FINAL TECHNICAL REPORT

GROUND-MOTION SITE EFFECTS IN THE WABASH VALLEY REGION FROM THE 18 APRIL 2008 MT. CARMEL, ILLINOIS EARTHQUAKE AND AFTERSHOCKS

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Principal Investigator: Edward W. Woolery

Department of Earth and Environmental Sciences
University of Kentucky
101 Slone Research Building
Lexington, KY 40506-0053
Phone: 859.257.3016
Fax: 859.257.1147
Email: woolery@uky.edu
http://www.as.uky.edu/EES/

Co-PI: Ron Street

13813 Werth Rd.
Hermosa, SD 57744
Email: bhrstreet@yahoo.com

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Edward W. Woolery, *Dept. of Earth and Environmental Sciences, University of Kentucky, 101 Slone Bldg., Lexington, Ky. 40506-0053, Tel: 859.257.3016, Email: woolery@uky.edu*

ABSTRACT

The peak particle velocities and response spectra (>2 Hz) for the 2008 earthquake sequence, as well as previous earthquake observations in the area (beginning with the 10 June 1987 southwestern Illinois event), were evaluated in order to quantify the practical reduction in variability of linear ground motions using conventional site response investigations. Specifically, the subsurface geometry and S-wave velocity models for forty-two sites were defined using refraction/reflection soundings; these data allowed us to approximate the associated linear sediment transfer function with a one-dimensional site response algorithm. Horizontal-to-vertical ambient noise measurements were also used as an alternate method for estimating the transfer function. Results indicate that the corrections reduced the range of spectral amplitude for frequencies greater than 2.5 Hz between 40 and 70 percent, as well as the spectral variation by approximately a factor of 4. In addition, the data suggest that a peak ground velocity of 1.2 cm/s defines a clear boundary separating Modified Mercalli intensities IV and V. These observations can be useful in scaling ground motions of historical seismicity, as well as predicting the effects of future events. We speculate these quantitative characteristics are likely representative for site effects throughout the lower Wabash River valley, except for the infrequent thick-sediment filled sites (> 30 m). This representative area includes southwestern Illinois, southeastern Indiana, and the adjoining area in Kentucky.

NONTECHNICAL SUMMARY

The southeastern Illinois earthquake sequence of April 2008 is the latest event, beginning with the 10 June 1987 **M**5.0 earthquake, during the past twenty years that have provided over 200 high-quality earthquake records from 42 blast monitors in the region of southern Indiana and Illinois, and north-central Kentucky. The resulting free-field records have provided a unique opportunity to assess the role of seismic hazard amplification in this area of the central United States. The assessment yielded quantitative ground-motion site characteristics and variation that will be useful for scaling historical seismicity and predicting the effects of future earthquakes in the Wabash Valley area of the central United States.

INTRODUCTION

It is well accepted that local site conditions can strongly affect the amplitude, frequency content, and duration of ground motions resulting from an earthquake. Factors contributing to site effects include a material's elastic properties, thicknesses and impedance contrasts within the sediment overburden and at the sediment/bedrock interface, surface topography, sediment/bedrock interface geometry (i.e. horizontal, irregular, dipping, etc.), ground motion amplitude (i.e. linear vs. nonlinear), and the existence of lateral and/or vertical velocity gradients in the sediment and/or bedrock. As a result, accurately quantifying the site effect is problematic. Even the most comprehensive site investigation can fail to fully account for all of the complex subtleties that affect the local ground motions (Bommer and Abrahamson, 2006). Many authors, including ourselves, have used vague terms like "profound" or "significant" to describe site condition effects; but what specifically does that mean for sites in the central United States?

Beginning with the **M**5.0 southeastern Illinois earthquake of 10 June 1987 and continuing through the southeastern Illinois earthquake sequence of 2008, over 200 earthquake records have been collected from blast-monitors in the lower Wabash River Valley area of southern Indiana and Illinois, and north central Kentucky. The earthquakes range in magnitude between **M**3 and 5.2. The locations of the events are shown in Figure 1. Paleoseismological evidence, historical earthquake accounts, and contemporary earthquake records indicate that the Wabash River Valley has a considerable seismic hazard (e.g., Obermeier et al., 1991; Pond and Martin, 1997; Munson et al., 1997; Pavlis et al., 2002; Kim, 2003; Herrmann et al., 2008). Historical and instrumental evidence have shown that small to moderate earthquakes occur in an area roughly coincident with the Wabash Valley fault system (Fig. 2). The low rate of seismicity and relatively sparse seismic network coverage has made correlating seismicity with specific geological structure problematic, however.

Table 1 gives the locations of the blast monitors and peak horizontal velocities (PHV's) recorded for the four largest earthquakes of April, 2008 sequence. Blast monitor locations and ground motion values for the 1987 and 2002 events are given in Street et al. (1988) and Street et al. (2005), respectively. The blast monitor locations and ground motions for the three smaller events that are indicated in Figure 1 by the filled circles are listed in the Appendix A. These events are not discussed in this report, but are included for completeness.

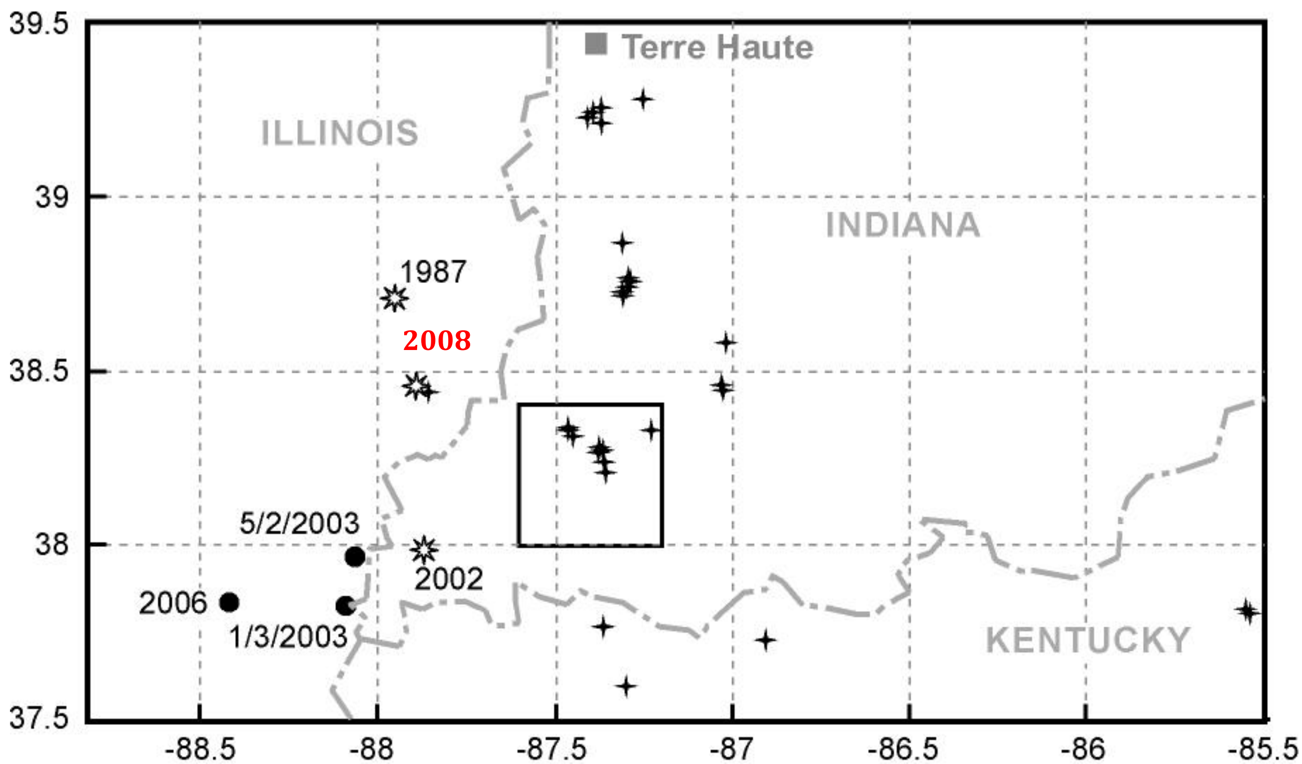


Figure 1. Locations of the blast monitor sites that recorded the 2008 earthquakes (+), the epicenters of the 1987, 2002, and 2008 earthquakes (*) that are discussed in the report, the locations of three small earthquakes (•) that are listed in the Appendix A, and the boundary of a 0.4^0x0.4^0 area discussed in the text (rectangle).

Our three primary objectives are to summarize the velocity recordings and site investigations for the April 2008 southeastern Illinois earthquake sequence, put the findings into context with previous work in the area, and to quantify the reduction in variability of ground motions that can be achieved with conventional site investigations and one-dimensional site effect approximations. For the latter objective, we focus on a $0.4^0 x 0.4^0$ area where there exists a relatively high density of blast monitor recordings of earthquakes, SH-wave refraction/reflection site investigations, and limited borehole information.

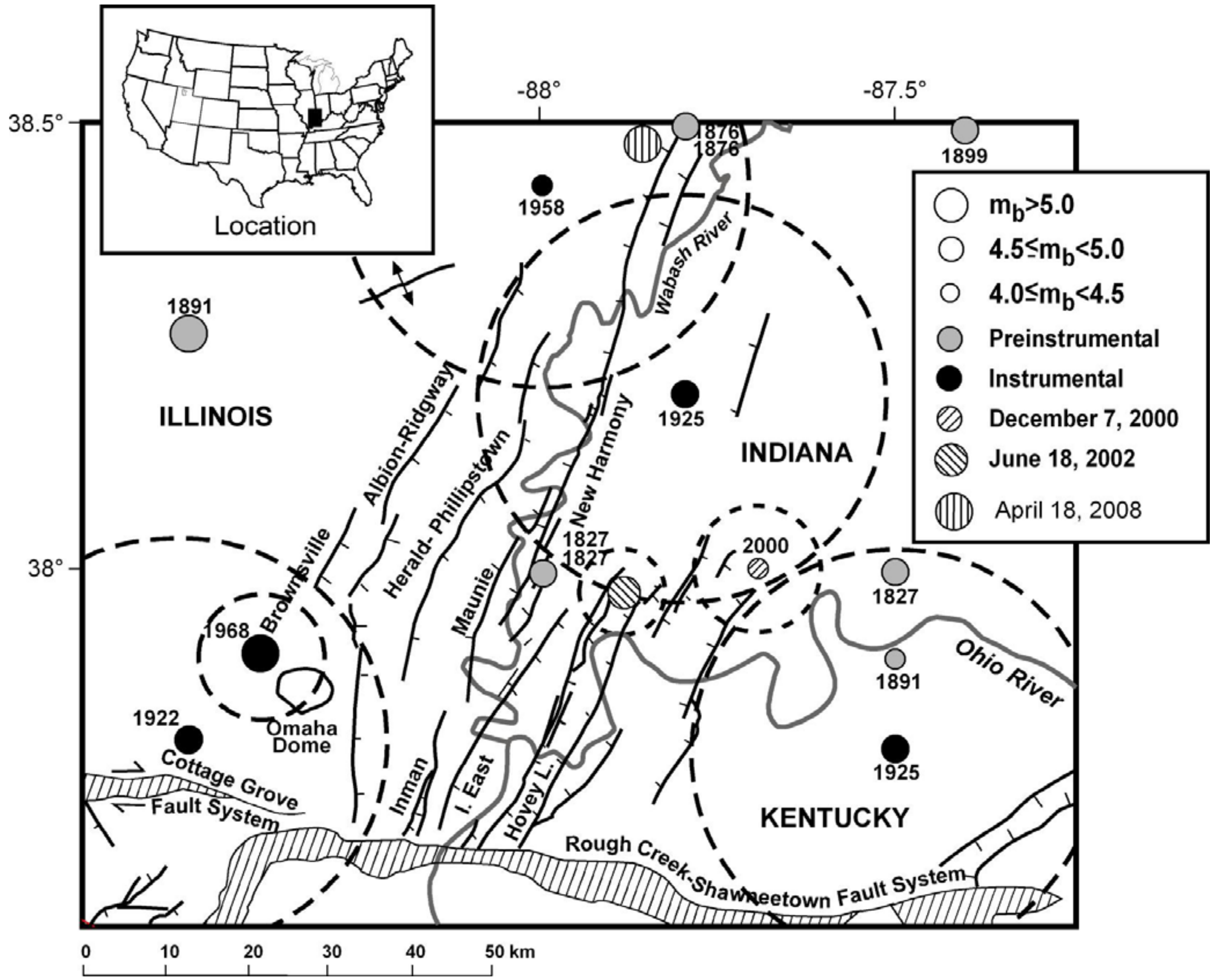


Figure 2. Location map that shows approximate location of significant historical and contemporary earthquakes in relation to the mapped structures of the Wabash Valley fault system (modified from Bear et al., 1997 and Woolery 2005). Dashed circles indicate uncertainty in the instrumentally derived epicenters (filled circles). Shaded circles represent historical epicenters from individual investigator's interpretation of intensity reports.

TABLE 1

1.1 **M**5.23; April 18, 2008, at 09:37:00 UTC

Site No.	Location ⁰N/⁰W	Dist. km	Azi. deg.	H1 cm/s	Vert. cm/s	H2 cm/s
1	38.336/87.461	39.3	109	4.656	0.356	1.829
2	38.334/87.462	33.0	120	1.473	0.381	2.057
3	38.279/87.377	48.3	113	1.041	0.302	1.893
4	38.268/87.378	48.7	114	2.159	0.305	1.194
5	38.241/87.364	51.1	117	2.743	0.432	1.448
6	38.277/87.374	48.6	113	0.864	0.330	1.372
7	38.212/87.358	53.0	117	0.737	0.229	1.372
8	38.317/87.452	40.7	111	0.889	0.279	0.457
9	38.331/87.228	58.9	103	0.762	0.254	0.940
10	38.868/87.309	68.2	47	0.787	0.229	0.914
11	37.810/85.542	216	108	0.191	([1])	0.368
12	37.821/85.547	215	108	0.267	0.178	0.152
13	37.821/85.548	215	108	0.241	0.152	0.229
14	39.243/87.391	97.6	26	1.194	0.254	0.699
15	39.226/87.406	95.3	26	0.787	0.127	1.143
16	38.450/87.026	74.8	90	0.686	0.457	1.803[2]
17	39.212/87.371	95.3	28	0.737	0.203	1.295
18	38.463/87.029	74.5	89	1.143	0.203	0.991
19	39.280/87.254	107	31	0.470	0.127	0.597
20	38.581/87.016	76.9	79	0.279	([1])	0.305
21	39.255/87.371	99.5	26	0.686	0.152	0.762
22	38.444/87.855	3.12[3]	106	3.318	0.750	3.180
23	37.769/87.366	88.0	149	0.813	0.191	0.610
24	38.727/87.311	58.6	58	0.851	0.165	0.711
25	38.760/87.286	62.4	56	0.648	0.267	0.889
26	38.742/87.297	60.5	58	0.749	0.241	0.470
27	38.760/87.289	62.2	56	1.168	0.254	1.067
28	38.731/87.310	58.5	59	0.914	0.305	0.914
29	37.732/86.906	117	133	0.914	0.191	1.219
30	40.598/86.681	259	23	0.152	0.191	0.165
31	37.600/87.305	107	151	0.216	0.083	0.222
32	38.282/87.377	48	113	0.787	0.152	1.448[2]
33	38.268/87.377	49	114	1.190	0.438	1.620

1. The peak ground motion of the vertical component is near or below the resolution of the system.
2. Blast monitor triggered on the P-wave and only a short segment of the Sg/Lg arrival and its coda is included in the record. Larger peak velocities could have occurred after the instrument shut down.
3. The hypocentral distance of the blast monitor for this site is 14.3 km.

1.2 **M**4.61, April 18, 2008, at 15:14:16 UTC

Site No.	Location °N/°W	Dist. km	Azi. deg.	H1 cm/s	Vert. cm/s	H2 cm/s
1	38.336/87.461	39.3	109	1.397	0.203	1.168
2	38.334/87.462	38.7	120	1.270	0.279	1.778
3	38.279/87.377	48.3	113	0.330	0.102	0.457
4	38.268/87.378	48.7	114	0.940	0.102	0.356
5	38.241/87.364	51.1	117	0.635	0.127	0.279
6	38.277/87.374	48.6	113	0.305	0.102	0.457
7	38.212/87.358	53.0	117	0.254	([1])	0.356
8	38.317/87.452	40.7	111	0.279	0.076	0.203
9	38.331/87.228	58.9	103	0.178	0.076	0.203
10	38.868/87.309	68.2	47	0.356	([1])	0.330
14	39.243/87.391	97.6	26	0.368	0.102	0.229
15	39.226/87.406	95.3	26	0.279	0.076	0.356
16	38.450/87.026	74.8	90	0.178	0.102	0.229
17	39.212/87.371	95.3	28	0.457	0.127	0.762
18	38.463/87.029	74.5	89	0.203	([1])	0.152
19	39.280/87.254	107	31	0.254	0.051	0.203
20	38.581/87.016	76.9	79	0.152	([1])	0.127
21	39.255/87.371	99.5	26	0.229	0.076	0.406
22	38.444/87.855	3.12[2]	106	2.438	1.092	3.251
23	37.769/87.366	88.0	149	0.222	0.051	0.247
24	38.727/87.311	58.6	58	0.813	0.178	0.305
25	38.760/87.286	62.4	56	0.279	0.127	0.330
26	38.742/87.297	60.5	58	0.368	0.114	0.305
27	38.760/87.289	62.2	56	0.508	0.165	0.330
28	38.731/87.310	58.5	59	0.406	0.203	0.914
29	37.732/86.906	117	133	0.414	0.071	0.465
32	37.89 / 89.24	135	242	0.451	0.089	0.286
33	38.282/87.377	48	113	0.267	0.070	0.046
34	38.268/87.377	49	114	0.160	0.038	0.073

1. The peak ground motion of the vertical component is near or below the resolution of the system.
2. The hypocentral distance of the blast monitor for this site is 14.5 km.
3. The hypocentral distance of the blast monitor for this site is 14.3 km. As a result of a transient spike in the recording, the PPV for this event is questionable.

1.3 **M**4.00, April 21, 2008, at 05:38:29 UTC

Site No.	Location $^0N/^0W$	Dist. km	Azi. deg.	H1 cm/s	Vert. cm/s	H2 cm/s
1	38.336/87.461	39.3	109	0.483	(1)	0.545
2	38.334/87.462	38.7	120	0.178	(1)	0.330
3	38.279/87.377	48.3	113	0.225	(1)	0.406
4	38.268/87.378	48.7	114	0.431	0.076	0.305
5	38.241/87.364	51.1	117	0.610	0.102	0.305
7	38.212/87.358	53.0	117	0.152	(1)	0.381
8	38.317/87.452	40.7	111	0.152	(1)	0.102
9	38.331/87.228	58.9	103	0.102	(1)	0.152
10	38.868/87.309	68.2	47	0.203	(1)	0.178
15	39.226/87.406	95.3	26	0.127	(1)	0.152
16	38.450/87.026	74.8	90	0.127	0.076	0.229
17	39.212/87.371	95.3	28	0.076	(1)	0.229
18	38.463/87.029	74.5	89	0.127	(1)	0.127
21	39.255/87.371	99.5	26	0.101	(1)	0.152
22	38.444/87.855	3.12[3]	106	0.470	0.159	0.654
23	37.769/87.366	88.0	149	0.203	(1)	0.318
25	38.760/87.286	62.4	56	0.279	0.127	0.330
26	38.742/87.297	60.5	58	0.368	0.114	0.305
27	38.760/87.289	62.2	56	0.508	0.165	0.330
29	37.732/86.906	117	133	0.414	0.071	0.465
33	38.282/87.377	48	113	0.203	(1)	0.318

1. The peak ground motions of the vertical component is near or below the resolution of the system.

1.4 **M**3.72, April 21, 2008, at 05:38:29 UTC

Site No.	Location $^0N/^0W$	Dist. km	Azi. deg.	H1 cm/s	Vert. cm/s	H2 cm/s
1	38.336/87.461	39.3	109	0.330	(1)	0.179
2	38.334/87.462	38.7	120	0.203	(1)	0.279
4	38.268/87.378	48.7	114	0.127	(1)	0.127
5	38.241/87.364	51.1	117	0.152	(1)	0.102
18	38.463/87.029	74.5	89	0.102	(1)	0.178
34	38.268/87.377	47	113	0.083	(1)	0.089

1. The peak ground motion of the vertical component is near or below the resolution of the system.

Blast Monitors Records for the April 2008 Southeastern Illinois Earthquake Sequence

Blast monitor records are digital velocity recordings typically acquired at a sampling rate of 512 or 1024 samples per second (sps), with a useable pass-band filter range between 2 and 100 Hz. The blast monitors that recorded the 2008 events were all set at a sampling rate of 1024 sps. Data are acquired with 3-component transducers that are located 20 to 35 cm beneath the ground surface. The instruments were programmed with 1-second of pre-event memory, and typically set to “trigger” at a peak velocity of approximately 0.1 cm/s; therefore, the instruments were usually triggered by the S-wave if the epicenter to the event was more than a couple of tens of kilometers from the blast monitor. Timing is maintained by an internal clock, but is not rigorously synchronized. Consequently, arrival times on the blast monitors are inadequate for phase studies. Nevertheless, blast monitors are well-maintained instruments. Recordings from the instruments are intended to serve in legal and regulatory proceedings, thus instrument response is regularly calibrated on shake tables. In addition, subsequent to each recording, calibration pulse for each component is recorded as a visual check on the instrument operation. Figure 3 is a typical blast monitor record (Site 6) for the **M**5.2 event. This record, like all but two recordings of the 2008 events, was triggered by the *S*-wave arrival. The epicentral distance between Site 6 and the **M**5.2 event is 48.6 km (Table 1). Each component on the records is labeled radial (R), vertical (V), and transverse (T), but the orientation of the horizontal axes of the transducer are not well documented with respect to the geographical axes. Consequently, the generic terms H1 and H2 are used to describe the horizontal components. The orientation of the radial component of a blast monitor is generally directed towards the blasting site.

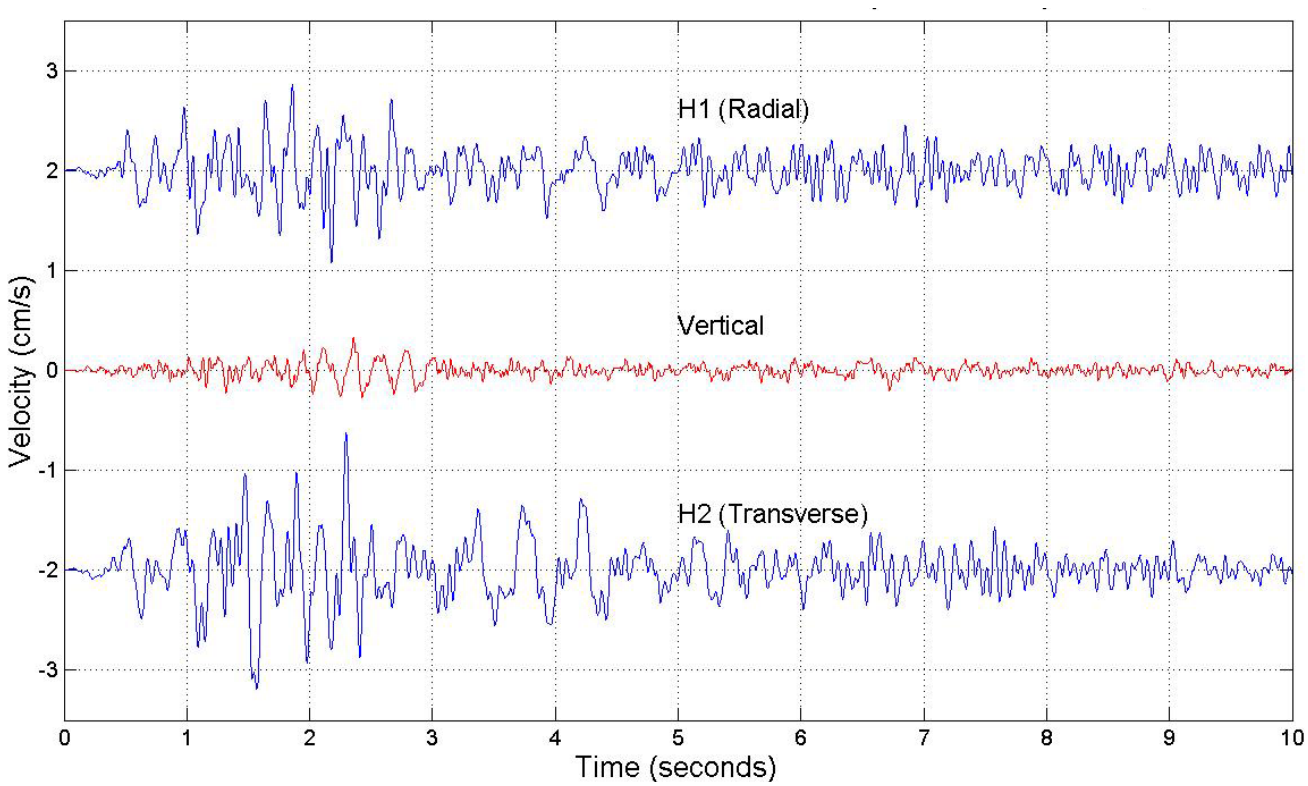


Figure 3. Blast monitor record at Site 6 of the M5.23 event on April 18, 2008. The horizontal traces are labeled H1 and H2 since their orientation with respect to the north-south axis is not well determined.

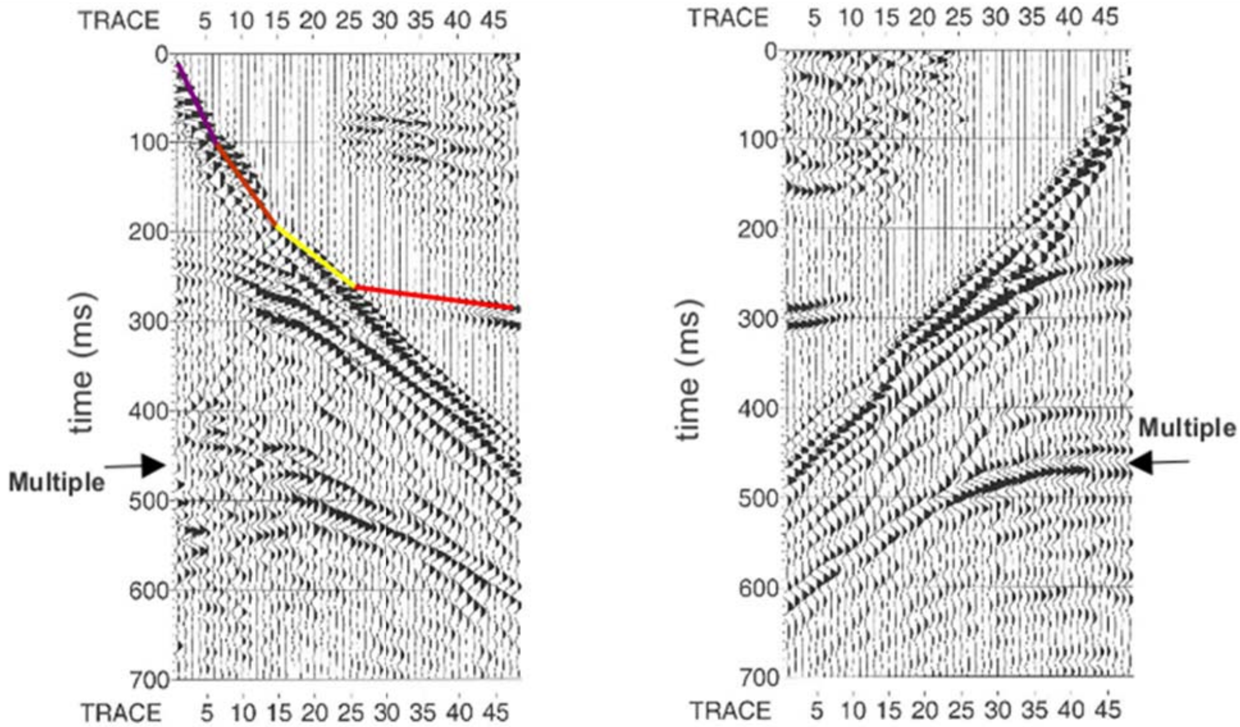
Unlike seismometers and most strong-motion installations, blast monitors are seldom placed at locations where the cultural site conditions are favorable with respect to low-noise conditions. In general, blast monitors are placed near homes or other structures where ground vibrations resulting from nearby blasting may be problematic. In addition, the geologic site conditions are not a consideration, i.e. the instruments may be situated on anything a thin residual soil veneer, alluvium, ridgelines, etc. Due to the wide-ranging geologic locations, the dynamic site conditions undoubtedly contribute to the variability of the recorded ground motions.

The permanency in a blast monitor's location is another variance with conventional seismic or strong-motion installations. Blast monitor locations follow mining activity, which in the case of the surface coal mines in the lower Wabash River Valley, frequently shifts. As a result, the locations of the blast monitors triggered by the various earthquakes during the two-decade time period of these observations changed; therefore, there is a lack of continuity in the recording sites which is generally not true for seismic and strong-motion stations.

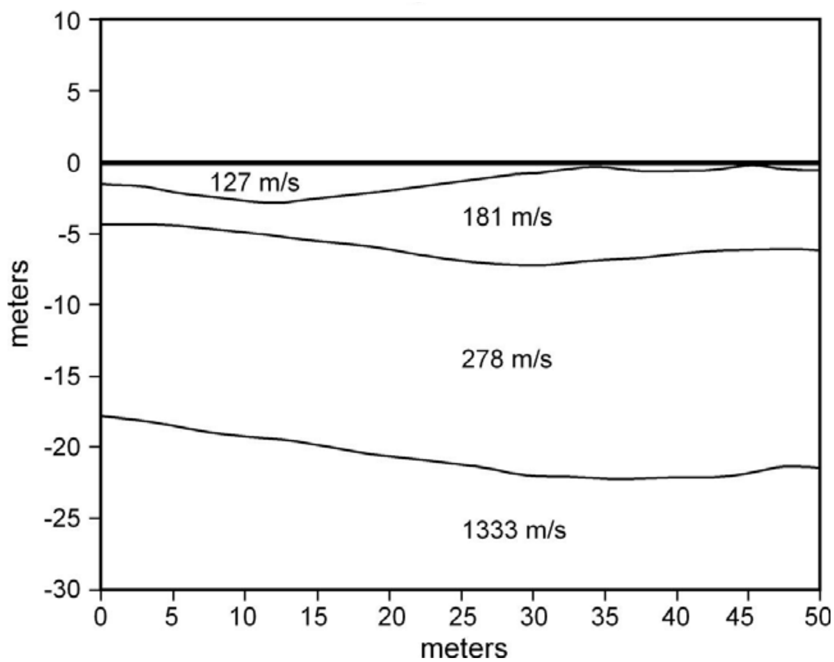
Site Conditions and 1-D Linear Site Responses

Dynamic site conditions have been determined at 18 of the blast monitor locations where the 2008 earthquakes were recorded. The conditions were investigated using conventional reversed SH-wave reflection/refraction seismic profiling and horizontal-to-vertical ambient noise ratios (H/V). Figure 4a shows an example of reversed SH-wave reflection/refraction seismic data. The seismic data were acquired with forty-eight, inline, 30 Hz, horizontally polarized geophones spaced at intervals of 2 m, with shot points at the zero offsets and the midpoint (data are not shown) of the geophone array, and a sampling interval of 0.25 ms. The energy source was a 1 lb hammer striking a steel I-beam perpendicular to the orientation of the geophone array. Data were vertically stacked as necessary. The polarity of the energy was reversed by striking the opposite side of the I-beam and symmetrically stacking. The data processing typically consisted of band pass filtering between 20 and 60 Hz, application of an appropriate AGC window, and the occasional application of an F-K filter to reduce off-line noise. Figure 4b shows the S-wave interpreted velocity model interpreted for the example site. The remaining site interpretations are shown in Figure 5.

Ambient noise samples were also collected for 17 of the 18 sites. The noise samples were collected at the midpoints of the seismic lines, using a three-component, 1-Hz Mark Products L-4C seismometer. At each site, three 15-sec ambient noise records were acquired at a sampling rate of 500 sps. The recording system included an active anti-aliasing 15-Hz high-cut filter that rolls off at 12 db. The high-cut filter is a limitation at the few sites where there is a relatively thin layer of low-velocity soils over a much higher velocity layer, such as at Site 16. The 1D linear approximations indicate that there are pronounced resonances at 15.8 and 21.4 Hz; frequencies well outside of the detection passband of the L-4C seismometer/high-cut filter.

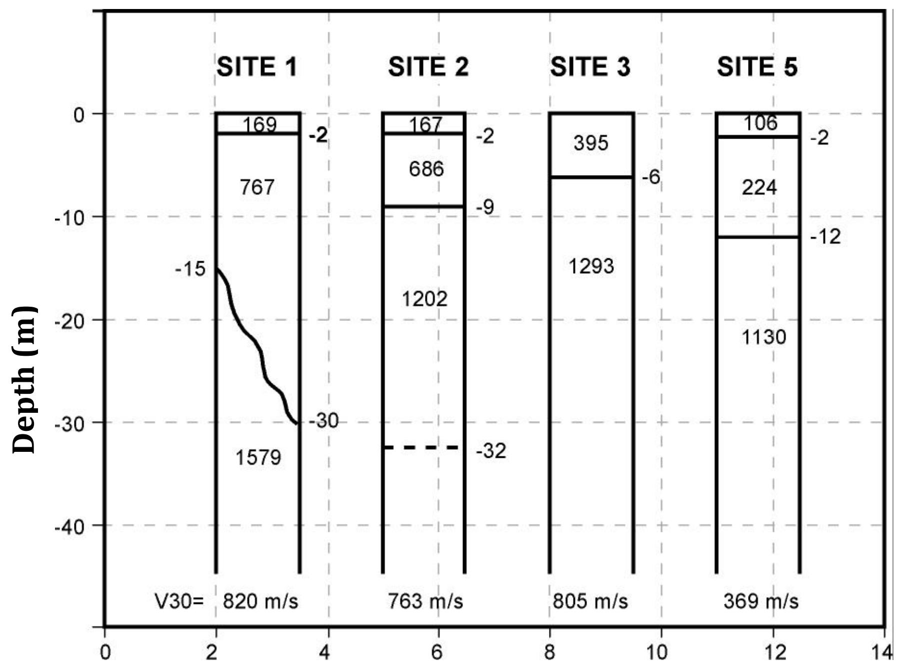


(a)

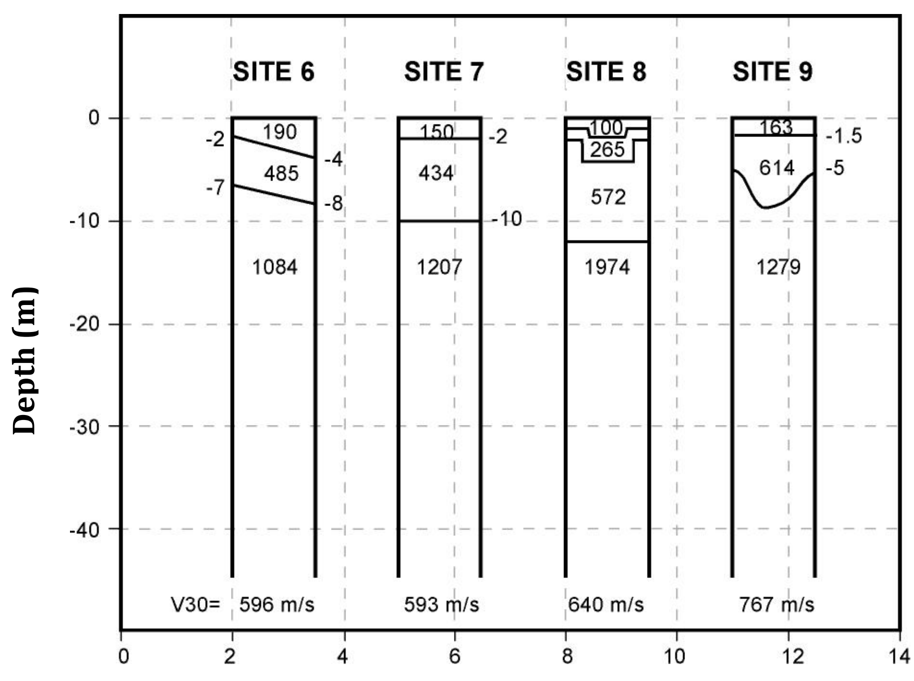


(b)

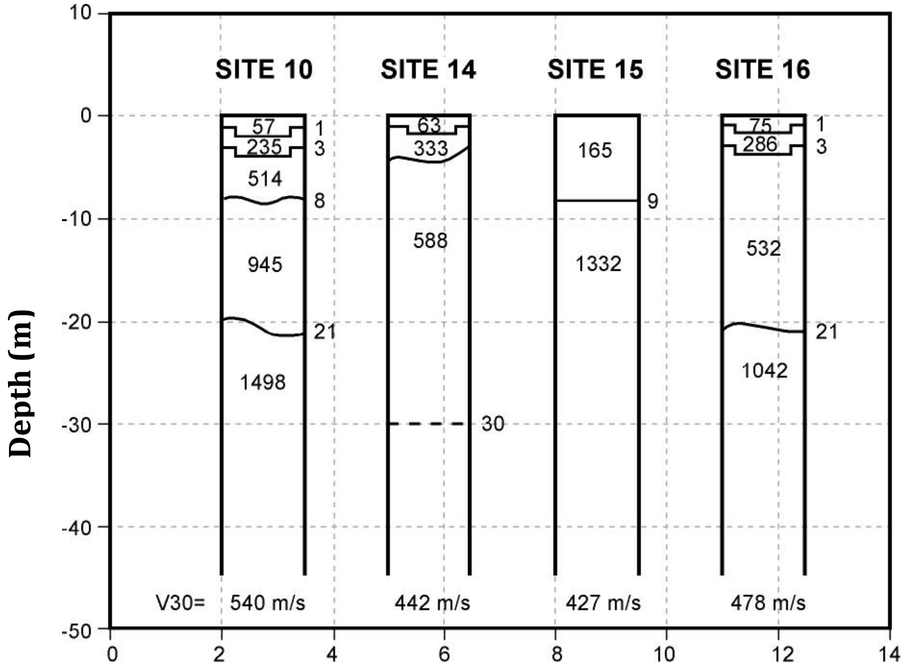
Figure 4. (a) Example of a processed, reversed SH-wave seismic refraction/reflection sounding. Data were collected at a sampling rate of 0.25 ms, from an inline spread of 48 geophones that were spaced at 2 m. (b) The *S*-wave velocity and depth model interpreted for the site from the first arrival times.


SITE 1
SITE 2
SITE 3
SITE 5
Depth (m)
169
767
1579
-2
-15
-30
167
686
1202
-2
-9
-32
395
1293
-6
106
224
1130
-2
-12
V30= 820 m/s
763 m/s
805 m/s
369 m/s

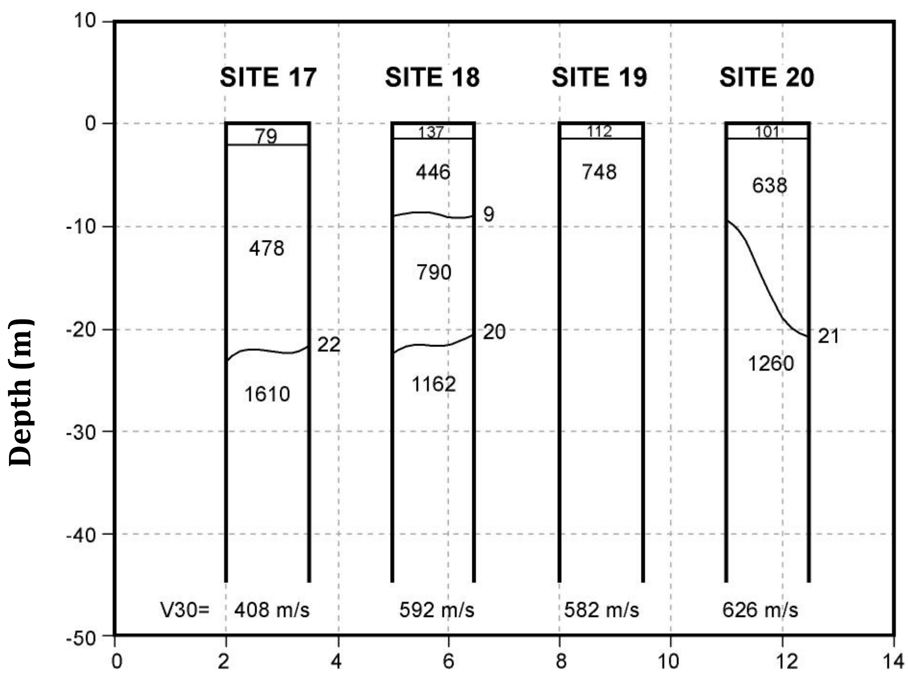
(a)


SITE 6
SITE 7
SITE 8
SITE 9
Depth (m)
190
485
1084
-2
-7
-4
-8
150
434
1207
-2
-10
100
265
572
1974
163
614
1279
-1.5
-5
V30= 596 m/s
593 m/s
640 m/s
767 m/s

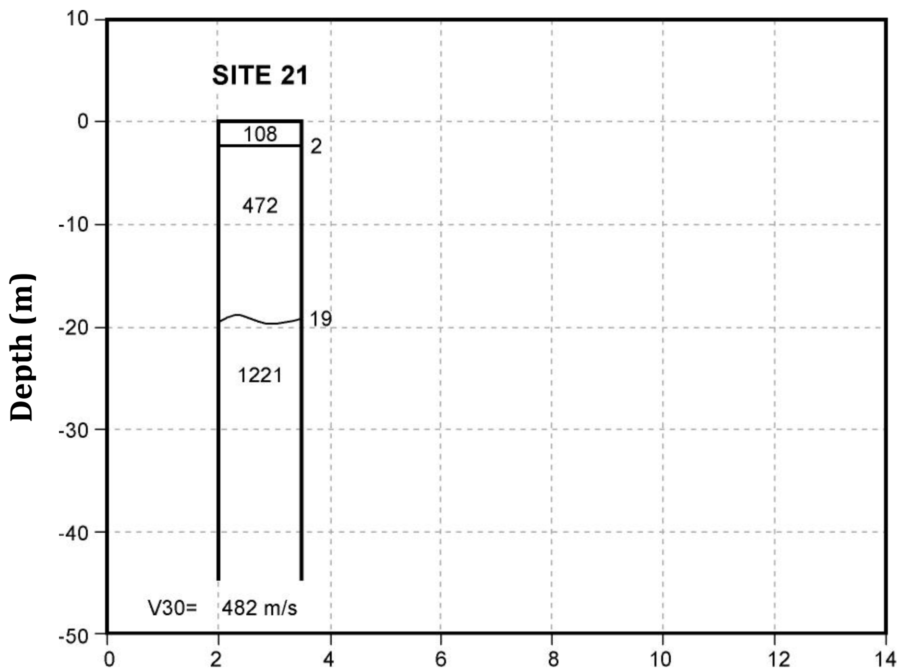
(b)


SITE 10
SITE 14
SITE 15
SITE 16
Depth (m)
57
235
514
945
1498
63
333
588
165
1332
75
286
532
1042
V30= 540 m/s
442 m/s
427 m/s
478 m/s

(c)


SITE 17
SITE 18
SITE 19
SITE 20
Depth (m)
79
478
1610
137
446
790
1162
112
748
101
638
1260
V30= 408 m/s
592 m/s
582 m/s
626 m/s

(d)

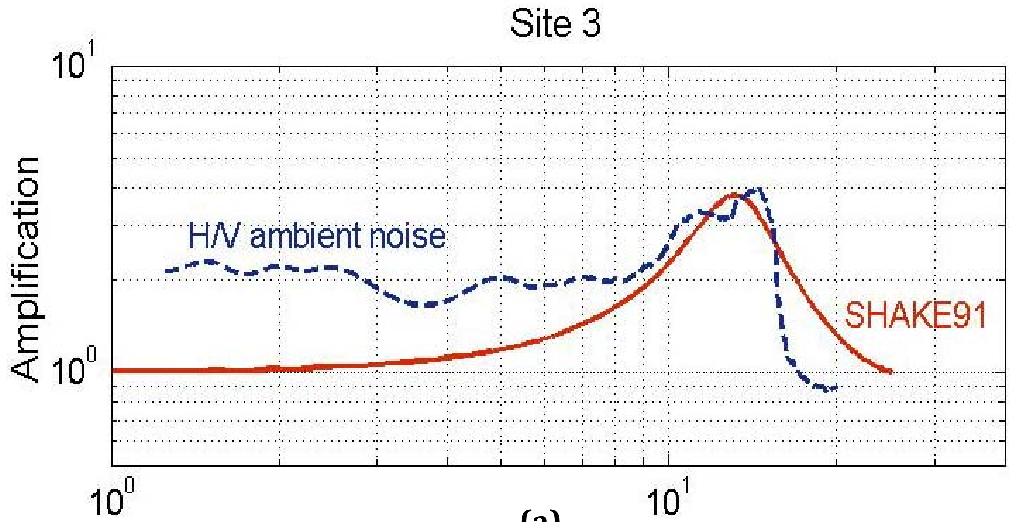


(e)

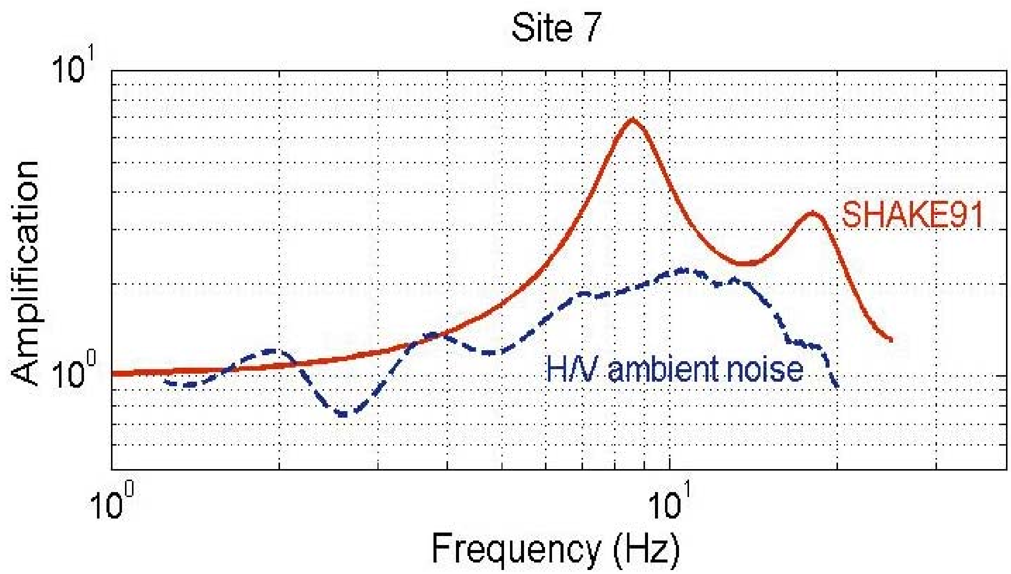
Figure 5. *S*-wave velocity and depth models interpreted for this study. Straight lines indicate planar contacts that are horizontal or dipping; wavy lines are used to indicate irregular contacts. Contacts are based on the interpretation from 48 geophones spread over 94 m; the horizontal width of the models, as compared to the vertical scale, is compressed by 10:1. The *S*-wave velocities (m/s) are shown in the interiors of the columns, and the scaling for the depths (m) is shown along the left side of the figure. V_{30}, the time-averaged shear-wave velocity of the upper 30 m of soil and rock, is indicated at the bottom of the velocity/depth models. The V_{30} velocities were calculated at the model mid-points.

Having stated this, ground motions above 12 Hz are generally of little engineering interest; therefore, the high-frequency system limitation for the ambient noise recordings is of little, if any, practical consequence.

The SH-wave velocity models and ambient noise samples were used to estimate the 1-D site responses and resonance frequencies of the sites with SHAKE91 and Nakamura's (1989) H/V spectral ratio method, respectively. Figure 6 shows the 1-D responses estimated by the two methods for sites 3, and 7; smoothing for the H/V ambient noise samples was done with the 20-point weighted function described by Konno and Ohmachi (1998). The spectra in Figure 6 illustrates a not uncommon experience we have



(a)



(b)

Figure 6. Example comparisons of the horizontal-to-vertical ambient noise (dashed line) with the one-dimensional linear approximation (SHAKE91) results (solid line) at sites 3 and 7. The falloff in the horizontal-to-vertical noise ratios in the plots at frequencies above 15 Hz is the result of an active filter in system used in acquiring the ambient noise data.

had in applying the H/V ambient noise method in the lower Wabash river valley area. Specifically, for relatively simple sites characterized by one or two layers of sediment overlying rock (e.g. Site 3), the H/V ambient noise ratio exhibits a clear indication of the resonance frequency, but at other sites with similar characteristics, the H/V ambient noise ratios are ambiguous (e.g. Site 7). The H/V ambient noise technique has another limitation as illustrated by Site 1 (Fig. 7). The H/V ambient noise result acquired at the Site 1 midpoint array agrees with the resonance frequency determined by the coincident midpoint 1D linear approximation; however, the seismic reflection/refraction model clearly indicates an irregular dipping bedrock surface. This apparently cannot be resolved by a single H/V ambient noise measurement. Other investigators have reported that dipping beds and small-scale irregularities with physical dimensions similar to the seismic wavelengths being considered can result in site effect variation (e.g. Ohtsuki and Harumi, 1983; Aki, 1988; Harmsen, 1997; Ghayamghamian, 2008; among others). Consequently, the irregular boundary at Site 1 may be the dominant factor in the overall characterization of the site effect.

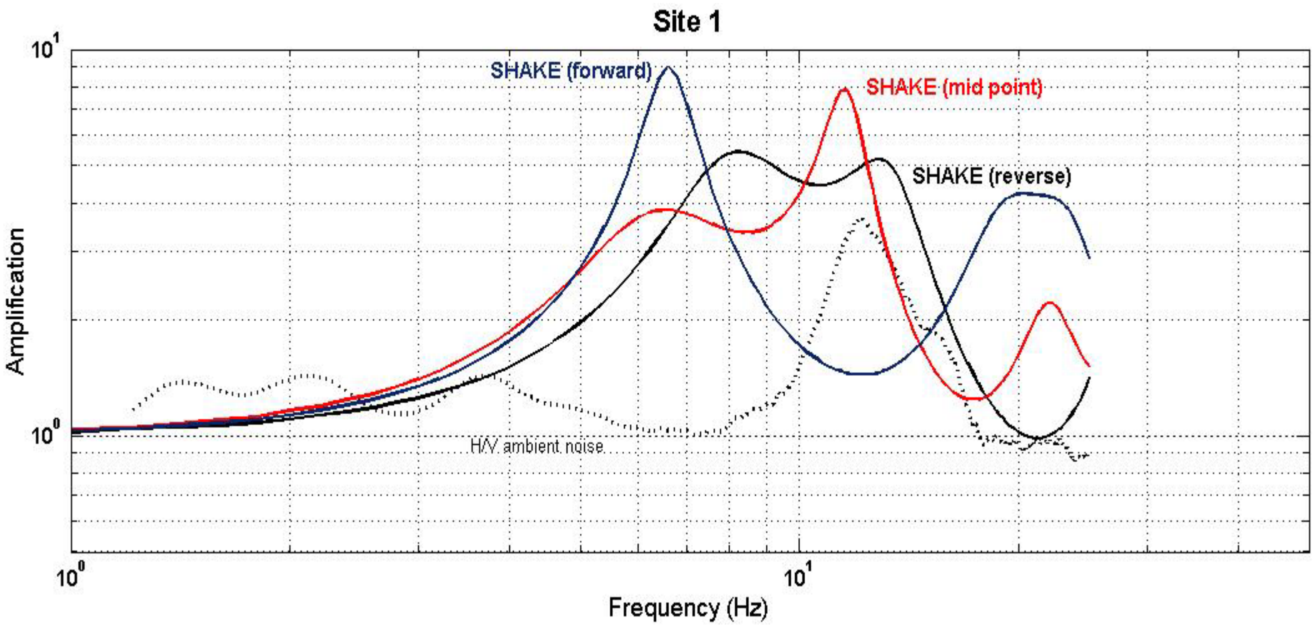


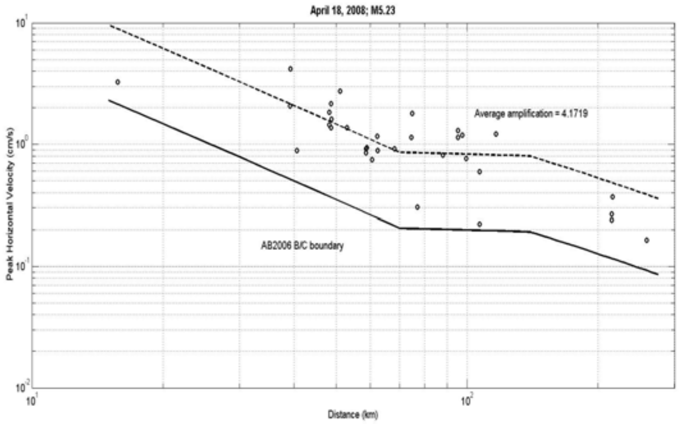
Figure 7. H/V ambient noise at the array midpoint (dashed line) compared with the 1D approximation (SHAKE91) results for the site conditions at either end [SHAKE (forward) and SHAKE (reverse)] and at the center [SHAKE (mid-point)] of the geophone spread. The lowermost velocity layer at Site 1 is a steeply dipping interface (Figure 4).

Peak Horizontal Velocities

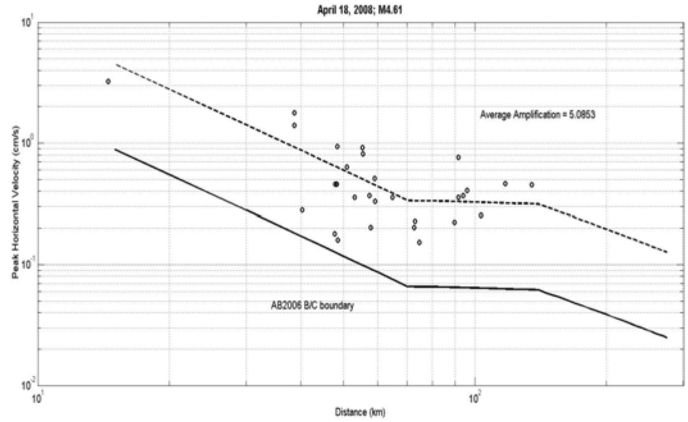
Figure 8 shows the plots of the peak horizontal velocities (cm/s) as a function of their epicentral distances (km) for the four largest (**M**5.2, **M**4.6, **M**4.0, and **M**3.7) earthquakes in the April 18, 2008 sequence. The solid lines indicate the peak horizontal velocity (PHV) predicted by the Atkinson and Boore (2006a, 2006b) ground-motion model for the B/C boundary (760 m/s). AB06 does not consider site effects and assumes a stress parameter of 140 bars (1.4×10^7 N/m^2). The B/C boundary model is used because the regional bedrock S-wave seismic velocities, determined in this and previous studies (e.g. Zhang, et al., 1991; Woolery, et al., 2009), indicate an averaged velocity for the lower Wabash River Valley of 1065 m/s. The dashed lines in the figures represent a least-squares fit of the B/C boundary model predictions with the PHV observations. The ratio of the dashed lines to the solid lines is shown at the top of the plots.

Similar analyses were performed for the **M**4.9 June 10, 1987 and **M**4.5 June 18, 2002 earthquakes (Fig. 9). The best fit of the observed to the predicted PHV residuals for the June 10, 1987, earthquake (Figure 9a) illustrates the difficulty in predicting ground motions. The average observed-to-predicted PHV ratio for the June 10, 1987 is 4.85 if the magnitude is assumed **M**4.9 (Herrmann and Ammon, 1997) with a stress parameter of 140 bars. Conversely, if the 290 bar stress parameter suggested by Atkinson and Boore (2006) is assumed, then the ratio of the observed to the predicted PHV's is 0.733. The latter value suggests that the typical 5 to 30 m sediment overburden at the 1987 observation sites generally result in an overall deamplification of the PHV's. This is unlikely, but the results demonstrate the importance in the choice of stress parameter when predicting ground motions for frequencies greater than f_{02} (i.e., the intersection of the T^0 and T^{-2} asymptotes for the source spectrum). The average observed-to-predicted PHV ratio for the 2002 event is 11.84 (Figure 9b).

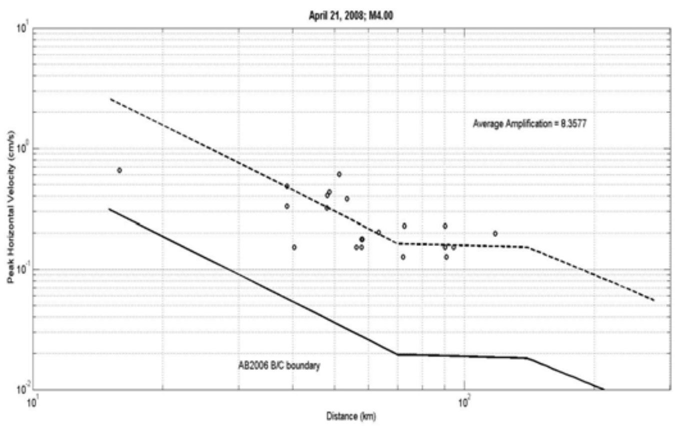
Figure 10 shows, with the exception of Site 22, the PPV's for the three largest 2008 events plotted as a function of epicentral distances. The PPV's for site 22 are plotted as a function of their hypocentral distances which is thought to be a more realistic value than the 3 km epicentral distances. The PPV for the **M**4.6 event at Site 22 is not included because of malfunction in the blast monitor (see Table 1.2).



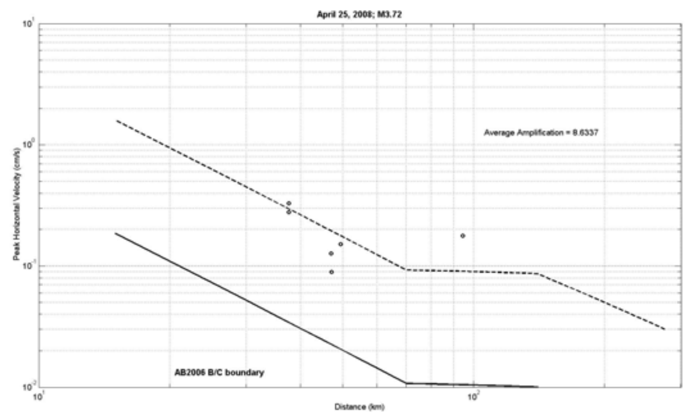
(a)



(b)



(c)



(d)

Figure 7. Peak horizontal velocities (o) recorded on blast monitors for four of the largest earthquakes in the southeastern Illinois sequence of April, 2008. The solid line is the Atkinson and Boore (2006) relationship for ground velocities at the B/C boundary and a stress parameter of 140 bars. The dashed line is the best fit of the solid line to the observed peak velocities. The "Average Amplification" is the ratio of the amplitude of the dashed line to the solid line.

(a)

(b)

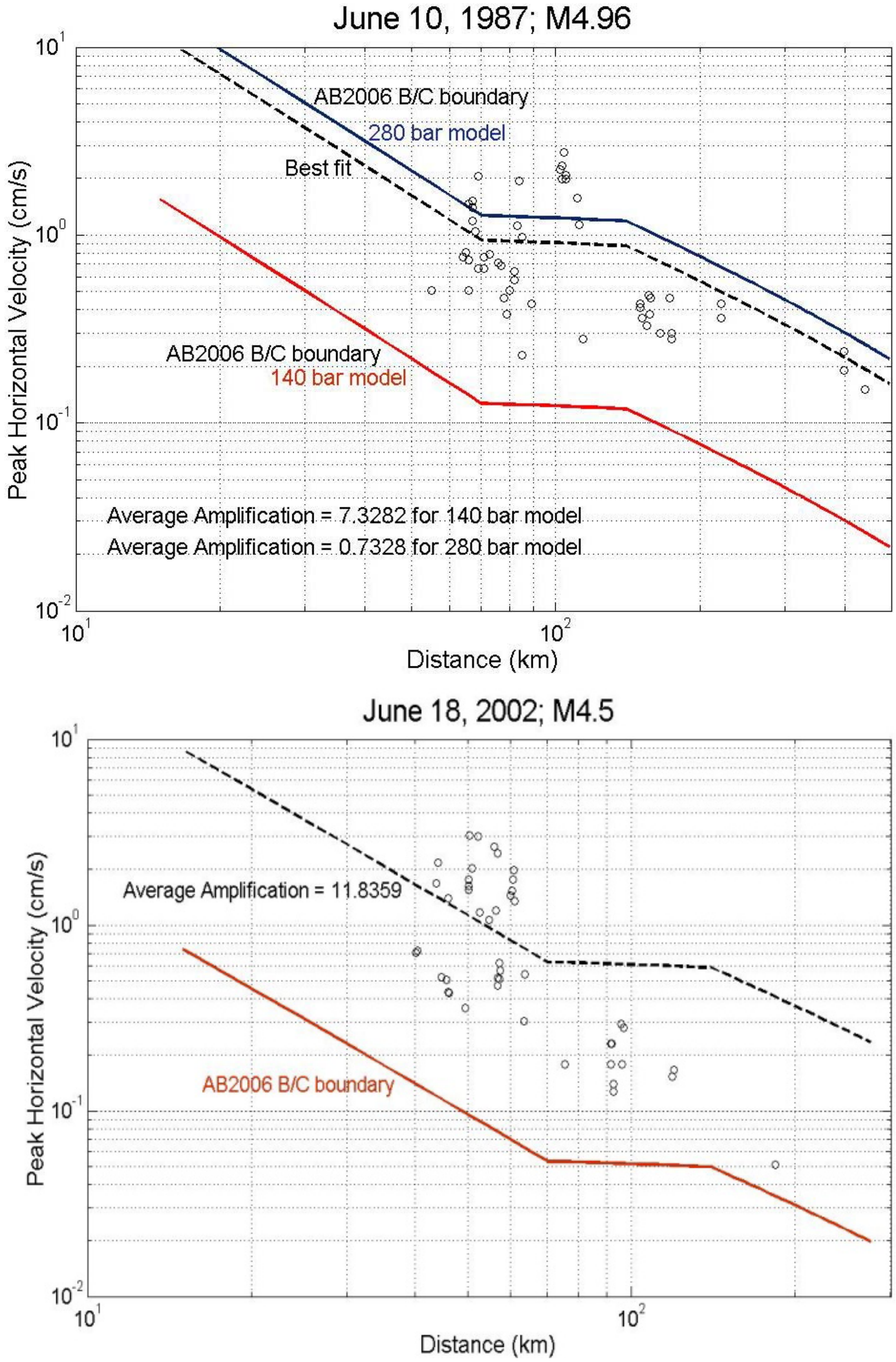


Figure 9. Peak horizontal velocities (o) recorded on blast monitors for the (a) M4.96 southeastern Illinois earthquake of June 10, 1987, and (b) M4.5 southwestern Indiana earthquake of June 18, 2002. The Atkinson and Boore (2006) for ground velocities at the B/C boundary and a stress parameter is indicated by solid lines and labeled in the plots, the dashed line is the best of the solid line to the observed peak velocities, and the "Average Amplification" is the ratio of the amplitude of the dashed line to the solid line. In (a), the second solid line labeled 290 bars is the stress parameter for the 1987 suggested by Atkinson and Boore (2006).

The PPV's are keyed to the specific event, and the colors correlate with the Modified Mercalli intensity (MMI) provided by the USGS intensity maps for the area. There are too few MMI VI's to define a distinct boundary between the intensity VI's and V's; however, the approximate PPV boundary between the V's and IV's is near the 1.2 cm/s line. The line is the approximate boundary separating the intensity V and IV given in Street et al. (2005) for the **M**4.5 southwestern Indiana earthquake of 18 June, 2002. Given the two data sets, it suggests that for the lower Wabash River Valley the PPV boundary between intensity IV and V is approximately 1.2 cm/s; an observation that could be useful in scaling ground motions of historic events in area, as well as predicting the effects of future events.

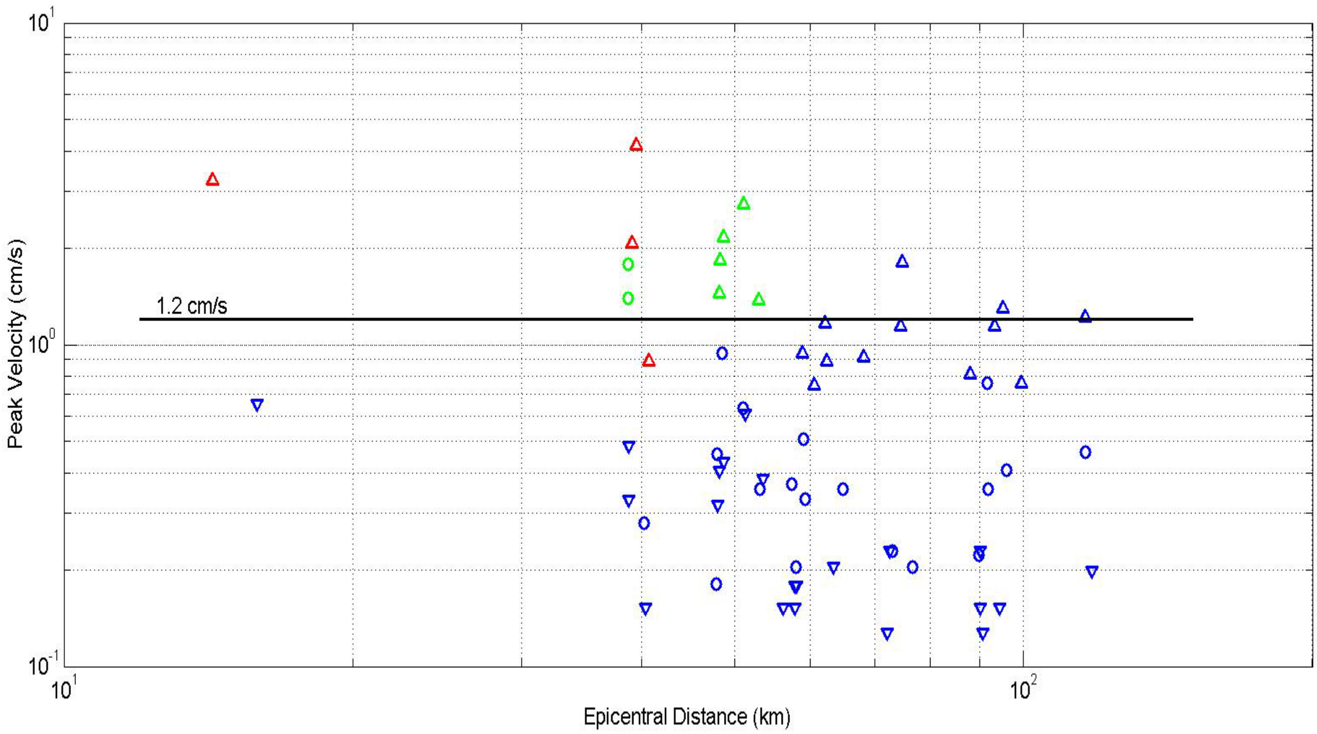


Figure 10. PPV's for the M5.2 (upright triangles), M4.6 (circles), and M4.0 (upside down triangles) events of April, 2008, plotted as a function of their epicentral distances, and their correlation to the reported Modified Mercalli intensities (MMI). The red symbols correspond to MMI's of VI, the green symbols correspond to the MMI's of V, and the blue symbols correspond to the MMI's of IV. The black horizontal line in the figure corresponds to the approximate PPV boundary between the MMI's of IV and V determined from Figure 10A of Street *et al*. (2005).

Observations and Site Conditions in a 0.4^0 x 0.4^0 Area of SW Indiana

A subset of 54 blast-monitor records from 46 sites within a 0.4^0 x 0.4^0 study area of southwestern Indiana are used to quantitatively investigate the meaning of the word phrase "profound" or "significant" for describing site effects in the lower Wabash Valley. The location of the area with respect to the epicenters of the 1987, 2002, and 2008 earthquakes is shown in Figure 1, and their distribution within the 0.4^0 x 0.4^0 area are shown in Figure 11. This area was selected because of the relatively high density of recordings, and the detailed information on the site conditions that has been developed within the area from the SH-wave seismic surveys. In addition, the northern boundary of the area is at approximately 38.4^0 and is coincident with the COCORP seismic line discussed in Bear et al. (1997). This seismic profile and associated velocity model crosses near the epicenters of the April, 2008 earthquakes. The earthquake observations available include the 10 June 1987 **M**5.0 southeastern Illinois, 18 June 2002 **M**4.5 southwestern Indiana, and the 18 April 2008 **M**5.2 and **M**4.6 southeastern Illinois events.

The regional topography is characterized by gently rolling hills and broad flat valley floors in the major drainage areas, with relief varying between 120 and 165 m above mean sea level. Site conditions typically consist between 2 and 30 m of low S-wave velocity loess, lacustrine deposits, or alluvium overburden unconformably atop Pennsylvanian bedrock. The bedrock consists of cyclic sequences of shale, sandstone, siltstone, and claystones, and thin units of limestone and coal, indicative of the deltaic, fluvial and coal swamp depositional environments that existed. Units within the bedrock tend to be highly variable in thickness and continuity, thus explaining the variability in the observed bedrock shear-wave velocities. The structural gradient for the bedrock is about 5 m per km dipping to the southwest, but localized irregular highs and lows in the surface of the bedrock are observed in the seismic surveys. The causes for the irregular surfaces could result from weathering and/or erosion and/or structure. Based on the seismic profiles and limited borehole data at 33 of the sites, the average depth to bedrock is -8 m. These site conditions are generally representative of many areas in the central United States with the exception of those sites that are underlain by thick (> 30 m) sediment deposits associated with the Mississippi embayment and other major river valleys.

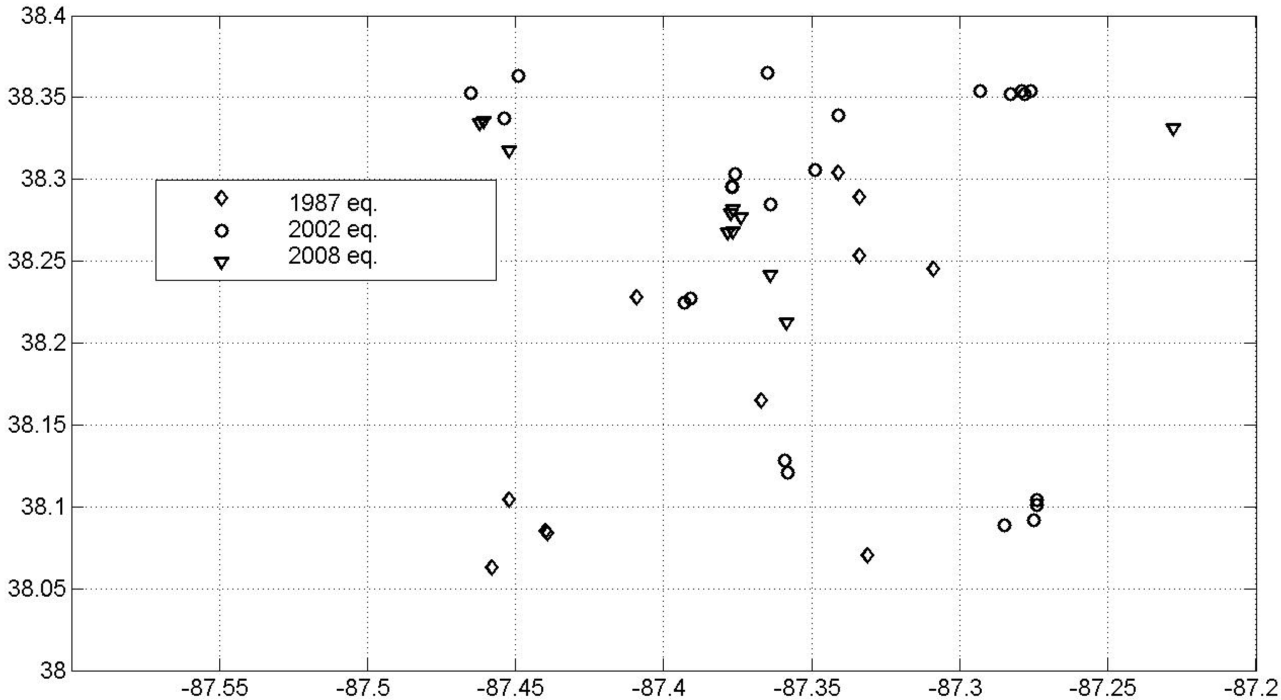


Figure 11. The blast-monitor locations within the 0.4^{0}x0.4^0 area that recorded the 1987, 2002, and one or more of the 2008 earthquakes.

Within the 0.4^0 x 0.4^0 there are 46 sites where observations were obtained for the1987, 2002, and 2008 events. Specifically, there were 11 sites for the1987 event, 24 for the 2002 event, and 11 for the 2008 **M**5.2 and **M**4.6 events. The earthquake record at Site 26 for the 2002 event was not included, because the instrument was located on engineered fill associated with the Wabash and Erie Canal (Street et al., 2005). Epicentral distances range between 69 and 89 km for the 1987 event, 40 and 61 km for the 2002 event, and 39 and 59 km for the 2008 event.

S-wave velocity models have been developed for 25 of the sites. Figure 12a shows 5% damped response spectra (pseudospectral velocities) at 16 of the 25 sites for the 2002 event. Figure 12b is a plot of damped response spectra for the event, scaled to a common distance of 50 km and corrected for path effect (Atkinson and Boore, 2006) with the exception of the S-wave velocity of the upper crust, which was assumed as 3.52 km/s.

Other spreading and attenuation models could have been used (e.g. Ou and Herrmann, 1990 or Erickon et al., 2004), but given the epicentral distance range and applying a common reference distance of 50 km, various models results in only small differences in the spectra.

The spectra shown in Figure 12c have been corrected for the linear one-dimensional site effect approximation using SHAKE91. The response spectra in Figure 12c are, like the response spectra in Figure 12b, scaled to a reference distance of 50 km. A comparison of the response spectra suggest that the range in scatter for the 2002 event has been reduced. At 4 Hz, for example, the scatter has been reduced by a factor of approximately 4. Figure 12d shows the means and plus one standard deviations of the spectra. The spectra in Figure 13 are similar to Figure 12, except that the 5% damped response are associated with the **M**4.6 2008 event. The eight spectra are for sites where site investigations, including S-wave profiling and ambient noise studies, were performed. The **M**4.6 event is used instead of the **M**5.2 event because at four of the nine sites where the main shock was recorded within the 0.4^0 x 0.4^0 area, the blast monitors were triggered by the P-wave and the preset recording length was such that the monitor turned off three or four seconds after the arrival of S-wave. The truncated records were judged to be too short for estimating response spectra. Figure 13d shows the means and plus one standard deviations for the response spectra shown in Figures 12b and 12c.

A similar analysis was not possible for the 1987 earthquake because the only available records for that event are low-gain paper copies that are inadequate for digitizing and determining response spectra. The peak ground velocities recorded at the sites are instrumentally stamped on the records at the time of the event, so the accuracy of those values is not in question.

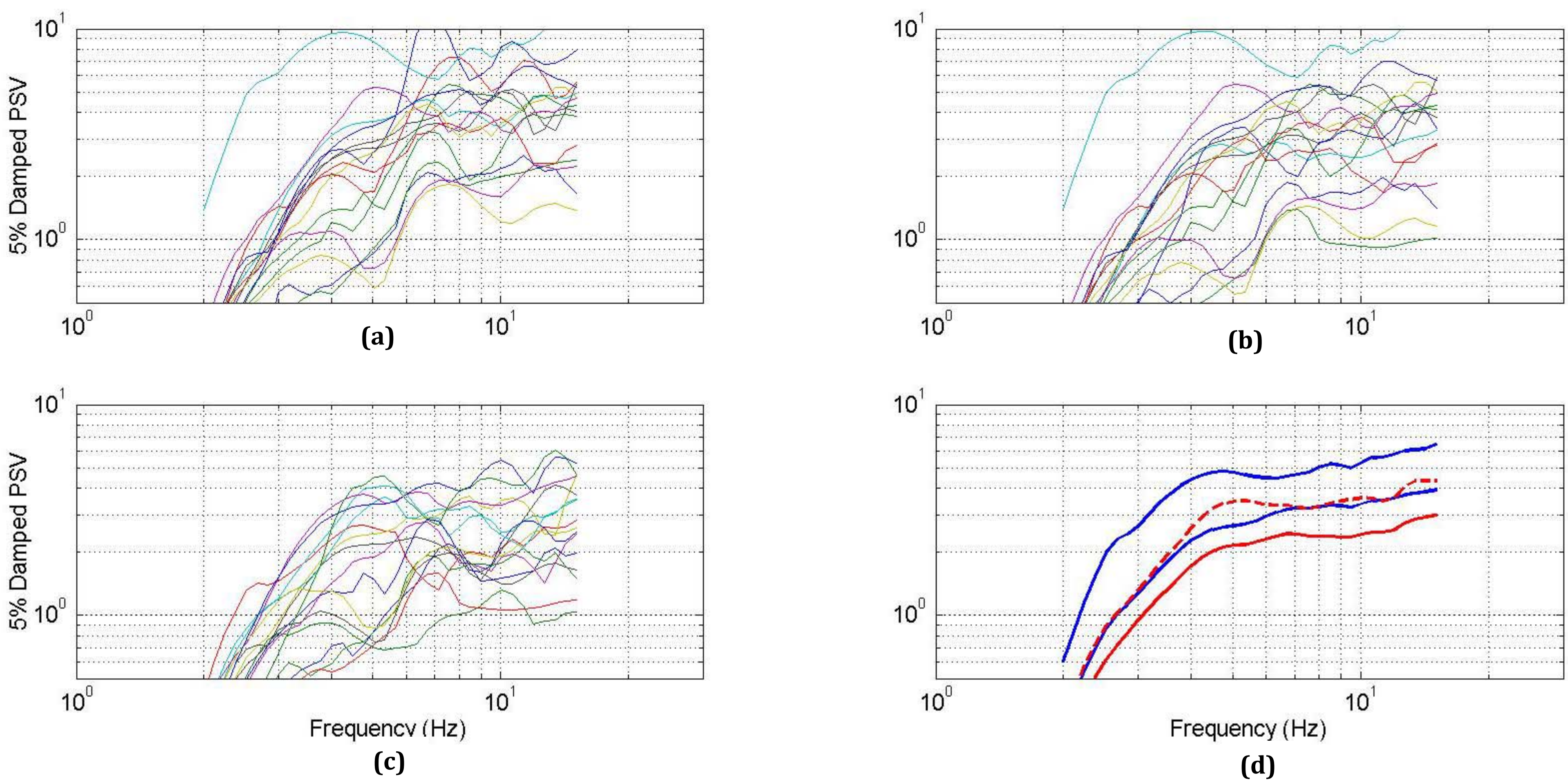


Figure 12. (a) 5% damped response spectra for the sites in the 0.4^0x0.4^0 area that recorded the June 18, 2002 southwestern Indiana earthquake. (b) Same response spectra, but corrected for attenuation and spreading, and scaled to a common 50 km epicentral distance. (c) Response spectra from (b), but corrected for the near-surface site effects as determined from the *S*-wave profiling and 1D linear approximation. (d) The means and mean-plus-one-standard-deviations for the normalized spectra (blue) shown in (b) and the site-corrected spectra (red) shown in (c).

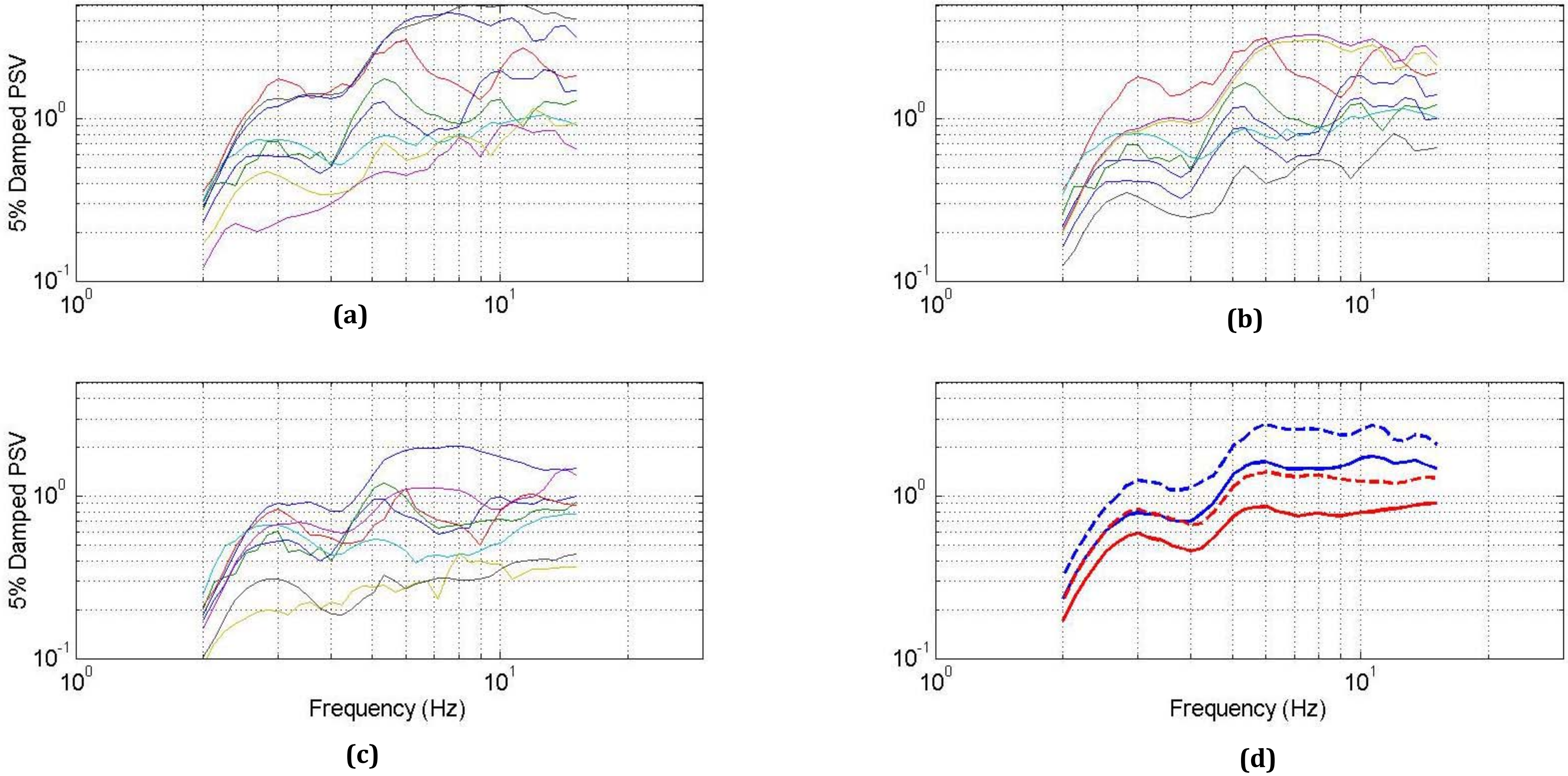


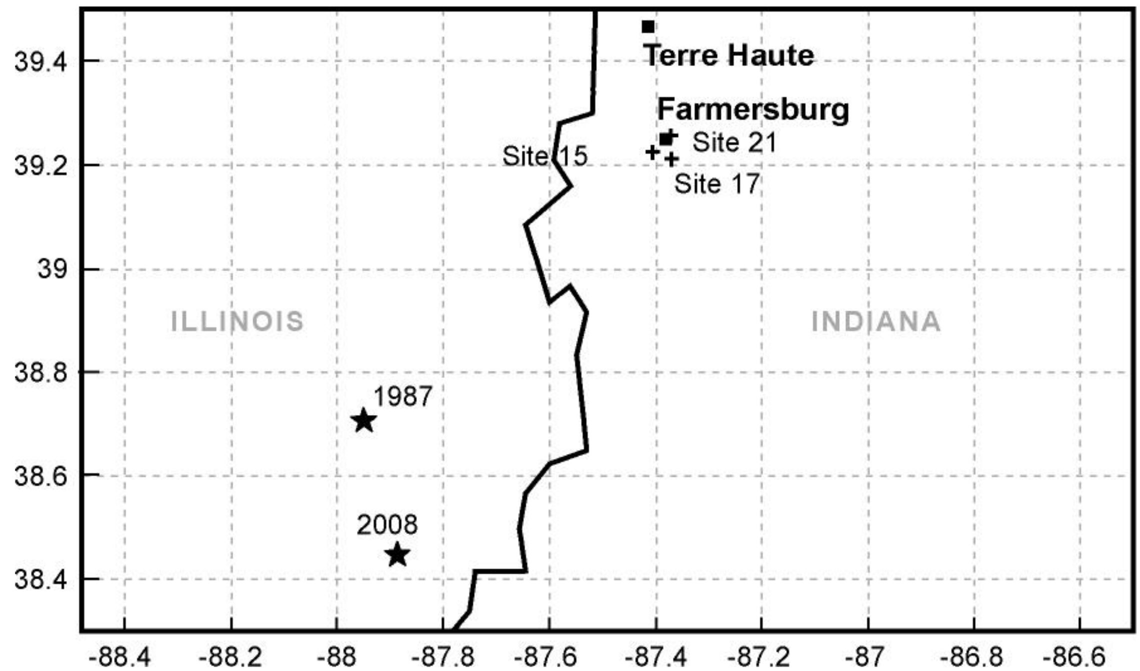
Figure 13. (a) 5% damped response spectra for the sites in the 0.4^0x0.4^0 area that recorded the M4.96 April 18, 2008 southeastern Illinois earthquake. (b) Same response spectra, but corrected for attenuation and spreading, and scaled to a common 50 km epicentral distance. (c). Same response spectra as (b), but corrected for near-surface site effects as determined from the *S*-wave profiling and 1D linear approximation. (d) The means and mean-plus-one-standard-deviations of the normalized spectra (blue) in (b) and the site-corrected spectra (red) in (c).

Discussion

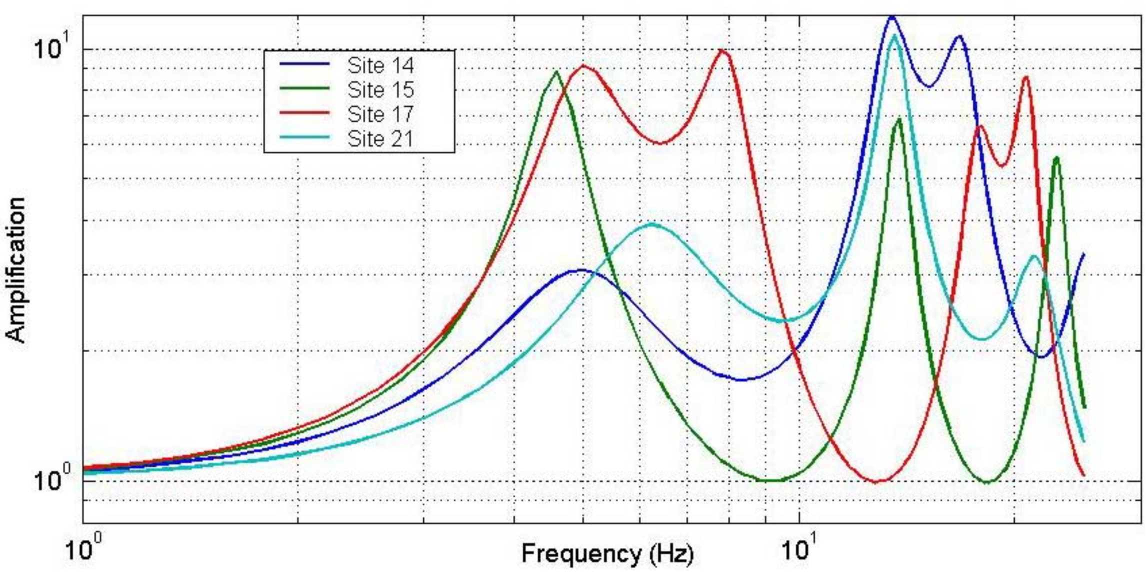
The lower Wabash River valley has a history of moderate (**M**5) earthquakes. For example, prior to the 10 June 1987 and 18 April 2008 earthquakes there were **M**5 or greater earthquakes in 1838, 1891, 1909, and 1968 (Stover and Coffman, 1993) (Fig. 2). In addition, Obermeier et al. (1991) and Munson et al. (1997) found liquefaction evidence of pre-historical earthquakes that Street et al. (2004) and Olson et al. (2005) concluded ranged between **M**6.2 and **M**7.3. Consequently, locally derived site conditions and ground motions are of interest for quantifying past, as well as possible future events.

Based on the observed PGV's for the 2008 earthquake sequence, the scatter in the amplitudes of the ground motions are similar to those recorded for the 1987 and 2002 earthquakes. The 2008 ground motions were recorded by the same instrument type and within the same geographical area and similar geological conditions as those recorded for the 1987 and 2002 earthquakes. Indeed, the site conditions are typical of sites found throughout much of the central United States with the exception of sites underlain by thick deposits of sediment overburden such as is found in the Mississippi embayment. Consequently, the role of the ground motion site effect for the 2008 earthquakes are likely typical of most events in the area.

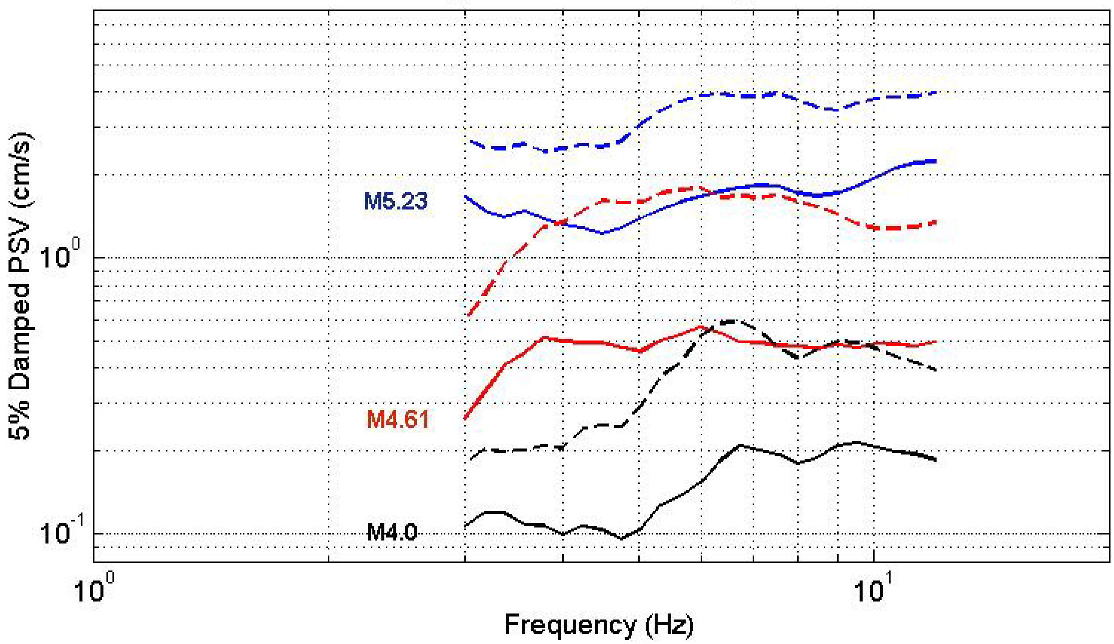
The spatial distribution of four blast monitors near Farmersburg, Indiana provides a basis for estimating the relative effectiveness of the linear one-dimensional site estimation. In the vicinity of this community, the **M**5.2, **M**4.6, and **M**4.0 earthquakes, with a single exception, were recorded at sites 14, 15, 17, and 21. All sites are within 5 km of one another (Figure 14a). The exception is Site 14 which failed to trigger for the **M**4.0 event. Based on the reflection/refraction seismic surveys, the depths to competent (i.e. SH-wave velocity >760 m/s) rock at the sites are 30, 9, 22, and 19 m, respectively (Fig. 4). Linear 1D site effect estimates for the sites are shown in Figure 14b. Figure 14c shows the average of the 5% damped response spectra at the four sites for the earthquakes with (solid lines) and without (dashed lines) site corrections. Spectra are corrected for spreading and attenuation to a common distance of 95 km. Figure 14d shows the ratio of the averaged site corrected to the uncorrected 5% damped response spectra for the earthquakes in Figure 14c. Based on the ratios, the 1D linear site corrections reduced the spectral amplitude for frequencies > 2.5 Hz between 40 and 70%.



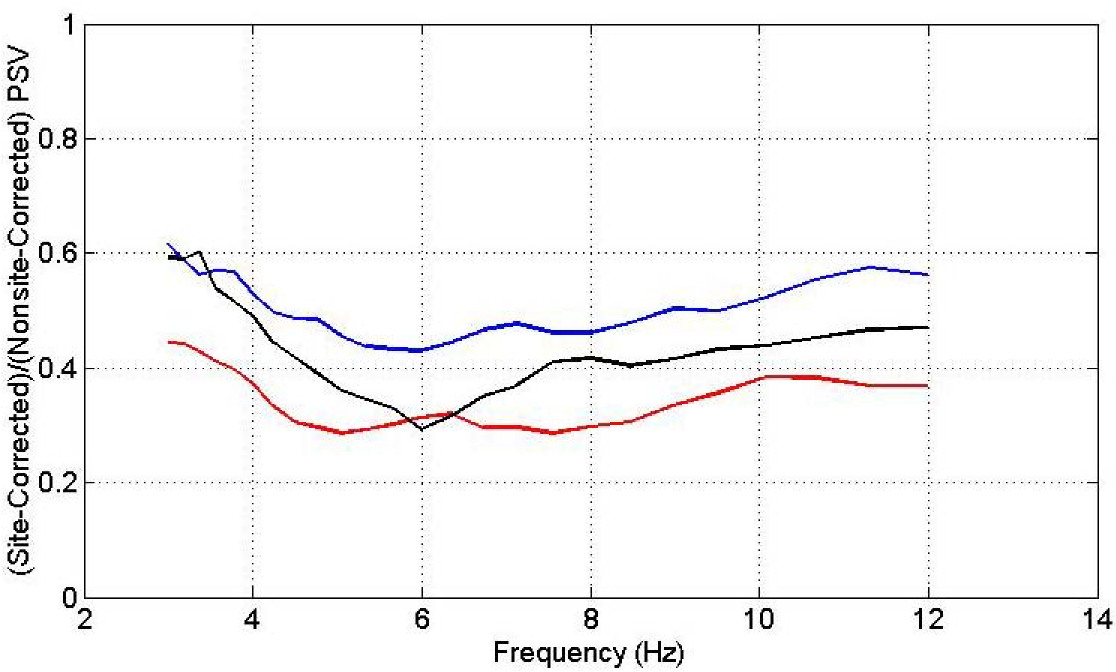
(a)



(b)



(c)



(d)

Figure 14. (a) Geographical location of sites 14, 15, 17, and 21. (b) SHAKE91 estimates of the site effects at the sites. (c) Average of the 5% damped response spectra, where the solid lines represent spectra that have been attenuation, spreading, and site effects. The dashed lines represent the same spectra corrected only for attenuation and spreading. All of the spectra have been corrected to a common epicentral distance of 95 km. (d) Ratios of the site corrected spectra to the non-site corrected spectra in (c).

Assuming that these sites are representative of what can achieved with site corrections using the traditional cost-effective non-invasive methods and one-dimensional modeling, more extensive site characterization is required. This includes more expensive and invasive methods that will collect material samples for laboratory testing of geotechnical index and dynamic properties, as well as provide subsurface exposure for higher-resolution in-situ field tests. Acquiring additional two-dimensional (or three-dimensional) seismic surveys at the site will provide a more defined image of near-field variation in the geological model. This also includes extending the imaging deeper into the bedrock, because, as noted by Abercombie (1977) and Thompson et al. (2009), crystalline rock can exhibit appreciable site response; therefore it is reasonable to speculate that multilayered sedimentary bedrock may also contribute to site effects not identified in this investigation. We must also consider the limitations of our methods. The primary limitation is the 3 to 100 Hz frequency range over which the data are valid. Because ground motions of engineering interest are typically limited to frequencies between 0.5 and 10 Hz, the frequency range of the blast monitor data imposes a constraint on the effectiveness at the low end.

Conclusions

The peak velocities and response spectra for the 2008 earthquake sequence are consistent with previous observations in the area, beginning with the 10 June, 1987 southwestern Illinois earthquake. This investigation and previous work for the southwestern Indiana 18 June, 2002 earthquake (Street et al., 2005, Woolery et al., 2009) include forty-two S-wave velocity soundings and associated linear one-dimensional site effect approximations. Results indicate that the corrections reduced the range of spectral amplitude for frequencies greater than 2.5 Hz between 40 and 70 percent, as well as the spectral variation by approximately a factor of 4. In addition, the data suggest that a peak ground velocity of 1.2 cm/s defines a clear boundary separating Modified Mercalli intensities IV and V. These observations can be useful in scaling ground motions of historical seismicity, as well as predicting the effects of future events. We speculate these quantitative characteristics are likely representative for site effects throughout the lower Wabash River valley, except for the infrequent thick-sediment filled sites (> 30 m). This representative area includes southwestern Illinois, southeastern Indiana, and the adjoining area in Kentucky.

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APPENDIX A

A1. Jan. 3, 2003, 16:17:07.00 UTC
37.83°N/88.09°W
M3.0 (unknown magnitude);

Location of Blast Monitor °N/°W	Epicentral Distance (km)	Azimuth (degrees)	Peak Horizontal Velocity (cm/s)
37.764/88.337	25.0	248	.8255
37.750/88.361	26.9	251	.1270
37.756/88.369	27.4	253	.1016
37.639/88.373	33.9	232	.1397
37.636/88.374	34.1	231	.1143

A2. May 2, 2003; 08:10:13 UTC
37.83°N/88.09°W
M3.0 (unknown magnitude)

Location of Blast Monitor °N/°W	Epicentral Distance (km)	Azimuth (degrees)	Peak Horizontal Velocity (cm/s)
37.756/88.369	34.1	134	.0635
37.779/88.337	34.5	128	.0953
37.750/88.361	35.0	134	.0508
37.639/88.373	43.8	146	.1270
37.636/88.374	44.1	147	.1143

A3. Jan. 2, 2006, 21:48:57 UTC
37.880°N/88.420°W
M3.6 (unknown magnitude)

Location of Blast Monitor °N/°W	Epicentral Distance (km)	Azimuth (degrees)	Peak Horizontal Velocity (cm/s)
37.7756/88.4080	7.18	173	.3874
37.7797/88.3920	7.10	160	.3115
37.7782/88.3810	7.63	153	.1461
37.7841/88.3560	8.33	138	.1334
37.7554/88.3683	10.3	154	.1905
37.7504/88.3608	11.2	152	.1524
37.7477/88.3626	11.4	154	.1143
37.7442/88.3710	11.4	158	.1842