Initial Assessment of the Intensity Distribution of the 2011 M_w 5.8 Mineral, Virginia, Earthquake

by Susan E. Hough

INTRODUCTION

The intensity data collected by the U.S. Geological Survey (USGS) "Did You Feel It?" (DYFI) Website (USGS, DYFI; http://earthquake.usgs.gov/earthquakes/dyfi/events/se/082311a/us/index.html, last accessed Sept 2011)for the M_w 5.8 Mineral, Virginia, earthquake, are unprecedented in their spatial richness and geographical extent. More than 133,000 responses were received during the first week following the earthquake. Although intensity data have traditionally been regarded as imprecise and generally suspect (e.g., Hough, 2000), there is a growing appreciation for the potential utility of spatially rich, systematically determined DYFI data to address key questions in earthquake ground-motions science (Atkinson and Wald, 2007; Hauksson *et al.*, 2008). DYFI intensities for the Mineral, Virginia, earthquake provide a unique opportunity to explore the variability of wave propagation and site amplification along the Atlantic seaboard of the United States, a region where instrumental recordings of moderate and large earthquakes are limited. The raw DYFI data suggest some general patterns, including not only significant apparent site response along Chesapeake Bay, in parts of the District of Columbia (D.C.), and elsewhere, but also source directivity and more efficient propagation along the predominantly southwest–northeast-striking tectonic fabric. These suggested patterns can be explored further by considering intensity residuals relative to an established intensity ground-motion-prediction equation (e.g., Atkinson and Wald, 2007).

The M_w 5.8 Mineral, Virginia, earthquake occurred at 17:51:04.59 UTC on 23 August 2011, one of the largest earthquakes in the region in historical times and the largest earthquake to strike the central/eastern United States (CEUS) in 70 yr. Shaking was widely felt through several major metropolitan areas, including the greater Washington, D.C., region, Philadelphia, and parts of New York State. The overall felt extent of the earthquake was enormous, with perceptible shaking reported as far west as Minnesota and as far south as Florida. To the northeast it was felt as far as Fredericton, New Brunswick, Canada (http://www.earthquakescanada.nrcan.gc.ca; last accessed March 2012). The felt extent was significantly bigger than that of the 1897 Giles County, Virginia, earthquake (Bollinger and Hopper, 1971; Nuttli *et al.*, 1979; Bollinger and Wheeler, 1983; Fig. 1a,b, present article). Nuttli *et al.* (1979) determine m_b 5.8 for this earthquake; the National Earthquake Information Center (NEIC) lists this event as M 5.9, the magnitude type not specified (http://earthquake.usgs.gov/earthquakes/states/events/1897_05_31.php; last accessed September 2011).

Instrumental recordings of the Mineral earthquake will provide valuable additional constraints for central/eastern North America ground-motion relations and potential site response. There is a unique opportunity to consider these data in combination with the intensity data collected for the earthquake. Compared to the relatively limited number of instrumental recordings of the earthquake, well-calibrated modified Mercalli intensity (MMI) values calculated from responses to the USGS Community Internet Intensity Map, also known as "Did You Feel It?," Web site (Wald *et al.*, 1999) provide orders of magnitude better spatial sampling. Although not an instrumental measure of ground motions, Atkinson and Wald (2007) show that DYFI intensities provide a surprisingly stable indicator of quantitative ground-motion parameters such as peak ground acceleration (PGA).

The DFYI response for the Mineral, Virginia, earthquake was unprecedented, with over 133,000 responses in the first week reported from a total of over 8600 ZIP codes in the United States and towns in Canada. DYFI responses can be geocoded to improve the spatial resolution of locations. However, for most of the area spanned by this data set population density is high enough that intensity responses averaged within ZIP codes provide good spatial correlation to the location of the actual reporting sites. Geocoded locations can also only be determined for a subset of the total responses. I thus used the averages within ZIP codes in our subsequent analysis. The DYFI-intensity locations, geocoded or otherwise, are fundamentally imprecise, which results in limitations in the ability to explore site response. Nonetheless, the unprecedented spatial richness of the DYFI data set for this earthquake provides a unique opportunity to explore the factors that influence wave propagation and therefore shaking intensity for a central/eastern North America earthquake, and to provide insight into analyses that might profitably be investigated further with available instrumental data.

ANALYSIS

To analyze the distribution of intensities for the Mineral, Virginia, earthquake, I first followed the approach of Hauksson *et al.* (2008), who considered the intensity distribution of the 2008 M_w 5.4 Chino Hills, California, earthquake. Hauksson

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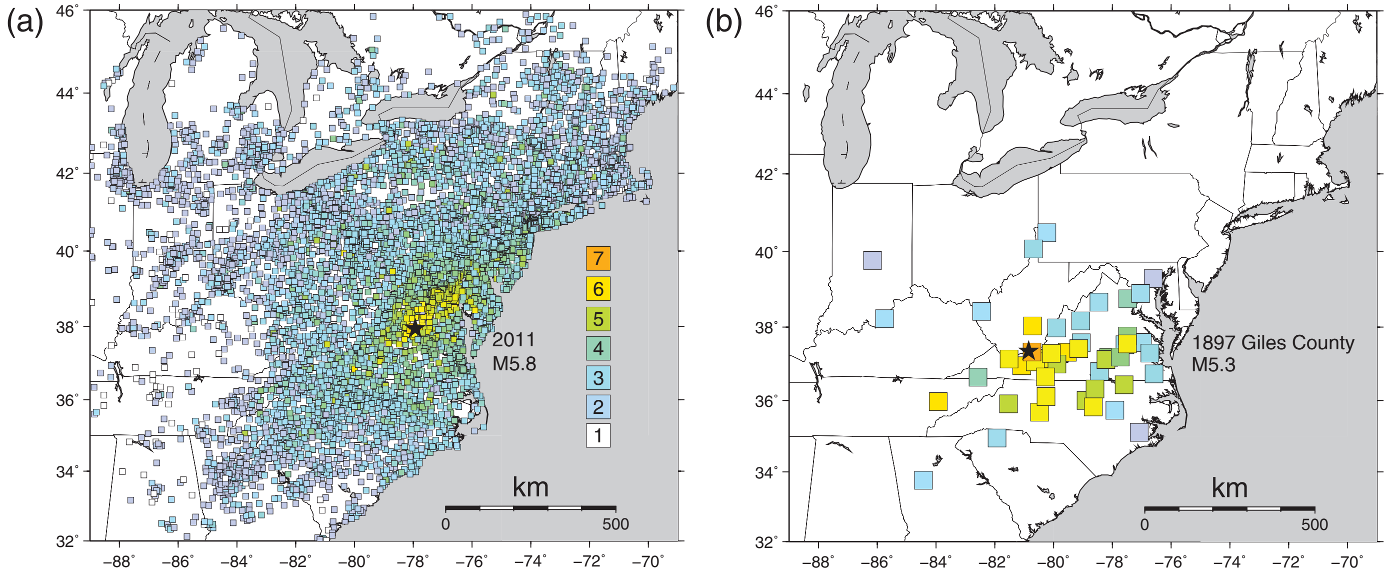


▲ **Figure 1.** (a) DYFI Intensities for the 23 August 2011 Mineral, Virginia, earthquake; each symbol is located at the geographical center of the ZIP code (or Canadian city) of the corresponding account (http://earthquake.usgs.gov/earthquakes/dyfi/events/se/082311a/us/index.html; last accessed September 2011). Color scale (also used for b) indicates MMI values plotted in discrete steps. For DYFI intensities, which are reported in decimal values, 1 = not felt; 2 = 2.0–3; 2.9; 3 = 3.0–3.9; etc. (b) Intensities for the 1897 Giles County, Virginia, earthquake; modified slightly from assignments by Hopper and Bollinger (1971).

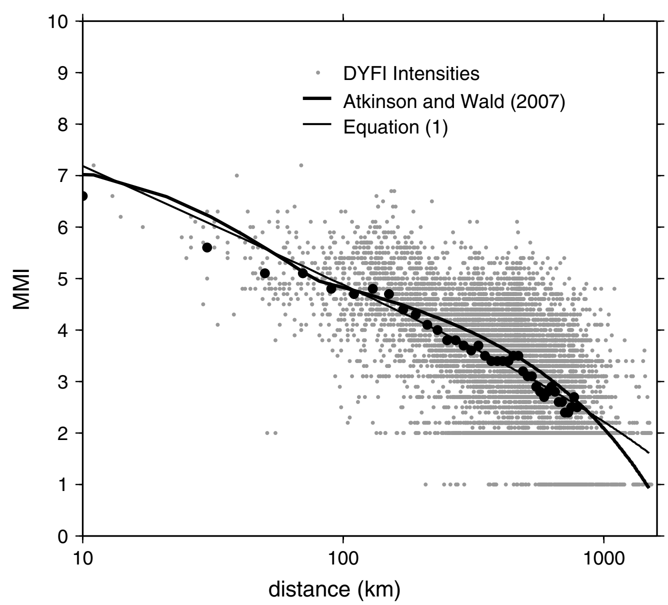
et al. (2008) fit a distance decay to the DYFI MMI(r) values using the standard functional form for the intensity-attenuation relation:

$$\mathrm{MMI}_p(r) = a - br - c \log_{10}(r), \tag{1}$$

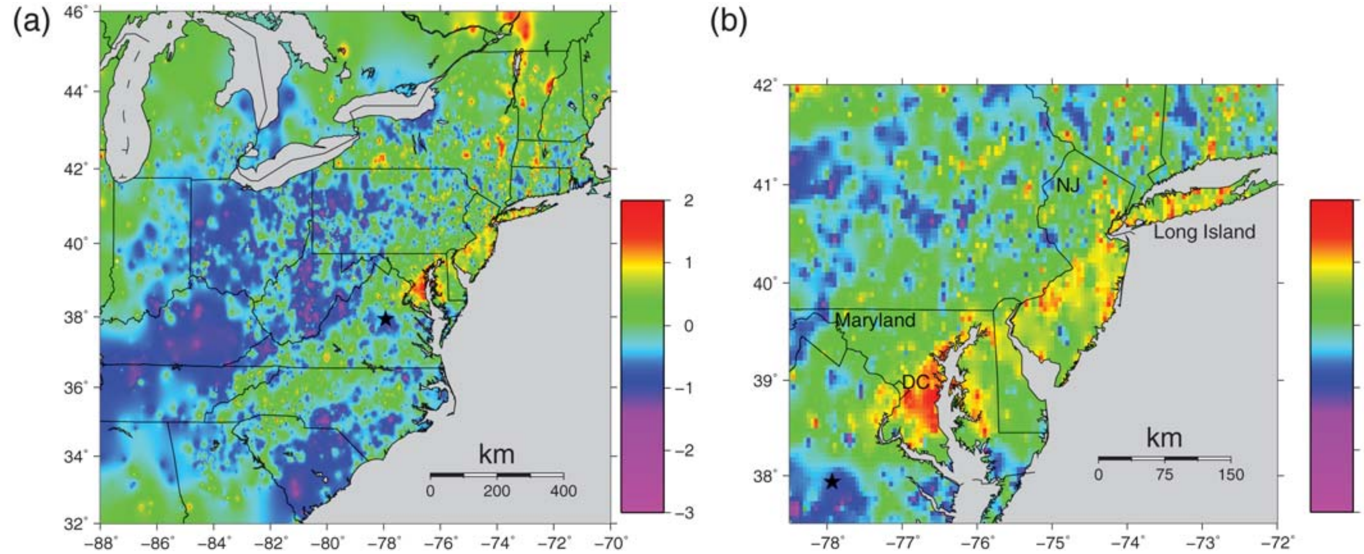
where a, b, and c are constants determined by a least-squares fit to the observations and r is epicentral distance. For the Chino Hills earthquake, the DYFI intensities revealed a generally smooth decay with distance. Hauksson *et al.* (2008) calculated residual intensities, δMMI, as the difference between the observed values and the prediction from equation (1) using parameters constrained by a least-squares regression.

For the Mineral, Virginia, earthquake, I first calculated residuals using the same procedure. A least-squares fit to the values yielded $a = 9.45$, $b = -0.00043$, and $c = -2.66$. I then calculated residuals relative to predicted MMI values for M 5.8 using the intensity-attenuation relationship developed for the central United States by Atkinson and Wald (2007). The latter differs from equation (1) in several respects; of particular importance here is that it includes a piecewise distance decay, which provides a better fit to the slow observed decay of MMI values between roughly 100 and 500 km (Fig. 2). For the subsequent analysis, I thus considered the intensity residuals, δMMI, relative to predicted values using the Atkinson and Wald (2007) relation, calculated simply as observed minus predicted values. Calculating residuals relative to predictions for a generic M 5.8 earthquake left open the possibility of a significant event term associated with this particular earthquake. The average residual, or event term, was small, −0.168, indicating the close correspondence between the observed intensities and predicted values for M 5.8. (The fit is slightly better, with an average residual = −0.04, for M 5.9).

Figure 3 presents a map of δMMI values, corrected for the event term, at the 8626 ZIP codes and cities for which DYFI intensities are available. The values were contoured using the surface utility of Generic Mapping Tools (GMT) (Wessel and



▲ **Figure 2.** DYFI intensities (gray circles), averaged values within 20-km bins out to a distance of 800 km (black circles) and predicted MMI using the relation of Atkinson and Wald (2007) (dark line) as well as equation (1) (light line).



▲ **Figure 3.** (a) Distribution of DYFI-intensity residuals relative to values predicted using Atkinson and Wald (2007) intensity-attenuation relationship, corrected for event term (see text). Swath of relatively high residuals SW of epicenter follows the Appalachians. Color scale indicates difference between observed and predicted values (MMI units). (b) Close-up view of (a).

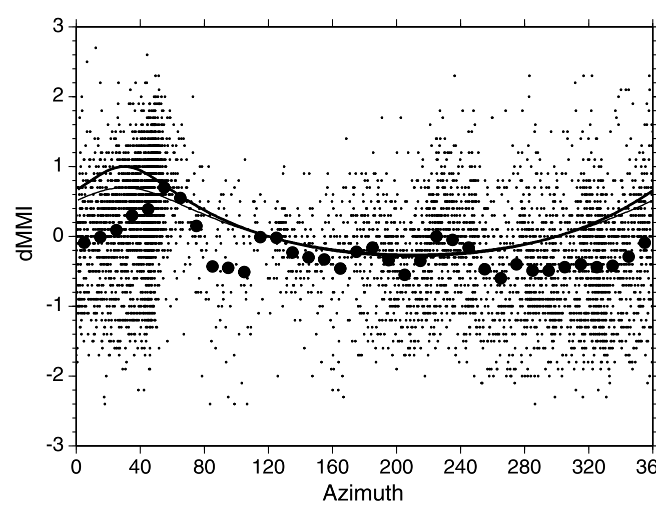
Smith, 1991). This algorithm uses a tension factor T to control the degree of curvature. The minimum curvature solution, $T = 0$, can generate oscillations, whereas $T = 1$ will generate a solution with no maxima or minima away from control points. Here I used $T = 1$, with a grid-spacing for plotting of 3 minutes.

The spatial distribution of residuals suggests several first-order patterns. First, generally higher residuals in the northeast quadrant are consistent with directivity to the northeast. Second, the distribution of residuals in the southwest quadrant suggests more efficient propagation along versus across the predominant strike of the tectonic fabric which follows the Appalachians (also evident in Fig. 1). Finally, the overall distribution of residuals provides *prima facie* evidence for significant site response in many locations. The striking concentration of high residuals in the D.C./Chesapeake region might also be partly due to energy from Moho bounce arrivals, which have been inferred to contribute significantly to shaking at distances of 70–200 km in eastern North America (Atkinson, 2004). The bin-averaged intensity values shown in Figure 2 reveal a small bump upward at distances of 130–150 km; however, several different factors clearly contribute to this small signal.

The individual and binned-average residuals are shown as a function of azimuth in Figure 4. Following Seekins and Boatwright (2010), I compared the distribution to predicted directivity, D, using the equation of Ben-Menaham (1961):

$$D = [1 - \nu/\beta \cos(\theta)]^{-1}, \tag{2}$$

where ν/β is the ratio of the rupture velocity to the shear-wave velocity and θ is the angle relative to the strike direction (28°). A comparison of observed residuals and predicted directivity for ν/β values of 0.8 and 0.9, assuming that intensities are proportional to the logarithm of the shaking intensity, is shown in Figure 4. The predicted curves match the trends in the observed residuals relatively well, but the influence of other factors is suggested as well. First, the highest residuals toward the northeast do not coincide exactly with the strike direction, but at azimuths of 40°–50°. This likely reflects the concentration of sediment sites, and therefore site amplification, toward the northeast. Second, the influence of preferred propagation along the predominant strike of the tectonic fabric is illustrated by the distribution of residuals in the southwest quadrant: residuals at azimuths of 220°–230° are elevated by approximately



▲ **Figure 4.** Intensity residuals as a function of azimuth for distances less than 600 km (small circles), including binned averages within 10-degree bins (large circles). Also shown is predicted directivity for assumed ν/β values of 0.8 (light line) and 0.9 (dark line).

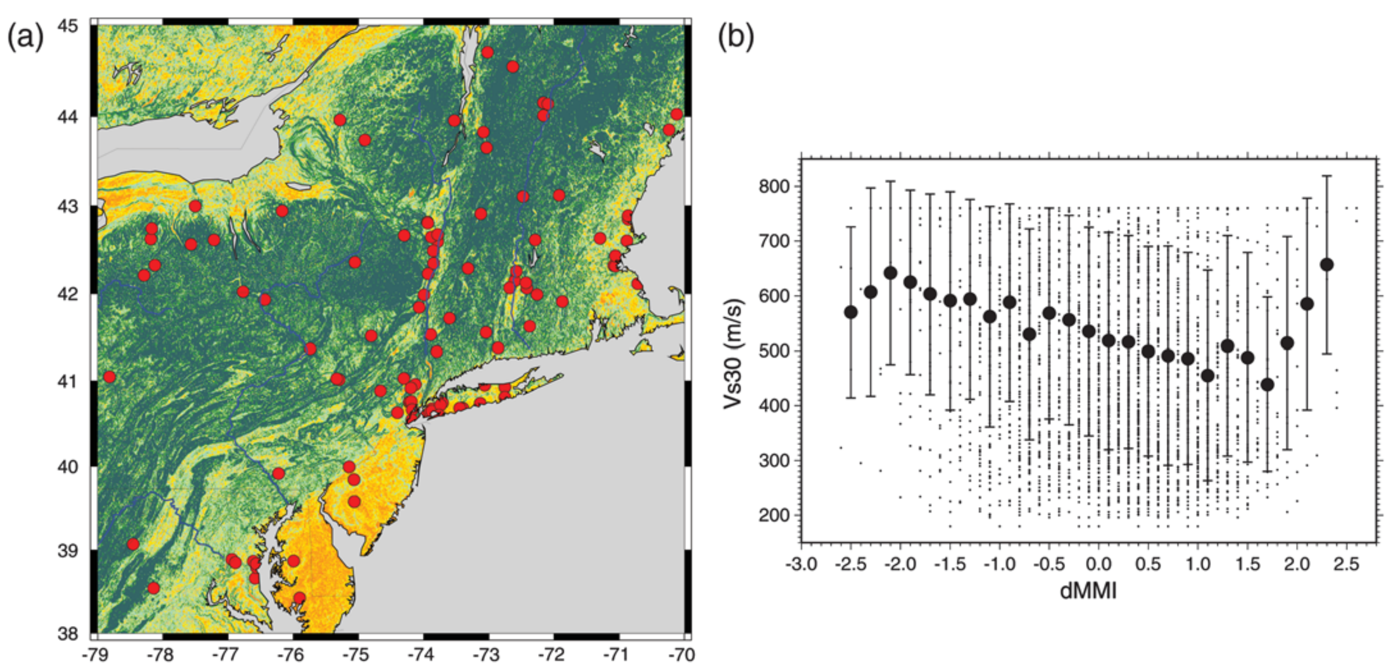
0.6 intensity units relative to those observed at other azimuths toward the southwest. It is perhaps surprising that the residuals do not reveal the expected amplification by the coastal plain sediments along the mid/south Atlantic coast (Chapman *et al.*, 1990); this signal might be obscured by the combined effects of the other two factors.

One can reasonably assume that amplification associated with preferred regional propagation is similar to the northeast as to the southwest. Figure 4 thus suggests that the effects of preferred propagation direction and directivity together serve to influence shaking intensities by as much as 1.0–1.2 units, with roughly comparable contributions from each effect.

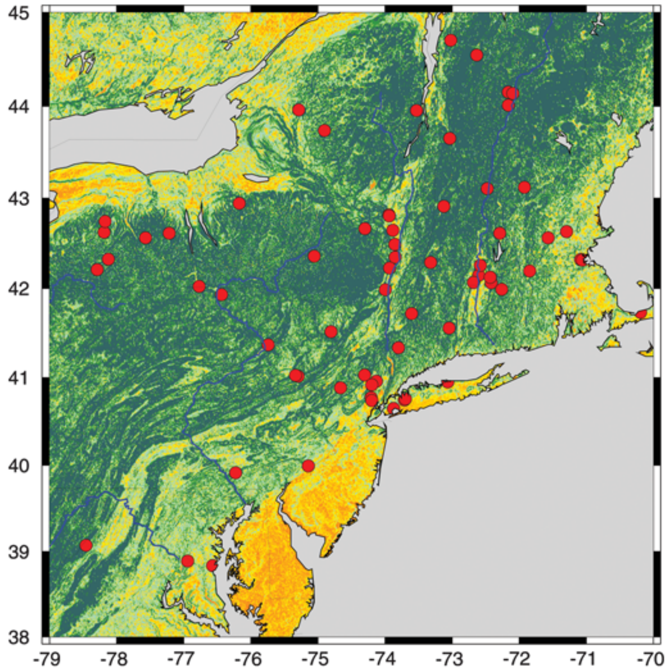
One might consider a joint inversion of residuals for site, propagation and directivity effects. However, such an approach would be complicated by the fact that the distribution of near-surface geology, and therefore expected residuals, is not uniform with azimuth. To focus on site response, I therefore focused on the northeast quadrant only (Fig. 3b), in an effort to isolate site effects from propagation and source effects, noting that the amplitudes of the residuals in this region are expected to reflect all three factors. The amplitude of the residuals might therefore not correspond directly to site response; the approach was used to explore the distribution of residuals. Figure 5a shows the locations of intensity residuals greater than 0.9 units superimposed on a map of V_{S30} determined using the topographic slope method of Allen and Wald (2009). The distribution suggests a correlation of high residuals and low V_{S30} values. Significant amplification is suggested, for example, around Chesapeake Bay, in the D.C. area, in the greater New York City region (including Long Island), in many coastal regions (including the Boston area), and along some major river valleys (including the Hudson and Connecticut). A comparison of individual residuals and V_{S30} values (Fig. 5b) reveals enormous scatter but a good correlation between bin-averaged intensity residuals and V_{S30}. For all but the highest residuals, the negative correlation is consistent with expectations; that is, stronger shaking at sites with low V_{S30} values.

The positive correlation between residuals greater than roughly 1.6 and V_{S30} values for the highest residuals is not expected and begs explanation. There are relatively few observations with residuals above 1.5, so the averages are relatively poorly constrained. It is possible, however, that the correlation is real: high intensities on relatively hard-rock sites could be due to topographic amplification effects, which many recent studies have shown can amplify shaking by upward of a factor of 2 at frequencies of engineering concern (e.g., Boore, 1973; Sánchez-Sesma, 1985; Lee *et al.*, 2009); alternatively, it is possible that shaking at hard-rock sites will be more strongly felt due to the relatively efficient propagation of high-frequency energy at hard-rock sites. It is also possible, however, that the apparent trend is simply due to a fundamental limitation of DYFI intensity analysis: the V_{S30} values, which are inherently uncertain, are determined for the center of the ZIP code and might not be representative of the V_{S30} values from which the DYFI reports are taken. Figure 6 shows the distribution of locations in the northeast quadrant for which estimated intensity residuals are greater than 1.6 and the estimated V_{S30} values are higher than 400 m/s.

Some of the locations shown in Figure 6 are plausible instances of misassigned V_{S30} values, for example, in Long



▲ **Figure 5.** (a) Residual MMI (dMMI) values greater than 0.9 (red circles) are superimposed on a V_{S30} map inferred using the topographic slope method of Allen and Wald (2009). (b) VS30 values plotted against residuals for individual locations (black circles), with binned averages (large circle; error bars indicate ± one standard deviation).



▲ **Figure 6.** Locations (red circles) where intensity residuals are greater than 1.6 and estimated V_{S30} values are greater than 400 m/s.

Island, coastal Massachusetts, and around Chesapeake Bay. Some locations cluster along the edge of topographic features, perhaps suggestive of topographic amplification. Given the fundamental and likely irreducible uncertainties, I conclude that it is not possible to draw meaningful conclusions about these values. I note that residuals greater than 1.6 are found for only 196 of the 8,626 locations; the bulk of the data thus reveals a good average correlation between residuals and V_{S30} values, albeit with enormous scatter.

REVISITING THE 1897 GILES COUNTY, VIRGINIA, EARTHQUAKE

In recent years a number of methods have been developed to quantitatively and rigorously analyze the intensity distributions of historical earthquakes (e.g., Bakun and Wentworth, 1997; Johnston, 1996). Analysis of historical earthquakes in central/eastern North America has long been hindered by a shortage of moderate and large instrumentally recorded calibration events. The rich intensity data set for the 2011 Mineral, Virginia, earthquake provides an opportunity to revisit key historical earthquakes. In particular, the intensity distribution can be compared with that of the 31 May 1897 Giles County, Virginia, earthquake, for which a magnitude of 5.9 had been estimated. Accounts of this earthquake were published by Hopper and Bollinger (1971); reports are summarized on the Virginia Tech (VaTech) Web site http://www.geol.vt.edu/outreach/vtso/Giles-Intensity.html (last accessed September 2011). I revisited the accounts and the intensity assignments and found that a reinterpretation of the intensity values was consistent with the assignments made by Hopper and Bollinger (1971) with only minor departures, although I assigned fewer values because not all accounts are judged to provide sufficient information (Table 1).

A common question with analysis of historical earthquakes is whether the low-intensity field is fully characterized by extant archival sources, because relatively subtle effects are presumably less likely to be reported. For the 1897 Giles County earthquake, the extent of the felt area appears to be fairly well constrained. For example, contemporary sources note the farthest locations that shaking was felt to the north and south: mid-Maryland and Savannah, Georgia, respectively. In Indianapolis, Indiana, an account specifically notes that the earthquake was most noticeable in tall buildings. A similar account is available concerning Baltimore.

Of particular note is the clear indication that the felt extent of the Giles County earthquake extended less far to the west than that of the Mineral, Virginia, earthquake, even though the former event was centered west of the latter (see Fig. 1). Overall, a comparison of intensity distributions suggests that the Giles County earthquake was smaller than M_w 5.8.

Interestingly, Figure 1b also suggests a qualitative difference between the two intensity distributions: the areal extent of moderate intensity values (V–VII) is larger relative to the overall felt extent for the Giles County earthquake than the Mineral, Virginia, event. The comparison of intensities for a historical earthquake and DYFI intensities does raise the question of consistency between the two methods for assigning intensity values. To explore this issue, I used the DYFI algorithm to assign intensity values for some of the more detailed accounts of the Giles County earthquake. For example, in Roanoke, Virginia, as well as in other locations, accounts describe a similar suite of effects: "crockery rattled, doors swung, furniture moved in many houses, several chimneys knocked down, people rushed outside, pictures shaken from walls and bottles from shelves." A key question with accounts like this is the extent of chimney damage: it is often unclear if only a small number of weak chimneys were damaged or if damage was more extensive. Translating these accounts into reasonable answers to the DYFI questionnaire generally confirmed the subjective MMI values of VII. Similarly, in a number of locations including Petersburg, Virginia, accounts describe more moderate effects: "houses shaken (but no damage), crockery rattled, small objects moved/toppled, people frightened." The DYFI algorithm assigned intensity values around V, again consistent with the subjective assignments. At a large number of locations, relatively modest effects are described; for example, in Washington, D.C., "chandeliers swayed, floors trembled, felt distinctly, mostly noticed in tall buildings." At some of these locations, accounts describe rattling of windows and/or hanging objects. For these accounts, although the subjective assignments are generally around IV, the DYFI algorithm assigns values of II. (The DYFI questionnaire asks if sounds were heard and if hanging pictures were moved or fell but does not ask specifically about rattling of windows or hanging objects.)

Table 1
Locations and Intensities for the 1897 Giles County, Virginia, Earthquake

Location	Latitude (°)	Longitude (°)	MMI*	MMI_d †	MMI_{HB} ‡
Pearisburg, VA	37.326	–80.735	7.5	7.5	8.0
Roanoke, VA	37.270	–79.941	6.5	6.5	8.0
Bedford City, VA	37.315	–79.533	6.5	6.5	7.0
Pulaski, VA	38.025	–80.717	6.5	6.5	7.0
Radford, VA	37.150	–80.555	6.5	6.5	7.0
Wytheville, VA	36.945	–81.089	6.5	6.5	6.5
Knoxville, VA	35.961	–83.921	6.0	6.0	6.5
Cristiansburg, VA	37.142	–80.401	6.0	6.0	6.0
Dublin, VA	37.059	–80.621	6.0	6.0	6.0
Lynchburg, VA	37.416	–79.146	6.0	6.0	6.0
Richmond, VA	37.541	–77.484	6.0	6.0	6.0
Rocky Mount, VA	36.998	–79.887	5.5	5.5	6.0
Salem, VA	37.288	–80.081	6.0	6.0	6.0
Stuart, VA	36.640	–80.274	6.0	6.0	6.0
Tazewell, VA	37.130	–81.520	6.0	6.0	6.0
Durham, NC	35.993	–78.898	5.5	5.5	6.0
Lenoir, NC	35.902	–81.535	5.5	5.5	6.0
Oxford, NC	36.309	–78.591	5.5	5.5	6.0
Raleigh, NC	35.820	–78.650	6.0	6.0	6.0
Salisbury, NC	35.670	–80.480	6.0	6.0	6.0
Weldon, NC	36.427	–77.602	5.5	5.5	6.0
Winston, NC	36.100	–80.260	6.0	6.0	6.0
Burkeville, VA	37.117	–78.217	5.5	5.5	5.5
Petersburg, VA	37.188	–77.736	5.0	5.0	5.5
Ashland, VA	37.753	–77.484	5.0	5.0	5.0
Gate City, VA	36.633	–82.573	4.5	4.5	5.0
Crewe, VA	37.179	–78.128	5.0	5.0	5.0
Manassas, VA	38.750	–77.480	4.5	4.0	5.0
Hot Springs, VA	37.971	–79.893	4.0	4.0	5.0
Chase City, VA	36.790	–78.440	4.0	3.0	4.5
Suffolk, VA	36.728	–76.584	4.0	2.0	4.5
Washington DC	38.895	–77.037	4.5	2.0	4.5
Amherst, VA	37.573	–79.058	4.0	3.0	4.0
Bon Air, VA	37.520	–77.569	4.0	3.0	4.0
Staunton, VA	38.160	–79.080	4.0	3.0	4.0
Williamsburg, VA	37.290	–76.730	4.0	3.0	4.0
Wheeling, WV	40.069	–80.688	4.0	3.0	4.0
Fincastle, VA	37.270	–79.950	4.0	3.0	4.0
Luray, VA	38.665	–78.459	3.5	3.0	3.5
Spartanburg, SC	34.950	–81.930	3.5	3.0	3.5
Lester Manor, VA	37.583	–76.983	3.5	3.0	3.0
Atlanta, GA	33.760	–84.430	3.0	3.0	3.0
Louisville, KY	38.220	–85.750	3.0	3.0	3.0

*MMI = assignment of this study.
†MMI_d = DYFI-informed intensity values.
‡MMI_{HB} = assignment of Hopper and Bollinger (1971; half-unit indicates range of values given.
(Continued next page.)

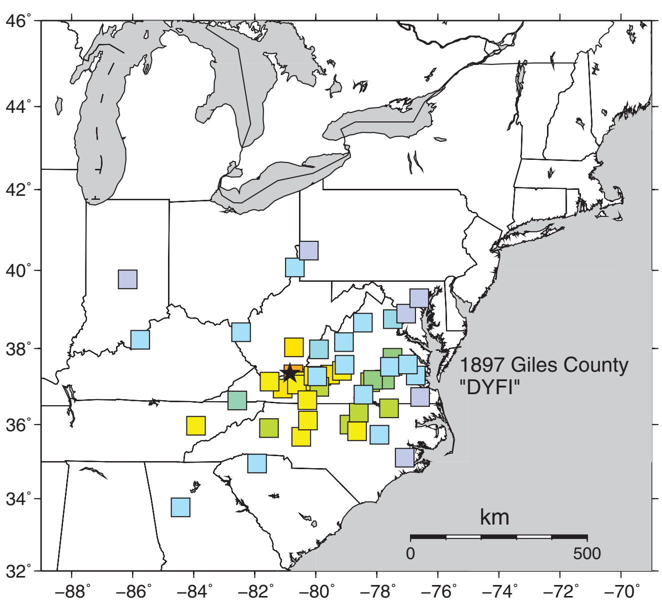
Location	Latitude (°)	Longitude (°)	MMI*	MMI_d†	MMI_{HB}‡
Wilson, NC	35.721	–77.915	3.0	3.0	3.0
Pittsburgh, PA	40.491	–80.233	3.0	2.0	3.0
Huntington, WV	38.419	–82.445	3.0	3.0	3.0
Indianapolis, IN	39.770	–86.160	2.0	2.0	2.5
Baltimore, MD	39.290	–76.610	2.0	2.0	2.5
New Bern, NC	35.110	–77.090	2.0	2.0	2.5

Table 1 (continued)
Locations and Intensities for the 1897 Giles County, Virginia, Earthquake

*MMI = assignment of this study.
†MMI_d = DYFI-informed intensity values.
‡MMI_{HB} = assignment of Hopper and Bollinger (1971; half-unit indicates range of values given.

In light of these discrepancies, I revisited the intensity assignments for the Giles County earthquake to make assignments that are not DYFI assignments *perse*, but rather are informed by the results of the DYFI algorithm. The results, shown in Table 1 and Figure 7, are different from those shown in Figure 1b, but not grossly so, and provide further support for the conclusion that the Giles County earthquake was significantly smaller than M 5.8. I do note that the intensity data set for the Giles County earthquake is relatively sparse: a full archival study would likely yield significantly more accounts given the location and date of the earthquake. Such a study would be timely in light of the new data from the 2011 Mineral earthquake.

To further compare the intensity distributions, one can estimate magnitude values using the method of Bakun and Wentworth (1997; hereinafter BW97) with the two CEUS attenuation relations presented by Bakun *et al.* (2003) and Bakun and Hopper (2004). Using the subjective intensity values for the Giles County earthquake and fixing the epicenter at the location of the most severe shaking (Pearisburg, Virginia), the preferred magnitude value is 5.3 using both attenuation relations. Using the DYFI-informed intensities, the preferred magnitude drops to 5.1 (5.13) using the model of Bakun *et al.* (2003) and to 5.2 (5.16) using the model of Bakun and Hopper (2004). The estimated magnitude values for the DYFI intensities for the Mineral earthquake are 5.4 using both attenuation relations. It is not clear why the BW97 approach underestimated the magnitude of the Mineral earthquake, but the consideration of the Giles County earthquake presented here suggests that DYFI-intensity values might be systematically lower than the subjectively determined intensities on which the intensity-attenuation relationships were based, in particular for the relatively large number of low-intensity values (II–IV).



▲ **Figure 7.** Intensity values for the 1897 Giles County earthquake recalculated based on results of the DYFI algorithm (see text). Color scale same as Figure 1.

To further explore this issue, one can consider the DYFI intensities for the 2008 Mt. Carmel, Illinois, earthquake, which also has a well-constrained M_w estimate (5.2; Herrmann *et al.*, 2008) and a spatially rich intensity distribution (over 42,000 responses from 3,575 locations). For this event the BW97 also underestimates the magnitude distribution, yielding estimates of 4.7–4.8 using the two attenuation relations. In recent years a number of studies have concluded that subjective-intensity assignments made in older studies are often inflated relative to assignments that follow currently accepted best practices (e.g., Ambraseys and Bilham, 2003; Hough and Page, 2011). The results presented here further suggest that DYFI intensities are systematically lower still than subjectively assigned intensities, even when the latter are assigned according to currently accepted practices.

Returning to the Giles County earthquake, I conclude the most direct comparison is between the DYFI-informed intensity values for this event with the DYFI intensities for the Mineral earthquake. Assuming that the biases associated with the use of DYFI intensities with the Bakun *et al.* (2003) and Bakun and Hopper (2004) attenuation relationship will be comparable for these two data sets, the BW97 method suggests that the former was approximately 0.3 units smaller than the latter, or M_w 5.5. This estimate remains relatively uncertain, of course, compared with instrumentally constrained magnitude values. The intensity distribution of a historical earthquake can be influenced, for example, by the rupture depth, which is unknown. However, the direct comparison of intensities suggests that the 1897 earthquake was smaller than the 2010 Mineral earthquake.

CONCLUSIONS

An extensive body of literature addresses the propagation and site effects that potentially control earthquake ground motions of engineering concern. Investigations generally rely on analysis of instrumentally recorded data and/or predictions of theoretical models. The strength of these approaches is their rigor; the common weakness is data limitation. The growing data set of well-constrained DYFI intensities offers the opportunity to undertake complimentary investigations, with different strengths and weaknesses. The most significant weaknesses are (1) imprecision of intensity values as an estimate of shaking amplitude, and (2) imprecision of the location of reported intensities. The analysis of Atkinson and Wald (2007) suggests that well-constrained intensity values do yield stable estimates of ground-motion parameters. The latter limitation, as discussed here, will limit the utility of DYFI data for detailed site-response investigations. The advantage of the approach, on the other hand, is the extraordinary spatial richness of the data, in particular for widely felt moderate and large earthquakes. The analysis presented in this report provides evidence for significantly anisotropic wave propagation due to the prevailing tectonic fabric of eastern North America. A similar conclusion was reached based on analysis of sparse instrumental data (e.g., Hough *et al.*, 1989) and is supported by theoretical modeling (e.g., Bostock and Kennett, 2007; Kennett, 1986.) The analysis further provides evidence that directivity significantly influences shaking intensities, even for an earthquake as small as M 5.8. Finally, site response is inferred to contribute a variability of roughly ± 1.5 intensity units.

Based on a first-principles consideration of intensity scales, Hough (2000) showed that each unit step of intensity increase corresponds, fairly robustly, to a factor of ~ 2 increase in PGA. (Intensity values must saturate at the highest shaking levels, but this issue is not relevant here). Thus, preferential propagation and directivity are inferred to influence PGA values on the order of $\pm 30\%$, while site response influences values by factors as high as 2–3. These results perhaps shed light on the high degree of aleatory uncertainty that have long plagued attempts to estimate or predict site response and its relationship to near-surface geology (e.g., Gibbs *et al.*, 1999; Tinsley *et al.*, 2004; Wald and Mori, 2000). A number of recent studies have explored the uncertainties associated with the use of proxy methods for site characterization (e.g., Allen and Wald, 2009; Wills *et al.*, 2000; Yong *et al.*, 2008). The results of this study further suggest that significant systematics are likely lurking within empirical estimates of site response. That is, in any earthquake, the combined effects of source directivity and azimuthally dependent propagation can influence shaking intensities as much as site response. With typically sparse instrumental data sets, these factors will be difficult to separate.

The growing collection of well-constrained DYFI-intensity values will also be invaluable for our understanding of key historical earthquakes, as illustrated here by the comparison between the 2011 Mineral earthquake and the 1897 Giles County, Virginia, earthquake. While there is a well-documented tendency for magnitudes of historical earthquakes to be overestimated (e.g., Ambraseys, 2004), this study points to a discrepancy between DYFI intensities and those determined subjectively using accepted current practices, in particular for low-intensity values (MMI II–IV). Thus, while analysis of DYFI-informed intensities might be a fruitful future approach, these values cannot be analyzed with methods and attenuation relations determined from traditionally determined intensity values. The results presented in this study suggest a potentially fruitful future direction will be to recalibrate the method of BW97 using well-constrained DYFI intensities for recent moderate events and to use these results to reexamine DYFI-informed intensities of key historical earthquakes. Following this general prescription, I estimate a preferred-magnitude of M_w 5.5 for the 1897 Giles County, Virginia, earthquake.

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REFERENCES

Allen, T. I., and D. J. Wald (2009). On the use of high-resolution topographic data as a proxy for seismic site conditions (V_{S30}), *Bull. Seismol. Soc. Am.* **99,** 935–943, doi 10.1785/0120080255.

Ambraseys, N. N. (2004). The state-of-the-art and practice of long-term seismicity, *Ann. Geophys.* **47,** 335–338.

Ambraseys, N., and R. Bilham (2003). Reevaluated intensities for the great Assam earthquake of 12 June 1897, Shillong, India, *Bull. Seismol. Soc. Am.* **93,** 655–673.

Atkinson, G. M. (2004). Empirical attenuation of ground motion spectral amplitudes in southeastern Canada and the northeastern United States, *Bull. Seismol. Soc. Am.* **94,** 1079–1095.

Atkinson, G. M., and D. J. Wald (2007). "Did You Feel It?" intensity data: A surprisingly good measure of earthquake ground motion, *Seismol. Res. Lett.* **78,** 362–368.

Bakun, W. H., and M. Hopper (2004). Magnitudes and locations of the 1811–1812 New Madrid, Missouri, and the 1886 Charleston, South Carolina, earthquakes, *Bull. Seismol. Soc. Am.* **94,** 64–75.

Bakun, W. H., and C. M. Wentworth (1997). Estimating earthquake location and magnitude from seismic intensity data, *Bull. Seismol. Soc. Am.* **87,** 1502–1521.

Bakun, W. H., A. C. Johnston, and M. G. Hopper (2003). Estimating locations and magnitudes of earthquakes in eastern North America from modified Mercalli intensities, *Bull. Seismol. Soc. Am.* **93,** 190–202.

Ben-Menahan, A. (1961). Radiation of seismic surface-waves from finite moving sources, *Bull. Seismol. Soc. Am.* **51,** 401–435.
Bollinger, G. A., and M. H. Hopper (1971). Virginia's two largest earthquakes—December 22, 1875 and May 31, 1897, *Bull. Seismol. Soc. Am.* **61,** 1033–1039.
Bollinger, G. A., and R. L. Wheeler (1983). The Giles County, Virginia, Seismic Zone, *Science* **219,** 1063–1065, doi 10/1126/science.219.4588.1063.
Boore, D. M. (1973). Effect of simple topography on seismic waves—implications for accelerations recorded at Pacoima Dam, San Fernando Valley, California, *Bull. Seismol. Soc. Am.* **63,** 1603–1609.
Bostock, M. G., and B. L. N. Kennett (2007). The effect of 3-D structure on *Lg* propagation patterns, *Geophys. J. Int.* **101,** 355–364, doi 10.1111/j.1365.246X.1990.tb06574.x.
Chapman, M. C., G. A. Bollinger, M. S. Sibol, and D. E. Stephenson (1990). The influence of the coastal plain sedimentary wedge on strong ground motions from the Charleston, South Carolina, earthquake, *Earthq. Spectra* **6,** 617–640, doi 10.1193/1.1585589.
Gibbs, J. F., J. C. Tinsley, D. M. Boore, and W. B. Joyner (1999). Seismic velocities and geological conditions at twelve sites subjected to strong ground motion in the 1994 Northridge, California, earthquake: A revision of *OFR 96-740*, *U.S. Geol. Surv. Open-File Rep., OF 99-446.*.
Hauksson, E., K. Felzer, D. Given, M. Giveon, S. E. Hough, K. Hutton, H. Kanamori, V. Sevilgen, A. Yong, and S. Wei (2008). Preliminary report on the 29 July 2008 Mw5.4 Chino Hills, eastern Los Angeles Basin, California, earthquake seqeunce, *Seismol. Res. Lett.* **79,** 855–866.
Herrmann, R. B., M. Withers, and H. Benz (2008). The April 18, 2008 Illinois earthquake: An ANSS monitoring success, *Seismol. Res. Lett.* **79,** 830–843, doi 10/1785/gssrl.79.6.830.
Hopper, M. G., and G. A. Bollinger (1971). *Earthquake History of Virginia*, 1774–1900, Dept. of Geol. Sciences, Virg. Polytechnic Inst and State Univ, Blacksburg, 87 pp.
Hough, S. E. (2000). On the scientific value of "unscientific" data, *Seismol. Res. Lett.* **71,** 483–485.
Hough, S. E., and M. Page (2011). Towards a consistent model for strain rate accrual and release for the New Madrid Seismic Zone, *J. Geophys. Res.* **116,** no. B03311, 17 pp., doi 10.1029/2010JB007783.
Hough, S. E., K. Jacob, and P. Friberg (1989). The 11/25/88, $M = 6$ Saguenay Earthquake near Chicoutimi, Quebec: Evidence for anisotropic wave propagation in northeastern North America, *Geophys. Res. Lett.* **16,** 645–648.
Johnston, A. C. (1996). Seismic moment assessment of earthquakes in stable continental regions I. Instrumental seismicity, *Geophys. J. Int.* **124,** 381–414.
Kennett, B. L. N. (1986). *Lg* waves and structural boundaries, *Bull. Seismol. Soc. Am.* **76,** 1133–1141.
Lee, S. J., Y. C. Chan, D. Komatitsch, B. S. Huang, and J. Tromp (2009). Effects of realistic surface topography on seismic ground motion in the Yangminshan region of Taiwan based on the spectral-element method and LiDAR DTM, *Bull. Seismol. Soc. Am.* **99,** 681–693.
Nuttli, O. W., G. A. Bollinger, and D. W. Griffiths (1979). On the relation between modified Mercalli intensity and body-wave magnitude, *Bull. Seismol. Soc. Am.* **69,** 893–909.
Sánchez-Sesma, F. J. (1985). Diffraction of elastic *SH* waves by wedges, *Bull. Seismol. Soc. Am.* **75,** 1435–1446.
Seekins, L. C., and J. Boatwright (2010). Rupture directivity of moderate earthquakes in northern California, *Bull. Seismol. Soc. Am.* **100,** 1107–1119, doi 10.1785/0120090161.
Tinsley, J., S. E. Hough, A. Yong, H. Kanamori, E. Yu, V. Appel, and C. Wills (2004). Geotechnical characterization of TriNet sites, *Seismol. Res. Lett.* **75,** 505–529.
USGS DYFI, http://earthquake.usgs.gov/earthquakes/dyfi/events/se/082311a/us/index.html (last accessed 9/2011).
VaTech, http://www.geol.vt.edu/outreach/vtso/Giles-Intensity.html (last accessed September 2011).
Wald, L. A., and J. Mori (2000). Evaluation of methods for estimating linear site-response amplifications in the Los Angeles region, *Bull. Seismol. Soc. Am.* **90,** S32–S42, doi 10.1785/0119970170.
Wald, D. J., V. Quitoriano, L. Dengler, and J. W. Dewey (1999). Utilization of the internet for rapid community intensity maps, *Seism. Res. Lett.* **70,** no. 6, 680–697.
Wessel, P., and W. H. F. Smith (1991). Free software helps map and display data, *Eos Trans. AGU* **72,** no. 441, 445.
Wills, C. J., M. Petersen, W. A. Bryant, M. Reichle, G. A. Saucedo, S. Tan, G. Taylor, and J. Treiman (2000). A site-conditions map for California based on geology and shear-wave velocity, *Bull. Seismol. Soc. Am.* **90,** S187–S208.
Yong, A., S. E. Hough, M. J. Abrams, H. M. Cox, and C. J. Wills (2008). Quantification of site characterization from remote-sensing data, *Bull. Seismol. Soc. Am.* **98,** 2679–2693.

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