship between depth and seismicity. This was particularly evident at the Soultz project, where the shallower reservoir had a significantly lower maximum magnitude than the deeper reservoir.

Based on the results discussed in Section 4.2, we can speculate that the higher temperatures at greater depth contribute to more friction weakening behavior. Another possible depth effect is that clays, which are generally associated with aseismic slip, are not stable at higher temperatures. For example, clay minerals have been observed to occur no deeper than certain horizons at the Coso and the Desert Peak geothermal fields (Kovac et al., 2005; Davatzes and Hickman, 2009, 2010; Lutz et al., 2010). The clay smectite was identified in the shallowest and lowest temperature project that we reviewed, Fjallbacka (Eliasson, 1990). At Soultz, illite was found in some fault zones (Evans et al., 2005), which may explain the lower magnitudes at shallower depth at Soultz. Davatzes and Hickman (2009) described how temperature, clay formation, and fault development interact to control the thermal and permeability regime of the Coso geothermal field.

Another obvious effect of depth is that deeper faults are under greater stress, which would tend to lead to greater stress drops and higher magnitudes. However, stress could account for only rather mild variations in induced magnitudes.
To fully detangle the effects of depth and fault zone development would require more data. Unfortunately, in our dataset there were no deep projects in granite with undeveloped fault zones nor were there shallow projects with well developed faults.

4.4 Alternative mechanisms of acoustic emission

It is widely acknowledged that earthquakes are caused by slip on preexisting faults, and rate and state friction is the leading theory to explain the frictional evolution that leads to seismic slip. However, in this section we discuss an alternative hypothesis to explain seismicity in locations where seismic events were very small. In this hypothesis, seismic slip was not favorable at these locations according to a rate and state mechanism. Instead, events may have been caused by alternative mechanisms. Rate and state is by far the most likely explanation for larger events, perhaps $M > 0$, but these alternative mechanisms may be reasonable hypotheses to explain smaller events.

This discussion is motivated especially by the observations at Rosemanowes, where joints with spatial extent up to 80m have been observed in outcrops, yet the maximum magnitude during stimulation was $M = 0.16$. As discussed Section 4.1, these small magnitudes imply that entire joints never slipped during a single event. Yet this is somewhat surprising, because if a seismic event was able to nucleate, the intense concentration of stress at the rupture front would tend to cause the rupture to continue to propagate (McClure and Horne, 2011a). There may be a rupture mechanics explanation for this phenomenon that is consistent with rate and state friction. The most obvious is that due to heterogeneity, seismic events were only frictionally favorable on small patches of the fractures. They nucleated on those patches and then arrested as they propagated into regions where seismic slip was not favorable. However, in this section we investigate the alternative hypothesis.

During triaxial compression tests, intact samples of rock are strained axially until failure while a confining radial stress is applied. Acoustic emissions are observed consistently during these experiments. Prior to failure, the acoustic emissions are distributed throughout the rock sample, but during failure they localize around the failure surface (Lockner et al., 1991; Lockner, 1993; Jaeger et al., 2007). Experiments with intact samples of granite (Lockner et all, 1991; Jaeger et al., 2007) show that deviatoric stresses of hundreds of MPa are necessary to induce the generation of a new shear fracture. The greater the confining pressure, the greater the deviatoric stress required to initiate failure (Jaeger et al., 2007). Because of the huge strength of intact rock, especially granite, it is unlikely that major fracturing occurs through the failure of intact rock during stimulation. To support that conclusion, zones of enhanced permeability are correlated with zones of preexisting fractures (Pearson et al., 1991; Wallroth et al., 1999; Ito and Kaieda, 2002; Evans et al., 2005). The tensile strength of granite, and all rocks, is far less than the compressive strength, on the order of MPa (Jaeger et al., 2007). Both failure of intact rock and failure of preexisting fractures can be described using Equation 1, the Coulomb failure criterion, but the cohesion, $S_o$, of intact rock is on the order of hundreds MPa, while the cohesion of an individual fracture is on the order of MPa (Jaeger et al., 2007). Because of these different strength thresholds, shear slip of preexisting fractures or propagation of tensile cracks would be expected to happen during injection, not extensive failure of intact rock.

Keeping in mind these strength considerations, we can speculate about alternative mechanisms that could lead to very small seismic events. One possibility is abrupt loss of cohesion, $S_o$. This might occur, for example, if a joint were sealed shut by mineralization. The chemical bonds of the mineralization might be abruptly broken in a section of a joint. The shear strength of a mineralized joint would be much lower than the strength of intact rock (Papaliangas et al., 1993; Armand et al., 1998). Flow is known to be highly channelized during flow within individual fractures (Auradou et al., 2006), and so events might be caused by the breaking of mineralization in low permeability regions of a fracture surrounded by higher permeability channels. Fluid pressure would be elevated in the higher permeability channels, leading to their slipping earlier and then subsequently triggering rapid slip on the surrounded lower permeability regions as the mineralization is broken. This mechanism involves slip on preexisting fractures but is quite different from the mechanism described by evolution of rate and state friction in Equation 5. Rate and state friction would allow seismic slip to occur repeatedly at a given location on a fracture (McClure and Horne, 2011a), which would make slip across an entire joint more likely. Abrupt failure of mineralization could only happen once at a location, making it possible that fractures could fail in a series of smaller events that never encompass the surface area of an entire large fracture.

Stress concentrations can be very strong locally near the tip of a fracture that has slipped. Concentrations of stress can generate both compressive and tensile forces in different locations relative to the tip (Sibson, 1986). These local stress heterogeneities might cause development of either tensile fracturing or possibly even compressive failure of intact rock.
Aki et al. (1977) theorized that jerky growth of a tensile fracture could result in microseismicity. Observations from triaxial tests demonstrate that compressive failure of intact rock can result in emission of acoustic waves (Lockner et al., 1991; Lockner, 1993).

These alternative mechanisms would not be able to generate significantly sized microseismic events. However, they could explain the occurrence of seismicity in places where magnitudes are very small. These mechanisms could explain microseismicity in places where fractures do not have rate and state frictional characteristics needed for rapid slip.

4.5 Background seismicity

Evans et al. (2012) noted that induced events tended to be more significant in places with stronger natural seismicity. It is natural that seismically active regions would tend to contain more significant fault zones because seismicity occurs because of slip on faults and over time seismicity helps form faults. However, low natural seismicity does not preclude the occurrence of significant induced seismicity. The natural seismic hazard at Soultz is rather low (Evans et al., 2012), yet relatively large events were triggered ($M > 2.5$). Even if modern strain rates are too low to generate seismicity and cause significant fault formation, faults may have formed in the past and could slip seismically if placed under elevated fluid pressure.

4.6 Outlier events and geological heterogeneity

At the Ogachi, Rosemanowes, and Fjallbacka projects, there were one or two events that were much greater in magnitude than any others. At Rosemanowes and Fjallbacka, these events occurred during long term unbalanced circulation experiments. These circulations involved the injection of more water than was produced, which means that they constituted long term injection (Evans et al., 2012; Kaieda et al., 2010). It is not possible to give a certain explanation for these events, but we speculate they may have been caused by geological heterogeneity.

In a statistically homogeneous region, a fault could not be present that was radically larger or otherwise different than the population of faults sampled in a statistically significant length of a wellbore. Power law distributions of fault sizes are commonly observed, and this has been theorized to be caused by the mechanism of fault growth by which smaller faults progressively link up to develop larger features (Cladouhos and Marrett, 1996). However, sharp discontinuities in statistical homogeneity are common in the earth. Because of heterogeneity, faults observed at a wellbore do not necessarily have to be representative of faults in the surrounding earth. In adjacent lithological units, potentially with different ages, and chemical and mechanical properties, faulting might develop differently. Even in a seemingly homogeneous region such as a granitic batholith, different units of granite can exist with significantly different fracture densities and orientations. For example, several different lithological units have been described in the granite at Soultz (Dezayes et al., 2010).

Large faults are often found at the interface between facies. One possible explanation is that interfaces are mechanically weaker, which would encourage strain localization. Another possible explanation is that large faults create large offsets, juxtaposing two different facies. For example, at Soultz, a large fault is located at the boundary between two types of granite (Dezayes et al., 2010). A fracture zone is reported at a lithological boundary at the EGS project at Desert Peak in Nevada (Davatzes and Hickman, 2009). A fault juxtaposing granite and volcanic rocks is inferred to be located roughly 500m from the Ogachi test project (Suzuki and Kaieda, 2000).

During long term injection, injected fluid could spread a greater distance away from the wellbore than during stimulation, increasing the possibility that the pressure could be perturbed on a fault that is statistically differently than the faults in the immediate vicinity of the wellbore. In addition, thermal contraction of the reservoir due to injection of cooler water would cause volumetric strain, which could perturb stress a considerable distance from the injector and producer wells (Segall, 1989; McGarr et al., 2002).

The circulation related event at Rosemanowes was around $M = 2.0$, and the magnitude of the circulation related event at Fjallbacka is unknown, but it was strong enough to be felt at the surface (Evans et al., 2012; Kaieda et al., 2010). During circulation tests in the deep reservoir at Soultz from 2005-2010, many seismic events occurred, with a maximum magnitude of $M = 2.3$. These circulation tests at Soultz were generally carried out with production rates equal to or greater than reinjection rates (Cuenot et al., 2011). Events during circulation were much larger than events during stimulation only for projects where the stimulation events were very small. For the case of Soultz, where the stimulation events were relatively large, the circulation events were of somewhat lesser magnitude.

4.7 Seismic hazard analysis

In this section, we discuss how this study fits into the broader topic of seismic hazard analysis. As discussed in the introduction, there are two important mechanisms of induced seismicity: direct pressure perturbation and volumetric contraction/expansion of
the field from poroelastic or thermoelastic forces. The spatial and time scales of these mechanisms may be different.

In EGS stimulation, direct pressure perturbation during injection occurs over a period of days in a region that is localized around a wellbore. In this case, induced seismicity is unmistakably correlated to the injection. The largest seismic event ever caused by EGS stimulation is probably $M = 3.7$ at the Cooper Basin project (Majer et al., 2007).

If long term EGS circulation is unbalanced, so that more fluid is injected than produced, then there may be a long term spreading of a pressure perturbation over greater time and length scales than during stimulation. Unbalanced EGS circulation could be compared to deep injection projects in crystalline rock such as at Rocky Mountain Arsenal, Colorado, Rangely, Colorado, and Ashtabula, Ohio (Nicholson and Wesson, 1990; McGarr et al., 2002). The maximum magnitudes associated with injection at these sites was $M = 4.85$, $M = 3.6$, and $M = 3.1$, respectively. These projects involved injection over months and years of much greater volumes of water than are injected during EGS stimulation.

Thermal contraction of a reservoir occurs slowly and the stress perturbation can be spread some distance from the region where the rock is actually cooled (Segall, 1989; McGarr et al., 2002). Induced seismicity from thermal contraction may also be correlated to the geothermal production, but it is not as obvious and direct. Hazard from thermal contraction induced seismicity is the same for EGS projects as it is for geothermal projects in general. The largest geothermal field in the world, the Geysers, California, has hosted some of the largest induced events, the largest being $M = 4.6$ (Majer et al., 2007). These are likely caused by thermal contraction from massive injection of cool water into the reservoir (Rutqvist and Oldenburg, 2008).

As noted in the Section 1.1, McGarr (1976) proposed that the seismic moment release during injection should be proportional to the volume of fluid injected. His prediction has generally been supported by subsequent experience comparing seismicity at different locations (McGarr et al., 2002; Nicol et al., 2011; Evans et al., 2012) and during injection at individual sites (Bommer et al., 2006; Baisch and Vörös, 2009, section 4.2).

To extend McGarr’s (1976) original idea somewhat, the induced seismic hazard scales with spatial extent of the stress perturbation, whether it is direct fluid injection or thermal contraction. With that in mind, it is evident that the scale of the stress perturbation caused by EGS stimulation is smaller than the scale of long term injection or thermal contraction.

Therefore, in the long term, seismic hazard from thermal contraction and unbalanced circulation should be greater than hazard from EGS stimulation. This suggests that long term unbalanced circulation in an EGS system would significantly increase seismic hazard, as it would for any subsurface project. In the case of long term balanced circulation, seismic hazard would arise mainly from thermal contraction. Seismic hazard from thermal contraction should be the same for EGS as for geothermal in general.

Both theoretical considerations and the empirical results from this study suggest that short term hazard from EGS stimulation are well predicted by characterization of the faults and fractures intersecting the wellbore. Observations of outlier events at Ogachi, Rosemanowes and Fjallbacka suggest that long term seismic hazard may be less well predicted by characterization of fracture networks. This is probably due to geological heterogeneity. Therefore, while investigation of local fault properties appears to be useful for stimulation induced seismic hazard analysis, it is not completely reliable, especially for long term hazard.

5. CONCLUSION

Comparative study of the effect of fault development on EGS has not previously been carried out. Our survey of fracture networks at different projects around the world suggests that fault and fracture characterization may be effective for estimation of hazard associated with EGS stimulation. A very good correlation was observed between the thickness of cataclasite zones and maximum induced event magnitude. Theoretically, this relationship might be expected, as ruptures would propagate more easily on spatially extensive and continuous slip surfaces with frictional properties favorable for seismic slip. Identification of mineralogy of fault infill was also identified as being important in order to identify the presence of clays or other minerals that would promote aseismic deformation.

In two places where only very tiny seismic events occurred during stimulation, long term injection led to a small number of relatively much larger events. This was probably due to geological heterogeneity.

At the Rosemanowes project, surface outcrop studies described joints with dimensions of tens of meters. However, very small observed magnitudes suggest that entire joints were not able to slip during individual seismic events. This suggests that seismic slip may have only been frictionally favorable on small patches within joints. An alternative hypothesis is that smaller magnitude events seen at Rosemanowes and other locations may not have been
caused by rate and state frictional sliding mechanisms.

The extreme variation in fault development at different historical EGS projects is striking, especially because this point has not been emphasized in the EGS literature. In some cases, it was challenging to find the information needed to characterize the degree of fault development. For some projects, it was not possible. Basic aspects of reservoir engineering and stimulation modeling would be affected by the degree of fault development. In different settings, fundamentally different reservoir behaviors might be expected and very different physical processes might be taking place. This study suggests characterization of the degree of fault development should be a priority in EGS field demonstrations.

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