CHARACTERIZATIONS OF LINEAR GROUND MOTION SITE RESPONSE IN THE NEW MADRID AND WABASH VALLEY SEISMIC ZONES AND SEISMICITY IN THE NORTHERN EASTERN TENNESSEE SEISMIC ZONE AND ROME TROUGH, EASTERN KENTUCKY

Nicholas von Seth Carpenter
University of Kentucky, seth.carpenter@uky.edu
Author ORCID Identifier: https://orcid.org/0000-0002-7664-9071
Digital Object Identifier: https://doi.org/10.13023/etd.2020.014

Recommended Citation
Carpenter, Nicholas von Seth, "CHARACTERIZATIONS OF LINEAR GROUND MOTION SITE RESPONSE IN THE NEW MADRID AND WABASH VALLEY SEISMIC ZONES AND SEISMICITY IN THE NORTHERN EASTERN TENNESSEE SEISMIC ZONE AND ROME TROUGH, EASTERN KENTUCKY" (2019). Theses and Dissertations--Earth and Environmental Sciences. 77.
https://uknowledge.uky.edu/ees_etds/77
STUDENT AGREEMENT:

I represent that my thesis or dissertation and abstract are my original work. Proper attribution has been given to all outside sources. I understand that I am solely responsible for obtaining any needed copyright permissions. I have obtained needed written permission statement(s) from the owner(s) of each third-party copyrighted matter to be included in my work, allowing electronic distribution (if such use is not permitted by the fair use doctrine) which will be submitted to UKnowledge as Additional File.

I hereby grant to The University of Kentucky and its agents the irrevocable, non-exclusive, and royalty-free license to archive and make accessible my work in whole or in part in all forms of media, now or hereafter known. I agree that the document mentioned above may be made available immediately for worldwide access unless an embargo applies.

I retain all other ownership rights to the copyright of my work. I also retain the right to use in future works (such as articles or books) all or part of my work. I understand that I am free to register the copyright to my work.

REVIEW, APPROVAL AND ACCEPTANCE

The document mentioned above has been reviewed and accepted by the student’s advisor, on behalf of the advisory committee, and by the Director of Graduate Studies (DGS), on behalf of the program; we verify that this is the final, approved version of the student’s thesis including all changes required by the advisory committee. The undersigned agree to abide by the statements above.

Nicholas von Seth Carpenter, Student

Dr. Edward W. Woolery, Major Professor

Dr. Edward W. Woolery, Director of Graduate Studies
CHARACTERIZATIONS OF LINEAR GROUND MOTION SITE RESPONSE IN THE NEW MADRID AND WABASH VALLEY SEISMIC ZONES AND SEISMICITY IN THE NORTHERN EASTERN TENNESSEE SEISMIC ZONE AND ROME TROUGH, EASTERN KENTUCKY

DISSERTATION

A dissertation submitted in partial fulfillment of the requirements for the degree of Doctor of Philosophy in the College of Arts and Sciences at the University of Kentucky

By

N. Seth Carpenter

Lexington, Kentucky

Co-Directors: Dr. Edward W. Woolery, Professor of Geophysics and Dr. Zhenming Wang, Professor of Seismology

Lexington, Kentucky

2019

Copyright © N. Seth Carpenter 2019
https://orcid.org/0000-0002-7664-9071
ABSTRACT OF DISSERTATION

CHARACTERIZATIONS OF LINEAR GROUND MOTION SITE RESPONSE IN THE NEW MADRID AND WABASH VALLEY SEISMIC ZONES AND SEISMICITY IN THE NORTHERN EASTERN TENNESSEE SEISMIC ZONE AND ROME TROUGH, EASTERN KENTUCKY

The central and eastern United States is subject to seismic hazards from both natural and induced earthquakes, as evidenced by the 1811-1812 New Madrid earthquake sequence, consisting of at least three magnitude 7 and greater earthquakes, and by four magnitude 5 and greater induced earthquakes in Oklahoma since 2011. To mitigate seismic hazards, both earthquake sources and their effects need to be characterized.

Ground motion site response can cause additional damage to susceptible infrastructure and buildings. Recent studies indicate that $V_s30$, one of the primary site-response predictors used in current engineering practice, is not reliable. To investigate site response in the New Madrid Seismic Zone, ratios of surface-to-bedrock amplitude spectra, $T_F$, from S-wave recordings at the two deep vertical seismic arrays in the sediment-filled upper Mississippi Embayment (i.e., VSAP and CUSSO) were calculated. The mean $T_F$ curves were compared with theoretical transfer functions; the results were comparable, indicating that $T_F$ estimates of the empirical, linear SH-wave site responses at these sites. The suitability of surface S-wave horizontal-to-vertical spectral ratios, $H/V$, for estimating the empirical site transfer function was also evaluated. The results indicate that mean S-wave $H/V$ curves are similar to $T_F$ at low frequencies (less than the fifth natural frequencies) at both CUSSO and VSAP.

SH-wave fundamental frequency, $f_0$, and fundamental-mode amplification, $A_0$, were evaluated as alternatives to the $V_s30$ proxy to estimate primary linear site-response characteristics at VSAP, CUSSO, and nine other seismic stations in the CEUS. In addition, calculated $f_0$ and $A_0$ were compared with the first peaks of S-wave $H/V$ spectral ratios. The $f_0$ and $A_0$ were found to approximate the 1-D linear, viscoelastic, fundamental-mode responses at most stations. Also, S-wave $H/V$ from weak-motion earthquakes can be used to measure $f_0$. However, S-wave $H/V$ does not reliably estimate $A_0$ in the project area. S-wave $H/V$ observations reveal site response within the frequency band of engineering interest from deeper, unmodeled geological structures.

Because damaging or felt earthquakes induced by hydraulic fracturing and wastewater disposal have occurred in the CEUS, characterizing background seismicity prior to new large-scale subsurface fluid injection is important to identify cases of and the
potential for induced seismicity. The Rogersville Shale in the Rome Trough of eastern Kentucky is being tested for unconventional oil and gas potential; production of this shale requires hydraulic fracturing, which has been linked to induced seismicity elsewhere in the CEUS. To characterize natural seismicity and to monitor induced seismicity during testing, a temporary seismic network was deployed in the Rome Trough near the locations of new, Rogersville Shale oil and gas test wells. Using the real-time recordings of this network and those of other regional seismic stations, three years of local seismicity were cataloged. Only three earthquakes occurred in the Rome Trough of eastern Kentucky, none of which was associated with the deep Rogersville Shale test wells that were stimulated during the time the network was in operation.

KEYWORDS: Seismic Hazard, Site Effect, Site Response, Rome Trough, Intraplate Earthquakes, Induced Seismicity
CHARACTERIZATIONS OF LINEAR GROUND MOTION SITE RESPONSE IN THE
NEW MADRID AND WABASH VALLEY SEISMIC ZONES AND SEISMICITY IN
THE NORTHERN EASTERN TENNESSEE SEISMIC ZONE AND ROME TROUGH,
EASTERN KENTUCKY

By
N. Seth Carpenter

Edward W. Woolery
Co-Director of Dissertation

Zhenming Wang
Co-Director of Dissertation

Edward W. Woolery
Director of Graduate Studies

12/4/2019
Date
DEDICATION

To my wife and best friend Sally, who never fails to love, encourage, and support me. Strength and honor are your clothing and your worth is far above rubies. And to my daughters Adeline, Esther, Naomi, and Greta who add abundant joy to my days, make me want to be a better person, and help me endeavor to finish my homework.
ACKNOWLEDGMENTS

The following dissertation has come about after many conversations on issues in seismic hazard that affect the central and eastern U.S. and discussions of strategies to address the issues with my Dissertation Chair, Ed Woolery, and Co-chair, Zhenming Wang. I am grateful to them not only for their insights and their constructive criticism of my research and writing, but also for their encouragement and thoughtfulness. I also wish to thank my other committee members, Dhananjay Ravat and Michael Kalinski, for their time discussing my research ideas at committee meetings and for the conversations and feedback in their classrooms and offices. I am also grateful for Scott Yost being willing to serve as the outside member of my dissertation committee.

I am grateful for the insights and feedback that I received from the technical reviewers of my manuscripts. I thank Martin Chapman (Virginia Tech. Univ.) and Charles Langston (Univ. of Memphis) especially who shared valuable insights on the methods to use and evaluate KSSMN borehole data. And I would like to acknowledge the insight and effort of Ron Street in installing the vertical seismic array in Paducah, Ky., and conceiving of the deep CUSSO vertical array.

I acknowledge KGS and the Department of Earth and Environmental Sciences at UK for continuing to support the operation of the Kentucky Seismic and Strong-Motion Network, whose data were critical for these investigations. I also express gratitude to the state of Kentucky and the Kentucky Geological Survey for funding the majority of seismicity characterization study in the Rome Trough. I also wish to acknowledge the help of KGS staff and interns, and UK visiting scholars and graduate students at who assisted
me with field work for that study and former KGS Associate Director Jerry Weisenfluh, who permitted and encouraged undertaking of the investigation.

Of even greater significance to me than the research support I have received, I acknowledge the devotion and labor of my wife, Sally. Any attempt I might make to describe the sacrifices she has made for me and our four daughters during the many times she was essentially parenting and teaching our children alone, would be a gross understatement of the reality through which she has persevered. She is worthy of much praise. I also acknowledge the encouragement and love my daughters give me and their feeding me with their inherent curiosity and excitement. And, I am thankful to Mom and Dad who have always encouraged and supported my education and development.

Of the greatest significance to me is the sustaining grace that God has given me through faith in His Son Jesus to complete the work presented in this dissertation. I am humbled and grateful for the privilege to study His creation and to marvel at the wonders He has done.
TABLE OF CONTENTS

ACKNOWLEDGMENTS ................................................................. iii
LIST OF TABLES .............................................................................. vii
LIST OF FIGURES ............................................................................ viii
LIST OF ADDITIONAL FILES .......................................................... xii
CHAPTER 1. INTRODUCTION ................................................................. 1
  1.1 SEISMIC HAZARDS INVESTIGATED IN THE CENTRAL AND
  EASTERN U.S. .............................................................................. 1
  1.2 GROUND MOTION SITE RESPONSE IN THE CENTRAL AND
  EASTERN U.S. .............................................................................. 5
  1.3 INDUCED SEISMICITY POTENTIAL AND SEISMIC SOURCE
  ZONES IN EASTERN KENTUCKY ............................................. 13
CHAPTER 2. GROUND MOTION SITE RESPONSE FROM SHEAR-WAVE
RECORDINGS AT DEEP BOREHOLES IN THE NEW MADRID SEISMIC ZONE . 17
  2.1 INTRODUCTION ......................................................................... 17
  2.2 SH-WAVE TRANSFER FUNCTION AND S-WAVE H/V ............. 18
    2.2.1 SURFACE-TO-BEDROCK SPECTRAL RATIOS ................... 19
    2.2.2 HORIZONTAL-TO-VERTICAL SPECTRAL RATIOS .......... 20
  2.3 VERTICAL ARRAYS AND DATASETS ........................................ 22
    2.3.1 VSAP ................................................................................. 23
    2.3.2 CUSSO ............................................................................. 25
  2.4 METHODS .................................................................................... 28
    2.4.1 DATA SELECTION AND PROCESSING ......................... 28
    2.4.2 SPECTRAL RATIOS ............................................................ 35
  2.5 RESULTS ..................................................................................... 35
  2.6 DISCUSSION .............................................................................. 38
    2.6.1 SH-WAVE TRANSFER FUNCTIONS ................................. 38
    2.6.2 S-WAVE H/V .................................................................... 39
    2.6.3 AMBIENT NOISE H/V ......................................................... 42
    2.6.4 ON THE APPLICABILITY OF S-WAVE H/V .................... 43
CHAPTER 3. PRIMARY SITE RESPONSE PARAMETERS IN THE CENTRAL AND
EASTERN U.S. AND THE IMPORTANCE OF EMPIRICAL SITE-RESPONSE
ESTIMATIONS .................................................................................. 45
  3.1 INTRODUCTION ......................................................................... 45
  3.2 PRIMARY SITE RESPONSE PARAMETERS .............................. 46
    3.2.1 SIMPLIFIED EXPRESSIONS OF $A_0$ AND $f_0$ ................... 46
    3.2.2 FULL-RESONANCE SITE RESPONSES ............................. 47
    3.2.3 ESTIMATING $A_0$ AND $f_0$ FROM S-WAVE H/V ........... 49
  3.3 DATA .......................................................................................... 52
    3.3.1 SEISMIC STATIONS AND S-WAVE RECORDINGS ........... 52
    3.3.2 INPUTS FOR EQUATIONS 3.1 and 3.2: V(z), $\rho$, and $\gamma$ ...... 58
    3.3.3 EMPIRICAL ESTIMATION OF $A_0$ and $f_0$ FROM S-WAVE H/V .68
  3.4 RESULTS ..................................................................................... 75
    3.4.1 SIMPLIFIED VERSUS FULL-RESONANCE $f_0$ and $A_0$ ....... 77
LIST OF TABLES

Table 2.1 Soil-profile parameters for site-response modeling at VSAP and CUSSO. α, P-wave velocity at CUSSO; β, S-wave velocity. ................................................................. 28
Table 2.2 Parameters for the earthquakes recorded by VSAP used in this study........... 29
Table 2.3 Parameters for the earthquakes recorded by CUSSO used in this study. .......... 30
Table 2.4 Data processing parameters for recordings at VSAP and CUSSO.................... 30
Table 3.1 Seismic station locations and distances to reflection/refraction surveys........... 53
Table 3.2 Station recording and H/V processing parameters, summary of earthquakes recordings used, and peak ground accelerations...................................................... 57
Table 3.3 Theoretical and observed $f_0$ and $A_0$ ............................................................ 76
Table 4.1 Long-term and temporary (stations with Off Date attributes) seismic stations used in real-time monitoring................................................................. 97
Table 4.2 Temporary seismic monitoring station locations, seismometers (T-40, Trillium 40; TC-PH2, Trillium Compact Posthole; MC-PH1, Meridian Compact Posthole), and operational time period. Stations with no Off Date were operating at the time of preparation of this chapter......................................................................................................................... 100
Table 4.3 Focal mechanism strikes, dips, and rakes of nodal plane one. ....................... 109
LIST OF FIGURES

Figure 1.1 USGS “Did You Feel It?” felt reports from one western U.S., 2016 moment magnitude (Mw) 6.0 Napa, Calif., and two central and eastern U.S. earthquakes of comparable, albeit slightly less, magnitudes.......................................................... 1

Figure 1.2 Epicenters of magnitude 3 and greater earthquakes from 1776 through 2008 (USNRC, 2012) and of the 2012 Perry County, Ky. earthquake and Precambrian faults from Hickman (2011). The boundary of the speculative Rogersville Shale play is also shown. 4

Figure 1.3 Schematic representation of the terms in equation 1.2 that produce ground accelerations experienced at surface receivers. ................................................................. 5

Figure 1.4 (a) The locations of four sites along the Ohio and Mississippi Rivers. Contours show depth to bedrock in the Mississippi Embayment. (b) Shear-wave velocity profiles for each site.......................................................................................................................... 9

Figure 1.5 Equivalent linear, 1-D spectral amplification functions f for the four sites along the Ohio and Mississippi Rivers in Figure 1.4 and a 0.1 g PGA input time history........... 12

Figure 1.6 Annual number of magnitude 3 and greater earthquakes in the central United States.................................................................................................................. 14

Figure 2.1 Locations of sensors (triangles) at the surface at soil (S) and rock-outcrop (R) sites, and beneath the soil in bedrock (B). H and V represent amplitude spectra of horizontal- and vertical-component recordings, respectively, at these locations. ............ 18

Figure 2.2 Vertical seismic arrays CUSSO and VSAP, in the northern Mississippi Embayment and the earthquakes they recorded (gray or black depending on the recording array). ................................................................. 22

Figure 2.3 Simplified stratigraphic column, sensor depths (stars), and shear-wave velocity structure at VSAP. ................................................................. 24

Figure 2.4 Simplified stratigraphic column, sensor depths (stars; locations with two sensors are labeled with a “2”), and shear-wave velocity structure at CUSSO................. 26

Figure 2.5 Time histories (left) and amplitude spectra (right) from the January 2, 2006, Mw 3.6 local earthquake recorded at VSAP (top row) and the Feb. 28, 2011, Mw 4.7 regional earthquake recorded at CUSSO (bottom row). ................................................................. 31

Figure 2.6 Polar-plot histogram of back-azimuths (azimuth from station to event) for events listed in Tables 2.2 and 2.3 for VSAP and CUSSO (a). Magnitude versus offset for all events recorded by VSAP and CUSSO (b). Lines corresponding to events listed in Tables 2.2 and 2.3 are tipped with large circles. ................................................................. 33

Figure 2.7 Mean spectral ratios from recordings at VSAP and CUSSO and theoretical Thomson-Haskell SH-wave transfer functions (TH_{SH}) for average bedrock incidence angles of 25° at VSAP and 15° at CUSSO................................................................. 36

Figure 2.8 Mean spectral ratios shown in Figure 2.7 (heavy black) and mean ±1 standard deviation regions (solid gray). Solid, horizontal line indicates a ratio of 1 in each plot. 37

Figure 2.9 H/V curves derived from five-hours of ambient noise, NVS_{noise} and EVS_{noise}, for the North- and East-components, respectively, recorded at VSAP and CUSSO. ......................................................................................................... 38

Figure 2.10 Vertical-component amplification, TF_{V}, and the ratio of spectral ratios TF_{T} to HVS at CUSSO and VSAP. ......................................................................................................... 41
Figure 2.11 Observed, \(TFV\), and the predicted, \(THSV\), vertical-component amplification for an SV-wave with an angle of incidence of 15° at CUSSO................................................. 42
Figure 3.1 Sites (triangles) on soil layers (S) or rock outcrops (R). H and V represent amplitude spectra of horizontal- and vertical-component recordings, respectively, at these locations. ................................................................. 46
Figure 3.2 Site responses at the fundamental frequency for each site shown in Figure 1.4. ........................................................................................................................................... 49
Figure 3.3 Seismic stations, colored by network code, and the epicenters of earthquakes used for this study. ........................................................................................................................................ 54
Figure 3.4 Shear-wave velocity structures developed for each station........................................... 55
Figure 3.5 Best-fitting shear-wave velocity structures at CUSSO determined through 10 trials of 10,000 grid-searches through the parameter space, colored by misfit. ....................... 59
Figure 3.6 Comparison of damping values predicted by five \(Q(Vs)\) relationships, using the time-weighted average velocity for each site and one \(Q(z)\) relationship. The mean value of these six estimates is also shown. ............................................................ 61
Figure 3.7 Comparison of full-resonance and fundamental-mode amplification from equation 1 for HEKY, when damping is increased \((\gamma_S + 0.009)\) and decreased \((\gamma_S - 0.009)\) from the mean damping ratio in each layer. ........................................................... 62
Figure 3.8 Layer damping ratios developed for each station from the mean value calculated from five \(Q_{ef}\)-velocity relationships and one \(Q_{ef}\)-depth relationship................................................. 62
Figure 3.9 (a) Compilation of densities and shear-wave velocities for various lithologies, and the corresponding statistical relationship of density \((\rho)\) versus S-wave velocity (red curve); modified from Boore (2016). \(Vs-\rho\) ordered pairs for the sediment layers at CUSSO and VSAP are shown for comparison with this compilation. (b) Comparison of the layer densities at CUSSO and VSAP with those predicted from the corresponding layer velocities using the Boore16 relationship. ........................................................................ 64
Figure 3.10 Comparison of amplification factors, \(A_{\rho}\), calculated from equation 3.1 using simplified, one-layer average velocity structures and full-resonance responses using measured densities and those predicted from the sediment and bedrock S-wave velocities using the Boore16 relationship. ........................................................................................................ 65
Figure 3.11 Layer densities developed for each station from the Boore (2016) density/shear-wave velocity relationship................................................................. 66
Figure 3.12 (a) Values of the terms in equation 3.1 calculated from average, one-layer-over-bedrock velocity structures. .................................................................................................................. 67
Figure 3.13 Comparison of mean S-wave H/V curves at each station for various SNR thresholds. ................................................................................................................................. 70
Figure 3.14 Example of the processing workflow for the recordings at HEKY of an M 3.4 earthquake, 98 km away. ........................................................................................................... 71
Figure 3.15 Theoretical full and simplified \((f_0A_0)\), and empirical (H/V) site responses and automatically picked peak responses. Observed spectral ratio peak frequencies from other studies are also shown................................................. 73
Figure 3.16 Comparisons of theoretical—full-resonance responses and simplifications in equations 3.1 and 3.2—and empirical (H/V) \(f_0\) and \(A_0\)................................................................. 77
Figure 3.17 H/V and theoretical site responses at USIN using the simplified expressions in equations 3.1 and 3.2, and full-resonance calculations.............................. 79
Figure 3.18 Comparisons of fundamental-mode S-wave H/V peaks with those from full-
resonance calculations (a) and those estimated using equation 3.1 (b). The best-fit lines
with 0 y-intercepts and through the points circled in red are shown and labeled by their
corresponding equations. .................................................................................................. 80
Figure 3.19 S-wave H/V and FR responses at stations that appear to have site-response
effects from deeper, unmodeled layers within the 1 to 3 Hz frequency range. ................. 86
Figure 3.20 S-wave H/V curves for candidate reference stations in or just outside of the
Illinois Basin. .................................................................................................................... 87
Figure 3.21 Standard spectral ratios at three Illinois Basin sites with respect to reference
site SLM from teleseismic S-wave recordings, and corresponding local- and regional-
earthquake S-wave H/V curves........................................................................................ 89
Figure 4.1 Precambrian (pC) faults, the Rome Trough boundary (heavy, dashed black
lines), area of possible Rogersville Shale production (gray, shaded), deep Rogersville Shale
test wells (Deep OG Well), wastewater-disposal wells (SWD well), -30 mGal and greater
Bouguer gravity anomalies (ECGH is the East Continent Gravity High), and seismicity
from 1900 through 1980 (open circles) and from 1980 to prior to the temporary network
(filled circles). Orientations of quality C and greater maximum horizontal stress ($S_{Hmax}$)
measurements from the World Stress Map (Heidbach et al., 2016) are also shown. ....... 93
Figure 4.2 Cumulative volume (left) and maximum monthly injection rate (right), in log
units, from all wastewater disposal wells with reported injection volumes in and near the
Rome Trough of eastern Kentucky. .................................................................................... 95
Figure 4.3 Well-based cross section based on drillers’ logs, with generalized stratigraphic
groups, across the Rome Trough through the study area in eastern Kentucky and
southwestern West Virginia (Fig. 4.1). ............................................................................. 95
Figure 4.4 Seismic stations used for event detection (colored by network code), the detected
earthquakes (colored by focal depth), and the ANSS ComCat during the same time period
(orange). ............................................................................................................................. 99
Figure 4.5 Vertical-component seismograms from temporary network (station codes,
labeling each trace, beginning with EK) and regional network stations from a local
magnitude 1.4 earthquake in the crust beneath the Rome Trough of eastern Kentucky
recorded at distances from 15 to 73 km from the hypocenter ................................. 102
Figure 4.6 Example vertical-component recordings (waveforms) of earthquakes beneath
(a) or outside (b) the Rome Trough bandpass filtered with a passband of 1 to 20 Hz, and
their corresponding STA/LTA characteristic functions............................................. 103
Figure 4.7 Velocity models used to determine earthquake locations. ......................... 104
Figure 4.8 P- and S-wave traveltime residuals versus hypocentral distance for all
earthquakes located in project area ............................................................................. 105
Figure 4.9 Distribution of detected event types.......................................................... 108
Figure 4.10 Cross section view of seismicity projected from 105 km perpendicular to the
either side of the section line N-S in Figure 3. .............................................................. 109
Figure 4.11 Lower-hemisphere focal mechanisms determined in this study and those
available in the literature .............................................................................................. 110
Figure 4.12 Lower hemisphere focal mechanisms, polarity observations, and direct P- and
S-wave amplitude ratios. All solutions that fit the observations are shown in thin lines. The
best-fitting solution, which minimized the misfit between the observed and predicted
amplitude ratios, are bold................................................. 111
Figure 4.13 Left: Attenuation correction function derived in this study (EK; equation 4.4) and those derived in other studies near the project area: Eastern Tennessee (B15; Bockholt et al., 2015), and the eastern U.S. (K98; Kim, 1998). Right: Distribution of observations with hypocentral distance (top) and magnitude residuals calculated as the event $M_L$ minus the station $M_L$ (using equation 4.3) versus hypocentral distance (bottom). ....................... 113
Figure 4.14 Gutenberg-Richter curves for the events located in this study when the entire temporary seismic network was operational (June 2016–October 2018) for the entire region in Figure 4.4 (all) and for those plotted on the cross section in Figure 4.10 (XS). .......... 116
Figure 4.15 Modeled nighttime (left) and daytime (right) minimum magnitude detection thresholds in the project area from Holcomb (2017). Seismicity located in this study overlay the map corresponding to the event origin times. ......................................................... 116
Figure 4.16 Focal depth versus root-mean-square traveltime misfit for the 02/20/2017 ML 0.3 earthquake, the only earthquake to have occurred within 5 km of a wastewater-injection well............................................................... 117
LIST OF ADDITIONAL FILES

[Supplemental Table 4.1 Parameters of Earthquakes Located in or near the Rome Trough, Eastern Kentucky] ........................................................................................................... [PDF 175 KB]
CHAPTER 1. INTRODUCTION

1.1 SEISMIC HAZARDS INVESTIGATED IN THE CENTRAL AND EASTERN U.S.

Damaging earthquakes in the central and eastern U.S. are much rarer than in active tectonic margins such as the western U.S. However, comparable ground motions reach greater distances in the central and eastern U.S. because the underlying older, harder crustal rocks attenuate seismic waves much less (Fig. 1.1). Strong seismic waves create ground-motion hazards that can cause damage or even result in collapses of buildings and other structures. Thus, investigating central and eastern U.S. earthquake sources and their effects has societal importance.

Figure 1.1 USGS “Did You Feel It?” felt reports from one western U.S., 2016 moment magnitude (Mw) 6.0 Napa, Calif., and two central and eastern U.S. earthquakes of comparable, albeit slightly less, magnitudes: the 2016 Pawnee, Ok. and 2011 Central Virginia Mw 5.8 events. Also shown are the felt reports from two recent smaller-magnitude earthquakes. Modified from https://www.usgs.gov/news/east-vs-west-coast-earthquakes (last accessed Nov. 19, 2019).
In the central and eastern U.S., most natural earthquakes fall into areas of concentrated seismicity, or seismic zones, distinguished based on geologic, geophysical, and seismological characteristics such as seismicity distributions, source focal mechanisms, geophysical anomalies, known faults, etc. Three central and eastern U.S. seismic zones—the Eastern Tennessee, New Madrid, and Wabash Valley Seismic Zones—lie within close proximity of one another in the region centered on Kentucky (Fig. 1.2). The damaging 1980 Sharpsburg, Ky., earthquake (Herrmann et al., 1982; Woolery et al., 2008) demonstrates that not all earthquakes of consequence are confined to established seismic zones, however.

Seismicity in the Eastern Tennessee Seismic Zone is associated with northeast-trending, en-echelon basement faults intersected by east-trending basement faults (Chapman et al., 1997). The largest instrumentally recorded earthquake in the ETSZ is the 2003 moment magnitude (Mw) 4.6 event near Fort Payne, Ala. (Dunn and Chapman, 2006), however the occurrences of earthquakes of magnitude 6 or greater have been inferred from paleoseismic investigations (Warrell et al., 2017). The New Madrid Seismic Zone is located within the Cambrian Reelfoot Rift. Three of the basement faults within the rift, reactivated under the current roughly east-west horizontal compression stress regime (Zoback, 1992), generated the largest historical earthquakes that have affected the region in Figure 1.2, i.e., the three magnitude 7 and greater earthquakes in the 1811-1812 New Madrid earthquake sequence (e.g., Hough, 2009). The Wabash Valley Seismic Zone is in a series of northwest-to-southeast oriented narrow grabens of Precambrian to early Cambrian age, located near the center of the Illinois Basin (e.g., Woolery et al., 2012). Paleoseismic investigations in this zone have evidenced multiple magnitude 6 and greater
earthquakes (e.g., Obermeier et al., 1991; Munson et al., 1995). Most recently, the 2008 Mw 5.2 Mount Carmel earthquake occurred in this zone (Hamburger et al., 2011).

In addition to hazards from natural earthquakes, since the onset of the shale gas boom in 2009, areas in the central and eastern U.S. with historically very low earthquake rates, such as central and northern Oklahoma, east central Arkansas, eastern Ohio, and western West Virginia, have experienced surges in earthquake activity, with some events producing shaking strong enough to cause structural damage to the built environment (e.g., the 2016 Mw 5.8 Pawnee, Ok. earthquake in Fig. 1.1). Figure 1.2 also shows the approximate outline of an unconventional hydrocarbon reservoir being tested in the Rome Trough of eastern Kentucky, the Rogersville Shale. Because the shale lies in the faulted Rome Trough, which is between the Eastern Tennessee Seismic Zone and the source region of the 1980 Mw 5.0 Sharpsburg earthquake, there is a potential for induced seismicity if large-scale production of this shale occurs. Thus, the region shown in Figure 1.2 is susceptible to hazards from natural and induced earthquakes.
The standard source-path-site convolution approximation for wave propagation from an earthquake to a surface receiver is used to predict ground motions assuming the earth behaves as a linear system for seismic wave propagation. In this approximation, the ground accelerations recorded at a site result from the convolution of time-domain representations of the earthquake source, the propagation path through the earth’s crust, and the effect of shallow, low-velocity layers beneath the site (Stein and Wysession, 2009):

\[ A(t) = S(t) * P(t) * G(t), \]  

where \( S(t) \) is the source term, \( P(t) \) is the crustal-path term, and \( G(t) \) is the site-term (i.e., effect of the shallow layers). Because multiplication is mathematically simpler and computationally faster than convolution, equation 1.1 is often expressed in its equivalent form in the frequency domain as
\[ A(f) = S(f) \cdot P(f) \cdot G(f). \]  

(1.2)

These terms are illustrated schematically in Figure 1.3 from a hypothetical thrust-faulting earthquake. Characterizing each term in equation 1.2 is needed to quantify ground-motion hazard at a particular site from a particular earthquake source.

Figure 1.3  Schematic representation of the terms in equation 1.2 that produce ground accelerations experienced at surface receivers. Although reflections off of each interface would produce down-going waves, only one down-going reflection—off of the free surface—is shown to illustrate resonance within the soil layers. This schematic also illustrates that sites underlain by sediment layers (“S” site) may experience higher accelerations than a site unerlain by rock (“R” site).

Chapters 2 and 3 in this dissertation present investigations of the site terms at deep borehole and surface seismic stations in the vicinity of the active seismic zones shown in Figure 1.2. Chapter 4 presents an investigation of earthquake sources in and near the Rome Trough, eastern Kentucky, where induced seismicity could become an issue.

1.2  GROUND MOTION SITE RESPONSE IN THE CENTRAL AND EASTERN U.S.

Near-surface, soft sediments alter the duration, frequency content, and amplitudes of strong ground-motions. This phenomenon, called ground motion site response, or site effect, can cause additional damage to susceptible buildings and infrastructure during
earthquakes. A classic example of such effects occurred in Mexico City during the 1985 Michoacán earthquake (M 8.1), during which ground motions were amplified by near-surface lake deposits (Seed et al., 1988). Another example is the Marina District of San Francisco, which incurred significant damage from amplified ground motion in the San Francisco Bay muds during the 1989 Loma Prieta earthquake (M 6.9) (Bonilla, 1991). Site response effects are common phenomena during strong earthquakes, and continue to be a significant subject for seismological research (see e.g., Woolery et al., 2016).

Linear site response results from two dominant factors as shown schematically in Figure 1.3 (where wave amplitudes are represented by arrow widths). First, energy conservation dictates that seismic waves amplify when they encounter a medium with a lower impedance (impedance = density times velocity) by a factor equal to the square root of the ratio of the higher to the lower impedances (Aki and Richards, 2002, equation 4.62). Second, when S-waves propagate into a medium of much lower impedance, the amplified waves can become to some degree trapped. In this situation, the waves constructively interfere at odd multiples of the fundament frequency, which is the reciprocal of twice the two-way vertical travel-time in the layer. This resonance effect is analogous to the so-called “organ pipe” resonance, which increases the amplification of seismic waves beyond that due to square root of the impedance ratio alone (Boore, 2013).

There are several empirical and theoretical methods in practice for characterizing site response. The standard spectral ratio (Borcherdt, 1970) became the first established approach, which estimates $G(f)$ as the ratio of the ground motion amplitude spectrum recorded at a soil site to that at a rock site, assuming the $S(f)$ and $P(f)$ terms in equation 1.2 are so similar that they are canceled by the spectral division, and that $G(f)$ at the rock
site is approximately unity at all frequencies. Borcherdt (1970) used recordings of nuclear explosions at regional distances (greater than 300 km) to assess amplification in the San Francisco bay area and at these large distances the effects of the source and the path are essentially the same at all stations within their study area. Later, surface-to-borehole spectral ratios were used to estimate $G(f)$, employing the same assumptions as the standard spectral ratio, where the borehole sensor installed in bedrock serves as the reference rock site under the surface sensor installed on sediments (e.g., Joyner et al., 1976; Archuleta et al., 1992; Margheriti et al., 2000).

The single-station technique, which estimates the site response as the ratio of the amplitude spectrum of ambient seismic noise recorded on the horizontal component to that on the vertical, H/V, was proposed in the late 1980s by Nakamura (Nakamura, 1989). Later, Lermo and Chavez-Garcia (1993) successfully applied the technique to earthquake recordings. This methodology assumes that vertical amplitude spectra recorded in the bedrock are similar to those from horizontal-component recordings, and that vertical-component spectra are largely unmodified by the overlying sediment column.

Site response is a complex 3-D wave-propagation phenomenon. Theoretical constructions of site response is therefore also complicated, and would most accurately be modeled by 3-D simulations. However, application of 3-D simulation is still limited because of limitations such as accuracy of the basin model, model resolution, low frequency, and nonlinearity. Therefore, the most common theoretical approaches are 1-D and include linear full resonance (e.g., Haskell, 1960), linear square-root impedance (Boore, 2013), equivalent linear (e.g., SHAKE; Schnabel et al., 1972), and nonlinear (e.g., Hashash et al., 2015).
Currently, one of the primary parameters for predicting site response in the central and eastern U.S. is the average shear-wave velocity for the top 30 m of surficial materials, Vs30 (e.g., Building Seismic Safety Council, 2009). Building on the work of Borcherdt (1994), the Vs30-based site-factors in current use were developed in California from the recordings of a single earthquake, the 1989 Loma Prieta earthquake (Dobry et al., 2000). Average amplification factors were derived from the standard spectral ratios of amplitude spectra recorded on various site conditions to those recorded on nearby reference rock conditions. These factors were grouped by the levels of input motion to account for potential nonlinear effects and correlated with Vs30. The so called site-factors were codified in 1994 and 1997 NEHRP Provisions and the 1997 Uniform Building Code (UBC) and refined in Dobry et al. (2000).

The shallow geologic conditions that affect site response in the western U.S. differ from those in the central and eastern U.S. And recent studies in the central and eastern United States (e.g., Hashash et al., 2008; Woolery et al., 2009; Hassani and Atkinson, 2016, 2017) and elsewhere (e.g., Castellaro et al., 2008; Cadet et al., 2010; Lee and Trifunac, 2010; Régnier et al., 2014) indicate that Vs30-based site factors may not reliably estimate site response, especially in regions like the central and eastern U.S., where resonance effects dominate the response of the shallow layers to incident seismic waves (Fig. 1.3).

For example, Figure 1.4 shows shear-wave velocity structures derived from surface reflection and refraction surveys and downhole tests at four sites along the Ohio and Mississippi Rivers (Li et al., 2013) near the New Madrid and Wabash Valley Seismic Zones. Vs30 was calculated for the sites and used to assign NEHRP site classifications. As shown in Figure 1.4b, all four sites are classified as site-class D, which implies that each
site should undergo the same level of amplification at short periods and the same level of amplification at long periods, even though the velocity structures at each site are significantly different. Spectral amplification functions derived from the 1-D equivalent-linear site-response model STRATA (Kottke and Rathje, 2008) shown in Figure 1.5 demonstrate that the site responses (i.e., the peak frequencies and ratios) are quite different for these four sites: base mode frequencies in particular have nearly an order of magnitude in variability, ranging from 0.3 to 2.3 Hz, and the peak amplifications range from 4.2 to 6.1, which exceed the amplification factors predicted by Vs30 (Dobry et al., 2000) by a factor of 2 or greater at short periods and at long periods. Figure 1.5 shows that site responses at these sites are controlled by S-wave resonances, which site-factors based on the Vs30 proxy did not capture.

Figure 1.4 (a) The locations of four sites along the Ohio and Mississippi Rivers. Contours show depth to bedrock in the Mississippi Embayment. (b) Shear-wave velocity profiles for each site. Time-weighted average shear-wave velocities from the surface to 30 m (Vs30; with the corresponding NEHRP site class) and to bedrock ($V_s^b$) are shown.

Figure 1.4b also shows that there are significant velocity contrasts between bedrock with shear-wave velocities greater than 1,100 m/s and the sediment column with average (time-weighted) velocities less than 546 m/s. There are strong velocity contrasts between
the sediments and bedrock throughout the New Madrid and Wabash Valley Seismic Zones (e.g., Street et al., 2001, 2004; Woolery et al., 2009, 2012), which have been attributed to large ground-motion amplifications observed in the central and eastern U.S. and elsewhere. For example, Singh et al. (1988) estimated amplifications up to 75 at the fundamental frequency, $f_0$, in the areas of Mexico City built on lake-bed sediments. Along the East Coast of the United States, large weak-motion amplifications at sites underlain by Atlantic Coastal Plain sediments of factors greater than 10 were observed in the Washington, D.C. (Pratt et al., 2017), and Boston (Baise et al., 2016) areas. Further, Pratt (2018) showed that site resonances from unlithified strata are extensive along the Atlantic Coastal Plain and in the Mississippi Embayment. Banab et al. (2012) observed weak-motion amplifications up to $\sim50$ at sites on soft soils in the Ottawa area and Woolery et al. (2008) calculated amplifications up to 8 at a site on Ohio River sediment in Maysville, Ky.

In each of those cases, the strong impedance contrasts between the sediments and underlying bedrock were considered to be major causes of the large amplifications. These studies indicate the importance of accounting for potential SH-wave resonance within low-velocity sediments in quantifying site-response. As demonstrated here, however, the Vs30 proxy does not account for the frequency-dependent site response due to resonance. Furthermore, Boore (2013) demonstrated that the square-root-impedance method underestimates the amplification at $f_0$, $A_0$, in the case of a strong sediment-bedrock impedance contrast. Thus, as other studies have recognized, there is a need for better estimation of site response for regional seismic-hazard assessment in the central and eastern U.S., which includes resonance effects. This is particularly important in the vicinity of the sources of high seismic hazard, including the New Madrid and Wabash Valley
Seismic Zones, and in thick sediment layers, such as the northern Mississippi Embayment (Fig. 1.4a). Also, several population centers in the central and eastern U.S. are near active seismic zones, such as Memphis, Tenn., Charleston, S.C., and Washington, D.C., with thick underlying sediment deposits.

Thus, Chapters 2 and 3 in this dissertation focus on assessing reliable empirical and theoretical quantifications of $G(f)$ in central and eastern U.S., including at sites underlain by thick sediments. In Chapter 2, the recordings from the two deep ($\geq 100$ m) vertical seismic arrays in the northern Mississippi Embayment, VSAP and CUSSO, are used to directly evaluate the site response in this deep-sediment setting using simultaneously recorded earthquakes at both vertical arrays consisting of downhole (i.e., bedrock) and surface sensors. Comparisons are made between the full-resonance responses and the empirical transfer functions. Chapter 2 also includes a comparison between the theoretical and empirical transfer functions and single-station H/V site response estimates.

Fundamental-mode (i.e., base mode) site resonance frequency, $f_0$, has been proposed to be of primary importance in accounting for site response (e.g., Hassani and Atkinson, 2016). As shown by Hassani and Atkinson (2017) and Cadet et al. (2010), however, site response variabilities can be further reduced when both $f_0$ and Vs30 parameterize site response. Because Vs30 may not reliably capture site amplifications in the central and eastern U.S., $A_0$, the spectral ratio at $f_0$, was evaluated in Chapter 3 as an alternative additional parameter. As shown in Figure 1.5, spectral amplification functions at four sites have several peak ratios—$A_0, A_1, A_2,...$—at corresponding frequencies of $f_0, f_1, f_2, ...$. Figure 1.5 shows that $A_0$ is larger than other peak ratios at three sites (i.e.,
Louisville, Owensboro, and Paducah), but it is slightly less than the second and third peak ratios at station CUSSO.

Figure 1.5 Equivalent linear, 1-D spectral amplification functions for the four sites along the Ohio and Mississippi Rivers in Figure 1.4 and a 0.1 g PGA input time history. Selected peak frequencies and magnitudes are labeled as \((f_n, A_n)\) ordered pairs; higher modes are only labeled for CUSSO. Also shown are mean long- \((F_v)\) and short-period \((F_a)\) site-class D amplification factors (from Dobry et al., 2000) for weak input motion (peak accelerations of 0.1 g and less).

As discussed in Chapter 3, the \(A_0\) and \(f_0\) parameters can be quantified using simplifications of analytical, full-resonance expressions that account for 1-D wave propagation in a stack of layers, include the effects of impedance contrasts and S-wave resonance, and are dependent on shallow earth models. The evaluation consisted of comparing these simplifications to the full-resonance responses at 11 seismic stations in the central and eastern U.S. In addition, the applicability of S-wave horizontal-to-vertical spectral ratios to provide empirical approximations of \(A_0\) and \(f_0\) was evaluated.
1.3 INDUCED SEISMICITY POTENTIAL AND SEISMIC SOURCE ZONES IN EASTERN KENTUCKY

Earthquakes can result from natural causes, including the sudden release of tectonic strain through earthquake cycles and from volcanic activity. They can also be caused by manmade activities such as the injection of fluids into deep boreholes. Most seismic events triggered or induced by human activity produce very low-level shaking (Ellsworth, 2013); however, some instances of wastewater injection have reactivated faults and caused felt earthquakes, some of which were large enough to cause structural damage in local communities (Taylor et al., 2017).

Since approximately 2009, the rate of felt earthquakes in the central United States has increased dramatically (Fig. 1.6). This increased rate correlates strongly in space and time with the increase in production of oil and gas, and resultant subsurface disposal of produced water (Weingarten et al., 2015; Langenbruch and Zoback, 2016). The principal cause of these events has been assigned to the injection of wastewater into subsurface formations (Horton, 2012; Keranen et al., 2013; Hornback et al., 2015); the largest earthquake likely induced by wastewater injection was the 2016 moment magnitude (Mw) 5.8 Pawnee, Okla., earthquake (Yeck et al., 2017). Hydraulic fracture stimulation of unconventional reservoirs, or fracking, has also induced felt earthquakes (Holland, 2013; Skoumal et al., 2015; Bao and Eaton, 2016; Brudzinski and Kozłowska, 2019); the largest event likely induced in North America by fracking was the 2015 Mw 3.9 Fox Creek earthquake in Alberta, Canada. Most cases of induced, felt earthquakes were the result of fluid injection into formations that are in hydraulic communication with the crystalline basement, which can lead to the rupture of preexisting, critically stressed basement faults (Zoback et al., 2002).
In the Rome Trough of eastern Kentucky, a deep formation with total organic carbon content sufficient for hydrocarbon generation, the Rogersville Shale, is currently being tested in exploration wells (Harris et al., 2015). Because of its low permeability, this shale is an unconventional reservoir, requiring high-volume and high-pressure fracking. And, because this deep formation is in close proximity to the faulted, crystalline basement in the Rome Trough, and bounded by seismically active regions, there is a potential for fracking-induced earthquakes by producing oil and gas from this shale.

In addition, produced wastewater has been injected in the eastern Kentucky Rome Tough since 1997 (Sparks and Curl, 2014; Carpenter et al., 2019). The injection formations in the Rome Trough are relatively shallow compared to the depth of the crystalline basement, and no injection-related events have been recorded by the regional seismic...
monitoring networks (Kentucky Geological Survey, 2014). If large-scale development of the Rogersville Shale occurs, and large volumes of brine are produced and injected, however, the risk of inducing earthquakes from wastewater disposal could increase.

The central and eastern U.S. hosts several active zones of natural seismicity (e.g., Thomas and Powell, 2017; Fig. 1.2), which are capable of producing damaging earthquakes. One of which, the Eastern Tennessee Seismic Zone, is adjacent to the eastern Kentucky Rome Trough to its south. Because the possibility of inducing earthquakes in the Rome Trough may increase if the Rogersville Shale becomes a productive hydrocarbon play, characterizing background microseismicity in the area is important. In regions of concurrent subsurface fluid injection and seismic activity, unequivocally discriminating between natural and induced earthquakes requires the analysis of multiple data sets, including at minimum fluid-injection volume histories of active wells and an earthquake catalog. When the timing of earthquakes that occur near wastewater-injection wells and fracture stimulations is strongly correlated with the injection history, the probability of a causal relationship between the two increases. Also, cataloged seismicity-rate changes permit estimating the probability that an increase in seismic activity is natural (Rubinstein et al., 2014). Natural and induced microearthquakes (earthquakes of magnitude less than 2.5) occur exponentially more frequently than larger, felt earthquakes, and are therefore a more sensitive indicator of variations in seismicity rate. Therefore, the background rates of microearthquakes are of particular importance for determining the likelihood that earthquake activity is induced. Permanent seismic stations monitoring the region around the Rome Trough are sparse, however, and if induced earthquakes were to occur from subsurface fluid injection, the existing regional stations would not permit the determination
of event locations with sufficient accuracy and precision to assess the potential association of earthquakes with subsurface injection activities.

Therefore, an evaluation of microseismicity in and around the Rome Trough, eastern Kentucky where seismic monitoring has been sparse, was conducted using a temporary seismic network to characterize natural seismic sources prior to large-scale development of the Rogersville Shale. The earthquake distribution is interpreted in its seismotectonic setting. The results of this investigation are presented in Chapter 4.
CHAPTER 2. GROUND MOTION SITE RESPONSE FROM SHEAR-WAVE RECORDINGS AT DEEP BOREHOLES IN THE NEW MADRID SEISMIC ZONE

2.1 INTRODUCTION

Site response is influenced by many factors, including lateral and vertical velocity gradients in the sediment and bedrock, impedance contrasts within the sediment overburden and at the sediment-bedrock interface, sediment thickness, sediment-bedrock interface geometry, incoming ground-motion amplitude (i.e., linear vs. nonlinear behavior), and surface topography. There are several established methods in practice for characterizing site effect, ranging between empirical and theoretical, but there are considerable attendant uncertainties (Steidl et al., 1996), particularly in regions with deep sediment deposits (>100 m) such as in the northern Mississippi Embayment of the central United States. A direct way to study site effect, of particular importance for sites overlying thick sediment layers, is to simultaneously record earthquakes with a vertical array of downhole (i.e., bedrock) and surface sensors (Archuleta et al., 1992; Margheriti et al., 2000). The recordings from the two deep vertical seismic arrays in the northern Mississippi Embayment, VSAP and CUSSO, permit direct evaluation of the site response in this deep-sediment setting.

This chapter presents the spectral analyses performed on weak-motion S-wave recordings from the only deep (≥ 100 m) vertical seismic arrays that penetrate the entire sediment column in the central and eastern U.S., VSAP and CUSSO, to determine empirical transfer functions. The mean empirical transfer functions were compared with theoretical transfer functions derived from the Thomson-Haskell plane-wave reflectivity model for SH-waves, focusing on frequencies of engineering interest. Also, the mean
single-station—i.e., horizontal-to-vertical—spectral ratios are compared with the empirical transfer functions at these borehole sites.

2.2 EMPIRICAL SH-WAVE TRANSFER FUNCTION AND S-WAVE H/V

The Fourier spectrum of ground acceleration for SH-waves at a given site can be modeled as the convolution of source, \( S(f) \), path, \( P(f) \), site response, \( G(f) \), and instrument response, \( I(f) \), terms as:

\[
A(f) = S(f) \cdot P(f) \cdot G(f) \cdot I(f)
\]  

(2.1)

In the following, the quantities in equation 2.1 were used to derive expressions for the empirical site transfer function, \( G \), from the SH-wave amplitude spectra, \( A \), of recorded ground motions. The subscripts of \( S \), \( R \), and \( B \) were added to equation 2.1 for horizontal (H) and vertical (V) amplitude spectra recorded at soil, rock outcrop, and borehole bedrock sites, respectively, as shown in Figure 2.1.

Figure 2.1 Locations of sensors (triangles) at the surface at soil (S) and rock-outcrop (R) sites, and beneath the soil in bedrock (B). H and V represent amplitude spectra of horizontal- and vertical-component recordings, respectively, at these locations.
2.2.1 SURFACE-TO-BEDROCK SPECTRAL RATIOS

If soil and rock-outcrop sites are proximal (i.e., differences in the source and path terms are negligible for both soil and rock sites—\( S(f) \cong S_R(f) \) and \( P_S(f) \cong P_R(f) \)), and if the site response at the rock site, \( G_R \), is assumed to be flat and to equal unity, then after removal of the instrument responses, the spectral ratio between soil and rock sites is

\[
G_S = \frac{A_S}{A_R} \tag{2.2}
\]

When horizontal, transverse motions of S-waves are being considered, equation (2.2) is the transfer function for SH-waves between the soil and rock sites, which is traditionally used to quantify empirical site response in earthquake engineering (Borcherdt, 1970).

Depending on the source mechanism and the relative positions of the soil and rock sites to the source, however, the requirements of equation (2.2) that differences in the source and path terms for rock and soil sites are negligible, might not be applicable (as is the case for the soil and rock sites in Fig. 1.3). An additional concern is that the rock site has its own site response, which is not accounted for in the above formulation (see e.g., Steidl et al., 1996). Another approach, which may abate these concerns, directly compares the acceleration spectra at the surface with those at the bedrock in a borehole (Fig. 2.1) (e.g., Joyner et al., 1976; Steidl et al., 1996). The ratio of the surface transverse-component amplitude spectrum to that in the bedrock for this configuration is

\[
TF_T = \frac{H_S}{H_B} \tag{2.3}
\]
which is the empirical SH-wave transfer function for a given angle of incidence from the bedrock.

Assuming that a plane wave model for SH-waves in an elastic, 1-D layered structure is appropriate for the seismic waves recorded by VSAP and CUSSO, equation (2.3) can be expressed analytically using Thomson-Haskell matrices (Haskell, 1953; Haskell, 1960), herein $TH_{SH}$

$$TH_{SH} = \frac{2 \mu_n r_{\beta_n}}{\mu_n r_{\beta_n} A_{11} + A_{21}}$$

where $\mu_n$ is the shear modulus of the bedrock, $r_{\beta_n}$ is the ratio of the vertical slowness to the ray parameter in the bedrock underlying the soil, and $A_{11}$ and $A_{21}$ are elements of the product layer matrix, $A$, which is formed by multiplication of the layer matrices for each of the $m$ soil layers i.e., $A = a_m a_{m-1} ... a_2 a_1$, where

$$a_m = \begin{bmatrix} \cos(kh_m r_{\beta_m}) & i (\mu_m r_{\beta_m})^{-1} \sin(kh_m r_{\beta_m}) \\ i \mu_m r_{\beta_m} \sin(kh_m r_{\beta_m}) & \cos(kh_m r_{\beta_m}) \end{bmatrix}$$

and $k$ is the horizontal wavenumber, $h_m$ is the layer thickness, and $r_{\beta_m}$ is the ratio of the vertical shear-wave slowness in the $m^{th}$ layer to the SH-wave ray parameter, which is equivalent to the cotangent of the angle of incidence at the base of the $m^{th}$ layer.

### 2.2.2 HORIZONTAL-TO-VERTICAL SPECTRAL RATIOS

The horizontal-to-vertical spectral ratio ($H/V$) was originally used to estimate site response using recordings of microtremors (Nakamura, 1989) and later from earthquake recordings (Lermo and Chavez-Garcia, 1993). $H/V$ is defined as the ratio of the horizontal, $H_S$ and the vertical amplitude spectra of ground motion, $V_S$, at the free-surface,
As previously stated, $H_S$ represents the amplitude spectra of the transverse component of motion.

Equation (2.4) can be expanded as,

$$HV_S = \frac{H_S}{V_S}$$

(2.4)

and equation (2.5) can be rewritten as

$$HV_S = \frac{H_S \cdot H_B \cdot V_B}{V_B \cdot V_S}$$

(2.5)

where $HV_B$ is the H/V in the bedrock and $TF_V$ is the transfer function of vertical ground motions for a particular ray parameter. Horizontally polarized SH-waves do not excite vertical motions. Therefore, the vertical-component time series contains SV and P arrivals within the S-wave window, rather than SH-waves.

Equation (2.6) shows that the surface H/V, $HV_S$, is equal to the SH-wave transfer function in equation (2.3), expressed theoretically by $TH_{SH}$, times the borehole H/V, $HV_B$, divided by the vertical-motion transfer function, $TF_V$. Nakamura (1989) observed that $HV_B$ is on average approximately unity for ambient noise. $HV_B$, determined from recordings windowed around S-waves, however, depends on the earthquake focal mechanism; therefore, because $TF_T$ and $TF_V$ are independent of the source, $HV_S$ will also depend on the source mechanism. The approach in this study was to calculate the mean $HV_S$ from multiple events to determine if on average $HV_S$ approximates $TF_T$. 

\[HV_S = TF_T \cdot HV_B \cdot \frac{1}{TF_V}\]

(2.6)
2.3 VERTICAL ARRAYS AND DATASETS

The settings of the vertical arrays used in this study differ in terms of un lithified sediment thicknesses, proximities to the edge of the embayment (Fig. 2.2) and age of the near-surface deposits. The geology, instrumentation, and recordings from VSAP and CUSSO are described briefly below.

Figure 2.2 Vertical seismic arrays CUSSO (blue star) and VSAP (red star), in the northern Mississippi Embayment and the earthquakes they recorded. Embayment depth-to-bedrock contours are labeled by depth below the surface in meters. Contours (in meters) are sediment thickness from Dart (1992) and Dart and Swolfs (1998), modified from Langston et al. (2009).
2.3.1 VSAP

VSAP was installed near Paducah, Ky., in the early 1990s (Street et al., 1997). The site is approximately 15 km from the edge of the northern Mississippi Embayment on a 100-m-thick sequence of unlithified to poorly lithified silts, sands, clays, and gravels of Late Cretaceous to Pleistocene age overlying Mississippian limestone bedrock (Harris, 1992). Four soil layers and the bedrock were identified by two orthogonal SH-wave refraction profiles and incorporated into the velocity model of VSAP (McIntyre, 2008; Table 2.1; Fig. 2.3). The fundamental-frequency, $f_0$, of S-wave resonance in the overburden at this site for vertical-incidence S-waves (Haskell, 1960) is 1.06 Hz using

$$f_n = \frac{V_S}{4h} (2n + 1), \quad n = 0,1,2 \ldots \quad (2.7)$$

where $V_S$ is the weighted-average S-wave velocity (i.e., layer velocities weighted by layer thickness), $h$ is the thickness of the sediment overburden, and $n$ is the resonance mode.
Figure 2.3 Simplified stratigraphic column, sensor depths (stars), and shear-wave velocity structure at VSAP. The average velocity shown (dashed line) was calculated for the entire sediment column, weighted by layer thickness.

VSAP’s recordings analyzed for this study were acquired from 1 May, 2005 through April, 2008. During this time, a 1/4g (i.e., ¼ times the acceleration due to gravity) strong-motion accelerometer was installed in the bedrock and, at various times, either a 1g or a 2g strong-motion accelerometer operated at free surface. The 1/4g and 2g sensors have flat-responses to ground acceleration from DC to nominally 50 Hz; the 1g sensor’s flat response extends from DC to a nominal 200 Hz. Data from the borehole and surface sensors were acquired at 200 samples-per-second.
2.3.2 CUSSO

Phased installation of the three-borehole, 21-component strong-motion array CUSSO—the Central United States Seismic Observatory—began in 2005 and was completed in 2008. The deepest borehole penetrates the entire soil-sediment overburden (585 m) and is terminated 9 m into Ordovician limestone bedrock. The stratigraphy, velocity model (Fig. 2.4; Table 2.1), and CUSSO’s instrumentation and recordings are described in Woolery et al. (2016). For this study, the recordings from the two medium-period seismometers (0.067 to 50 Hz flat-response passbands), installed at the surface and at 587 m depth, each acquired at 200 samples-per-second, were used. From the S-wave velocity structure at CUSSO, $f_0$ is 0.23 Hz using equation 2.7.
Figure 2.4 Simplified stratigraphic column, sensor depths (stars; locations with two sensors are labeled with a “2”), and shear-wave velocity structure at CUSSO. The average velocity shown (dashed line) was calculated for the entire sediment column, weighted by layer thickness.
The bedrock S-wave velocity at CUSSO increases rapidly with depth from 1,452 m/s to 1,810 m/s in one meter (McIntyre, 2008), and it is uncertain if the velocity at the depth of the borehole sensor falls within this range. Also, it is unknown if the steep velocity gradient continues below this deepest measurement to produce the site’s high-impedance boundary. Due to these unknowns, and the observations of large SH-wave amplifications, the maximum bedrock S-wave velocity observed at the nearby New Madrid test well 1-X (NMTW-1-X), 26 km southwest of CUSSO, as CUSSO’s bedrock velocity was adopted. The depth to bedrock at the NMTW-1-X site is 616 m and, similar to CUSSO, Sexton et al. (1986) reported S-wave velocities that increase rapidly with depth in the bedrock: 1,200 m/s was observed at the top of bedrock, and the maximum of 2,132 m/s occurred four meters deeper.

Although the actual bedrock S-wave velocity at CUSSO is uncertain, the observed bedrock velocity at the NMTW-1-X well produces theoretical amplifications that are much more consistent with the observations than those from a slower velocity. Therefore, this velocity was used as the bedrock velocity in the site’s velocity model. However, the theoretical SH-wave transfer function calculated for CUSSO is provisional until this velocity is validated with an independent method.
Table 2.1 Soil-profile parameters for site-response modeling at VSAP and CUSSO. α, P-wave velocity at CUSSO; β, S-wave velocity.

<table>
<thead>
<tr>
<th>Site</th>
<th>Thickness (m)</th>
<th>α (m/s)</th>
<th>β (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>VSAP</td>
<td>4</td>
<td>150</td>
<td></td>
</tr>
<tr>
<td></td>
<td>8</td>
<td>248</td>
<td></td>
</tr>
<tr>
<td></td>
<td>33</td>
<td>400</td>
<td></td>
</tr>
<tr>
<td></td>
<td>55</td>
<td>485</td>
<td></td>
</tr>
<tr>
<td></td>
<td>163</td>
<td>1630</td>
<td></td>
</tr>
<tr>
<td>CUSSO</td>
<td>15</td>
<td>1000</td>
<td>160</td>
</tr>
<tr>
<td></td>
<td>25</td>
<td>1550</td>
<td>280</td>
</tr>
<tr>
<td></td>
<td>45</td>
<td>1600</td>
<td>390</td>
</tr>
<tr>
<td></td>
<td>50</td>
<td>1650</td>
<td>515</td>
</tr>
<tr>
<td></td>
<td>155</td>
<td>1850</td>
<td>600</td>
</tr>
<tr>
<td></td>
<td>205</td>
<td>1900</td>
<td>650</td>
</tr>
<tr>
<td></td>
<td>90</td>
<td>2300</td>
<td>875</td>
</tr>
<tr>
<td></td>
<td>3669</td>
<td>2132</td>
<td></td>
</tr>
</tbody>
</table>

2.4 METHODS

2.4.1 DATA SELECTION AND PROCESSING

Both CUSSO and VSAP recorded few events each (Fig. 2.2) due to their brief operational timespans. In addition, no strong ground motions (i.e., greater than 50 cm/s^2) were recorded by these stations. Therefore, selecting high-quality recordings of the weak-motions is imperative to avoid contaminating the spectral-ratios and their averages with noise. The quality assessments included inspection of waveforms, the corresponding amplitude spectra, and signal-to-noise calculations in the frequency domain. Records that contained instrument glitches or spikes in the S-wave window were excluded from the analyses and only recordings with pre-P-wave noise and signal-to-noise ratios exceeding 1.5 for each component and for frequencies from the site \( f_0 \) to 25 Hz were analyzed.
Parameters of the local and regional earthquakes used for this study are in Tables 2.2 and 2.3.

Table 2.2 Parameters for the earthquakes recorded by VSAP used in this study. Lat, latitude; Lon, longitude; Dep, depth; Mag, event magnitude and type: w, moment magnitude; d, duration magnitude; l, mb-Lg; Dist, epicenter-VSAP offset; BAZ, VSAP epicenter back azimuth; iB, S-wave incidence angle at the bedrock sensor; iS = S-wave incidence angle at the surface.

<table>
<thead>
<tr>
<th>Date</th>
<th>Time</th>
<th>Lat (°N)</th>
<th>Lon (°E)</th>
<th>Dep (km)</th>
<th>Mag</th>
<th>Dist (km)</th>
<th>BAZ (°)</th>
<th>iB (°)</th>
<th>iS (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>5/1/2005</td>
<td>12:37</td>
<td>35.83</td>
<td>-90.15</td>
<td>10.0</td>
<td>4.2w</td>
<td>187</td>
<td>220</td>
<td>26</td>
<td>1</td>
</tr>
<tr>
<td>6/2/2005</td>
<td>11:35</td>
<td>36.15</td>
<td>-89.47</td>
<td>15.0</td>
<td>4.0w</td>
<td>124</td>
<td>209</td>
<td>14</td>
<td>2</td>
</tr>
<tr>
<td>6/20/2005</td>
<td>02:00</td>
<td>36.94</td>
<td>-88.99</td>
<td>7.7</td>
<td>2.7d</td>
<td>27</td>
<td>216</td>
<td>15</td>
<td>2</td>
</tr>
<tr>
<td>6/20/2005</td>
<td>12:21</td>
<td>36.92</td>
<td>-89.00</td>
<td>18.7</td>
<td>3.6w</td>
<td>28</td>
<td>216</td>
<td>12</td>
<td>2</td>
</tr>
<tr>
<td>6/27/2005</td>
<td>15:46</td>
<td>37.63</td>
<td>-89.42</td>
<td>9.6</td>
<td>3.0l</td>
<td>77</td>
<td>316</td>
<td>15</td>
<td>2</td>
</tr>
<tr>
<td>1/2/2006</td>
<td>21:48</td>
<td>37.84</td>
<td>-88.42</td>
<td>7.3</td>
<td>3.6l</td>
<td>86</td>
<td>204</td>
<td>15</td>
<td>2</td>
</tr>
<tr>
<td>4/18/2008</td>
<td>9:36</td>
<td>38.45</td>
<td>-87.89</td>
<td>14.2</td>
<td>5.2w</td>
<td>168</td>
<td>29</td>
<td>14</td>
<td>2</td>
</tr>
<tr>
<td>4/18/2008</td>
<td>15:14</td>
<td>38.46</td>
<td>-87.87</td>
<td>15.5</td>
<td>4.7w</td>
<td>169</td>
<td>29</td>
<td>14</td>
<td>2</td>
</tr>
<tr>
<td>4/21/2008</td>
<td>5:38</td>
<td>38.45</td>
<td>-87.88</td>
<td>18.3</td>
<td>4.0w</td>
<td>168</td>
<td>29</td>
<td>14</td>
<td>2</td>
</tr>
<tr>
<td>3/2/2010</td>
<td>19:37</td>
<td>36.79</td>
<td>-89.36</td>
<td>8.2</td>
<td>3.7l</td>
<td>61</td>
<td>232</td>
<td>14</td>
<td>2</td>
</tr>
</tbody>
</table>
Table 2.3 Parameters for the earthquakes recorded by CUSSO used in this study. Lat, latitude; Lon, longitude; Dep, depth; Mag, event magnitude and type: w, moment magnitude; d, duration magnitude; I, mb-Lg; Dist, epicenter-CUSSO offset; BAZ, CUSSO-epicenter back azimuth; iB, S-wave incidence angle at the bedrock sensor; iS = S-wave incidence angle at the surface.

<table>
<thead>
<tr>
<th>Date</th>
<th>Time</th>
<th>Lat (°N)</th>
<th>Lon (°E)</th>
<th>Dep (km)</th>
<th>Mag</th>
<th>Dist (km)</th>
<th>BAZ (°)</th>
<th>iB (°)</th>
<th>iS (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>5/30/2010</td>
<td>2:24</td>
<td>36.55</td>
<td>-89.72</td>
<td>9.2</td>
<td>3.1d</td>
<td>34</td>
<td>269</td>
<td>19</td>
<td>1</td>
</tr>
<tr>
<td>6/9/2010</td>
<td>18:40</td>
<td>36.25</td>
<td>-89.40</td>
<td>5.2</td>
<td>2.5d</td>
<td>34</td>
<td>191</td>
<td>19</td>
<td>1</td>
</tr>
<tr>
<td>2/17/2011</td>
<td>10:49</td>
<td>35.28</td>
<td>-92.36</td>
<td>6.5</td>
<td>3.8w</td>
<td>308</td>
<td>243</td>
<td>15</td>
<td>1</td>
</tr>
<tr>
<td>2/18/2011</td>
<td>4:59</td>
<td>35.26</td>
<td>-92.37</td>
<td>5.0</td>
<td>3.9w</td>
<td>310</td>
<td>243</td>
<td>14</td>
<td>1</td>
</tr>
<tr>
<td>2/18/2011</td>
<td>8:13</td>
<td>35.27</td>
<td>-92.38</td>
<td>6.2</td>
<td>4.1w</td>
<td>310</td>
<td>244</td>
<td>15</td>
<td>1</td>
</tr>
<tr>
<td>2/28/2011</td>
<td>5:00</td>
<td>35.27</td>
<td>-92.35</td>
<td>3.1</td>
<td>4.7w</td>
<td>308</td>
<td>243</td>
<td>15</td>
<td>1</td>
</tr>
<tr>
<td>3/4/2011</td>
<td>8:45</td>
<td>35.28</td>
<td>-92.34</td>
<td>3.0</td>
<td>2.8d</td>
<td>306</td>
<td>243</td>
<td>15</td>
<td>1</td>
</tr>
<tr>
<td>4/8/2011</td>
<td>3:27</td>
<td>35.26</td>
<td>-92.39</td>
<td>5.5</td>
<td>3.2l</td>
<td>311</td>
<td>243</td>
<td>15</td>
<td>1</td>
</tr>
<tr>
<td>4/8/2011</td>
<td>14:56</td>
<td>35.26</td>
<td>-92.36</td>
<td>6.2</td>
<td>3.9w</td>
<td>309</td>
<td>243</td>
<td>14</td>
<td>1</td>
</tr>
</tbody>
</table>

Each triggered waveform file was converted to SAC format, the mean and trend were removed, and the instruments responses were deconvolved to yield ground acceleration time histories (seismometer recordings at CUSSO were converted from ground velocity to acceleration), using the processing parameters shown in Table 2.4. And the surface and borehole horizontal-component recordings were rotated to radial and transverse orientations. Figure 2.5 shows example accelerograms, and their corresponding amplitude spectra.

Table 2.4 Data processing parameters for recordings at VSAP and CUSSO. t0, window start time prior to SH arrival; twin, window length around SH wave; Taper, percentage of window length tapered with Hanning window; f_\text{lo} / f_\text{hi}, low and high corner frequencies for two-pole, zero-phase Butterworth bandpass filter; f\text{smooth}, -length of running-average smoothing filter used to smooth amplitude spectra.

<table>
<thead>
<tr>
<th>Station</th>
<th>t0 (s)</th>
<th>twin (s)</th>
<th>Taper (%)</th>
<th>f_{\text{lo}} / f_{\text{hi}} (Hz)</th>
<th>f_{\text{smooth}} (Hz)</th>
</tr>
</thead>
<tbody>
<tr>
<td>VSAP</td>
<td>0.25</td>
<td>5</td>
<td>5</td>
<td>0.07 / 40</td>
<td>0.5</td>
</tr>
<tr>
<td>CUSSO</td>
<td>1</td>
<td>20</td>
<td>5</td>
<td>0.07 / 40</td>
<td>0.1</td>
</tr>
</tbody>
</table>
Figure 2.5 Time histories (left) and amplitude spectra (right) from the January 2, 2006, Mw 3.6 local earthquake recorded at VSAP (top row) and the Feb. 28, 2011, Mw 4.7 regional earthquake recorded at CUSSO (bottom row). Surface (upper three traces in each row) and bedrock (lower three traces in each row) traces are shown and labeled by channel name. Amplitude spectra are calculated from the windowed portion (dashed lines) of each waveform; noise spectra were determined from the waveforms prior to the first P-wave arrival.

Because both sites are over thick layers of unconsolidated sediment, relatively long S-wave windows of five times the sites’ fundamental periods (i.e., $f_0^{-1}$) were used to resolve the amplification at each site’s fundamental frequency. Shorter windows do not provide adequate resolution at low-frequencies due to the weak-motions recorded by these arrays. For local events (offset < ~100 km), windowed, transverse-component recordings
will principally contain direct SH-waves with some scattered SH-wave and Lg-wave phases. At larger offsets, the transverse-component S-wave windows, which avoid Rg waves, can contain arrivals from Sn, direct-SH and Lg-waves. Although this diversity of phases are potentially included in spectral ratios from individual events, most peaks in the mean $TF_T$ curves from local events and from regional events occur at the frequencies predicted by $TH_{SH}$, calculated for average incidence angles (Tables 2.2 and 2.3). This consistency indicates that the arrivals included in the S-wave windows produce resonances in the soil columns consistent with direct SH-wave resonance. For vertical-component recordings, energy in the S-wave windows comes primarily from incident SV-waves, which are transmitted as P- and SV-waves, as demonstrated for CUSSO in the Discussion section.

Figure 2.6 summarizes the dataset in terms of the spatial distribution of the events with respect to the stations. The largest surface ground motion at VSAP, 23.0 cm/s$^2$, was produced by a moment magnitude ($M_w$) 3.6 earthquake 29 km southwest of VSAP. Three of the earthquakes recorded at VSAP were associated with the 2008 $M_w$ 5.2 Mount Carmel, Ill., earthquake sequence (Hamburger et al., 2011), including the main shock. All but two of the earthquakes recorded at CUSSO listed in Table 2.3 occurred in Arkansas and were associated with the Guy-Greenbriar sequence between 2010 and 2011 (Horton, 2012). The largest ground motion recorded at the surface at CUSSO was 1.0 cm/s$^2$, from the duration magnitude ($M_d$) 3.1 2010/05/30 local earthquake.
At both stations, the orientations of borehole horizontal components were not measured at the time of installation, although they were estimated for CUSSO for a brief period of its operation (Woolery et al., 2016). The azimuths of the bedrock sensors’ horizontal components at both VSAP and CUSSO for the time periods of the recordings used for this study were estimated using local earthquake P-wave arrivals (see e.g., Aster and Shearer, 1991). For each event, the azimuth of the sensor’s Y-component was found using the known station-event back azimuth, BAZ, and the angular displacement, $\theta$, required to minimize the $L^2$-norm of one cycle of the P-wave on the rotated X-component recording. Simultaneously requiring a positive correlation between the rotated Y-component recording and the vertical component guarantees that the true transverse direction was achieved for the X-component, and not the transverse-orientation plus 180°. The azimuth of the Y-component is obtained by subtracting $\theta$ from the sum of BAZ and 180° (and subtracting 360° if this sum exceeds 360°). The X-component azimuth is the azimuth of the Y-component plus 90° (again, correcting for azimuths exceeding 360°). The
average of the azimuths was used for the absolute orientations of the borehole sensors’ horizontal components.

At VSAP, P-waves from seven earthquakes were used with a standard deviation of the calculated azimuths of 9.7°. At CUSSO, two time spans of bedrock sensor installations were covered by the recordings; as such, two sets of horizontal-component orientations had to be determined. The first orientation was constrained by the P-wave recordings from a single, impulsive, local earthquake. The orientation uncertainty cannot be estimated for this time period using the recordings from a single earthquake. However, the orientation is reasonable because, for both local earthquakes in this dataset that occurred during this period, the rotated surface and bedrock horizontal component recordings show consistent first-arrival transverse-component polarities at the surface and bedrock. For the second time period, P-waves from four local and four teleseismic events were used and the standard deviation of the component azimuths is 7.0°.

H/V from continuous recordings of ambient noise at the free surface, $H/V_{S,\text{noise}}$, were calculated to evaluate its ability to resolve the SH-wave transfer function, per the Nakamura (1989) approach. Satoh et al. (2001), among others, reported differences between $H/V_5$ and $H/V_{S,\text{noise}}$ in terms of the frequency of maximum amplification, $f_{\text{peak}}$, and amplification levels. Five hours of night-time (to reduce cultural noise) ambient noise, uncontaminated by earthquakes or blasts, recorded by CUSSO’s existing surface seismometer and from a temporary broadband seismometer co-located with VSAP’s surface accelerometer were used.
2.4.2 SPECTRAL RATIOS

Using the bedrock and surface amplitude spectra for the events listed in Tables 2.2 and 2-3, the mean spectral ratios $TF_T$, $TF_Y$, $HV_S$, and $HV_B$, and their standard deviations were calculated at VSAP and CUSSO. Individual spectral ratios were calculated by spectral division of the smoothed amplitude spectra. The amplitude spectra of all recordings at the free-surface were divided by a factor of two to remove the effect of free surface amplification for equivalency with traditional spectral ratios (equation 2.2) and for comparison with $HV_S$.

The mean $HV_{S,noise}$ spectral ratios were determined by first averaging the spectral ratios of smoothed (moving window lengths in Table 2.4) amplitude spectra calculated from the five-minute-long, 50 percent-overlapping windows of five hours of continuous recordings. Because the sources of ambient noise are likely from a suite of azimuths, recorded on both horizontal components, $HV_{S,noise}$ curves were calculated from the average spectrum of both horizontal components divided by the average vertical-component spectrum both horizontal components.

2.5 RESULTS

Figures 2.7 and 2.8 show mean $TF_T$, $TF_Y$, $HV_S$, and $HV_B$ curves and the corresponding standard deviations for VSAP and CUSSO, and the predicted SH-wave responses from Thomson-Haskell propagator matrices, $TH_{SH}$ (divided by two for consistency with $TF_T$ and $HV_S$). In general, there is remarkable consistency between $TF_T$ and $HV_S$, particularly within the frequency band of engineering interest (0.1-10 Hz), in terms of the peak frequencies. And at the first peak frequency (herein referred to as
observed-$f_0$), amplifications implied by both $TF_T$ and $HV_S$ are very similar. Furthermore, Figure 2.7 shows that peak frequencies of both $TF_T$ and $HV_S$ correspond closely with the fundamental and higher-mode resonances predicted by equation (2.7) for vertical-incidence S-waves and to $TH_{SH}$, calculated at average incidence angles.

Figure 2.7 Mean spectral ratios from recordings at VSAP and CUSSO and theoretical Thomson-Haskell SH-wave transfer functions ($TH_{SH}$) for average bedrock incidence angles of 25° at VSAP and 15° at CUSSO. Inverted triangles correspond to the fundamental and 10 next higher natural frequencies from equation (2.7). Solid, horizontal line indicates a ratio of 1 in each plot.
Figure 2.8 Mean spectral ratios shown in Figure 2.7 and mean ±1 standard deviation regions (solid gray). Solid, horizontal line indicates a ratio of 1 in each plot.

The mean ± one standard deviations (Fig. 2.8) demonstrate that the spectral-ratios’ peak frequencies are generally consistent between events, regardless of distance (as also observed by Zandieh and Pezeshk, 2011) for $HV_S$. However, $HV_S$ has greater variability and resolves observed-$f_0$ with less resolution than $TF_T$, based on $HV_S$ having broader half-widths of the lowest-frequency peaks. It is possibility that some of the variability in the peak frequencies is due to nonlinear responses. Rubinstein (2011) reported evidence of nonlinear response for ground accelerations as low $34 \text{ cm/s}^2$, which is comparable with the largest acceleration observed at VSAP of $24 \text{ cm/s}^2$. The spectral ratios from individual events were examined and it was observed-$f_0$ does not decrease with PGA, which is interpreted as evidence that no nonlinearity occurred.

The $H/V$ curves calculated from recordings of ambient noise, $HV_{S,noise}$, at both sites are plotted in Figure 2.9. There are important differences between S-wave spectral ratios and $HV_{S,noise}$ curves for frequencies greater than observed-$f_0$, as discussed in the Discussion section.
Figure 2.9 H/V curves derived from five-hours of ambient noise, $N_{V_{S,noise}}$ and $E_{V_{S,noise}}$, for the North- and East-components, respectively, recorded at VSAP and CUSS0. For comparison, S-wave H/V, $H_{V_{S}}$, and the theoretical Thomson-Haskell SH-wave transfer function, $T_{SH}$, are also shown.

2.6 DISCUSSION

2.6.1 EMPIRICAL AND THEORETICAL SH-WAVE TRANSFER FUNCTIONS

At both sites, $T_{F_{T}}$, the empirical SH-wave transfer function is similar to the theoretical SH-wave transfer function from the elastic Thomson-Haskell propagator matrix method, $T_{SH}$, for average and for vertical bedrock incidence angles. Evidently, the large impedance contrast between the northern Mississippi Embayment sediment overburden and the bedrock bends transverse-wave arrivals from a range of bedrock incidence angles to nearly vertical incidence at the surface. And consequently, averaging the spectral ratios from transverse-component recordings of direct-, head-, and Lg-waves reveals empirical SH-wave site responses suitable for engineering purposes. Furthermore, the similarities between $T_{F_{T}}$ and $T_{SH}$ indicate that 2-D and 3-D effects do not contribute significantly to the site responses at the VSAP and CUSS0 sites. However, 1-D site response models might not be applicable nearer to the edge of the Embayment, due to basin-edge effects as
observed by Ramírez-Guzmán et al (2012) and modeled by MacPherson et al. (2010), or in settings with complicated subsurface structures. In addition, the similarity between $TF_T$ at VSAP and CUSSO and the corresponding theoretical responses, which do not include anelasticity, also supports the observations of relatively low intrinsic attenuation for body waves in the Mississippi Embayment made by Langston (2003).

The $TF_T$ curves show significant SH-wave amplification at peak frequencies from the fundamental to higher than the $10^{th}$ natural frequency at each site. The maximum observed amplification factors from the $TF_T$ curves are $8.5\pm6.2$ at 12.9 Hz ($7^{th}$ natural frequency) at VSAP and $15.0\pm4.8$ at 1.3 Hz ($3^{rd}$ natural frequency) at CUSSO. The theoretical SH-wave transfer functions predict amplifications of 10.1 at VSAP and 8.3 at CUSSO for the peaks nearest 12.9 Hz and 1.3 Hz, respectively. At observed-$f_0$, amplification at VSAP is $7.8\pm5.0$ and $4.6\pm2.5$ at CUSSO. For comparison, an amplification of 4.8 is predicted by the theoretical SH-wave transfer functions at $f_0$ at both sites. The theoretical responses at CUSSO are provisional and require validation of the bedrock S-wave velocity used in this study. Nevertheless, the bedrock velocity employed is apparently reasonable as evidenced by the similarities between the observed amplifications and the theoretical SH-wave response.

2.6.2 S-WAVE H/V

Peak amplifications implied by $HV_S$ are similar to peak $TF_T$ amplifications: the maxima are $8.3\pm7.0$ at VSAP at 1.1 Hz and $11.1\pm8.7$ at 1.7 Hz at CUSSO. The theoretical SH-wave transfer functions predict amplifications of 4.8 at VSAP and 7.2 at CUSSO for the peaks nearest 1.1 Hz and 1.7 Hz, respectively. And at observed-$f_0$, amplifications are
8.3±7.0 at VSAP and 6.1±5.1 at CUSSO; 4.8 is the theoretical amplification at both sites at $f_0$.

Below a site-specific frequency, mean $TF_T$ and $HV_S$ curves are similar for both VSAP and CUSSO and they resemble the theoretical SH-wave transfer functions at each site. However, there are differences between $TF_T$ and $HV_S$, which their ratio makes clear (Fig. 2.10): for frequencies above approximately the fifth natural frequencies (~9 Hz and ~2.0 Hz at VSAP and CUSSO, respectively), $HV_S$ is consistently less than the observed SH-wave transfer function at both sites. This difference is much greater at the deeper-soil site, CUSSO. At lower frequencies, the ratios of $TF_T$ to $HV_S$ tend to oscillate about one. At these frequencies, the differences are due in large part to slight differences in the peak frequencies, rather than differences in amplification; $HV_S$ peaks occur at slightly lower frequencies than $TF_T$. For example, at both stations Figure 2.10 suggests that $HV_S$ yields greater amplification than $TF_T$ by a factor of two to three for frequencies near $f_0$. However, the differences of the amplifications at the respective observed-$f_0$ are much less: 7% at VSAP and 25% at CUSSO.
Figure 2.10 Vertical-component amplification, $TF_V$, and the ratio of spectral ratios $TF_T$ to $HV_S$ at CUSSO and VSAP. The fifth natural frequency, below which $HV_S$ approximates $TF_T$, is labeled as $f_a$. Inverted triangles correspond to the resonance frequencies in equation 2.7.

The curves in Figure 2.10 also indicate that the differences between $HV_S$ and $TF_T$ are due to vertical-component amplification. The influence of $TF_V$ on $HV_S$ is shown in equation (2.6): $HV_S$ is indirectly related to $TF_V$, and when $HV_B$ is nearly one, as at CUSSO (Fig. 2.8), the ratio of $TF_T$ to $HV_S$ should be $TF_V$. At VSAP, $HV_B$ is more complicated than at CUSSO, and is generally greater than one. As such, the ratio of $TF_T$ to $HV_S$ is generally greater than $TF_V$. However, at both stations, the two curves in Figure 2.10 are correlated, demonstrating the strong control of the vertical-component transfer function on $HV_S$. Therefore, the ability of $HV_S$ to approximate $TF_T$ depends on $TF_V$. 
$TF_V$ is consistent with the vertical response predicted by Thomson-Haskell propagator matrices for incident SV-waves, $TH_{SV,V}$ (Haskell, 1953; Haskell, 1962), as shown in Figure 2.11. Therefore, the major differences between $TF_V$ and $HV_S$ (Fig. 2.10) are explained by the amplification of transmitted SV-waves and converted P- and SV-waves. Furthermore, $HV_S$ will more accurately approximate the SH-wave transfer function when corrected for $TF_V$ (equation 2.6), which can be calculated by plane-wave propagation matrices. The similarity of the curves in Figure 2.11 also suggests that for the steeply ascending waves recorded by both arrays, it appears to be reasonable to correct for free-surface amplification on the vertical component by division by a factor of two, which was done to be consistent with $TF_V$. A thorough treatment of this particular topic is beyond the scope of this paper.

![Figure 2.11 Observed, $TF_V$, and the predicted, $TH_{SV,V}$, vertical-component amplification for an SV-wave with an angle of incidence of 15° at CUSSO.](image)

2.6.3 AMBIENT NOISE H/V

Figure 2.9 compares $HV_{S,noise}$ curves with $HV_S$ and the theoretical SH-wave responses, and reveals that $HV_{S,noise}$ clearly identifies the fundamental site frequency, as
observed in numerous studies (see e.g. Nakamura, 1989; Bodin and Horton, 1999; Langston, et al., 2009). At both stations, the amplification of the first peak of $HV_{S,noise}$ from either horizontal component is similar to $HV_s$. Therefore, $HV_{S,noise}$ is effective in both identifying the site $f_0$ and indicating the level of amplification at or near the site $f_0$. Higher mode resonances, however, are not clearly identified with $HV_{S,noise}$ suggesting the presence of additional phase arrivals with energetic vertical motions. Therefore, although this methodology may be useful to calculate an average shear-wave velocity model, it does not reveal the frequencies at which peak amplifications occur ($7^{th}$ and $3^{rd}$ natural frequencies at VSAP and CUSSO, respectively), nor their magnitudes, and is not suitable for studies of detailed velocity structure or site response.

2.6.4 ON THE APPLICABILITY OF S-WAVE H/V

Observed spectral ratios in this study suggest that the ability of S-wave and ambient-noise H/V to approximate the site transfer function in the northern Mississippi Embayment at frequencies of engineering importance, depends on the site’s natural frequencies. Both $HV_s$ and $HV_{S,noise}$ approximate site response at $f_0$. However, if higher modes occur at frequencies of engineering interest, they will not be revealed by $HV_{S,noise}$ and may be underestimated by $HV_s$ due to the amplification of high-frequency vertical motions. This is important because $f_{peak}$ may not correspond with $f_0$ in the Embayment, as at VSAP and CUSSO, and therefore maximum amplification may not be observable by H/V. However, $HV_s$ estimates the site response for frequencies up to the fifth natural frequency, which may be sufficient for sites over thinner ($< \sim 100$ m) sediment layers or that have faster sediment S-wave velocity structures.
In addition, Rong et al. (2016) demonstrated that $HV_s$ curves estimate the nonlinear site transfer function in cases of strong ground motions. Therefore, $HV_s$ may be useful for estimating the nonlinear site transfer function in the Embayment, because $HV_s$ reliably approximates the site response at lower frequencies and because high frequency responses are decreased due to nonlinear effects (e.g., see Rong et al., 2016). Evaluating this will be possible when strong-motions are recorded by VSAP and CUSSO.
3.1 INTRODUCTION

Fundamental frequency $f_0$ and associated amplification, $A_0$, from approximations of full-resonance expressions, were evaluated as alternatives to the site factors determined from the Vs30 proxy. The $f_0$ and $A_0$ include the effects of impedance contrasts and SH-wave resonance, and are dependent on shallow earth models. Thus, these parameters should provide more reliable estimation of primary site response parameters. The evaluation presented in this chapter consisted of comparing these simplifications to the full-resonance responses at 11 seismic stations in the central and eastern U.S. In addition, encouraged by the favorable H/V results in Chapter 2, $A_0$ and $f_0$ measured from S-wave horizontal-to-vertical spectral ratios were compared with the theoretical $A_0$ and $f_0$ at these sites, which are on a variety of site conditions (e.g., the soil site, and weathered and unweathered rock sites in Fig. 3.1). Furthermore, S-wave H/V curves developed at the sites were interpreted in light of possible site response from deeper, unmodeled impedance boundaries.
Figure 3.1 Sites (triangles) on soil layers (S) or rock outcrops (R). H and V represent amplitude spectra of horizontal- and vertical-component recordings, respectively, at these locations. \( R_1 \) and \( R_2 \) are possible reference sites, installed on weathered rock or unweathered bedrock, respectively.

3.2 PRIMARY SITE RESPONSE PARAMETERS

3.2.1 SIMPLIFIED EXPRESSIONS OF \( A_0 \) AND \( f_0 \)

Theoretically, site-effect amplification peaks are caused by the constructive interference of SH-waves trapped between the free surface and stiff bedrock (Haskell, 1960), where the frequencies of the peaks are determined by the sediment velocity structure. As originally presented in Okamoto (1973; as reviewed in Sánchez-Sesma and Crouse, 2015), the amplification at \( f_0 \) at a soil site relative to a nearby reference site can be approximated by

\[
\tilde{A}_0 \approx \frac{1}{\left(\frac{\rho_s V_s}{\rho_b V_b}\right) + \frac{\pi \gamma_s}{2}}, \tag{3.1}
\]

where \( \rho \) is density, \( V \) is shear-wave velocity, \( \gamma \) is the shear-wave damping ratio and the \( s \) and \( b \) subscripts are for sediment overburden and bedrock, respectively. This expression has been compared with full-resonance site responses more recently by Dobry et al. (2000) and Banab et al. (2012).
For a single, slow layer subjected to a vertical-incidence SH-wave from a faster underlying bedrock, the fundamental resonance frequency (Haskell, 1960) is

\[ f_0 = \frac{V}{4 \cdot Z_b}, \] (3.2)

where \( Z_b \) is the depth to stiff bedrock.

To calculate \( A_0 \) and \( f_0 \) using equations 3.1 and 3.2, one-layer, average earth models are used, where average velocities are the time-weighted average of the shear-wave velocity structures \( V = \frac{Z_b}{\sum h_i V_i} \), where \( Z_b \) is the depth to the fastest layer in each model and \( h_i \) and \( V_i \) are the layer thicknesses and shear-wave velocities, respectively) and density and damping ratios are the layer thickness-weighted average values.

### 3.2.2 FULL-RESONANCE SITE RESPONSES

\( \tilde{A}_0 \) and \( \tilde{f}_0 \) were compared with full-resonance (FR), linear site responses for the multiple-layer velocity structures and vertically incident SH-waves calculated using Thomson-Haskell propagator matrices (Haskell, 1953, 1960; equations 2.4 and 2.5). Comparisons in the Chapter 2 demonstrated that the 1-D full-resonance responses calculated by the Thomson-Haskell matrix method are reasonable for two sites in this study area—namely CUSSO and VSAP—by comparing the theoretical responses with borehole surface-to-bedrock spectral ratios. Thus, the same methodology was employed in this study.

Viscoelastic effects were included in the calculations, modeled as Kelvin-Voigt solids, through complex shear moduli in each layer. This assumes frequency-independent
damping, as is the case in common site-response codes such as SHAKE91 and STRATA. Because the FR response is the ratio of the amplitude spectrum of motion at the surface to that of the input motions from bedrock into the stack of overlying layers, it includes free-surface amplification, which is a factor of 2 for SH-waves. Finally, the full-resonance responses were divided by 2 for compatibility with equation 3.1 and empirical observations made at the free surface only.

Figure 3.2 shows that linear, fundamental-mode, FR site responses for the sites in Figure 1.4b are consistent with $\tilde{A}_0$ and $\tilde{f}_0$, suggesting the appropriateness of both the expressions given in equations 3.1 and 3.2 and of using simplified, one-layer earth models to approximate $f_0$ and $A_0$ at these sites. Similar to Figure 1.5, Figure 3.2 also compares these theoretical responses with the site-class D site coefficients (Dobry et al., 2000). This comparison indicates that $\tilde{A}_0$ and $\tilde{f}_0$ more closely estimate the FR response than do the amplifications predicted by the Vs30 proxy.
Figure 3.2 Site responses at the fundamental frequency for each site shown in Figure 1.4. Points plotted show the amplification predicted by equation 3.1 at $f_0$, calculated from equation 3.2, using a simplified one-layer earth model over a bedrock half-space. The soil model shear-wave velocity was calculated by the time-weighted average, and density and damping, calculated from statistical relationships with shear-wave velocity, were layer-thickness-weighted averages. Short- and long-period amplification factors for site class D sites and weak-motion input are also shown.

3.2.3 ESTIMATING $A_0$ AND $f_0$ FROM S-WAVE H/V

Frequency-dependent amplifications due to near-surface geologic conditions are measured with respect to a rock reference site (Borcherdt, 1970; i.e., $H_S/H_R$ in Figure 3.1). However, depending on the source mechanism and the relative positions of the soil and rock sites to the source, differences in the source and path terms for rock and soil sites may not be negligible, (e.g., soil and rock sites in Fig. 1.3) as is the case for sites far from the edge of broad basins such as the Mississippi Embayment and Illinois Basin. An additional concern is that weathered and fractured rock outcrops contribute to the site response (e.g., Steidl et al., 1996; and reference site $R_2$ in Figure 3.1), which must be accounted for when estimating site response from the spectral ratios of soil- to rock-site spectral ratios.
Using single-station H/V spectral ratios ($H_s/V_s$ in Figure 3.1) has become a popular empirical method to estimate site response because of ease of use and low cost. Spectral ratios of horizontal-to-vertical component recordings of ambient noise (e.g., Nakamura, 1989) are particularly attractive because of the ubiquity of the microtremor energy source and associated minimal recording time needed. However, several studies have demonstrated that spectral ratios of ambient noise do not always produce reliable estimates of site response (e.g., Satoh et al., 2001; Woolery et al., 2012; Perron et al., 2018), particularly for modes above the fundamental (e.g., Carpenter et al., 2018; Wang et al., 2019). Lermo and Chávez-García (1993), using earthquake recordings from the Mexico City area, were the first to demonstrate that single-station spectral ratios formed by the ratio of the horizontal-component S-wave spectrum to that of the vertical component, H/V, can estimate site response. Field and Jacob (1995) found that peak frequencies (i.e., the frequencies corresponding to peaks in the site-response transfer functions) were revealed by S-wave H/V in the Bay Area, California, and that the corresponding implied amplifications were similar to those revealed from traditional soil-site/reference-site spectral ratios. Laurendeau et al. (2017) observed the same in the Quito Basin of Ecuador.

As a corollary, Field and Jacob (1995) found that established reference sites had relatively flat S-wave H/V curves, further supporting the reliability of H/V curves in estimating site response. Zhao et al. (2006) used S-wave H/V observations from Japan for site classifications and demonstrated that these spectral ratios revealed the fundamental frequencies at more than 600 K-net stations. Furthermore, they showed that $f_0$ and $A_0$ measured from H/V curves, $f_{0,H/V}$ and $A_{0,H/V}$, respectively, have little to no dependence on station-event offset, or event magnitude or focal depth. Closer to the area of this study,
Murphy and Eaton (2005) observed that S-wave H/V curves at sites in southern Ontario, Canada, provided reliable fundamental-mode site-response estimates at stations underlain by sediments, but they did not observe responses at modes higher than the fundamental one.

For the central and eastern U.S., Woolery et al. (2009) found that S-wave H/V inconsistently approximated 1-D responses calculated at sites in the Wabash Valley. However, they noted that their spectral ratios were calculated from only one earthquake recording, which may be insufficient to reliably capture the site response. In contrast, Zandieh and Pezeshk (2011), supported by S-wave H/V observations made by Sedaghati et al. (2018), demonstrated that this technique provided first-order approximations of the site response in the Mississippi Embayment. The observations of Zandieh and Pezeshk (2011) agreed with those of Zhao et al. (2006) that the earthquake H/V curves showed little dependence on the source parameters, including event-station offset. Likewise, Carpenter et al. (2018) showed that S-wave H/V approximates the empirical transfer functions at low frequencies (up to the fifth resonance mode) at the bedrock-penetrating borehole observatories in the Mississippi Embayment, CUSSO and VSAP (the Paducah site shown in Figure 1.4b). In addition, Yassminh et al. (2019) calculated H/V of regional Lg-phase recordings at the EarthScope Transportable Array stations in the central and eastern U.S. and observed that H/V peak magnitudes correlated positively with site amplifications estimated by the reverse two-station method at center frequencies of 1, 2, 3, and 4 Hz using the same earthquakes. The strength of the correlation increased in large basins (e.g., the Illinois Basin), which supports the observations of Laurendeau et al. (2017) and Mendoza and Hartzell (2019) that S-wave H/V captures not only site responses from shallow layers,
but also from deeper basin rock layers, which can affect site response in the frequency band of primary engineering interest, 0.1 to 20 Hz.

Because of the success of using earthquake S-wave H/V to estimate site response in the Mississippi Embayment and elsewhere, the technique was evaluated in the project area. However, the success of this approach has not been demonstrated unambiguously in shallow-soil sites in the central and eastern U.S., and the results of Woolery et al. (2009) showing that H/V inconsistently estimates site response warrant additional analysis in the region. Therefore, seismic stations that have recorded local and regional earthquakes in and around sources of high seismic hazard in the central and eastern United States—the Wabash Valley and New Madrid Seismic Zones—along and which have existing shear-wave velocity profiles were used to further evaluate S-wave H/V as an estimator of site response at a variety of site conditions.

3.3 DATA
3.3.1 SEISMIC STATIONS AND EARTHQUAKE S-WAVE RECORDINGS

The locations of seismic stations used in this study are shown in Figure 3.3. Stations that recorded local and regional earthquakes and for which near-surface velocity profiles have been developed and are available in the literature were selected. The 11 stations included in this study are listed in Table 3.1. Velocity structures developed for each station are shown in Figure 3.4.
Table 3.1 Seismic station locations and distances to reflection/refraction surveys. Offset: Distance from the seismic station to the approximate center point of the reflection/refraction survey used for developing the velocity structure; N/A: indicates the offset was not available.

<table>
<thead>
<tr>
<th>Station</th>
<th>Latitude (°N)</th>
<th>Longitude (°E)</th>
<th>Offset (m)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>CUSSO</td>
<td>36.552</td>
<td>–89.330</td>
<td>45</td>
<td>Woolery et al. (2016)</td>
</tr>
<tr>
<td>EVIN</td>
<td>37.971</td>
<td>–87.530</td>
<td>175</td>
<td>Odum et al. (2010)</td>
</tr>
<tr>
<td>HAIL</td>
<td>37.752</td>
<td>–88.437</td>
<td>60</td>
<td>Odum et al. (2010)</td>
</tr>
<tr>
<td>HEKY</td>
<td>37.815</td>
<td>–87.592</td>
<td>0</td>
<td>Li et al. (2013)</td>
</tr>
<tr>
<td>MVKY</td>
<td>38.646</td>
<td>–83.761</td>
<td>60</td>
<td>Li et al. (2013)</td>
</tr>
<tr>
<td>OLIL</td>
<td>38.734</td>
<td>–88.099</td>
<td>20</td>
<td>Odum et al. (2010)</td>
</tr>
<tr>
<td>S46A</td>
<td>37.685</td>
<td>–87.715</td>
<td>0</td>
<td>Li et al. (2013)</td>
</tr>
<tr>
<td>SCMO</td>
<td>38.762</td>
<td>–90.641</td>
<td>N/A</td>
<td>Williams et al. (2007)</td>
</tr>
<tr>
<td>USIN</td>
<td>37.965</td>
<td>–87.666</td>
<td>55</td>
<td>Odum et al. (2010)</td>
</tr>
<tr>
<td>VSAP</td>
<td>37.131</td>
<td>–88.813</td>
<td>35</td>
<td>McIntyre (2008)</td>
</tr>
</tbody>
</table>
Figure 3.3 Seismic stations, colored by network code, and the epicenters of earthquakes used for this study. Circled stations were candidate reference stations—i.e., stations with little site response. The Wabash Valley (WBSZ), New Madrid (NMSZ), and Eastern Tennessee (ETSZ) Seismic Zones are delineated by the shaded ellipses. Precambrian faults from Hickman (2011) are plotted as thin gray lines; particular fault zones mentioned in the text are the Cottonwood Grove (CGFZ) and New Harmony (NHFZ) Fault Zones. Outlines of the Mississippi Embayment and the Illinois Basin are also shown.
Figure 3.4 Shear-wave velocity structures developed for each station. The plot on the right uses different depth and velocity scales for detailed view of the shallower velocity structures.

The study area includes several active seismic zones; of preeminence is the New Madrid, which was responsible for the 1811-12 sequence of large (M ≥ 7) earthquakes (e.g., Hough, 2009). Just north of the New Madrid Seismic Zone, at least five moderate (M ~5) earthquakes occurred in the Ste. Genevieve Fault Zone and Wabash Valley Seismic Zone in the past half century, which exceeds the number of moderate earthquakes observed in the New Madrid region during the same period (Yang et al., 2014).

The earthquake epicenters plotted in Figure 3.3 correspond to the events whose shear waves were used to calculate mean S-wave H/V curves for each site. Most earthquakes used in this study occurred in the New Madrid and Wabash Valley Seismic
Zones, but events in the Eastern Tennessee Seismic Zone were also used for stations MVKY in northeastern Kentucky and T47A in southwestern Kentucky. Previous studies from in this study area or nearby (e.g., Zandieh and Pezeshk, 2011; Carpenter et al., 2018) and elsewhere (e.g., Zhao et al., 2006) have shown that mean earthquake H/V curves do not depend strongly on event-station offsets or source parameters. In addition, as Parolai and Richwalski (2004) demonstrated through modeling synthetic seismograms, for a single 30-m-thick soil layer over a bedrock half-space model, where both the soil and bedrock velocities are comparable to those used in this study, $f_{0,H/V}$ approaches $f_{0,FR}$ determined by full-resonance calculations, $f_{0,FR}$, at offsets greater than 35 km, whereas at smaller offsets, $f_{0,H/V}$ decreases with respect to $f_{0,FR}$. Therefore, because of the relatively large event-station offsets in most of this dataset (Table 3.2), the impact of variable offsets was expected to be minor and thus all available recordings of earthquakes of magnitude 2.5 and greater at local and regional offsets—within 3° of each station—recorded throughout each station’s operational history were used.
Table 3.2 Station recording and H/V processing parameters, summary of earthquakes recordings used, and peak ground accelerations. Stations with an asterisk were only used as reference sites. $f_s$: Sampling frequency. $f_{0,i}$: Initial estimate of the empirical site fundamental frequency. $t_{\text{win}}$: Window length around the direct S-wave. $f_{\text{smooth}}$: Length of running-average smoothing filter used to smooth amplitude spectra. Neq: Number of earthquake. $M_{\text{max}}$: largest magnitude. $D_{\text{q1}}$: First quartile of the event-station offsets. $D_{\text{q2}}$: Median event-station offset. $D_{\text{q3}}$: Third quartile of the event-station offsets. PGA$_{\text{max}}$: Maximum peak ground motion within the windowed seismograms.

<table>
<thead>
<tr>
<th>Station</th>
<th>$f_s$ (Hz)</th>
<th>$f_{0,i}$ (Hz)</th>
<th>$t_{\text{win}}$ (s)</th>
<th>$f_{\text{smooth}}$ (Hz)</th>
<th>Neq</th>
<th>$M_{\text{max}}$</th>
<th>$D_{\text{q1}}$ (km)</th>
<th>$D_{\text{q2}}$ (km)</th>
<th>$D_{\text{q3}}$ (km)</th>
<th>PGA$_{\text{max}}$ (m/s$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BLO*</td>
<td>100</td>
<td>1.0</td>
<td>5.0</td>
<td>0.5</td>
<td>27</td>
<td>4.0</td>
<td>147</td>
<td>226</td>
<td>273</td>
<td>4.41e-03</td>
</tr>
<tr>
<td>CUSSO</td>
<td>200</td>
<td>0.25</td>
<td>20.0</td>
<td>0.125</td>
<td>31</td>
<td>4.7</td>
<td>37</td>
<td>308</td>
<td>310</td>
<td>1.24e-02</td>
</tr>
<tr>
<td>EVIN</td>
<td>100</td>
<td>1.8</td>
<td>5.0</td>
<td>0.9</td>
<td>73</td>
<td>4.1</td>
<td>186</td>
<td>246</td>
<td>266</td>
<td>3.45e-02</td>
</tr>
<tr>
<td>HAIL</td>
<td>100</td>
<td>2.0</td>
<td>5.0</td>
<td>1.0</td>
<td>123</td>
<td>4.1</td>
<td>149</td>
<td>184</td>
<td>214</td>
<td>3.57e-02</td>
</tr>
<tr>
<td>HEKY</td>
<td>200</td>
<td>2.6</td>
<td>5.0</td>
<td>1.3</td>
<td>60</td>
<td>4.4</td>
<td>220</td>
<td>237</td>
<td>282</td>
<td>2.53e-02</td>
</tr>
<tr>
<td>MVKY</td>
<td>100</td>
<td>2.0</td>
<td>5.0</td>
<td>1.2</td>
<td>21</td>
<td>4.2</td>
<td>176</td>
<td>233</td>
<td>250</td>
<td>7.89e-02</td>
</tr>
<tr>
<td>OLIL</td>
<td>100</td>
<td>1.2</td>
<td>5.0</td>
<td>0.6</td>
<td>145</td>
<td>4.1</td>
<td>238</td>
<td>280</td>
<td>304</td>
<td>3.62e-02</td>
</tr>
<tr>
<td>S46A</td>
<td>40</td>
<td>2.4</td>
<td>5.0</td>
<td>1.2</td>
<td>15</td>
<td>3.5</td>
<td>118</td>
<td>215</td>
<td>227</td>
<td>1.00e-03</td>
</tr>
<tr>
<td>SCMO</td>
<td>100</td>
<td>3.6</td>
<td>5.0</td>
<td>1.8</td>
<td>75</td>
<td>4.1</td>
<td>225</td>
<td>256</td>
<td>277</td>
<td>9.33e-02</td>
</tr>
<tr>
<td>SLM*</td>
<td>100</td>
<td>1.0</td>
<td>5.0</td>
<td>0.5</td>
<td>189</td>
<td>4.1</td>
<td>219</td>
<td>250</td>
<td>280</td>
<td>9.90e-03</td>
</tr>
<tr>
<td>T47A*</td>
<td>100</td>
<td>1.0</td>
<td>5.0</td>
<td>0.5</td>
<td>158</td>
<td>4.0</td>
<td>226</td>
<td>253</td>
<td>294</td>
<td>1.76e-02</td>
</tr>
<tr>
<td>USIN</td>
<td>100</td>
<td>1.8</td>
<td>5.0</td>
<td>0.9</td>
<td>159</td>
<td>4.1</td>
<td>223</td>
<td>242</td>
<td>268</td>
<td>4.43e-02</td>
</tr>
<tr>
<td>VSAP</td>
<td>200</td>
<td>1.0</td>
<td>5.0</td>
<td>0.5</td>
<td>11</td>
<td>5.2</td>
<td>42</td>
<td>86</td>
<td>168</td>
<td>2.40e-01</td>
</tr>
<tr>
<td>WCI*</td>
<td>100</td>
<td>1.0</td>
<td>5.0</td>
<td>0.5</td>
<td>13</td>
<td>4.4</td>
<td>143</td>
<td>215</td>
<td>312</td>
<td>4.68e-03</td>
</tr>
<tr>
<td>WVIL</td>
<td>100</td>
<td>1.6</td>
<td>5.0</td>
<td>0.8</td>
<td>69</td>
<td>4.1</td>
<td>209</td>
<td>266</td>
<td>282</td>
<td>3.58e-01</td>
</tr>
</tbody>
</table>

The stations are installed on a variety of site conditions; the majority are on unlithified sediment units, primarily loess, glacial till, or Quaternary alluvium, with thicknesses ranging from a few meters to tens of meters (Fullerton et al., 2003; Odum et al., 2010) overlying bedrock or weathered bedrook. Sediment thicknesses exceed 30 m (Gray, 1989; Moore et al., 2009) at several of these sites. Most stations are in the Illinois Basin, which is delineated as the approximate extent of large coalbed gas- or oil-bearing reservoirs in Pennsylvanian or Upper Mississippian sandstone in the basin (Swezey, 2007). In addition, VSAP and CUSSO are in the Mississippi Embayment, where sediment thicknesses are 100 and 585 m, respectively. Also, S-wave H/V curves at four additional...
sites that lacked velocity models but were installed on bedrock or weathered bedrock (Fig. 3.3) were calculated. These four stations were evaluated as reference sites, to investigate potential amplifications due to deeper layers in the Illinois Basin.

3.3.2 INPUTS FOR EQUATIONS 3.1 and 3.2: $V(z)$, $\rho$, and $\gamma$

The velocity structures shown in Figure 3.4 were derived primarily from surface SH-wave reflection and refraction surveys (Table 3.1). In addition, the layer thicknesses in the velocity structures at stations CUSSO, HEKY, and VSAP (Fig. 3.3) were determined by borehole observations. Therefore, these sites have better-constrained velocity structures than other stations used in this study. The deeper (> 135 m) velocity structure at CUSSO was determined from a combination of sonic logging and from earthquake S-wave travel times between bedrock and the surface (Woolery et al., 2016). The bedrock S-wave velocity of 2,079 m/s was determined by a grid-search fit of the FR response to the surface-to-bedrock spectral ratios, using a modified version of OpenHVSR (Bignardi et al., 2016). This velocity is ~30 percent larger than the top-of-bedrock velocity reported in Woolery et al. (2016) of 1,452 m/s, but within three percent of the velocity used for CUSSO in Chapter 2 (Carpenter et al., 2018) of 2,132 m/s. Figure 3.5 shows the S-wave velocity structures that produced the best-fit to the empirical transfer function at CUSSO for 10 trials.
Figure 3.5 Best-fitting shear-wave velocity structures at CUSSO determined through 10 trials of 10,000 grid-searches through the parameter space, colored by misfit. Wider lines are the velocity structure used in Chapter 2 (black), the median of all velocity structures with error bars scaled by the median-absolute deviations (purple), and the structure that produced the lowest misfit (colored by misfit).

These velocity structures were used to estimate $\gamma_s$ using statistical relationships between $Q_{ef}$, the effective shear-wave quality factor that includes the effects of frequency-independent intrinsic attenuation and frequency-dependent scattering, and shear-wave velocity, which have been developed for the Mississippi Embayment by Wang et al. (1994) and Campbell (2009). Following a standard approach for quantifying viscoelastic effects for predicting site responses in geotechnical applications, $Q_{ef}$ was interpreted to contain both frequency-dependent and -independent attenuation effects, which is consistent with borehole observations (e.g., Abercrombie, 1997) and which is commonly employed in 1-
D site-response software (e.g., STRATA; Kottke and Rathje, 2008). Thus, $Q_{\text{ef}}$ and $\gamma_s$ are related by the following equation (Campbell, 2009):

$$\gamma_s = \frac{1}{2} Q_{\text{ef}},$$  \hspace{1cm} (3.3)

which can be used to relate shear-wave velocity to effective shear-wave damping ratio via statistical $Q_{\text{ef}}(V)$ relationships.

The $Q_{\text{ef}}(V)$ relationships came from equations 20, 21, and 22 in Wang et al. (1994; herein Wang94$_{p}$, Wang94$_{B}$, and Wang94$_{A}$, respectively) and models 1 and 3 in Campbell (2009, equations 14 and 15; herein Campbell09$_{1}$ and Campbell09$_{3}$, respectively). In addition, the relationship between $Q_{\text{ef}}$ and depth developed for the Mississippi Embayment (www.eas.slu.edu/People/RBHerrmann/MAEC/maecgnd.html; last accessed July 2019), $Q_{\text{ef}}(z) = 6 \times z^{0.24}$, was used to calculate an additional estimate of $Q_{\text{ef}}$ for each layer. The $Q_{\text{ef}}$ assigned to each layer in each station’s velocity structure was the mean value of the six estimates—five velocity-dependent models and one depth-dependent model—except at CUSSO and VSAP. For CUSSO and VSAP, the $Q_{\text{ef}}$ models specifically developed by Wang et al. (1994) for those sites (namely, Wang94$_{B}$ and Wang94$_{p}$, respectively) were exclusively used. The $\gamma_s$ of each layer was then determined from the layer’s $Q_{\text{ef}}$ using equation 3.3. Figure 3.6 shows the $\gamma_s$ structures determined from the time-weighted average velocity structures using each of the six $Q_{\text{ef}}$ models, as well as the mean $Q_{\text{ef}}$. The range of $\gamma_s$ estimates is similar at each site and the median of the ranges is 0.019; thus, the uncertainty of $\gamma_s$ was assigned to be half of that range, or 0.009. Because damping values are typically small (~0.05) compared to the impedance ratio, the damping term in equation
3.1 exerts less influence on the fundamental-mode amplification calculated from this expression compared with the other terms. For example, at HEKY, where the largest amplification is predicted, the variation in both $\tilde{A}_0$ and full-resonance fundamental-mode amplification, $A_{0,FR}$, responses when damping ratios in each layer are increased and decreased by 0.009, relative to the response calculated for the mean damping values, are only 12 and 16 percent, respectively (Fig. 3.7). Figure 3.8 shows the layer damping ratios for each station, calculated from the stations’ velocity structures.

Figure 3.6 Comparison of damping values predicted by five $Q(V_s)$ relationships, using the time-weighted average velocity for each site and one $Q(z)$ relationship. The mean value of these six estimates is also shown.
Figure 3.7 Comparison of full-resonance and fundamental-mode amplification from equation 1 for HEKY, when damping is increased ($\gamma_S+0.009$) and decreased ($\gamma_S-0.009$) from the mean damping ratio in each layer.

Figure 3.8 Layer damping ratios developed for each station from the mean value calculated from five $Qef$-velocity relationships and one $Qef$-depth relationship. The plot on the right uses a different depth scale for detailed view of the shallower layers.
Because shear-wave velocity structures are commonly determined by noninvasive techniques, layer densities are not often measured. Measured densities were not available at most sites, and therefore the layer densities at those sites were estimated from the statistical relationship between density and shear-wave velocity in Boore (2016; herein, Boore16) at those sites. Figure 3.9a shows that the measured layer densities at stations VSAP and CUSSO, where each layer contains sands and various amounts of clays, silts, and gravels (Street et al., 1997; Woolery et al., 2016; Carpenter et al., 2018), are within the range of densities observed globally for these bulk lithologies and the range of observed layer velocities. Furthermore, Figure 3.9b shows that the densities predicted by shear-wave velocity for each layer using the Boore16 relationship are generally within 10 percent of the measured values at these sites.
Figure 3.9 (a) Compilation of densities and shear-wave velocities for various lithologies, and the corresponding statistical relationship of density ($\rho$) versus S-wave velocity (red curve); modified from Boore (2016). $Vs$-$\rho$ ordered pairs for the sediment layers at CUSSO and VSAP are shown for comparison with this compilation. (b) Comparison of the layer densities at CUSSO and VSAP with those predicted from the corresponding layer velocities using the Boore16 relationship. Dashed line shows equivalence; dotted lines are ±10 percent of equivalence.

The fundamental-mode amplifications estimated by equation 3.1 and calculated from densities predicted by Boore16 are slightly reduced with respect to those calculated from the measured densities; the reduction in $A_0$ at both sites—11 percent and 12 percent at CUSSO and VSAP, respectively—is primarily the result of the measured bedrock
densities being larger than those predicted by Boore16. The difference between $A_{0,FR}$ at VSAP determined using the measured and predicted densities is the same as that determined using equation 3.1; at CUSSO, the difference is only 4 percent (Fig. 3.10). The relatively small differences between the theoretical amplifications at $f_0$ suggest that the Boore16 model produces reasonable densities for amplification calculations in the study area. Figure 3.11 shows the layer densities for each station, calculated from the stations’ velocity structures.

Figure 3.10 Comparison of amplification factors, $A_0$, calculated from equation 3.1 using simplified, one-layer average velocity structures and full-resonance responses using measured densities and those predicted from the sediment and bedrock S-wave velocities using the Boore16 relationship. Error bars are scaled by the maximum and minimum $A_0$ calculated from the maximum and minimum sediment-bedrock density contrasts, respectively, where densities are increased or decreased from their predicted value by 1 $\sigma$. Standard deviations were calculated from the densities in Figure 3.9a in logarithmically spaced bins spanning the range of velocities in Figure 3.4, and range between 0.12 g/cm$^3$ and 0.20 g/cm$^3$ (consistent with the standard deviation of all residuals of 0.13 g/cm$^3$ reported in Boore, 2016). $A_0$ varies by less than 15 percent due to 1 $\sigma$ density variabilities.
As shown in Figure 3.12a, although the layer damping ratios and densities are not measured values, but are estimated from statistical relationships (except densities at CUSSO and VSAP), of all the terms in equation 3.1, the ratio of the bedrock to the average sediment velocities exerts the most control on fundamental mode amplification. Therefore, the velocities themselves (the bedrock-sediment velocity ratio, in particular) are most important for estimating $\tilde{A}_0$, and significant ($>20$ percent) absolute differences between the modeled and actual site responses due to the absolute uncertainties in the density or the damping-ratio estimates ($<0.20 \text{ g/cm}^3$ and 0.009, respectively) are not expected.
Figure 3.12 (a) Values of the terms in equation 3.1 calculated from average, one-layer-over-bedrock velocity structures. The upper plot shows absolute values; the values in the lower plot are normalized with respect to the maximum of each term. (b) The effect of 30 percent uncertainty in the bedrock S-wave velocity on $A_0$. Error bars are scaled by the change in $A_0$ due to a 30 percent increase (above the point) and a 30 percent decrease (below) in $V_b$. The percent differences due to these changes are plotted using the y-axis on the right.

Implicit in this evaluation is that the velocity structures are reasonably accurate. However, the effect of uncertainties in the bedrock S-wave velocity, $V_b$, can be larger than those of the other terms used to calculate $A_0$. Errors in the S-wave velocity structure of the
sediment column might be reduced or canceled out by the time-weighted averaging used to calculate $V_s$ in equation 3.1, but errors in $V_b$ would not be. As Carpenter et al. (2018) observed at CUSSO, such errors can occur at sites with a relatively thin weathered zone overlying a faster, more competent bedrock that controls whole-sediment-column resonance. For example, the velocity corresponding to the bedrock layers responsible for the site response at CUSSO was estimated to be ~30 percent larger than the bedrock velocity just below the sediment-bedrock contact.

Thus, although the actual uncertainties of $V_b$ are unavailable, the effect of a ±30 percent error in $V_b$ was evaluated. As demonstrated in Figure 3.12b, such an error, propagated though the corresponding bedrock density and damping estimates, resulted in average differences in $\tilde{A}_0$ of -30 percent and 25 percent for decreases and increases in $V_b$, respectively. Therefore, it is particularly important to reduce uncertainties in bedrock S-wave velocities in order to reliably predict site amplification factors. Furthermore, underestimating $V_b$ has a larger impact on the error in $\tilde{A}_0$ than overestimating $V_b$. Assuming a ±30 percent uncertainty in $V_b$ is representative of this dataset, then $\tilde{A}_0$ has an associated estimated uncertainty of ~±30 percent.

3.3.3 EMPIRICAL ESTIMATION OF $A_0$ and $f_0$ FROM S-WAVE H/V

The magnitudes and frequencies of the first prominent peaks of earthquake S-wave H/V spectral ratios were compared with $A_{0,FR}$ and $f_{0,FR}$. Following the processing steps in Carpenter et al. (2018), mean S-wave H/V curves were determined from transverse- and vertical-component seismograms of multiple earthquake shear-waves. Because of the limited operational span of the sensors at VSAP and CUSSO (Woolery et al., 2016;
Carpenter et al., 2018), all events recorded by both of these arrays during periods when the instrumentation was functioning properly were included in the spectral ratios.

To process the recordings, the Fourier spectra of the windowed, de-trended, demeaned, and tapered (5 percent of window-length cosine tapers) waveforms were calculated. Time windows began prior to the direct-S arrival and window lengths of 5.0 s at all stations—which avoids surface-wave contamination—were used except at CUSSO, which required a longer window of 20.0 s to capture multiple S-wave reverberations in the thick sediment column. Instrument responses were removed from these waveforms, and the amplitude spectra were calculated. Spectra were then smoothed with running Hanning windows whose lengths were site dependent: Lengths of one-half the frequency corresponding to the first observed peak were used. These processing parameters are given in Table 3.2 for each station.

Unreliable signal was excluded at each frequency via signal-to-noise ratios calculated in the frequency domain. Noise amplitude spectra were estimated from the Fourier spectra of the pre-P-wave noise windows of lengths equal to those used for the S-wave windows. SNRs were calculated by spectral division and used to filter out low-quality spectral estimates. Only S-wave spectra at frequencies for which both the transverse and vertical components had SNRs of at least 2.5 were retained, although as shown in Figure 3.13 the mean S-wave H/V curves are very similar, regardless of SNR threshold. Figure 3.14 is an example of processing a single earthquake recording, from windowing the waveforms to selection of the usable parts of the H/V curve.
Figure 3.13 Comparison of mean S-wave H/V curves at each station for various SNR thresholds. At least three observations with SNR greater than the given threshold were required to calculate H/V at each frequency.
Figure 3.14 Example of the processing workflow for the recordings at HEKY of an M 3.4 earthquake, 98 km away. After preprocessing the recordings (discussed in the text), noise (orange) and signal (green) windows are taken on the transverse- (HNT) and vertical-component (HNZ) recordings (left column). Amplitude spectra are calculated for both noise and signal windows and signal-to-noise ratios are formed for both components (middle column). The H/V spectral ratio for this recording is calculated from the spectral division of HNT by HNZ, but only for frequencies with SNR ratios of 2.5 or greater (right column).

Carpenter et al. (2018) obtained stable, mean S-wave H/V curves from 10 events with SNRs of 1.5 and greater for all frequencies on both the horizontal and vertical components. Hassani and Atkinson (2017) required even fewer recordings, a minimum of three, to obtain reliable H/V curves. The mean H/V curves in this study are calculated from tens of events for most stations at most frequencies. However, H/V estimates at particular frequencies at several stations were obtained from only a few events. Mean S-wave H/V were calculated only for frequencies with at least three individual-event observations having SNR ≥ 2.5. Table 3.2 summarizes the events used to calculate the mean spectral ratios at each site and their distribution with respect to the recording stations. Although the noise windows are relatively short, large variations in the noise amplitude spectra from
prior to the P-wave through the S-wave window are not expected because the S-to-P times of the earthquakes used in this study are relatively small (< 40 s). In addition, in the unexpected case that the noise characteristics changed substantially from prior to the P-wave through the S-wave window, systematic effects on the mean S-wave H/V ratios are not anticipated because the recordings of numerous earthquakes (> 10) were used for each station. Finally, because the maximum PGA is only 36 gal (at station WVIL; Table 3.2) all recordings were weak motion and therefore not expected to be modified by any nonlinear effects. Mean S-wave H/V curves are shown in Figure 3.15.
Figure 3.15 Theoretical full and simplified ($f_0, A_0$), and empirical (H/V) site responses and automatically picked peak responses. The H/V mean ± one standard deviation curves are also plotted as thin, dotted lines. For stations in the Illinois Basin (IB), the frequency band from 1 Hz to 3 Hz is highlighted, and except for sites along the Ohio River (EVIN and HEKY), H/V peaks were picked outside of this band; the picks ignored within this band are plotted as open circles ($A_{0,H/V,IB}$) for comparison. Peak H/V picks are also plotted ($A_{\text{peak},H/V}$, diamonds). Predicted responses from equation 3.1 are plotted at the corresponding frequencies predicted by equation 3.2 are indicated by asterisks. When two FR peaks are plotted, the peak response within the band 0.1 Hz to 20 Hz occurred at a frequency greater than $f_0$. Observed spectral ratio peak frequencies from other studies are indicated with vertical arrows and labeled with the corresponding study and data type: e=earthquake, n=ambient noise, and r=reflection soundings. C18=Carpenter et al. (2018). M15=McNamara et al. (2015), Odum et al. (2010). W07=Williams et al. (2007). Y19=Yassminh et al. (2019).
Peaks from the mean S-wave H/V curves were automatically picked using the MATLAB function FINDPEAKS and stipulated that peaks have a prominence of at least 1.0. In other words, if a peak’s height above either of the adjacent troughs was less than 1.0, it was not considered. Two peaks were considered to be of particular relevance to this study: the first peak \( A_{0,H/V} \) with corresponding frequency \( f_{0,H/V} \), which ostensibly corresponds with the fundamental frequency, and the maximum peak, \( A_{peak,H/V} \) with corresponding frequency \( f_{peak,H/V} \) within the band of engineering interest, from 0.1 to 20 Hz (Fig. 3.16).

The magnitudes and frequencies of the first prominent peaks at frequencies greater than 3 Hz were assigned as \( A_{0,H/V} \) and \( f_{0,H/V} \), respectively, at all but two sites in or adjacent to the Illinois Basin. The frequencies of the first peaks were much lower than those predicted by the FR responses and by equation 3.2 at these stations. And, as discussed in the Discussion section, these lower-frequency peaks may be related to resonances caused by large impedance contrasts within the sedimentary rock layers in the Illinois Basin and not to the responses of the shallow layers included in the site velocity profiles. Thus, H/V peaks below 3 Hz at stations in the basin were ignored except for EVIN and HEKY, whose peak frequencies are consistent with those predicted from the velocity structures shown in Figure 3.4. Similar to other Illinois Basin stations, however, at EVIN and HEKY the H/V peaks are notably broader than predicted. At SCMO, the first H/V peak is also broad and has similar character to those observed at the Illinois Basin stations. The next H/V peak at SCMO is more consistent with the FR response and equation 3.2 predictions, and the H/V curve from this peak to higher frequencies has a shape similar to the theoretical FR response. Furthermore, this peak nearly coincides with the resonant frequency predicted
from two-way travel times observed by Williams et al. (2007). Therefore, this second peak frequency and its corresponding magnitude were assigned as $f_{0,H/V}$ and $A_{0,H/V}$, respectively, at SCMO in the comparisons.

### 3.4 RESULTS

Figure 3.15 shows the theoretical FR and simplified site responses and the mean S-wave H/V ratios, as well as the peaks picked on both the theoretical and empirical curves. Figure 3.16 compares these theoretical and observed picks, amplifications, and corresponding frequencies, which are also listed in Table 3.3. Observed peak frequencies ($f_{\text{peak}}$) from other studies at stations used in this study are also shown on Figure 3.15. McNamara et al. (2015) measured $f_{\text{peak}}$ using the ratio of ambient noise power spectral densities at stations OLIL and USIN. Odum et al. (2010) estimated $f_{\text{peak}}$ from amplitude spectra, corrected for path effects, of S-wave recordings from the 2008 moment-magnitude 5.2 Mount Carmel earthquake at stations EVIN, HAIL, OLIL, USIN, and WVIL. They did not correct the spectra for source effects, and so their observations permit only qualitative comparisons, particularly in terms of amplitude. Williams et al. (2007) used zero-offset, two-way travel times from surface-derived bedrock reflections to estimate $f_0$ at station SCMO. And Yassminh et al. (2019) measured $f_{\text{peak}}$ at S46A using H/V of ambient noise and from multiple regional Lg-phase recordings.
Table 3.3 Theoretical and observed $f_0$ and $A_0$.

<table>
<thead>
<tr>
<th>Station</th>
<th>$f_{0,FR}$ (Hz)</th>
<th>$f_{0,eqn(2)}$ (Hz)</th>
<th>$f_{0,H/V}$ (Hz)</th>
<th>$A_{0,FR}$</th>
<th>$A_{0,eqn(1)}$</th>
<th>$A_{0,H/V}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>CUSSO</td>
<td>0.3</td>
<td>0.2</td>
<td>0.3</td>
<td>4.6</td>
<td>4.7</td>
<td>8.2</td>
</tr>
<tr>
<td>EVIN</td>
<td>1.2</td>
<td>1.8</td>
<td>1.5</td>
<td>5.8</td>
<td>6.3</td>
<td>12.2</td>
</tr>
<tr>
<td>HAIL</td>
<td>18.9</td>
<td>12.3</td>
<td>7.4</td>
<td>3.7</td>
<td>3.2</td>
<td>6.4</td>
</tr>
<tr>
<td>HEKY</td>
<td>2.7</td>
<td>2.5</td>
<td>2.7</td>
<td>10.2</td>
<td>10.0</td>
<td>7.9</td>
</tr>
<tr>
<td>MVKY</td>
<td>2.0</td>
<td>1.9</td>
<td>2.1</td>
<td>5.6</td>
<td>5.6</td>
<td>12.0</td>
</tr>
<tr>
<td>OLIL</td>
<td>5.2</td>
<td>4.1</td>
<td>3.6</td>
<td>2.6</td>
<td>2.7</td>
<td>4.6</td>
</tr>
<tr>
<td>S46A</td>
<td>12.5</td>
<td>12.5</td>
<td>13.3</td>
<td>5.7</td>
<td>5.7</td>
<td>6.1</td>
</tr>
<tr>
<td>SCMO</td>
<td>9.3</td>
<td>7.3</td>
<td>7.6</td>
<td>9.7</td>
<td>8.9</td>
<td>5.8</td>
</tr>
<tr>
<td>USIN</td>
<td>13.0</td>
<td>7.4</td>
<td>9.5</td>
<td>6.3</td>
<td>3.6</td>
<td>6.1</td>
</tr>
<tr>
<td>VSAP</td>
<td>1.2</td>
<td>1.0</td>
<td>1.0</td>
<td>4.8</td>
<td>5.0</td>
<td>8.3</td>
</tr>
<tr>
<td>WVIL</td>
<td>12.7</td>
<td>7.2</td>
<td>7.7</td>
<td>6.9</td>
<td>4.3</td>
<td>7.5</td>
</tr>
</tbody>
</table>
Figure 3.16 Comparisons of theoretical—full-resonance responses and simplifications in equations 3.1 and 3.2—and empirical (H/V) $f_0$ and $A_0$. (a) Frequency comparisons. (b) Absolute value of the percent differences of the frequencies in (a). (c) Amplification comparisons. (d) Absolute value of the percent differences in the amplifications.

3.4.1 SIMPLIFIED VERSUS FULL-RESONANCE $f_0$ and $A_0$

Figures 3.15 and 3.16 show that $\tilde{A}_0$ and $\tilde{f}_0$ correspond closely with the full-resonance responses at most stations. However, at two Wabash Valley sites, USIN and WVIL, $\tilde{A}_0$ underestimates $A_{0,FR}$ by ~40 percent and $\tilde{f}_0$ underestimates $f_{0,FR}$ by ~44 percent. A very strong impedance increase, associated with the contact between overlying Quaternary deposits and underlying weathered Pennsylvanian sedimentary rocks (Odum et al., 2010) distinguishes these sites from the other sites. At both sites, the impedance
ratios of the Quaternary sediment layers in contact with the weathered sedimentary units are nearly the same or greater than the impedance ratios of the weathered units to the underlying bedrock with the highest impedances. In addition, the weathered bedrock layers are relatively thick, more than half of the total thickness of the velocity profiles at both sites.

As shown in Figure 3.17, all layers contribute to the site response; therefore, both the weathered bedrock and underlying rock half-space were included in the analyses. However, the constructive and destructive interferences of waves reflecting from the free surface and these intermediate-depth layers, along with those waves between the surface and the base of the measured velocity profiles, yield a complex response that is more poorly approximated by the simplified factors $\tilde{A}_0$ and $\tilde{f}_0$. Station HAIL, where $f_{0,FR}$ is underestimated by 35 percent by equation 3.2, appears to have a similarly complex response. HAIL has a large, shallow impedance contrast in the Quaternary deposits, with thickness of approximately half the depth to bedrock. Therefore, the simplifications $\tilde{A}_0$ and $\tilde{f}_0$, although valid for most sites in this study, underestimate the more realistic FR responses for sites with at least one large impedance contrast corresponding to a relatively thick layer above fastest bedrock.
Figure 3.17 H/V and theoretical site responses at USIN using the simplified expressions in equations 3.1 and 3.2, and full-resonance calculations. FR responses are shown for the original structure in Figure 3.4 (FR\text{orig.}); for various subsets of that structure (FR\text{2lyr}, and FR\text{1lyr}), where the immediate layer below the layer designated by the subscript is treated as the half-space and layers below that are ignored; and for average structures: down to the deepest bedrock depth (FR\text{ave. str}) and the average of the first and second layers, with layer three taken as the half-space (VR\text{Ave. 1,2}).

3.4.2 THEORETICAL AND EMPIRICAL $f_0$

The S-wave H/V observations were compared with the FR responses. Figure 3.15 shows that S-wave H/V $f_{\text{peak}}$ generally corresponds with $f_{0,H/V}$. Figures 3.15 and 3.16 also show that predicted and observed $f_0$ are consistent at most stations: At CUSSO, HEKY, MVKY, S46A, and VSAP, $f_{0,H/V}$ is within 20 percent of $f_{0,FR}$ and within 40 percent of $f_{0,FR}$ at all sites but HAIL and SCMO. At stations OLIL, SCMO, USIN, and WVIL, H/V curves approximate the shape and relative separation of the FR response peaks, but are shifted to lower frequencies. The empirical and theoretical responses have little similarity at HAIL.
3.4.3 THEORETICAL AND EMPIRICAL $A_0$

As with $f_0$, S-wave H/V observations were compared with the FR responses. Figure 3.15 shows that observed peak amplifications occur at $f_0$ at all stations except CUSSO (as reported in Carpenter et al. (2018) for CUSSO). Figures 3.15, 3.16, and 3.18 show that $A_{0,H/V}$ is generally greater than the amplifications predicted by 1-D theory: $A_{0,H/V}$ exceeds $A_{0,FR}$ by more than 20 percent at all but three sites and by more than 40 percent at all but four sites. The greatest exception to this is station SCMO, for which Williams et al. (2007) report a very high bedrock S-wave velocity of 3,800 m/s, where $A_{0,FR}$ exceeds $A_{0,H/V}$ by 44 percent. In addition, $A_{0,FR}$ exceeds $A_{0,H/V}$ by 22 percent at HEKY. Also, the stations with the largest $A_{0,H/V}$, EVIN and MVKY, have the greatest differences between observations and predictions.

Figure 3.18 Comparisons of fundamental-mode S-wave H/V peaks with those from full-resonance calculations (a) and those estimated using equation 3.1 (b). The best-fit lines with 0 y-intercepts and through the points circled in red are shown and labeled by their corresponding equations. HEKY and SCMO appear to be outliers and were not included in the regressions.
3.5 DISCUSSION

3.5.1 SIMPLIFIED VERSUS FULL-RESONANCE \( f_0 \) and \( A_0 \)

The comparisons in Figure 3.15 indicate that equations 3.1 and 3.2 approximate the more realistic full-resonance responses at most stations. However, at deep-soil sites such as CUSSO, where \( f_{\text{peak}} \) occurs at a frequency higher than \( f_0 \), a full-resonance analysis may be required to accurately reflect the site response.

Also, at the sites with strong impedance contrasts between thick layers above the base of the sites’ measured velocity profiles, \( \tilde{A}_0 \) and \( \tilde{f}_0 \) can underestimate both \( A_{0,FR} \) and \( f_{0,FR} \). As Figure 3.1 illustrates and as discussed for stations USIN and WVIL, weathered bedrock, overlying intact or unweathered bedrock, can affect seismic-wave propagation and site response; see Hashash et al. (2014) for a discussion of weathered versus intact rocks as reference sites in central and eastern North America. Figure 3.17 shows that the two soil layers (FR\(_{2\text{lyr}}\)) over weathered bedrock account for most (84 percent) of the magnitude of the FR response from all layers in the measured velocity profile at USIN, but yield a 23 percent larger \( f_0 \). Empirical \( f_0 \) measurements, from the S-wave H/V curve and as observed in other studies (Fig. 3.15), correspond better with \( f_0 \) determined by the FR response from all layers than that predicted by the shallower layer, which suggests that the weathered bedrock should be included in site-response calculations at this and similar sites. Therefore, accounting for large impedance contrasts that affect site response within the frequency band of engineering interest is important, regardless of lithology.
Figure 3.15 shows that $f_{0,H/V}$ provides first-order approximations to $f_{0,FR}$ at most sites. The observations from other studies provide insights on the $f_{0,H/V}$ comparisons with $f_{0,FR}$ (Fig. 3.15). For example, at station USIN, $f_{0,H/V}$ is consistent with $f_{peak}$ observations from both earthquakes (Odum et al., 2010) and ambient noise (McNamara et al., 2015). The agreement between these empirical observations and $f_{0,H/V}$ in this study suggests that inaccuracies in the velocity structure or 3-D effects are the cause of the 27 percent difference between $f_{0,H/V}$ and $f_{0,FR}$ at this site. At WVIL, the two largest S-wave H/V peaks occurred at 7.0 and 9.0 Hz, with the latter being the largest (Fig. 3.15). These peak frequencies are remarkably consistent with the observations of Odum et al. (2010) from the 2008 Mount Carmel earthquake (the two largest peaks are at 7.7 and 9.2 Hz, the former being the largest by a nominal amount). Resolving whether the discrepancies between the empirical observations—both the observations in this study and those of Odum et al. (2010)—and predicted site responses are caused by an inaccurate velocity structure or the effects of the nearby geologic structure, or both, will require more detailed 3-D site characterization and perhaps 3-D site-response modeling. But the similarity between $f_{peak}$ observed by Odum et al. (2010) and $f_{0,H/V}$ in this study supports the reliability of S-wave H/V to capture the empirical site response, even perhaps the responses caused by faults or other 3-D structures. WVIL and HAIL coincide with the locations of large surface faults that extend into Precambrian basement—the New Harmony and Cottonwood Grove Fault Zones, respectively (Fig. 3.3) (Odum et al., 2010)—which could introduce or enhance 3-D propagation effects.
3.5.3 THEORETICAL AND EMPIRICAL $A_0$

Although $f_{0,H/V}$ approximates $f_{0,FR}$ at many sites, $A_{0,H/V}$ does not approximate $A_{0,FR}$ nearly as well. The differences in observed and theoretical amplifications could arise from a number of factors including 1-D velocity structure inaccuracies, which in turn include bedrock S-wave velocities; vertical motions in the S-wave window that influence the H/V ratios; the responses of 3-D or deeper, unmodeled structures; or some combination of those factors. However, there appears to be a positive correlation between $A_{0,H/V}$ and $A_{0,FR}$, as shown in Figure 3.18. Therefore, because the sediment-bedrock velocity ratio exerts the most control on fundamental mode amplification (Fig. 3.12), $A_{0,H/V}$ is likely related to the sediment-bedrock velocity contrast. Further, Figure 3.18 also shows that at most sites, $A_{0,H/V}$ is 1.7 times greater than $\tilde{A}_0$. This suggests the possibility of deriving a correction to $A_{0,H/V}$ measurements for use with estimating $\tilde{A}_0$.

The diffuse field theory interpretation of body-wave H/V (Kawase et al., 2011; Rong et al., 2017) provides candidate formulation for such a derivation. Equation 24 in Kawase et al. (2011) relates the vertical-incidence S-wave transfer function, $|TF_{Hv}|$, with H/V, assuming equipartitioning of the S-wave energy on both horizontal components, but includes the vertical-component transfer function of vertical-incidence P-waves, $|TF_{Pv}|$, and a constant of proportionality dependent on the bedrock P- ($\alpha_b$) and S-wave ($\beta_b$) velocities. Similar to the derivation in Rong et al. (2017), equation 24 in Kawase et al. (2011) for $|TF_{Hv}|$ can be expressed as:
\[ |TF_H| = \left( \frac{2\alpha_b}{\beta_b} \right)^{-\frac{1}{2}} |TF_V| \cdot \frac{H}{V}, \]  

which shows that for H/V to estimate the site transfer function, it needs to be reduced by a factor of \( \sqrt{\frac{2\alpha_b}{\beta_b}} \). Furthermore, H/V must be corrected by \( |TF_V| \). Applying equation 3.4 to S-wave H/V observations made in this study is not straightforward because transverse-component recordings, rather than both horizontal components, were used. However, because the stations used in this study and numerous seismic stations have recorded earthquake S-waves in the central and eastern U.S. since the EarthScope projects’ Transportable and Flexible Arrays, future investigations of the use of S-wave H/V to estimate \( A_0 \) using equation 3.4 are warranted.

### 3.5.4 SITE RESPONSE FROM DEEP LAYER(S) IN THE ILLINOIS BASIN

An additional factor that may contribute to differences between theoretical and empirical \( A_0 \) and \( f_0 \) is the potential for contributions to the site response from deeper rock layers, which may be observed in S-wave H/V ratios but which are not modeled in the theoretical responses. At each of the sites in the Illinois Basin—EVIN, HAIL, HEKY, OLIL, S46A, USIN, and WVIL—\( f_{0,FR} \) picks did not always coincide with the first prominent peak (compare \( A_{0,H/V,\text{(IB)}} \) with \( A_{0,H/V} \) in Fig. 3.15). The first S-wave H/V peak observed at these sites is broad, sometimes followed by a second peak of similar character (OLIL and WVIL), but generally with a peak magnitude of ~5. Furthermore, except for EVIN and HEKY where the \( A_{0,H/V} \) are greater than 5, the peak frequencies are not predicted by the shallow velocity structures shown in Figure 3.4. Spectral ratio peaks within the same
frequency band were observed at S46A and OLIL by Yassminh et al. (2019) and McNamara et al. (2015), respectively, using ambient-noise H/V.

These first peaks are highlighted in Figure 3.15 by shaded rectangles that delineate a 1-to-3 Hz frequency band. Because these peaks were seen at all Illinois Basin stations, it is possible that they result from at least one deep, large impedance contrast in the basin, but below the depth to fastest (lowest) bedrock in each site’s velocity structure. At station SCMO, adjacent to the basin, the first peak of the S-wave H/V curve has a similar, broad character as do those observed in the corresponding curves from basin stations. Because the corresponding frequency is inconsistent with theoretical \( f_0 \) predictions but the second peak frequency is consistent it seems plausible that, as with stations in the Illinois Basin, deeper structures (hundreds of meters deep) not included in the site velocity profile are responsible for the first H/V peak at this station.

Figure 3.19 adds evidence to these first S-wave H/V peaks being caused by resonances by comparing higher-modes predicted assuming SH-wave resonance (i.e., \( f_n = (2n+1) \cdot f_{0,H/V} \)) with H/V peaks at higher frequencies. For example, at EVIN the fundamental resonance of the soil layers may coincide with the fundamental resonance from the deeper basin structure. Therefore, these two resonances may superimpose, resulting in larger and broader \( A_{0,H/V} \) peaks at \( f_0 \) than those predicted by the FR response for the shallow layers alone. Also, at OLIL where the S-wave H/V has two peaks in the 1-to-3 Hz frequency range, \( f_0 \) and \( f_0' \), additional H/V peaks appear at frequencies corresponding to \( f_2, f_1', f_3, f_2', \) and \( f_3' \).
Figure 3.19  S-wave H/V and FR responses at stations that appear to have site-response effects from deeper, unmodeled layers within the 1 to 3 Hz frequency range. The first H/V peak, presumably the fundamental one, is highlighted in blue and the next three predicted modes are highlighted with the same color scheme in each plot. At stations OLIL and WVIL, the second of the two peaks that occur within the same frequency range are highlighted by vertical dashed lines. Manually picked H/V peaks that correspond with the highlighted higher modes are indicated with diamonds.

The profound effect of such an intrabedrock layer on site response was postulated by Laurendeau et al. (2017) to explain large low-frequency responses in the Quito Basin of Ecuador and in the central and eastern U.S. by Yassmin et al. (2019), including in the Illinois Basin, and in Oklahoma by Mendoza and Hartzell (2019). This hypothesis was tested in this setting by first comparing the H/V curves in Figure 3.14 with those at four other seismic stations in or near the Illinois Basin (circled stations in Figure 3.3), as shown in Figure 3.20. The only station that shows a distinct peak in the 1-to-3 Hz band is T47A, which is within the basin boundary. Stations near the boundary—SLM, WCI, and BLO—do not show this response peak, perhaps because of thinning, shallowing, or disappearance of the causal layer or layers near the edge of the basin.
Figure 3.20 S-wave H/V curves for candidate reference stations in or just outside of the Illinois Basin, shown in Figure 3.3.

These S-wave H/V peaks were compared with peaks determined using standard spectral ratios. The flat S-wave H/V curve for station SLM indicates that this site is a suitable reference site (Field and Jacob, 1995); this was also observed by Williams et al. (2007) from recordings of a local earthquake. Stations BLO and WCI likewise have little to no response in the 1-to-3 Hz band, but there were not enough high-quality S-wave recordings to use these stations as reference sites. Because SLM is quite far from the nearest Illinois Basin station (185 km from OLIL), used shear-wave recordings from teleseismic earthquakes were used to minimize differences in the path and source effects similar to Pratt (2018). The recordings of all magnitude 5 and greater earthquakes with focal depths of 100 km and greater, and at select offset ranges between 20 and 90°, corresponding to offsets at which the predicted S-wave arrivals did not coincide with P-phases or surface waves within a ±1-min window were collected. Smoothed (0.5-Hz running Hanning windows) amplitude spectra from 25-s windows on the instrument-corrected transverse-component recordings starting before the S-wave arrival were calculated and mean spectral ratios at three stations in the Illinois Basin—OLIL, T47A, and USIN—were formed by
spectral division, using SLM as the reference site. As with the S-wave H/V processing, S-wave recordings at both the basin and reference sites were required to have SNRs of 2.5 or greater. Because of the limited number of usable pairs of recordings, only two recordings at each frequency were required to calculate the mean.

Although only 12 earthquakes were used and the achievable frequency band was limited to ~3.0 Hz and below, the resulting spectral ratios contain peaks within the 1 to 3 Hz band (Fig. 3.21), supporting the hypothesis that the S-wave H/V peaks result from a deep geologic interface. The interface is possibly at least 250 m beneath the surface, based on equation 3.2 and an $f_0$ of ~2 Hz, assuming an average shear-wave velocity of 2,000 m/s for the layers to that depth and that near-vertical S-wave resonance is the cause of the peak.
Figure 3.21 Standard spectral ratios at three Illinois Basin sites with respect to reference site SLM from teleseismic S-wave recordings, and corresponding local- and regional-earthquake S-wave H/V curves. The range of the number of teleseismic observations used to calculate the mean SSR at each frequency is labeled on each plot.

The ostensible correspondence of resultant higher-mode resonance peaks in the S-wave H/V curves (Fig. 3.19) complicates the comparisons of S-wave H/V $f_0$ and $A_0$ with those from the FR responses. However, Figures 3.15 and 3.16 suggest the effect on $f_{0,H/V}$ is not significant. Furthermore, the nearly constant ratio of $A_{0,H/V}$ to $A_{0,FR}$ at almost all sites, both those within and outside of the Illinois Basin, suggests that the higher modes also do not greatly effect $A_{0,H/V}$. 
CHAPTER 4. SEISMICITY IN AND AROUND THE ROME TROUGH, EASTERN KENTUCKY

4.1 INTRODUCTION

The central and eastern United States has experienced a significant increase in earthquakes induced by hydrocarbon-related hydraulic fracturing or wastewater injection during the past decade (Fig. 1.6; Ellsworth, 2013; Brudzinski and Kozłowska, 2019). The principal cause of these events has been assigned to the injection of wastewater into subsurface formations (Horton, 2012; Keranen et al., 2013; Hornbach et al., 2015). Hydraulic fracture stimulation of oil and gas reservoirs, or fracking, has also induced felt earthquakes (Holland, 2013; Skoumal et al., 2015; Bao and Eaton, 2016). Most cases of induced, felt earthquakes were the result of fluid injection into formations that are in hydraulic communication with the crystalline basement, which can lead to the rupture of preexisting, critically stressed basement faults (Zoback et al., 2002).

Consequently, soon after the first deep Rogersville Shale test well was completed in the Rome Trough in eastern Kentucky (May 2014), a temporary seismic network was deployed to monitor the most likely areas of oil and gas production within the Rogersville Shale and the area around clusters of existing wastewater-injection wells. The seismicity observations from the first three years and three months of network operation (June 2015 through October 2018) are the focus of this chapter.
4.2 PROJECT SETTING

4.2.1 THE ROME TROUGH AND THE ROGERSVILLE SHALE

The study area is in a stable continental region of the central Appalachian foreland basin of North America. It is underlain by a series of grabens that are collectively part of a more extensive Early and Middle Cambrian interior failed rift system within the Eastern Granite-Rhyolite province basement, and are associated with the breakup of the supercontinent Rodinia (Gao et al., 2000; Hickman, 2011). The Rome Trough, a northeast-trending graben system extending from eastern Kentucky northeastward across West Virginia and Pennsylvania into southern New York (Harris et al., 2004; Hickman, 2011), is part of this larger failed rift system and contains a thick sequence of Cambrian sedimentary rocks. The Rogersville Shale is one of six formations recognized within the Middle and Late Cambrian Conasauga Group and the only one that shows evidence of greater than 1 percent total organic carbon content (Hickman et al., 2015), generally accepted as the minimum needed to generate oil or gas.

The boundary faults of the westernmost Rome Trough trend southward into Tennessee and bound a zone of thickened crust (Yang et al., 2017) and a prominent Bouguer gravity anomaly known as the East Continent Gravity High, a high-density, mafic body extending from near the surface to mid- (Keller et al., 1982; Powell et al., 2014) to lower- (Mayhew et al., 1982) crustal depths. In West Virginia, the Rome Trough takes a ~25° bend, trending more north-northeasterly than in eastern Kentucky. This study focused on the length of the eastern Kentucky Rome Trough, and its adjacent crust to the northwest and southeast (boxed region in Fig. 4.1), that trends nearly linearly at N65°E, and that hosts the Rogersville Shale being tested. This section of the Rome Trough experienced
transtensional stresses during Cambrian rifting, and therefore underwent both strike-parallel and extensional strain (Hickman, 2011). The current maximum horizontal regional stress in this section of the trough is compressive and oriented predominantly southwest-northeast, at N63°E from the average of stress measurements in the Rome Trough (in Heidbach et al., 2016).
Figure 4.1 Precambrian (pC) faults, the Rome Trough boundary (heavy, dashed black lines), area of possible Rogersville Shale production (gray, shaded), deep Rogersville Shale test wells (Deep OG Well), wastewater-disposal wells (SWD well), -30 mGal and greater Bouguer gravity anomalies (ECGH is the East Continent Gravity High), and seismicity from 1900 through 1980 (open circles) and from 1980 to prior to the temporary network (filled circles). Orientations of quality C and greater maximum horizontal stress \( S_{Hmax} \) measurements from the World Stress Map (Heidbach et al., 2016) are also shown. Blue line is the location of the cross section shown in Figure 4.2. The area within the box is the study area.
4.2.2 SEISMICITY

Seismicity in the project area is characterized by a diffuse distribution of small-magnitude events outside of the Rome Trough of eastern Kentucky (Fig. 4.1). Noteworthy larger, recorded earthquakes in the region are the 1980 Mw 5.0 Sharpsburg, Ky., earthquake (Herrmann et al., 1982) to the north of the Rome Trough and the 2012 Mw 4.2 Perry County earthquake to the south, which is in the northern part of the Eastern Tennessee Seismic Zone (Carpenter et al., 2014). The ETSZ produces the second highest moment release from natural earthquakes in the central and eastern U.S. (Powell et al., 1994). Because of the high seismicity rates and its capability to produce $M \geq 6$ earthquakes (Warrell et al., 2017), the ETSZ has been the focus of numerous seismological, geodynamical, and geomorphological studies (e.g.; Chapman et al., 1997; Levandowski and Powell, 2018; Gallen and Thigpen, 2018). In contrast, the crust beneath the EKRT, only two earthquakes since 1900 are reported in the ANSS Comprehensive Catalog (ANSS ComCat; earthquake.usgs.gov/earthquakes/search; last accessed 09/20/2019); Fig. 4.1).

4.2.3 INDUSTRIAL ACTIVITY: SUBSURFACE FLUID INJECTION

Class II wastewater-injection wells have operated in the Rome Trough of eastern Kentucky since at least 1997 (Fig. 4.1; Carpenter et al., 2019). However, the injection rates and total volumes of injected wastewater are modest compared with wells associated with induced earthquakes (Fig. 4.2): the median injection rate of these wells in the study area is 1,500 barrels/month; their median cumulative injection volume is 65,600 barrels. More importantly, the injection intervals are shallow and typically several km above basement
(Fig. 4.3). Therefore, inducing earthquakes from wastewater injection in this region is unlikely.

Figure 4.2 Cumulative volume (left) and maximum monthly injection rate (right), in log units, from all wastewater disposal wells with reported injection volumes in and near the Rome Trough of eastern Kentucky reported in Carpenter et al. (2019).

Figure 4.3 Well-based cross section based on drillers’ logs, with generalized stratigraphic groups, across the Rome Trough through the study area in eastern Kentucky and southwestern West Virginia (Fig. 4.1). Areas marked in red refer to depths of waste-injection targets within 30 km of the section. The Rogersville Shale, the target horizon for deep fracking, is outlined in green. Figure courtesy of John B. Hickman, Kentucky Geological Survey.
In contrast, tests of the Rogersville Shale occur much deeper, generally ~1 km above the Precambrian surface (Fig. 4.3). Therefore, due to its proximity to basement, there is a risk of inducing earthquakes from hydraulic fracturing this formation (e.g. Skoumal et al., 2018). The first Rogersville well in the project area was completed in 2014; the second was completed in June 2015, four days after the installation of the first monitoring station. The remaining three wells in Figure 4.1 were completed between October 2015 and January 2017.

4.3 DATA AND METHODS

4.3.1 TEMPORARY MONITORING NETWORK

Table 4.1 lists long-term seismic stations operating in the vicinity of the EKRT. The estimated detection thresholds of these stations range from moment magnitude (Mw) 1.5 in the southwestern EKRT to Mw 1.9 in the northeastern EKRT (Carpenter et al., 2019). With the addition of EarthScope (Central and Eastern United States Network) stations, which were operational during the time of the monitoring project, the detection thresholds range from Mw 1.5 to 1.7 in the same areas, respectively.
Table 4.1 Long-term and temporary (stations with Off Date attributes) seismic stations used in real-time monitoring. EK, temporary Rome Trough stations; KY, University of Kentucky; N4, Central and Eastern United States Network; OH, Ohio Seismic Network; US, ANSS-NEIC network; XO, OIINK EarthScope Flexible Array experiment (Yang et al., 2017); Off Date, Date of uninstallation. No date indicates that the station was operational through the end of the analysis period.

<table>
<thead>
<tr>
<th>Station</th>
<th>Network</th>
<th>Latitude (°N)</th>
<th>Longitude (°E)</th>
<th>Off Date</th>
</tr>
</thead>
<tbody>
<tr>
<td>BHKY</td>
<td>KY</td>
<td>38.0344</td>
<td>-84.5032</td>
<td></td>
</tr>
<tr>
<td>FLKY</td>
<td>KY</td>
<td>38.4261</td>
<td>-83.7506</td>
<td></td>
</tr>
<tr>
<td>HZKY</td>
<td>KY</td>
<td>37.2511</td>
<td>-83.2067</td>
<td></td>
</tr>
<tr>
<td>PKKY</td>
<td>KY</td>
<td>38.3830</td>
<td>-83.0341</td>
<td></td>
</tr>
<tr>
<td>ROKY</td>
<td>KY</td>
<td>37.9091</td>
<td>-83.9257</td>
<td></td>
</tr>
<tr>
<td>P51A</td>
<td>N4</td>
<td>39.4818</td>
<td>-83.0601</td>
<td></td>
</tr>
<tr>
<td>P53A</td>
<td>N4</td>
<td>39.4868</td>
<td>-81.3896</td>
<td></td>
</tr>
<tr>
<td>Q51A</td>
<td>N4</td>
<td>39.0260</td>
<td>-83.3456</td>
<td></td>
</tr>
<tr>
<td>Q52A</td>
<td>N4</td>
<td>38.9622</td>
<td>-82.2669</td>
<td></td>
</tr>
<tr>
<td>R49A</td>
<td>N4</td>
<td>38.2916</td>
<td>-85.1714</td>
<td></td>
</tr>
<tr>
<td>R50A</td>
<td>N4</td>
<td>38.2816</td>
<td>-84.3274</td>
<td></td>
</tr>
<tr>
<td>R53A</td>
<td>N4</td>
<td>38.3307</td>
<td>-81.9522</td>
<td></td>
</tr>
<tr>
<td>S51A</td>
<td>N4</td>
<td>37.6392</td>
<td>-83.5935</td>
<td></td>
</tr>
<tr>
<td>S54A</td>
<td>N4</td>
<td>37.7997</td>
<td>-81.3114</td>
<td></td>
</tr>
<tr>
<td>T50A</td>
<td>N4</td>
<td>37.0204</td>
<td>-84.8384</td>
<td></td>
</tr>
<tr>
<td>U54A</td>
<td>N4</td>
<td>36.5209</td>
<td>-81.8204</td>
<td></td>
</tr>
<tr>
<td>SSFO</td>
<td>OH</td>
<td>38.6953</td>
<td>-83.1972</td>
<td></td>
</tr>
<tr>
<td>TZTN</td>
<td>US</td>
<td>36.5439</td>
<td>-83.5490</td>
<td></td>
</tr>
<tr>
<td>KH50</td>
<td>XO</td>
<td>37.4170</td>
<td>-84.4633</td>
<td>Oct. 2015</td>
</tr>
<tr>
<td>KH54</td>
<td>XO</td>
<td>37.4149</td>
<td>-84.1600</td>
<td>Oct. 2015</td>
</tr>
<tr>
<td>KI51</td>
<td>XO</td>
<td>37.1857</td>
<td>-84.5075</td>
<td>Oct. 2015</td>
</tr>
<tr>
<td>KI53</td>
<td>XO</td>
<td>37.1845</td>
<td>-84.2061</td>
<td>Oct. 2015</td>
</tr>
<tr>
<td>KJ50</td>
<td>XO</td>
<td>37.0462</td>
<td>-84.5808</td>
<td>Oct. 2015</td>
</tr>
<tr>
<td>KJ52</td>
<td>XO</td>
<td>36.9186</td>
<td>-84.2500</td>
<td>Oct. 2015</td>
</tr>
<tr>
<td>KK50</td>
<td>XO</td>
<td>36.8694</td>
<td>-84.8024</td>
<td>Oct. 2015</td>
</tr>
</tbody>
</table>

To characterize seismicity at a lower magnitude threshold in the project area, a temporary network of 13 telemetered, broadband seismic stations (herein EK) was established in the EKRT (Fig. 4.4). Approximately half of the stations were installed in vaults and half in postholes and all seismometers were set on or within bedrock. EK station installations began in June 2015, and the final real-time station was installed in June 2016.
(Table 4.2). The average station spacing was 20 km, but the station distribution was slightly denser in the eastern part of the project area, where the Rogersville Shale is being tested and most likely to be produced.

Station installation locations satisfied multiple criteria needed for the successful, year-round operation of telemetered, autonomous, broadband seismographs including:

- southern exposure to the sky for year-round battery charging using solar panels
- sufficient cellular signal strength for data transmission
- seismometer burial depth to at least 1 m to avoid seismometer tilts from diurnal temperature fluctuations
- seismometer installation on or in bedrock or other original geologic materials (e.g., bedrock residuum)
- lateral separation from tall objects apt to induce ground motions under windy conditions, such trees and built structures, by at least the height of the object
- separation from sources of cultural noise such as frequently traveled roads, train tracks, power lines, pumps, and generators
- lack of visibility to prevent vandalism
- the identification of consenting landowners.
Figure 4.4 Seismic stations used for event detection (colored by network code), the detected earthquakes (colored by focal depth), and the ANSS ComCat during the same time period (orange) for comparison. Stations marked with an ‘*’ were used only in the analysis of detected events (additional stations used for analyses only are outside the map region and are not shown). Events within the box subdivided by line N-S are projected onto the cross section in Figure 4.10. The $\geq -30$ mGal Bouguer gravity anomalies are also shown.
Table 4.2 Temporary seismic monitoring station locations, seismometers (T-40, Trillium 40; TC-PH2, Trillium Compact Posthole; MC-PH1, Meridian Compact Posthole), and operational time period. Stations with no Off Date were operating at the time of preparation of this chapter.

<table>
<thead>
<tr>
<th>Station</th>
<th>Latitude (°N)</th>
<th>Longitude (°E)</th>
<th>Seismometer</th>
<th>Sample Rate (Hz)</th>
<th>On Date</th>
<th>Off Date</th>
</tr>
</thead>
<tbody>
<tr>
<td>EK12</td>
<td>38.1287</td>
<td>–83.1042</td>
<td>T-40</td>
<td>100</td>
<td>09/30/2015</td>
<td>5/30/2019</td>
</tr>
<tr>
<td>EK13</td>
<td>38.2301</td>
<td>–82.8286</td>
<td>T-40</td>
<td>100</td>
<td>10/18/2015</td>
<td>6/11/2019</td>
</tr>
<tr>
<td>EK14</td>
<td>38.2996</td>
<td>–82.7037</td>
<td>TC-PH2</td>
<td>200</td>
<td>06/03/2015</td>
<td></td>
</tr>
<tr>
<td>EK20</td>
<td>37.7332</td>
<td>–83.8661</td>
<td>T-40</td>
<td>100</td>
<td>11/06/2015</td>
<td>5/30/2019</td>
</tr>
<tr>
<td>EK21</td>
<td>37.8160</td>
<td>–83.5315</td>
<td>MC-PH1</td>
<td>200</td>
<td>09/02/2015</td>
<td>5/22/2019</td>
</tr>
<tr>
<td>EK22</td>
<td>37.9152</td>
<td>–83.2508</td>
<td>MC-PH1</td>
<td>200</td>
<td>08/25/2015</td>
<td></td>
</tr>
<tr>
<td>EK23</td>
<td>37.9213</td>
<td>–82.9004</td>
<td>T-40</td>
<td>100</td>
<td>09/29/2015</td>
<td>5/30/2019</td>
</tr>
<tr>
<td>EK25</td>
<td>38.1359</td>
<td>–82.8145</td>
<td>TC-PH2</td>
<td>200</td>
<td>06/10/2015</td>
<td></td>
</tr>
<tr>
<td>EK26</td>
<td>38.0704</td>
<td>–82.5810</td>
<td>TC-PH2</td>
<td>200</td>
<td>06/04/2015</td>
<td></td>
</tr>
<tr>
<td>EK32</td>
<td>37.6198</td>
<td>–83.3024</td>
<td>MC-PH1</td>
<td>200</td>
<td>09/02/2015</td>
<td>6/03/2019</td>
</tr>
<tr>
<td>EK33</td>
<td>37.7582</td>
<td>–83.1249</td>
<td>T-40</td>
<td>100</td>
<td>11/04/2015</td>
<td>6/02/2019</td>
</tr>
<tr>
<td>EK34</td>
<td>37.7056</td>
<td>–82.7496</td>
<td>T-40</td>
<td>100</td>
<td>06/06/2016</td>
<td>6/11/2019</td>
</tr>
<tr>
<td>EK35</td>
<td>37.8569</td>
<td>–82.7147</td>
<td>TC-PH2</td>
<td>200</td>
<td>06/09/2015</td>
<td>6/11/2019</td>
</tr>
</tbody>
</table>

Instrumentation for most of the network was purchased by the Kentucky Geological Survey, with support from the University of Kentucky Department of Earth and Environmental Sciences. Cimarex Energy Co. contributed six complete stations to the monitoring network, and Nanometrics, the manufacturer of the instruments used, contributed support for one station. Instrumentation in the microseismic monitoring network consists of broadband ground-motion sensors (corner periods of 40 s or lower and high-frequency cutoff of 85 Hz or higher) and 24-bit data loggers.

4.3.2 AUTOMATIC EVENT DETECTION

Telemetered waveforms from the EK network and existing regional stations in Table 4.1 were acquired and processed in near-real-time using Earthworm (http://www.isti.com/products/earthworm; last accessed Oct., 2019). Broadband
waveforms were band-pass filtered from 1 to 20 Hz—the maximum frequency in this passband coincided with the lowest Nyquist frequency of the stations used for detections (stations in the XO network in Table 4.1)—and arrival detection was conducted on filtered waveforms using conventional STA/LTA detectors. Earthworm’s CARLSUBTRIG coincidence-trigger algorithm was used for event triggering through the duration of the EK network deployment to associate vertical-component detections. In an attempt to reduce the amount of analyst time dedicated to the laborious task of earthquake-blast discrimination, ten months after initiating event triggering a grid-based association of vertical- and horizontal-component detections, determined from waveforms filtered with a slightly narrower passband of 4 to 20 Hz, was implemented using Earthworm’s BINDER.

Candidate arrival detections were declared for STA/LTA ratios ≥ 3.5 and events were declared via coincidence triggering when four detections arrived within a coincidence window of 35 s, which is the S-wave travel time from a surface-focus event across the maximum separation between EK stations. The detector was configured primarily to detect S-wave rather than P-wave arrivals because direct S-waves on average (assuming a well-sampled focal sphere) have similar or larger amplitudes than direct P-waves at offsets (Figs. 4.5 and 4.6). An event was declared by BINDER when at least four candidate P-wave or S-wave arrivals from at least three stations clustered within a time window of 2 s after removing the travel times from a candidate hypocenter to the corresponding stations.
Figure 4.5 Vertical-component seismograms from temporary network (station codes, labeling each trace, beginning with EK) and regional network stations from a local magnitude 1.4 earthquake 16.5 km deep in the crust beneath the Rome Trough of eastern Kentucky recorded at distances from 15 to 73 km from the hypocenter. The onsets of identifiable P- and S-phase arrivals are indicated with blue and red vertical lines, respectively.
Figure 4.6 Example vertical-component recordings (waveforms) of earthquakes beneath (a) or outside (b) the Rome Trough bandpass filtered with a passband of 1 to 20 Hz, and their corresponding STA/LTA characteristic functions. For both examples shown, although the P-wave arrivals were not detected at the STA/LTA threshold of 3.5, the S-wave arrival were. Plots are labeled by the station codes and hypocentral offsets (Δ), and the earthquakes’ magnitudes.

A minimum of four detections were required to declare an event from either triggering method to ensure that observations from enough stations were available to calculate event hypocenters. Both the STA/LTA ratio and four-detection thresholds were selected to balance reducing the number of false triggers with detecting and locating small earthquakes. Detected events were manually separated into the categories of local,
regional, or teleseismic and the local events were further categorized as earthquakes or probable mine or quarry blasts.

4.3.3 EARTHQUAKE ANALYSIS

Phase arrivals from local earthquakes were manually picked and events were located using SEISAN (Havskov and Ottemöller, 1999): P-wave arrivals and polarities were picked on vertical-component recordings and S-wave arrivals on transverse, horizontal components. Events were located using HYPOCENTER (Lienert and Havskov, 1995) and one of two regional velocity models (Fig. 4.7): the HAMBURG model (Herrmann and Ammon, 1997), modified to remove low-velocity zones for use with HYPOCENTER, and the model in Vlahovic et al. (1998) for the Eastern Tennessee Seismic Zone. Both models were tested for each event location, and the model producing the lowest root-mean-square arrival-time residuals was used to determine the final hypocenter.

Figure 4.7 Velocity models used to determine earthquake locations. Vp is P-wave velocity; Vs is S-wave velocity. H and C subscripts correspond to the modified HAMBURG (Herrmann and Ammon, 1997) and Vlahovic et al. (1998) velocity models, respectively.
The appropriateness of the velocity models used in this study area is demonstrated by the travel-time residuals (observed minus predicted travel-time) from all earthquakes within the project area in Figures 4.8. The scatter of the residuals increases with distance, which is expected because of reduced signal quality with distance as a result of attenuation effects along the travel path. This scatter results in increased errors in the arrival-time picks with distance. But the lack of a systematic offset or trend in the residuals with distance, indicated by the best-fitting line through the residuals (average residual is zero to two significant digits regardless of distance), shows that the velocity structures used for the locations are appropriate, with no systematic bias due to unmodeled velocity structures.

Figure 4.8 P- and S-wave traveltime residuals versus hypocentral distance for all earthquakes located in project area. The equation of the best-fitting line through these residuals is shown.

Also, double-couple focal mechanisms were calculated by FOCMEC (Snoke, 2003) for events with 10 or more P-wave first-motion polarities. A grid search was performed for trial strike, slip, and rake parameter values that were consistent with all
picked polarities. Because the number of polarities was modest—the maximum used for any event was 13—numerous focal mechanisms were consistent with a given set of polarity observations. Optimal mechanisms were constrained using direct-P-, SV- and SH-wave amplitude ratios SV/P, SH/P, and SV/SH, where P- and SV- amplitudes were measured on the vertical component and SH-amplitudes were measured on the transverse component. Because uncertainties with amplitude measurements are greater than those of the polarities, the fewest amplitude ratios needed to produce fewer than 100 fault-plane solutions with 2° grid spacing for strikes, dips, and rakes, consistent with polarity and amplitude-ratio observations, were used. The final, best-fitting solution is the one with the lowest RMS misfit between the observed and predicted amplitude ratios, and which fits all P-wave polarities.

Magnitudes were calculated for all located events using both an amplitude-based scale ($M_L$) and a signal-duration-based scale ($M_C$). The $M_C$ scale was calculated for events when triggered waveforms included the entire coda. Duration magnitudes were estimated using the relationship of Chapman et al. (2002):

$$M_c = -3.45 + 2.85 \log_{10}(D)$$

where D is the coda length in seconds. The median of the individual station values is reported for the event.

The $M_C$ scale is susceptible to producing variable magnitude estimates due to variations in site noise levels, particularly for small events. Therefore, amplitude-based local magnitudes were also calculated for each located event, which are more consistent with the energy-based moment-magnitude scale than duration magnitudes, based on
preliminary findings in Holcomb (2017). An $M_L$ scale was developed for the project area using amplitude measurements on both horizontal components for each station with clear S- or Lg-wave arrivals. Coefficients that account for attenuation in the region were calibrated by SEISAN’s inversion algorithm MAG2 so that an $M_L$ 3.0 earthquake produces a displacement of 1 mm at 100 km offset, consistent with the definition of local magnitude (Richter, 1935). Station-correction terms were also derived in this procedure.

4.4 RESULTS

4.4.1 SEISMICITY AND FOCAL MECHANISMS

The EK network and contributing regional stations detected 56,127 events from June 11, 2015, through Aug. 16, 2018. A total of 32,008, or 57 percent, of the triggers were identified as seismic events (Fig. 4.9); the remainder were triggered by transient noise sources. A total of 28,679 triggers from local mine blasts were recorded, which constitute 90 percent of the seismic events. Of the remaining triggers, 1,857 were from teleseismic earthquakes and 1,314 were from regional events. Less than 1 percent of the seismic events (160) were from local earthquakes. One of the local events was not detected by the Earthworm triggers, but rather by visual inspection of the waveforms. This $M_L$ 0.3 earthquake occurred on Feb. 2, 2017, and is the smallest event located in this study.
Within the approximately 2° by 2° box centered on the EK network and aligned with the EKRT’s N65°E trend in Figure 4.3, 56 earthquakes were located (Supplemental Table 4.1). Those 56 events are included on the cross section and depth histograms in Figure 4.10. A chief observation from the map and cross section views is a lack of seismicity in the crust beneath the EKRT compared to surrounding regions. Also, Figure 4.10 shows differences between the depth distributions north and south of the EKRT. To the north, most (56 percent) focal depths are shallower than 15 km, whereas to the south only 15 percent of the earthquakes occurred shallower than 15 km. Focal mechanisms for seven earthquakes in the crust outside and beneath the EKRT were also determined. The solutions that fit all the polarities and best-fit the amplitude ratios are plotted in Figure 4.11 and listed in Table 4.3. All solutions fitting the observations are plotted in Figure 4.12. These mechanisms show that predominantly oblique-slip faulting occurs in the project area. However, one mechanism north of the EKRT is predominantly thrust.
Table 4.3 Focal mechanism strikes, dips, and rakes of nodal plane one; the strikes and dips of the associated P- and T-axes; and the number of observations used to constrain the best-fitting solution: Evt, event number; Pol, the number of polarities; AR, the number of amplitude ratios.

<table>
<thead>
<tr>
<th>Evt</th>
<th>Date</th>
<th>Pol</th>
<th>AR</th>
<th>Strike (°)</th>
<th>Dip (°)</th>
<th>Rake (°)</th>
<th>P-strike (°)</th>
<th>P-dip (°)</th>
<th>T-strike (°)</th>
<th>T-dip (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>11/29/2015</td>
<td>12</td>
<td>5</td>
<td>298</td>
<td>44</td>
<td>44</td>
<td>239</td>
<td>10</td>
<td>133</td>
<td>58</td>
</tr>
<tr>
<td>2</td>
<td>12/22/2015</td>
<td>12</td>
<td>8</td>
<td>123</td>
<td>79</td>
<td>12</td>
<td>257</td>
<td>0</td>
<td>347</td>
<td>16</td>
</tr>
<tr>
<td>3</td>
<td>04/26/2016</td>
<td>11</td>
<td>7</td>
<td>117</td>
<td>51</td>
<td>56</td>
<td>230</td>
<td>1</td>
<td>322</td>
<td>64</td>
</tr>
<tr>
<td>4</td>
<td>08/06/2016</td>
<td>12</td>
<td>10</td>
<td>95</td>
<td>59</td>
<td>-20</td>
<td>59</td>
<td>35</td>
<td>323</td>
<td>9</td>
</tr>
<tr>
<td>5</td>
<td>09/22/2016</td>
<td>11</td>
<td>9</td>
<td>299</td>
<td>58</td>
<td>16</td>
<td>253</td>
<td>12</td>
<td>155</td>
<td>33</td>
</tr>
<tr>
<td>6</td>
<td>09/04/2017</td>
<td>11</td>
<td>11</td>
<td>112</td>
<td>55</td>
<td>-15</td>
<td>78</td>
<td>34</td>
<td>337</td>
<td>15</td>
</tr>
<tr>
<td>7</td>
<td>08/02/2018</td>
<td>10</td>
<td>8</td>
<td>303</td>
<td>39</td>
<td>33</td>
<td>251</td>
<td>18</td>
<td>137</td>
<td>52</td>
</tr>
</tbody>
</table>

Figure 4.10 Cross section view of seismicity projected from 105 km perpendicular to the either side of the section line N-S in Figure 3. Inverted, black triangles delineate the approximate boundaries of the Rome Trough. Red dashed lines show potential bounds of rifting-related faults through the crust. Error bars show the formal errors in focal depth calculated by the location algorithm. Depth histograms show the distribution of focal depths in the southern half (left) and northern half (right) of the section. Focal depths are with respect to mean sea level.
Figure 4.11 Lower-hemisphere focal mechanisms determined in this study. Additional mechanisms available in the literature are labeled by year of occurrence: 1980 Mw 5.0 Sharpsburg (Herrmann et al., 1982); 1988 Mw 4.1 Sharpsburg (Street et al., 1993); 2012 Mw 4.2 Perry County (Carpenter et al., 2014). Projection of P-axes onto the horizontal plane are shown as thick lines for focal mechanisms from this (EK; black) and other studies (blue). Orientations of maximum horizontal stress ($S_{Hmax}$) from the World Stress Map (Heidbach et al., 2016) are also shown. Inset histogram shows the distribution of P-axis and $S_{Hmax}$ trends; the red dot labeled RT plots the approximate trend of the EKRT (~65°N).
Figure 4.12 Lower hemisphere focal mechanisms, polarity observations, and direct P- and S-wave amplitude ratios. All solutions that fit the observations are shown in thin lines. The best-fitting solution, which minimized the misfit between the observed and predicted amplitude ratios, are bold. Numbers beneath each mechanism correspond to the Event numbers in Table 4.3.

4.4.2 MAGNITUDE SCALES AND ATTENUATION

The following calibrated $M_L$ scale was developed using 5,269 amplitudes measured from all located earthquakes:

$$M_L = \log_{10}(A) + 1.1911 \log_{10}(r) + 0.00008 r - 2.0717 + s,$$

where $A$ is the displacement amplitude in nanometers, $r$ is the hypocentral distance in kilometers, and $s$ is the station term. The final result, event $M_L$, is the median value of the horizontal-component magnitudes determined with equation 4.3. The terms in equation 4.3
that account for regional attenuation from geometric spreading and intrinsic attenuation, and a static correction accounting for the sensitivity of a standard Wood-Anderson seismometer and the calibration magnitude of 3.0, are:

\[-\log_{10}(A) = 1.1911 \log_{10}(r) + 0.00008 r - 2.0717.\]  \hspace{1cm} (4.3)

As shown in Figure 4.13, this attenuation correction is comparable with other attenuation correction functions derived for the nearby regions: Eastern Tennessee (Bockholt et al., 2015), and the eastern U.S. (Kim, 1998).

Magnitude residuals, i.e., event $M_L$ minus station $M_L$, are plotted versus hypocentral distance in Figure 4.13. Long wavelength variations, uncorrected by this log-linear scale, indicate that the actual attenuation contains complexities that this model cannot accommodate. S- and Lg-wave amplitudes increase at distances from approximately 80 km to approximately 200 km. This increase has been observed in other studies in the eastern U.S. (e.g., Burger et al., 1987) and has been attributed to the post-critical shear-wave reflections off the Moho. However, because the median of the station magnitudes measured over a range of offsets is reported as the event magnitude, and the bulk of the observations (first, second, and third quartiles are indicated in Figure 4.13) include both positive and negative $M_L$ residuals, systematic biases in the local magnitudes are not expected.
4.5 DISCUSSION

The spatial distribution of seismicity observed in this study is consistent with the long-term seismicity catalog for the region (compare Figures 4.1 and 4.4): earthquakes occurred mostly in the crust outside of the Rome Trough, and were distributed throughout the region, except for a dearth of activity around the East Continent Gravity High. There are no obvious lineations in the observed seismicity, in contrast with the observations of Chapman et al. (1997) and Powell and Thomas (2016) in the central Eastern Tennessee Seismic Zone where longer-term, relocated seismicity catalogs revealed clusters of seismicity trending predominantly northeast-southwest or east-west.

The seismicity rate in the 2° by 2° area centered on the EK network in Figure 4.4 is very low, but also spatially variable with different numbers of events north of (33), south of (20), and beneath (3) the EKRT (Fig. 4.10). The variable rate warrants determining
frequency-magnitude relationships separately for each of these sub-regions. However, the already modest number of events must be further reduced (by nine) to reflect the time period when the network sensitivity was relatively constant: the sensitivity changed significantly from June 2015 to June 2016 as stations were incrementally installed (the deployment time was stretched over such a long period due to instrument availability). Thus, there are too few events in the study area to calculate Gutenberg-Richter plots for each sub-region. Figure 4.14 shows the Gutenberg-Richter curves from earthquakes that occurred between July 2016 and October 2018 for the whole region (Fig. 4.4) and for the study area. These curves represent time- (daytime and nighttime) and area-weighted averages across these sub-regions with b-values of 1.1 and 1.3 for the whole region and study area, respectively. For comparison, Bockholt et al. (2015), estimated a much lower b-value of 0.93 for the Eastern Tennessee Seismic Zone. We also noted that the magnitude of completeness of the study area, estimated from the lowest magnitude of the linear part of the Gutenberg-Richter curve, of $M_L 1.5$ is much larger than expected based on the detection threshold maps, as discussed below, and higher than $M_L 1.3$ that Bockholt et al. (2015) estimated for eastern Tennessee using a coarser network of stations. Therefore, more earthquakes are needed to reduce b-value and completeness-magnitude uncertainties, which ideally would be determined from Gutenberg-Richter curves for sub-regions in the study area.
Figure 4.14 Gutenberg-Richter curves for the events located in this study when the entire temporary seismic network was operational (June 2016–October 2018) for the entire region in Figure 4.4 (all) and for those plotted on the cross section in Figure 4.10 (XS). The b-values for each curve are labeled.

The detection threshold maps (Fig. 4.15) indicate that the monitoring network reduced Mmin in the EKRT and that the network was most sensitive in the crust beneath the EKRT. Therefore the dearth of seismicity beneath the EKRT is almost certainly not an artifact related to insufficiently sensitive event detection. Although the maps, created for daytime and nighttime noise conditions (because multi-stage fracking completions can occur throughout the day), are useful for comparing the earthquake detection sensitivity in the study area before and during the EK network deployment, seismicity observations indicate that the maps may overestimate the actual sensitivity of the monitoring network. The calculated magnitudes of nearly one third of the events during both the daytime—five of 11 (ignoring the M_L 0.3 that was detected visually, and not by Earthworm)—and nighttime—four of 15—are less than the modeled Mmin at the corresponding latitudes and
longitudes. Nevertheless, more earthquake observations within the EKRT are needed for a thorough assessment of the utility of these maps.

![Figure 4.15 Modeled nighttime (left) and daytime (right) minimum magnitude detection thresholds in the project area from Holcomb (2017). Seismicity located in this study overlay the map corresponding to the event origin times. Epicenters are colored by whether or not the calculated magnitude is less than the modeled $M_{min}$. Seismic stations are colored by network code.](image)

Although several earthquakes occurred near the EKRT boundary faults, only three occurred more than 5 km from the faults, well within the crust that they bound in map view. Only one of these events occurred near—i.e., within 10 km—a subsurface injection well: its epicenter was 9 km from a Rogersville Shale test well and 2.5 km from a wastewater disposal well. However, this event’s focal depth and its timing with respect to the most recent injection activities at each well suggest the unlikelihood that this event was induced. The focal depth is well constrained by nearby P- and S-wave arrivals at 10.6±1.7 km, which is much deeper than the nearby Rogersville Shale test well’s stimulation depth of 3.5 km and the 0.3 km depth of the nearby wastewater disposal well’s injection interval (all depths are given with respect to the surface). Also this earthquake occurred five days after the
Rogersville Shale well’s completion finished and no injection volumes have been reported since 2007 for the disposal well (Carpenter et al., 2019).

This M$_L$ 0.3 earthquake was detected visually on digital helicorders and not by the real-time system. Visual inspection of helicorders occurred only occasionally for station performance assessment throughout the deployment, but daily during this Rogersville Shale well’s completion and for one week afterward. Its occurrence suggests that other undetected, small-magnitude events likely occurred beneath the EKRT. More sophisticated detectors (e.g., developed through machine learning techniques) may facilitate detecting these events form the continuous, archived waveforms. In addition, template matching (e.g., Skoumal et al., 2015) may enhance our event catalog both beneath and outside of the EKRT.

Figure 4.16 Focal depth versus root-mean-square traveltime misfit for the 02/20/2017 M$_L$ 0.3 earthquake, the only earthquake to have occurred within 5 km of a wastewater-injection well. The best-fitting focal depth is indicated by the star; the formal error in this depth determination is also shown by the vertical error bars.

To the south and to the north, the seismicity nearly truncates at the EKRT boundary. This contrast in seismic activity suggests a difference in the seismogenic potential of the faults beneath the Rome Trough compared to those in the surrounding crust. Hurd and
Zoback (2012) and Levandowski et al. (2018) observed that maximum horizontal stress, $S_{Hmax}$, orientations were consistent with the P-axes orientations from earthquake focal mechanisms in the central and eastern U.S., suggesting that P-axes estimate the orientation of $S_{Hmax}$ in the region. This consistency was also observed in this study, which provides a potential explanation for the very low rate of seismicity observed in the crust beneath the EKRT. As shown in Figure 4.11, P-axes are oriented within $\pm 15^\circ$ of the general trend of the Rome Trough of N65°E. If most faults large enough to produce detectable earthquakes within the EKRT trend subparallel to this general trend at seismogenic depths, these faults would not be favorably aligned for failure in this stress condition: faults striking parallel to $S_{Hmax}$ will not rupture. The single mechanism determined for an earthquake in the crust beneath the EKRT, however, indicates that not all of the faults within the trough are parallel to the trough.

If most faults beneath the EKRT at seismogenic depths are parallel to the trough’s trend, then a substantial contrast in the crustal fabric, in terms of fault orientations, exists between the crust beneath and outside of the EKRT. Also, if EKRT faults predominantly trend parallel to $S_{Hmax}$, then the EKRT may serve as a northern boundary to the Eastern Tennessee Seismic Zone. Figure 4.10 shows potential boundaries of the zone of extended crust with faults trending trough-parallel, based on the seismicity depth profile. This zone coincides with broad, positive receiver function amplitudes observed by Chen et al. (2018) in the crust beneath the EKRT that extend through the crust, and that are bound by negative amplitudes to the north and south of the trough’s boundary faults (see Figure 8 in their paper). It is possible that the Chen et al. (2018) receiver functions image the zone of predominantly EKRT-parallel faults.
To the west, seismicity patterns observed in this study and the historical catalogs suggest a similar abrupt truncation of the seismicity occurs near the East Continent Gravity High. This dense mafic body, most likely consisting of rift-related volcanic rocks (Keller et al., 1982; Powell et al., 2014), may be less seismogenic because it lacks the inherited faults extant in the surrounding eastern Granite-Rhyolite crust, or because it has not experienced extensive deformation since emplacement, which was the case in other eastern North American seismic zones (Thomas and Powell, 2017).

As within the EKRT, seismicity in the Rome Trough east of the boxed region in Figure 4.4 is rare. However, earthquakes have historically been more frequent in this section of the Rome Trough compared to the EKRT (Fig. 4.1). The trend of this section of the trough is ~25° more northerly than the EKRT, which would place the faults in the crust beneath at an oblique angle with respect to the orientation of $S_{\text{Hmax}}$, and would make them more inclined to failure than in the EKRT.

Although determining the mechanisms responsible for the seismic quiescence in the EKRT is outside the scope of this study, the observations reported herein have practical implications related to seismic hazard, and warrant additional investigation. Principally, the Eastern Tennessee Seismic Zone, which extends into the project area, is capable of producing $M \geq 6$ earthquakes. Understanding the probable spatial extent of damaging ETSZ earthquakes could help reduce uncertainties in regional seismic hazard analyses.
CHAPTER 5. CONCLUSIONS

5.1 SITE RESPONSE FROM DEEP BOREHOLES IN THE NEW MADRID SEISMIC ZONE

Weak-motion S-wave recordings at the two deep vertical seismic arrays in the northern Mississippi Embayment, VSAP and CUSSO, were used to estimate site responses using the spectral ratio method. The maximum observed amplification factors from the mean empirical SH-wave transfer functions are $8.5 \pm 6.2$ at $12.9 \text{ Hz}$ at VSAP and $15.0 \pm 4.8$ at $1.3 \text{ Hz}$ at CUSSO. Comparing the spectral ratios with Thomson-Haskell propagator matrices reveals that, although only 10 S-wave recordings at each array were suitable for analysis, the frequencies of the theoretical site response peaks were consistent with those from observed SH-wave surface-to-bedrock spectral ratios, $TF_T$, from local and regional earthquakes, thus indicating that $TF_T$ represents an empirical SH-wave transfer function for weak-motions. Theoretical and observed amplifications were also comparable, which indicates the appropriateness of 1-D site-response modeling at these sites, but the theoretical levels of amplification at CUSSO are provisional because the bedrock S-wave velocity is uncertain.

$TF_T$ curves were also used to evaluate the appropriateness of surface S-wave H/V, $HV_S$, to estimate the empirical site transfer function. The observed $HV_S$ curves are similar to the $TF_T$ spectral ratios at frequencies below approximately the fifth natural frequency at each site, indicating that the $HV_S$ curves can be used as single-station, empirical approximations of the S-wave transfer functions for low-frequency analyses. For higher frequencies, vertical-component amplifications of incident SV-waves, and the converted P- and SV-wave system, reduce $HV_S$, and cause it to deviate from observed SH-wave
amplification at both VSAP and CUSSO. Therefore, the applicability of $H_V$ to approximate $T_F_T$ is site-specific and depends on a site’s vertical-component transfer function.

Finally, H/V curves from ambient-noise recordings, $H_{V,noise}$, imply amplification levels that are consistent with those indicated by the observed and theoretical SH-wave transfer functions. However, $H_{V,noise}$ curves at both sites decreases rapidly with frequency, and do not contain important peaks in the SH-wave transfer functions at either site. Most importantly, $H_{V,noise}$ fails to reveal the frequencies at which the maximum amplifications occur in the frequency band of engineering interest (i.e., from 0.1 to 10 Hz) and the corresponding amplification levels; the largest amplifications observed by the S-wave spectral ratios occur at resonances higher than the sites’ fundamental frequencies. Therefore, it appears that ambient noise H/V cannot be used for detailed site-response analyses in the northern Mississippi Embayment.

5.2 CENTRAL AND EASTERN U.S. PRIMARY SITE-RESPONSE PARAMETERS

The wave propagation-based site-response parameters fundamental frequency, $f_0$, and fundamental-mode SH-wave amplification, $A_0$, were evaluated as alternatives to the Vs30-based site factors to estimate primary linear site-response characteristics at 11 seismic stations in the central and eastern U.S. $A_0$ and $f_0$ calculated from realistic, full-resonance site-response calculations, $A_{0,FR}$ and $f_{0,FR}$, respectively, were compared with simplifications ($\bar{A}_0$ and $\bar{f}_0$ respectively) calculated from a single soil layer over an elastic bedrock half-space, where the soil layer is assigned the average of the dynamic properties from each layer in the site models, and where layer densities and damping ratios were
estimated from site shear-wave velocity structures. Also, empirical estimates of $A_0$ and $f_0$
from earthquake S-wave H/V ratios, $A_{0,H/V}$ and $f_{0,H/V}$, respectively, were compared with the
theoretical full-resonance response calculations.

$\tilde{A}_0$ and $\tilde{f}_0$ generally approximate $A_{0,FR}$ and $f_{0,FR}$ at most sites. In particular, $\tilde{A}_0$ was
within 20 percent of $A_{0,FR}$ at all but two sites, USIN and WVIL. These two sites have at
least one large impedance contrast above the base of the sites’ measured velocity profiles,
and the resultant FR response is too complicated to be adequately approximated by
simplified response factors calculated from one layer over a half-space. Although the
differences between the $\tilde{f}_0$ and $f_{0,FR}$ exceed 20 percent at all but five sites, Figure 3.16a
shows that equation 3.2 provides first-order approximations of $f_0$ compared with FR
calculations.

S-wave H/V was found to provide first-order approximations of $f_{0,FR}$ at all but two
sites. Furthermore, $A_{0,H/V}$ correlates positively with $A_{0,FR}$, but $A_{0,H/V}$ is generally larger. The
strong, positive correlation of $A_{0,H/V}$ with $\tilde{A}_0$ indicates that $A_{0,H/V}$ is controlled by the
sediment-bedrock impedance ratio and damping ratio, but generally is a factor of 1.7 times
greater than $\tilde{A}_0$, suggesting the possibility of deriving a correction factor to estimate $\tilde{A}_0$
from $A_{0,H/V}$.

Potential primary sources of the differences between the observed and theoretical
amplifications are: (1) the 1-D wave propagation assumption, because some stations are in
river valleys or basins and some are in or near fault zones. (2) Velocity model uncertainties,
including S-wave bedrock velocities, which exert the greatest control on $A_0$. For example,
a 30 percent error in the bedrock shear-wave velocity would result in an approximately 30 percent error in the predicted \( A_0 \), with the largest errors resulting from underestimating the velocity. And (3) the estimated uncertainties in the densities and damping ratios predicted using statistical relationships dependent on shear-wave velocity structure (up to 10 percent and 16 percent, respectively).

The \( f_{0,H/V} \) measurements are consistent with \( f_0 \) observations at the same stations in other studies and using different datasets and methods; observations from the methods that used earthquake recordings (Odum et al., 2010; Yassmin et al., 2019) were the most consistent with ours. This consistency suggests that empirical estimations of site fundamental frequencies, including S-wave H/V, provide valuable observations for quantifying site response, and are not limited by velocity-structure inaccuracies or the 1-D wave propagation assumption.

H/V observations at relatively low frequencies—between 1 and 3 Hz—in the Illinois Basin also indicate that at least one large impedance contrast resides likely hundreds of meters beneath the base of the sites’ measured velocity profiles. The S-wave H/V observations were supported by standard spectral ratios using a reference site adjacent to the basin and S-wave recordings from teleseismic earthquakes. Of importance, at all sites but one (HAIL), their apparent amplifications measured from the H/V curves have similar (S46A, USIN, and WVIL) or greater (OLIL) magnitudes compared to the peaks related to the shallow layers included in the velocity structures. The influence of these layers on the site responses at Illinois Basin stations EVIN and HEKY is unclear, because resultant peak frequencies coincide with the base-mode resonance frequencies of the soil columns at these sites.
These deep layers were not included in any of the site velocity profiles, but appear to produce resonances within the frequency band of engineering concern, and may be sources of unmodeled earthquake site-effect hazard in the event of a strong earthquake in the region. These observations, along with the consistencies between the S-wave H/V observations and those of other studies using earthquake data, highlight the importance of collecting empirical site-response estimates, as well as the need for developing velocity models that extend deeper into the underlying rock layers, particularly through weathered layers, as recommended in Hashash et al. (2014). This investigation suggests that empirical S-wave H/V curves can identify sites where such deep velocity investigations need to be conducted.

5.3 SEISMICITY IN AND AROUND THE ROME TROUGH

The Rogersville Shale is a deep formation in the Rome Trough of eastern Kentucky that is being tested for oil and gas production. Because of the formation’s close proximity to the crystalline basement (~1 km above) and its location within the faulted Rome Trough, however, there is a potential for inducing earthquakes when using hydraulic fracturing to stimulate well production in this shale.

To characterize natural seismicity in the areas where the Rogersville Shale is most likely to be tested or produced and near active wastewater disposal wells, a temporary network of seismic stations was installed in the Rome Trough of eastern Kentucky. This network improved the monitoring sensitivity in the vicinity of wastewater-injection wells and deep oil and gas wells testing the Rogersville Shale by an estimated 0.3 to 0.8 magnitude units.
In the first 39 months of recordings, 160 local earthquakes were detected. A local magnitude scale was developed from S- and Lg-wave arrivals. Only three earthquakes occurred well within in the crust beneath the eastern Kentucky Rome Trough—i.e., more than 5 km from the boundary faults in map view—and none appears to be associated with the deep Rogersville Shale test wells or with wastewater-injection wells.

P-axes from seven focal mechanism determined in the project area are consistent with the mechanisms from larger earthquakes determined both north and south of the EKRT. The P-axes are also consistent with both the orientations of nearby maximum horizontal compressive stress measurements and with the trend of the EKRT. Therefore, if most seismogenic faults of sufficient size to produce detectable earthquakes in the crust beneath the EKRT trend subparallel to the trough’s boundary faults, then they are not favorably oriented for failure in the current stress field. This provides a testable explanation for the seismic quiescence in the crust beneath the EKRT and indicates that the EKRT may be the northern boundary of the Eastern Tennessee Seismic Zone.
REFERENCES


Lee, V.W., and Trifunac, M.D., 2010. Should average shear-wave velocity in the top 30 m of soil be used to describe seismic amplification? Soil Dynamics and Earthquake Engineering, 30(11), pp.1250–1258.


VITA

Education:
M.S. in Geophysics, Oregon State University, Dec. 2010.
B.S. in Physics; Concentration in Applied Mathematics, Davidson College, May 1999.

Professional Experience:
Graduate Research Assistant, Oregon State University, Corvallis, Or., Apr. 2004 – Dec. 2006
Physics Teacher, Episcopal School of Acadiana, Cade, La., Aug. 1999 – May 2003

Professional Publications:


Based Nuclear Explosion Monitoring Technologies, NNSA, Air Force Research Laboratory, p. 166-175.

