Recordings from the Deepest Borehole in the New Madrid Seismic Zone

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INTRODUCTION

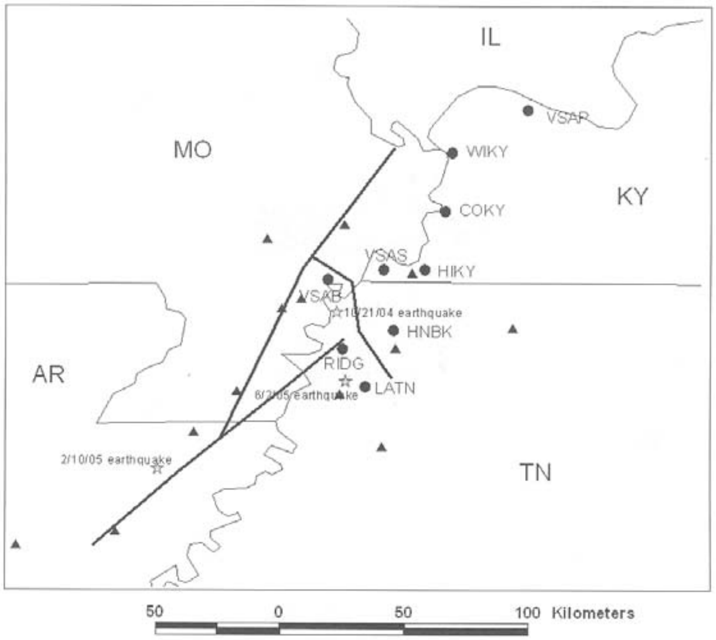
The New Madrid Seismic Zone (NMSZ) is the most active region in the central and eastern United States. At least three large earthquakes with magnitude greater than 7.0 occurred along the New Madrid faults (Figure 1) between winter 1811 and spring 1812 (Nuttli, 1973; Johnston, 1996; Hough *et al.*, 2000; Bakun and Hopper, 2004). The paleo-liquefaction records (Tuttle *et al.*, 2002) suggest that large earthquakes similar to the 1811–1812 events occurred twice in the past 1,200 years in the NMSZ. Although the causes of these large intraplate earthquakes are not well understood (Zoback, 1992; Newman *et al.*, 1999; Kenner and Segall, 2000), they pose certain hazards and risks because of their proximities to population centers such as Memphis, Tennessee, and St. Louis, Missouri.

Even though the 1811–1812 New Madrid events have been investigated intensively (Nuttli, 1973; Johnston, 1996; Hough *et al.*, 2000; Bakun and Hopper, 2004), the differences among the magnitude estimates by different researchers are significant, about 1.0 magnitude unit, because of the lack of instrumental recordings of the events. Our limited understanding of ground-motion attenuation and site amplification in the region further hinders efforts to assess magnitudes based on macroseismic effects. Therefore, obtaining strong ground-motion recording in the NMSZ remains a high priority for seismic hazard and risk assessment. In the fall of 1989, the University of Kentucky (UK) began to install and operate a strong-motion network in the northeastern part of the NMSZ (Figure 1). The Kentucky Seismic and Strong-Motion Network has recorded more than two hundred earthquakes with magnitudes ranging from 1.5 to $5.2m_{b,Lg}$ (Street and Wang, 2003; Wang *et al.*, 2003). The Lamont-Doherty Earth Observatory (LDEO) of Columbia University had previously installed and operated a strong-motion network in the New Madrid Seismic Zone between 1995 and 2000 (Street and Wang, 2003). The LDEO decommissioned the network in 2000 and gave two stations, HNBK and RIDG (Figure 1), to UK.

The University of Memphis and St. Louis University are the principal NMSZ regional broadband (Figure 1) and other network operators. Historically, their research has been focused generally on the fundamental seismotectonics of the central United States. Since the late 1980s, the research associated with the UK network has been primarily strong-motion seismology of engineering interest. Currently, the University of Kentucky operates a strong-motion network of nine stations in the New Madrid Seismic Zone (Figure 1). All the stations are accessed through dial-up telephone lines and equipped with 3- to 12-channel K2 digital recorders by Kinemetrics Inc.

A unique feature of the network is the inclusion of vertical strong-motion arrays, each with one or two downhole accelerometers. The first vertical strong-motion recording in the central and eastern United States was recorded at station VSAP from the 5 February 1994 southern Illinois earthquake (Street *et al.*, 1997). The deepest borehole array is 260 m (850 ft) below the surface at station VSAS in Fulton County, Kentucky (Figure 1). The vertical accelerometer array at VSAS consists of three 3-component accelerometers, recorded on a 24-bit, 12-channel K2 digital recorder equipped with GPS timing. The deep accelerometer (FBA-23DH) is at the bottom of the hole. The second (ES-152DH) is at the bottom of a 30-m geotechnical hole. The third (Episensor) is a free-field surface installation. The installation of the vertical strong-motion array at VSAS was completed in late 2003 (Woolery and Wang, 2002).

A preliminary surface seismic refraction and reflection survey was conducted at the site before drilling the hole at VSAS (Woolery and Wang, 2002). The depth to the Paleozoic bedrock at the site was estimated to be approximately 595 m, and the depth to the first very stiff layer (*i.e.*, Porters Creek Clay) was found to be about 260 m. These depths and stratigraphic interpretations correlated well with a proprietary seismic reflection line and the Ken-Ten Oil Exploration No. 1 Sanger hole (Schwalb, 1969), as well as our experience in the area (Street *et al.*, 1995; Woolery *et al.*, 1999). The Porters Creek Clay consists predominantly of montmorillonitic clay with minor amounts of interbedded sand and clay in the upper part of the formation (Olive, 1980). A significant shear-wave velocity increase appears at approximately 260 meters in depth. This stiff boundary correlates well with the anticipated Eocene-Paleocene stratigraphic horizon. A mud-rotary drilling company was contracted to drill the hole, and a 102-mm steel-cased borehole was completed to



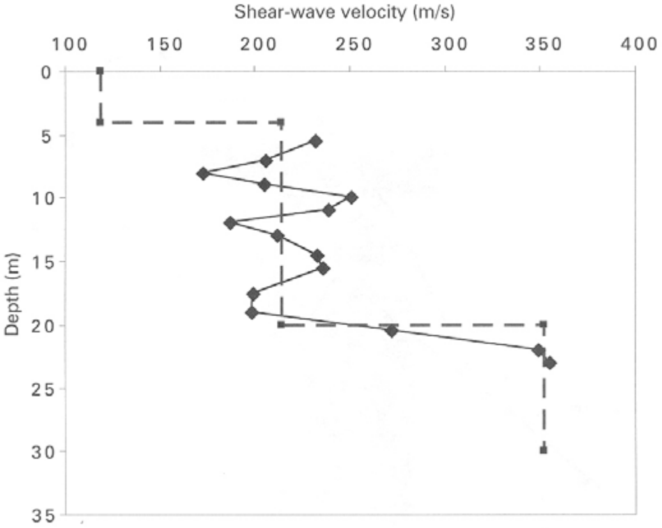
▲ **Figure 1.** The location of the Kentucky Seismic and Strong-Motion Network stations (filled circles) in relation to the New Madrid faults (heavy black lines, modified from Johnston and Schweig, 1996) and recent earthquakes (stars) recorded by the deep vertical strong-motion array, VSAS. The regional broadband station locations are shown by the filled triangles.

a depth of 260 meters (Woolery and Wang, 2002). The 30-m geotechnical hole was drilled with a 102-mm PVC-casing. The topmost 40 meters of sediment at the site were found to consist of loose to dense sands (SW/SP) and gravels (GW/GP). A downhole seismic survey was performed, and the result is shown in Figure 2.

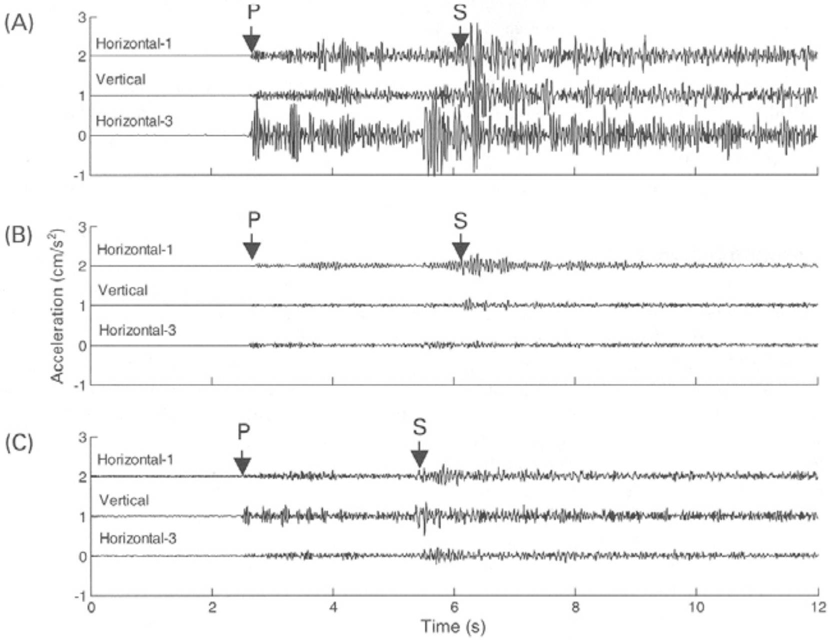
DEEP VERTICAL ARRAY RECORDINGS

Three earthquakes, the 21 October 2004 earthquake (M_d 2.5) near Tiptonville, Tennessee; the 10 February 2005 Arkansas earthquake (M_w 4.1); and the 2 June 2005 earthquake (m_{bLg} 4.0) near Ridgely, Tennessee, have triggered the array at VSAS. Table 1 lists the source parameters for the events. Figure 3 shows the acceleration-time histories at the surface, at a depth of 30 m, and at a depth of 260 m, respectively, from the 21 October 2004 earthquake. The recordings were filtered with a bandpass of 5.0 to 50 Hz. The arrival times for the direct *P*-wave are 2.645, 2.625, and 2.500 s at the surface, 30 m deep, and 260 m deep, respectively. The average *P*-wave velocity between the surface and 30 m is 1,500 m/s, between the surface and 260 m is 1,793 m/s, and between 30 m and 260 m is 1,840 m/s. These velocities are quite similar to those estimated from the *P*-wave reflections (Woolery and Wang, 2002). The arrival times for the direct *S*-wave are about 6.25, 6.10, and 5.35 s at the surface, 30 m, and 260 m, respectively. The average *S*-wave velocity between the surface and 30 m is 200 m/s, between the surface and 260 m is 289 m/s, and between 30 m and 260 m is 307 m/s. The average shear-wave velocity (200 m/s) of the top 30 m is very close to that shown on the downhole log (Figure 2) and the estimates from the *SH*-wave refractions and reflections (Woolery and Wang, 2002). However, the average shear-wave velocities (289 and 307 m/s) between the surface and 260 m and between 30 m and 260 m are significantly lower than those estimated from the *SH*-wave refractions and reflections (Woolery and Wang, 2002).

Figure 4 shows the acceleration-time histories at the surface, 30 m deep, and 260 m deep, respectively, from the 10 February 2005 Arkansas earthquake. The recordings were filtered with a bandpass of 1.0 to 30 Hz. No *P*-wave arrival was recorded because of insufficient pre-event memory. The arrival times for the *S*-wave are 5.00, 4.90, and 4.50 s at the surface, 30 m, and 260 m, respectively. The average *S*-wave velocity between the surface and 30 m is 300 m/s, between the surface and 260 m is 520 m/s, and between 30 m and 260 m is 575 m/s. The average shear-wave velocity (300 m/s) of the top 30 m is very close to that shown on the downhole log (Figure 2) and the estimates from the *SH*-wave refractions and reflections (Woolery and Wang, 2002). The average shear-wave velocities (520 and 575 m/s) between the surface and 260 m and 30 m and 260 m also are similar to those estimated from the *SH*-wave reflections (Woolery and Wang, 2002).



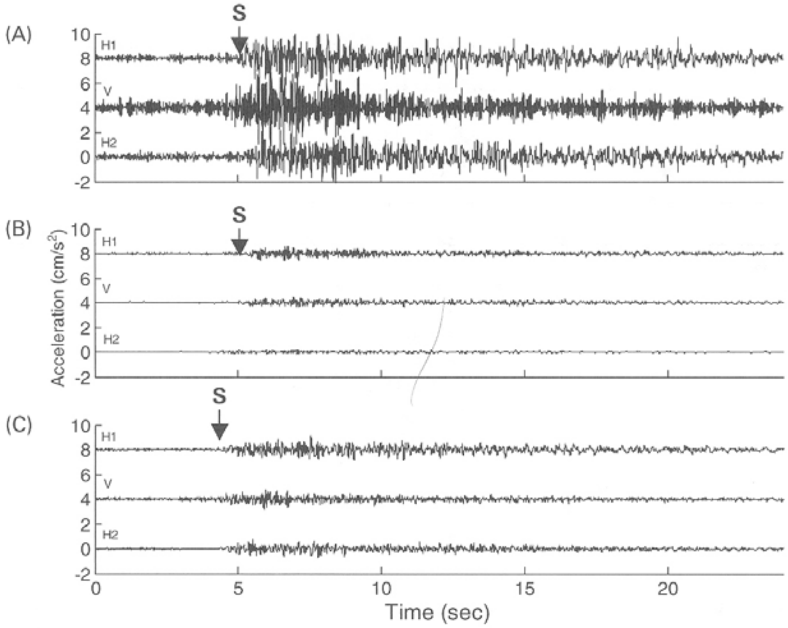
▲ **Figure 2.** Downhole shear-wave seismic log with a 3-point smoothing function applied (solid line) and the average shear-wave velocity (dashed line). Data were acquired with a three-component, 14-Hz geophone.



▲ **Figure 3.** Acceleration recordings from the 21 October 2004 Tiptonville, Tennessee, earthquake at vertical strong-motion array VSAS. (A) surface, (B) 30 m deep, (C) 260 m deep.

TABLE 1
Source Parameters for the Tennessee and Arkansas Earthquakes

Date	Time (UTC)	Magnitude	Latitude/Longitude	Depth (km)	Distance (km)
10/21/2004	11:58:38	M_d 2.5	36.40°N / 89.50°W	5.9	22
2/10/2005	14:04:54	M_w 4.1	35.76°N / 90.25°W	15.5	110
6/2/2005	11:35:11	m_{blg} 4.0	36.15°N / 89.47°W	14.9	47

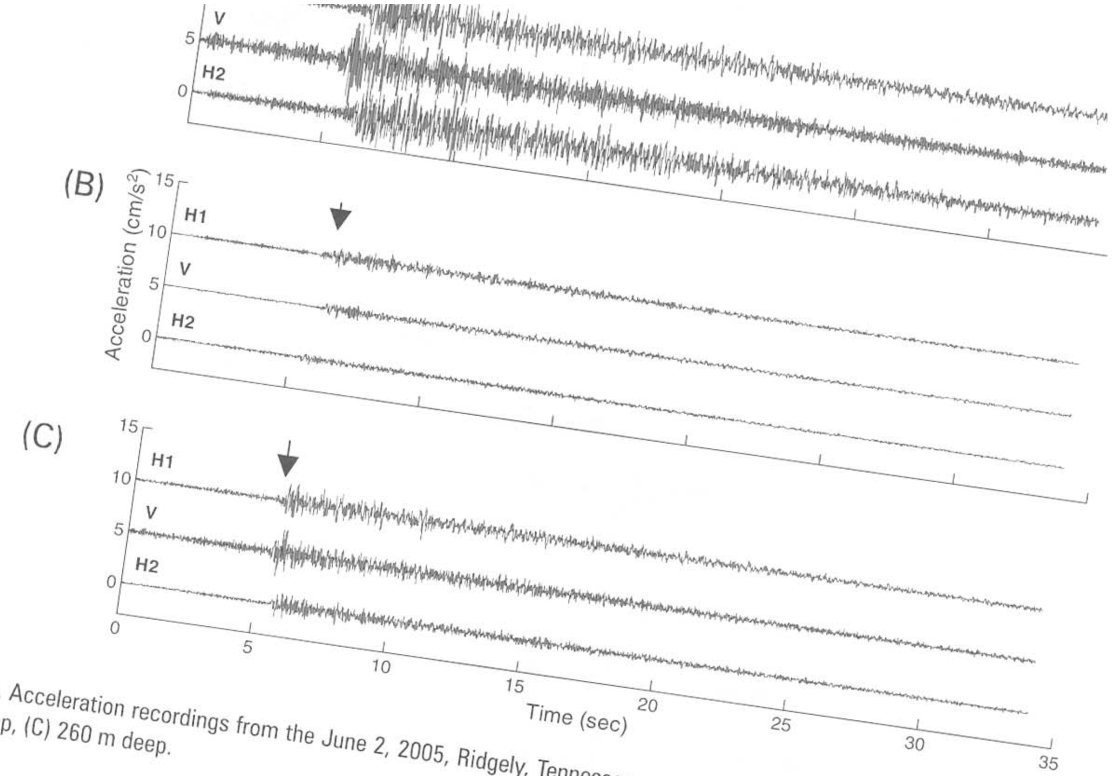


▲ **Figure 4.** Acceleration recordings from the February 10, 2005, Arkansas earthquake at vertical strong-motion array VSAS. (A) surface, (B) 30 m deep, (C) 260 m deep.

Figure 5 shows the acceleration- and displacement-time histories at the surface, 30 m deep, and 260 m deep, respectively, from the 2 June 2005 earthquake. The recordings were filtered with a bandpass of 1.0 to 40 Hz. *P*-wave arrivals are not clear on the records. The arrival times for the *S*-wave are 6.20, 6.10, and 5.60 s at the surface, 30 m, and 260 m, respectively. The average *S*-wave velocity between the surface and 30 m is about 300 m/s, between the surface and 260 m is 433 m/s, and between 30 m and 260 m is 460 m/s. The average shear-wave velocity (300 m/s) of the top 30 m is also very close to that shown on the downhole log (Figure 2) and the estimates from the *SH*-wave refractions and reflections (Woolery and Wang, 2002). The average shear-wave velocities (433 and 460 m/s) between the surface and 260 m and 30 m and 260 m are lower than those estimated from the *SH*-wave reflections (Woolery and Wang, 2002).

SUMMARY

The recordings at the deepest vertical strong-motion array (VSAS) from three small events, the 21 October 2004 Tiptonville, Tennessee, earthquake; the 10 February 2005 Arkansas earthquake; and the 2 June 2005 Ridgely, Tennessee, earthquake show some interesting wave-propagation phenomena through the soils: the *S*-wave is attenuated from 260 m to 30 m depth and amplified from 30 m to the surface. The *S*-wave arrival times from the three events yielded different shear-wave velocity estimates for the soils. These different estimates may be the result of different incident angles of the *S*-waves due to different epicentral distances. The epicentral distances are about 22 km, 110 km, and 47 km for the Tiptonville, Arkansas, and Ridgely earthquakes, respectively.



e 5. Acceleration recordings from the June 2, 2005, Ridgely, Tennessee, earthquake at vertical strong-motion array VSAS. deep, (C) 260 m deep.

e recordings show the usefulness of the borehole tion array. The vertical strong-motion arrays oper- University of Kentucky have started to accumulate that will provide a database for scientists and engi- dy the effects of the near-surface soils on the strong ion in the New Madrid Seismic Zone. More infor- ut the Kentucky Seismic and Strong-Motion n be found at www.uky.edu/KGS/geologichazards. cordings are available at ftp://kgsweb.uky.edu.

EDGMENTS

es, a Fulton County, Kentucky, landowner, has iversity of Kentucky a 5-year right of entry (with tend) to a small parcel of land for the purpose and operation of the vertical strong-motion SAS. The U.S. Geological Survey provided ng the boreholes through the NEHRP con- 101. The maintenance and operation of the and Strong-Motion Network are supported wealth of Kentucky. The U.S. Geological es partial support for the network through ct 04HQAG0013.

REFERENCES

Bakun, W. H. and M. G. Hopper (2004). Magnitudes and lo the 1811–1812 New Madrid, Missouri, and the 1886 C South Carolina, earthquakes, *Bulletin of the Seismological America* **94**, 64–75.

Hough, S., J. G. Armbruster, L. Seeber, and J. F. Hough (2000). Modified Mercalli intensities and magnitudes of the 1811/18 Madrid, central United States earthquakes, *Journal of Geo Research* **105**, 23,839–23,864.

Kenner, S. J. and P. Segall (2000). A mechanical model for int earthquakes: application to the New Madrid Seismic Zone, **28,** 2,329–2,332.

Johnston, A.C. (1996). Seismic moment assessment of earthquakes i ble continental regions—III. New Madrid 1811–1812, Charl 1886, and Lisbon 1755, *Geophysical Journal International* 314–344.

Johnston, A.C. and E.S. Schweig (1996). The Enigma of the New Ma Earthquakes of 1811–1812, *Annual Review of Earth and Plane Sciences* **24**, 339–384.

Newman, A., S. Stein, J. Weber, J. Engeln, A. Mao, and T. Dixon (199 Slow deformation and low seismic hazard at the New Madrid seism zone, *Science* **284**, 619–621.

Nuttli, O. W. (1973). The Mississippi Valley earthquakes of 1811 1812: intensities, ground motion and magnitudes, *Bulletin of th Seismological Society of America* **63**, 227–248.

Olive, W. W. (1980). Geologic maps of the Jackson Purchase region, Kentucky: U.S. Geological Survey Miscellaneous Investigation, I-1217, 1 sheet and 11-page pamphlet.

Schwalb, H. R. (1969). Paleozoic geology of the Jackson Purchase Region, Kentucky: Kentucky Geological Survey, ser. 10, Report of Investigation 10, 40 pp.

Street, R. and Z. Wang (2003). Analysis of strong-motion records from the University of Kentucky accelerometers in the New Madrid Seismic Zone: 1990 through 2001: Final report prepared for the U.S. Geological Survey, NEHRP Award 02HQGR0016, 8 pp.

Street, R., E. Woolery, Z. Wang, and J. Harris (1995). A short note on shear-wave velocities and other site conditions at selected strong-motion stations in the New Madrid Seismic Zone, *Seismological Research Letters* **66**, 56–63.

Street, R., Z. Wang, E. Woolery, J. Hunt, and J. Harris (1997). Site effects at a vertical accelerometer array near Paducah, Kentucky, *Engineering Geology* **46**, 349–367.

Tuttle, M. P., E. S. Schweig, J. D. Sims, R. H. Lafferty, L. W. Wolf, and M. L. Haynes (2002). The earthquake potential of the New Madrid seismic zone, *Bulletin of the Seismological Society of America* **92**, 2,080–2,089.

Wang, Z., E. Woolery, and J. Schaeffer (2003). A Short Note on Ground-Motion Recordings from the June 18, 2002, Darmstadt, Ind., Earthquake, *Seismological Research Letters* **72**, 148–152.

Woolery, E., R. Street, Z. Wang, J. Harris, and J. McIntyre (1999). Neotectonic structures in the central New Madrid Seismic Zone: Evidence from multimode seismic reflection data, *Seismological Research Letters* **70**, 554–576.

Woolery, E. and Z. Wang (2002). A comprehensive geotechnical investigation and installation of a borehole accelerometer array in the New Madrid Seismic Zone: Final report prepared for the U.S. Geological Survey, NEHRP Award 02HQGR0101, 13 pp.

Zoback, M. L. (1992). Stress field constraints on intraplate seismicity in eastern North America, *Journal of Geophysical Research* **97**, 11,761–11,782.

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