Large 19th Century Earthquakes in Eastern/Central North America: A Comparative Analysis

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Foreword For the understanding of seismogenesis as well as seismic hazard assessment in the North American mid-continent, two historical events are of paramount importance: the 1811–1812 New Madrid, central U.S,. sequence and the 1886 Charleston, South Carolina earthquake. Published estimates of magnitudes of the four principal New Madrid earthquakes have ranged from M$\sim$7–8.75. In contrast, published estimates of the magnitude of the Charleston earthquake have almost all been within a range of Mw6.8–7.3. Upon cursory inspection, the macroseismic effects of the New Madrid mainshocks appear to be more severe at regional distances than those of the Charleston mainshock. I compare the intensity distributions more carefully, focusing on key indicators rather than the poorly constrained overall distribution of intensities. I conclude that the primary difference between the intensity distributions of the Charleston and New Madrid earthquakes is that the former has much better sampling, in particular of the low intensity field. These results suggest that the largest New Madrid mainshocks were not substantially larger than the Charleston earthquake.

1 Introduction

The earthquake sequence that struck the New Madrid region of the North American mid-continent in 1811–1812 had remarkably far-reaching effects. By some accounts the principal events in this sequence are among the largest—if not the largest—earthquakes to have ever occurred in a so-called Stable Continental Region (SCR, Johnston, 1996). Ground motions from the three principal events were felt by individuals as far away as Canada, New England, and at a number of locations along the Atlantic coast (Mitchill, 1815; Bradbury, 1819; Fuller, 1912; Nuttli, 1973; Penick, 1981; Street, 1984; Johnston, 1996). Contemporary accounts document three principal mainshocks: approximately 0215 local time (LT) on 16 December

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1811; around 0900 LT on 23 January, 1812, and approximately 0345 LT on 7 February 1812 (henceforth NM1, NM2, and NM3, respectively). All three events were felt throughout much of the central and eastern United States. Additionally, a large aftershock to NM1 (NM1-A) occurred near dawn on 16 December 1811. Available accounts also document substantial aftershock activity following all three mainshocks (Drake, 1815; McMurtrie, 1819; Fuller, 1912; Penick, 1981).

The Charleston earthquake of 1 September 1886—9:50 p.m. LT on 31 August 1886—was the primary event in a more conventional earthquake sequence: a single large mainshock preceded by a small number of foreshocks and followed by a conventional, if perhaps widespread, aftershock sequence (Dutton, 1889; Seeber and Armbruster, 1987).

Paleoseismic investigations suggest a repeat time of the order of 400–500 years for both the New Madrid sequence and the Charleston earthquake (Talwani and Schaeffer, 2001; Tuttle et al., 2002); they also suggest that the New Madrid seismic zone tends to produce prolonged sequences with multiple, distinct mainshocks, the magnitudes of which are comparable to those of the 1811–1812 events (e.g., Tuttle and Schweig, 1996; Tuttle et al., 2002). Thus, the magnitudes of these earthquakes are a critical issue for the quantification of regional hazard in central North America. A repeat of the 1811–1812 sequence would clearly have a tremendous impact. Because of low regional attenuation, the New Madrid seismic zone (NMSZ) contributes a nontrival component of seismic hazard in relatively distant, large midwestern U.S. cities such as St. Louis, Missouri (Frankel et al., 1996). The Charleston seismic zone also contributes significantly to regional as well as local hazard.

A second impetus to investigate the 1811–1812 sequence stems from its implications for general issues related to intraplate earthquake processes. The NMSZ is among the best-understood intraplate source zones in the world, largely because it has been so active throughout the historic and recent prehistoric past. This relative abundance of data affords the opportunity to explore critical unanswered scientific questions regarding large SCR earthquakes, most notably the questions of why such events occur in certain regions but (apparently) not in others, why and to what extent large earthquakes are clustered, and the nature (i.e., scaling) of large intraplate earthquakes.

In a sense, the importance of the New Madrid earthquakes—both scientifically and for hazard–correlates with their magnitudes, yet these values remain grossly uncertain. Considerable effort has been invested in gleaning quantitative information from the limited available data. Available data include (1) paleoliquefaction features preserved by the sediments within the Mississippi embayment (e.g., Tuttle and Schweig, 1996); (2) the present-day distribution of seismicity in the NMSZ, which is generally assumed to be a long-lived aftershock sequence that illuminates the principal fault zones (e.g., Gomberg, 1993; Johnston, 1996; Mueller et al., 2004); (3) first-hand reports ("felt reports") of the shaking and/or damage caused by the events over the central/eastern United States (e.g., Nuttli, 1973; Street, 1984).

While the size of both inferred mainshock ruptures and liquefaction features provides some constraint on magnitude, such estimates are invariably less well-constrained than those based on macroseismic effects. Determination of magnitudes for the 1811–1812 mainshocks thus hinges on the felt reports and their interpretation for modified Mercalli intensity (MMI) values. In a seminal investigation, Nuttli (1973) drew isoseismal contours based on his compilation and interpretation of approximately 40 archival accounts. He determined body-wave magnitude, m_b, values of 7.2, 7.1, and 7.4 for NM1, NM2, and NM3, respectively, based on a relationship between ground motion and intensities from smaller and more recent instrumentally recorded earthquakes in the central United States. With an exhaustive archival search, Street (1984) greatly expanded the number of reports (to approximately 100 for NM1) and assigned his own intensity values. Street (1982, 1984) used these new data and the same method used by Nuttli, (1973) to obtain m_b of 7.1 and 7.3 for NM2 and NM3 and 7.0 for the 0715 LT aftershock of December 16, 1811. Street (1982) determined these values by assuming the m_b value for NM1 determined by Nuttli (1973) and comparing the relative isoseismal areas of the other events.

Following the introduction of the moment-magnitude scale in 1979 (Hanks and Kanamori, 1979), attempts were made to convert earlier m_b values to moment-magnitude, M_w. It was at this time that the magnitude estimates grew to very large values, with estimates as high as 8.75 (Nuttli, 1979). Even as these estimates were made, it was recognized that they were based on extrapolations of data from smaller earthquakes and thus were highly uncertain. The lack of true calibration events from central/eastern North America led Johnston (1996) to undertake a comparison between intensity distributions and moment magnitudes M_w for large earthquakes in stable continental regions worldwide. He compared areas within isoseismals of discrete intensities with instrumentally measured moment magnitudes. On the basis of this calibration, he assigned Mw values of 8.1+−0.31, 7.8+/−0.33, and 8.0+/−0.33 for NM1, NM2, and NM3, respectively.

Hough et al. (2000) reinterpreted the accounts compiled by Nuttli (1973) and Street (1984), identifying a small number of outright transcription errors in the study of Nuttli (1973) and a larger number of inappropriately high intensity values that had apparently been assigned based on subjective perceptions of shaking. This study also addressed the bias due to early American settlement patterns, namely the fact that observers of the earthquakes were concentrated along major river valleys where substantial sediment-induced amplification is expected (e.g., Singh et al., 1988), and was in fact documented (e.g., Drake, 1815).

Hough et al. (2000) did not correct MMI values for site-response. Rather, the MMI values were assigned based on a careful consideration of the overall macroseismic effects as described by available archival accounts. In their interpretation, Hough et al. (2000) considered site response biases, in effect not allowing biased values to inappropriately control inferred isoseismal areas. Using the method of Johnston (1996), Hough et al. estimated Mw values of 7.2–7.3, 7.1, and 7.4–7.5 for NM1, NM2, and NM3, respectively.

The method of Johnston (1996) was developed using MMI values for a set of instrumentally recorded calibration earthquakes in so-called Stable Continental Regions (SCR) world-wide. If there are biases in the MMI values for the calibration earthquakes, or if other SCR regions are not perfect analogs for central/eastern North America, then the application of the Johnston (1996) method will introduce biases that are difficult to quantify. For this reason, the comparison between the New Madrid earthquake magnitudes and the magnitude of the Charleston earthquake is especially illuminating. That is, while analysis of the New Madrid intensity values alone might be fraught with uncertainty, a direct comparison with intensity values for the Charleston earthquake can help constrain the relative sizes of the events.

More recently, Bakun and Hopper (2004) estimated magnitude values for the New Madrid mainshocks using a new method, one in which intensity versus distance observations are used together with attenuation relationships developed from instrumentally recorded earthquakes in central/eastern North America (Bakun et al., 2003). His preferred estimates are 7.6, 7.5, and 7.8 for NM1, NM2, and NM3, respectively. These estimates are described as M-I, indicating that they are derived from intensity data. Because the attenuation relationships are derived using Mw values, it is generally assumed that M-I represents Mw. The approach of Bakun et al. (2003) does not require isoseismal contours and is thus less subjective than the method of Johnston (1996). However, Bakun's method reintroduces the problem that Johnston (1996) attempted to solve with his SCR compilation—namely, the lack of true calibration events for the largest historical earthquakes. This re-introduces the need for extrapolation, and its attendant uncertainties. For example, Bakun and Hopper (2004) consider two different extrapolation techniques, the one that leads to the preferred values and a second technique that yields values about 0.3 units smaller. A further potential difficulty is that Bakun et al. (2003) use the 1929 Mw7.3 Grand Banks earthquake (Bent, 1995) to develop their attenuation relationship—the only Mw>7 earthquake in their dataset. However, this event was located offshore from Newfoundland, Canada, arguably in a very different tectonic setting than the New Madrid events. Also, because the event occurred several hundred kilometers off-shore, its macroseismic effects are not well documented.

Investigation of the Charleston earthquake dates back to the immediate post-earthquake investigations led by Clarence Dutton, an Army officer detailed to the U.S. Geological Survey. This effort culminated in the publication of one of the earliest comprehensive, scientific reports of a large earthquake (Dutton, 1889). The so-called "Dutton Report" includes thorough and consistent compilations of near-field geological effects of the earthquake and far-field macroseismic effects. Whereas about 100 or fewer intensity values are available for each of the New Madrid mainshocks, the Dutton Report provides the basis for assignment of over 1000 intensity values. In a comprehensive interpretation of these accounts, Bollinger (1977) assigned almost 800 intensity values based on the 1337 intensity reports tallied by Dutton (1889). Bollinger (1977) estimated an m_b value of 6.8–7.1 using the same techniques that Nuttli (1973) used to estimate magnitudes for the New Madrid earthquakes.

Whereas an initial review of earlier intensity assignments for the New Madrid earthquakes suggested immediate biases (and a small handful of outright mistakes) that led to the reinterpretation by Hough et al. (2000), a initial review of the intensity assignments by Bollinger (1977) reveals the values to have been assigned carefully and in keeping with modern conventions. Each of the accounts were evaluated independently by three individuals and any discrepancies in assignment were evaluated and reconciled.

The intensity values determined by Bollinger (1977) have provided the basis for later investigation using increasingly modern methodology. Johnston (1996) estimated Mw 7.3+/−0.26 for the Charleston earthquake. Bakun and Hopper (2004) report a preferred Mw value of 6.9.

Published Mw values for the Charleston earthquake have thus been relatively consistent: the U.S. National Hazards Mapping project currently assumes a range between 6.8 and 7.5, with highest weight given to a value of 7.3 (Frankel et al., 2002). In contrast, Mw values for the New Madrid mainshocks have varied from 7.2 to 8.75, and the National Hazard Mapping project currently uses a range of 7.3–8.0, with highest weight assigned to 7.7 (Frankel et al., 2002). Considering the long-term strain-accrual rate, Newman et al. (1999) suggested an even lower value (Mw7.0), although Kenner and Segall (2000) showed that the long-term strain accrual rate provides at best only a weak constraint on the short-term rate of earthquakes generated by a local stress perturbation.

The enormous range of Mw values for the New Madrid events reflects the fundamental difficulty in interpreting sparse macroseismic effects for an earthquake for which no modern calibration event exists. With estimates varying by over a full magnitude unit, the task of reducing the uncertainties is clearly daunting. Yet it is important to not lose sight of the original, documented macroseismic effects of the earthquakes. In the following section I consider these in detail, including a comparison with the better documented effects of the Charleston earthquake. I focus on the first New Madrid mainshock, NM1, because the largest number of intensity values are available for this event.

2 Macroseismic Effects

Figures 1 and 2 present ShakeMap representations of intensities from the 16 December 1811 mainshock and the 1 September 1886 Charleston earthquake. These figures are generated using intensity data from Hough et al. (2000) and Hough (2004) and, for the Charleston earthquake, from Bollinger (1977). Obviously any comparison of intensities for the two events will depend critically on which intensity data one uses for NM1. Using the earlier intensity values of Nuttli (1973), Street (1982, 1984), or those in the official NOAA database (http://www.ngdc.noaa.gov/seg/hazard/int_srch.shtml), a comparison would look quite different. The arguments in favor of the reinterpreted values are discussed at length in Hough et al. (2000) and Hough (2004). Figures 1 and 2 are generated with identical color palettes and interpolation schemes; the greater sampling for the Charleston earthquake is manifest in both the number of sample points and the resolution of small-scale details in the intensity field.

Converting the intensity values to MMI (r) assuming epicentral locations of 35.8N and −90.1 W for NM1 and 32.4 N, −79.5 W for Charleston, one obtains

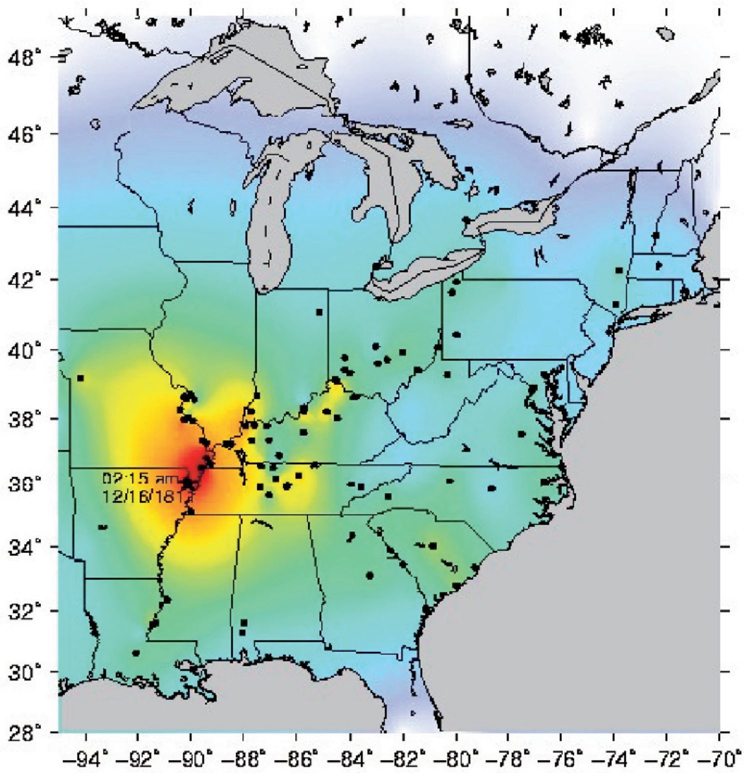

Fig. 1 Intensity map for the 16 December 1811 New Madrid mainshock constrained by intensity values determined by Hough et al. (2000) and Hough (2004). Intensity values are constrained for locations indicated by solid circles; between these locations intensity values are interpolated. The decay of the far-field intensity pattern is artificially imposed

the values shown in Fig. 3b. Figure 3b also shows the average distance at which each intensity level is observed for both earthquakes. Intensity values, I, for a given earthquake can typically be fit by

$$I = \mathrm{A} - \mathrm{B}(r) - \mathrm{C}\log(r) \tag{1}$$

Where A, B, and C are constants and r is epicentral distance. Fitting this equation to the intensity values for Charleston and NM1 yields the curves shown in Fig. 3a. These curves suggest that, on average, New Madrid intensity values are systematically about 1 unit larger than Charleston values any given distance. However, focusing on the average values within distance bins (black and gray stars), it is clear that average values of moderate intensities (V–VIII) are very similar. The overall amplitude of the NM1 curve is strongly controlled by, first, a couple of high intensity values at ∼100 km, and, second, high values at distances around 1000 km. The former are very poorly constrained given the paucity of well-built structures within

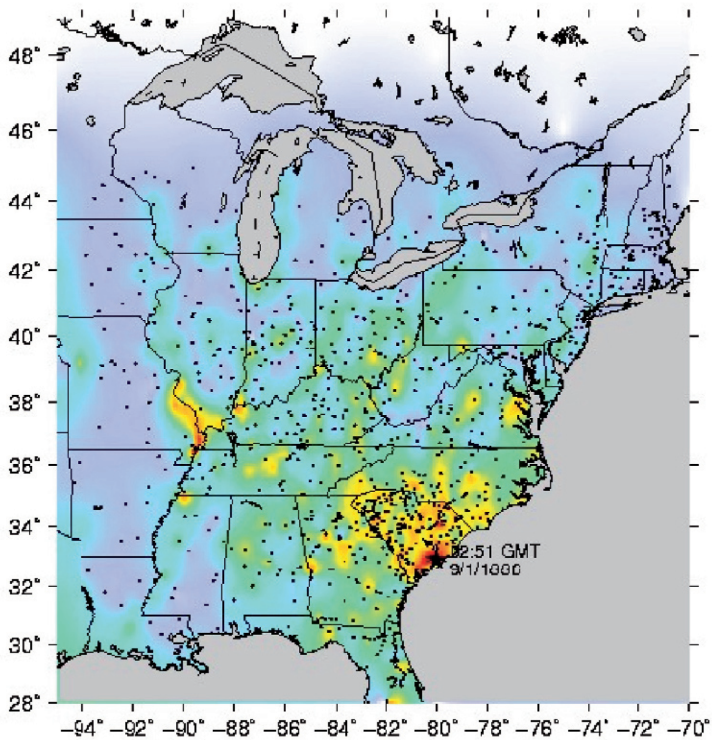

Fig. 2 Intensity map for the 1886 Charleston mainshock constrained by intensity values determined by Bollinger (1977) using accounts compiled by Dutton (1889). Intensity values are constrained for locations indicated by solid circles; between these locations intensity values are interpolated. The decay of the far-field intensity pattern is artificially imposed

100 km of NM1; the latter are also poorly constrained, as I discuss below. Focusing on average values for moderate MMI levels, Fig. 3b reveals that the two intensity distributions are quite similar for distances < 600 km, but that at greater distances higher intensities are suggested for the New Madrid event.

However, the low intensity values (II–IV) assigned for NM1 require careful consideration. A key distinction between the Charleston earthquake and NM1 is that the former struck at 9:50 p.m. LT whereas the latter occurred around 2:15 a.m. LT. Bollinger (1977) assigned values of II and III at locations where the shaking was described as felt by only those at rest and generally felt by those indoors, respectively. Assuming that NM1 was felt only by those who were awakened by the shaking, Hough et al. (2000) assigned MMI values of IV to accounts that described the shaking in any detail, making the conservative assumption that witnesses were asleep and were awakened by the shaking. Values of III were assigned to those locations where it was only noted that the shaking was felt. Values of II–IV are thus clearly difficult to distinguish for NM1.

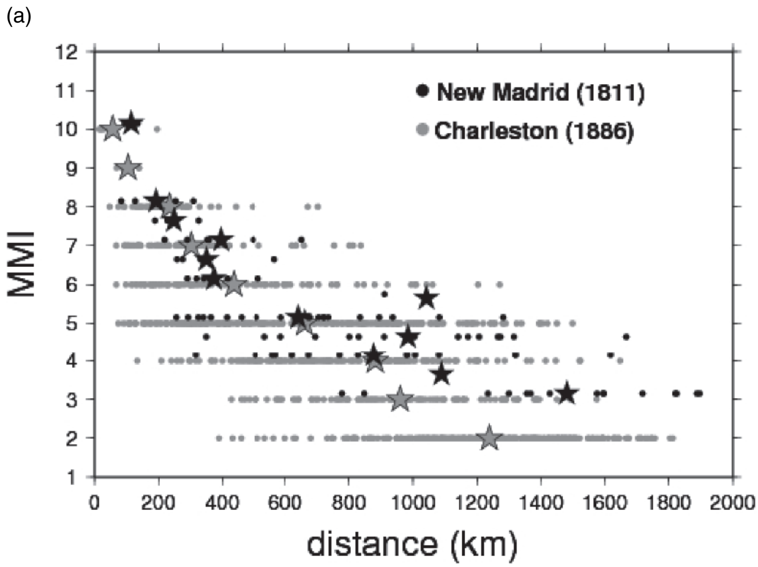

Fig. 3a Intensity versus distance for the 16 December 1811 New Madrid mainshock (*black circles*) and the 1 September 1886 Charleston mainshock (*gray circles.*) New Madrid values are shifted slightly up for clarity. Stars indicate average distance at which each intensity level is observed

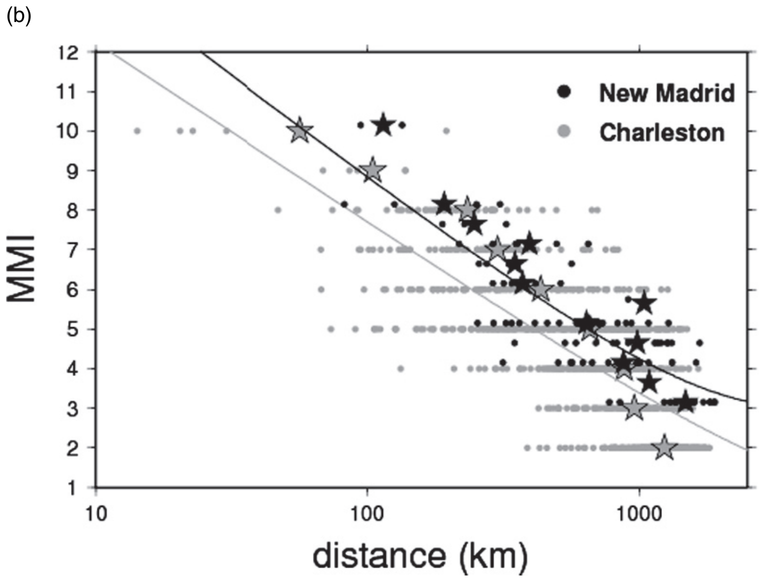

Fig. 3b Black and gray lines indicate regression curves fit to NM1 and Charleston intensity values, respectively. Intensity values for New Madrid (*black circles*) are shifted up by 0.1 units for clarity

To explore this issue further, one can hypothesize that the values of III and IV assigned by Hough (2000) in fact represent a mix of values between II and IV. One can then consider the distribution of distances at which this these values were observed for both of the earthquakes (Fig. 4). The distributions are similar. NM1 was weakly felt at *relatively* more locations at distances of 1800+ km. This may, however, reflect a relative concentration of population centers along the Altantic seaboard at the time of the New Madrid earthquakes. To test this one can consider the population distributions as revealed by the 1820 and 1880 US censuses. Assigning each state population to the average latitude and longitude for that state, one can examine state population as a function of distance from the Charleston and New Madrid earthquakes, respectively (Fig. 5a). The relative number of potential eyewitnesses to the New Madrid earthquake was indeed higher at distances greater than 1000 km than at smaller distances (Fig. 5b.) Thus the apparently large relative number of intensity values (i.e., felt reports) for NM1 at large distances may in fact be due to the low number of potential witnesses at closer distances.

It is clearly impossible to interpret the macroseismic effects of the New Madrid earthquakes without an appreciation for historical context, including settlement

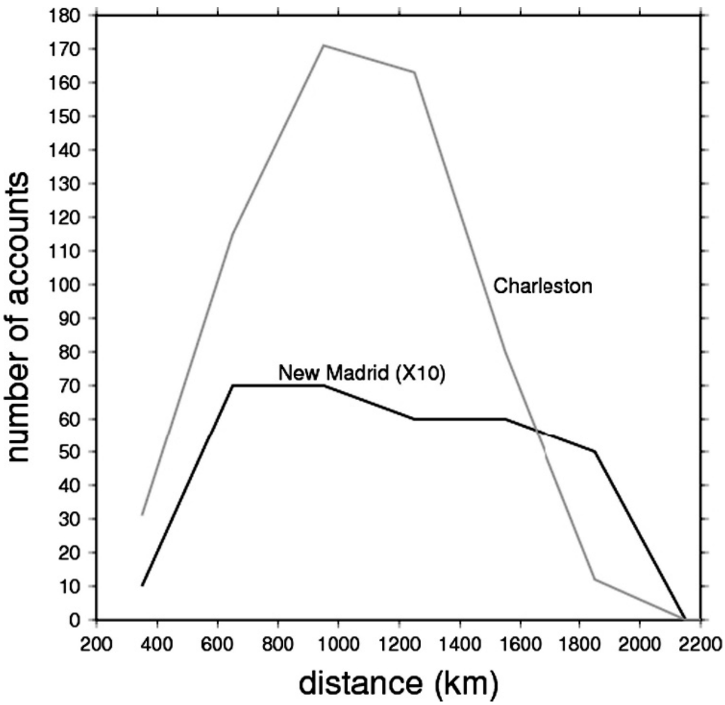

Fig. 4 Distance distribution showing the number of available accounts of weak shaking during the Charleston earthquake (*gray line*) and 16 December 1811 New Madrid mainshock (*black line*) as a function of epicentral distance. Values for the New Madrid event are amplified by a factor of 10; only about 100 total values are available for this event

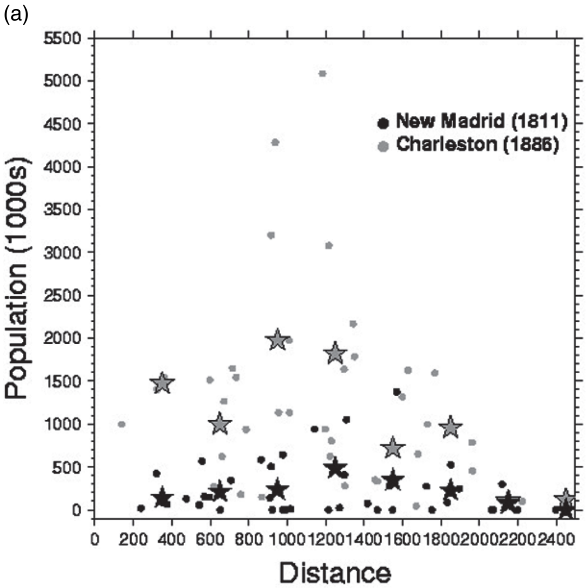

Fig. 5a Population (in 1000s) as a function of distance from the Charleston earthquake (*gray circles*) and NM1 (*black circles*), from the 1880 and 1820 census, respectively. Stars indicate average values within distance bins

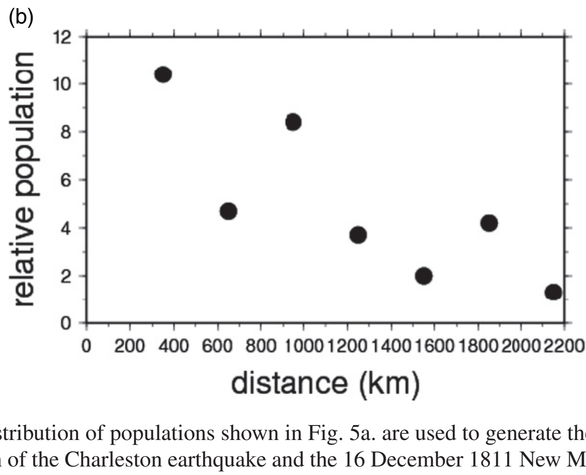

Fig. 5b The distribution of populations shown in Fig. 5a. are used to generate the relative population distribution of the Charleston earthquake and the 16 December 1811 New Madrid mainshock. Values shown here represent the ratio of the gray versus the black stars in Fig. 5a, and illustrate that, while roughly comparable populations experienced the 16 December 1811 New Madrid and Charleston mainshocks at distances of 2000 km, about a factor of 10 more people were living close to the latter event compared to those who were around to experience the former

patterns as well as overall population figures. The population of the United States was approximately 7,000,000 in 1811, with sizable numbers in the states of Tennessee, Kentucky, and the region including the present-day states of Missouri and Louisiana. The 1810 Census gives the population for several districts for which felt reports are considered, including the District of St. Louis (population 5667), Cincinnati (2540), New Orleans (24,552), Louisville (1357), and New Madrid (2103). By 1811 some towns had grown beyond simple frontier villages, with solidly constructed houses appearing by the turn of the century. The oldest brick building west of the Mississippi was built in the town of Ste. Genevieve in 1804; this town is along the Mississippi River valley north of New Madrid. This house and approximately 50 others that predate the New Madrid sequence, are still standing today.

Although the New Madrid earthquakes were likely felt by hundreds of thousands of people, spatial sampling of the intensity field was far from uniform. Especially throughout the mid-continent, early American settlements clustered in proximity to major river valleys. Significant amplification of shaking is expected at such sites, and was in fact explicitly documented by several eye-witnesses to the New Madrid earthquakes. As discussed at length by Hough et al. (2000), while every macroseismic data set will include some effects that reflect sediment-induced amplification effects, special care is necessary when interpreting 1811–1812 intensities because of the especially biased nature of the data set.

In contrast, by 1880 the population of the United States had grown to over 50 million, and settlement patterns had changed dramatically, largely due to the development of the U.S. railroad system. Railroad construction began in the U.S. in the late 1820s and the first commercial lines began in the early 1830s. In 1838 the railroads were designated as “post roads” by the U.S. Post Office; from this time onward the railroads were used to move U.S. mail. This provided further impetus for development of the rail system to the mid-continent and the West. As a consequence of these developments, as well as the growing overall population, settlement became more uniform throughout the former frontier regions. By the 1920s, early settlers had also begun to recognize the pitfalls associated with life on the immediate river banks, which included poor drainage, floods, and disease. The very earliest settlements of the late 1700s and very early 1800s often were on fluvial sites, immediately adjacent to rivers. New Madrid was built so close to the river bank that even before the earthquakes, parts of the town regularly gave way under the continued assault of river currents (Penick, 1981). By the time the Charleston earthquake occurred, settlements had migrated inland, away from waterways.

Given the disparate size and distribution of the populations in 1811–1812 versus 1886, it is appropriate to consider the intensity distributions in more detail. As a simple experiment, I winnow the 1886 intensity distribution down to only those locations that are within 10 km of a point location for which an intensity is available for NM1 (Figs. 6 and 7). Although one would ideally like use only the precise set of locations for which an intensity value is available for NM1, in fact the locations of early intensity observations are rarely precisely known. Using a buffer of 10 km, the list of winnowed values for Charleston is about the same size as the number available for NM1. In effect, this provides an indication of what the Charleston

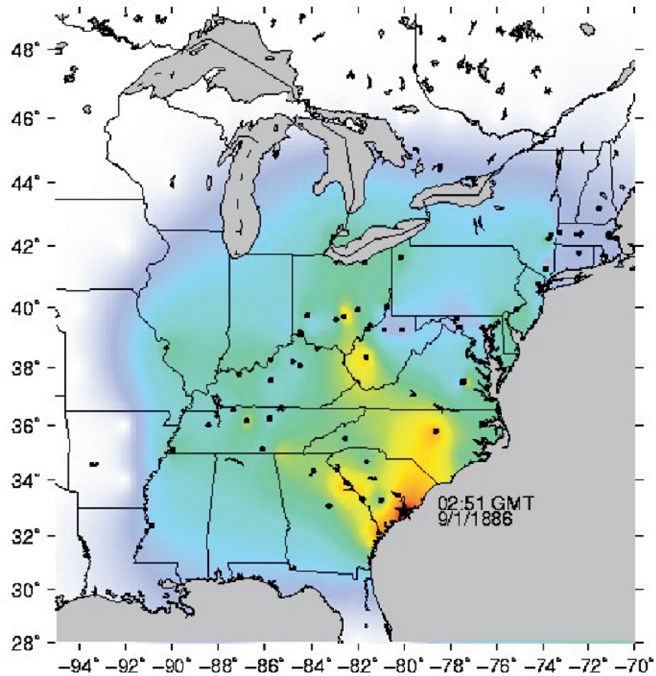

Fig. 6 The intensity distribution shown in Fig. 2, calculated using only those locations that are within 10 km of a location for which there is an account of NM1. This illustrates what the Charleston intensity distribution would have looked like, had the earthquake occurred in 1811 rather than 1886

intensity distribution might have looked like if the earthquake had occurred in 1811. One still finds higher intensity values for NM1 at distances greater than 800 km, but the winnowed values at closer distances are generally lower than the intensity values for NM1.

One can further consider key indicators of the intensity field for NM1 versus that for the Charleston earthquake: the maximum distance at which light damage occurred, and the nature of shaking at hard rocks sites. The latter comparison is difficult because so few observations are available from locations that are known to be hard-rock sites. However, a few reliable observations are available. In Cincinnati, Ohio, physician Daniel Drake described light damage in town along the river valley, but noted that on the elevated ridges away from the river, many families slept through the shock. (Drake went on to attribute this discrepancy to the fact that strata in the river valley were "loose" compared to the nearby limestone hills, one of the earliest observations of, and explanations for, site response (Drake, 1815)). This indicates a MMI no higher than IV for hard rock sites, as V is the level at which sleepers are generally awakened.

Another key hard-rock observation is available from Sainte Genevieve, Missouri, which had been moved to higher ground approximately a mile from the river after a

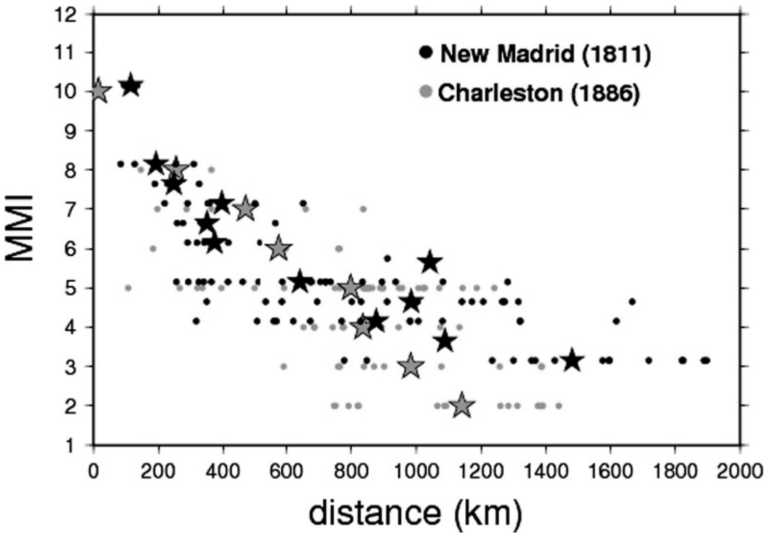

Fig. 7 Same as Fig. 3, but generated using only the winnowed Charleston data set shown in Fig. 6. Intensity values for New Madrid earthquake are shifted upwards by 0.1 units for clarity

flood in the late 1700s resulted in substantial erosion of the river bank upon which the town had originally been built (Brackenridge, 1817). According to a historian whose father lived in Ste. Genevieve at the time of the earthquakes, many shocks were felt in the town but they caused no damage (Rozier, 1890). This indicates MMI values no higher than V for any of the events. Ste. Genevieve provides a further illustration of the biases that can be associated with early archival records: the fact that dramatic effects are more likely to be documented than less dramatic effects. No account of the earthquakes from Ste. Genevieve is included in the compilation of Street (1984). The brief account in Rozier (1890) was discovered by the author following a focused archival search (Hough, 2004).

The accounts from Ste. Genevieve and Cincinnati thus suggest maximum credible MMI values of V and IV for hard-rock sites at distances of 160 and 560 km, respectively. Considering the distribution of MMI values estimated by Bollinger (1977), presumably at a given distance range, the low values provide an indication of intensities at hard-rock sites (Fig. 8). At a distance range of 100–199 km, MMI values range from a high of IX to a low of V, with just a single assignment of IV. At a distance range of 500–599 km, values range from VIII to II-III, with just 7 values of II-III versus 19 for MMI IV. I thus suggest that V represents an estimate of intensities at hard-rock sites at distances of 100–200 km, and an intensity of III–IV for distances of 500–600 km. In both cases the estimated hard-rock intensities are comparable to those available for the New Madrid events.

One can also consider the maximum distance at which NM1 and the Charleston earthquake caused damage. As discussed by Bollinger (1977), the Charleston earthquake caused plaster to fall from walls in Chicago, Illinois, and Valparaiso, Indiana,

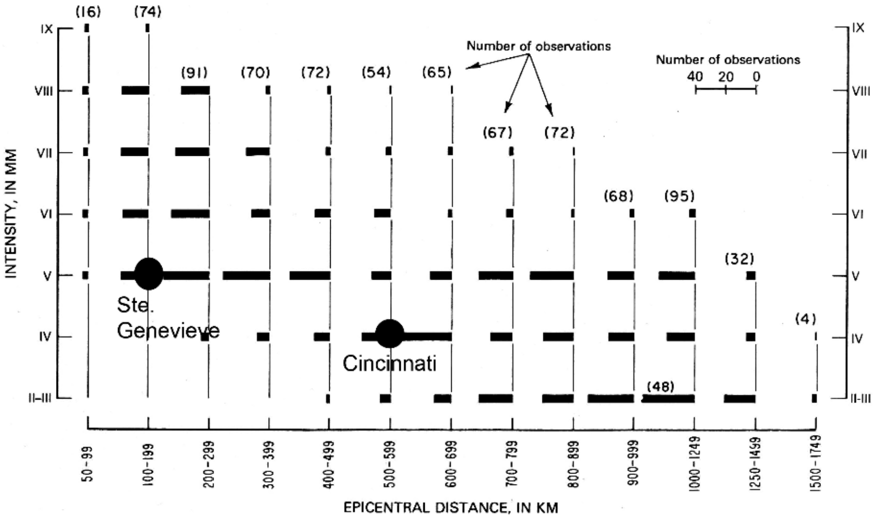

Fig. 8 Original figure from Bollinger (1977) shows the distribution of intensity values as a function of epicentral distance for the Charleston earthquake. Black circles indicate inferred hard-rock intensity values for NM1 based on especially reliable accounts from Ste. Genevieve, Missouri, and Cincinnati, Ohio

both at a distance of approximately 1200 km. In Terra Haute, Indiana, at a distance of ~1000 km, plaster fell from walls of the Opera House. Following NM1, plaster fell from walls in Columbia, South Carolina, and a church bell rang at Charleston, South Carolina, at distances of ~830 and ~960 km, respectively. The Charleston intensities might have been associated with swaying of tall buildings: plaster fell in one building in Chicago, Illinois, only above the fourth floor. However, the dramatic effects following NM1 occurred in a college dormitory building in Columbia, South Carolina, and in a church steeple in Charleston, South Carolina, also two especially tall and large structures. Again, the observations for NM1 appear to be comparable to those for the Charleston earthquake.

3 Conclusions

Although there has always been good agreement regarding the magnitude of the 1886 Charleston earthquake, there has been considerable disagreement about the magnitudes of the principle 1811–1812 New Madrid events; there has also been a prevailing conventional wisdom that the latter events were much larger than the former. Part of this impression might be rooted in the especially dramatic effects that the largest New Madrid events produced along riverbanks, in particular along the Mississippi River. The Charleston earthquake, in contrast, caused widespread liquefaction and ground failure, but did not have the same impact on a major river. Nonetheless, considering carefully the intensity distributions of NM1 versus the Charleston mainshock, one concludes that the former was not significantly larger than the latter. Available evidence in fact suggests the two to have been comparable in size—or at least to have produced comparable intensity fields.

I have focused on NM1 because more complete intensity data set is available for this event compared to the other large earthquakes in the sequence. The magnitude of NM2 is especially uncertain because some evidence suggests a location outside of the New Madrid Seismic Zone (Mueller et al., 2004; Hough et al., 2005). If the event did occur in the northern NMSZ, previous studies suggest a magnitude 0.1–0.2 units smaller than NM1. However, using the method of Bakun et al. (2003) and the intensities of Hough et al. (2000), one obtains a preferred value of 6.8 for the location suggested by Mueller et al. (2004).

In contrast, the location of NM3 is the best constrained of any of the New Madrid events: the earthquake created uplift across the Mississippi River and therefore can be associated with confidence with the Reelfoot thrust fault (e.g., Russ, 1982; Odum et al., 1998). Hough et al. (2000) and Bakun et al. (2003) both conclude that NM3 was approximately 0.2 magnitudes larger than NM1. Thus, while the magnitude of NM2 remains especially uncertain, one can estimate magnitudes of NM1 and NM3 relative to the magnitude of the Charleston earthquake. The latter value is, of course, itself uncertain; but magnitude estimates have been much more consistent for this event than for the principal New Madrid earthquakes. The most recent iteration of the U.S. National Hazard Mapping project assigned a maximum weight to a value of 7.3 for the Charleston earthquake; this implies values of 7.3 and 7.5 for NM1 and NM3, respectively. If the Charleston event was smaller than M7.3, the values for NM1 and NM3 would drop by a corresponding amount.

M_w values of 7.3 and 7.5 for NM1 and NM3 are consistent with the estimates obtained by Hough et al. (2000). These are in turn consistent with other lines of evidence, including rupture area as inferred from geomorphology, aftershock distribution, and stress-transfer theory, assuming standard scaling relations (Mueller et al., 2004). In particular, given the inferred dimensions of the rupture areas of NM1 and NM3, one need not postulate especially high stress drop values.

The M_w estimate of NM3 is slightly smaller than the instrumentally determined magnitude of the 2001 Bhuj, India earthquake: Mw7.6 (Antolik and Dreger, 2003). While this event is generally regarded as the best modern calibration event for the largest New Madrid mainshocks, Bakun and McGarr (2002) conclude that both intensity and weak motion data reveal lower attenuation in central/eastern North America than in other SCR regions around the world, including India. For this reason, while Hough et al. (2002) show that Bhuj mainshock and the largest New Madrid events produced comparable intensity distributions, they conclude that the magnitude of the Bhuj earthquake represents a credible upper bound for the largest New Madrid mainshocks. This conclusion is, again, consistent with the results of this study.

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