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Seismic diffusivity and the influence of heterogeneity on injection-induced seismicity

Ryan Haagenson[1] and Harihar Rajaram[1]

[1]Johns Hopkins University

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Abstract

The spatiotemporal patterns of injection-induced seismicity (IIS) are commonly interpreted with the concept of a triggering front, which propagates in a diffusion-like manner with an associated diffusivity parameter. Here, we refer to this diffusivity as the "seismic diffusivity". Several previous studies implicitly assume that seismic diffusivity is equivalent to the effective hydraulic diffusivity of the subsurface, which describes the behavior of the mean pressure field in heterogeneous porous media. Seismicity-based approaches for hydraulic characterization or simulations of IIS using domains of homogeneous equivalent porous media are implicitly based on this assumed equivalence. However, seismicity is expected to propagate with the threshold triggering pressure, and thus not be controlled by the evolution of the mean pressure field. We present numerical simulations of fluid injection to compare the seismic and effective hydraulic diffusivities in heterogeneous formations (including fractured rock). The numerical model combines uncoupled, linear pressure diffusion with the Mohr-Coulomb failure criterion to simulate IIS. We demonstrate that connected pathways of relatively high hydraulic diffusivity in heterogeneous media (particularly in fractured rock domains) allow the threshold triggering pressure to propagate more rapidly than predicted by the effective hydraulic diffusivity. As a result, the seismic diffusivity is greater than the effective hydraulic diffusivity in heterogeneous porous media, possibly by an order of magnitude or more. Additionally, we present a case study of IIS near Soultz-sous-Forêts where seismic diffusivity is found to be at least one order of magnitude larger than the effective hydraulic diffusivity.

Seismic diffusivity and the influence of heterogeneity on injection-induced seismicity

Ryan Haagenson[1,2] **and Harihar Rajaram**[1,2]

[1]Environmental Health and Engineering, Johns Hopkins University, 615 N Wolfe St, Baltimore, MD, 21205, USA

[2]Civil, Environmental, and Architectural Engineering, University of Colorado Boulder, 1111 Engineering Dr, 422 UCB, Boulder, CO, 80309, USA

Key Points:

- We present numerical simulations to investigate the influence of heterogeneous hydraulic properties on induced seismicity
- We find the seismic diffusivity associated with propagation of induced seismicity to be distinct from the effective hydraulic diffusivity
- The seismic diffusivity can be an order of magnitude larger than the effective hydraulic diffusivity in fractured rock

Corresponding author: Ryan Haagenson, `rhaagenson@jhu.edu`

Abstract

The spatiotemporal patterns of injection-induced seismicity (IIS) are commonly interpreted with the concept of a triggering front, which propagates in a diffusion-like manner with an associated diffusivity parameter. Here, we refer to this diffusivity as the "seismic diffusivity". Several previous studies implicitly assume that seismic diffusivity is equivalent to the effective hydraulic diffusivity of the subsurface, which describes the behavior of the mean pressure field in heterogeneous porous media. Seismicity-based approaches for hydraulic characterization or simulations of IIS using domains of homogeneous equivalent porous media are implicitly based on this assumed equivalence. However, seismicity is expected to propagate with the threshold triggering pressure, and thus not be controlled by the evolution of the mean pressure field. We present numerical simulations of fluid injection to compare the seismic and effective hydraulic diffusivities in heterogeneous formations (including fractured rock). The numerical model combines uncoupled, linear pressure diffusion with the Mohr-Coulomb failure criterion to simulate IIS. We demonstrate that connected pathways of relatively high hydraulic diffusivity in heterogeneous media (particularly in fractured rock domains) allow the threshold triggering pressure to propagate more rapidly than predicted by the effective hydraulic diffusivity. As a result, the seismic diffusivity is greater than the effective hydraulic diffusivity in heterogeneous porous media, possibly by an order of magnitude or more. Additionally, we present a case study of IIS near Soultz-sous-Forêts where seismic diffusivity is found to be at least one order of magnitude larger than the effective hydraulic diffusivity.

1 Introduction

The injection of fluid into the subsurface is a common practice in several industries such as geothermal energy production, wastewater disposal, or hydraulic stimulation. In some cases, changes to the subsurface fluid pressure distribution or stress state may trigger seismic events, with potential risks to the nearby population and infrastructure depending on the magnitude of the induced seismicity (Chase et al., 2019; Ellsworth, 2013; Majer et al., 2012; Rutqvist et al., 2014; Weingarten et al., 2015). The first recorded incident of injection induced seismicity (IIS) was near Denver, Colorado in 1962 (Healy et al., 1968). The authors concluded that an increase in the fluid pressure within the injection formation led to a decrease in the effective normal stress acting along previously dormant faults and fractures, thus prompting increased seismic activity in the area. In the United States, IIS has been mostly associated with the disposal of coproduced wastewater from oil and gas operations (EPA, 2016; Weingarten et al., 2015); although, IIS is common in numerous other industries, including geologic carbon sequestration and geothermal energy (Catalli et al., 2016; Cuenot et al., 2008; Majer et al., 2012; Riffault et al., 2018). Several studies have investigated the behavior and physical mechanisms of IIS in numerical, statistical, and field studies (Chang & Segall, 2016; Hajati et al., 2015; Langenbruch & Zoback, 2016; Rudnicki, 1986; Segall & Lu, 2015; Shapiro et al., 1997; Talwani & Acree, 1985).

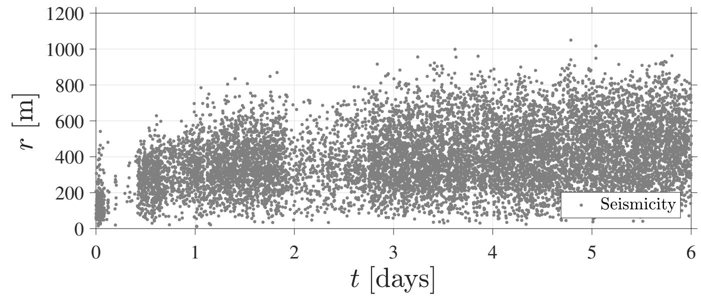

Figure 1. The recorded seismic events from the 2000 fluid injection experiment near Soultz-sous-Forêts, France plotted as r-t plot (Leptokaropoulos et al., 2019).

Complex changes to the subsurface stress state – either through poroelastic effects or static stress transfers – certainly impact the behavior of seismicity during fluid injection (Catalli et al., 2016; Chang & Segall, 2016; Chang & Yoon, 2020; Haagenson et al., 2020; Jha & Juanes, 2014; Rutqvist et al., 2013; Schoenball et al., 2012; Segall & Lu, 2015; Zhai & Shirzaei, 2018; Zhai et al., 2019). However, most studies assume that the decrease of effective normal stress acting along a fault or fracture due to the increase of pore fluid pressure from fluid injection is the dominant physical mechanism causing seismicity (Brown et al., 2017; Dempsey & Riffault, 2019; Hummel & Shapiro, 2013; Keranen et al., 2014; Langenbruch et al., 2018; Nakai et al., 2017; Shapiro et al., 1997, 2002; Shapiro & Dinske, 2009b; Shapiro, 2015; Rothert & Shapiro, 2003; Talwani & Acree, 1985). If we neglect changes to the subsurface stress state, the triggering of seismic events during fluid injection is expected to be predominantly controlled by fluid flow. Here, we model fluid flow using an uncoupled diffusion equation of fluid pressure, based on the clear precedence in the classical theory of groundwater flow (Bear, 1972; Charbeneau, 2006; De Marsily, 1986; Verruijt, 2013) and of linear poroelasticity in cases of irrotational deformation (Cleary, 1977; Rice & Cleary, 1976; Rudnicki, 1986).

At sites prone to IIS, seismic events are observed to occur at increasingly larger radial distances from the injection location as time progresses. A convenient approach to understanding the spatiotemporal behavior of IIS is to plot the distance between the injection location and the seismic event hypocenter against the event occurrence time. These plots are commonly referred to as r-t plots. As an example, the recorded seismic dataset from the fluid injection experiment near Soultz-sous-Forêts, France from the year 2000 is shown as an r-t plot in Figure 1 (Delepine et al., 2004; Leptokaropoulos et al., 2019). We will further discuss the results of this fluid injection experiment in Section 4.

Several previous studies have employed r-t plots to investigate the diffusive propagation of the so-called triggering front (Antonioli et al., 2005; Chen et al., 2012; Ingebritsen & Manning, 2010; Goebel et al., 2017; Goertz-Allmann et al., 2017; Haffener et al., 2018; Hummel & Shapiro, 2013; Shapiro et al., 1997, 2002; Shapiro & Dinske, 2009a; Yong et al., 2018; Yu et al., 2019). In the most general terms, the triggering front is the three-dimensional surface propagating away from the injection location (which need not be radially symmetric) at which the onset of seismicity occurs (Rothert & Shapiro, 2003; Shapiro et al., 1997; Shapiro & Müller, 1999; Shapiro et al., 2002; Shapiro, 2015). If we assume that the onset of seismicity is induced by a threshold triggering pressure increment (p_t) (Dempsey & Riffault, 2019; Gischig & Wiemer, 2013; Goebel et al., 2017; Keranen et al., 2014; Shapiro, 2015), then the triggering front may be alternatively viewed as the isobaric surface with a pressure increment value equal to p_t. While r-t plots directionally aggregate the three-dimensional seismic dataset and thus only reflect variations of seismicity in the radial direction, they are still helpful for tracking the location of the triggering front. As previous studies have shown (Delepine et al., 2004; Shapiro et al., 1997, 2002; Shapiro, 2015), the location of the triggering front can be approximately identified by the upper envelope of the seismic data cluster in an r-t plot such as Figure 1, if the subsurface is assumed to be homogeneous and isotropic. This represents the farthest radial distance from the injection location where seismicity has occurred at any given time.

It is well established that the propagation of seismicity and the triggering front (as observed in r-t plots) is approximately described by a diffusion-like process (Shapiro et al., 1997; Shapiro, 2015; Talwani & Acree, 1985). As such, expressions describing the diffusive propagation of the triggering front can be fit to the upper envelope of the seismic data cluster, producing an estimate of diffusivity (Delepine et al., 2004; Hummel & Müller, 2009; Hummel & Shapiro, 2012, 2013; Rothert & Shapiro, 2003; Segall & Lu, 2015; Shapiro et al., 1997; Shapiro & Müller, 1999; Shapiro et al., 2002; Shapiro & Dinske, 2009a, 2009b; Shapiro, 2015). For clarity, we refer to this estimate of diffusivity as the "seismic diffusivity", borrowing the term from Talwani and Acree (1985). There, the authors suggested that estimates of seismic diffusivity (associated with the spatiotemporal patterns of IIS) are in fact accurate estimates of the effective hydraulic diffusivity of the injection formation. If true, this indicates that methods for estimating seismic diffusivity are in fact seismicity-based approaches for subsurface hydraulic characterization. Since Talwani and Acree (1985), numerous other studies have also implicitly assumed that seismic diffusivity and effective hydraulic diffusivity are equivalent, and thus the term "seismic diffusivity" is not widely used (Haagenson et al., 2018; Haagenson & Rajaram, 2020).

The seismic diffusivity and effective hydraulic diffusivity are no doubt equivalent in homogeneous and isotropic porous media, where isobaric surfaces (including the triggering front) will be radially symmetric. When employing r-t plots, the location of the triggering front is described as a single radial distance from the injection location, which hence implicitly assumes that the subsurface is homogeneous and isotropic. Indeed, most studies of the spatiotemporal patterns of IIS, using both analytical models (Antonioli et al., 2005; Chen et al., 2012; Goebel et al., 2017; Goertz-Allmann et al., 2017; Haffener et al., 2018; Hummel & Shapiro, 2012, 2013; Ingebritsen & Manning, 2010; Segall & Lu, 2015; Shapiro et al., 1997; Shapiro & Dinske, 2009a, 2009b; Yong et al., 2018; Yu et al., 2019) and numerical models (Brown et al., 2017; Catalli et al., 2016; Dempsey & Riffault, 2019; Keranen et al., 2014; Langenbruch & Zoback, 2016; Langenbruch et al., 2018; Pollyea et al., 2019; Riffault et al., 2018), assume a homogeneous injection formation.

An important question that arises is whether the aforementioned equivalence between seismic diffusivity and effective hydraulic diffusivity holds in a heterogeneous porous medium. Fluid flow and pressure diffusion in subsurface formations are influenced by heterogeneity of hydraulic and mechanical properties in natural earth materials; this influence has been investigated in an extensive body of literature over the last 40 years and synthesized in several textbooks (Dagan, 1989; Gelhar, 1993; Zhang, 2001). When heterogeneity is represented as a spatially correlated random field with a well-defined correlation length, it is generally accepted that (at scales much larger than the correlation length) the behavior of the mean pressure and fluid flux fields can be described by effective permeability and hydraulic diffusivity tensors. Similar frameworks exist for defining block-scale effective conductivities for large grid blocks in field-scale flow models based on the underlying heterogeneity structure (Renard & De Marsily, 1997; Sanchez-Vila et al., 2006; Wen & Gómez-Hernández, 1996) or fracture network topology (Botros et al., 2008; Long et al., 1982; R. Zimmerman & Bodvarsson, 1996). The above body of literature provides a basis for representing flow behavior in heterogeneous media in terms of equivalent homogeneous media with effective properties determined by the spatial structure of heterogeneity. Most field-scale modeling of fluid flow in hydrogeology implicitly adopts this view and only represents material property variations across distinct geological units explicitly. Moreover, field-scale hydraulic tests are commonly interpreted based on analytical solutions for fluid flow in homogeneous media and the properties estimated from these tests are assumed to reflect effective properties (such as the effective permeability) at the test scale (Gelhar, 1993). In radial flow, theoretical and computational analyses in heterogeneous media (e.g. Naff (1991) or Guadagnini et al. (2003)) demonstrate that apparent conductivities approach constant values equal to the effective conductivities within a few correlation lengths from the well. Similarly, analytical and numerical modeling studies of IIS and frameworks for interpreting the spatiotemporal patterns of IIS may often be viewed as implicitly adopting equivalent homogeneous representations of hydraulic properties (Brown et al., 2017; Catalli et al., 2016; Dempsey & Riffault, 2019; Keranen et al., 2014; Langenbruch & Zoback, 2016; Langenbruch et al., 2018; Pollyea et al., 2019; Riffault et al., 2018; Segall & Lu, 2015; Shapiro et al., 1997; Shapiro & Müller, 1999; Shapiro & Dinske, 2009a, 2009b; Shapiro, 2015). As mentioned above, these approaches rely on the assumed equivalence between the seismic diffusivity and effective hydraulic diffusivity of the subsurface (Haagenson et al., 2018; Haagenson & Rajaram, 2020).

Our specific goal in this paper is to demonstrate that in highly heterogeneous porous media, there is a clear distinction between the seismic diffusivity and the effective hydraulic diffusivity. In reality, the triggering front is not radially symmetric for cases of heterogeneous domains. This is because pressure increments are expected to propagate preferentially through pathways of relatively high hydraulic diffusivity in heterogeneous media. As a result, the farthest radial distance to which the threshold triggering pressure increment propagates (which we denote as $r_{max}(p = p_t)$ and is synonymous with the triggering front in the context of r-t plots) will be farther than the radial distance at which the directionally averaged pressure increment equals the threshold triggering pressure increment (which we denote as $r(p_{avg} = p_t)$ and propagates according to the effective hydraulic diffusivity). Therefore, the triggering front observed in r-t plots will propagate more rapidly than predicted by the effective hydraulic diffusivity. Put another way, seismicity will propagate more rapidly in a realistic, heterogeneous domain than it would in an equivalent homogeneous domain that employs the effective hydraulic diffusivity, since the latter would inevitably underestimate the farthest radial distance reached by the threshold triggering pressure increment (p_t) at any given time. We investigate this possible distinction, the degree to which these two quantities may differ, and what subsurface conditions may exacerbate the difference.

We investigate these questions using numerical simulations of fluid flow and induced seismicity. After providing a description of the computational framework (in Section 2), we present results of simulated fluid injection and induced seismicity in two different types of heterogeneous domains: smoothly varying fields of random hydraulic diffusivity derived using Sequential Gaussian Simulation (i.e. SGS domains described in Section 2.1) and three-dimensional discrete fracture networks in rock matrix (i.e. DFNM domains described in Section 2.2). In Section 4, we present a case study of the fluid injection experiment near Soultz-sous-Forêts, France from the year 2000, which further illustrates the potential distinction between the seismic and effective hydraulic diffusivity (Dezayes et al., 2010; Dorbath et al., 2009; Genter et al., 2010; Meller & Ledésert, 2017).

2 Computational Framework

Here, we investigate the behavior of fluid flow and IIS in correlated random fields of hydraulic diffusivity as well as a domain of highly fractured rock. The random fields are generated using a Sequential Gaussian Simulation (SGS) algorithm, producing a smoothly varying field of hydraulic diffusivity (which are a common but relatively simple representation of heterogeneity in the subsurface) (Müller & Schüler, 2020). We refer to the domain of highly fractured rock as a discrete fracture network and matrix (DFNM) domain, because we explicitly model fluid flow in both the fracture network and surrounding rock matrix (Birdsell et al., 2018; Haagenson et al., 2018; Haagenson & Rajaram, 2020). This is a more realistic representation of the kind of formation in which IIS typically occurs.

For simplicity, we consider only heterogeneous media with large-scale, isotropic effective permeability and diffusivity tensors (i.e. statistically isotropic correlation structures in the SGS domains and uniform orientation distributions for fractures in the DFNM domains). These models are readily generalizable to represent large-scale anisotropy resulting from either anisotropic spatial correlation in the SGS domains or non-uniform orientation distributions (e.g. families of similarly aligned fractures) in the DFNM domains. However, our purpose here is mainly to illustrate the influence of heterogeneity on the propagation of the triggering front, which is expected to occur with or without anisotropy.

We model fluid flow through a porous medium in response to a point source of fluid injection based on the classical pressure diffusion equation (Bear, 1972; Charbeneau, 2006; Cleary, 1977; De Marsily, 1986; Rice & Cleary, 1976; Rudnicki, 1986; Verruijt, 2013), given as

$$\rho_f \left(\phi\beta_f + \beta_m\right) \frac{\partial p}{\partial t} - \nabla \cdot \left(\frac{\rho_f \kappa}{\mu} \nabla p\right) = Q_m \tag{1}$$

where ρ_f is the fluid density, ϕ is the porosity of the porous medium, β_f and β_m are the compressibilities of the fluid and porous medium respectively, p is the increment of pore fluid pressure (with respect to the initial, static pressure distribution), κ is the intrinsic permeability of the porous medium, μ is the dynamic viscosity of the fluid, and Q_m is the source or sink of fluid mass. Although some of these parameters (particularly permeability) can exhibit pressure dependence, we use constant parameter values such that equation (1) becomes a linear diffusion equation. Our goal is to highlight the influence of heterogeneity on patterns of IIS within the framework of linear diffusion. We intend to investigate the nuanced behavior of nonlinear fluid flow and IIS in future work. In this form, equation (1) represents a diffusion equation of pressure increment, where the well-known hydraulic diffusivity (Bear, 1972; De Marsily, 1986) is defined as

$$D_h = \frac{\kappa}{\mu \left(\phi\beta_f + \beta_m\right)}. \tag{2}$$

Flow through porous media (either in the SGS domain or in the rock matrix portion of the DFNM domain) is governed by equation (1). For the fractures, the model employs an alternative form of the general fracture flow equation:

$$\frac{\partial \left(\rho_f b\right)}{\partial t} - \nabla \cdot \left(\frac{\rho_f b^3}{12\mu} \nabla p\right) = Q_f - L_m \tag{3}$$

where b is the fracture aperture, Q_f is the source or sink of fluid mass per unit area of the fracture and L_m is the fluid leak-off rate per unit area from the fracture into the surrounding rock matrix (Chaudhuri et al., 2013; Murphy et al., 2004). Equation (3) employs the well-known local cubic law for fracture transmissivity (R. W. Zimmerman & Bodvarsson, 1996). Although there are limitations to the local cubic law at smaller scales, it is widely used in DFN models (Adler et al., 2013; Frampton et al., 2019; J. Hyman et al., 2015) with a constant aperture (b) within individual fractures (which may also be interpreted as the equivalent hydraulic aperture). In Section 2.2.2 below, we further describe how equation 3 can be expressed in the form of a diffusion equation and solved in combination with the pressure diffusion equation in the rock matrix given in equation (1).

Table 1. Summary of fluid injection model parameters.

Parameter	Description	Value	Unit
ρ_f	fluid density	998	kg/m^3
μ	fluid viscosity	$8.9 \cdot 10^{-4}$	Pa $\cdot$ s
β_f	fluid compressibility	$4.4 \cdot 10^{-10}$	Pa^{-1}
H_{inj}	depth of injection location	4,000	m
Q	injection rate of fluid	25	m^3/hr
t_e	injection period	12	hours
H_{wt}	depth of water table	100	m
ρ_s	density of overburden	2,300	kg/m^3

For simulations of fluid injection, we have developed a numerical model to solve equations (1) and (3) using FEniCS – a general-purpose, open-source finite element method (FEM) software (Alnæs et al., 2015), which has been previously applied to a wide range of geoscientific problems including subsurface fluid flow and generalized poroelasticity (Haagenson et al., 2020). Each domain used in this study is cubic, with sides measuring two kilometers in length. Fluid is injected at the center vertex of the numerical mesh (as a point source of fluid injection) at a constant rate (Q) of 25 cubic meters per hour for a period (t_e) of 12 hours. These parameters were selected to reflect a realistic scenario of fluid injection (Shapiro et al., 1997, 2005; Shapiro & Dinske, 2009b), while also minimizing potential boundary effects by employing a sufficiently large domain (with domain boundaries located a distance of at least $12\sqrt{D_{eff}t_e}$ away from the injection location, where D_{eff} is the effective hydraulic diffusivity of the heterogeneous domain). The initial condition and all boundary conditions are set to a hydrostatic pressure field. A summary of the parameters used in the numerical simulations (which apply to simulations of both the SGS and DFNM domains) are given in Table 1.

Following the approach of Rinaldi and Nespoli (2017), Catalli et al. (2016), and Shapiro (2015), we track seismicity using a set of weak points, which are seeded randomly within one kilometer of the injection location. Each weak point represents a potential location for seismicity. In the SGS domain, each weak point is randomly assigned a strike and dip angle (to represent a hypothetical fracture at that location), whereas the weak points in the DFNM domain are seeded exclusively along fractures and are assigned the corresponding fracture's strike and dip angle. The weak point triggering pressure increment (p_t^{wp}), which is the increase in fluid pressure above the initial fluid pressure that would trigger a seismic event, is found using the well-known Mohr-Coulomb failure criterion:

$$f\left(\sigma_n - (p_t^{wp} + p_i)\right) - |\tau| \leq 0 \tag{4}$$

where f is the coefficient of static friction of the fracture, p_i is the initial fluid pressure at the weak point, and σ_n and τ are the normal compressive stress and shear stress acting on the fracture respectively. The coefficient of friction is allowed to vary between 0.6 and 0.7 to capture a realistic range of values (Jaeger, 1959; Talwani & Acree, 1985). The initial pressure profile is assumed to be hydrostatic, and is found using a constant fluid density ($\rho_f = 998$ kg/m^3) and a water table located 100 meters below the ground surface. The depth of the fluid injection (H_{inj}) and is stipulated as four kilometers. The stresses σ_n and τ can be found using the expressions

$$\sigma_n = (\boldsymbol{\sigma} \cdot \vec{n}) \cdot \vec{n} \tag{5}$$

and

$$\tau = (\boldsymbol{\sigma} \cdot \vec{n}) \times \vec{n} \tag{6}$$

where $\boldsymbol{\sigma}$ is the local stress tensor and $\vec{n}$ is the normal vector of the fracture plane. We have assumed that the local stress tensor is defined by a normal faulting, lithostatic condition where the vertical stress component of $\boldsymbol{\sigma}$ is defined as $\sigma_v = \rho_s g H_{inj}$ and both horizontal components as $\sigma_h = 0.6\sigma_v$ (Zoback, 2010). The rock density of the overburden (ρ_s) is assumed to be 2,300 kilograms per cubic meter.

Commonly, pressure increments associated with induced seismicity are assumed to lie in the range of 0.01 and 0.1 MPa, where a pressure increment less than 0.01 MPa is either not significant enough to trigger seismicity or is often met by processes other than fluid injection (such as background fluid flow or tidal forcing) (Dempsey & Riffault, 2019; Gischig & Wiemer, 2013; Goebel et al., 2017; Keranen et al., 2014; Shapiro, 2015). Thus, weak points with a triggering pressure increment outside this range are removed from the final set of weak points used in the simulation. It is clear then, that the threshold triggering pressure increment associated with the triggering front will in fact be the minimum of the range of weak point triggering pressure increment values (i.e. $p_t = min(p_t^{wp}) =$ 0.01 MPa).

To track seismicity in the numerical simulations, each weak point is evaluated for potential failure in each time step of the simulation. If the local pressure increment due to fluid injection at the weak point location exceeds the weak point triggering pressure increment (indicating that the weak point fails the Mohr-Coulomb criterion), then the weak point's location is recorded as the hypocenter of a seismic event at that given time. This approach results in a synthetic IIS dataset, which allows us to produce r-t plots in order to investigate the spatiotemporal patterns of seismicity and estimate the seismic diffusivity of the heterogeneous domain (following the approach described later in Section 2.4).

The following sections describe the specifics related to the SGS and DFNM domains, including how each domain was generated and spatially discretized, the set of weak points

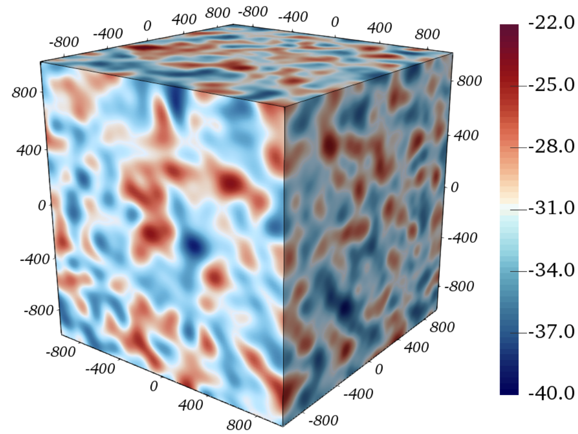

Figure 2. Example of a three-dimensional lnκ field from Sequential Gaussian Simulation (SGS) with isotropic covariance function using Var(lnκ) = 5.3. The permeability (κ) field and thus hydraulic diffusivity fields are readily obtained from the lnκ field.

used in each domain type, and our treatment of fluid flow in the fracture network of the DFNM domain. In Section 2.3, we describe the approach used to estimate the effective hydraulic diffusivity based on pressure diffusion and in Section 2.4, we describe the approach used to estimate the seismic diffusivity from r-t plot analysis.

2.1 SGS Domains

There is a long history of previous work on flow and transport in heterogeneous porous media based on numerical simulations in computer-generated spatially correlated random fields (Deutsch & Journel, 1998; Tompson et al., 1989). To generate such a field for hydraulic diffusivity, we first generate a logarithm of permeability field (lnκ, where κ has units of m^2) with an isotropic, Gaussian covariance function using the GeoStat-Framework (i.e. GSTools module in Python) (Müller & Schüler, 2020). We simulate ten realizations of the SGS domain at four different values of the variance of lnκ (Var(lnκ) = 0.53, 1.325, 2.65 and 5.3), for a total of 40 SGS domain realizations. Note that a value of zero for Var(lnκ) entails a perfectly homogeneous domain.

The SGS domains are discretized using a structured, tetrahedral mesh with vertices uniformly spaced at 20 meters apart. The autocorrelation length of the SGS covariance function is 100 meters (five times larger than the grid spacing of the numerical mesh) and the sides of the cubic domain are two kilometers (20 times larger than the autocorrelation length). An example of the resulting lnκ field is shown in Figure 2. A correlated, random field of permeability (κ) for use in equations (1) or (2) is readily obtained by taking the exponential of the lnκ field.

After generating the field of permeability, the correlated, random field of hydraulic diffusivity is found using the well established definition of hydraulic diffusivity given in

Table 2. Summary of parameters for SGS domains.

Parameter	Description	Value or Range	Unit
E(lnκ)	mean value of lnκ	-31.0	–
Var(lnκ)	variance of lnκ	0.53 - 5.3	–
ϕ	porosity	0.05	–
β_m	compressibility of porous medium	$7.0 \cdot 10^{-11}$	Pa^{-1}
$D_{h,SGS}$	local hydraulic diffusivity	$5.2 \cdot 10^{-5}$ - $3.4 \cdot 10^{3}$	m/s^2

equation (2). When evaluating this expression, we have assumed typical, constant values for μ ($8.9 \cdot 10^{-4}$ Pa $\cdot$ s), ϕ (0.05), β_f ($4.4 \cdot 10^{-10}$ Pa^{-1}), and β_m ($7.0 \cdot 10^{-11}$ Pa^{-1}) that are reflective of a highly fractured, low permeability rock in which IIS most often occurs (Bear, 1972; Charbeneau, 2006; De Marsily, 1986; Freeze & Cherry, 1987). Note that permeability can vary spatially by several orders of magnitude within a heterogeneous domain, whereas porosity varies within a relatively narrow range in a given formation (Gelhar, 1993). Therefore, using a spatially variable permeability along with a uniform porosity is an acceptable approximation that is common in studies of fluid flow in heterogeneous porous media. The field of spatially variable permeability along with uniform values of all other parameters in equation (2) produces a random, spatially correlated field of hydraulic diffusivity, as desired. A summary of the model parameters specific to the SGS domains are given in Table 2. As mentioned previously, the weak points are seeded throughout the SGS domains and assigned a hypothetical fracture orientation. The final set of weak points for the SGS domains is shown in Figure 3, colored based on the weak point triggering pressure increment (p_t^{wp}).

2.2 DFNM Domains

In a domain of highly fractured, low permeability rock, the fracture network represents a well-connected pathway of relatively high hydraulic diffusivity through which pressure increments can propagate rapidly. Since most cases of IIS occur in fractured rock, it is critical to consider models of subsurface heterogeneity that explicitly represent these connected fracture networks. In the following sections, we describe the methodology specifically employed for the fractured rock domain.

2.2.1 Fracture Network and Numerical Mesh Generation

The DFNM domain is fully three-dimensional, with the fracture network represented by penny-shaped fractures of finite thickness embedded within the rock matrix. This methodology has been previous presented (Birdsell et al., 2018; Haagenson et al., 2018; Haagenson & Rajaram, 2020) and is also referred to as an upscaled discrete fracture matrix model (UDFM) by Sweeney et al. (2020). The primary benefits of this approach, as opposed to modeling the fracture network alone, is that we can directly account for fluid flow in

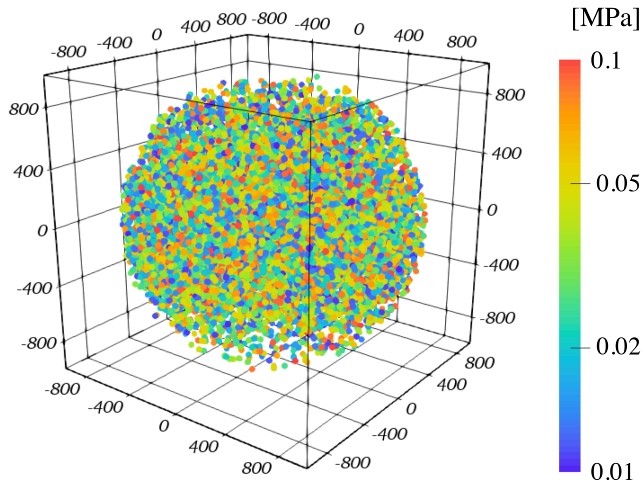

Figure 3. The locations of the weak points for the SGS domains colored by their respective triggering pressure increment p_t^{wp}. Note that weak points are randomly distributed within one kilometer of the injection location and have triggering pressure increment values limited to the range of 0.01 and 0.1 MPa.

the rock matrix (including leak-off from the fracture network) while keeping the mathematical and numerical formulations relatively simple.

The discrete fracture network (DFN) is initially generated as a three-dimensional network of two-dimensional fractures using dfnWorks – a fracture network generator and flow solver developed at Los Alamos National Laboratory (J. D. Hyman et al., 2015). The DFN is generated using a truncated power law distribution for fracture radius (ranging between 200 and 400 meters) and a uniform distribution for fracture orientation. The fracture apertures (b) are linearly related to the fracture radii and range between 0.2 and 0.5 millimeters. The DFN used in this study was chosen for illustrative purposes, and the approach can readily be extended to include alternative fracture network characteristics (such as fracture families with preferred orientations). The three-dimensional DFN used in this study is presented in Figure 4(a).

To build the final three-dimensional mesh, the DFN is overlaid onto a structured tetrahedral mesh. Mesh cells overlapped by a fracture are then recursively refined using an octree-based mesh refinement method (Sweeney et al., 2020). The number of refinement steps is stipulated, with more steps leading to finer mesh resolution near the mesh elements that represent the DFN. This process is depicted in Figure 4(b). The final result is a three-dimensional tetrahedral mesh containing cells to represent the fractures and the rock matrix separately. The octree-based mesh refinement method produces a DFNM domain that accurately reflects the topology of the original DFN, although with "staircase" shaped fractures as a secondary feature. A portion of the full mesh is

Table 3. Summary of parameters for DFNM domain.

Parameter	Description	Value or Range	Unit
κ	permeability of rock matrix	10^{-17}	m^2
ϕ	porosity of rock matrix	0.05	–
β_m	compressibility of rock matrix	$2.0 \cdot 10^{-10}$	Pa^{-1}
$D_{h,m}$	hydraulic diffusivity of rock matrix	$5.0 \cdot 10^{-5}$	m^2/s
b	fracture aperture	0.2-0.5	mm
K_n	normal stiffness of fractures	$3.0 \cdot 10^{-11}$	Pa/m
$D_{h,frac}$	hydraulic diffusivity of fractures	$2.2 \cdot 10^2 - 3.3 \cdot 10^3$	m^2/s

shown in Figure 4(c) and the final DFN after the octree-based mesh refinement method in Figure 4(d). For the DFNM domain in this study, the initial structured mesh is a cube with side lengths of two kilometers and vertices spaced uniformly at 100 meters apart. There are four iterations of octree-based mesh refinement around the fracture cells, leading to a final model fracture cell width of 6.25 meters (which we refer to as b_p) and a high degree of mesh refinement around the fracture network. Clearly, in a field-scale model, fracture cell widths cannot be refined down to the true fracture aperture (i.e. b), which is typically in the sub-millimeter range. Interface finite element formulations can handle narrow fractures treated as discontinuities (Abushaikha et al., 2015; Berre et al., 2019; Fumagalli & Scotti, 2013; Geiger et al., 2010; Odsæter et al., 2019); however, these approaches typically require highly specialized numerical methods beyond the current capabilities of most general-purpose FEM software (e.g. FEniCS) and limit the domain to either two dimensions or simple DFN geometries in three dimensions (Sweeney et al., 2020). Alternatively, a practically reasonable model fracture width (b_p) that is much larger than the true fracture aperture (b) may be used, so long as the model representation of the fracture is hydraulically equivalent to the true fracture and leak-off fluxes from the fracture to the adjacent rock matrix are calculated accurately. Such an approach has been employed in previous work (Birdsell et al., 2015, 2018; Bower & Zyvoloski, 1997; Chaudhuri et al., 2013; Pandey & Rajaram, 2016; Pandey et al., 2017; Sweeney et al., 2020). In Section 2.2.2, we further discuss the hydraulic equivalence between the model representation of fractures and the true fracture. A summary of the model parameters specific to the DFNM domain is presented in Table 3.

The zones of relatively high hydraulic diffusivity in the SGS realizations with $\mathrm{Var}(\ln\kappa) = 5.3$ (i.e. the largest variance considered) have hydraulic diffusivity values comparable to those found in the fractures of the DFNM domain. The major difference between the domains is that zones of high hydraulic diffusivity are well-connected in the DFNM domain. To illustrate this, we plot the distribution of hydraulic diffusivity within the fracture network of the DFNM domain in Figure 5(b), and compare it with the locations of equivalently high hydraulic diffusivity in an example of the SGS domains (with $\mathrm{Var}(\ln\kappa) =$

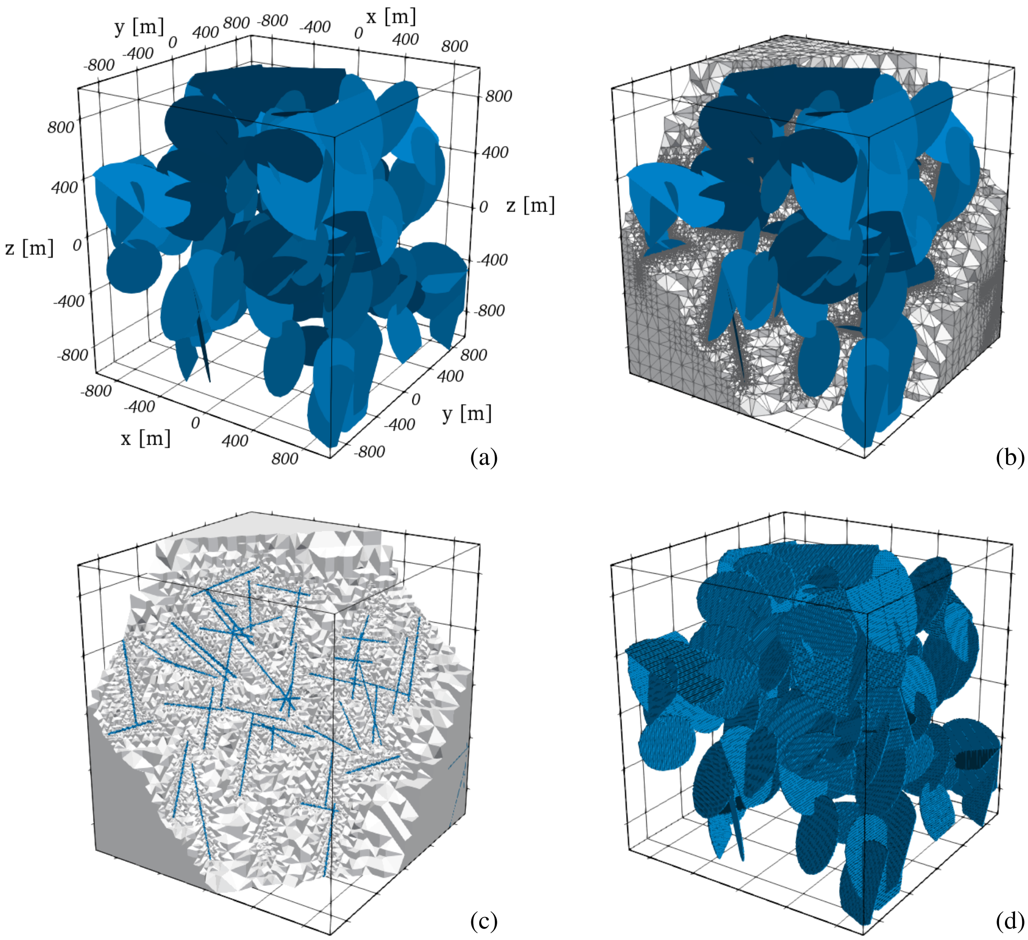

Figure 4. A visualization of how the DFNM numerical mesh is generated. The matrix cells are shown in white and the fracture cells are shown in blue. (a) The initial DFN of two-dimensional fractures generated using dfnWorks. (b) The original DFN overlaid onto the final three-dimensional, tetrahedral mesh. The cell edges are highlighted in dark grey to clearly visualize the high degree of mesh refinement near the fracture cells. (c) A portion of the final three-dimensional mesh, showing locations of the fracture cells within numerical mesh. (d) The final DFN of the three-dimensional mesh.

Figure 5. A visual comparison of the spatial distribution of the upper range of hydraulic diffusivity values (i.e. values of hydraulic diffusivity found in the DFN) for (a) the SGS domain and (b) the DFNM domain. Note that relatively high hydraulic diffusivity values are exclusively found in the fracture network for the DFNM case, which creates well-connected pathways throughout domain.

5.3) in Figure 5(a). Notice that zones of relatively high hydraulic diffusivity are not well-connected in the SGS case, whereas they are in the DFNM domain. We expect that the enhanced connectivity in the DFNM domain will cause pressure increments to diffuse rapidly within the fracture network, leading to more rapid propagation of seismicity compared to the SGS domains.

2.2.2 Flow and Seismicity in a Fracture Network

Fluid flow in the rock matrix portion of the DFNM domain is governed by equation (1). Here, we discuss the treatment of equation (3) for modeling fluid flow in the fracture network. The DFNM domain contains both the fracture network and rock matrix (as discussed in Section 2.2.1), allowing us to drop L_m from equation (3) as leak-off from the fracture network is automatically considered by allowing fluxes across fracture-matrix interfaces. The storage term in equation (3) may be expanded following the approach of Rutqvist et al. (1998), giving

$$\rho_f \left(\rho_f b \beta + 1/K_n \right) \frac{\partial p}{\partial t} - \nabla \cdot \left(\frac{\rho_f b^3}{12 \mu} \nabla p \right) = Q_f \qquad (7)$$

where K_n is the normal stiffness of the fracture, which can be measured using laboratory experiments (Bandis et al., 1983). Following the approach of Chaudhuri et al. (2013) and others (Birdsell et al., 2015, 2018; Bower & Zyvoloski, 1997; Pandey & Rajaram, 2016; Pandey et al., 2017), we define the permeability of the fractures as ure, as we see in Figure 13(f). These are examples of the more subtle influences of fracture networks on the behavior of IIS.

If we look at the synthetic dataset of seismicity from both simulations as an r-t plot (shown in Figure 14), we clearly see very different behavior between the equivalent homogeneous and DFNM domains. As expected, seismicity has propagated much farther in the DFNM domain than in the equivalent homogeneous domain. This suggests that equivalent homogeneous domains (which are commonly used in studies of IIS (Brown et al., 2017; Catalli et al., 2016; Dempsey & Riffault, 2019; Keranen et al., 2014; Langenbruch & Zoback, 2016; Langenbruch et al., 2018; Pollyea et al., 2019; Riffault et al., 2018)) may be limited in their ability to replicate the spatiotemporal patterns of IIS as observed in r-t plots from real-world or heterogeneous domains. Next, we examine the behavior of the triggering front in each domain, using equation (13) as described in Section 2.4. As expected, the triggering front in the equivalent homogeneous domain is well described using the effective hydraulic diffusivity. In this case, the triggering front is radially symmetric and thus there should not be any distinction between the seismic diffusivity and the effective hydraulic diffusivity. In the DFNM domain, however, we see that the location of the triggering front is vastly underestimated by equation (13) using the effective hydraulic diffusivity of the domain, as was the case in the highly heterogeneous SGS domains. This again suggests that the seismic diffusivity associated with the diffusive propagation of the triggering front (which was found to be 3.0 m^2/s in the DFNM domain) is larger than the effective hydraulic diffusivity in heterogeneous porous media (which was found to be 0.29 m^2/s in the DFNM domain).

Finally, we consider the ratio of D_s/D_{eff} as we did in the study of SGS domains (see Figure 11(c) for the comparison of this ratio from the SGS domains), which produces a value of 10.2 in the DFNM domain. This shows that in the case of fractured rock, the distinction between the seismic and effective hydraulic diffusivity can be up to one order of magnitude. Moreover, this is considerably larger than the largest ratio found in the SGS cases (i.e. $D_s/D_{eff} = 3.4$), which indicates that the well-connected pathways of the fracture network allow pressure increments to propagate more rapidly in the DFNM domain than in any of the SGS domains considered in this study.

To be thorough, we also performed a simple sensitivity analysis to gain insight into the potential uncertainty associated with the estimate of seismic diffusivity of the DFNM domain. Figure 15 shows the location of the triggering front according to equation (13) using $D_s = 3.0$ m^2/s along with various other values of seismic diffusivity: $0.1D_s$, $0.5D_s$, $2D_s$, and $10D_s$. It is clear from Figure 15 that even relatively small variations in the seismic diffusivity (i.e. $0.5D_s$ and $2D_s$) alter the behavior of the triggering front enough that it no longer fits well to the upper envelope of the seismic data cluster. This suggests that we can be reasonably confident in the values of seismic diffusivity estimated with the triggering front plotted in Figure 14(b).

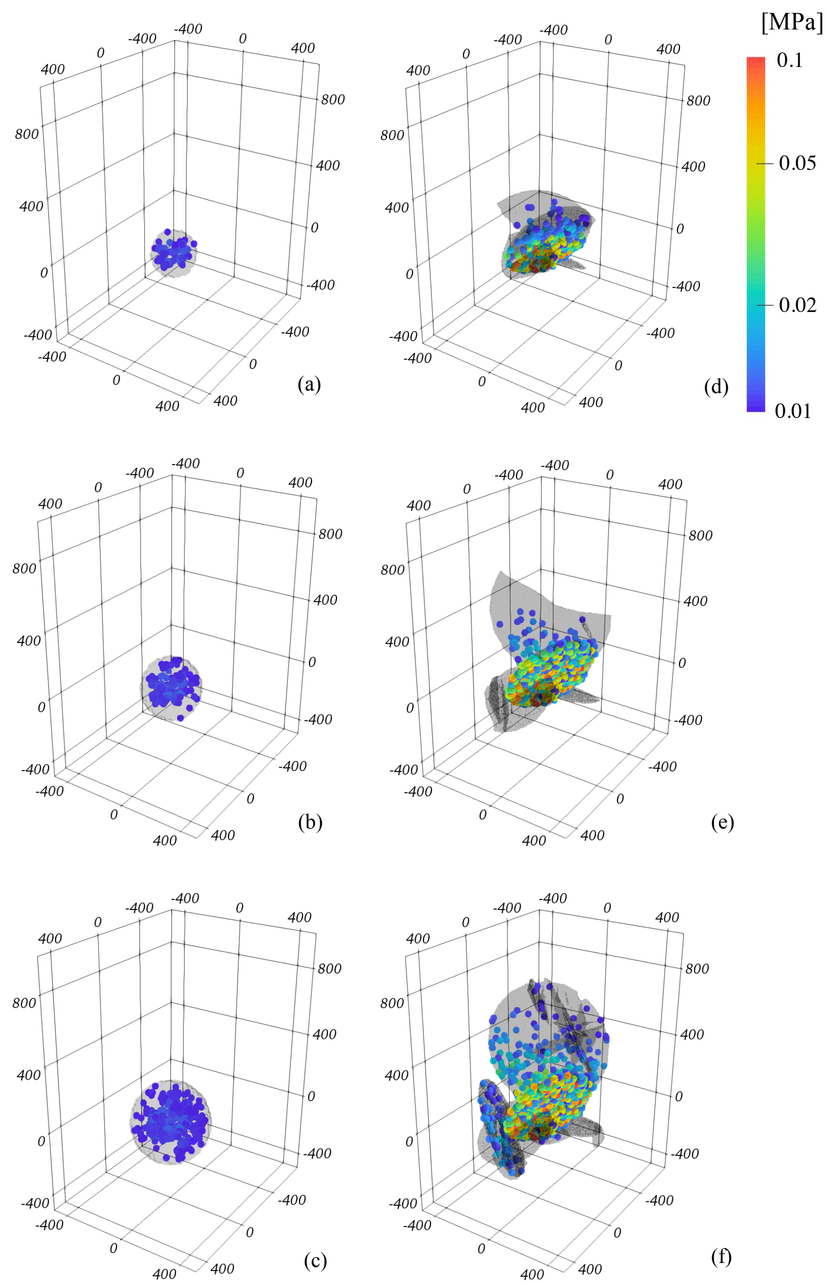

Figure 13. Plots of seismicity during the injection simulation at three different times during the injection simulations ($t = 3$, 6 and 12 hours) for: (a) - (c) the homogeneous domain with uniform hydraulic diffusivity equal to the effective hydraulic diffusivity of the DFNM domain and (d) - (f) the DFNM domain. The seismic events are colored based the weak point triggering pressure increment (p_t^{wp}). The spatial extent to which threshold triggering pressure increment (p_t) extends (which corresponds to the plotted portions in Figure 12) is shaded in dark grey.

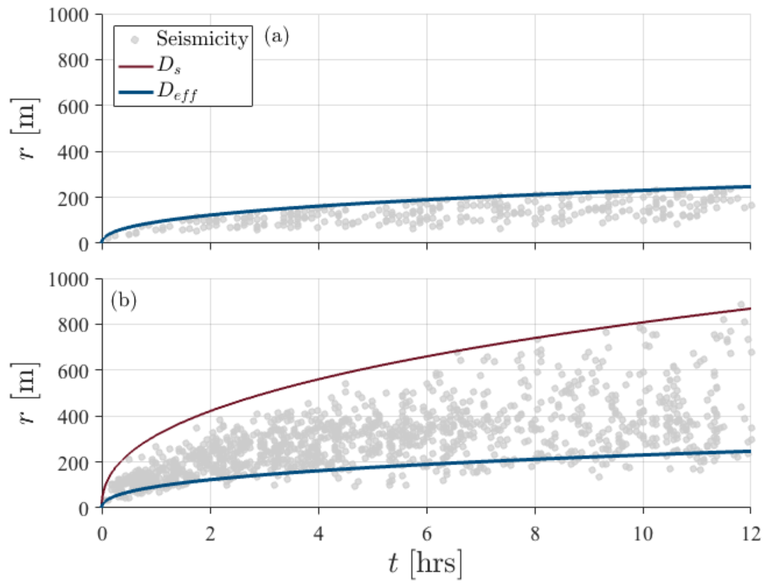

Figure 14. An r-t plot using the synthetic dataset of seismicity from the fluid injection simulation for (a) the homogeneous domain with uniform hydraulic diffusivity equal to the effective hydraulic diffusivity of the DFNM domain and (b) the DFNM domain. Each r-t plot is overlaid with equation (13) using the effective hydraulic diffusivity of the DFNM domain (D_{eff}). Equation (13) is also plotted with the fitted estimate of seismic diffusivity (D_s) of the DFNM domain.

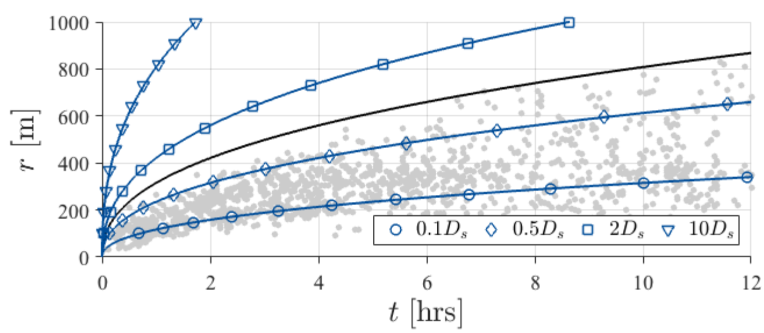

Figure 15. An r-t plot using the synthetic dataset of seismicity from the fluid injection simulation for the DFNM domain, overlaid with equation (13) using the estimated value of seismic diffusivity (D_s) as well as $0.1D_s$, $0.5D_s$, $2D_s$, and $10D_s$ to illustrate the sensitivity of the triggering front to the estimate of the seismic diffusivity.

4 Case Study: Soultz-sous-Forêts

Numerous studies of IIS have previously indicated that tracking the location of the triggering front as observed in r-t plots may produce reasonable estimates of the effective hydraulic diffusivity at the injection site, mostly using the approach described by Shapiro et al. (1997) (Antonioli et al., 2005; Chen et al., 2012; Delepine et al., 2004; Goebel et al., 2017; Goertz-Allmann et al., 2017; Haffener et al., 2018; Hummel & Müller, 2009; Hummel & Shapiro, 2012, 2013; Ingebritsen & Manning, 2010; Improta et al., 2015; Rothert & Shapiro, 2003; Segall & Lu, 2015; Shapiro & Müller, 1999; Shapiro et al., 2002; Shapiro & Dinske, 2009a, 2009b; Yong et al., 2018; Yu et al., 2019). Although these previous studies have implicitly assumed that the seismic diffusivity is essentially equivalent to the effective hydraulic diffusivity, the numerical study presented here has clearly shown that these two quantities are likely distinct in heterogeneous domains. To further illustrate this potential distinction, we analyze the seismic dataset collected during the fluid injection experiment near Soultz-sous-Forêts, France in the year 2000.

The lower reservoir of the enhanced geothermal system (EGS) site near Soultz-sous-Forêts, France consists of three wells (called GPK2, GPK3, and GPK4) located in a highly fractured, crystalline rock formation approximately 4,000 - 5,000 meters below the ground surface (Dezayes et al., 2010; Genter et al., 2010; Meller & Ledésert, 2017). The site was selected for EGS due to the presence of a thermal anomaly, with downhole temperatures reaching 200° C in each of the three wells (Meller & Ledésert, 2017). Hydraulic stimulation was performed at each well at different times during the construction of the EGS site: GPK2 in 2000, GPK3 in 2003, and GPK4 in 2004 and 2005 (Dorbath et al., 2009; Meller & Ledésert, 2017). A full description of the site geology and wellbore details are given by Meller and Ledésert (2017) and Dezayes et al. (2010). Fluid injection and seismicity data have been presented by Genter et al. (2010) and Cuenot et al. (2008), in addition to being freely available through the EPOS-IP Anthropogenic Hazards data repository (Leptokaropoulos et al., 2019). Here, we present a case study of the fluid injection experiment performed at the GPK2 well in the year 2000.

Water was injected at an average rate of 44.6 liters per second (see Figure 16(a)) for a period of approximately six days. The screened interval of the injection well ranged from 4,400 to 5,100 meters below the ground surface. Using numerous surface and downhole seismometers, seismicity was tracked in space and time during the fluid injection experiment, which reached up to one kilometer away from the injection location. During just the first day of fluid injection, seismicity occurred as far as 700 meters away from the injection location. The r-t plot of this seismic dataset was shown earlier in Figure 1.

In order to characterize the hydraulic diffusivity of the EGS reservoir, Delepine et al. (2004) previously studied the spatiotemporal patterns of IIS found in this seismic dataset using the expression

$$r_t = \sqrt{4\pi D t} \tag{14}$$

which was originally suggested by Shapiro et al. (1997). By fitting equation (14) to the upper envelope of the seismic data cluster shown in Figure 16, the authors estimated the seismic diffusivity to be 0.15 m^2/s. (Note that the authors did not use the term "seismic diffusivity", but instead assume that the spatiotemporal patterns of IIS and the propagation of the triggering front is associated with the effective hydraulic diffusivity of the subsurface.) Equation (14) suggests that the triggering front propagates with $\sqrt{t}$ (Shapiro et al., 1997; Shapiro & Müller, 1999; Shapiro & Dinske, 2009b; Shapiro, 2015). This behavior is commonly found in IIS datasets and corresponds to a two-dimensional, diffusional process. The upper envelope of the seismic data cluster in Figure 16, though, appears to indicate a different behavior in the propagation of the triggering front. More complex triggering front behavior has been previously explained using nonlinear diffusion in two dimensions (Hummel & Shapiro, 2012, 2013; Shapiro & Dinske, 2009a, 2009b). While nonlinear diffusion may help to explain the triggering front behavior observed in this dataset (and its apparent deviation from the common $r_t \propto \sqrt{t}$ relationship), we suggest that the triggering front propagation is more appropriately described using a three-dimensional diffusion solution given by equation (13). The underlying porous medium is in fact three-dimensional and the diffusive propagation of the triggering front according to equation (13) does not necessarily conform to the $r_t \propto \sqrt{t}$ relationship and may therefore provide a better fit with the Soultz-sous-Forêts dataset. When fitting equation (13) to the upper envelope of the seismic data cluster (shown in Figure 16(b)), we assumed the fluid viscosity to be $2.0 \cdot 10^{-4}$ $\mathrm{Pa} \cdot \mathrm{s}$ to reflect water at approximately 200° C and the threshold triggering pressure increment to be 0.04 MPa, which is the minimum value used by Keranen et al. (2014) and well within the range suggested by others (Dempsey & Riffault, 2019; Gischig & Wiemer, 2013; Goebel et al., 2017; Shapiro, 2015).

The triggering front according to equation (13) fits the upper envelope of the seismic data cluster more accurately than equation (14). Notice that the rapid spread of seismicity during early times is captured well by equation (13) while underestimated by equation (14). The propagation rate of the triggering front at late time appears to be properly estimated by equation (13) as well. To evaluate the fit of each expression, we calculated both NRMSE and the NSE for equations (13) and (14). The reference dataset for the location of the triggering front was found by first binning the data into time periods of 0.1 days and then using the 10% of seismic events that occurred farthest away from the injection location within each time period. This set of seismic events are highlighted in Figure 16(b) in dark grey for reference. The NRMSE for equations (13) and (14) are 0.05 and 0.09 respectively. The NSE for equations (13) and (14) are 0.56 and -0.42 respectively. This indicates that equation (13) is a considerably better fit to the observed data than equation (14), and may therefore describe the propagation of the trig-

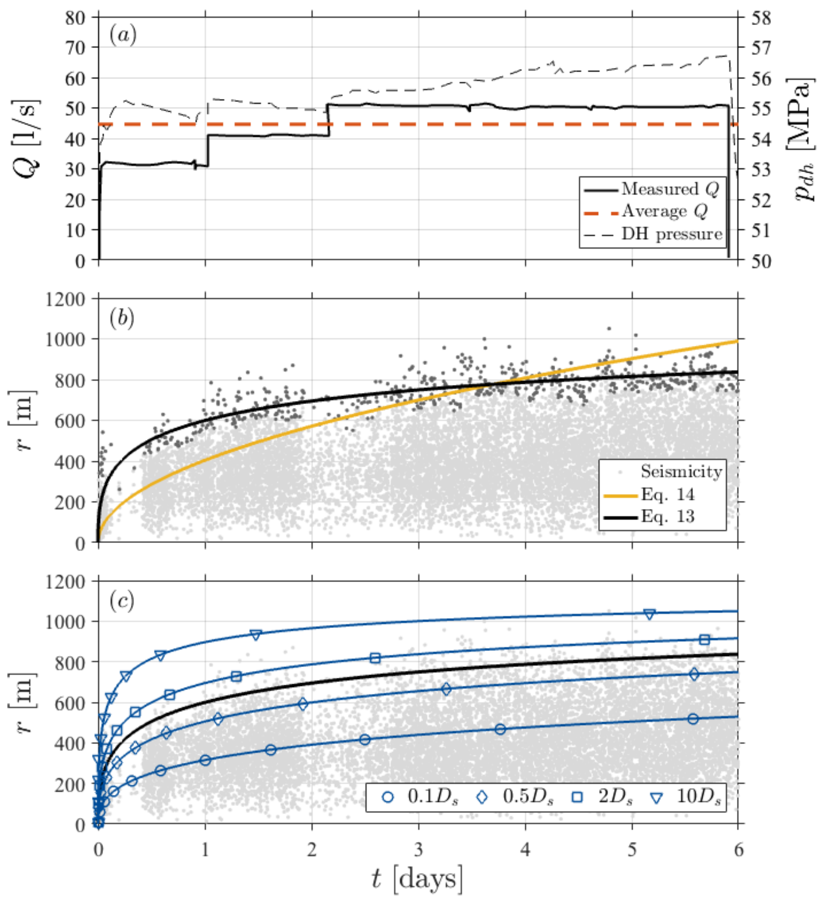

Figure 16. (a) Injection rate (Q) and downhole injection pressure (p_{dh}) from the fluid injection experiment near Soultz-sous-Forêts, France, with the calculated average injection rate overlaid. (b) The r-t plot for the fluid injection experiment. The seismic events are shown as light grey dots, with the farthest 10% of seismic events that occurred in each 0.1 day period plotted in dark grey (which were used to evaluate the fit of equation (13)). The seismic dataset has been fit with two expressions for the location of triggering front: (yellow) equation (14) using the diffusivity value of 0.15 m^2/s (Delepine et al., 2004) giving a normalized-root-mean-square-error of 0.09 and a Nash-Sutcliffe model efficiency coefficient of -0.42 and (black) equation (13) using the diffusivity value of 4.6 m^2/s giving a normalized-root-mean-square-error of 0.05 and a Nash-Sutcliffe model efficiency coefficient fo 0.56. (c) The r-t plot is overlaid with equation (13) using the estimated value of seismic diffusivity (D_s) as well as 0.1D_s, 0.5D_s, 2D_s, and 10D_s to illustrate the degree of uncertainty in the estimate of seismic diffusivity using equation (13).

gering front more accurately and produce a more reliable estimate of seismic diffusivity.

For the Soultz-sous-Forêts dataset, the best estimate of seismic diffusivity is found to be 4.6 m^2/s according to the fit of equation (13) shown in Figure 16(b). To evaluate the potential uncertainty related to this estimate of seismic diffusivity, we plot the location of the triggering front according equation (13) using the estimated value of seismic diffusivity (D_s) as well as $0.1D_s$, $0.5D_s$, $2D_s$, and $10D_s$ in Figure 16(c). Visual inspection indicates that the triggering front fits the upper envelope of the seismic data cluster using seismic diffusivity values ranging between $0.5D_s$ and $2D_s$ (i.e. 2.3 m^2/s and 9.2 m^2/s). Note that the estimated value of seismic diffusivity depends on the set of seismic events selected when fitting the location of the triggering front. Here, we have elected to use the 10% of events that occurred the farthest away from the injection location within each time period. Of course, an alternative set (such as the farthest 5% or even 1% of seismic events) would produce different estimates of the seismic diffusivity, where sets of seismic events located farther from the injection location would naturally produce larger estimates of seismic diffusivity. Our final estimate of 4.6 m^2/s, then, represents a conservatively low estimate of the seismic diffusivity.

Other studies have attempted to characterize the subsurface in the vicinity of the GPK2 well (producing estimates of effective permeability) using various approaches including a laboratory experiment on core samples (Hettkamp et al., 1998), a pre-stimulation slug test (Weidler, 2001), an inverse modeling study using downhole pressure and injection rate data (McClure & Horne, 2011), a circulation test between wells GPK2 and GPK3 (Ledésert & Hébert, 2012), and a tracer experiment between wells GPK2, GPK3 and GPK4 (Vogt et al., 2012). To compare our value of seismic diffusivity to these studies, we first convert the estimates of effective permeability to effective hydraulic diffusivity using equation (2). For this, we assume that viscosity of the injected water at 200° C is $\mu = 2.0 \cdot 10^{-4}$ Pa $\cdot$ s, the porosity of the fractured, crystalline rock formation is $\phi = 0.05$ (Freeze & Cherry, 1987), and the typical value for the compressibility of water is $\beta_f = 4.4 \cdot 10^{-10}$ Pa^{-1}. For the bulk compressibility of the fractured crystalline rock (β_m), Vogt et al. (2012) employs a value of 10^{-8} Pa^{-1} while Freeze and Cherry (1987) report typical values ranging between 10^{-10} Pa^{-1} and 10^{-8} Pa^{-1}, where values less than 10^{-10} Pa^{-1} pertain to intact crystalline rock. Here, we use a value of 10^{-8} Pa^{-1} when converting to effective hydraulic diffusivity for the estimate associated with Vogt et al. (2012). For the other studies, we estimate a range of effective hydraulic diffusivity using the end member values of bulk compressibility reported by Freeze and Cherry (1987). The resulting estimates of effective hydraulic diffusivity from each study are shown in Table 4.

Comparing the value of seismic diffusivity found in this study (i.e. 4.6 m^2/s) to the values of effective hydraulic diffusivity given in Table 4, it is clear that the seismic diffusivity is one to five orders of magnitude larger than the effective hydraulic diffusivity. This case study thus provides potential empirical evidence suggesting that the seismic diffusivity associated with spatiotemporal patterns of IIS is distinct from the effective

Table 4. Estimates of the hydraulic diffusivity from various previous studies. Ranges shown here account for the broad range in the bulk compressibility of fractured rock (β_m) (Freeze & Cherry, 1987).

Study	Effective Hydraulic Diffusivity [m^2/s]
Hettkamp et al. (1998)	$1.9 \cdot 10^{-4}$ - $1.5 \cdot 10^{-2}$
Weidler (2001)	$2.5 \cdot 10^{-5}$ - $5.3 \cdot 10^{-3}$
McClure and Horne (2011)	$5.0 \cdot 10^{-3}$ - $4.1 \cdot 10^{-1}$
Ledésert and Hébert (2012)	$5.0 \cdot 10^{-4}$ - $4.1 \cdot 10^{-1}$
Vogt et al. (2012)	$5.0 \cdot 10^{-3}$

hydraulic diffusivity in heterogeneous porous media, such as the highly fractured, crystalline rock formation at the Soultz-sous-Forêts EGS site.

5 Conclusions and Discussion

In this paper, we analyzed the difference between the diffusivities associated with the propagation of seismicity (i.e. the seismic diffusivity) versus mean pressure diffusion (i.e. the effective hydraulic diffusivity) in heterogeneous subsurface formations. Often, the spatiotemporal patterns of IIS are interpreted with r-t plots and the concept of a triggering front, which has been previously shown to propagate in a diffusion-like manner with an associated diffusivity parameter (Segall & Lu, 2015; Shapiro et al., 1997; Shapiro, 2015). Here, we refer to this diffusivity as the "seismic diffusivity" following the terminology of Talwani and Acree (1985). If we assume that the onset of seismicity is induced by a threshold triggering pressure increment (p_t) (Dempsey & Riffault, 2019; Gischig & Wiemer, 2013; Goebel et al., 2017; Keranen et al., 2014; Shapiro, 2015), then the seismic diffusivity is associated with the spatiotemporal evolution of the farthest radial distance to which the threshold triggering pressure increment has propagated (i.e. $r_{max}(p = p_t)$). It is well established that effective hydraulic properties such as the effective hydraulic conductivity and effective hydraulic diffusivity describe the behavior of mean pressure fields (i.e. p_{avg}) in heterogeneous porous media (Dagan, 1989; Gelhar, 1993; Sanchez-Vila et al., 2006; Zhang, 2001). In a homogeneous porous medium, the seismic diffusivity is undoubtedly equivalent to the effective hydraulic diffusivity. However, due to the rapid preferential propagation of pressure increments through pathways of relatively high hydraulic diffusivity, this equivalence is not expected to hold in heterogeneous subsurface formations, since the propagation of $r_{max}(p = p_t)$ and $r(p_{avg} = p_t)$ are unlikely to coincide.

We have presented numerical simulations of fluid injection and IIS in heterogeneous domains in order to investigate the possible distinction between the seismic and effective hydraulic diffusivity. The fluid flow model (based on uncoupled, linear pressure diffusion) simulated IIS by evaluating the Mohr-Coulomb failure criterion at randomly seeded weak points throughout the domain. We considered two forms of subsurface heterogeneity: spatially correlated, random fields of hydraulic diffusivity (i.e. the SGS domains) and highly fractured, low permeability rock (i.e. the DFNM domain). Results of the SGS simulations show that the location of $r_{max}(p = p_t)$ does indeed exceed $r(p_{avg} = p_t)$ in all cases at all times, with the distance between them increasing with the variability in the hydraulic permeability. Thus, we found that the seismic diffusivity (which describes the propagation of $r_{max}(p = p_t)$ and the spatiotemporal patterns of IIS) is in fact distinct from and greater than the effective hydraulic diffusivity of the SGS domains. After calculating the seismic diffusivity (D_s) and effective hydraulic diffusivity (D_{eff}) of each SGS domain, we found that the ratio D_s/D_{eff} was always above one and systematically increased with the degree of heterogeneity (i.e. with Var(lnκ)). For the largest value of Var(lnκ) considered (i.e. 5.3), the ratio D_s/D_{eff} had an ensemble mean of approximately 1.9 and the maximum value encountered in any realization was approximately 3.4. In contrast, results of the injection simulations in the DFNM domain found that the ratio D_s/D_{eff} was approximately 10.2, showing an order of magnitude difference between the seismic diffusivity and effective hydraulic diffusivity. The fracture network in the DFNM domain provides well-connected pathways of relatively high hydraulic diffusivity through which pressure increments can rapidly propagate. Moreover, we compared the DFNM simulation results to those from a homogeneous domain with the effective hydraulic properties of the DFNM domain. As expected, we found that the triggering front propagates much more rapidly in the DFNM domain and thus the equivalent homogeneous domain underestimates the extent of the seismically active region during fluid injection.

Overall, the results of this study indicate that the influence of subsurface heterogeneity creates spatiotemporal patterns of IIS that are not well described by the effective hydraulic diffusivity of the heterogeneous domain. Rather, the propagation of the triggering front and spatiotemporal patterns of IIS are controlled by the so-call seismic diffusivity (Talwani & Acree, 1985). This result suggests that estimates of hydraulic diffusivity from seismicity-based approaches (Delepine et al., 2004; Hummel & Müller, 2009; Hummel & Shapiro, 2012, 2013; Rothert & Shapiro, 2003; Segall & Lu, 2015; Shapiro et al., 1997; Shapiro & Müller, 1999; Shapiro et al., 2002; Shapiro & Dinske, 2009a, 2009b; Shapiro, 2015) likely over-estimate the true effective hydraulic diffusivity of the subsurface. Another critical implication is that modeling fluid injection operations with homogeneous domains using the effective hydraulic properties of the injection formation (Brown et al., 2017; Catalli et al., 2016; Dempsey & Riffault, 2019; Keranen et al., 2014; Langenbruch & Zoback, 2016; Langenbruch et al., 2018; Pollyea et al., 2019; Riffault et al., 2018) may underestimate the rate of propagation of seismicity and the size of the seismically active region. Alternatively, when employing the seismic diffusivity as a substitute for the effective hydraulic diffusivity, such models would produce the correct spatiotemporal patterns of IIS, yet other hydraulic processes may be inaccurately represented (such as inter-well connectivity or reservoir pressurization) (Birdsell et al., 2018; Haagenson et al., 2018; Haagenson & Rajaram, 2020).

Even if seismic diffusivity is not an accurate estimate of the effective hydraulic diffusivity in heterogeneous porous media, it may still be helpful by providing a simple descriptor of the rate at which seismicity appears to spread at a particular location. Since both the seismic diffusivity and the effective hydraulic diffusivity of subsurface formations are influenced by heterogeneity, estimation of the statistical properties of the underlying heterogeneity based on estimation of the seismic diffusivity may provide an indirect approach to estimating the effective hydraulic diffusivity. Conversely, hydraulic characterization at a particular site may indirectly provide estimates of the seismic diffusivity, allowing operators to evaluate the potential for the rapid propagation of seismicity.

Our approach involves approximations, which implies certain limitations. Foremost, the subsurface stress state is considered static during the simulation and only impacts the calculation of the Mohr-Coulomb failure criterion. This approach is similar to numerous previous studies that also neglect potential mechanical effects on the behavior of IIS (Brown et al., 2017; Dempsey & Riffault, 2019; Hummel & Shapiro, 2013; Keranen et al., 2014; Langenbruch et al., 2018; Nakai et al., 2017; Rothert & Shapiro, 2003; Shapiro et al., 1997, 2002; Shapiro & Dinske, 2009b; Shapiro, 2015; Talwani & Acree, 1985). Recently, there is a growing body of research investigating the influence of mechanical coupling on the behavior of IIS through both poroelastic stressing (Chang & Segall, 2016; Jha & Juanes, 2014; Rutqvist et al., 2013; Segall & Lu, 2015; Zhai & Shirzaei, 2018; Zhai et al., 2019) and static stress transfer following a seismic event (Catalli et al., 2016; Schoenball et al., 2012), which will certainly lead to improved insights on the physical nature of IIS phenomenon. In addition, we have neglected the dilation of fractures due to the change in normal effective stress (Bandis et al., 1983) or due to shear failure along a fracture (Rong et al., 2016; Ye & Ghassemi, 2018). We fully expect that including fracture dilation would only hasten the propagation of the triggering front and thus enhance the distinction between seismic diffusivity and the effective hydraulic diffusivity in formations of fractured rock. The influence of mechanical coupling is more difficult to predict, as changes to the stress state could be either stabilizing or destabilizing depending on the orientation of any given fracture. Future studies on this topic could investigate the potential influence of mechanical coupling or fracture dilation on the spatiotemporal behavior of IIS in three-dimensional domains of fractured rock. It may also be insightful to further explore the critical implications of the study's findings – particularly the potential for connecting the spatiotemporal patterns of IIS and estimates of the seismic diffusivity to various properties of the underlying fracture network. This could provide a highly beneficial tool for subsurface characterization or predicting the seismic response at potential fluid injection sites.

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