Spectral Ground Motion Characteristics of Induced Earthquakes in The Fort Worth Basin, Texas

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SPECTRAL GROUND MOTION CHARACTERISTICS OF INDUCED EARTHQUAKES IN THE FORT WORTH BASIN, TEXAS

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SPECTRAL GROUND MOTION CHARACTERISTICS OF INDUCED EARTHQUAKES IN THE FORT WORTH BASIN, TEXAS

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in Partial Fulfillment of the Requirements for the degree of
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by
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Spectral Ground Motion Characteristics of Induced Earthquakes in The Fort Worth Basin, Texas

Advisor: Professor Brian Stump

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Concurrent with the development of unconventional oil and gas production, seismicity in the central United States has dramatically increased. Previous studies in many locations suggest that the earthquakes reactivate pre-existing faults, as a result of changing subsurface stresses due to injection of fluids from wastewater disposal wells. This sudden rise in seismicity rate motivates an assessment and possible mitigation of seismic hazard due to the proximity of induced seismicity to metropolitan areas like the Dallas-Fort Worth, Texas.

Successful assessment and mitigation of earthquake hazards requires estimates of ground motion parameters representing path attenuation, site effects and source characteristics constrained by observed ground shaking. The generalized inversion technique (GIT) is a well-known spectral method that separates the three contributions based on a database composed of accelerometric recordings from many small to moderate sized earthquakes.

This dissertation aims to better understand seismic hazards in the Fort Worth Basin (FWB) and source physics of shallow-depth earthquakes that have been shown to be induced by regional wastewater injection. The GIT is used to separate the three contributions (path, site, and source) from the observed ground motions to yield seismic hazard assessment and source
discrimination of induced earthquakes. Since there are no hard-rock recording sites within the FWB, a new method is developed for the application of GIT under this circumstance, called a site correction method. Here, the efficacy of modified GIT is validated and compared against similar estimates using the empirical Green’s functions technique, which isolate source property using co-located large and small events.

In these studies, we find that the GIT derived seismic attenuation suggests the presence of a mid-crustal boundary and partially fluid-saturated material. The GIT site amplification functions document maximum amplification as a high as a factor of 5, slightly larger than the amplification of 3 estimated at the resonant frequency of the fundamental wavelength corresponding to 30 m depth. The site amplification (3 or 5) may be consistent with the thick sequences of sediments in the basin. At the resonant frequency, the GIT amplification is validated against horizontal-to-vertical ratio site functions and synthetic site responses from $V_{S30}$ data and correlates with geologic conditions. Average stress drop estimates from FWB earthquakes are ~5 MPa, similar to stress drops from tectonic intraplate earthquakes that range from 1 to 10 MPa. All of these results lead to enhanced earthquake hazards for induced earthquakes in the basin structure.

The Dallas-Fort Worth Airport sequence, which occurred shortly after the initiation of injection on a fault close to the well, shows a lower mean stress drop (~1 MPa) than other FWB earthquake sequences (~5 MPa). The Airport stress drops increase with distance from the injector within the first 1.5 km, suggesting a possible fluid effect on the induced earthquake rupture predicted by direct triggering of earthquake via rapid pore pressure diffusion. This effect suggests that stress drop variations with respect to short distances from an injector could be used to discriminant between injection-induced earthquakes and tectonic events.
TABLE OF CONTENTS

LIST OF FIGURES ................................................................................................................. xii

LIST OF TABLES ...................................................................................................................... xv

CHAPTER 1 INTRODUCTION ................................................................................................. 1

FIGURES ................................................................................................................................. 8

REFERENCES .......................................................................................................................... 9

CHAPTER 2 FUNDAMENTAL SEISMOLOGICAL THEORY .................................................... 11

2.1 The Elastodynamic Equation of Motion and Far-Field Radiation ................................. 12

2.2 Earthquake Source Models .......................................................................................... 15

2.2.1 Haskell Source Model ............................................................................................... 15

2.2.2 The Brune Source Model .......................................................................................... 18

2.2.3 Source Parameters from Brune Source Model .......................................................... 20

2.2.4 Discussion of Rupture Complexity ......................................................................... 21

2.2.5 Role of Fluid on Earthquake Source ....................................................................... 22

2.3 Path Attenuation ............................................................................................................ 23

2.3.1 Geometrical Spreading ............................................................................................ 23

2.3.2 Inelastic Attenuation and Scattering ....................................................................... 23
CHAPTER 4 STRESS-DROP ESTIMATES FOR INDUCED SEISMIC EVENTS IN THE
FORT WORTH BASIN, TEXAS................................................................. 100

Abstract ........................................................................................................ 100

4.1 Introduction ............................................................................................. 101

4.2 Seismicity of the Fort Worth Basin, Texas .............................................. 104

4.3 Datasets .................................................................................................. 106

4.4 Methodology .......................................................................................... 107

4.4.1 The GIT for Cleburne Sequence ....................................................... 107

4.4.2 Stress Drop Calculation ................................................................. 108

4.4.3 The Empirical Green's Function (EGF) method ................................ 108

4.5 Results .................................................................................................. 111

4.6 Discussion ............................................................................................. 116

4.7 Conclusions .......................................................................................... 121

FIGURES ...................................................................................................... 123

TABLES ........................................................................................................ 136

APPENDIX B ............................................................................................... 142

REFERENCES ............................................................................................. 147
6.6 Conclusions ......................................................................................................................... 197

FIGURES ........................................................................................................................................ 199

TABLES ........................................................................................................................................ 206

REFERENCES .................................................................................................................................. 210

CHAPTER 7 SUMMARY AND CONCLUSIONS ............................................................................. 214

REFERENCES .................................................................................................................................. 219
LIST OF FIGURES

Figure 1.1. Major oil and gas regions in Texas ................................................................. 8
Figure 2.1. The Haskell source model ............................................................................... 27
Figure 2.2. Seismic spectrum of two boxcars .................................................................. 28
Figure 2.3. The Brune source model - the finite source effect on the near-field displacement .... 29
Figure 3.1. Map showing earthquakes and seismic stations in the Fort Worth Basin ........... 62
Figure 3.2. Data distributions ......................................................................................... 63
Figure 3.3. Examples of accelerograms .......................................................................... 64
Figure 3.4. Site amplification of reference site ................................................................ 65
Figure 3.5. Nonparametric attenuation functions for 4 frequencies ................................. 66
Figure 3.6. Nonparametric attenuation functions for all frequencies ............................... 67
Figure 3.7. Example observed displacement spectra at local and regional distance .......... 68
Figure 3.8. $Q(f)$ model for $S$- and $P$-waves ............................................................... 69
Figure 3.9. Site response functions for 12 stations .......................................................... 70
Figure 3.10. Histogram of maximum amplification of site effects .................................... 71
Figure 3.11. Comparison between average spectral ratio and GIT H ratio ........................ 72
Figure 3.12. Four example of source spectra .................................................................... 73
Figure 3.13. Stress drop estimates for $S$-waves ............................................................... 74
Figure 3.14. Stress drop estimates for $P$-waves ............................................................... 75
Figure 3.15. Comparison of three magnitudes ($M_L$, $M_w$, and $m_{bLg}$) .......................... 76
Figure 3.A1. Test of three distance bins (0.5, 2, and 5 km) .................................................. 84
Figure 3.A2. Site amplification with a single reference and average site constraints .......... 85
Figure 3.A3. Nonparametric attenuation functions from initial and site corrected reference ..... 86
Figure 3.A4. Comparison for 4 distance limits (20, 40, 50, and 80 km) for Q estimates ........ 87
Figure 3.A5. Total site amplifications of GIT H ................................................................. 88
Figure 3.A6. Total site amplifications of GIT HV .............................................................. 89
Figure 3.A7. Total site amplifications of GIT V ................................................................. 90
Figure 3.A8. Comparison of short and longer time window for P-waves ......................... 91
Figure 3.A9. Stress drop estimates for normalized P-waves ............................................ 92
Figure 4.1. Map showing four sequences and seismic stations in the Fort Worth Basin ........ 123
Figure 4.2. Details of four study areas ............................................................................. 124
Figure 4.3. Example time series and S-wave spectra ...................................................... 125
Figure 4.4. Nonparametric attenuation functions and reference site effects for Cleburne ...... 126
Figure 4.5. Process of retrieving EGF source spectra ...................................................... 127
Figure 4.6. Comparison of GIT and EGF results ............................................................... 128
Figure 4.7. EGF corner frequency comparison from Boatwright and Brune models .......... 129
Figure 4.8. Analyses of GIT source parameters ............................................................... 130
Figure 4.9. Spatial distribution of stress drops for four sequences ................................... 131
Figure 4.10. Temporal stress drop distribution for four sequences ................................... 132
Figure 4.11. Stress drop estimates versus event depth ..................................................... 133
Figure 4.12. Comparison of stress drops in CUS, OK, FWB, and DFW areas ................. 134
Figure 4.13. Stress drops versus distance from nearest injector and pore pressure increases .... 135
Figure 4.B1. Comparison of unsmoothed and 1-Hz smoothed spectra ............................. 144
Figure 4.B2. Site response for Cleburne sequence ................................................................. 145
Figure 4.B3. Examples of normalized spectral ratio for validation of 2 km separation ........... 146
Figure 5.1. Map with 5 sequences and faults and downthrown ............................................. 165
Figure 5.2. Lateral stress drops for 5 sequences ........................................................................ 166
Figure 5.3. Duration between injection initiation and onset seismicity ................................... 167
Figure 5.4. Stress drops versus $M_w$, depth, and time for 5 sequences ................................. 168
Figure 5.5. Stress drops versus distance from the nearest injector ........................................... 169
Figure 5.6. Change in pore pressure in the FWB ................................................................. 170
Figure 5.7. Rate of pore pressure increase in the FWB .......................................................... 171
Figure 5.C1. DFW catalog selection criteria ........................................................................... 178
Figure 5.C2. Site response functions for DFW Airport sequence ........................................... 179
Figure 6.1. Map showing geologic proxy $V_{S30}$ and in-situ $V_{S30}$ ........................................ 199
Figure 6.2. Site amplifications from multiple sensors ............................................................ 200
Figure 6.3. Histogram of site amplifications .......................................................................... 201
Figure 6.4. Site amplification factor from 3 techniques with geology .................................... 202
Figure 6.5. Site amplifications of 16 stations showing good fit for GIT H and QWA .......... 203
Figure 6.6. Ten site amplifications showing little agreement .................................................. 204
Figure 6.7. Comparison of peak frequency and peak amplification from 3 techniques ......... 205
LIST OF TABLES

Table 3.1. Stations and sensors used in Chapter 3 ................................................................. 77
Table 4.1. Source parameters for 14 target events using EGF method .................................. 136
Table 4.2. GIT source parameters for 79 earthquakes from 2009 to 2018. .......................... 137
Table 4.3. Injection wells near each sequence ................................................................. 141
Table 5.1. GIT source parameters for 10 Dallas-Fort Worth earthquakes ............................ 172
Table 5.2. Mean and median stress drops for 5 sequences. ................................................. 173
Table 5.3. Pore pressure changes at depth 4.5 km from Ogwari et al. (2018). ........................ 174
Table 6.1. Station classification in geology ........................................................................ 206
Table 6.2. Station information used in Chapter 6 ............................................................. 207
Table 6.3. Site amplification results ............................................................................... 209
This is dedicated to my beloved wife and daughter, my family.
CHAPTER 1

INTRODUCTION

Natural hazards from earthquakes present serious dangers to many areas around the world. A few memorable large events provide some context such as the 1994 Northridge earthquake (57 dead, 22-86 billion US$ loss), the Sumatra-Andaman earthquake with Indian Ocean-wide tsunami in 2004 (about 225000 dead and estimated economic loss of approximately 15 billion US$), and the 2010 Haiti earthquake (230000 dead and 7.8-8.5 billion US$ loss) (Oth, 2007; Bartholomew, 2014; Amadeo, 2020). In particular, the 1811-1812 earthquakes in the New Madrid Seismic Zone (equivalent potential damage simulated in 2008: 7.2 million people displaced from their homes, 300 billion US$ loss) and the 2011 New Zealand Christchurch earthquake (185 dead, 1500-2000 injuries, 40 billion NZ$ loss) are examples pulled from similar tectonic or hazard settings of this dissertation; namely significant intraplate events or events under sedimentary basins, motivating the study of propagation path effects, site effects, and earthquake source process in order to improve hazard assessment (Oth, 2007; Elnashai et al., 2009; Potter et al., 2015).

Earthquakes are generally triggered by stress perturbations on faults due to natural process (e.g., static stress, dynamic stress, and Earth tides, etc.) or anthropogenic activity that perturbs natural pre-existing stresses in the crust (e.g., injecting fluid for waste disposal, enhanced oil recovery, and hydraulic fracturing, etc.). Human-induced earthquakes can increase
hazard and risk, particularly where infrastructure and populations are not prepared for earthquakes. In the Central and Eastern United States (CEUS), wastewater disposal, hydraulic fracturing, and CO$_2$ injection associated with the development of unconventional oil and gas production have led to increases in earthquake rates (Ellsworth, 2013; Holland, 2013; Keranen et al., 2014; Frohlich et al., 2016). Due to the increase in seismicity, earthquake hazards estimates were increased in the 2014 USGS National Seismic Hazard Model (NSHM) for areas in Oklahoma, Kansas, Colorado, New Mexico, central Arkansas, and north-central Texas near Dallas by more than a factor of 3 (Petersen et al., 2016). These sudden changes raised concerns of earthquake damage by residents, engineers, and public officials. Weingarten et al. (2015) report that more than ~60% of all CEUS seismicity (M 3.0+) is associated with injection wells. Before the year 2000, an average of ~20% of all CEUS seismicity was associated with injection wells while the yearly percentage of associated earthquakes rose sharply to ~87% from 2011 to 2014 (Weingarten et al., 2015). Many induced earthquakes in the CEUS are observed to be associated with wastewater disposal, with maximum magnitudes exceeding 5, as illustrated by the M5.8 Pawnee, M5.1 Fairview, and M5.0 Cushing earthquakes in Oklahoma in 2016 (McGarr, 2014; McGarr and Barbour, 2017; Foulger et al., 2018).

The oil boom in Texas rose during the early 20th century beginning with the discovery of a large petroleum reserve near Beaumont, Texas. Development of shale wells technologies (horizontal drilling and hydraulic fracturing) led to numerous other large oil fields in Texas since the mid-2000s (The Academy of Medicine, Engineering and Science of Texas [TAMEST], 2017). As a result, Texas produces oil and gas now exceeds one-third of the US’s total oil production (U.S. Energy Information Administration [US EIA], 2017). Current oil and gas activity in Texas includes the Anadarko Basin in the Texas Panhandle region, the Barnett Shale
in North Texas (study area in this dissertation), the Eagle Ford in South Texas, the Haynesville area of East Texas, and the Permian Basin in West Texas (Figure 1.1). Similar to other areas in CEUS, the shale gas and oil production results in an increase in the rate of seismicity in Texas. Frohlich et al. (2016) report that the earthquakes of magnitude 3 and above across Texas has occurred at about 2 events per year before 2008. In contrast, there have been about 12 earthquakes per year between 2008 and 2016.

The Fort Worth Basin (FWB) in north Texas had experienced little to no significant deformation due to faulting over the past 300 Ma (Magnani et al., 2017), but since 2008, the area has experienced over 30 felt earthquakes (Frohlich et al., 2016). Temporal and spatial evidence supports the conclusion that FWB earthquakes are linked to wastewater injection disposal (Hornbach et al., 2015; Ogwari et al., 2018; Quinones et al., 2019; Scales et al., 2017). Thus, this dissertation will focus on the wastewater injection mechanism, not hydraulic fracturing and CO₂ gas injection. Most felt earthquakes in the FWB have occurred on faults rooted in the crystalline basement at focal depths of ~4.5 km and extending upward through the overlying Ordovician Ellenburger and Mississippian Barnett formations (Magnani et al., 2017; Quinones et al., 2018). Injection deep intervals are slightly shallower than many of the earthquake focal depths. Most FWB seismicity occurs in swarm-like patterns on northeast (NE)-southwest (SW) trending basement faults and within the overlying Ellenburger dolomitic limestone formation. When the Azle-Reno sequence began in 2013, a group of residents from Azle-Reno were upset and requested an explanation from the state on the seismicity. In excess of 800 people attended a 2014 meeting sponsored by the Texas Railroad Commission, illustrating the local concern for the recently increased hazard associated with the induced seismicity (Villafranca, 2014).
This dissertation aims to improve seismic hazard assessment and source discrimination in the FWB associated with induced seismicity by conducting through source, path, and site analyses for these events with measured magnitudes as large as 4.0. A least squares inversion approach usually referred as the generalized inversion technique (GIT) provides the basis for the separation of source, path and site spectra, crucial components needed to estimate ground shaking and its possible damage from future earthquakes. The results of such studies ultimately provide input for appropriate building codes and construction of infrastructure (e.g., bridges, tunnels, roads, and power supplies, etc.) designed to mitigate the effects of future earthquakes. Moreover, if the source properties of the induced earthquakes differ from events within active tectonic regions, then this study can provide tools to discriminate human-related events from natural earthquakes.

Chapter 2 introduces the reader to the fundamental seismological concepts developed and applied in this dissertation. The chapter includes a short section on the elastodynamic equation of motion, the representation theorem and the source description of earthquakes. The Haskell and Brune models are subsequently presented. The role of fluids, which potentially reduces the effective stress or stress drop on the fault in the FWB, are discussed. After completing the discussion on the earthquake source, the focus turns to the path and site effects, which can significantly modify the level of ground motions at a given receiver location.

Chapter 3 is a reformatted version of an article published in Bulletin of the Seismological Society of America: Jeong, S. J., B. W. Stump, and H. R, DeShon (2020), Spectral characteristics of ground motion from induced earthquakes in the Fort Worth Basin, Texas using the generalized inversion technique. This study focuses on propagation path attenuation, site effects, and source characteristics (i.e., stress drop) estimated from 90 FWB earthquakes for Azle-Reno,
Irving-Dallas, and Venus seismic sequences. The GIT is used to separate the three contributions from the observed acceleration spectra of the FWB earthquakes. The details of the GIT process are presented in this chapter and the modification I designed and implemented to make the technical applicable to use in the sedimentary basins with no existing hard-rock reference sites. The main results including travel path, site amplification, and source parameters are described with an introduction to the tectonic and geologic structure of the FWB region.

Chapter 4 is a reformatted version of an article published in the *Bulletin of the Seismological Society of America*: Jeong, S. J., B. W. Stump, H. R. DeShon, and L. Quinones (2021), Stress-drop estimates for induced seismic events in the Fort Worth Basin, Texas. Source parameters (stress drop, seismic moment, and source radius) are estimated using both empirical Green’s function (EGF) and the revised GIT for four sequences in the FWB: Cleburne 2009, Azle-Reno 2013, Irving-Dallas 2014, and Venus, Texas 2015. Here, the traditional EGF approach is applied to well-recorded 14 moderate FWB earthquakes for comparison to source estimates using the modified version of GIT. Stress drop, magnitude scaling and spatiotemporal correlation are analyzed and compared for the established EGF and the GIT. In this study, both approaches produce similar estimates of average stress drop (> 5 MPa) that are consistent with values for tectonic intraplate earthquakes. In addition, stress drops using each approach appear to be invariant with seismic moment, providing no differences from previous natural earthquakes. Results are discussed in relation to the impact of fluid and pore pressure change.

In Chapter 5, stress drop estimates are extended to the Dallas-Fort Worth (DFW) Airport sequence. This sequence is unique in that it was the first sequence in the FWB and produced earthquake locations that migrate away from the single-source well with time, resulting in higher b-values (Ogwari et al., 2018). The stress drop estimates in this case study are significantly lower
than those from other FWB sequences and increase with increasing distance from the single injector, suggesting a direct fluid effect on earthquakes rupture. In this one case, the seismicity began shortly after injection was initiated on or within 100s m of the causative fault, consistent with a direct link between the fluids, the faults and the earthquakes.

In Chapter 6, the site amplifications derived from the GIT in the FWB are compared with other common approaches to investigate appropriateness of site amplification and the role of surface geology (rock types and geologic age). These other techniques include the horizontal-to-vertical ratio site function and a quarter wavelength approach estimated by average shear wave velocities in the upper 30 m ($V_{S30}$) data. The different estimates of site response are roughly in agreement with GIT site functions at resonant frequency, which is corresponding to 30 m depth. The average amplification is ~3, less than peak amplification (a factor of 5). Three site amplifications generally correlate with the known geological structure. The results suggest that each of the three methods has equally applicability to constrain site amplification at the resonant frequency. Thus, $V_{S30}$ data widely used in engineering seismology could provide a good estimate for site function at resonance frequency from 2.5 to 10 Hz, consistent with fundamental vibration frequency for 1 to 8 story buildings (2-10 Hz).

Finally, Chapter 7 ends the dissertation with a summary of the results and the main conclusions drawn from these integrated studies to address two questions: (1) seismic hazard assessment; (2) source discrimination for the FWB earthquakes. Path attenuation may be affected by mid-crustal boundary and partially fluid-saturated material. Site amplification at resonant frequency correlates with geologic conditions, producing an average amplification of ~3, but slightly less than peak amplification of ~5. FWB Stress drops are generally consistent with those of tectonic intraplate earthquakes. These results have a significant impact on ground motion.
prediction and seismic hazard assessment. The lower stress drops observed in DFW Airport events, possibly influenced by fluid injection, could be the key to improve source discrimination between induced earthquakes and tectonic earthquakes.
Figure 1.1. Major oil and gas regions in Texas. Barnett shale area in north central Texas is the study region in this dissertation. Figure from TAMEST (2017).
REFERENCES


CHAPTER 2
FUNDAMENTAL SEISMOLOGICAL THEORY

Observed ground motions result from multiple contributions, commonly source, propagation path, and site effects. These contributions are considered as a linear filter system, thus they are independent of each other. The instrument corrected waveform, \( u(t) \), can be simplified with convolution as:

\[
u(t) = s(t) * a(t) * i(t),
\]

(2.1)

where \( s(t) \) is the source term, which is the contribution from the earthquake rupture process, \( a(t) \) is the propagation path attenuation, \( i(t) \) is the site effect, and \( * \) represents the convolution operator. The seismic waveform can be described in frequency domain as a complex multiplication:

\[
U(f) = S(f) \cdot A(f) \cdot I(f),
\]

(2.2)

where \( S(f) \) is the source, \( A(f) \) is the propagation path, \( I(f) \) is the site spectra. Note that we restrict our analysis to Fourier amplitude spectrum. \( S(f) \) is used to estimate source parameters such as seismic moment, corner frequency (source radius), and stress drops, which is the released stress during earthquake rupture. \( A(f) \) can be separated into geometrical spreading and inelastic attenuation (including scattering) known as the quality factor \( Q \). \( I(f) \) consist of the site amplification and site diminution at high frequency.
This chapter briefly explains the basic seismological concepts that are used to estimate the source characteristics of the earthquakes based on the elastodynamic equations of motion. The equations for a continuous medium are used to develop the representation theorem, which describes the ground motion in terms of source spectra, propagation path, and site transfer function. Subsequent to this representation two source models, the Haskell (1964) source model for rectangular faults and the Brune’s circular crack model (1970) are used to interpret the waveforms from earthquakes. A brief description of how fluids might influence the earthquake source characteristics follows. The chapter ends with development of the path and site effects illustrating their impacts on the ground motions.

2.1 The Elastodynamic Equation of Motion and Far-Field Radiation

Far-field seismic waves can be described by a single fundamental vector equation (i.e., the equations of motion), which connect forces associated with the source to observable displacements in a continuous medium for a source with spatial dimensions that are small compared to the wavelengths of observed waves. The equations of motion are based on Newton’s second law (i.e., $F = ma$ where $F =$ force, $m =$ mass, and $a =$ acceleration) under the assumption that the earth is a continuous medium. Newton’s second law or conservation of momentum under these circumstances can be represented as:

$$\frac{d}{dt} \iiint \rho \frac{du}{dt} dV = \iiint f dV + \iint T(n) dS, \quad (2.3)$$

where $\rho$ is density, $u$ is the displacement vector with $x = (x_1, x_2, x_3)$, $f$ is the body force term, and surface traction, $T(n)$, within volume $V$ with subsurface $S$ (Aki and Richards, 2002). Here, $T(n) = T_i = \sigma_{ij} n_j$, where $\sigma_{ij}$ is the stress tensor and $n$ is the normal vector. If we assume very small particle displacements, using Gauss’s divergence theorem to transform the surface integral
to a volume integral and considering the forces per unit of volume, the equation is reduced to local elastodynamic equations:

\[ \rho \frac{\partial^2 u_i}{\partial t^2} = f_i + \frac{\partial \sigma_{ij}}{\partial x_j}. \]  

(2.4)

Hooke’s law is applied to equation (2.4) to simplify the equation in terms of spatial derivatives of displacement:

\[ \sigma_{ij} = C_{ijkl} \epsilon_{kl}, \]  

(2.5)

where \( \epsilon_{ij} = \frac{1}{2} \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) \) is strain and \( C_{ijkl} \) is stiffness tensor or elastic constants. In an isotropic medium, the stress is simplified to the two elastic moduli, Lame moduli \( \lambda \) and shear moduli \( \mu \):

\[ \sigma_{ij} = \lambda \epsilon_{kk} \delta_{ij} + 2\mu \epsilon_{ij}. \]  

(2.6)

Using this relation, the elastodynamic equations are rewritten as:

\[ \rho \frac{\partial^2 u_i}{\partial t^2} = \rho f_i + (\lambda + 2\mu) \nabla (\nabla \cdot u_i) - \mu \nabla \times \nabla \times u_i, \]  

(2.7)

where the body force distribution per unit volume, \( \rho f_i \), is

\[ \rho f_i = F(t) \delta(r) \mathbf{a}, \]  

(2.8)

which represents the source with \( F(t) \) becoming source time history, \( \delta(r) \) is delta function, and \( \mathbf{a} \) is a unit vector in the direction of the force with time \( t \) and source-to-receiver distance \( r \). After some manipulation (e.g., Lay and Wallace, 1995), the total displacement field becomes:

\[ u_i(r, t) = \frac{1}{4\pi \rho} \left[ \mathbf{a} \cdot \nabla \left( \frac{1}{r} \right) \right] \int_{r/\alpha}^{r/\beta} \tau F(t - \tau) d\tau \]
\[ + \frac{1}{4\pi \rho \alpha^2} \frac{1}{r} (a \cdot \nabla r) \nabla r F \left( t - \frac{r}{\alpha} \right) \]
\[ + \frac{1}{4\pi \rho \beta^2} \frac{1}{r} [a - (a \cdot \nabla r) \nabla r] F \left( t - \frac{r}{\beta} \right), \]  
(2.9)

where \( \alpha \) and \( \beta \) are velocities for \( P \)- and \( S \)-waves. The first integral term is the near-field term, where the amplitude attenuates with \( 1/r^2 \) and the other terms are far-field terms decaying as \( 1/r \), with the far-field terms dominating the farther the wave propagates. Since most seismic observations are made at sufficient distance from faults, we focus on the far-field terms (Shearer, 2009). Generalization of the far-field radiation from a point force buried in an elastic half-space becomes:

\[ u_c = \frac{R_{\theta \phi}}{4\pi \rho c^2} \frac{1}{r} F \left( t - \frac{r}{c} \right). \]  
(2.10)

Equation (2.10) illustrates that the seismic waves attenuate as \( 1/r \) and propagate with velocity \( c \). Realistically, seismic sources are not point sources. The source process from a pure shear dislocation can be represented by a system of force couples perpendicular to each other (i.e., double couple) distributed over the spatial extent of the fault. Thus, the force term \( F(t) \) convert to moments \( M(t) = F(t)d \), where \( d \) is the distance between two point sources in the equation (2.10) distributed across the fault surface. If we assume a very small distance \( d \) along the axes of the forces, then \( M(t) \sim F(t) \). Under these conditions (see, e.g., Lay and Wallace, 1995), the far-field radiation from a double couple point source is represented as:

\[ u_c = \frac{R_{\theta \phi}}{4\pi \rho c^3} \frac{1}{r} \dot{M} \left( t - \frac{r}{c} \right), \]  
(2.11)

where \( \dot{M} \) is the time derivative of the moment tensor.
2.2 Earthquake Source Models

An earthquake can be represented as sudden shear displacement along a fault plane. Within the source region, the strain is large and Hooke’s law (linear relationship between stress and strain) can be broken. In this case, the earthquakes can be represented as a displacement discontinuity or slip. The finite fault model suggests that point sources summed with time lags to give the total time history is an adequate kinematic representation although it does not fully capture the dynamics of failure and rupture. In this chapter, we utilize these kinematic source representations to interpret the observed wavefield focusing on the two main models proposed by Haskell (1964) for rectangular fault model and Brune (1970) for a circular crack model.

2.2.1 Haskell Source Model

Haskell (1964) provided a simple solution for the far-field radiation from a rectangular fault of length $L$ and width $w$ (Figure 2.1) with a constant fracture rupture velocity. If the particle displacement takes a finite length of time to achieve its total offset, (i.e., rise time), the time history term, $\dot{M}(t - \frac{r}{c})$, in equation (2.11) becomes:

$$\dot{M}(t - \frac{r}{c}) = \mu w dx \dot{D}(t - \frac{r}{c}),$$  \hspace{2cm} (2.12)

where $\mu$ is shear modulus, $dx$ is the individual segments of the fault (Figure 2.1) and $\dot{D}(t - \frac{r}{c})$ is the far-field source time function. $\dot{D}(t - \frac{r}{c})$ can be rewritten in a more useful form by using shift property of the delta function $\delta$ as:

$$\dot{D}(t - \frac{r}{v_{r}}) = \dot{D}(t) \otimes \delta \left(t - \frac{r}{v_{r}}\right),$$  \hspace{2cm} (2.13)
where $\dot{D}(t)$ is the boxcar function with a duration of rise time $\tau_r$. Note that $c$ is replaced to $v_r$, rupture velocity. Equation (2.13) is rewritten using the fact that time function of individual point on finite fault area is summarized as:

$$u_c = \frac{R_{\theta\phi}\mu}{4\pi\rho c^3} w \int_0^x \dot{D}(t) \delta \left( t - \frac{r}{v_r} \right) dx,$$

(2.14)

where $x$ is length of the fault and $R_{\theta\phi}$ is radiation patterns for either $P$- or $S$-waves. Since $\dot{D}(t)$ is independent of $x$ and assumed to be the same everywhere on the fault, it can be moved outside of the integral as:

$$u_c = \frac{R_{\theta\phi}\mu}{4\pi\rho c^3} w \dot{D}(t) \int_0^x \delta \left( t - \frac{r}{v_r} \right) dx,$$

(2.15)

The integral of the delta-function by substituting $z = t - x/v_r$ leads to:

$$\int_0^x \delta \left( t - \frac{r}{v_r} \right) dx = \int_{t}^{t-x/v_r} -v_r \delta(z) dz$$

$$= -v_r [H(z) \big|_{t}^{t-x/v_r}]$$

$$= v_r \left[ H(t) - H \left( t - \frac{r}{v_r} \right) \right]$$

$$= v_r B(t; \tau_c),$$

(2.16)

where $B(t; \tau_c)$ is a boxcar with duration of the faulting process, $\tau_c = x/v_r$, reflecting the rupture time, noting that the integration of the delta-function is the Heaviside step function, $H(t)$. Thus, the displacement of unilateral slip from Haskell model is given by the convolution of two boxcars with lengths representative of the source rise time $\tau_r$ and rupture time $\tau_c$:
where \( B(t; \tau_r) = \dot{D}(t) \). The resulting source time function including these two characteristic time scales is a trapezoid function (see also Lay and Wallace, 1995):

\[
S(t) = M_0 (B(t; \tau_r) \otimes B(t; \tau_c)),
\]

where seismic moment \( M_0 \) is defined as the area of trapezoid function, \( M_0 = \int_{-\infty}^{\infty} \dot{M}(t) dt \), which is the sum of the slip of all particles on the fault. In the frequency domain, the convolution of two boxcars are equivalent to multiplication between two sinc function \((\sin[x]/x)\), and the equation (2.18) becomes:

\[
S(\omega) = M_0 \left| \frac{\sin(\omega \tau_r/2)}{\omega \tau_r/2} \right| \cdot \left| \frac{\sin(\omega \tau_c/2)}{\omega \tau_c/2} \right|.
\]

Figure 2.2 shows that the spectrum has three different regions as:

\[
S(\omega) \approx \begin{cases} 
M_0 & \omega < \frac{2}{\tau_c} \\
\frac{M_0}{\omega \tau_c} & \frac{2}{\tau_c} < \omega < \frac{2}{\tau_r} \\
\frac{M_0}{\omega^2 (\tau_r \tau_c)^2} & \omega > \frac{2}{\tau_r}
\end{cases}
\]

The intersect points of these three regions in the spectra are defined as corner frequencies for the rupture time and rise time, respectively. The spectrum is flat at low-frequency before encountering the corner associated with rupture time and constrains the level of \( M_0 \). Beyond this point the spectra decay as \( \omega^{-1} \) between the rise time and the rupture time and at high frequencies the spectra decay as \( \omega^{-2} \) capturing the destructive interference introduced by both rise time and rupture time.
2.2.2 The Brune Source Model

Brune’s circular crack model is a simple source model associated with slip on a fault as a result of changes in stress. The model assumes that the slip on the fault is driven by stress drop, which is the difference between stress before and after rupturing, inside the fault (i.e., temporal history of stress drop). In the dynamic crack model, stresses built with no slip on the fault until the critical value of shear stress exceeds the rock strength, which is the stress required to break the atomic bond in a crystal lattice or reinitiate slip on an existing fault, and subsequently sudden slip occurs associated with the stress drop. Slip occurs episodically with stress drop occurring simultaneously. Experimental and theoretical studies of rock failure (Byerlee, 1978) quantify the relationship between stress and rock failure. Here, we apply this fracture mechanics to the earthquake faulting with assumption of episodic slip representing the earthquake. Rupturing in the case of an earthquake is defined as a process of strain energy release associated with the creation of a new crack representing the fault.

For a finite faulting event, we define the static stress drop as the stress drop over the entire fault area (Brune, 1970). Using Hooke’s law, we estimate the static stress drop as (Kanamori and Anderson, 1975):

\[ \Delta \sigma = C \mu \left( \frac{D}{L} \right), \]  \( \text{(2.21)} \)

where \( D \) is the average displacement, \( L \) is a characteristic rupture dimension and a nondimensional constant \( C = 7\pi/16 \) for a circular fault (Scholz, 1990). Since \( L = \beta t \) for \( S \)-wave, the instantaneous strain \( \epsilon \) becomes:

\[ \epsilon = \frac{S(t)}{\beta t} = \frac{\Delta \sigma}{\mu}, \]  \( \text{(2.22)} \)
then, the source time function of near-field displacement can be rewritten as:

\[
S(t) = \frac{\Delta \sigma \beta t}{\mu}.
\]

(2.23)

Equation (2.23) should have a ramp-like time function with nearly constant \(\Delta \sigma\). The particle velocity is:

\[
\dot{S}(t) = \frac{\Delta \sigma \beta}{\mu}.
\]

(2.24)

Equation (2.24) is directly proportional to the stress drop. To approximate the near-field displacement including the diffracted waves, which are a near-field effect of the finite size of the dislocation, Brune (1970) multiplied equations (2.23) and (2.24) (Figure 2.3) by an exponential function. This results in a reduced particle velocity or far-field displacement:

\[
\dot{S}(t) = \frac{\Delta \sigma \beta}{\mu} e^{-\left(t/\tau\right)},
\]

(2.25)

where \(\tau\) is a time constant given by \(L/\nu_r\). Equation (2.25) forms a step with a decaying tail (Figure 2.3). The Fourier amplitude spectrum of the equation (2.25) becomes as:

\[
S(\omega) = \frac{\Delta \sigma \beta L}{\mu} \frac{1}{\omega^2 + \omega_c^2}.
\]

(2.26)

Equation (2.26) is simplified with a single corner frequency defined by the intersection of the \(\omega^0\) and \(\omega^{-2}\) asymptotes (Shearer, 2009) as:

\[
S(\omega) = \frac{\Omega_0}{1 + (\omega/\omega_c)^2},
\]

(2.27)

where \(\Omega_0\) is the low-frequency flat level and \(\omega_c\) is the corner frequency. The source model is then characterized by only two independent parameters: the seismic moment \(M_0\), which is
proposal to $\Omega_0$, and the corner frequency $\omega_c$. The spectrum has a $\omega^{-2}$-shape similar to the Haskell model. The Brune (1970) model has been shown to provide good agreement with observations of many tectonic events over large range of magnitudes (Frankel, 2019).

2.2.3 Source Parameters from Brune Source Model

The source properties of events are dependent on physical parameters such as seismic moment, source radius, and stress drop. The seismic moment can be directly obtained from equation (2.27) using the relationship, $M_0 = (4\pi \rho c^3 R \Omega_0)/F$ where $F$ is the average double-couple radiation coefficient, $R$ is the hypocentral distance, $\rho$ is the density, and $c$ is velocity. The source radius $r_0$ is estimated using the assumption that corner frequency $\omega_c$ is equivalent source radius or fault area:

$$r_0 = \frac{K_0 \beta}{\omega_c},$$

(2.28)

where $K_0$ is a constant depending on fault geometry.

Stress drop is calculated from the relation that scales the seismic moment by source radius using a circular crack model (Eshelby, 1957):

$$\Delta\sigma = \frac{7M_0}{16r_0^3}.$$  

(2.29)

Finally, Hanks and Kanamori (1979) introduce moment magnitude derived by seismic moment as:

$$M_W = \frac{2}{3} \log_{10} M_0 - 6.03,$$

(2.30)

where unit of $M_0$ is Nm.
2.2.4 Discussion of Rupture Complexity

Aki (1967) and Brune (1970) proposed an $\omega^{-2}$ model for source spectra. However, Beresnev (2019) argued that there is no reason for spectral fall-off to naturally comply with the $\omega^{-2}$ model and argues for the $\omega^{-2.5}$ model. Frasier and North (1978) suggest that the $\omega^{-3}$ source model is more appropriate based on analysis of the Rat Island seismic data. Kaneko and Shearer (2014, 2015) obtained various fall-off rates between $\omega^{-2}$ and $\omega^{-3}$. Based on these results, there remains debate as to whether the $\omega^{-2}$ law or the $\omega^{-n}$ model is better to describe the source spectra (Wang, 2019). Aki (1987) and Papageorgiou (1988) proposed that the high frequency variations are a result of incoherencies in the rupture process in a complex faulting process that predicts high-frequency deviations from $\omega^{-2}$ (Koyama, 1994).

In order to describe the systematical high-frequency changes, the barrier model (Das and Aki, 1977) and the asperity model (Kanamori and Stewart, 1978) were suggested. If this more complex rupture process is included, patch corner frequencies as a result of barrier/asperity size are appropriate as suggested by Papageorgiou and Aki (1983a, b). The second corner frequency results from the average rupture time of random fault patches, strictly related to the number of random pulses. Even though barrier/asperity models may be applicable to larger earthquakes similar source effects are reported in some small events. Archuleta and Ji (2016) simulated two corner-frequency spectra resulting from total rupture time and rise time for earthquakes between magnitude 3.3 and 5.3. In their study, the existence of rise time for the smaller earthquakes results from the initial crack (i.e., patch). In addition, Uchide and Imanishi (2016) observed two corner frequencies in source spectra for small and shallow earthquakes ($M_w$ 3.2-4.0) and suggest that asperity-like earthquakes may be the reason of the bump in spectra. They speculated that the lower and higher corner frequencies correspond to the entire rupture size and the strong patch
size, respectively. For injection-induced earthquakes, Fan and McGuire (2018) observed double humps at high frequencies in source spectra for Oklahoma Wavefield Experiment earthquakes ($M_w 2.3$). They interpreted these results to be caused by multiple subevents comprising the one main event (i.e., complex faulting process) combined with strong rupture directivity.

2.2.5 Role of Fluid on Earthquake Source

Wastewater associated with unconventional oil and gas production is disposed by injection into deeper formations, which can induce seismicity. The main mechanism responsible for triggering injection-induced earthquakes may be an increased pore pressure on critically stressed faults. Since the pore pressure and stress are often coupled (Hillis, 2000), changes in pore pressure can decrease the effective normal stress (or stress drop), effectively unlocking the fault and allowing slip initiation (Ellsworth, 2013). Goertz-Allmann et al. (2011) show a dependence of stress drop with radial distance from injection well in Basel, Switzerland. Their study also shows an inverse relationship between stress drop and changes in pore pressure. In the Salton Sea geothermal field, Chen and Shearer (2011) found increased stress drops with distance from the nearest injection well, which is consistent with the fact that the pore pressure decreases with distance away from the injector.

According to the effective stress law (Hubbert and Rubey, 1959), the static fault shear strength is:

$$\tau_f = \tau_0 + \mu (\sigma_n - P_f) \quad (2.31)$$

where $\tau_0$ is the cohesive strength of the fault, $\mu$ is the friction coefficient, $\sigma_n$ is the normal stress acting on the fault, and $P_f$ is the pore pressure within the fault zone. Equation (2.31) suggests that changes in pore pressure will increase the probability of failure by reducing effective normal
stress and fault frictional strength when the stress acting on the fault exceeds the threshold to sliding.

2.3 Path Attenuation

Since the Earth is not perfectly elastic material, energy is lost due to inelastic attenuation (wave energy is converted into heat), scattering (heterogeneities of propagation medium) and geometrical spreading. In order to interpret source properties from an observed spectrum, the path attenuation must be constrained.

2.3.1 Geometrical Spreading

When seismic waves propagate through the Earth, the size of wavefront increase with distance and waveform amplitude decreases with distance due to conservation of energy (known as Huygen's principle). Geometrical spreading is commonly defined as \( G(x) = r^{-\gamma} \) and dependent on the wave type. In general, body waves decay with \( \gamma =1 \) and surface waves decay with \( \gamma =0.5 \). Since \( Lg \) wave is commonly generated at \(~100\) km, the geometrical spreading becomes as \( G(x) = \frac{1}{r} \) for \( r <100 \) and \( G(x) = \sqrt{\frac{1}{100r}} \) for \( r \geq 100 \). The geometrical spreading is usually frequency independent.

2.3.2 Inelastic Attenuation and Scattering

Both inelastic attenuation and scattering lead to decrease in amplitude of seismic waves as they propagate in space and time. Note that inelastic attenuation leads to an effective loss of energy whereas scattering redistributes the wave energy. Both inelastic attenuation and scattering effects are commonly quantified as the quality factor \( Q \). The amplitude of seismic waves at time, \( t \), can be represented with the quality factor \( Q \) as follows:
\[ A(f, t) = A_0 e^{-\pi f t / Q(f)}, \]  
(2.32)

where \( A_0 \) is initial amplitude and \( f \) is frequency. \( Q(f) \) is the frequency dependent quality factor and describes as \( Q(f) = Q_0 f^a \), when \( f > 1 \text{ Hz} \). In practice, \( Q \) is constant for 0.1-1.0 Hz.

Alternatively, the equation (2.32) is represented as a function of frequency as:

\[ A(f, r) = A_0 e^{-\pi f r / Q(f)c}, \]  
(2.33)

where \( r \) is hypocentral distance and \( c \) is average velocity along the propagation path. Equation (2.33) can be represented as:

\[ A(f, t) = A_0 e^{-\pi f t^*}, \]  
(2.34)

where \( t^* = \int dt / Q(r, f) = T/Q_{av}(f) \), in which \( T \) is total travel time and \( Q_{av} \) is average \( Q \) along the path. \( t^* \) is typically used to represent teleseismic attenuation.

2.4 Site Effect

Site effects can strongly impact ground motions and play an important role in seismic hazard assessment. Site effects generally result from the near-surface effects of soft sedimentary layers, local topography, water table or sedimentary basins in vicinity of the receiver (Anderson, 2003). Site effects can be highly frequency dependent. Kane et al. (2011) suggest that small-scale site effects can lead to a minimum uncertainty of up to 30 % in stress drop estimates illustrating the importance of characterizing these effects for stress drop determination.

2.4.1 Site Amplification

Upward propagating S-waves are amplified when the seismic waves propagate into a lower velocity medium as a result of energy conservation. In addition, a continuous reverberation
of trapped seismic waves in low-velocity layers can increase destructive and constructive interference in distinct frequency bands (Bard, 1999). Site amplification leads to enhanced amplitudes at the resonance frequencies and subsequently distorts the shape of source spectra. There are two common approaches to estimating the site amplification spectrum: (1) Reference Site Method (RSM) and (2) Horizontal-to-Vertical Spectral ratio (HVSR). Both rely on spectral division, but RSM uses a hard rock site as a reference while the HVSR use the vertical component of S-wave as the reference under the assumption that the vertical component does not have strong site amplification.

2.4.2 Site Diminution

The acceleration source spectrum at frequencies above the corner frequency theoretically is expected to be flat as a function of frequency based on the $\omega^{-2}$ source model, but practical observations show a dissipation at high frequencies. This spectral decay is related to the effects of near-surface weathered layers (Anderson and Hough, 1984). This effect can be addressed by empirically accounting for the high-frequency decay in acceleration spectra using the parameter $\kappa$. If the acceleration spectra fall off exponentially with frequency then we can describe the parameters as follows (Anderson and Hough, 1984):

$$a(f) = A_0 \exp(-\pi \kappa f) \quad \text{for } f < f_E,$$  \hspace{1cm} (2.35)

where a constant $A_0$ depends on source properties, epicentral distance, and other factors, $\kappa$ is the high-frequency decay parameter, and $f_E$ reflects a frequency point above which the decay is no longer dominant or critical. Taking a natural logarithm of both sides of the equation (2.35), the formulation can be rewritten as:

$$\ln[a(f)] = \ln A_0 - \pi \kappa f.$$  \hspace{1cm} (2.36)
From the equation (2.36), the spectral decay parameter $\kappa$ can be estimated from the slope ($-\pi \kappa$) as a function of frequency using a linear least-squares fit within an effective frequency bandwidth from an initial frequency of exponential decay to $f_E$. Caution should be applied in estimating $\kappa$, since the $\kappa$ and path attenuation $Q$ factor can be trade off with each other.
Figure 2.1. The Haskell source model geometry. A one-dimensional fault plane of width $w$ ruptures through length $L$ with constant rupture velocity $V_r$. The length $dx$ is the individual segments of the fault. Figure from Lay and Wallace (1995).
Figure 2.2. Seismic spectrum of two boxcars. The intersection points ($\omega_{c1}$ and $\omega_{c2}$) of the asymptotes to the low-frequency and high-frequency portions of the spectrum defines two corner frequencies related to rupture time and rise time. Figure from Lay and Wallace (1995).
Figure 2.3. The Brune source model with the effect of finite source dimension on the near-field displacement. The particle velocity $\dot{u}$ will decrease and smoothly approach zero at times (red box). In the figure, $u$ is near-field displacement, $\sigma$ is the effective stress (stress drop), $\mu$ is the rigidity, $\tau$ is a time constant, $\beta$ is shear wave velocity, and $r$ is source radius. Figure from Brune (1970).
REFERENCES


CHAPTER 3

SPECTRAL CHARACTERISTICS OF GROUND MOTION FROM INDUCED EARTHQUAKES IN THE FORT WORTH BASIN, TEXAS USING THE GENERALIZED TECHNIQUE

Abstract

A generalized inversion technique (GIT) is applied to local seismic data from 90 induced earthquakes ($M_L$ 2.0-3.9) in the Fort Worth Basin (FWB) of north Texas, to separate path, site and source characteristics and to improve local seismic hazard assessment. Seismograms from three earthquake sequences on spatially separated basement faults are recorded on 66 temporary stations. Due to the lack of hard-rock recording sites within the sedimentary basin, we developed a site correction method for the appropriate GIT process. At about 30 km distance from the hypocenters, we observed a change in spectral attenuation and thus focus data analysis within this distance range. The estimated quality factors for $S$- and $P$-waves result in a $Q_S$ that is larger than $Q_P$, which we interpret as a result of concentrations of crustal pore fluids or partial fluid-saturated material along the path; an interpretation consistent with fluid-rich sedimentary rocks in the FWB. Strong site amplifications as much as 5 times on horizontal components reflect the thick sediments in the basin. A limited number of sites exhibit amplification or de-amplification on the vertical component that limits the use of horizontal-to-vertical spectral ratio methods for characterizing the site effect relative to the site effects estimated by GIT. Stress drops for all earthquakes range from 1.18 and 21.73 MPa with a mean of 4.46 MPa, similar to values reported for tectonic intraplate events. The stress drop values suggest that strong motion and seismic
hazard from the injection-induced earthquake in the FWB are comparable to those for tectonic earthquakes. The strong site amplification and fluid effects on propagation attenuation may be crucial factors to take into account for estimating seismic hazards of induced earthquakes in sedimentary basins.

3.1 Introduction

The Fort Worth Basin (FWB) in north Texas experienced little to no significant fault-driven deformation for the past 300 Ma (Magnani et al., 2017), but since 2008 the area has seen a sharp increase in earthquakes (Frohlich et al., 2016; Quinones et al., 2019). The majority of these events occur in five earthquake sequences: 2008-2009 Dallas-Fort Worth (DFW) International Airport (Frohlich et al., 2011; Janská and Eisner, 2012), 2009-2010 Cleburne (Justinic et al., 2013), 2013-2015 Azle-Reno (Hornbach et al., 2015), 2014-2016 Irving-Dallas (Quinones et al., 2018), and 2015 Venus sequence (Scales et al., 2017), which included the largest earthquake ($M_w$ 4.0) in the FWB. Many of these earthquakes are located on pre-existing faults rooted in the crystalline basement (see Magnani et al., 2017; Horne et al., 2020). The sudden increase in seismicity across the FWB has accompanied wastewater injection associated with oil and shale gas extraction from the Barnett Shale (Justinic et al., 2013; Frohlich et al., 2016; Hornbach et al., 2016; Scales et al., 2017). The spatiotemporal correlation between seismicity and fluid injection, supported by numerical pore pressure modeling, suggests that perturbations in pore pressure may stimulate failure on pre-existing faults when the faults are critically stressed and optimally oriented relative to regional estimates of maximum horizontal compressive stress (Figure 3.1a; Frohlich, 2012; Hornbach et al., 2015; Lund Snee and Zoback, 2016; Quinones et al., 2018; Ogwari et al., 2018; Zhai and Shirzaei, 2018; Horne et al., 2020), though other triggering mechanisms continue to be explored (e.g., Chen et al., 2018; Fan et al., 2019).
As the earthquake epicenters lie close to and within a metropolitan area (Figure 3.1), the seismic hazard assessment can benefit from an improved understanding of the observed ground motions. Earthquake ground motions and resultant damage are dependent upon propagation path attenuation from the medium properties and local site amplification. Previous studies suggest complex path attenuation and strong site effects in sedimentary basins (e.g., Hauksson and Shearer, 2006; Ahmadzadeh et al., 2019; Drwila et al., 2019). In the Mississippi sedimentary embayment over the New Madrid Seismic Zone (NMSZ), propagation path attenuates P-waves more than S-waves (Pezeshk et al., 2018) with local site amplification due to the thick unconsolidated sediments (1-2 km) 3-7 times that of stations located on hard-rock (Sedaghati et al., 2018). Like the embayment, the FWB geologic characteristics include a crystalline basement covered by a deep sedimentary basin (approximately 3.7 km) (Montgomery et al., 2005; Pollastro et al., 2007), which motivates this study of path attenuation and site amplification.

Kinematic source parameters for induced earthquakes, such as stress drop, source radius and moment also contribute to seismic ground motions and in turn to the seismic hazard. Some previous studies suggest that stress drops accompanying induced earthquakes are smaller than those of natural tectonic earthquakes (e.g., Hough 2014, 2015; Sumy et al., 2017) and that lower apparent stress drop may be linked to pore pressure changes (Abercrombie and Leary, 1993). Other studies do not find reduced stress drop for induced earthquakes (e.g., Goertz-Allmann et al., 2011; Justinic et al., 2013; Huang et al., 2016; Wu et al., 2018; Fan and McGuire, 2018). If, when compared to tectonic earthquakes, induced earthquakes exhibit systematically lower stress drop, then one might expect less high-frequency energy, and reduced seismic hazard relative to tectonic earthquakes. In the FWB, preliminary studies of source properties for earthquakes near DFW airport (Reiter et al., 2012) and Cleburne (Justinic et al., 2013) focused on a limited
number of earthquakes (Figure 3.1a) and did not directly address the impact of local site effects on these estimates. Thus, the effects of path attenuation and local site amplification across the sedimentary basin on estimates of source properties, with the continued debate about stress drop for induced earthquakes, motivates this comprehensive study of injection-induced earthquake source parameters in the FWB.

The generalized inversion technique (GIT) as introduced by Andrews (1986) has been used to provide estimates of source-receiver path effects, near-surface site characteristics and source properties. The method has subsequently been improved and applied to earthquakes with a wide range of magnitudes and tectonic settings, including events in Mexico (Castro et al., 1990), Romania (Oth et al., 2008, 2009), Japan (Oth et al., 2011; Oth, 2013), China (Wang et al., 2017) and New Zealand (Oth and Kaiser, 2014) as well as earthquakes comparable in size to those in the FWB. Pacor et al. (2016) and Ameri et al. (2011) characterized magnitude 3.0 to ~5.5 aftershocks of the 2009 L’Aquila earthquake (central Italy); Ahmadzadeh et al. (2017) investigated smaller earthquakes ($M_L$ 1.5-4.4) using frequencies from 0.5 to 35 Hz in Western Iran; and Picozzi et al. (2017) reported on induced-earthquakes ($M_W$ 2-3.8) in the Geysers geothermal area, California. In this study, GIT is applied to characterize the ground shaking from the FWB earthquake sequences ($M_L$ 2.0-3.9).

GIT requires site or source constraints in order to resolve the undetermined degree of freedom implicit in separating these effects (Andrews, 1986). One approach is to utilize a hard-rock reference site to minimize model trade-offs. Unfortunately, no hard-rock recording sites exist in the FWB. To address this issue, we used the approach of Moya et al. (2000) to correct the site response at a reference station within the basin. The site effect was then estimated by
subtracting a Brune’s source model (Brune, 1970, 1971) from the path-corrected spectra with the assumption that source spectra follow a \( \omega^{-2} \) spectral shape.

We used the GIT with the reference site correction to separate path, site and source characteristics for the induced earthquakes near Azle-Reno, Irving-Dallas and Venus in the FWB (Figure 3.1a). Seismic attenuation functions were parameterized in terms of geometrical spreading and \( Q \) factor. The site transfer functions estimated by GIT were compared to horizontal-to-vertical spectral ratio (HVSR) site responses, which were estimated directly from the observed spectra using the same number of earthquake records as were used in the GIT process. This comparison was done to assess the applicability of the two approaches in a sedimentary environment. Finally, source spectra were interpreted using the \( \omega^{-2} \) source model (Brune, 1970, 1971) providing kinematic source estimates that constrain source scaling over the narrow range of magnitudes for the FWB earthquakes (\( M_L \) 2.0-3.9).

3.2 Geology and Tectonics in the Fort Worth Basin, Texas

The FWB is a north-south elongated sedimentary basin associated with orogenic structures resulting from collisional tectonics during the later Paleozoic in north central Texas (Montgomery et al., 2005). The basin is approximately 63,000 km\(^2\) (Figure 3.1a). Important structures include the major Ouachita thrust-fold to the east and the high angle normal faults of the Newark East Fault Zone (NEFZ) trending northeast (NE)-southwest (SW) (Figure 3.1; Ewing, 1990; Magnani et al., 2017; Horne et al., 2020). Many of the reactivated faults host earthquakes with normal-faulting mechanisms consistent with NE-SW trending faults that parallel the NEFZ (Figure 3.1a; Justinic et al., 2013; Quinones et al., 2018). The generalized geologic column for the basin includes a relatively thick section of preserved Mesozoic and
Paleozoic sedimentary strata with structures, such as faults, rooted in the underlying Precambrian crystalline basement (see Magnani et al., 2017). The sediments reach a maximum depth of approximately 3.7 km (Montgomery et al., 2005) divided into three intervals: (1) Ordovician carbonates with a thickness of 1150-1400 m (Ellenburger, Viola, and Simpson groups) overlying the Precambrian crystalline basement; (2) Mississippian and Lower-Middle Pennsylvanian shales (Barnett group), and carbonates with a thickness of 120-300 m; and (3) Middle Pennsylvanian-lowermost Permian strata of 1000-1800 m thickness (Bend, Wichita, and Strawn groups) deposited as interlayered marine carbonate and deltaic fluvial siliciclastic sedimentary rocks (Pollastro et al., 2007). Above the Paleozoic strata, Cretaceous coastal plain sediments unconformably overlay the eastern portion of the basin (Alsalem et al., 2017). The NE-SW trending basement faults do not offset units younger than 300 Ma (Magnani et al., 2017) and extend both through the Ordovician Ellenburger unit, primarily used for wastewater injection, and the Mississippian Barnett Shale, used for gas production. Focal depth estimates for the FWB earthquakes range from 2 to 8 km, which corresponds to the Precambrian crystalline basement and the Ellenburger formation (see Quinones et al., 2019 for a summary).

### 3.3 Datasets

In response to the earthquakes, Southern Methodist University (SMU) installed multiple temporary seismic networks with 3-component short-period, broadband, and strong motion sensors across the FWB to supplement the existing regional seismic network surrounding the basin (Figure 3.1b; Frohlich et al., 2011; Justinic et al., 2013; DeShon et al., 2018). Due to the rapid deployment following felt earthquakes in an urban setting, the sensors were installed on building floors or in backyards within the uppermost layer of the sedimentary basin (DeShon et al., 2018). Thus, there were no hard-rock reference sites within the deployments.
We analyzed three well-recorded and well-located earthquake sequences that occurred after significant seismic network expansion in 2013 (DeShon et al., 2018; Quinones et al., 2019). After 2017, deployment of the Texas Seismic Network (TexNet) provided additional broadband sensors in the basin (Savvaidis et al., 2019) useful to this study. The sample rates of the data from these networks are 100 or 200 sample/s. We distinguished stations by combining location code and station name (e.g., AZLE__ and AZLE01, see Table 3.1) in order to track changes in sensors or configurations over time. Ninety earthquakes (2.0 < \( M_L \) < 3.9) recorded on 66 different stations are included in the study; the Azle-Reno, Irving-Dallas, and Venus sequences yield 1203, 1037 and 1185 measurements, respectively, for \( S \)-waves on horizontal components and \( P \)- and \( S \)-waves on vertical components. Individual recordings were selected by visual inspection of seismograms with an average signal-to-noise ratio (SNR) >3 between 5 and 25 Hz, which is the high SNR frequency band empirically determined by Quinones et al. (2018). Figure 3.2a shows the distribution of selected records from each station sorted by chronological event order.

To investigate the low-frequency capability of the network’s short-period sensors (mostly Mark Products L28 with a 4.5 Hz low cut frequency; see Table 3.1), the average spectral ratio between a short-period station, AZDA01, and the closest broadband station, FW0600 (separation of ~150 m), was plotted in Figure 3.2b after instrument correction. Since the spectral ratio was estimated using recordings from the same earthquakes and paths, these effects are eliminated. The empirical ratio indicates that the low-frequency amplitudes recorded by both the short-period and broadband sensors are similar and approach one from 1 to 10 Hz. Slight variations at frequencies above 10 Hz may reflect subtle site effect differences between the two stations; AZDA01 is closer to a road and creek than station FW0600. As a result of this successful
comparison, low-frequency data from the network’s instrument corrected short-period sensors were used in the subsequent analysis. Based on the result, the lower limit of the analysis frequency band was determined to be 1 Hz.

Earthquake locations and magnitudes were taken from the North Texas Earthquake Study (NTXES) catalog (Quinones et al., 2019). The hypocenters are based on manual $P$ and $S$ arrivals and a 1D velocity model developed with data from a 5 km local borehole to capture the local geology. The source-to-receiver ranges include both upgoing and downgoing arrivals providing robust depth estimates (Figure 3.2c). The local magnitude methodology was designed to be consistent with the Advanced National Seismic System (ANSS) Comprehensive Catalog (ComCat) magnitudes. The Venus $M_W$ 4.0 event is estimated to be $M_L$ 3.9 in the NTXES catalog.

In order to perform spectral analysis, we applied instrument response corrections and selected time windows, $T$, for $P$- and $S$-waves as suggested by Ottemöller and Havskov (2003):

$$OT + r/V_{\text{max}} \leq T \leq OT + r/V_{\text{min}},$$

(3.1)

where $OT$ is origin time, $r$ is hypocentral distance in km, and $V_{\text{max}}$ and $V_{\text{min}}$ are maximum and minimum velocity to provide an adequate window length. In equation (3.1), $V_{\text{max}}$ and $V_{\text{min}}$ are unknown. $V_{\text{max}}$ was estimated from the travel time of $P$ and $S$ waves, which was either calculated using arrival time directly picked by analysts or estimated using the Taup-toolkit (Buland and Chapman 1983; Crotwell et al., 1999) with the same velocity models used to locate the earthquakes (Quinones et al., 2019). $V_{\text{min}}$ was chosen using $V_{\text{max}} - 1.0$ km/s for $P$-waves and $V_{\text{max}} - 0.6$ km/s for $S$-waves based on group velocity range for the western United States proposed by Taylor (1996). $P$- and $S$-wave windows were tapered with a 10% cosine function with the window starting before the arrival time by 10% of the total extracted window length.
Typical windows ranged from 3 to 6 s for $S$-waves (min: 3 s and max: 14 s) and 1 to 4 s for $P$-waves (min: 1 s and max: 5 s); minimum window lengths ensure adequate spectral resolution at the lower frequencies. Zero-padding was applied to increase spectral resolution. Signals were differentiated to acceleration in order to minimize numerical effects at high frequencies. The Fourier amplitude spectra were estimated and smoothed with a 1.0 Hz moving window on a logarithmic scale. Noise spectra were computed for the same length window before the $P$ onset time. The two horizontal $S$-wave spectra were combined into a root-mean-square average. $P$-wave spectra were estimated using the vertical component. Signal processing was implemented in the Seismic Analysis Code software (Goldstein et al., 2003).

Example seismograms with estimated window lengths at local (hypocentral distance $r = 8$ km) and near regional ($r = 62$ km) distances are shown in Figure 3.3. Compared to local seismograms (Figures. 3.3e,g) regional seismograms (Figures. 3.3a,c) are more complex, due to a combination of direct and reflected arrivals. Local seismograms on the vertical component show scattered waves including $P$-coda and $S$-$P$ converted phases observed as $S$-wave energy before the $S$-wave arrival time (Oth et al., 2008, 2009; Hrubcová et al., 2016; Huang et al., 2016) (Figure 3.3g). As a result, in some cases $P$-wave windows may include these secondary arrivals although the shorter windows minimize these effects. Thus, a minimum time window length is an unavoidable compromise in order to maintain spectral resolution while suppressing the secondary arrivals. The right column in Figure 3.3 displays the resulting spectra. In Figure 3.3b, impulses observed at high frequency above 40 Hz in both signal and noise spectra may be local site noise. In Figure 3.3f, spectral shapes change slightly above 30 Hz, an observation similar to the source finiteness effects reported by Fan and McGuire (2018) for similar-sized Oklahoma
earthquakes. As a result of these spectral complexities, kinematic source interpretations will be limited to below 25 Hz for S-waves and below 40 Hz for P waves.

3.4 Methodology

3.4.1 The Generalized Inversion Technique (GIT)

The generalized inversion technique (GIT) was introduced to separate frequency-dependent propagation path effects, site response and source characteristics in observed Fourier amplitude spectra (e.g., Castro et al., 1990; Oth et al., 2008, 2009; Wang et al., 2017). Following the original two-step GIT (Castro et al., 1990), a one-step GIT was developed in order to produce a more stable separation of the three contributions for earthquakes in Japan (Oth et al., 2011). The one-step GIT has subsequently been applied to a range of earthquake sizes including small, induced earthquakes (e.g., Ameri et al., 2011; Pacor et al., 2016; Picozzi et al., 2017).

Assuming the earth can be represented as a linear filter, the instrument corrected seismic spectrum \( U(f) \) can be represented in the frequency domain as:

\[
U_{ij}(f, r_{ij}) = A(f, r_{ij})I_j(f)S_i(f),
\]

where \( A(f, r_{ij}) \) is the distance-dependent seismic attenuation, \( I_j(f) \) is the site response, and \( S_i(f) \) is the source spectra for the \( i \)th earthquake at the \( j \)th station. \( r \) is the source-to-receiver distance in km and \( f \) is frequency. Equation (3.2) can be linearized by taking logarithms of both sides:

\[
\log_{10} U_{ij}(f, r_{ij}) = \log_{10} A(f, r_{ij}) + \log_{10} I_j(f) + \log_{10} S_i(f),
\]

which conforms to \( Gm = d \), where \( d \) is the data vector, \( m \) is the vector of model parameters and \( G \) is the system matrix relating data and model vectors, represented in matrix form below:
In equation (3.4), the attenuation \( A \), site \( I \) and source \( S \) terms are stacked vertically and separated by the horizontal dashed lines in the far right-hand vector. In the system matrix, the left, middle and right submatrices separated by the two vertical dashed lines are consistent with the attenuation parameters \( A \), site \( I \) and source \( S \) characteristics. \( w_1 \) is the weighting parameter to constrain amplitudes such that \( A(f, r_0) = 1 \) where \( r_0 \) is the reference distance corresponding to the average focal depth of 4 km for the FWB earthquakes.

We tested three distance bins in the parameterization with widths of 0.5, 2, and 5 km covering distances from 4 to 100 km (Figure 3.A1) in order to determine the optimum width for attenuation \( A \) estimation. If the bins are chosen too large, the smoothing constraint suppresses spatial variations and leads to an inappropriate attenuation function (Ameri et al., 2011). As a result of this analysis, we selected a 2 km bin width to estimate nonparametric attenuation functions (Figure 3.A1). \( ND, NS \) and \( NE \) denote the number of distance intervals (4-100 km with 2 km bin width), stations and earthquakes (i.e., \( ND = 49, NS = 66, \) and \( NE = 90 \)). The \( w_2 \) is the degree of smoothness in the solution. The \( w_3 \) is the weight used to resolve the one undetermined degree of freedom between site effects and source spectra (Andrews, 1986). Previous studies have addressed this trade-off by fixing a single reference rock site or the average of a set of rock sites to be unity. To determine an appropriate reference condition, we ran a test inversion using a
single site and one using the average of all sites as a site constraint in the GIT for both horizontal and vertical components (Figure 3.A2). Site AZDA01 was chosen as the single reference site for GIT analysis because the station deployment duration covers all sequences (Figure 3.2a). From the test, we found that site amplifications derived from both the single site and all site reference are roughly similar to each other. In addition, we observed that when compared to HVSR the site amplification on horizontal components is smaller, suggesting a necessity of reference site correction, while vertical amplification estimates are well constrained and no site correction required (discussed in the Appendix A). Consequently, as suggested by Pacor et al. (2016), we implemented GIT on vertical components by constraining all sites to unity, while for the horizontal component we use a single reference site (i.e., AZDA01) with the site correction method introduced in the next section.

Finally, a least-squares solution $\mathbf{m} = (\mathbf{G}^T \mathbf{G})^{-1} \mathbf{G}^T \mathbf{d}$ was formed for each of 40 log space frequency bins in the 1 to 40 Hz band. In order to evaluate the stability of the inversion, we performed 100 bootstrap analyses at each frequency point following Parolai et al. (2000, 2004) and estimated the mean and standard deviation from these iterations.

3.4.2 Site Correction for Reference Station

Since AZDA01 is not a hard-rock site, a site effect needs to be incorporated into the GIT estimates. Bindi et al. (2004) suggest the use of HVSR as a correction for the reference site under conditions similar to this study. HVSR is generally used to assess site amplification under the assumption that the vertical component does not have strong site amplification. Previous studies suggest that HVSR and GIT produce similar estimates of peak frequencies, but that HVSR underestimates site amplification when there is vertical site effect (e.g., Ameri et al., 2011; Oth et
To account for this situation, we introduce a technique to correct site effects for the reference station.

To estimate reference site effects, we adopted an approach from Moya et al. (2000), which assumes that all sources follow the Brune’s source model (i.e., $\omega^{-2}$ shape) and estimate the site effect as the residual between path-corrected spectra and the best fit of the source model to the spectra. First, we performed an initial GIT estimate using spectra for AZDA01 without a site correction. Based on equation (3.2), for the reference station $j = ref$, the path-corrected spectra $U_{iref}^{corr}(f, r_{iref})$ are estimated by dividing the instrument corrected spectra recorded at the reference site $U_{iref}(f, r_{iref})$ by the initial path attenuation $A(f, r_{ij})$ as:

$$U_{iref}^{corr}(f, r_{iref}) = \frac{U_{iref}(f, r_{iref})}{A(f, r_{ij})} = S_i(f)I_{ref}(f),$$

where $I_{ref}(f)$ is the reference site effect. Equation (3.5) can be reorganized as:

$$I_{ref}(f) = \frac{U_{iref}^{corr}(f, r_{iref})}{S_i(f)}.$$  

We estimate $S_i(f)$ with a Brune source model (Brune, 1970, 1971) in acceleration, $S_i^{Brune}(f)$, using the following equation:

$$S_i^{Brune}(f) = (2\pi f)^2 \Omega_0 / (1 + (f / f_c)^2),$$

where $\Omega_0$ is the source low-frequency level and $f_c$ is the corner frequency. In order to obtain an appropriate estimate of $S_i^{Brune}(f)$, we used a nonlinear inversion using the trust-region-reflective (TR) optimization in MATLAB (command lsqnonlin, Moré and Sorensen, 1983), a Newton step-based method designed to find a global minimum with upper and lower limits that exhibits better performance than a Levenberg-Marquart search (Berghen, 2004). The Snoke
approach (Snoke, 1987), a mathematical method to estimate source parameters using
displacement and velocity energy fluxes, was followed for initial input estimates for the
nonlinear inversion. The upper limit of $\Omega_0$ is constrained by the local magnitude from the NTXES
catalog where the local magnitude is assumed to be consistent with moment magnitude $M_w$ and
the relationship $M_w = \frac{\log_{10} M_0 - 9.05}{1.5}$ from Hanks and Kanamori (1979) is used to retrieve seismic
moment $M_0$. $\Omega_0$ is proportional to seismic moment as:

$$\Omega_0 = \frac{M_0 F}{4\pi \rho c^3 r_0}. \quad (3.8)$$

The average double-couple radiation coefficient $F$ is taken as 0.52 and 0.63 for $P$ and $S$
waves (Boore and Boatwright, 1984), $r_0$ denotes the reference distance (= 4 km), density, $\rho$, is
2.68 g/cm$^3$ based on Trigg Well No. 1 data (Geotechnical Corporation, 1964) (Figure 3.1b), and
$c$ is velocity taken from the 1D velocity model used for the hypocenter estimate. The lower limit
was set to $\Omega_0/2$ or $\Omega_0/3$ depending on the earthquake. Limits for $f_c$ are set from 1 to 25 Hz. The
final form of the estimated reference site effect is written as:

$$I_{ref}(f) = \frac{U_{iref}^{corr}(f, r_{iref})}{S_{Brune}^B(f)}. \quad (3.9)$$

Ninety $I_{ref}(f)$ are estimated from all earthquakes in the dataset and averaged to obtain
the final reference site response. The true corner frequency for each earthquake remains
unknown, and each $S_{Brune}^B(f)$ represents an apparent corner frequency. The averaged reference
site response from the 90 earthquakes was used to minimize the possible errors introduced by the
use of the apparent corner frequency rather than the true corner frequency. Using the site-
corrected AZDA01 spectra, we then calculated new estimates of the path, site and source
contributions using the one-step GIT scheme. Note that Oth et al. (2011) suggest that the one-
step GIT provides the same attenuation functions regardless of the reference condition imposed by either the site response or source spectra in their sensitivity tests. We compared attenuation functions derived by GIT using non-site corrected and site-corrected AZDA01 spectra and observed consistent path attenuations from both GIT processes (Figure 3.A3). We chose to apply the single station approach for the remainder of the study, in part to complete the proof of concept example for a region with no hard-rock site.

For reference site AZDA01, Figure 3.4 shows GIT site amplifications for S-wave horizontal components (GIT H) with ± one standard deviation (gray shaded area) estimated from all 90 earthquakes, GIT V (GIT site responses calculated from vertical S-wave), the ratio of GIT H and GIT V (GIT HV), and HVSR. Below 4 Hz, GIT H variation is large but at higher frequencies variation is relatively small. Both average GIT H and GIT HV estimated site amplifications are similar to those estimated using HVSR, consistent with the relatively stable and flat GIT V. Although above 10 Hz there is a slight amplification on GIT V, HVSR and GIT HV are within one standard deviation of GIT H. The agreement between GIT H and HVSR associated with the negligible amplification on GIT V indicates that GIT H provides an appropriate reference condition for the GIT process. Note that since we use the Brune’s source model to correct for reference site effects, the inverted source spectra are expected to follow the $\omega^{-2}$ shape.

3.5 Results

3.5.1 Propagation Path Attenuation

Using the reference site correction developed in this study, we estimated nonparametric attenuation functions using the GIT approach. The inverted nonparametric attenuation curves for
S-waves at four distinct frequencies are displayed in Figure 3.5. The standard deviation estimated from the bootstrap analysis is small, suggesting that the attenuation functions are well constrained. At distances within 30 km, the attenuation functions decay rapidly with distance and then flatten at greater distances, a possible effect of a complex propagation path. Previous studies suggest that Moho depth ranges from 35 to 40 km in Texas (Prewitt, 1969; Wang, 2013). Based on these depths, one would expect the attenuation transition distance to be from 50 to 60 km or 1.5 times of crustal thickness (e.g., Wang et al., 2017; Burger et al., 1987). The shorter 30 km attenuation transition distance observed for the FWB may be indicative of reflections and refractions from shallower crustal interfaces (e.g., Bindi et al., 2004; Ameri et al., 2011).

In general, attenuation functions monotonically decay with increasing frequencies (e.g., Ahmadzadeh et al., 2017). In this study, the attenuation functions for S-waves observed on horizontal and vertical components at frequencies above 25 Hz exhibit decreased decay relative to lower frequencies (Figures 3.5 and 3.6b,d); the attenuation curves for 25 Hz and below exhibit roughly monotonic decay with frequencies (Figure 3.6a). For the P-waves, path-dependent attenuation curves show monotonically decaying shapes at high frequencies up to 40 Hz (Figure 3.6c). We compared two normalized spectra recorded at hypocentral distance of 4 km and 62 km in order to illustrate these differences/similarities (Figure 3.7). The normalized spectra are from AZDA01 (reference station) for the same local magnitude $M_L$ 2.8. Thus, the path effect should be the primary factor contributing to differences in the shapes of the two spectra. The local S-wave spectrum decays more rapidly above 25 Hz than the regional spectrum (Figure 3.7a). In contrast, the P-wave spectrum at a local distance decays similarly to the regional spectrum at high frequencies (Figure 3.7b). In order to analyze the high-frequency issue, we simply estimated an exponential slope of the high-frequency fall-offs between 25 and 35 Hz.
The slope of the local $S$-wave spectrum is $-0.1724$, steeper than the regional value of $-0.0903$ while the local $P$-wave shows a flatter value of $-0.0076$ than the value of $-0.0587$ from the regional $P$-wave spectrum despite the visually similar high-frequency fall-offs. The results of this analysis suggest that $S$-wave decay above 25 Hz may not completely reflect path and site effects and thus as noted earlier, these higher frequencies are removed from the subsequent analysis.

Parametric attenuation functions were estimated using the quality factor $Q$ and geometrical spreading. Before estimating the $Q$ factor, we performed a sensitivity test for attenuation transition distance using multiple distances ranging from 20 to 80 km (Figures 3.8 and 3.A4). These tests indicate that a transition distance between 30 and 50 km does not significantly impact the estimate of $Q$ factor. The frequency-dependent attenuation function $Q(f)$ was estimated using the following relationship with nonparametric function $A(f,r)$:

$$A(f,r) = G(r)\exp\left[-\frac{\pi f(r-r_0)}{Q(f)V}\right],$$

(3.10)

where $V$ is the velocity of $P$- or $S$-wave (5.8 km/s for $P$-wave and 3.3 km/s for $S$-wave) and $G(r)$ is the geometrical spreading assumed as $G(r) = r_0/r$ for both $P$-wave and $S$-wave. Taking the common logarithm at both side of equation (3.10) the equation becomes:

$$\log_{10} A(f,r) - \log_{10} G(r) = -\frac{\pi f(r-r_0)}{2Q(f)V},$$

(3.11)

where $\log_{10}(e) = 1/2.3$. $Q(f)$ can be computed as a slope from a linear least-square fit to the left-hand side of equation (3.11), residual between nonparametric attenuation and geometrical spreading. Since the quality factor $Q$ trades-off with the assumed geometrical spreading, both parameters must be used together.
$Q$-models for $P$- and $S$-waves were estimated as $Q_P(f) = 9f^{0.90}$ and $Q_S(f) = 20f^{1.01}$ with the same geometrical spreading within 30 km. The parametric attenuations follow a linear trend with frequency for both body waves (Figure 3.8). The empirical ratio between $Q_S$ and $Q_P$ is greater than 1 from 1 to 25 Hz, suggesting higher attenuation for $P$-waves than $S$-waves.

For north Texas, Cramer et al. (2014) estimated a regional $S$-wave attenuation model $Q_S(f) = 274f^{0.70}$ with a geometrical spreading of $r^{-0.5}$. The regional attenuation is weaker than local attenuation estimated within the FWB reported here (Figure 3.8). Quinones et al. (2019) also suggest that the attenuation curve using peak amplitudes within 50 km denotes higher attenuation than the regional out to 450 km. The low $Q$'s may be related to the shallow layers in the sedimentary basins that dominate the wave paths in this study (Hauksson and Shearer, 2006; Drwiła et al., 2019).

3.5.2 Site Response

After extracting path effects, the resulting site responses for the $S$-wave horizontal components (GIT H) were compared to HVSR for 12 stations in Figure 3.9. All sites in Figure 3.9 reveal decreased amplitude starting at 10 Hz for both HVSR and GIT H site effect estimates. For sites where GIT V is approximately 1 across the frequency band, GIT H and HVSR produce consistent site effect estimates (AZWR__, FW0900, FW1100, and IPD1V01). GIT V is not 1 across all frequencies at a number of sites (AZNH01, IFCF00, AZHS01, AFDA__, AZLE01, IPD1__, AFDA01, and AZLE__). For cases where GIT V is not 1, site response estimates using GIT H and HVSR diverge and the shapes of GIT HV site estimates are much closer to HVSR than GIT H (Figure 3.9), consistent with previous studies (e.g., Oth et al., 2009; Ameri et al., 2011; Oth et al., 2011; Ahmadzadeh et al., 2019). GIT site effects for all seismic stations are included in the Appendix A (Figures 3.A5-3.A7).
These results illustrate the importance of assessing both GIT V and GIT H in order to completely quantify site effects as opposed to relying on estimates such as HVSR. GIT V may depart from 1 as result of $S$-$P$ conversions, 2D or 3D effects at the edge of sedimentary basin, and presence of $P$-wave contamination in the $S$-wave window (Ameri et al., 2011). In this study, the $S$-$P$ converted phases are observed on the vertical component of the local seismogram, as shown in Figure 3.3g. Thus, the presence of $P$-waves in the $S$-wave window can contaminate the shape of vertical $S$-wave spectra.

To demonstrate the significance of site amplification in the basin, we show the maximum amplification of site transfer functions from 1.5 to 20 Hz (Figure 3.10). The maximum GIT H range from 1.6 to 19.8 with a mean of 6.1 while the maximum amplification from HVSR range from 2.0 to 23.4 with mean 5.8. Standard deviations of GIT H and HVSR are 3.7 and 2.9, respectively. A significantly large amplification in HVSR is observed for LLRK02, which may result from a strong de-amplification of GIT V (Figure 3.A7). For GIT H, larger amplifications of 15 or greater are observed at three stations (Figure 3.10a) (AFDA__, AZLE01, and IPD1__ shown in Figure 3.9). Over the course of the deployments, these sites had changes in instrumentation (to AFDA01, AZLE__ and IPD1V01 shown in Figure 3.9) that produce site amplifications from 6.4 to 6.7 and are more consistent with other stations across the network. This result provides insight into the differences in spectra that can be recorded at the same site but using different instrumentation.

Fortunately, at IPD1 we have recordings of the same earthquakes (15 events on IPD1__ and IPD100 and 3 events on IPD100 and IPD1V01). Using these earthquakes, we estimated direct spectral ratios (same plot with Figure 3.2b) and GIT H ratios for IPD1__/IPD100 and IPD100/IPD1V01 using instrument-corrected spectra (Figure 3.11). Although all contributions
(path, site, and source effects) are thought to be corrected for the direct spectral ratio, the maximum amplitudes between 1 and 10 Hz are 1.87 and 1.33 for IPD1__/IPD100 and IPD100/IPD1V01, respectively. For GIT H ratios, the maximum amplitudes are 2.31 and 1.86 for IPD1__/IPD100 and IPD100/IPD1V01, respectively, with shapes consistent with those of the spectral ratios. Consequently, the IPD1__ spectra are 3 or 4 times larger than those of IPD1V01.

We cannot unwrap factors affecting the differences in spectra recorded on various sensors at the same place. Based on the agreement between the GIT H ratio and spectral ratio, however, we suggest that the trade-offs generated from the GIT do not produce the amplitude gap between different seismometers at IPD1.

Since these outliers can increase the mean value, we took median values for GIT H and HVSR, which are 5.1 and 5.5 from GIT H and HVSR (Figure 3.10). As a result, the average site amplification factor in FWB is about 5. The FWB site amplification is close to that observed in the Mississippi embayment (Sedaghati et al., 2018). Similarly, Ahmadzadeh et al. (2019) observe maximum site amplification factor as large as 7 for a station installed in a basin with water-saturated layers and an average amplification of 2.4 for 9 bedrock sites based on GIT H site responses in Alborz, Iran. The larger amplifications may reflect thick sediments (e.g., Sedaghati et al., 2018) and near-surface sensor installations (Oth et al., 2011).

3.5.3 Source Spectra and Source Parameter

The nonparametric source contributions are easily identified in the path and site corrected earthquake source spectra. Using the source spectra, we estimated source parameters (e.g., seismic moment, rupture radius and stress drop). The seismic moment $M_0$ is estimated from $\Omega_0$ in equation (3.8). Stress drop, $\Delta\sigma$, is calculated from the relation that scales the seismic moment by source radius, assuming a circular crack model (Eshelby, 1957):
\[
\Delta \sigma = \frac{7M_0}{16R_0^3}, \quad (3.12)
\]

\[R_0 = \frac{K_0 \beta}{2\pi f_c}, \quad (3.13)\]

where \(K_0\) is a constant depending on fault geometry, 2.01 for \(P\)-wave and 1.32 for \(S\)-wave, assuming a rupture velocity of \(0.9 \beta\), where \(\beta\) is shear wave velocity (Madariaga, 1976). The best parameters and 95% confidence intervals are estimated for corner frequency and moment magnitude using the nonlinear inversion with the TR algorithm (Moré and Sorensen, 1983).

Figure 3.12 displays four \(S\)-wave acceleration source spectra using GIT and the standard deviations obtained from a 100 iteration bootstrap analysis. The mean spectral shapes are consistent with the Brune model, which satisfy the prerequisite for the reference site correction that source spectra follow the \(\omega^{-2}\) shape. The source model for the small earthquake (\(M_L 2.0\)) produces a higher corner frequency of \(13.90 \pm 0.45 \text{ Hz}\) compared to the fit for the larger earthquake (\(M_L 3.3\)) with a corner frequency of \(5.13 \pm 0.14 \text{ Hz}\) (Figures 3.12a,b).

In the FWB, \(S\)-wave stress drops range from 1.18 to 21.73 MPa with a mean stress drop 4.46 MPa (Figure 3.13a). The mean stress drop is comparable to average stress drop estimates of \(~1 \text{ MPa}\) for the DFW International Airport earthquakes using coda spectral method (Reiter et al., 2012) and 4.3 MPa for Cleburne earthquakes using a Snoke time domain method (Justinic et al., 2013), as well as average stress drops for intraplate earthquakes from 1 to 10 MPa (Kanamori and Anderson, 1975). These kinematic source estimates suggest that within the errors of the estimates, the slip properties of these injection-induced earthquakes are no different from those of natural earthquakes.
In order to explore earthquake scaling, corner frequencies were plotted against seismic moment on a logarithmic scale (Figure 3.13b). We took all events with $M_w$ 2.0 or greater, which yields a total magnitude range of 1.6. For $S$-waves, a least-square linear fit of log corner frequency to log moment has a slope of $-0.28$ (Figure 3.13b), which is close to the value of $-0.33$ for self-similarity. Earthquakes with magnitudes above 3 show a slight increase in stress drop with the seismic moment for the total dataset (Figures 3.13a,b). Boyd et al. (2017) observed a slope of about $-1/3$ in the United States with the level changes of stress drop at $\sim M$ 4. Wu et al. (2018) suggest similar scaling patterns for stress drops from induced-earthquake sequences in Oklahoma. Although the range of magnitudes is limited, the resulting stress drops as a function of $M_w$ are consistent with self-similar scaling. As a result of this limited magnitude range extrapolating these results to larger magnitudes must be done with caution.

Usually stress drop is statistically interpreted due to the large uncertainties that accompany the estimates (Allmann and Shearer, 2007). Previous global studies have shown that stress drop varies over at least three orders of magnitude (e.g., Shearer et al., 2006; Allmann and Shearer, 2007, 2009). In the FWB, we observed that the resulting $S$-wave source spectra retrieved from the GIT for the same local magnitude document some variation in estimated corner frequencies while the estimated moment magnitudes are approximately identical (Figures 3.12c,d). The total variability of stress drop estimates is more than one order magnitude (Figures 3.13a,c). These new results exhibit smaller variability than the global scale studies over larger magnitude ranges with a log-normal distribution (Figure 3.13c). Corner frequencies ranged from 3.1 to 24.2 Hz. This result is consistent with estimates from previous studies, 8-11 Hz for earthquakes ($M$ 2.0-2.5) in the DFW (Reiter et al., 2012) and 3-18 Hz for earthquakes ($M$ 2.0-2.8) in the Cleburne (Justinic et al., 2013). Figure 3.13d documents that the relationship between
local magnitude and moment magnitude estimated from the source spectra is in good agreement, but the distribution seems to be separated into two groups, which are located on upper and lower parts of the one-to-one trend. This effect will be explored in the discussion section.

*P*-wave stress drops were estimated and compared to those from *S*-waves (Figure 3.14). For spectral analysis of *P*-waves, we used no site correction under the assumption that vertical site effects are small, consistent with many of the GIT V estimates. Using this approach, the estimated stress drops of *P*-waves were approximately 3 times larger than those from *S*-wave (Figure 3.14a). Figure 3.14b plots the *P*-wave corner frequencies against moment with a similar slope of $-0.27$. Corner frequencies estimated from *P*-waves are larger than those from *S*-wave data. The average $P/S$ corner frequency ratio is 1.4, which is consistent with a theoretical value of 1.5 derived by Magariaga (1976). A ratio of 1.5 is estimated for shallow and small tectonic earthquakes (M 0.1-3.7) in Italy (Zollo et al., 2014) and the average ratio of $P/S$ corner frequencies is reported as 1.2 for induced earthquakes in Oklahoma (Huang et al., 2016). The histogram of logarithmic stress drops for *P*-waves also shows a log-normal distribution similar to that of stress drop estimates from *S*-wave although for *P*-waves the variability is slightly larger (Figure 3.14c). The moment magnitude estimated from *P*-wave deviates (~0.38) from local magnitude (Figure 3.14d) and suggests a larger moment than those estimated from *S*-wave based on the observation in Figure 3.13d. Under the hypothesis that converted phases could increase the estimated seismic moment for *P*-wave, we compared two *P*-wave spectra estimated using short time windows (0.5 s) to minimize inclusion of coda waves. The longer window produced larger spectral estimates from 10 to 20 Hz (Figure 3.A8). At low frequencies where the moment is estimated (Figure 3.A8) the amplitude difference is small and we conclude that converted phases may not in fact cause the larger moments calculated from *P*-wave data. The differences
between $P$ and $S$ moments and stress drops may be related to the assumption of no $P$-wave site effect and may warrant further investigation. Using a $P$-wave amplitude correction of 3.0 produces $P$-wave seismic moments that match the local magnitude (Figure 3.A9). The resulting average stress drop is 4.95 MPa with individual values ranging from 0.49 to 45.26 MPa, comparable to the stress drops estimated for the $S$-waves. The estimated corner frequencies before and after reducing the $P$-wave-derived moments are nearly the same.

3.6 Discussion

Using GIT with site corrections for soft rock sites, we calculated attenuation functions, site transfer functions, and source spectra from the data recorded by 66 temporary stations for 90 shallow-depth induced earthquakes in the FWB. In this discussion, we first focus on unique features noted in the nonparametric attenuation functions and moment magnitudes. We then broaden the discussion to consider the characteristics of induced earthquakes and the sedimentary basins in which they occur. Finally, we discuss the appropriateness of this study’s reference site correction method.

The empirically characterized 30 km transition distance for nonparametric attenuation functions (Figures 3.5 and 3.6) is interpreted to be due to contributions of reflections or refractions from a crustal boundary located above the Moho. Using regional refraction data, Keller and Hatcher (1999) imaged a mid-crustal boundary between 20 and 22 km along the Ouachita thrust. Similarly, using first-arrival times from FWB stations, such a mid-crustal boundary was estimated at 18 km in North Texas (Frohlich et al., 2011; Quinones et al., 2019). Based on the observations and models, we suggest that the mid-crustal boundary, which should be a regional feature along the Ouachita thrust front, may be responsible for the observed transition distance in the attenuation curves in north Texas.
For S-waves, we observe attenuation complexity at frequencies above 25 Hz (Figures 3.5 and 3.6). This high-frequency decay above 25 Hz does not follow general attenuation theory (Figure 3.7). Similarly, for injection-induced earthquakes, complex spectral shapes and complex rupture have been observed (e.g., Fischer, 2005; Holmgren et al., 2019). In Oklahoma, with the largest number of induced earthquakes in the central United States, Fan and McGuire (2018) observe spectral complexity that includes double humps in source spectra (similar to Figure 3.3f in this study) and suggest that multiple subevents contribute to the complex spectral shapes. Wu et al. (2019) use numerical modeling to conclude that four subevents accompany the 2015 $M_w$ 4.0 Guthrie earthquake in Oklahoma. These observations and models suggest that induced earthquakes may have complicated rupture histories (Fan and McGuire, 2018). Since we cannot uniquely assess the reason for the S-wave spectral anomalies above 25 Hz observed in this study, further work possibly complemented by numerical models, on the high-frequency characteristics is warranted (e.g., Fischer, 2005; Wu et al., 2019).

The ratio of the quality factor of $P$- and $S$-wave, $Q_s/Q_p$, is larger than 1 for the FWB estimates (Figure 3.8), which can be interpreted as the impact of partial fluid saturation in the crust (Hauksson and Shearer, 2006; Zollo et al., 2014). The higher $P$-wave attenuation than $S$-wave attenuation is often observed in sedimentary basins. We note, however, that the absolute value of the attenuation ratio $Q_s/Q_p$ is dependent on the chosen geometrical spreading (Morozov, 2008) and scattering attenuation characterized by the scale of heterogeneity (Yoshimoto et al., 1993). That said, in the NMSZ, Pezeshk et al. (2018) observe $Q_s > Q_p$, assuming $1/r$ geometrical spreading for frequencies from 4 to 24 Hz and suggest that the crust beneath the NMSZ is partially fluid-saturated. In the Delaware basin in west Texas, also an area with induced seismicity, Drwiła et al. (2019) observe that $P$-waves are attenuated more than $S$-waves using
frequency independent attenuation in the peak frequency method and suggest that the result is consistent with attenuation of a saturated sedimentary basin. The presence of fluid along the path is one of several possible mechanisms impacting the observed $P$- and $S$-wave attenuation. A seismic survey performed in the FWB demonstrates a spatial correlation between the high production areas and high attenuation regions (Li et al., 2016) further suggesting that a hydraulically fractured Barnett Shale layer may also contribute to scattering attenuation.

In the analysis of $P$-wave source parameters, we found larger $P$-wave moments than those of $S$-waves (Figure 3.14d). As noted earlier the assumption that there is no vertical $P$-wave site effect may influence these estimates. An additional possibility is that a constant $P$-wave velocity, $V_P$, adopted from well logs may lead to overestimation in cases where $V_P$ near the source has changed due to wastewater injection activity. Laboratory experiments by Barrière et al. (2012) report that both $V_P$ and $Q_P$ decrease in a non-consolidated porous media with water saturation while the $S$-wave is weakly dependent on the saturating fluid. In the FWB, the 2005-2017 cumulative injected volume into the Ellenburger formation was $\sim$318 million m$^3$ (2 billion U.S. barrels) (Quinones et al., 2019). The fluid is co-produced with gas from the Barnett Shale and some production wells may also access brines within the underlying Ellenburger (Gao et al., 2019). It is estimated that 70-90% of fluid from the Barnett Shale layer is moved into the Ellenburger formation and that some Ellenburger brine recycling also contributes to the total wastewater injection volumes (Gao et al., 2019). However, due to a lack of study to support this assumption, the effect on seismic velocity changes associated with wastefluid injection is a possible subject of research in the FWB.

In Figures 3.13d and 3.14d, the distribution of $M_L$ as a function of $M_W$ shows significant variation. The variation appears to be separated as two groups of events based on 1:1 line. With
the hypothesis that the number of stations can impact magnitude resolution, we plot the residual between $M_l$ and $M_w$ as a function of the number of stations used to make the estimates (Figure 3.15a). The events with a small number of observations primarily produce positive values while events with a large number of observations (>10) produce dominantly negative residuals. In order to investigate the cause of the issue, we compared two event clusters to $m_{blg}$ estimated from the ANSS ComCat (Figures 3.15b,c). Note that only 29 events in ANSS ComCat are identical to the events used in this study. Using the ComCat data, we found a similar dependence on the number of stations between local magnitude and $m_{blg}$ while GIT estimated $M_w$ shows no such trend with $m_{blg}$. We suggest that the $M_l$ to $M_w$ dependence on station distribution may result from the local magnitude estimate provided by the NTXES catalog. We infer that site amplification may affect the difference between $M_l$ and other magnitudes. $m_{blg}$ is estimated from vertical motion, assuming limited site effects (Rigsby et al., 2014) while $M_w$ provides site-corrected seismic moments.

Since the technique introduced to estimate the site correction for a non-hard-rock reference station relies on a Brune’s model estimate, the observed variation in local magnitude and the use of apparent corner frequency may affect the site transfer function for the reference station (e.g., the large error bounds for low frequencies (< 4 Hz) shown in Figure 3.4). By using events covering $M_l$ from 2.0 to 3.9 for the 90 earthquakes, we used data from a range of sizes and average the effects of corner frequencies and moments on the estimate. The effectiveness of this assumption is illustrated by comparing the GIT H and HVSR site estimate for the reference station AZDA01 (Figure 3.4). Theoretically, HVSR should not be affected by source effects as HVSR is calculated from the same source using the horizontal and vertical components. The
similarity of GIT H and HVSR for AZDA01 suggests that the effect of corner frequency misfit and local magnitude variation are minimized by averaging across the events.

3.7 Conclusions

Near-source spectral analyses of P- and S-waves from earthquakes in three named induced earthquake sequences within the FWB covering a magnitude range for $M_l$ from 2.0 to 3.9 are reported. Recordings on 66 temporary stations from 90 earthquakes are used to investigate attenuation characteristics, local site response functions and source parameters using a nonparametric one-step GIT with a site correction method developed in this study.

Attenuation curves document consistent propagation effects to distances of 30 km, where wave propagation complexities possibly consistent with reflected or refracted waves from mid-crustal interaction are expected within the FWB. The ratio of the quality factor estimates for P- and S-waves is larger than one. This ratio may reflect the effect of partially fluid-saturated materials.

Site transfer functions document strong site amplifications (a median value of 5) for most of the stations. This effect may result from the thick sedimentary basin. In addition, we find that vertical S-waves document site amplification for several sites. Thus, the basic assumption for application of HVSR fails for these sites. Estimated GIT H, GIT V and GIT HV are consistent with this interpretation.

Mean stress drop estimated from S-wave source spectra is 4.46 MPa ranging from 1.18 to 21.73 MPa, similar to those for tectonic earthquakes and previous earthquakes in the FWB. Stress drops show a very slight departure from self-similarity primarily due to enhanced values for the largest events above magnitude 3.0+. However, the narrow 1.6 magnitude range in this
study limits the ability to extrapolate to larger magnitudes. Stress drops are normally distributed on a logarithmic scale with less variability than those estimated from global earthquake studies. The corner frequency ratio between $P$- and $S$-wave is approximately 1.4, which is similar to the theoretical value of 1.5. $P$-wave derived stress drops are 3 times greater values than those from $S$-waves. However, because the moment magnitudes derived from $P$-wave data are 3.8 larger than those derived directly from a local magnitude relationship, we suggest that larger moment estimates from $P$-wave are leading to the larger $P$-wave derived stress drop values. The larger $P$ moment may either be indicative of a change in $P$-wave velocity in and around the source region as injection has continued or due to the assumption that vertical $P$-waves have no receiver effect.

The source properties estimated from $S$-waves suggest that strong motions and seismic hazard from the injection-induced earthquake will be similar to those for tectonic intraplate events. The effect of basin sediments, mid-crustal interface, $Q_s/Q_P > 1$ and large $P$-wave moment analyzed in this study are thought to be important factors contributing to the local seismic propagation.
Figure 3.1. (a) Map illustrating the earthquakes (red circles) across the Fort Worth Basin in Texas. Characteristic focal mechanisms for each sequence are from Justinic et al. (2013) and Quinones et al. (2018). The primary tectonic features and regional faults mapped in the injection volume and crystalline basement (solid black lines) are from Horne et al. (2020): NEFZ: Newark East Fault Zone. Cities or facilities (black circles) are used to name the earthquake sequences. Plus symbols indicate wastewater-injection wells. The map inset illustrates the extent of the Barnett Shale (gray shaded area). (b) Network map showing locations of strong motion (triangles), broadband (squares), and short-period (diamonds) stations. Some broadband and short-period sensors are installed at nearly identical locations. Maximum horizontal stress ($S_{H\ max}$) orientation (red bar) is from Lund Snee and Zoback (2016). The black star denotes the location of Trigg Well No. 1 (Geotechnical Corporation, 1964). The figure was produced using the Generic Mapping Tools (GMT) (Wessel et al., 2013).
Figure 3.2. (a) Temporal distribution of seismograms from the 66 temporary stations recording the 90 events in this study. Data ranges from 01/2014 to 02/2018. The reference station AZDA01 (black symbols) recorded all sequences, while other stations (gray) recorded portions of the total dataset. The two horizontal sold lines separate the short-period sensors (top), strong motion stations (middle) and broadband sensors (bottom). The two vertical lines indicate the first event of Irving-Dallas and Venus sequences in this dataset. (b) The average spectral ratio between the near co-located short-period AZDA01 and broadband FW0600 based on 15 earthquakes (EQs) following instrument correction. The spatial separation (∆r) between the two instruments is ~150 m. Gray shaded area indicates ± one standard deviation. (c) Distribution of local magnitude with hypocentral distance in km.
Figure 3.3. Examples of accelerograms for an Irving-Dallas earthquake on 01/20/2015 ($M_L = 3.1$, depth 5.7 km), showing horizontal (EW and HN3) and vertical (Z) components (left column) and the respective spectra (right column). (a)-(d) Regional data ($r = 62$ km, where $r$ is hypocentral distance) are recorded at AZDA01 (short-period). (e)-(f) Local data ($r = 8$ km) are recorded at NLKCP01 (strong motion). In left column, horizontal lines above the time series represent time windows for $P$-wave (gray) and $S$-wave (black). In right column, the arrow in (f) indicates the starting frequency of spectral shape change. Horizontal components are used to estimate $S$-wave spectra while vertical components are used to calculate both $S$-wave (black curve) and $P$-wave spectra (gray curve). Dashed spectra denote noise.
Figure 3.4. Site amplification functions for station AZDA01 used as a reference site. White continuous curve is the mean GIT site amplification for horizontal (GIT H) while the gray shaded area denotes the mean ± one standard deviation estimated from the 90 earthquakes (EQs). Black continuous, dashed lines and black line with circles are HVSR, vertical S-wave amplification (GIT V) and horizontal-to-vertical ratio between GIT H and GIT V (GIT HV), respectively. Both GIT H and GIT HV are in approximate agreement with HVSR due to the effect of GIT V.
Figure 3.5. Nonparametric attenuation functions (black lines) derived for four different frequencies using the horizontal $S$-waves. The gray shaded area denotes the mean ± one standard deviation calculated using bootstrap resampling. Black asterisks are the spectral amplitudes estimated after correcting for the site and source contribution. Black dashed lines indicate geometrical spreading $G(r) = 1/r$. 
Figure 3.6. Nonparametric attenuation functions as a function of hypocentral distance for all 40 frequencies, color-coded in legend. Frequencies above 25 Hz are represented as black thin dashed curves. Black bold dashed lines illustrate geometrical spreading $G(r) = 1/r$. The attenuation curves for: (a) Horizontal $S$-waves with frequencies from 1 to 25 Hz; (b) Horizontal $S$-waves with a frequency range from 1 to 40 Hz; (c) $P$-waves and (d) vertical $S$-waves with frequencies from 1 to 40 Hz.
Figure 3.7. Example observed displacement spectra at local (black lines) and regional distance (gray lines) recorded at AZDA01 station for (a) S-waves and (b) P-waves. The amplitudes are normalized in order to compare the spectral shapes. The two events have the same magnitude ($M_L$ 2.8) but are from different sequences (Azle-Reno and Irving-Dallas) and hypocentral distances (4 and 62 km for local and regional distances). Vertical dashed lines represent 25 Hz.
Figure 3.8. $Q(f)$ model for both horizontal $S$-wave (circles) and $P$-wave (triangles) derived from the nonparametric attenuation curves in Figure 3.6. The least-square fits are denoted as black lines. Distances are restricted to 4-30 km in the analysis. For the horizontal $S$-waves, the frequency band is 1 to 25 Hz due to high-frequency variability shown in Figure 3.6b. For comparison, regional $Q_s(f)$ obtained by Cramer et al. (2014) is plotted.
Figure 3.9. Site response functions estimated using GIT compared to HVSR for 12 stations. The GIT site effects on horizontal (GIT H) (white lines) and vertical (GIT V) (dashed lines) components, and the horizontal-to-vertical ratio (GIT HV) (black line with circles) are plotted with HVSR (black lines). The gray shaded area denotes the mean ± one standard deviation of the GIT H from the bootstrap analysis. The station name with location code and the number of earthquakes (EQs) used are noted in each plot. See Figures 3.A5-3.A7 in the Appendix A for all stations.
Figure 3.10. Histogram of maximum amplification estimates for (a) GIT H and (b) HVSR. The maximum values are estimated for frequencies ranged from 1.5 to 20 Hz. The bin width is one amplification unit.
Figure 3.11. Comparison between the average spectral ratio (black line) directly estimated from instrument-corrected spectra (that is similar to Figure 3.2b) and GIT H site amplification ratio (black line with circles) for (a) IPD1__ and IPD100 and (b) IPD100 and IPD1V01. Gray shaded area indicates ± one standard deviation of the spectral ratio.
Figure 3.12. Four examples fits between inverted source spectra (white lines) with ± one standard deviation from bootstrap analysis (shaded area) and the Brune source model (black lines). Estimated corner frequency (inverted triangles) and $M_w$ are written in the plot with the 95% confidence intervals of the parameters. The two examples in (a) and (b) are from a small ($M_L 2.0$) and large ($M_L 3.3$) event. The two earthquakes in (c) and (d) have the same local magnitude ($M_L 2.4$). The earthquakes in (c) and (d) have nearly the same $M_w$, but significantly different corner frequencies.
Figure 3.13. Stress drop estimates from the three sequences (Azle-Reno, Irving-Dallas and Venus) for $S$-waves. (a) Logarithmic stress drop versus moment magnitude $M_W$. White squares are the mean values for 0.1 magnitude bins with error bars denoting ± one standard deviation. The diagonal dashed line in the upper left represents the corner frequency limit of 25 Hz. (b) Relationship between corner frequency $f_c$ with the error bars and seismic moment $M_0$. Black dashed lines represent a constant stress drop relation. The least-square fit has a slope of $-0.28$. (c) Histogram of stress drops. (d) Comparison of $M_W$ derived from source spectra with local magnitude. Black continuous line is regression prediction when the slope is fixed to 1. Black dashed lines indicate 95% confidence interval.
Figure 3.14. Same as Figure 3.13 but for $P$-waves. (a) The diagonal dashed line in the upper left represents the corner frequency limit of 30 Hz. Mean stress drop and its variation are larger than estimates from $S$-waves while other stress drop properties shown in (b) and (c) are consistent with the results from Figure 3.13. (d) The estimated moment magnitudes from $P$-wave are 0.38 larger than the local magnitude.
Figure 3.15. Comparison of three magnitudes ($M_L$, $M_w$, and $m_{bLg}$). (a) Residual between $M_L$ and $M_w$ estimated by GIT versus the number of stations used in this study. Asterisks reveal the positive values and circles are negative residual. (b) $M_L$ versus $m_{bLg}$ provided by ANSS ComCat and (c) $M_w$ versus $m_{bLg}$ for 29 events. In (b) and (c), asterisks are the events with seismic records from less than 10 stations while circles denote the events that have more than 10 stations.
### Table 3.1. Stations and sensors used in this study.

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APPENDIX A

The supporting information provides further details on: Test results documenting the effects of bin widths of 0.5, 2, and 5 km on estimation of nonparametric attenuation; Comparison of site amplifications estimated from the generalized inversion technique (GIT) using a single site and an average of all sites as a reference condition; Comparison of nonparametric attenuation functions estimated from GIT using a non-site corrected and site-corrected AZDA01 as a reference; Assessment of the effect of assumed distance range on the estimate of the $Q$ factor; The entire set of amplification functions estimated from the GIT for horizontal (GIT H), horizontal-to-vertical ratio (GIT HV), and vertical (GIT V) for $S$-wave; Comparison of spectral estimates using short and long time windows for $P$-waves; and Stress drops for $P$-waves after magnitude adjustment.

Test of Bin Width in Nonparametric Attenuation Estimates

Estimates of nonparametric attenuation require the selection of a spatial bin width to construct the system matrix in the GIT process. Here, we test bin widths of 0.5, 2, and 5 km. The nonparametric attenuation functions estimated from 0.5 and 2 km bin widths at 1 Hz produce similar attenuation estimates (Figures 3.A1a,b). In contrast, the nonparametric attenuation using a 5 km bin width deviates from the others with lower resolution for the closest data (Figure 3.A1c). Estimates of quality factors from the three-bin widths produce similar values for 0.5 and 2 km widths, but $Q$ values for the 5 km bin width are as much as a factor of 2 higher (Figure 3.A1d). Since the study utilizes data from the closer distance ranges, the 0.5 and 2 km bin widths
are thought to be more appropriate. We choose the 2 km bin width for ND in equation (3.4) as the attenuation curve for the 0.5 km bin produces fluctuations near 60 km while the attenuation curve for the 2 km bin is smoother (Figures 3.A1a,b).

**Comparison of GIT Results Using a Single Site and An Average of All Sites as References**

GIT results are illustrated using a single reference site (AZDA01) and an average of all sites (Figure 3.A2). Two example stations (FW0900 and IFCF00) are used for illustration. Site amplification estimated using the single site constraint are slightly larger than those using the station average as a reference with both estimates roughly similar to one another for horizontal (GIT H) and vertical site amplification (GIT V). The ratio of GIT H and GIT V (GIT HV) is smaller than HVSR. This test result suggests that either GIT H is underestimated or GIT V is overestimated. GIT V for FW0900 is near 1 as a function of frequency consistent with the expectation of small site effects on vertical components. In contrast, the GIT H is significantly underestimated for both FW0900 at low frequency (< 10 Hz) and IFCF00 at high frequency (> 7 Hz) compared to HVSR. Based on these observations, we interpret that vertical site amplifications are well constrained from both the single site and average site references but the horizontal components may need some improved reference site estimate in order to provide an improved GIT H estimate for consistency with HVSR. Thus, we developed the site correction method introduced in Chapter 3.4.2.

**Comparison of Initial Attenuation Estimates and Site-Corrected Attenuation Estimates**

In order to check the independence of attenuation function from the site constraint as suggested by Oth et al. (2011), attenuation estimates were made using both site amplifications estimated using non-site corrected GIT and site-corrected GIT based on AZDA01 (Figure 3.A3).
The two attenuation functions are identical and therefore, we conclude that the attenuation curves are independent of reference conditions used in the one-step GIT.

**Assessment of the Effect of Assumed Distance Range on the Estimate of the $Q$ Factor**

The effect of assumed maximum distance range on the estimate of $Q$ factors is investigated. Five different distance limits of 20, 30, 40, 50, and 80 km are applied to $P$- and $S$-wave data with a minimum distance of 4 km (Figures 3.8 and 3.A4). Results estimated using transition distances of 40 and 50 km are consistent with the $Q$ factors using 30 km, which is chosen in this study (Figure 3.8). The $Q_S$ value estimated from 4 to 20 km is slightly higher than $Q_S$ from 30 km. The $Q$ factor associated with a distance of 80 km is a factor of 2 larger than that for 30 km.

**Comparison of GIT Site Amplification and HVSR**

Many stations in this study produce similar frequency-dependent site amplification for GIT $H$ and HVSR (Figure 3.A5), but GIT HV site estimates improve the comparison to HVSR (Figure 3.A6). Observed site amplification on the vertical component (Figure 3.A7) controls either the agreement or disagreement between GIT $H$ and HVSR. Note that several stations (e.g., AZHL00, IFSC00, and VSAB00) have a very limited number of waveforms due to temporary deployments but still show good agreement between the GIT curves and the HVSR.

**Comparison of Spectral Estimates Using Short and Long Time Windows for $P$-waves**

In order to investigate the effect of window length, two $P$-wave spectra are estimated from a short time window (0.5 s), which minimizes coda waves, and a longer window (1.0 s) (Figure 3.A8). The spectrum with the longer window produces larger spectral estimates from 10 to 20 Hz, but the amplitude difference is small at low frequencies where the moment is estimated
Thus, time window length appears to not strongly affect the moment estimates from P-waves.

**Stress Drops of P-wave After Magnitude Adjustment**

We document stress drops from P-waves that are three times larger than those from S-waves (Figures 3.13-3.14) assuming no site effect for vertical P-waves. Since the corner frequency ratio between P- and S-waves is 1.4, which is similar to the theoretical value of 1.5, larger moment estimates for P-wave produce higher stress drops. We normalize the P-wave amplitudes by a factor of 3 in order to match P-wave seismic moments with the local magnitudes (Figure 3.A9). The resulting average stress drop is 4.95 MPa with individual values ranging from 0.49 to 45.26 MPa, comparable to the stress drops estimated for the S-waves. Thus, the stress drops of P- and S-waves are similar on average when the moments are normalized to local magnitude. As noted in Chapter 3.6 the cause of the enhanced P-waves moments is still a matter of additional research.
Figure 3.A1. Comparison of spectral amplitude estimated after correcting for the source and site contribution (black circles) and nonparametric attenuation curves at 1 Hz using three-bin widths of (a) 0.5 km (black), (b) 2 km (blue), and (c) 5 km (red). (d) Estimated $Q$ factors for each bin width.
Figure 3.A2. Site amplification estimated from GIT with a single reference site and average site constraints for FW0900 (top) and IFCF00 (bottom). Left, middle, and right columns denote the site amplifications on horizontal, vertical, and the ratio of horizontal and vertical components, respectively. For comparison, HVSR estimates are plotted (left and right columns).
Figure 3.A3. Nonparametric path attenuation estimated from (a) initial GIT using no site correction for AZDA01 and (b) using the site correction developed for AZDA01.
Figure 3.A4. Comparison of four distance limits (20, 40, 50, and 80 km) used to estimate $Q$ factors for $P$- and $S$-wave. The results of 40 and 50 km are consistent with 30 km.
Figure 3.A5. Site amplifications using HVSR (black line) and horizontal S-wave (GIT H) for all 66 stations. GIT H estimates include the mean (white curves) and mean ± one standard deviation estimated by 100 bootstrap analyses (gray shaded area) except for AZDA01 (reference site). The site response function and one standard deviation of AZDA01 are the same as Figure 3.4.
Figure 3.A6. Comparison of GIT site amplifications for the horizontal-to-vertical ratio of $S$-wave (GIT HV) (black line with circles) and HVSR for all 66 stations. HVSR estimates include the mean (white curves) and mean $\pm$ one standard deviation for each site (gray shaded area).
Figure 3.A7. Same as Figure 3.A6 but for GIT site amplifications for the vertical S-wave (GIT V) (dashed lines) and HVSR.
Figure 3.A8. (a) Same as Figure 3.3g, but with a shorter time scale. Horizontal lines above the waveform denote time windows for longer windows (black), which include coda waves, and shorter windows (gray) without the scattered waves. (b) Resulting spectra for $P$-waves estimated from the short (gray) and long windows (black). The longer-window spectrum produces larger amplitude estimates in the frequency band from 10 to 20 Hz, but low-frequency amplitudes are similar to one another.
Figure 3.A9. Same as Figures 3.13 and 3.14 but with the normalized moment for the $P$-waves (factor of 3). (a) Stress drops are slightly larger than those from $S$-wave, with a mean value of 4.95 MPa, which compares to 4.46 MPa for $S$-wave. The diagonal dashed line denotes a limit to corner frequency of 30 Hz. Stress drops shown in (b) and (c) are also similar to those of $S$-waves. (d) Moment magnitude is consistent with local magnitude in this case after normalizing by a factor of 3.
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CHAPTER 4

STRESS-DROP ESTIMATES FOR INDUCED SEISMIC EVENTS IN THE FORT WORTH BASIN, TEXAS

Abstract

Earthquakes in the Fort Worth Basin (FWB) have been induced by the disposal of recovered wastewater associated with extraction of unconventional gas since 2008. Four of the larger felt earthquakes, each on different faults, prompted deployment of local distance seismic stations and recordings from these four sequences are used to estimate the kinematic source characteristics. Source spectra and the associated source parameters including corner frequency, seismic moment and stress drop are estimated using a modified generalized inversion technique (GIT). As an assessment of the validity of the modified GIT approach, corner frequencies and stress drops from the GIT are compared to estimates using the traditional empirical Green’s function (EGF) method for 14 target events. For these events, corner frequency residuals (GIT − EGF) have a mean of −0.31 Hz with a standard deviation of 1.30 Hz. We find consistent mean stress drops using the GIT and EGF methods, 9.56 MPa and 11.50 MPa, respectively, for the common set of target events. The GIT mean stress drop for all 79 earthquakes is 5.33 MPa, similar to estimates for global intraplate earthquakes (1-10 MPa) as well as other estimates for induced earthquakes near the study area (1.7-9.5 MPa). Stress drops exhibit no spatial or temporal correlations or depth dependency. In addition, there are no time or space correlations between estimated FWB stress drops and modeled pore pressure perturbations. We conclude that
induced earthquakes in the FWB occurring on normal faults in the crystalline basement, release pre-existing tectonic stresses and that stress drops on the four sequences targeted in this study do not directly reflect perturbations in pore fluid pressure on the fault.

4.1 Introduction

Fluid disposal, production of geothermal fields, reservoirs formed by dams, and hydraulic fracturing are all known to be sources of human-induced seismicity (e.g., National Research Council [NRC], 2013). In the Fort Worth Basin (FWB), Texas, the rate of wastefluid disposal associated with the production of unconventional shale gas from the Mississippian Barnett shale formation increased significantly beginning around year 2005 and a commensurate seismicity rate increase in 2008 (Hennings et al., 2019). Previous studies suggest that FWB sequences are induced by stress changes caused by this disposal (e.g., Frohlich, 2012; Frohlich et al., 2016; Quinones et al., 2019). Between 2006 and 2018, more than 2 billion U.S. barrels (~318 million m$^3$) of fluid were injected into the Ordovician Ellenburger dolomite formation, which underlies the Barnett (Quinones et al., 2019). Many studies focused on the FWB have concluded that the increased pore fluid pressure or poroelastic stresses accompanying injection diffuses and ultimately reduces the shear strength of pre-existing faults to promote slip (Hornbach et al., 2015, 2016; Magnani et al., 2017; Zhai and Shizaei, 2018; Fan et al., 2019; Hennings et al., 2019; Quinones et al., 2019).

Although the FWB events have been small in magnitude ($M \leq 4$ and below), the close proximity to the large Dallas-Fort Worth (DFW) metropolitan area increases the risk of damage from possible future earthquakes (Figure 4.1a). In order to assess and mitigate the seismic hazard associated with injection-induced earthquakes, it is important to obtain accurate estimates of earthquake source parameters, such as source radius, moment, corner frequency and stress drop.
Furthermore, the source parameters can provide a better understanding of the similarities and differences between reactivated intraplate faults linked to wastewater disposal and faults in tectonically active seismic zones. Finally, source parameters provide constraints on rupture processes and possible maximum magnitude associated with seismic source scaling.

Previous studies suggest that stress drops for induced earthquakes broadly related to oil and gas extraction, geothermal sites and wastewater disposal are smaller than natural tectonic earthquakes (e.g., Fehler and Phillips, 1991; Abercrombie and Leary, 1993; Hough, 2014, 2015; Picozzi et al., 2017; Sumy et al., 2017; Trugman et al., 2017). Hough (2014) argues that stress drops estimated for induced earthquakes in the central and eastern United States based on empirical intensity data from felt reports are lower than expected. The lower stress drops might reflect conditions associated with fluid disposal such as pore fluid pressure changes or may be a function of the shallow focal depths (Abercrombie and Leary, 1993; Hough, 2014, 2015).

Stress drop estimates associated with induced earthquakes exhibit a range of spatiotemporal variations (e.g., Fehler and Phillips, 1991; Lengliné et al., 2014; Chen and Shearer, 2011; Goertz-Allmann et al., 2011; Agurto-Detzel et al., 2017; Picozzi et al., 2017; Sumy et al., 2017; Yoshida et al., 2017). Based on analysis of events at the Basel geothermal site, Switzerland, Goertz-Allmann et al. (2011) suggest that stress drop increases by about a factor of five with radial distance from the fluid injection wellhead and is inversely proportional to linear pore pressure perturbation within ~300 m of the injection point. Huang et al. (2017) and Trugman et al. (2017) argue for decreased stress drop with decreasing depth for shallow induced earthquakes (< 5 km below sea level) in the central United States (CUS). Chen and Abercrombie (2020) document temporal changes of stress drop, with lower stress drops at the start of the sequence indicating that pore pressure changes caused the weaker segments of the fault to fail
first. If induced earthquakes exhibit systematically lower stress drops then high-frequency energy from such sources should be reduced, possibly changing estimates of local ground shaking relative to tectonic earthquakes.

Increases in pore fluid pressure associated with injection might lead to fault weakening with slip (Malagnini et al., 2014), resulting in a possible break in earthquake self-similarity (Kane et al., 2011). In comparison, natural tectonic earthquakes have been found to be self-similar as stress drop scales with seismic moment (e.g., Abercrombie and Leary, 1993; Abercrombie, 1995; Ide and Beroza, 2001; Shearer et al., 2006; Allmann and Shearer, 2007, 2009; Baltay et al., 2011). While some injection-induced earthquake studies suggest constant stress drop as a function of earthquake size (Goertz-Allmann et al., 2011), others find stress drops from induced earthquakes that increase with magnitude (Fletcher, 1982; Abercrombie and Leary, 1993; Mandal et al., 1998; Lengliné et al., 2014; Trugman et al., 2017). Determining if self-similarity holds for earthquakes in a particular region is needed to correctly forecast potential maximum magnitudes and ground motions in areas where larger earthquakes may be possible.

In this study, we estimate source parameters using the Fourier amplitude spectra of S-waves recorded from FWB earthquakes occurring since 2009. Previous studies of FWB seismicity produce mean stress drop estimates ranging between ~1 and 4.5 MPa using a variety of techniques focused on individual faults (Reiter et al., 2012; Justinic et al., 2013; Jeong et al., 2020). These stress drops are comparable to average stress drops expected for intraplate earthquakes, typically from 1 to 10 MPa (e.g., Kanamori and Anderson, 1975). Here, two types of spectral analysis are implemented and compared in order to estimate source parameters. First, we use the generalized inversion technique (GIT) with site correction for a single reference
station following Jeong et al. (2020) to isolate the source spectra from the observed spectra. This modified approach separates source spectra from propagation path and site contributions in situations where there are no hard-rock sites with the assumption that source spectra follow the Brune-type $\omega^{-2}$ spectral model (Brune, 1970). Next, we test the assumptions of Jeong et al. (2020) by directly comparing GIT source spectra and corner frequency estimates with values produced using the empirical Green’s function (EGF) method. The EGF method has been applied in studies that report lower stress drops for induced earthquakes based on a fundamentally different theoretical approach to the calculation of source parameters (e.g., Agurto-Detzel et al., 2017), thus motivating the source comparisons in this paper for a common set of earthquakes. We also compare GIT seismic moments to $M_w$ derived from $m_{bLg}$, the $Lg$ body wave magnitude from the U.S. Geological Survey (USGS) Advanced National Seismic System (ANSS) Comprehensive Catalog (ComCat), using the empirical relation of Rigsby et al. (2014). These analyses provide a basis for assessing the appropriateness of path and site estimates intrinsic in the GIT approach. The comparative study provides insight into estimates of mean stress drop and uncertainty based upon methodology. These results can be used to test self-similarity, and possibly identify temporal-spatial variations that might produce unique characteristics of induced earthquakes relative to natural earthquakes.

4.2 Seismicity of the Forth Worth Basin, Texas

Many mid-magnitude and larger induced earthquakes ($M > 3.5$) in the CUS are linked to wastefluid disposal associated with oil and gas production (Ellsworth, 2013; Rubinstein and Mahani, 2015). Frohlich et al. (2016) report that the earthquake rate for magnitude 3 and above across Texas has increased from about 2 events per year to 12 events per year since 2008. In the FWB, unconventional oil and gas extraction techniques applied to the Barnett shale produce
wastewater, or brine, that is re-injected into the deeper Ellenburger dolomite formation, which is an ~1 km thick unit whose top lies at 2.00 to 2.74 km (Pollastro et al., 2007; Smye et al., 2019). Earthquake sequences in the FWB have occurred on faults rooted in the underlying Precambrian crystalline basement that extend upward into the Ordovician Ellenburger and Mississippian Barnett formations (Magnani et al., 2017), with most felt earthquakes occurring in the basement (Quinones et al., 2019). The ComCat and Texas Seismic Network report all events above magnitude 2.0 in the basin while the North Texas Earthquake Study (NTXES) networks provide data to lower magnitudes along specific targeted faults (Figure 4.1; DeShon et al., 2018; Quinones et al., 2019). Here, we focus on four, well-studied earthquake sequences recorded by these local seismic networks: 1) Cleburne from June 2009 to June 2010 (Justinic et al. 2013); 2) Azle-Reno beginning November 2013 (Hornbach et al., 2015); 3) 2014-present Irving-Dallas sequence (Magnani et al., 2017; Quinones et al., 2018); 4) the Venus sequence with the largest $M_w$ 4.0 (ComCat) May 2015 earthquake (Scales et al., 2017).

Earthquakes occur on steeply dipping normal faults in the crystalline basement or in limited cases within the injection unit (see Quinones et al., 2019 for a summary). Reported focal depths range from 2 to 8 km (Figure 4.2). Focal mechanisms are consistent with normal faulting along the north-northeast (NNE) to south-southwest (SSW) trending faults (Figure 4.1a; see Justinic et al., 2013 and Quinones et al., 2018), consistent with modern maximum horizontal stress ($S_{H_{max}}$) orientations (Fig. 1b; Lund Snee and Zoback, 2016). Most seismicity occurs on regional faults optimally oriented for failure in the modern stress region, with the notable exception of the N-S striking Cleburne fault (Figure 4.2; Hennings et al., 2019). The Azle-Reno, Irving-Dallas, and Venus sequences include multiple larger earthquakes (magnitude 3.5+), whereas the 2009-2010 Cleburne earthquakes range from magnitude 2.0 to 2.9. Of the sequences
studied here, the Cleburne, Azle-Reno and Venus sequences occur within 5 km of one or more wastewater disposal wells while the Irving-Dallas sequence is more than 10 km from an injection well (Figure 4.2).

4.3 Datasets

Ninety-five earthquakes ($M_L \geq 2.0$) that occurred in the FWB from 2009 to 2018 and recorded on 72 different broadband, strong motion and short-period stations are analyzed (Figure 4.1). The datasets are a combination of 90 earthquakes from the Azle-Reno, Irving-Dallas, and Venus sequences used in Chapter 3 and 5 earthquakes in Cleburne. Hypocentral parameters for the Cleburne sequence are taken from Justinic et al. (2013), using ComCat magnitudes for the 5 events. Detailed station information of Cleburne stations is documented in Justinic et al. (2013).

Data processing to produce spectrum from time series is the same with those from Chapter 3. In this study, we only use $S$-wave to analyze the source properties of FWB earthquakes. Additionally, in the Appendix B, we investigate influences of the 1-Hz smoothing window on Fourier spectral estimates (Figure 4.B1). The result suggests that the smoothing window produces a good agreement between unsmoothed and smoothed spectra at frequencies above 1 Hz, which is the low-frequency limit of this study.

Examples of three-component seismograms documenting the window length and resulting spectra are shown in Figure 4.3. As discussed in Jeong et al. (2020) and Chapter 3, scattered energy and strong $S$-wave contributions on the vertical component before the $S$-wave arrival time are observed on some local records as a result of $P$-coda and shallow $S$ to $P$ conversions (Figure 4.3e). Increased variability in spectral estimates above 25 Hz (Figures 4.3b,d,f) is also observed, possibly an effect of complex rupture propagation and thus these higher frequencies are excluded from additional processing.
4.4 Methodology

4.4.1 The GIT for Cleburne Sequence

We use a modified GIT to correct site effects for a reference station in order to isolate the earthquake source spectra (Jeong et al., 2020). The details of the method are documented in Chapters 3.4.1 and 3.4.2.

Since the seismic network changed as a function of time, we perform two GIT inversions. First, 90 earthquakes from the Azle-Reno, Irving-Dallas, and Venus sequences are used in the GIT process with the reference site AZDA. Results are discussed in Chapter 3 and Jeong et al. (2020), which document seismic attenuation characteristics and site effects in the FWB. We use the path functions calculated from the initial GIT analysis as the propagation characteristics for the FWB (Figure 4.4a). After correcting for these path effects, a second GIT is carried out for the 5 events recorded on 6 stations from the Cleburne sequence. For the Cleburne GIT, we select the station CLEF2 as a reference site based on low site amplification on the vertical components (GIT V) (Figure 4.4b). The horizontal site spectra estimated by GIT (GIT H) are consistent with horizontal-to-vertical spectral ratios of observed spectra (HVSR) at low-to-intermediate frequencies while GIT H decays slightly at high frequency, possibly reflecting near-surface site attenuation (Oth et al., 2011). The ratio between GIT H and GIT V (GIT HV) is not much different from GIT H, suggesting that GIT V is close to unity. All site effects for the Cleburne sequence are documented in Figure 4.B2. 100-bootstrap inversions are performed to estimate the uncertainty in the GIT estimates following Parolai et al. (2000, 2004) (see Figure 4.B2). The average of all site effects for vertical components is constrained to unity as suggested by Pacor et al. (2016) under the assumption that vertical site effects are small.
4.4.2 Stress Drop Calculation

Source corner frequency, seismic moment and stress drop are extracted from the isolated source spectra by applying the following source model:

\[ S(f) = \Omega_0 \left(1 + \left(\frac{f}{f_c}\right)^n\right)^{1/\gamma}, \]

where \( \Omega_0 \) is the constant low-frequency level, \( f_c \) is the corner frequency, and \( f \) is frequency. \( f \) ranges from 1 to 25 Hz, and then the frequency band is adjusted by an analyst for both the EGF and GIT methods. \( n \) is the high-frequency falloff rate and \( \gamma \) is a parameter to control the sharpness of the decay with frequency in the spectrum around the corner frequency. \( \gamma = 1 \) and \( n = 2 \) follows Brune (1970) and \( \gamma = 2 \) and \( n = 2 \) with the sharper decay follows Boatwright (1980).

The seismic moment \( M_0 \) is estimated from \( \Omega_0 \) using:

\[ M_0 = \frac{4\pi\rho c^3 r_0^2 \Omega_0}{F}, \]

where \( F \) is the average double-couple radiation coefficient, 0.63 for \( S \) waves (Boore and Boatwright, 1984), \( r_0 \) is the reference distance (4 km in this study), \( \rho \) is the density, 2.68 g/cm\(^3\) based on Trigg-Well Number 1 data (Figure 4.1b; Geotechnical Corporation, 1964), and \( c \) is velocity taken from the velocity model based on the earthquake hypocenter in the catalogs. The moment magnitude is estimated from the relationship, \( M_W = (\log_{10} M_0 - 9.05)/1.5 \), where \( M_0 \) is defined in newton meters, as described by Hanks and Kanamori (1979). Finally, stress drop and source radius are calculated with the process explained in Chapter 3.5.3.

4.4.3 The Empirical Green's Function (EGF) Method

The EGF method is an alternative approach that isolates source contributions by calculating the spectral ratio between a target earthquake and smaller EGF events under the
assumption that the path and site effects are the same for collocated events observed at the same receiver (e.g., Hough, 1997; Huang et al., 2016; Boyd et al., 2017; Shearer et al., 2019). Many induced earthquakes occur in clusters and so the EGF method has been widely used to estimate source parameters (e.g., Huang et al., 2016; Wu et al., 2018). In contrast, GIT has mostly been applied to broader study areas and larger earthquakes (e.g., Oth et al., 2011; Pacor et al., 2016). Comparison of GIT source spectra to estimates using the EGF method can be used to assess and possibly validate the completeness of the GIT approach applied to the FWB induced earthquake sequences.

Based on equation (3.2), a spectral ratio between a target event $U_1$ and a smaller EGF event $U_2$ is defined as:

$$\frac{U_1(f)}{U_2(f)} = \frac{A(f)\ell(f)S_1(f)}{A(f)\ell(f)S_2(f)} = \frac{S_1(f)}{S_2(f)}.$$  

(4.3)

The source spectral ratio, $\frac{S_1(f)}{S_2(f)}$, can be modeled using the ratios based on equation (4.1):

$$\frac{S_1(f)}{S_2(f)} = \frac{M_{01}}{M_{02}} \left( \frac{1+(f/f_{c2})^\gamma}{1+(f/f_{c1})^\gamma} \right)^{1/\gamma}$$  

(4.4)

where $M_{01}$ and $f_{c1}$ are the seismic moment and source corner frequency of the target event while $M_{02}$ and $f_{c2}$ represent the EGF event.

EGF events are selected based on four criteria. (1) The separation between target and EGF events should be no greater than 2 km. In this study, use of events with separation distances of 2 km or less does not lead to significantly different individual spectral ratios as illustrated by data examples from north Texas (Figure 4.B3). (2) Average spectral ratios are estimated based
on at least five stations recording one common target and at least three EGF events. (3) The magnitude difference between the target and the EGF events is greater than 0.8 magnitude unit. (4) We require the EGF earthquakes have average SNR greater than 5 in two frequency bands (1-5 and 5-25 Hz) to ensure clear spectral shapes. Note that this is a stricter criteria than applied to the GIT dataset (SNR > 3 between 5 and 25 Hz). Individual spectral ratios are stacked in order to average directivity effects as well as other source complexities (Ross and Ben-Zion, 2016). Kane et al. (2011) indicate that averaging multiple stations reduces stress drop uncertainties by 30% or more relative to single station estimates. In order to compare the source spectral shape between GIT and EGF methods, source spectra are subsequently derived from the EGF estimates using a Brune $\omega^{-2}$ spectral model following Uchide and Imanishi (2016) (Figure 4.5). First, the frequency band is set to 1.0 to 25 Hz for all recordings of EGF events used to estimate the corner frequency of the target event. This initial frequency range is then manually adjusted using the SNR criteria and spectral shape (Figure 4.5a). We note that the frequency band used for estimating the EGF corner frequency of the target event is different from the frequency band used for EGF event selection. For each EGF event, the spectral ratios estimated for individual stations are stacked and normalized on a log scale (Figure 4.5b). The multiple spectral ratios are then used to calculate a median value in order to produce the final spectral ratio for the target event. This spectral ratio is fit using equation (4.4) to estimate the model parameters ($f_{c1}$, $f_{c2}$, and $M_{01}/M_{02}$). The comparison spectrum between the data spectral ratios and the source model is subsequently estimated (Figure 4.5c). Finally, the product of the $\omega^{-2}$ model (Brune, 1970; Boatwright, 1980) in acceleration with $f_{c1}$ and the comparison spectrum (Figure 4.5c) yields the source spectrum (Figure 4.5d).
4.5 Results

GIT corrects for path and site effects in cases where small magnitude events, reduced earthquake rates, event dissimilarities or lack of permanent sites may limit the applicability of EGF or other relative spectral comparisons. The local seismic networks in the FWB were deployed following significant felt earthquakes such that many of the felt earthquakes do not have local seismic recordings, especially prior to ~2015 (DeShon et al., 2018). Application of GIT to data from the FWB produced 79 source spectra for events with corner frequencies below 25 Hz. These results are compared to source spectra estimates using the EGF method.

Only 11 target events larger than $M_L$ 3.1 have accompanying EGF earthquakes of at least $M_L$ 2.0 with adequate SNR meeting the criteria listed above. All are associated with the 2015 Irving-Dallas sequence, which was best captured by local seismic networks in the basin. In order to ensure that conclusions drawn from EGF analyses can be applied basin wide, we chose 3 additional events, one each from Cleburne, Azle-Reno, and Venus sequences, but have to carefully relax one or more of the EGF selection conditions. A target event in the Cleburne satisfies all EGF selection conditions except for the small magnitude ($M_L$ 2.4); thus, we have to take even smaller EGF events ($< M_L$ 2.0) from Quinones et al. (2019) catalog to maintain a 0.8 magnitude difference. We note that the Cleburne EGF events all have a SNR > 5 in frequency ranges 1-5 Hz and 5-25 Hz. The target event in Azle-Reno ($M_L$ 3.0) satisfies most EGF selection conditions, but was only recorded by 2 stations. The Venus target event (ComCat $M_w$ 4.0 mainshock) has 1 EGF event with a 2.8 km separation between the target and EGF events; this is because the mainshock was recorded by stations deployed in the Azle-Reno and Irving-Dallas area and has a higher location uncertainty compared to its EGF events recorded after the local
Venus network was deployed. Earthquake information and the source parameters of the 14 target events estimated using the EGF method are included in Table 4.1.

Since the EGF source spectra are derived from spectral ratios, the corner frequencies estimated directly from the spectral ratios are nearly equal to those from the EGF source spectra. Here, we take the EGF source parameters estimated from the EGF source spectra, which produce smaller 95% confidence intervals than those directly calculated from the spectral ratios. Figures 4.6a and 4.6b compare two example acceleration source spectra and corner frequencies estimated by both EGF and GIT. The shapes of the source spectra from the two methods are similar. For the 14 earthquakes used in this comparison, the corner frequency estimates from the GIT and EGF methods generally follow a line with a slope of 1 (Figure 4.6c) with error bars representing the 95% confidence interval estimates. Corner frequency residuals, calculated as GIT minus EGF, have a mean of $-0.31$ Hz ($\text{EGF } f_c > \text{GIT } f_c$) and a standard deviation of 1.30 Hz (Figure 4.6d). Note that the corner frequency residuals from the 11 Irving-Dallas target events that fully satisfy all EGF selection criteria had a mean of $-0.54$ Hz and a standard deviation of 1.29 Hz, which is not significantly different. When we estimate EGF stress drop using the EGF corner frequency and GIT seismic moment, the mean stress drop is 11.50 MPa, roughly consistent with GIT mean stress drop of 9.56 MPa for the 14 target events. These results suggest that the GIT correction isolates the source spectrum as effectively as the EGF method. There is scatter between the two estimates, as illustrated by the two source spectra in Figure 4.6b which shows an example producing a large corner frequency residual of $\sim 2.0$ Hz. In this instance, above the corner frequency the GIT source spectrum exhibits a slightly higher high-frequency falloff rate than the EGF source spectrum and results in a higher corner frequency estimate. Variations between the two methods may reflect data limitations for the EGF method where the selection
criteria limits the number of events in cases with lower rates of seismicity and limited spatiotemporal network coverage. Even though multiple stations are stacked in the EGF method, limited data can result in poor azimuthal coverage. The EGF process uses one to four fewer stations than the GIT process for the target earthquakes, and the Azle-Reno target event only had 2 stations in the stack (Table 4.1). In the case of GIT, all earthquakes and stations are used and therefore can under some circumstances produce more robust source estimates. Furthermore, if the EGF event corner frequency is not high enough (i.e., close to a corner frequency of target event), the secondary corner frequency can impact the estimate of the corner frequency for the target event (Shearer et al., 2019). Lastly, slightly different frequency bands used to fit the $\omega^{-2}$ to the spectral ratio and GIT source spectra may impact the resulting source spectra and corner frequency estimates.

Since GIT source spectra are estimated using the Brune source model, the EGF source spectra are interpreted with the same model. Huang et al. (2016) suggest that the EGF source estimates using the Boatwright model produce stress drop estimates that are 3 times lower than those estimated using the Brune model due to the limited bandwidth and fixed falloff rates. We compare corner frequencies calculated using both the Brune and Boatwright models in the interpretation of the EGF source spectra in Figure 4.7 (Boatwright spectral model on the y-axis). The EGF corner frequencies from the Boatwright model are systematically lower than those of Brune model but parallel the 1:1 line (Figure 4.7a). The mean residual in corner frequency (Boatwright − Brune) is −1.53 Hz with a standard deviation of 0.72 Hz (Figure 4.7b). Estimates of stress drop using the Boatwright model for the EGF corner frequencies and GIT seismic moments produce a mean value of 5.08 MPa, smaller than GIT stress drop estimates. Based on
these limited examples from the FWB, we infer that the Boatwright spectral model produces
source estimates 2 times smaller than estimates using the Brune model.

Moment magnitudes estimated from the GIT source spectra are compared to \( m_{bLg} \) magnitudes based on \( Lg \) arrivals from ComCat for the 34 larger events (Figure 4.8a). The resulting relation is \( M_W = 0.79m_{bLg} + 0.67 \) for a \( m_{bLg} \) range from 1.7 to 4.0. This relation provides an assessment of the moment estimates with comparison to the empirical relationship \( M_W = 0.67m_{bLg} + 1.03 \) for \( 2.0 < m_{bLg} < 4.3 \), for eastern north America (ENA) as suggested by Rigsby et al. (2014) (Figure 4.8a). The two relations are consistent with each other within the 95% confidence interval (Figure 4.8a). We conclude that the GIT \( M_w \) is comparable to \( M_W \) based on the scaling using ComCat \( m_{bLg} \) as suggested by Rigsby et al. (2014).

Corner frequencies are plotted against seismic moments estimated from GIT on a logarithmic scale to assess the source parameters (Figure 4.8b). Some events produce saturated corner frequencies (~25 Hz) in the moment magnitude range from 1.6 to 1.9. We therefore remove events below \( M_W \) 1.9 for subsequent analysis leaving a total of 79 earthquakes from the original 95. The mean and standard errors of the resulting stress drops are 5.04±2.18 MPa (Cleburne), 4.43±1.92 MPa (Azle-Reno), 6.53±1.79 MPa (Irving-Dallas), 5.46±2.78 MPa (Venus), and 2.58±1.69 MPa (other areas in the FWB not associated with the 4 sequences shown in Figure 4.1a). The mean stress drop for the total dataset is 5.33 MPa, with individual estimates ranging from 1.37 to 27.35 MPa. These values are consistent with the expected range of tectonic intraplate stress drops from 1 to 10 MPa. The detailed GIT source parameters are included in Table 4.2.

The slope of corner frequencies as a function of seismic moment for all FWB earthquakes is \(-0.29 \) (Figure 4.8b), approximating self-similar scaling with an expected slope of \(-1/3 \). Since
the total magnitude range is limited to 1.7 ($M_w$ 1.9 to 3.6), extrapolation to larger earthquake magnitudes could be problematic. High-frequency spectral complexity limits analysis to 25 Hz (Figure 4.3), which eliminates events below a moment magnitude of 1.9 based on corner-frequency saturation. High-frequency spectral variability has been observed in other source studies of larger induced earthquakes and interpreted as either a complex rupture history or strong directivity effects (e.g., Fan and McGuire, 2018; Holmgren et al., 2019; Wu et al., 2019). However, the cause of high-frequency spectral variability remains unclear in this study.

Spatial variations of stress drop are investigated by plotting each event scaled by its moment, the location of its causative fault (Quinones et al., 2019; Horne et al., 2020) and near-by injection wells to see if we observe any spatial patterns (Figure 4.9). The locations and details of individual injection wells are from the Texas Railroad Commission and documented in Table 4.3. These plots produce no clear spatial correlations between injection well location and stress drop for individual earthquakes. In the case of the Azle-Reno sequence, one high stress-drop event is located at the northeast end of the fault (Figure 4.9a) occurring late in the sequence (Figure 4.10), possibly reflecting stress concentrations on the edge of the fault (Oth and Kaiser, 2014) or a stress change with time, although the limited number of events makes this observation highly speculative. Irving-Dallas stress drops, where data is more numerous due to the larger number of high-quality broadband stations, are randomly distributed across the fault (Figure 4.9b) and over time (Figure 4.10). The Cleburne earthquakes cover a small spatial area and occur over a very short time span, precluding assessment of spatiotemporal stress drop differences (Figures 4.9c and 4.10). In the Venus sequence, the mainshock (ComCat $M_w$ 4.0 and GIT $M_w$ 3.5) has higher stress drop (27.35 MPa) than its aftershocks, which ranged from 2.56 to 8.40 MPa (Figures 4.9d and 4.10). No temporal changes in stress drop are observed in the other three
sequences, though we again note that local stations were deployed after the first felt events in all cases. Seismicity in the FWB has not followed typical mainshock-aftershock patterns but rather consists of swarms of small earthquakes (Hornbach et al., 2016; Quinones et al., 2019). The 2015 $M_w$ 4.0 (ComCat) Venus earthquake was the largest event and triggered deployment of a local network, but it occurred 7 years after the first earthquake on the fault (Scales et al., 2017). If temporal patterns were present, as documented for Oklahoma (Chen and Abercrombie, 2020), one might expect to see evidence in the well-documented Irving-Dallas sequence (Figure 4.10), but we do not.

Estimated stress drops for individual earthquake sequences are plotted against hypocenter depth to investigate possible vertical variations in stress drop (Figure 4.11). No correlations exist with event depths ranging from 2 to 9 km. Azle-Reno stress drops may hint at a decrease with depth but there are too few earthquakes to be definitive. There is no apparent change in stress drop with increasing source depth for the Irving-Dallas sequence, which includes many more events. These results are distinct from other studies that suggest that stress drop increases with depth, either as a result of increasing overburden pressure (e.g., Boyd et al., 2017) or decreasing pore pressure (e.g., Huang et al., 2017; Trugman et al., 2017). This lack of correlation with source depth may be due to the small number of earthquakes in our study or reflect the limited depth range of earthquakes, which all occur within the shallow crystalline basement.

4.6 Discussion

Applying GIT with a site correction based on a soft rock reference, we isolate source spectra in order to estimate source parameters and investigate stress release of earthquakes occurring in the FWB. Here, we compare our new stress drops to previous estimates for events near the study area. Earthquake scaling and estimates of the possible maximum magnitude are
subsequently discussed, recognizing that the narrow range of observed magnitudes limits the ability to extrapolate to larger events. Lastly, we discuss the relation between source characteristics of FWB earthquakes and estimated pore fluid pressure changes associated with wastewater disposal.

In order to compare the estimated FWB stress drops from this study to estimates of other induced earthquakes, the underlying model assumptions applied in these other estimation procedures must first be assessed. For example, in equation (3.13), we used the Madariaga (1976) model to define the relationship between source radius and corner frequency, which depends on source geometry ($K_0=1.32$). If we used the original Brune model parameter ($K_0=2.32$) then the mean stress drop decreases by a factor of 5.5. In addition, shear-wave velocity at the source is an important piece of information in order to be able to compare stress drop estimates from different studies. Since the stress drop scales with the cube of velocity, small velocity changes will have a large impact on stress drop estimates. We re-estimate stress drops for induced earthquakes in the CUS and Oklahoma using procedures consistent with those applied in this paper using the corner frequencies and seismic moments provided by Boyd et al. (2017) and Wu et al. (2018) (Figure 4.12). Boyd et al. (2017) provide S-wave velocity at the earthquake hypocenter and we use these velocities for CUS stress drop estimates. In order to estimate stress drop associated with Oklahoma seismicity, we use the 1D velocity model created for the state of Oklahoma (Darold et al., 2015). For the DFW Airport area we use the FWB velocity model. The comparison suggests that the stress drops for the FWB induced earthquakes are consistent with those from the other studies. The mean stress drops from CUS and Oklahoma are estimated to be 2.4 and 9.5 MPa. Stress drop variations in these other studies are larger than in the FWB study, possibly attributable to differences in focal mechanisms and the broader
extent of the study areas. The focal mechanisms of FWB earthquakes are mostly normal-faults
(Figure 4.1a) while Oklahoma earthquakes have primarily strike-slip mechanisms (Wu et al.,
2018). Reiter et al. (2012) calculate source parameters from coda-derived source spectra for the
DFW earthquakes, which were the first felt earthquakes associated with oil and gas production in
the FWB, but they do not include corner frequency estimates in their paper. They state that all
events exhibit similar corner frequencies near 8-10 Hz, with the exception of two events with
slightly lower corner frequencies. We estimated apparent corner frequencies from their published
coda source spectra with values from 8 to 10 Hz and recalculated stress drops that are added to
Figure 4.12. The minimum DFW stress drop estimate is 0.4 MPa with a mean value of 1.7 MPa,
similar to ~1 MPa from Reiter et al. (2012). Although DFW stress drops are somewhat lower,
they are consistent with the lower end of stress drops expected for global intraplate earthquakes
and within the stress drop ranges recalculated from Boyd et al. (2017) and Wu et al. (2018)
(Figure 4.12).

Although FWB earthquakes have a limited magnitude range of 1.7, the nearly constant
source scaling of stress drop with magnitude suggests a method to estimate anticipated ground
motion from larger magnitude events. Fault length estimates for the FWB active seismic
sequences studied here range from 3.5-5.5 km and a downdip width of ~4 km (Magnani et al.,
2017; Hennings et al., 2019). A Brune circular source with maximum radius of 3 km replicates
the potential source area which in turn produces an estimate for the largest moment magnitude of
5.6 based on a mean stress drop of 5.33 MPa from the data analysis. The uncertainty of the
maximum magnitude is ±0.1 with the standard deviation of GIT stress drop estimates (~2 MPa).
This estimate is slightly larger than the magnitude 5 estimate using the empirical scaling
relationship derived by Wells and Coppersmith (1994) and a maximum fault area of 24.7 km².
from the Venus study (Scales et al., 2017). We note that this simple estimate assumes slip in a single earthquake, which is not consistent with the activity in the basin to date, and it ignores important local effects such as pre-existing stress state and pore fluid pressure changes and the limited magnitude range used to assume self-similarity. However, many FWB regional faults in the basin are longer than the 4 active faults studied here (see Hennings et al., 2019; Horne et al., 2020), so it is not an unfeasible estimate of an upper bound on earthquake size. If we take into account estimates of site amplification, which are as great as 5 in the study area (Jeong et al., 2020), the potential damage from such an event in the metropolitan areas of the FWB increases.

In order to explore any possible relationship between stress drop and pore pressure perturbation, we plot stress drops as a function of distance from nearest injection wells for individual earthquakes sequences (Figure 4.13a). The analysis is motivated by the hypothesis that events closest to the injection points may be more strongly influenced by pore pressure changes. Injection depth is estimated by averaging over the injection interval (Table 4.3). For the Irving-Dallas sequence, the injector near DFW Airport is used to calculate the radial distance although the distance from the events is more than 10 km as mentioned earlier (Figure 4.1a). The closest separation distance is 1.5 km for the Cleburne earthquakes. These separations are larger than the 300 m distance suggested to be influenced by pore pressure changes by Goertz-Allmann et al. (2011) but smaller than the 2.5 km from Chen and Shearer (2011). We find no correlation between stress drop and the radial distance associated with the four well-developed sequences studied here.

Subsurface pore fluid pressure estimates are dependent on a number of geomechanical parameters that vary between studies. We can directly compare FWB stress drop estimates to bounds on pore pressure changes needed to initiate slip based on stress inversions of the focal
mechanism for 33 earthquakes in Azle-Reno, Irving-Dallas and Venus as documented in Quinones et al. (2018) (Figure 4.13b). The pore pressure change estimates include fault-plane uncertainties (quality from A to D) associated with each mechanism. This comparison results in no systematic relationship between stress drop and the bounds on pore fluid pressure perturbation. Pore pressure estimates in the FWB based on numerical models and data on cumulative fluid injection volumes produce significantly lower pore pressure perturbations (< 1 MPa) (e.g., Gono et al., 2015; Hornbach et al., 2015; Ogwari et al., 2018; Zhai and Shirzaei, 2018; Gao et al., 2019) than the estimates from Quinones et al. (2018) based on focal mechanisms (shown in Figure 4.13b).

The lack of correlation between earthquake stress drop and estimates of pore fluid pressure change suggests that tectonic stresses control slip and stress drop during the induced earthquakes in the FWB, as articulated by Huang et al. (2017). Foulger et al. (2018) suggest that the stress released in an induced earthquake is not necessarily equal to the anthropogenic stress but rather the effect of pre-existing tectonic stress. A caveat here is that the FWB dataset does not contain the first events in each sequence, making it difficult to document changes with respect to time and fluid diffusion rates (e.g., Chen and Abercrombie, 2020). Studies combining continuously recorded seismic data, fluid-flow modeling, geomechanical analysis, and geologic characterization may be required to explore possible direct physical links between induced earthquake source characteristics in the crystalline basement rocks and pore fluid pressure changes that begin within the injection volume which may be hydraulically linked to the basement rocks.
4.7 Conclusion

This study assesses source characteristics of earthquakes occurring in four earthquake sequences in north Texas believed to be induced by human activity. Source parameters are estimated using S-wave source spectra following corrections for path and site effects using the modified GIT introduced by Jeong et al. (2020) under conditions where there are no hard-rock reference sites. Corner frequency estimates from GIT compare favorably with those estimated using the EGF method. The mean corner frequency residual between the GIT and EGF method is $-0.31$ Hz. This small variation suggests that the two methods produce similar estimates. The GIT moment magnitudes are also found to be consistent with the ENA $M_w$ from the relation of Rigsby et al. (2014). These comparisons validate the path and site corrections developed in the modified GIT and the subsequent source spectra estimates.

The mean stress drop estimate for all FWB earthquakes is $5.33$ MPa, consistent with average stress drops for intraplate earthquake which vary from 1 to 10 MPa (e.g., Kanamori and Anderson, 1975). The results are also consistent with previous stress drops estimates, corrected for a Brune source model, for earthquakes from the DFW Airport (Reiter et al., 2012), CUS (Boyd et al., 2017) and Oklahoma (Wu et al., 2018). These studies report intraplate earthquakes stress drops of 1.7, 2.4 and 9.5 MPa, respectively.

Stress drops for the FWB earthquakes in this study show no correlation with time and location, focal depth, distance from nearest injection well, and pore pressure changes estimated from focal mechanisms. These earthquakes also follow self-similar scaling, although the magnitude range in this study is very limited. Based on these results, the induced intraplate stress drops in this study suggest that while subsurface stress perturbations from wastefluid injection promote slip on these faults, the stress drops of events occurring well into the sequence evolution
are a function of the pre-existing tectonic stresses. For this reason, the FWB earthquakes source characteristics are not distinct from tectonic intraplate earthquakes at the same depths. Hence, seismic ground shaking estimates from the injection-induced earthquakes should be assessed to be similar to tectonic intraplate earthquakes, while taking into account any basin specific propagation path effects and local site amplification.
Figure 4.1. (a) Map with earthquakes (circles) and focal mechanisms for Cleburne, Azle-Reno, Irving-Dallas, and Venus sequences in the Fort Worth Basin (FWB), Texas (Justinic et al., 2013; Quinones et al., 2018). The main tectonic features and faults (solid black lines) including the Newark East Fault Zone (NEFZ) and the Ouachita thrust front, are from Horne et al. (2020). Population centers with named earthquake sequences (DFW Airport, Cleburne, Azle-Reno, Irving-Dallas, Venus, and Fort Worth) are marked (black dots). Cross symbols denote wastewater injection wells. Rectangles denote the expanded areas depicted in Figure 4.2. The inset map outlines the Barnett Shale (gray) in the FWB. (b) Map with the seismic network including strong motion (triangles), broadband (squares), and short-period (diamonds) instruments deployed since 2009. Maximum horizontal stress ($S_{H_{\text{max}}}$) orientation from Lund Snee and Zoback (2016) is denoted by the gray bar. The white star is the location of Trigg Well (Geotechnical Corporation, 1964) used to constrain the velocity model.
Figure 4.2. Details of earthquake locations in the four study areas denoted by rectangles in Figure 4.1a including: (a) Azle-Reno; (b) Irving-Dallas; (c) Cleburne; and (d) Venus areas with the hypocentral depth denoted by the color scale to the right.
Figure 4.3. An example three-component recording (left panels) and estimated S-wave spectra (right panels) from the station UDFB (epicentral distance 2.9 km) for an Irving-Dallas earthquake on 01/20/2015 ($M_L$ 3.2, depth 5.7 km). Horizontal lines in the left column represent time windows for S waves. Dashed rectangles in (e) identify secondary waves following the direct P-arrivals. In the right column, black and gray lines indicate signal and noise estimates. The inverted black triangles indicate the starting point of spectral shape change near 30 Hz.
Figure 4.4. (a) Nonparametric attenuation functions as a function of hypocentral distance for selected frequencies from Jeong et al. (2020). Black bold dashed line represents assumed geometrical spreading of $G(r)=1/r$. (b) Site amplification function for reference station (CLEF2) from the Cleburne sequence. The mean GIT H (white line) with ± one standard deviation (gray) was estimated using data from 5 earthquakes and is compared to HVSR (continuous line). Dashed curve and black line with circles are the estimates of the vertical (GIT V) and the ratio between GIT H and GIT V (GIT HV) for these events.
Figure 4.5. An example showing the process of retrieving EGF source spectra. (a) SNR for recordings of all EGF events. Vertical dashed lines delineate the effective frequency range used to estimate corner frequency. The horizontal dashed line denotes SNR = 5. (b) Normalized spectral ratios estimated for individual EGF events (gray lines) and the final median spectral ratio (black line). The dashed curve depicts the Brune spectral ratio model with the corner frequency of the target event (triangle). (c) The comparison spectrum between the median spectral ratio and the synthetic model in (b). (d) The $\omega^{-2}$ model plotted in acceleration with corner frequency estimated in (b) of the target event multiplied by the comparison spectrum in (c) as normalized EGF source spectrum (dashed line). The continuous line indicates the normalized GIT source spectrum.
Figure 4.6. Comparison of acceleration source spectra and corner frequencies estimated using the GIT and EGF methods applied to data from common earthquakes. (a) The first example documents similar corner frequency estimates using the GIT (continuous) and EGF methods (dashed) and (b) the second example produces slightly different corner frequency estimates using the two methods. Amplitudes are normalized to compare spectral shape. (c) Comparison of corner frequencies with 95% confidence intervals from GIT and EGF. Events from the four sequences are distinguished by the different symbols where open symbols represent events shown in (a) and (b). (d) Corner frequency residuals between the GIT and EGF estimates. Black continuous lines in (c,d) represent 1:1 slope and zero offset.
Figure 4.7. Corner frequency estimates from EGF using both the Boatwright and Brune spectral models. (a) Comparison of EGF corner frequencies with 95% confidence intervals and (b) corner frequency residual between the Boatwright and Brune models. Events from the four sequences are distinguished by the different symbols. Black continuous lines in (a,b) represent 1:1 slope and zero offset.
Figure 4.8. Analyses of GIT source parameters. (a) Comparison of $M_W$ derived from GIT source spectra and ComCat $m_{blg}$ magnitudes. Black line is a regression prediction with 95% confidence intervals (dashed lines). Events from the four sequences and the remainder of the catalog (named Other) are distinguished by the different symbols in the legend. The gray line is the relation $M_W = 0.67m_{blg} + 1.03$ for eastern north America (Rigsby et al., 2014). (b) Relationship between corner frequency $f_c$ and seismic moment $M_0$ (also $M_W$ upper axis) estimated from GIT for the FWB earthquakes. The least-squares fit produces a slope of $-0.29$. Black dashed lines represent constant stress drop levels.
Figure 4.9. Spatial distributions of stress drops for all events in this study ($M_w > 1.9$). Individual figures include the earthquakes (circles), injection wells (crosses), and causative faults (Horne et al., 2020) (black lines) for (a) Azle-Reno, (b) Irving-Dallas, (c) Cleburne, and (d) Venus. Circle diameters represent moment magnitude with color indicative of the logarithm of stress drop.
Figure 4.10. Temporal stress drop distribution with 95% confidence intervals for the Cleburne series in 2009 (top) and the Azle-Reno, Irving-Dallas, Venus, and other areas in 2014-2018 (bottom). The black circle denotes the Azle-Reno event with highest stress drop located at edge of the fault as shown in Figure 4.9a.
Figure 4.11. Stress drop estimates with 95% confidence intervals from GIT analysis plotted against event depth.
Figure 4.12. Comparison of source parameters estimated for earthquakes in the central United States (CUS) (Boyd et al., 2017) (gray asterisks), Oklahoma (Wu et al., 2018) (black asterisks), the FWB (this study) (gray circles), and DFW area (Reiter et al., 2012) (white circles). (a) Stress drops as a function of moment magnitude. The stress drops are calculated using the same estimation procedure and source model for direct comparison. (b) Relationship between $f_c$ and $M_0$ ($M_w$ upper axis), directly provided by each study.
Figure 4.13. Stress drop estimates with 95% confidence intervals from GIT plotted against (a) distance from the nearest injection point and (b) pore pressure changes ($\Delta P_P$) needed for slip based on fault plane solutions (Quinones et al., 2018). For the Irving-Dallas sequence, the injector near DFW Airport is used to calculate the radial distance. Injection depths are approximated as an average value between the top and bottom of the injection interval in Table 4.3.
Table 4.1. Catalog information and source parameter estimates for the 14 target events using the EGF method

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Table 4.2. GIT source parameters for the 79 Fort Worth Basin earthquakes from 2009 to 2018.

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The supporting information provides details on the following topics: Comparison between unsmoothed and 1-Hz smoothed spectra; Site effects estimated from generalized inversion technique (GIT) and horizontal-to-vertical spectral ratio (HVSR) estimates for the Cleburne sequence; and Validation of the 2 km separation criteria used for the selection of events used in the empirical Green's function (EGF) analysis.

Comparison between Unsmoothed and 1-Hz Smoothed Spectra

We compare unsmoothed spectra to 1-Hz smoothed spectra used for main spectral analysis (Figure 4.B1). There are discrepancies at low-frequency amplitude (< 1 Hz) between unsmoothed and smoothed spectra (Figure 4.B1 top). However, for frequencies above 1 Hz, both spectra are well matched. Thus, we conclude that the smoothing window does not strongly affect the spectral shape within the effective frequency band in this study from 1 to 25 Hz. Additionally, we investigate the effect of the smoothing window on the spectral ratio (Figure 4.B1 bottom). Note that the spectral ratio is a direct division between the smoothed spectra for target earthquake and EGF events. Similarly, the unsmoothed spectral ratio and the 1-Hz smoothed spectral ratio show no significant difference in spectral shape (Figure 4.B1 bottom).

Site Effects for Cleburne

Jeong et al. (2020) and Chapter 3 document site effects for Azle-Reno, Irving-Dallas, and Venus sequences using the modified GIT, which is applied to the study area where there is no
hard-rock reference site. Here, we illustrate additional site effects for the unique stations that recorded the Cleburne sequence. Figure 4.B2 displays the GIT site effect from the horizontal component (GIT H), vertical components (GIT V), the ratio of GIT H and GIT V (GIT HV), and horizontal-to-vertical ratio from observed spectra (HVSR) for the 6 unique local stations deployed in the Cleburne area. 100 bootstrap inversions were performed to estimate the mean and standard deviation of GIT H (gray area in Figure 4.B2) except for CLEF2 (reference station). For the reference site, the mean and standard deviation of GIT H are estimated from the reference site correction method introduced in Chapter 3. For most stations, GIT H, HVSR and GIT HV are in general agreement with GIT V estimates approaching unity except for CLELK. In the case of CLELK, GIT H is larger than HVSR with GIT V indicating an amplification of 2. For all stations, GIT H decays slightly faster at high frequency than HVSR possibly due to near-surface site diminution.

Validation of 2 km Separation Criteria in EGF Event Selection

Target and EGF events are separated by no more than 2 km in this study. Abercrombie (2015) recommends that the EGF method not be used when wavelengths are shorter than the EGF earthquake source dimension. Based on the wavelength argument, the 2 km separation may approach a distance that exceeds the limits at the highest frequencies. To analyze the separation effect, we compare spectral ratios for the closest and farthest distances (≤ 2 km) using EGF for events in the Irving-Dallas sequence (Figure 4.B3). The comparisons indicate that the two spectral ratios are similar up to a separation of 2 km, supporting the separation criteria used for the EGF method in this study.
Figure 4.B1. Comparison of unsmoothed and 1-Hz smoothed acceleration spectra (top) recorded at stations AFDA (left) and AZDA (right) for an Irving-Dallas earthquake on 01/07/2015 02:11:17 (ML 3.7, depth 4.9 km). Dashed vertical lines indicate 1 Hz, which is the minimum frequency used in this study. The same comparison for the spectral ratio (bottom) is plotted using EGF event on 01/07/2015 02:47:40 (ML 2.3, depth 6.0 km).
Figure 4.B2. Site response functions estimated using GIT and HVSR for the 6 unique local stations that recorded the Cleburne sequence. The GIT site effects from the horizontal (GIT H; white lines), vertical (GIT V; black dashed lines), and the ratio of GIT H and GIT V (GIT HV; black line with black circles) are compared to HVSR (black lines). The gray shaded area denotes the mean ±1 standard deviation of the GIT H from the bootstrap analysis except for CLEF2 (reference site highlighted in bold). The site response function and one standard deviation of CLEF2 are the same as Figure 4.4b, which is estimated using the reference site correction method. The station name and the number of earthquakes (EQs) are noted in each plot.
Figure 4.B3. Examples of normalized spectral ratios using the closest EGF event (gray curves) and the most distant EGF event (black curves) at IPD1 and AZWP for an earthquake in Irving-Dallas sequence ($M_L$ 3.7, depth 4.9 km).
REFERENCES


CHAPTER 5
STRESS DROP VARIATIONS OF INDUCED EARTHQUAKES AT THE DALLAS-FORT WORTH AIRPORT, TEXAS

Abstract

Stress drops for injection-induced earthquakes near the Dallas-Fort Worth International Airport in the Fort Worth Basin (FWB), Texas are estimated. The mean stress drop for the Airport sequence is lower than other FWB earthquake sequences. Airport earthquake stress drops increase with radial distance from the injection point, but only over the first 1.5 km; estimates for more distant earthquakes are spatially independent, consistent with other FWB sequences. The low stress drop Airport events occurred shortly after the initiation of injection on a fault extremely close to the well, and we suggest that here we uniquely captured the initial low stress drops predicted by direct triggering of earthquake via rapid pore pressure changes.

5.1 Introduction

Stress drop is a source parameter defined as the difference in shear stress on a fault before and after earthquake slip that provides a measure of rupture energy. Stress drop is proportional to the strain energy released as seismic waves over a characteristic fault area and scales with area similar to seismic moment in that more slip yields higher stress drop than less slip on the same fault area. Stress drop estimates are essential as they quantify variations in slip behavior across different geology and faults, which contribute to seismic hazard calculations.
Shear resistance during frictional sliding on a fault plane is related to changes in pore pressure through the modification of effective normal stress (i.e., Hubbert and Rubey, 1959). In areas of anthropogenically induced earthquakes associated with wastefluid injection, the increased pore pressure in the subsurface has been hypothesized to reduce effective normal stress on pre-existing faults which can trigger shear failure within the background stress field (e.g., Zoback, 2007; Ellsworth, 2013; Lund Snee and Zoback, 2016; Yeck et al., 2016). Numerous earthquake source studies suggest that pore pressure changes and reduction of effective normal stress associated with the fluid injection directly influences stress drop, but consensus has not been reached on a single quantitative relationship. Using earthquakes near geothermal sites in the Salton Trough, Chen and Shearer (2011) observed that stress drop increases with radial distance from an injection point for short distances but beyond 2.5 km pore pressure perturbations were not high enough to reduce stress drop. Similarly, Yu et al. (2020) document lower stress drops (~0.1 to 1 MPa) for events that occurred within 1 km of a hydraulic fracturing well in the Montney oil and gas production area compared to those associated with earthquakes at further distances (~1 to 10 MPa); this study attributes the finding to increased fracture density and/or elevated pore pressure near the well. Although these two studies suggest different distance limits of lateral variations in stress drop, they both indicate that the spatial distribution of stress drop can be modified within a region of high pore pressure perturbation. However, in studies of earthquakes in the Geysers geothermal field, Staszek et al. (2017) suggest that stress drop changes are inversely related to the injection rate and find no relationship with distance from injector. Injection-induced earthquakes in Oklahoma and Kansas show low stress drops with variation as a function of magnitude but no correlation with distance (Sumy et al., 2017; Trugman et al., 2017). Lastly, in the Soultz-sous-Forets geothermal reservoir, Lengliné et al.
(2014) observe stress drop variations as large as a factor of 300 but do not report any correlation with injection data.

Here, we estimate stress drop using S-waves in a data-driven spectral inversion for well-recorded induced earthquakes in the Fort Worth Basin (FWB) of north Texas (Figure 5.1). Previous source studies in the FWB reported stress drops consistent with values expected for tectonic events with no discernable spatiotemporal patterns (Justinic et al., 2013; Jeong et al., 2020, 2021). In contrast, the Dallas-Fort Worth (DFW) Airport sequence, which produced the first felt earthquakes associated with oil and gas production in the FWB, yielded a lower mean stress drop of ~1 MPa (Reiter et al., 2012). Additionally, the Airport sequence is the only FWB sequence that has produced earthquake locations that migrate away from a single-source injection well with time (Ogwari et al., 2018); all other FWB sequences are associated with multiple wells or no close-in injection wells. Here, we recompute and compare the Airport stress drops to the FWB stress drop catalog (Jeong et al., 2020, 2021), holding methodology fixed. In the case of the Airport sequence, we document a connection between the source characteristics of the earthquakes and the modeled pore pressure increase within the first 1.5 km from the injector.

5.2 Seismicity of Dallas-Fort Worth, Texas

The FWB has experienced little to no significant deformation due to faulting over the past 300 Ma (Magnani et al., 2017), but between 2008 and 2020, the population has experienced over 30 magnitude 3 and above earthquakes with 114 felt events (e.g., Frohlich et al., 2016; U.S. Geological Survey [USGS] Comprehensive Catalog [ComCat], last accessed December 2020). FWB earthquakes have been linked to unconventional oil and gas extraction techniques (Hornbach et al., 2015; Scales et al., 2017; Ogwari et al., 2018; Quinones et al., 2019) (Figure
5.1). Here, wastefluid from primarily gas production in the Mississippian Barnett shale is re-injected into the underlying Ordovician Ellenburger dolomite, which has an average thickness of ~1 km and lies unconformably on the Precambrian basement (Pollastro et al., 2007; Smye et al., 2019). Most felt earthquakes in the FWB have occurred on faults rooted in the crystalline basement that extend upward through the Ellenburger and Barnett formations (Magnani et al., 2017). For perspective, the east-dipping top of Ellenburger occurs from 2.00-2.74 km while the earthquakes occur between ~4.5 km to 8 km below sea level (see review in Quinones et al., 2019 and Smye et al., 2019). Thus, the injection depth interval of interest is shallower than many of the earthquake focal depths but the regional faults provide connectivity between units (Figure 5.1). Most FWB seismicity occurs on the northeast-southwest trending basement faults that appear optimally oriented for failure within the regional stress field (e.g., Hennings et al., 2019; Horne et al., 2020) (Figures 5.1 and 5.2). Seismogenic faults are generally less than 8 km in length, while the fault length in the DFW Airport area is significantly larger, exceeding 50 km (Horne et al., 2020) (Figures 5.1 and 5.2).

FWB seismicity has been recorded by a series of seismic networks since 2008 (for review see DeShon et al., 2018). The five best-studied earthquake sequences include: (1) DFW International Airport (Frohlich et al., 2011; Ogwari et al., 2018); (2) Cleburne (Justinic et al., 2013); (3) Azle-Reno (Hornbach et al., 2015); (4) Irving-Dallas (Magnani et al., 2017; Quinones et al., 2018); (5) the 2015 $M_w$ 4.0 Venus earthquake (Scales et al., 2017) (Figure 5.1). The Airport sequence, which we focus on here, generated eight events large enough to be reported by USGS, with magnitudes from 2.6 to 3.0, beginning in October 2008. Subsequently, 11 events with magnitude from 1.7 to 2.3 were well-recorded using a local seismic network installed by Southern Methodist University (SMU) between 20 November and 2 December 2008. We use
these 11 events to estimate Airport stress drops. These 11 hypocenters occurred within 1 km of a then newly operating injection well located at the southern end of the Airport property and hypocenters traced a ~1 km linear feature that coincided with a mapped fault (Figure 5.2b). Fluid injection at the nearest well began 7 weeks before the earthquakes (September 2008) and was subsequently shut down in August 2009 (Figure 5.3). Ogwari et al. (2018) showed that subsequently seismicity continued to migrate primarily to the northeast consistent with an evolving pore pressure change front. Unlike the Airport earthquakes, the other FWB sequences over the last decade show relatively long time spans (2-5 years) between initial fluid injection at nearby (<5 km) wells and the onset seismicity (Figure 5.3). Some studies of FWB earthquakes suggest a role for the production wells (Hornbach et al., 2015; Chen et al., 2020) and cumulative regional changes (Hornbach et al., 2016) over simple well-earthquake distances. In this vein, it is important to note that the Irving-Dallas sequence is located >10 km from the nearest injector and remains an enigmatic sequence (Hornbach et al., 2016).

5.3 Data and Methods

Earthquake location and magnitude estimates for the Azle-Reno, Irving-Dallas, and Venus sequences come from the North Texas Earthquake Study (NTXES) catalog (Quinones et al., 2018) and those for the Cleburne sequence are taken from Justinic et al. (2013). Magnitudes from both catalogs’ match estimates found in the Advanced National Seismic System Comprehensive Catalog (ComCat). For the locally recorded 11 Airport events, there are no corresponding origins in ComCat but three studies report catalogs for these smaller events (Frohlich et al., 2011, Reiter et al., 2012, and Quinones et al., 2019). In order to choose which study to use as our primary reference, we assessed the magnitudes and location differences, as described in Appendix C. The locations are roughly consistent between the three catalogs. In
contrast, there are a difference in the magnitudes between the catalogs. As a result, the Frohlich et al. (2011) magnitudes best match with our magnitude assessment method in Appendix C. Thus, for the stress drop computation, we chose the hypocenters and magnitudes reported in Frohlich et al. (2011).

Data processing to make spectral estimates follows methods outlined in Chapter 3.3. We only use S-waves to investigate stress drop characteristics. In order to isolate S-wave source spectra from observed spectra, the modified generalized inversion technique (GIT) is used following the procedures outlined in Chapters 3.4.1 and 3.4.2.

The inhomogeneity of the seismic networks as a function of time (DeShon et al., 2018) necessitated the performance of three GITs with unique reference stations. The Airport and Cleburne sequences did not overlap in time or space and require sequence-specific references. A common set of stations recorded the Azle-Reno, Irving-Dallas, and Venus sequences and 1203 records from 90 earthquakes were used in the GIT process with the site AZDA providing the reference (Jeong et al., 2020; Chapter 3). We adopt the path term from Figure 3.6 as the propagation characteristics for the FWB. After correcting for these path effects, Jeong et al. (2021) and Chapter 4 applied GIT to the 175 recordings from the 5 events for the Cleburne sequence using CLEF2 as the reference. In a similar fashion, after correcting for the FWB path attenuation, we report results for GIT performed using 48 recordings from 10 well-recorded Airport earthquakes using AFMOM as the reference site. Here, one event among the 11 Airport earthquakes was removed due to no data at the reference site. A detailed discussion of the site effects for DFW Airport stations is presented in the supporting information. Based on the extracted source spectra, we estimate seismic moment, corner frequency, and stress drop by using a Brune source model as documented in Chapter 3.5.3.
5.4 Results

We estimate stress drops for the FWB earthquakes as a function of spatiotemporal distribution and moment magnitude (Figures 5.2 and 5.4). The stress drops for Cleburne, Azle-Reno, Irving-Dallas, and Venus sequences are generally invariant with respect to time and location relative to the injectors. The one set of outliers are the lower stress drops from the Airport sequence. The Airport stress drops have a mean of 0.72 MPa with a standard error of 0.38 MPa, similar to ~1 MPa from Reiter et al. (2012). Stress drops for the rest of the FWB with a mean of 5.33 MPa with a standard deviation of 2.00 MPa. The median value of Airport stress drop is 0.61 MPa with a median absolute deviation (MAD) of 0.29 MPa while the stress drops of other sequences are a median of 5.49 with a MAD of 1.82 MPa. The details of the stress drop values are documented in Tables 5.1 and 5.2. Since the stress drop differences between the DFW Airport and other sequences are more than 4 MPa for both mean and median and exceeds the uncertainty of total FWB stress drops (~2 MPa), we suggest that the Airport stress drops are lower than those for the other sequences. The magnitudes and depths associated with the Airport events are not appreciably shallower than in the other FWB sequences, and depth alone fails to correlate with the lower stress drop (Figures 5.4a and 5.4b). While the Airport earthquakes are the first events in the basin, comparison of the time history of stress drops across the last decade does not support a time-varying change as the subsurface pore fluid pressures evolve with gas development in the basin (Figure 5.4c).

We find a correlation between stress drop and the distance from the nearest injection wellhead for Airport events (Figure 5.5). The locations and details of individual injection wells are available from the Texas Railroad Commission (see also Table 4.3). Since injection takes place over discrete intervals open within a well over the 1 km thickness of the Ellenburger, we
assign the average value of the intervals as ‘injection depth’ in this study and then calculate radial distance from well to earthquake. The Airport stress drops increase with increasing radial distance up to 1.5 km (Figure 5.5), consistent with the findings in other studies (Chen and Shearer, 2011; Goertz-Allmann et al., 2011; Kwiatek et al., 2014; Yu et al., 2020). Stress drops for Cleburne, Azle-Reno, and Venus sequences, however, appear invariant to the radial distance from the nearest injection point. These other sequences suggest that stress drops for events beyond 1.5 km from the injector may be less affected by direct stress perturbation from the injected fluid flow, producing higher stress drops consistent with tectonic intraplate earthquakes from 1 to 10 MPa (Kanamori and Anderson, 1975). We note that the distance from injection points to earthquake hypocenter includes the earthquake location errors (~300 m) and injection depth uncertainty.

5.5 Discussion

We estimate stress drops for FWB earthquakes and find that the DFW Airport events produce statistically lower stress drops. The low stress drop Airport earthquakes uniquely occur within 1.5 km of the injector on a high-throw (>200 m) regional fault (Horne et al., 2020) thought to be crossed or very near the injector (Ogwari et al., 2018), and the stress drop values increase with radial distance from the injection depth. We do not find a temporal correlation, probably due to the short time period spanned by Airport earthquakes in this study (Figure 5.4c). Note that the very small magnitude earthquakes derived from template approaches used to show distant (6+ km) migration from the injector (Ogwari et al., 2018) could not be used for this study. The Airport sequence otherwise shares most tectonic characteristics with FWB earthquakes: earthquakes occur on a pre-existing normal fault well-oriented for failure at or below the Ellenburger basement interface in a series of moderate to small events (e.g., Frohlich et al., 2011;
Ogwari et al., 2018; Quinones et al., 2019), suggesting that the variation in the stress drop distribution may not be driven by the tectonic setting including focal mechanism and directivity effects (Kanamori et al., 1993; Hardebeck and Aron, 2009). Here, we discuss possible factors generating stress drop differences between the DFW Airport sequence and the other FWB area events.

Previous studies suggest there are different b-values for the Airport sequences and the other FWB earthquakes. Ogwari et al. (2018) estimated a higher b-value of 1.43 for the Airport earthquakes from October 2008 to December 2008 while Justinic et al. (2013) and Quinones et al. (2019) suggested the lower b-values from 0.67 to 1.01 for the Cleburne, Azle-Reno, Irving-Dallas and Venus sequences. The SMU NTXES catalog in its entirety (2008-presents) exhibits a b-value of 1. Bachmann et al. (2012) observed that b-values are higher at a close distance to the injection point and lower at a farther distance in the Basel geothermal field. They suggest that b-value variations are related to the reduction of effective stress. Moreover, the variation of b-values may be related fracture types (e.g., new and pre-existing fractures for higher and lower b-values, respectively) (Goertz-Allman and Wiemer, 2013) or higher pressurization rates near the injection point, which induce microseisms within the region of high-rate pressurization (Ogwari et al., 2018). In the FWB, where earthquakes are shown to occur on pre-existing faults, we would favor the latter interpretation.

Earthquake sequences in the FWB experience varying cumulative injection volumes, providing unique study opportunities (Figure 5.2). Gao et al. (2019) estimated pore pressure changes in the FWB based on a 3D hydrogeologic model built using extensive geologic inputs, reservoir fluid flowing properties, and fault geometries (e.g., Smye et al., 2019; Horne et al., 2020). Their modeled pore pressure values are calculated at the sedimentary/basement boundary.
and provide an average pore pressure value along the active faults at the sedimentary/basement boundary. The pore pressure changes at the initiation of the Airport sequence is ~0.03 MPa (Figure 5.6). This estimate is 1 order of magnitude smaller on average than other sequences (Gao et al., 2019; Figure 5.6). The lower pore pressure changes reported in the regional hydrogeologic model for the Airport reflect the short time and moderate volume of the south injector but spreads the injection pressures over time and space steps. Similarly, Zhai and Shirzaei (2018) estimated Coulomb failure stress changes in the FWB using injection data that are 0.005 MPa for the Airport and 0.35-0.4 MPa for Cleburne and Azle-Reno, where larger volumes are injected. These regional studies indicate that the absolute pore pressure change at the Airport well was low (Figure 5.2). These results suggest that the absolute value of pore pressure change may not directly tie to earthquake rupture characteristics like stress drop.

Airport seismicity occurred much closer to the injector and at shorter time lags following the onset of fluid injection than any of the other FWB sequences. These two facts, along with the lower stress drop estimates, suggest the possibility of direct triggering by rapid fluid pressure increases via a high-permeability hydraulic pathway (e.g., Peña Castro et al., 2020). The rate of pore pressure increase at the start time of the Airport sequence based on the results of Gao et al. (2019) is 0.0043 MPa/weeks at the sediment/basement interface (Figure 5.7). The modeled rates for the other FWB sequences range from 0.0012 to 0.0027 MPa/weeks (Figure 5.7). In the Airport area, however, Ogwari et al. (2018) estimated the pore pressure changes on the basement fault (4.5 km depth) at earthquake source depths (see Table 5.3). Here, the calculated rate of pore pressure increase is 0.004 MPa/weeks between 20 November and 27 November, which is consistent with the rate of pore pressure change from Gao et al. (2019). Thus, we suggest that the
rate of the pressurization increase may more directly affect the earthquake source properties than the absolute value of pore pressure perturbation.

The low stress drop Airport earthquakes could also reflect some degree of aseismic slip (e.g., Wei et al., 2015; Bhattacharya and Viesca, 2019; Cappa et al., 2019). Lengliné et al. (2014) suggest that the reduction of effective stress promoted by wastewater injection contributes to frictional stability and aseismic motion on the fault plane. Gugliemi et al. (2015) conducted a large-scale field experiment to estimate changes in fault displacement and fluid pressure following artificial water injection into a pre-existing fault in southeastern France. Their results suggest that when pore pressure significantly increases, pre-existing faults experience primarily aseismic slip then typical induced fault slip. During the aseismic period, observed seismicity releases only a portion of the total seismic moment (i.e., lower effective normal stress and stress drop). For events that occurred in this experiment, Huang et al. (2019) estimate significantly low stress drops (0.01 MPa) and suggest that fluid pressure perturbation and aseismic deformation influence the earthquake source properties. Similarly, Harrington et al. (2020) observe lower stress drops near injection wells in the Montney Play in British Columbia, Canada and suggest that the lower values result from aseismic slip. In this study, we do not have enough data in time or space to resolve the role of aseismic processes.

5.6 Conclusions

We observe anomalously low stress drops for an earthquake sequence near DFW Airport compared to other earthquakes recorded in the FWB over the last decade. The DFW Airport stress drops increase with radial distance from the causative injector for a distance of 1.5 km. The low stress drop earthquakes began within two months of injection initiation. Earthquakes extending up to 6 km distance along the same fault through 2015 but were too small to resolve
stress drop. Other earthquake sequences in the FWB occur at distances greater than 1.5 km from injectors and at much longer times following the initiation of injection, and these events show no spatial or temporal correlations and exhibit significantly higher stress drops consistent with natural intraplate earthquakes. We conclude that the rate of pore pressure increase for the DFW Airport events, related to the abrupt onset of a moderate volume well very near a large regional fault, explains the lower stress drops rather than the absolute perturbation associated with this moderate to low volume well. Based on our results, we hypothesize two modes of earthquake triggering in the FWB: (1) low stress drops reflect a direct influence on source linked to pressurization close to the injector; and (2) higher (normal) stress drops consistent with intraplate earthquakes which reflect the release of pre-existing strain on faults well-oriented for failure within the tectonic background stress regime. Seismic moments, focal depths, and absolute pore pressure changes (or cumulative amount of the injected wastewater) may not be the only controlling factors contributing to stress drop variations; consideration of the rate of pore pressure change and distance from injector to fault may be equally important.
Figure 5.1. The SMU North Texas Earthquake catalog showing the DFW Airport (red circle), Cleburne (blue), Azle-Reno (green), Irving-Dallas (magenta), Venus (cyan), and other areas (yellow) for the Fort Worth Basin, Texas (modified from Quinones et al., 2019). Boxes indicate the boundaries used in Figure 5.2. Focal mechanisms are from Justinic et al. (2013) and Quinones et al. (2018). Regional faults (solid black lines) with the downthrown hanging-wall block (black dots) are from Horne et al. (2020) and plotted at the top of the crystalline basement. Bold line indicates the seismogenic fault for the DFW Airport sequence. Wastefluid injection wells (gray inverted triangles) and the maximum horizontal stress ($S_{H_{\text{max}}}$) orientation from Lund Snee and Zoback (2016) (red bar) are also shown. The Trigg Well site (Geotechnical Corporation, 1964) (black star) provides primary stratigraphic and velocity control of the Airport study area. Inset: Texas and the Barnett Shale distribution in the FWB (gray).
Figure 5.2. The 5 named earthquake sequences include: (a) Azle-Reno, (b) DFW Airport, (c) Irving-Dallas, (d) Cleburne, and (e) Venus. Events are color-coded by stress drop variation. Injection wells are scaled by total volume from 2005 to 2018 (blue inverted triangles). Seismogenic faults are highlighted as bold lines and black dots indicate the downthrown hanging-wall block for each fault (Horne et al., 2020). Sequence locations are shown in Figure 5.1.
Figure 5.3. Duration between fluid injection initiation at nearest wells and onset seismicity for 4 sequences. The duration for DFW Airport area is significantly shorter (7 weeks) than other sequences (2-5 years).
Figure 5.4. Stress drop estimates versus (a) moment magnitude $M_W$, (b) event depth, and (c) time. Error bars are 95% confidence intervals of stress drop estimates. In (c), note the time breaks after Airport and Cleburne sequences. Color indicates sequence and following legend in (c).
Figure 5.5. Stress drops (circles) versus distances from the nearest injection points for the Airport (red), Cleburne (blue), Azle-Reno (green), and Venus (cyan) earthquake sequences. The Airport earthquakes document an increase in stress drop with range, while stress drops from the other sequences do not correlate with distance. The Irving-Dallas sequence locate at >10 km from the nearest well and can be viewed in Figure 5.4.
Figure 5.6. Change in pore pressure ($\Delta P_p$) estimated by Gao et al. (2019) at the starting time of each sequence and at each fault. The $\Delta P_p$ is estimated in the sedimentary layer on top of basement. The $\Delta P_p$ for DFW Airport area denotes the lower value than those from other areas.
Figure 5.7. Rate of pore pressure increase, ratio of $\Delta P_p$ and duration from Figure 5.6 and Figure 5.3, for 4 earthquakes sequences in the study area. DFW Airport area represents the rapid pore pressure increase due to closer distance to the injector and shorter time lag following the onset of fluid injection, suggesting direct effect of fluid on the fault.
Table 5.1. GIT source parameters for the 10 Dallas-Fort Worth earthquakes.

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<th>Longitude</th>
<th>Depth (km)</th>
<th>$M_L$</th>
<th>$M_W$</th>
<th>$f_c$ (Hz)</th>
<th>STD (MPa)</th>
<th># of STA</th>
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<td>2.3</td>
<td>2.19</td>
<td>5.72</td>
<td>0.4555</td>
<td>4</td>
<td>DFW</td>
</tr>
<tr>
<td>2008-12-01T21:26:33</td>
<td>32.8594</td>
<td>-97.048</td>
<td>4.46</td>
<td>2.3</td>
<td>2.3</td>
<td>7.57</td>
<td>1.5712</td>
<td>4</td>
<td>DFW</td>
</tr>
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</table>
Table 5.2. Mean and median stress drops with the errors for the FWB earthquakes

<table>
<thead>
<tr>
<th>Region</th>
<th>Mean (MPa)</th>
<th>Standard deviation (MPa)</th>
<th>Median (MPa)</th>
<th>Median absolute deviation (MPa)</th>
<th>Stress drop range (MPa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>DFW Airport</td>
<td>0.72</td>
<td>0.38</td>
<td>0.61</td>
<td>0.29</td>
<td>0.36 - 1.57</td>
</tr>
<tr>
<td>FWB except for DFW</td>
<td>5.33</td>
<td>2.00</td>
<td>5.49</td>
<td>1.82</td>
<td>1.37 - 27.35</td>
</tr>
<tr>
<td>Cleburne</td>
<td>5.04</td>
<td>2.18</td>
<td>8.38</td>
<td>1.96</td>
<td>1.83 - 9.10</td>
</tr>
<tr>
<td>Azle-Reno</td>
<td>4.43</td>
<td>1.92</td>
<td>4.39</td>
<td>1.57</td>
<td>1.37 - 10.46</td>
</tr>
<tr>
<td>Irving-Dallas</td>
<td>6.53</td>
<td>1.79</td>
<td>7.27</td>
<td>1.63</td>
<td>1.79 - 15.15</td>
</tr>
<tr>
<td>Venus</td>
<td>5.46</td>
<td>2.78</td>
<td>2.94</td>
<td>2.26</td>
<td>2.56 - 27.35</td>
</tr>
<tr>
<td>Other FWB area</td>
<td>2.58</td>
<td>1.69</td>
<td>2.18</td>
<td>1.46</td>
<td>1.64 - 10.52</td>
</tr>
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</table>
Table 5.3. Pore pressure changes between November 20 and December 1, 2008 from Ogwari et al. (2018)

<table>
<thead>
<tr>
<th>Date</th>
<th>Top of basement (MPa)</th>
<th>4.5 km depth (MPa)</th>
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</thead>
<tbody>
<tr>
<td>20-Nov-2008</td>
<td>0.0585</td>
<td>0.0088</td>
</tr>
<tr>
<td>21-Nov-2008</td>
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<td>0.0093</td>
</tr>
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<td>22-Nov-2008</td>
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<td>0.0099</td>
</tr>
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<td>23-Nov-2008</td>
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<td>0.0623</td>
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<td>29-Nov-2008</td>
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<tr>
<td>30-Nov-2008</td>
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<td>0.0149</td>
</tr>
<tr>
<td>1-Dec-2008</td>
<td>0.0689</td>
<td>0.0156</td>
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</tbody>
</table>
Appendix C

Supporting information provides further details of the selection of the Dallas-Fort Worth (DFW) catalog (Figure 5.C1) and resulting site effects from the DFW sequence (Figure 5.C2).

Selection of DFW catalog

A generalize inversion technique (GIT) isolates path, site, and source contributions using a hard-rock reference site. When no hard-rock site exists, GIT site effects for a horizontal component (GIT H) at a reference station can be corrected using a site correction method introduced in Jeong et al. (2020) and Chapter 3. In this method, the site amplification for the reference station is estimated as a residual from path-corrected spectra and Brune $\omega^{-2}$ spectral model (Brune, 1970) using seismic moment derived from a magnitude that matches the Advanced National Seismic System Comprehensive Catalog (ANSS ComCat). Jeong et al. (2020) suggest that GIT H estimates show good agreement with horizontal-to-vertical ratio (HVSR), which are estimated directly from the observed spectra, when the GIT site effect on a vertical component (GIT V) is a flat response as a function of frequency. Based on the result, we find an optimal catalog for this study.

Since the ANSS ComCat does not report the locally recorded 10 earthquakes in the DFW Airport study, we assess three different magnitude estimates from the literature for these events (Frohlich et al., 2011; Reiter et al., 2012; Ogwari et al., 2018) in order to determine appropriate magnitudes for the DFW Airport sequence. First, GIT V is estimated by constraining the average
of all sites to unity under the assumption that there should be small site amplifications on vertical components. We find the least variation in GIT V for AFMOM and select this station as a single reference site for the DFW earthquake inversion (Figure 5.C1a). We estimate the site effects for AFMOM using three different local magnitudes (Figure 5.C1a). The Frohlich et al. (2011) magnitudes produce the most consistent GIT H site amplification within the standard deviation of HVSR. Frohlich et al. (2011) state that their magnitudes are estimated to mimic the ComCat magnitudes. Site amplification functions estimated using the magnitudes of Reiter et al. (2012) are slightly underestimated (i.e., will produce larger moments) while the magnitudes from Ogwari et al. (2018) overestimate the site effect (i.e., will produce smaller moments).

Event locations from the three catalogs are compared in Figure 5.C1b. They are consistent with one another and follow the causative fault provided by Horne et al. (2020). The earthquakes in the catalog from Ogwari et al. (2018) have a depth distribution similar to the reference catalog (i.e., Frohlich et al., 2011). The mean depth difference between the two catalogs is 0.02 km with a standard deviation of 0.22 km. The catalog from Reiter et al. (2012) documents shallower depths than those of Frohlich et al. (2011). The mean depth difference between the two catalogs is 0.45 km with a standard deviation of 0.24 km. Considering location uncertainty of 0.3 km, the hypocenters of DFW Airport events from the three catalogs are roughly consistent with one another. As a result, the magnitudes and hypocenter locations from Frohlich et al. (2011) catalog are used for the DFW Airprot source study.

**DFW site amplifications**

We investigate GIT H, GIT V, GIT HV (ratio of GIT H and GIT V), and HVSR estimates for the DFW local stations (Figure 5.C2). The site effect of reference site AFMOM for the DFW Airport events is estimated using site correction method suggested by Jeong et al. (2020) and
documented in Chapter 3.4.2. GIT site effects for the other stations are derived from GIT using the site corrected reference site AFMOM. For the GIT H estimates, 100-bootstrap inversions are performed to estimate the uncertainty. Most GIT V estimates approach unity except for AFDAD, which shows strong site amplification on both the horizontal and vertical components and so we excluded AFDAD from reference sites for GIT V calculation. Frohlich et al. (2011) observed locally large ground shaking in the DFW Airport study, possibly reflecting the strong amplification at AFDAD. GIT H site effects decay slightly at higher frequencies compared to HVSR, suggesting site attenuation from near-surface conditions. GIT HV estimates are in agreement with HVSR, suggesting that vertical site effects control the deviation between GIT H and HVSR. As a result, GIT H and GIT HV for the DFW Airport stations are generally comparable to HVSR consistent with a GIT V that is close to one, supporting the selection of the magnitude estimates from Frohlich et al. (2011).
Figure 5.C1. DFW Airport event catalog selection criteria. (a) Site response functions of station AFMOM from the DFW Airport sequence. The GIT site effects from the horizontal components estimated by multiple publications are illustrated as continuous lines in different colors. Vertical component (GIT V; dashed line), HVSR (white lines) with the mean ± one standard deviation (gray shaded area), the station name and the number of earthquakes (EQs) are displayed. (b) Map showing epicentral locations from different catalogs. Black lines are the location of faults from Horne et al. (2020). The seismogenic fault is highlighted as the thicker line. Black dots indicate the downthrown hanging-wall block for each fault. The blue inverted triangle represents the injection well.
Figure 5.C2. Site response functions estimated from GIT and HVSR for the 5 local stations from the DFW Airport sequence. GIT site effects from the horizontal (GIT H; white lines), vertical (GIT V; black dashed lines) components and the ratio of GIT H and GIT V (GIT HV; black lines with circles) are compared to HVSR (black lines). The gray shaded area denotes the mean ± one standard deviation for the GIT H from the 100-bootstrap resampling except for the reference site AFMOM. The AFMOM site response function and the mean ± one standard deviation are estimated using the reference site correction method. The station name and the number of earthquakes (EQs) are noted in each plot. The name of the reference station (AFMOM) is highlighted in bold.
REFERENCES


CHAPTER 6

SITE AMPLIFICATIONS FROM EARTHQUAKE DATA AND $V_{S30}$ IN THE FORT WORTH BASIN, TEXAS

Abstract

Following the development of unconventional oil and gas production across the Fort Worth Basin in Texas, a rapid increase in seismicity across the basin began in 2008 that grew to include earthquakes affecting a substantial portion of the urban metropolitan area. To assess and mitigate the seismic hazard in the area, impacted by the thickness of the sedimentary basin and accompanying soft soil layer, we estimate site effects at 22 seismic stations that recorded these events. Site responses are derived using three different approaches: (1) a modified generalized inversion technique (GIT), (2) horizontal-to-vertical spectral ratio (HVSR), and (3) the quarter-wavelength approximation (QWA). The site effects from GIT and HVSR are based on the observed S-wave Fourier amplitude spectra from earthquakes, while the QWA is calculated using estimates of average shear wave velocities in the upper 30 m ($V_{S30}$). We find that site amplification estimates based on the three techniques are consistent with one another over resonance frequencies from 2.5 to 10 Hz. The mean amplification values from the three site factors are found to be on average about 3 times larger than the vertical site response derived from GIT, which is averaged to unity. These site amplification estimates also correlate with the geology proxy $V_{S30}$ determined by rock types and geologic age, suggesting lower site
amplification for old and harder rocks (e.g., Pennsylvania limestone and sandstone) and higher amplification for young and soft rocks (e.g., Quaternary alluvium).

6.1 Introduction

During earthquake shaking, local site effects can enhance the amplitude of ground motion at resonance frequencies and hence accurate knowledge of this effect can impact building design and risk assessment. Additionally, these local site effects can impact the estimation of magnitude and kinematic source model parameters. The quantification of site response is challenging due to the factors that modify the seismic wavefield, such as anisotropic or heterogeneous characteristics of soft soil deposits, effects of local topography, depth of the water table or sedimentary basin, and near-surface attenuation (Anderson, 2003). Additionally, site amplification estimates are often poorly constrained in areas where the earthquake hazard is historically low and only recently changed with the occurrence of anthropogenically induced earthquakes.

Recent increases in earthquake rates within the Dallas-Fort Worth metropolitan area (population of nearly 7 million), North Texas, have focused interest in earthquake hazards (Frohlich et al., 2016; Petersen et al., 2016a, 2016b). Since 2008, following the development of unconventional oil and gas reserves associated with the Barnett Shale in the Fort Worth Basin (FWB), there have been a series of well-recorded earthquakes with a maximum magnitude reaching M4 (see Quinones et al., 2019 for review). The earthquakes have been shown to be induced by the disposal of wastewater associated with the natural gas production (Hornbach et al., 2015; Scales et al., 2017; Ogwari et al., 2018; Quinones et al., 2019) and occur on discrete faults that neither reach the surface nor were previously mapped (Horne et al., 2020). This recent
seismicity motivates the study of site-dependent amplification in this area to better inform hazard and risk assessments in a previously aseismic region.

The FWB geology includes a relatively thick section of preserved Mesozoic and Paleozoic sedimentary strata (see Smye et al., 2019), reaching a maximum depth of approximately 3.7 km (Montgomery et al., 2005). This sedimentary basin amplifies ground shaking due to the velocity contrast between basement and sediments as well as the resonance effects which include trapped body and surface waves within the basin (e.g., Castro et al., 2004; Sedaghati et al., 2018; Ahmadzadeh et al., 2019). Jeong et al. (2020) suggest a maximum site amplification of 5 from estimates based on a modified generalized inversion technique (GIT) and horizontal-to-vertical spectral ratio (HVSR) using S-wave amplitude spectra for 66 stations in the FWB. This previous study did not associate these site effects with site geology and rock type. Zalachoris et al. (2017) estimated the average shear wave velocity of the upper 30 meters ($V_{S30}$) using P-wave seismogram. Zalachoris and Rathje (2019) subsequently calculated the ground motion model using the $V_{S30}$ for central and eastern North America including Texas, Oklahoma, and Kansas. These two studies suggest that the ground motion as a function of $V_{S30}$ was roughly in agreement with mapped geologic units. However, the studies did not focus on detailed site amplification levels and response characteristics.

We investigate site characteristics in the FWB using both seismic waveform data and $V_{S30}$ measurements. The earthquake-specific site functions are estimated using the modified GIT and HVSR following Jeong et al. (2020). The site amplification from $V_{S30}$ data is estimated using the quarter-wavelength approach (QWA) developed by Joyner et al. (1981). Li et al. (2020) provide $V_{S30}$ estimates from in situ measurements and geologic conditions for Texas, including measurements at FWB stations, providing a good opportunity for this comparative study.
Comparison of site responses from these three techniques using two different datasets provides the basis for a comprehensive analysis, cross-validation and detailed assessment of site-dependent seismic hazard including the correlation with near-surface geologic materials and structures. Furthermore, we assess the potential of $V_{S30}$ to provide an initial proxy for local site amplification, which is an input parameter used to estimate the appropriate ground motion model for hazard assessment (e.g., Hassani and Atkinson, 2015; Zalachoris and Rathje, 2019).

6.2 Datasets

In order to estimate site amplification, we use both earthquake records and $V_{S30}$ data. The seismic data is the same as used in Chapter 3, which includes ground motion records from 90 earthquakes ($M_l \geq 2.0$) occurring near Azle-Reno, Irving-Dallas, and Venus areas in the FWB from 2013 to 2018. The data processing used to make spectral estimates including window lengths, windowing and smoothing are also described in Chapter 3.

Li et al. (2020) provide multiple estimates of $V_{S30}$, based on P-wave seismograms, in situ measured $V_{S30}$ values, and a geologic proxy $V_{S30}$ which relies on rock type and geologic age based on Cox et al. (2017) and Zalachoris et al. (2017). Here, we use the 22 in situ $V_{S30}$ estimates that correspond to the locations of the seismic stations in the FWB (Figure 6.1). Sites are divided into four groups based on mean geologic proxy $V_{S30}$ values (Figure 6.1a and Table 6.1): (1) Group A has a high $V_{S30}$ that ranges from 756 to 981 m/s and is located to the west of the Azle-Reno area with Pennsylvanian limestone and sandstone at 2 sites; (2) Group B has a reduced $V_{S30}$ of 727 m/s situated in Cretaceous limestone and chalk at 4 sites; (3) Group C has an even slower $V_{S30}$ of 543 m/s consistent with Cretaceous sandstone, chalk, and shale at 11 sites; (4) Finally, Group D has a $V_{S30}$ of 363 m/s, composed of Quaternary alluvium, mostly young and soft sediments at 5 sites. Figure 6.1 illustrates that in situ $V_{S30}$ estimates are not completely consistent
with $V_{S30}$ values based on geological settings, consistent with the findings of Zalachoris and Rathje (2019). For example, station FW0100 (#11 in Figure 6.1) has the highest $V_{S30}$ value (1230 m/s) based on field measurements, yet this station belongs to the geologic Group C. Station FW0500 (#6 in Figure 6.1) is assigned geology Group B but has a lower $V_{S30}$ value (486 m/s) than FW0100. Among 22 sites, three sites (AFDA, IPD1, and EML1) had multiple seismic instruments, which were installed at the same place but during different time periods (Figure 6.2). Note that the multiple stations have the same site numbers in Figure 6.1. To track changes in sensor or configuration as a function of time, we distinguish seismic stations by combining location code and station name (e.g., AFDA__ and AFDA01, see Table 6.2). Consequently, 26 seismic stations are analyzed and compared to 22 in situ $V_{S30}$ measurements. The detailed parameters (station locations, $V_{S30}$, instrument, and deployment condition) for each station are documented in Table 6.2.

### 6.3 Methodology

The generalized inversion technique (GIT) site response is estimated following the methodology described in Chapter 3. HVSR is calculated using the average spectral ratio between the horizontal and vertical $S$-wave spectra, as used in the GIT inversion, under the assumption that the vertical component has little or no site amplification. The QWA is used to estimate a simple synthetic site amplification (Joyner et al., 1981). Boore and Joyner (1997) argue that QWA provides a good estimate of the mean value of the response for various soil types classified by $V_{S30}$. A resonance frequency is given as:

$$f(z) = \frac{\tilde{\beta}(z)}{4z},$$  

(6.1)
where $\bar{\beta}(z)$ is average shear wave velocity and $z$ is the thickness between two layers (generally surface and a particular depth). Equation (6.1) is based on constructive interference with the resonance frequency at wavelengths that are 4 times layer thickness (i.e., quarter-wavelength condition). The QWA site amplification factor $I^{QWA}(f)$ is estimated as:

$$I^{QWA}[f(z)] = \frac{\sqrt{\rho_s \beta_s}}{\sqrt{\bar{\rho}(z)\bar{\beta}(z)}}.$$  (6.2)

where $\rho_s$ and $\beta_s$ are the density and the shear wave velocity at seismic source (i.e., reference rock in deeper depth). $\bar{\rho}(z)$ and $\bar{\beta}(z)$ are an average density and S-wave velocity between the surface and depth $z$.

In this study, $\bar{\beta}(z)$ is taken to be $V_{S30}$. The quarter-wavelength frequency corresponds to the 30 m depth at each seismic station (i.e., $z = 30$ m). Because most earthquakes used in this study occur in the basement rooted faults, we assume the basement rock is the reference rock; thus, $\beta_s$ is 3.46 km/s for the FWB as suggested by Quinones et al. (2019). To assess the effect of reference shear wave velocity in the calculation of the QWA, we also apply a slower shear wave velocity of 3.0 km/s consistent with the reference rock condition in central and eastern North America (Hashash et al., 2014). We assume that there is no significant change in densities between the reference and surface rocks. The velocity contrasts dominate the synthetic site amplification estimates under these assumptions. The density in the basement is known to be 2.65 g/cm$^3$ (Geotechnical Corporation, 1964). If we take densities that ranged from 2.3 to 2.65 g/cm$^3$ based on surface geology, the density differences are estimated from 0.93 to 1 ($= \sqrt{2.3/2.65}$ to $\sqrt{2.65/2.65}$), suggesting a negligible effect. Similarly, previous studies have calculated QWA amplifications without considering the density terms because of either small density contrasts or few reliable density estimates (e.g., Poggi et al., 2011; Hashash et al., 2014).
6.4 Results

Amplifications at multiple sites are estimated based on the modified GIT, HVSR, and QWA using the earthquake ground motions and the *in situ* $V_{S30}$ data. The results from the different approaches are compared to investigate the site amplification levels and response characteristics in the FWB including a thick sedimentary basin and increased seismicity associated with oil and gas production.

Before general site analysis, site amplification estimates using multiple instruments installed at the same location are compared (Figure 6.2). The first column (from left) in Figure 6.2 depicts the GIT H site functions for the stations with multiple seismometers - AFDA, IPD1, and EML1. GIT H amplitude at the resonance frequency for AFDA01 is smaller than that for AFDA__ but similar to the QWA site response. For station IPD1, we observe the largest GIT H amplification for IPD1__, mid-level for IPD100, and the smallest site amplification for IPD1V01. Here, the IPD1V01 is closest to the QWA value. The GIT H site amplification from EML100 and EML101 are consistent, whereas EML101 more closely matches the QWA value. The distinct amplification factors for multiple sensors installed in the same site are also shown for GIT V site functions with trends similar to GIT H (second column in Figure 6.2). By contrast, HVSR produces a similar shape and amplification between multiple instruments (third column in Figure 6.2). HVSR for IPD1 and EML1 are in good agreement with QWA, whereas, for station AFDA, HVSR is larger than QWA. Finally, we estimate the ratio of GIT site function between horizontal and vertical components (GIT HV) for the three sites (fourth column in Figure 6.2). The pattern of GIT HV is similar to HVSR for all stations. These results imply that some possibly common factor amplifies the GIT site effects of AFDA__, IPD1__, and IPD100 on both horizontal and vertical components. The magnifications are eliminated using the ratio of
horizontal and vertical components (e.g., HVSR and GIT HV). We compare the site amplification from AFDA and IPD1 with nearby posthole stations, FW0400, FW1100 and FW0900 (Savvaidis et al., 2019). The FW0400 and FW1100 are 12.02 km and 13.36 km from AFDA while the FW0400 and FW0900 are 4.69 km and 4.17 km from IPD1, respectively (Table 6.2). Because FW0900 has no $V_{S30}$ data, we estimate the resonance frequency using $V_{S30}$ from IFBF00 due to the close separation (~0.1 km). The GIT H site amplification from FW0400 and FW1100 are more consistent with AFDA01 than AFDA__ (Table 6.3). The GIT H of FW0900 is 2.28 at the resonance frequency, suggesting that, among IPD1 stations, the IPD1V01 site value is closest to both FW0400 and FW0900 site functions. Based on the fact that GIT H amplification estimates are consistent with QWA and the posthole stations, we select the stations AFDA01, IPD1V01 and EML101 as the representative stations for these sites. This result suggests that QWA estimates can be used to assess consistent instrument responses when multiple sensors are deployed at the same location.

For the 22 sites, the mean amplification factors of GIT H, HVSR, and QWA are 3.02, 3.41, and 2.82 with one standard deviation of 1.59, 1.57, and 0.46, respectively (Figure 6.3a-c). The histograms of GIT H and HVSR document larger scatter than QWA consistent with the standard deviation estimates. The median values of GIT H, HVSR, and QWA are 2.95, 2.79, and 2.87, similar to the mean value comparisons. The average values indicate that the three site response techniques produce consistent estimates with a factor of 3 amplification at the resonance frequency, less than the peak amplification values of GIT H and HVSR as high as 5 in the study from Chapter 3. Vertical site effects for GIT average close to unity, 1.02 and 0.98 for mean and median values (Figure 6.3d). All site amplification values are documented in Table 6.3. The QWA site amplification with a smaller reference velocity of 3.0 km/s from central and
eastern North America (Hashash et al., 2014) shows that the mean and median values are 0.2 smaller than the QWA using the shear wave velocity of 3.46 km/s, illustrating the weak influence of the reference velocity assumption (Figure 6.3a).

Individual site amplifications from the QWA, GIT H, HVSR and GIT V are associated with the four geologic classifications (Figure 6.4a). Both GIT H and HVSR site functions for AZWR__ show large scatter even though the site belongs to the geologic Group A. Site amplification in Group B are relatively stable for all approaches. In Group C, several stations (e.g., AZE2__, AFDA01, and IFS300) reveal stronger amplification for GIT H or HVSR, whereas FW0100 shows the lowest site amplifications from all techniques, corresponding to the highest $V_{S30}$ value. In Group D, HVSR at IFBF00 and IFDF00 are amplified. Figure 6.4b displays the average site response for each geologic classification. The QWA site amplification based on $V_{S30}$ measurements clearly correlates with the local geological conditions, although in situ $V_{S30}$ departs somewhat from the geologic proxy $V_{S30}$ (Figure 6.1). GIT V amplification shows a linear increase with geologic conditions except for the slightly larger values in Group A. Also, we observe a linear correlation for Group B to D with large variations in Group A for the GIT H and HVSR amplifications. Considering that there are only two stations in Group A, we conclude that the site amplifications estimated by the three techniques roughly correlate with geological conditions.

Next, we investigate the details of 26 site functions including multiple sensors. We find that the GIT H for 16 stations are in good agreement with those from QWA at the quarter-wavelength frequency (Figure 6.5). HVSR estimates are also consistent with the QWA except for several stations (e.g., AFDA01, IFS300, IFBF00, and IFDF00). Estimates of site responses
for AZDA01 and ILCC00 show that GIT H and HVSR site estimates differ from the QWA by less than 0.5 amplification units.

For 8 stations, HVSR better fits the QWA values than the GIT H estimates at the resonance frequency (Figure 6.6 top and middle rows). The difference between GIT H and HVSR may depend on the type of reference. GIT H is estimated as a relative response to the single reference station (AZDA01) whereas HVSR uses vertical site spectra as the reference. Stations FW0500, ITSC00, and FW0100 with larger HVSR all have estimates of GIT V that are smaller than 1 at the resonance frequency. In contrast, stations AZE2__, FW0400, IPD1__, and IPD100 have smaller site amplification estimates based on HVSR compared to GIT H, which appears to be related to higher GIT V site estimates (Table 6.3). Similarly, stations AFDA01 and IFS300 have a significantly different HVSR from GIT H, as a result of the smaller GIT V estimate (Figures. 6.4a and 6.5). Based on these results, we suggest that GIT V controls the documented HVSR variability. Thus, for these stations, QWA may not properly capture the vertical site amplification. For the case of ITL100, GIT H is smaller than HVSR, but GIT V is greater than 1, inconsistent with the hypothesis that these differences are a result of GIT V. Also, IFBF00 and IFDF00 show higher HVSR with a unity of GIT V (Figures 6.4a and 6.5).

At stations AZWR__ and AFDA__ GIT H and HVSR are amplified relative to QWA (Figure 6.6 bottom row). For these two stations, both GIT H and HVSR are amplified at frequencies from 3 to 10 Hz. The anomaly of AFDA__ may result from the unknown amplification factor (see Figure 6.2) because the GIT H of AFDA01 is consistent with the QWA (Figure 6.5). For AZWR__, we suggest that the QWA does not include multiple peaks and valleys that can result from the interference of multiple reflected waves (Kwak et al., 2017). The
significant peaks from AZWR__ also impact to relationship between geology and site amplifications (Figure 6.4).

The GIT H site estimates decay at high frequency relative to the HVSR estimates (e.g., FW1100, IFS300, EML101, ITL100, etc.) even though the GIT V exhibits a stable shape. Because the modified GIT site effect assumes a Brune (1970)’s source model, the GIT site estimate includes any remaining attenuation at high frequency that decays more rapidly than $\omega^{-2}$ at high frequencies (Oth et al., 2011). This high-frequency decay may be related to local near-surface attenuation at or near the receiver reflecting low-velocity materials directly below the seismic station facilitating high-frequency attenuation (e.g., Anderson and Hough, 1984; Ktenidou et al., 2015).

6.5 Discussion

Twenty-two site amplifications corresponding to the 26 seismic stations including multiple seismometers at the same site are estimated in the FWB using three site response techniques including the modified GIT, HVSR, and QWA with $V_{S30}$. The site amplifications derived from different approaches show roughly consistent values producing an average amplification of 3 at the resonance frequency for the 30 m layer. The average amplifications correlate with geological characteristics and rock types. In contrast, we identify some differences between site response functions estimated from the three approaches (Figures 6.5 and 6.6). Here, we discuss the cause of the discrepancy and the appropriateness of $V_{S30}$ as a standard site approach.

Site response spectra were estimated in a limited number of cases where different instruments were installed at nearly the same location (Figure 6.2). Both horizontal and vertical
components of the GIT spectra for stations AFDA__, IPD1__, and IPD100 are amplified compared to estimates for AFDA01 and IPD1V01. The study from Chapter 3 (Jeong et al., 2020) reports possible instrumental calibration values at IPD1__, which produce frequency-dependent amplitude anomalies (Figure 3.11a). Except for IPD1__, the cause of magnification observed on other sensors remains unclear. AFDA01 and IPD1V01 are L28 sensors, which are attached to concrete using some modeling clay or putty, while the broadband seismometers were directly installed on the concrete-floored building. This installation difference may impact the estimate of the site response function. Site noise as a function of installation conditions or cultural sources are often found to be negligible (Ringler et al., 2014). However, this study suggests that individual site noise may impact the interpretation of the site amplification and under such conditions QWA may be used to evaluate the quality of the estimates.

We find that GIT V spectral variability mirrors differences between GIT H and HVSR (Figures 6.5 and 6.6). Previous studies suggest that vertical amplification results can be impacted from converted phases such as $S$-$P$ or $P$-$S$ conversions (e.g., Oth et al., 2011; Huang et al., 2016; Hrubcová et al., 2016). Converted phases are observed on the vertical components of the seismic waveform recorded in the FWB (Figures 3.3g and 4.3e). Castro et al. (2004) suggest that the location of the site (border or middle on the basin) can also produce significant GIT V variations. In Figure 6.6 the results in the upper two rows are mostly stations located near the edge of the basin in the eastern FWB except for AZE2__, which is located in the middle of the basin (Figure 6.1). Waves recorded on the stations near the eastern edge of the basin have to pass through the metasedimentary units associated with the Ouachita thrust front, which intersects with the Ellenburger formation underlying the Barnett Shale (see Magnani et al., 2017). These local geologic and geomorphologic influences can increase the variability in HVSR estimates.
(Sokolov et al., 2005). For the cause of the larger GIT V at AZE2__ compared to nearby stations with similar surface geology, we suggest that either the station location close to a road or its different installation condition (installed on a ceramic tile; see Table 6.2) may contribute to the site response, but additional study is needed to support this argument.

Site amplification at station ITL100, IFBF00, and IFDF00 are not easily explained by differences in the GIT V site spectrum. In the ITL100 case, HVSR shows good agreement with QWA; thus GIT site amplification may have other contributions. The GIT site amplification is estimated using an averaged path attenuation function estimated for the total dataset and hence depth-dependent path attenuation changes (Oth et al., 2011; Abercrombie et al., 2020) or geological discrepancies (Oth et al., 2008) are not well addressed. Since most events used to estimate the GIT site amplification at ITL100 are from local recordings within 7.5 km of the hypocenter, it is possible that this local dataset may produce biased estimates. In contrast, for IFBF00 and IFDF00, HVSR are more amplified than GIT H and the QWA with almost unity of GIT V. Parolai and Richwalski (2004) suggest that sources need to be azimuthally distributed around a station at a variety of distance for reliable HVSR estimates as HVSR depends on the incidence angle of the waves (Lermo and Chavez-Garcia, 1993). Because the local seismic networks are placed over clustered earthquakes sequences in the FWB (DeShon et al., 2018), ray paths are not well distributed and therefore the recordings produce a narrow range of crossing ray paths for one station. For IFBF00 and IFDF00, azimuths have narrow ranges, 169 to 180 degrees and 355 to 3 degrees, respectively. Finally, the three stations were deployed next to a human-made structure (e.g., levee and fire department). Thus, the structure may impact site noise.
Despite the result of this study, the use of $V_{S30}$ to estimate site amplification factors remains controversial (Poggi et al., 2012). Castellaro et al. (2008) suggest that $V_{S30}$ appears to be a weak proxy for seismic amplification arguing the site amplification cannot be quantified by a single parameter such as $V_{S30}$. Park and Hashash (2004) suggest that the site response is impacted by the deeper soil (>30 m). Figure 6.7 illustrates the comparison of peak site responses and resonance values for 30 m deposits. The QWA resonant frequencies and amplifications are clustered, but those from GIT and HVSR site spectra show greater scatter. The average peak frequencies are 5.53 and 5.02 Hz with one standard deviation of 2.27 and 3.20 Hz for both GIT H and HVSR. These are slightly higher than the resonance frequency with a mean of 4.01 Hz and standard error of 1.8 Hz for QWA. For the GIT H and HVSR, the commensurate amplifications are 5.20 and 5.84 on average, with one standard deviation of 2.01 and 2.06. Thus, the average peak amplitudes of the site effects are larger than the QWA values of ~3. We observe that the resonance frequencies usually occur at frequencies lower than the location of the spectral bump in HVSR or GIT H (Figures 6.5 and 6.6). These results suggest that the resonance frequency based on the assumed thickness of 30 meters is not good for estimating the maximum amplification at each site. The shallower or deeper velocity profile may be more proper to estimate peak site amplitude. Although the resonance frequency is not identical to the peak frequency, the utility of $V_{S30}$ is still useful as the resonance frequencies ranged from 2.5 to 10 Hz, which overlaps with the fundamental vibration frequency for 1 to 8 story buildings (2-10 Hz) (Castellaro et al., 2008).

6.6 Conclusions

In order to improve the estimate of earthquake hazard in the FWB area following increases in seismicity associated with unconventional oil and gas production, site amplification
functions are estimated using three different techniques (modified GIT, HVSR, and QWA) based upon seismic records and a $V_{S30}$ dataset for 22 sites (26 seismic stations). Although there are some discrepancies between estimates using these three site response techniques, probably due to possible local effects and installation impacts, we conclude that site amplification factors using the three techniques are roughly consistent with each other. Site amplification estimates generally correlate with geologic age and rock types (lower site amplification for old and harder rocks and higher amplification for young and soft rocks) with an average site amplification of approximately a factor of 3 at the resonant frequency. Although the resonant frequencies are not good for estimating the largest amplifications, $V_{S30}$ parameters provide a good estimate for site effects at quarter-wavelength frequencies from 2.5 to 10 Hz, based on an assumed 30 m depth estimate of the resonant layer. In addition, the QWA site functions derived using $V_{S30}$ provide insight for more reliable seismic installations when comparing multiple sensors deployed at the same location.

Li et al. (2020) provide $V_{S30}$ data for other seismic stations in both west and east Texas. These additional characterizations illustrate an opportunity to further extend this study to wider regions across Texas including a broader range of surface conditions.
Figure 6.1. Map illustrating (a) the geologic proxy $V_{S30}$ (average shear wave velocities in the upper 30 m) and (b) in situ measured $V_{S30}$ provided by Li et al. (2020). The $V_{S30}$ values are spatially interpolated using the Kriging method based on a grid of points. The sites used in this study are illustrated as squares with numbers. The site numbers increase from geologic Group A to D (see Table 6.1 for details). Site numbers 14, 15, 16, 17, 19, and 21 are not noted due to their proximity to #22. The GMT was used to make the figure (Wessel et al., 2013).
Figure 6.2. Comparison of site amplification for sites with multiple sensors: AFDA (top), IPD1 (middle), and EML1 (bottom). Horizontal site functions from a generalized inversion technique (GIT H), vertical GIT site amplification (GIT V), horizontal-to-vertical spectral ratio (HVSR), and the ratio of GIT H and GIT V (GIT HV) are plotted from left to right for each station. The red circle notes the quarter-wavelength approximation (QWA) using $V_s^{30}$ and the horizontal line is provided for visual reference. BP, broadband sensor; SP, Short-period sensor; SM, strong-motion accelerometer.
Figure 6.3. Histograms of site amplification estimates from (a) QWA, (b) GIT H, (c) HVSR, and (d) GIT V at the resonance frequency using a layer thickness of 30 m. The default velocity for basement is taken from Quinones et al. (2019). In (a), the darker histogram represents the distribution assuming the shear wave velocity of 3.00 km/s for reference (Hashash et al., 2014).
Figure 6.4. Site amplification factors at the resonance frequency from QWA (red circles), GIT H (green squares), HVSR (magenta diamonds), and GIT V (blue inverse triangles) for (a) 22 sites and (b) 4 geologic classifications. In (a), vertical lines and capital letters document the geologic groups. In (b), averaged site amplification coefficients and ± one standard deviations are shown for each geologic classification.
Figure 6.5. Site amplification estimates for 16 stations using the QWA, GIT H, HVSR, and GIT V. For these stations, GIT H and QWA produce similar estimates. Capital letters and numbers in parentheses denote geologic group and site numbers in Figure 6.1.
Figure 6.6. Site amplifications for the 8 stations (top and middle rows) where HVSR provides better agreement with QWA estimates than GIT H site functions. The 2 stations in the bottom row show little similarity between QWA and GIT H or HVSR.
Figure 6.7. Comparison of (a) peak frequency and (b) peak amplification for GIT H (green squares) and HVSR (magenta diamonds) site response estimates as a function of QWA values. Lines represent a ratio of 1:1.
Table 6.1. Station classifications based on geologic proxy $V_{S30}$.

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<tr>
<th>Geologic group</th>
<th>Geologic proxy $V_{S30}$ (m/s)</th>
<th>Rock Type</th>
<th>Period</th>
<th>Site # in Figure 6.1</th>
<th># of sites (stations)</th>
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<tr>
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<td>543</td>
<td>Sandstone, chalk, shale</td>
<td>Lower and Upper Cretaceous</td>
<td>7-17</td>
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<td>D</td>
<td>363</td>
<td>Alluvium</td>
<td>Quaternary</td>
<td>18-22</td>
<td>5 (8)</td>
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<td># in Fig. 6.1</td>
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<td>Longitude (°)</td>
<td>$V_{S30}$ (m/s)</td>
<td>Instrument (Type*)</td>
</tr>
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* BP = Broadband, SP = Short-period, SM = Strong-motion accelerometer
Table 6.3. Site amplification results in this study

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<th>QWA ((\beta_s=3.4\text{km/s}))</th>
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<th>HVS R</th>
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REFERENCES


CHAPTER 7
SUMMARY AND CONCLUSIONS

Spectral characteristics of ground motion from induced earthquakes associated with oil and gas production in the Fort Worth Basin (FWB), Texas are analyzed in this study separating path, site, and source effects. A generalized inversion technique (GIT) is used to separate these contributions from observed seismic data. Since there is not a hard-rock site in the FWB, a novel approach was developed to correct for the site effects based on a reference station. Path and site effects estimated in this study are consistent with previous studies in similar basin structures. The earthquake source characteristics and especially stress drops estimated from events in the Cleburne, Azle-Reno, Irving-Dallas, and Venus sequences are comparable to those for tectonic earthquakes. In contrasts, stress drops for the Dallas-Fort Worth (DFW) earthquakes are statistically lower with distinct spatial variations relative to the nearest injector.

In Chapter 3, ground motion parameters including path, site, and source spectra are separated using the modified GIT procedure for induced earthquakes in Azle-Reno, Irving-Dallas, and Venus, Texas. The original GIT relies on one or more reference sites (usually hard-rock site) to resolve the trade-off between source and site functions (e.g., Oth et al., 2011). Due to the lack of hard-rock site in the FWB, a modified GIT is developed with implementation of a single site correction method. Path attenuation curves document a transition at 30 km, which appears to correspond to a mid-crustal boundary (Keller and Hatcher, 1999). Path attenuation is
parameterized to include geometrical spreading \((1/r)\) and quality factor \(Q\). Estimates of the ratio of the \(Q_s/Q_p\) is larger than one, suggesting the attenuation in the basin is impacted by partially fluid-saturated materials (e.g., Drwila et al., 2019). Site effects document strong amplification as large as a factor of 5 for most of the stations, reflecting the thick sedimentary basin (e.g., Sedaghati et al., 2018). S-wave stress drops are averaged as 4.46 MPa and range from 1.18 to 21.73 MPa, similar to those for tectonic earthquakes, indicating roughly self-similarity as a function of event magnitude, although the range of magnitudes is narrow.

Chapter 4 includes the detailed S-wave stress drop analysis for all earthquakes occurring in the four earthquake sequences adding the Cleburne sequence to the three earthquake sequences analyzed in the Chapter 3. In order to validate the GIT approach, results are compared to source estimates using the relative EGF method where a near-by small earthquake is used as an empirical Greens function. Corner frequency differences between the GIT and EGF methods are 0.31 Hz on average, suggesting that the two methods produce similar estimates. GIT moment magnitude estimates are also found to be consistent with eastern north America \(M_w\) from the relation of Rigsby et al. (2014). These comparisons validate the path and site corrections developed in the modified GIT and the subsequent source spectra estimates. The mean stress drop estimate for all FWB earthquakes is 5.33 MPa, consistent with average stress drops for intraplate earthquake which vary from 1 to 10 MPa (e.g., Kanamori and Anderson, 1975) and for injection-induced earthquakes near the study area (Reiter et al., 2012; Boyd et al., 2017; Wu et al., 2018). Furthermore, stress drops show no correlation with time, space, and pore pressure changes provided by Quinones et al (2018). The stress drops follow self-similar scaling, although only over a very limited range of magnitude. Based on these results, the FWB earthquakes source characteristics are determined to be similar to tectonic earthquakes at the same depths.
In Chapter 5, one anomaly in the stress drop estimates is explored where lower stress drops are observed for the DFW Airport earthquake sequence. In addition, DFW Airport stress drops increase with radial distance from the injection wellhead over the initial 1.5 km distance (all other sequences are much further from injectors). Unlike the other earthquake sequences in the FWB, the Airport earthquakes began within two months of the initiation of wastewater injection, suggesting that immediate pore pressure increases may be related to the lower stress drops. Pore pressure models from Gao et al. (2019) suggest that the absolute pore pressure changes in DFW area are lower than for other FWB area earthquake sequences. As a result, absolute pore pressure changes (or cumulative amount of the injected wastewater), seismic moments, and focal depths may not be the only controlling factors contributing to stress drop variations; consideration of the rate of pore pressure change and distance separation from injector to fault may be important.

Finally, in Chapter 6, a more detailed study of site amplification is conducted than in Chapter 3. Three site amplification estimates are developed including the modified GIT, horizontal-to-vertical spectral ratio (HVSR), and quarter-wavelength approximation (QWA) using seismic records and average shear wave velocities in the upper 30 m ($V_{S30}$) data (Li et al., 2020) for 22 of the sites (26 seismic stations) in the FWB. Site amplification estimates using the three techniques are roughly consistent with one another with a few exceptions (e.g., AZWR and ITL100). These site amplifications generally correlate with geologic age and rock types with an average site amplification of about 3 at the resonance frequency, slightly smaller than peak amplifications (factor of 5 on average) estimated from GIT H and HVSR. This study suggests that surface geology may influence the site effect. Also, consistent results among the three
different approaches validates the use of $V_{s30}$, which is the simplest technique to estimate site amplification.

In conclusions, this dissertation represents a complete study of the induced earthquakes in the FWB designed to explore two topics: (1) seismic hazard assessment (measure of ground shaking); (2) source discrimination between injection-induced events and tectonic intraplate earthquakes. To achieve the goal, the ground motion records are separated into path attenuation, site amplification, and source processes using the modified GIT, providing reference site constraint to regions having no hard-rock stations.

The stress drops for most of the induced earthquakes in this study suggest that subsurface stress perturbations from wastewater injection promote slip on faults whereas the stress drops are a function of pre-existing tectonic stresses. Hence, seismic ground shaking estimates from the injection-induced earthquakes should be assessed with this stress drop level, while taking into account any basin specific propagation path effects (mid-crustal boundary and partially fluid saturated material) and local site amplification (factor of 5). Understanding these combined effects is key to improving seismic hazard assessment and represents an initial step towards better risk mitigation for the FWB earthquakes that occur within a metropolitan area with a population of 7 million.

This unique result suggests that at least two modes of wastewater injection triggered seismicity exist in the FWB. Most events are triggered on faults located at distances less than 1.5 km from wells through diffusion of pore pressure and/or poroelastic stress; these events release energy left on pre-existing faults and hence have stress drops in line with tectonic intraplate earthquakes. The DFW Airport example, however, shows that earthquakes triggered on faults near injection wells (<1.5 km for FWB) exhibit relatively lower stress drop values and a direct
dependence on the rate of pore pressure change. These unique results suggest that stress drop can be an important tool in distinguishing causal physical mechanism in areas of induced seismicity. Stress drop values alone cannot distinguish between induced and tectonic earthquakes. If the relationship between the lower stress drop and pore pressure increase as a function of distance is further validated, it could provide the physical basis for identification of the induced seismicity and restriction of the volume of wastewater injected in the seismically active areas. In Oklahoma, a 40% reduction of wastefluid injection has led to decreases in seismicity (Langenbruch and Zoback, 2016).

This dissertation includes a newly developed GIT which is applicable to areas with no hard-rock reference site. GIT source spectra compare favorably to those estimated using the EGF source spectra. GIT site functions are roughly consistent with HVSR and QWA. These results validate the appropriateness of path and site corrections developed in the new method.

The approaches used on this dissertation can be applied to other study areas (e.g., West Texas showing significant induced earthquakes). Such detailed source studies can be used to document the spatial-temporal evolution of source parameters and possibly infer additional effects of interevent times (e.g., Cochran et al., 2018) and source directivity (e.g., Ameri et al., 2020). Additionally, $V_{S30}$ data provided by Li et al. (2020) for other Texas seismic stations in Texas can be compared to site effects covering a greater range of geologic environments across Texas.
REFERENCES


