Evidence for Strain Accrual in the Eastern Tennessee Seismic Zone from Earthquake Statistics

by Will Levandowski and Christine A. Powell

ABSTRACT

The Eastern Tennessee Seismic Zone is roughly ten times more active than the average central and eastern United States, second only to the New Madrid Seismic Zone. Yet unlike New Madrid, no large earthquake is documented historically or paleoseismologically. Nonetheless, some of the ongoing seismicity could represent aftershocks of an unknown prehistoric event rather than as the result of focused long-term strain accrual. We compare modern seismicity rates with those expected after a large earthquake. We first adopt an extreme scenario in which the entire 250-km zone ruptured in an **M** 7.9 event on the eve of European settlement of the region 500 yrs ago, and we model its aftershocks with deterministic, probabilistic, and stochastic approaches. Each method shows that modern rates are significantly higher (to at least 97.3% confidence) than this upper bound on aftershock rates. Using a more reasonable set of assumptions, far less than 10% of modern seismicity could be ascribed to an ongoing aftershock sequence. We thus conclude that at least 90% of seismicity reflects time-independent rates most readily interpreted in terms of localized release of long-term strain in eastern Tennessee. The current distinction in estimated seismic hazard—based on historical and instrumental earthquake rates—between eastern Tennessee and its surroundings is warranted, but what localizes modern strain release in eastern Tennessee is still an open question.

INTRODUCTION

Intraplate seismicity clusters in space and time (e.g., Clark *et al.*, 2012), so modern loci of strain release may not highlight all areas of elevated future hazard. In particular, aftershock sequences do not reflect concentrated long-term strain accrual, and separating such sequences from background seismicity is a key challenge in understanding long-term hazard in the central and eastern United States (CEUS). This separation is difficult because geodetic strain rates are often insignificantly different from 0, and intraplate aftershock sequences can last for hundreds of years (Omori, 1894; Stein and Liu, 2009). In fact, some researchers have suggested that many of the earthquakes in areas such as the Eastern Tennessee Seismic Zone (ETSZ, Ebel, 2008) and New Madrid (Stein and Liu, 2009; Boyd *et al.*, 2015) may solely reflect ongoing aftershocks. If so, future ongoing seismic hazard in these zones may not be as high as currently estimated (e.g., Petersen *et al.*, 2014).

The New Madrid region hosts the highest rate of moment release in the CEUS and suffered four $M \gtrapprox 7$ earthquakes in 1811–1812. Paleoseismic records indicate similar earthquake sequences around 1450 C.E. and 900 C.E., as well as older sequences (Tuttle *et al.*, 2002, 2005). Although Page and Hough (2014) conclude that modern earthquake statistics are incompatible with the hypothesis that all New Madrid earthquakes are aftershocks, if the characteristic timescale for aftershock decay from $\mathbf{M} \sim 8$ events is as long as several hundred years (Stein and Liu, 2009), then at least some of the modern seismicity could indeed be related to postseismic process such as afterslip (e.g., Boyd *et al.*, 2015). (Numerous studies have focused on geodetic and myriad other evidence for and against ongoing strain accrual and long-term strain localization in the New Madrid Seismic Zone; we will not wade into that debate here but instead focus on earthquake statistics in the ETSZ.)

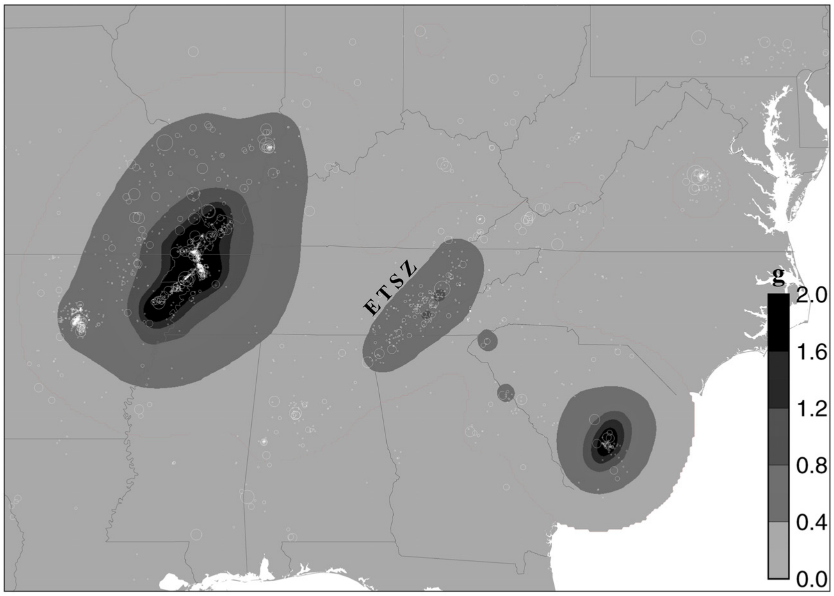
The ETSZ hosts the second highest rate of moment release in the CEUS. The present annual rate N of ETSZ earthquakes of magnitude $\geq M_{\mathrm{L}}$ approximately follows the Gutenberg–Richter recurrence relationship:

$$\log_{10}(N) = a + bM_{\mathrm{L}} = 3.198 - 0.937M_{\mathrm{L}}. \quad (1)$$

This relationship is derived by Bockholt *et al.* (2015) specifically for local magnitude, M_{L} in the ETSZ: for example, rates of $M_{\mathrm{L}} \geq 2$ and $M_{\mathrm{L}} \geq 3$ are 21 and 2.4 per year, respectively.

In contrast to New Madrid, however, no large historical shock in the ETSZ is known. The largest recorded are the 1973 m_{b} 4.7 Vonore, Tennessee, and 2003 M_{w} 4.6 Fort Payne, Alabama, earthquakes. Paleoseismic evidence is limited to one locality in the ETSZ (Warrell *et al.*, 2017), where deformed terraces suggest substantial earthquakes > 112, ~22, and ~15 ka. Maximum offset is ~1 m (Hatcher *et al.*, 2012; Warrell *et al.*, 2017), implying that one event could have been $\mathbf{M} \geq 6.5$ (using the maximum displacement-to-magnitude scaling of Wells and Coppersmith, 1994). Even generous estimates of the longevity of intraplate aftershock sequences (e.g., Stein and Liu, 2009) do not suggest that modern seismicity could be the result of one

 doi: 10.1785/0220180052



▲ **Figure 1.** Seismicity and seismic hazard in the southeastern United States. Shading is 2% probability of exceedance in 50 yrs for peak ground acceleration at 2 Hz (Petersen *et al.*, 2014), expressed as a fraction of gravitational acceleration. Seismicity is the declustered catalog of Petersen *et al.* (2014). ETSZ, Eastern Tennessee Seismic Zone. The color version of this figure is available only in the electronic edition.

of these events. Nevertheless, identifying paleoseismic targets in the ETSZ is challenging because of surface geology (e.g., few liquefaction-prone units), steep topography, and rapid geomorphic activity, but equation (1) implies a recurrence interval of 2352 yrs for $M_L \geq 7$.

We use earthquake statistics and epidemic-type aftershock sequence (ETAS) modeling to investigate the extent to which modern seismicity could possibly represent aftershocks of a large earthquake that occurred just before European and Cherokee settlement in the region (~500 yrs ago). First, deterministic equations for aftershock decay illustrate "best-guess" values for the maximum proportion of modern seismicity that could be aftershocks. Next, uncertainties in the relevant parameters are incorporated. Finally, we develop (stochastic) ETAS models to provide the most robust bounds on time-dependent earthquake rates.

GEOLOGIC SETTING

The ETSZ trends ~250 km northeast–southwest along the southern Appalachian region (Fig. 1), but earthquakes are not associated with Paleozoic structures related to the orogeny. Instead, colocation of ETSZ seismicity with potential field anomalies suggests association of the seismic zone with ancient basement structure; the most concentrated seismic activity is bounded on the northwest by the prominent 1600-km-long New York–Alabama (NYAL) magnetic lineament and associated Bouguer gravity lows. The NYAL is interpreted as a Proterozoic transform fault on the basis of crosscutting relationships with shallower, detached structures (King and Zietz, 1978), geochronological and geochemical data (Loewy *et al.*, 2003; Tohver *et al.*, 2004), and paleomagnetic polar-wander paths (D'Agrella-Filho *et al.*, 2008).

Even though the basement features associated with the seismicity extend ~1600 km, the ETSZ is the only location along the extensive NYAL magnetic lineament that is currently very seismogenic. Moreover, modern seismicity does not occur on the major northeast–southwest-trending structures but rather on *en echelon* north–south and east–west-striking faults (Chapman *et al.*, 1997). Given the modern approximately east-northeast–west-southwest horizontal maximum compression direction in the CEUS (e.g., Zoback, 1992; Levandowski *et al.*, 2018), these faults—which are dominantly near-vertical—are suitably oriented for strike slip, but the steep northeast–southwest basement structure is not.

The distinction between currently active and inherited structure bears on the maximum magnitude $M_{\max}$ that could be expected for a Holocene earthquake in the ETSZ. Following Wells and Coppersmith (1994) and assuming strike-slip motion (Chapman *et al.*, 1997), a fault of length L km could produce:

$$M_{\max} = 4.33 + 1.49 \log_{10}(L) \pm 0.24. \qquad (2)$$

bounding the possible rupture length with the entire 250-km northeast–southwest trend of concentrated epicenters, $M_{\max} = 7.90$.

Nevertheless, the steep basement fault that underlies the ETSZ—or any other structure parallel to it—is unlikely to slip in the modern stress field. Modern preferential fault orientations and those associated with most studied earthquakes are approximately east–west or north–south; the maximum extent across the ETSZ in these directions is ~125 km (Fig. 1). Following equation (1), therefore, a reasonable $M_{\max}$ is 7.45, which is similar to that currently assigned in the National Seismic Hazard Model (Petersen *et al.*, 2014).

DETERMINISTIC PALEOSEISMICITY

The paleoseismicity approach outlined by Ebel *et al.* (2000) allows deterministic estimates of the expected rates of aftershocks as a function of mainshock magnitude and time. To be conservative, we begin with $M_{\max} = 7.90$.

If aftershocks of the hypothetical $M_{\max}$ 7.9 account for all modern seismicity, then the time because this mainshock can be determined by combining Omori law describing the temporal decay of aftershocks of magnitude ≥ **M** with time and the Gutenberg–Richter relationship (Reasenberg and Jones, 1989):

$$\log_{10}(t + c) = (1/p)[a + b(M_{\max} - M) - \log_{10}(N)], \qquad (3)$$

in which t is the time in years since the mainshock; p is the temporal decay exponent in Omori law; a represents overall aftershock productivity; b is the b-value from the Gutenberg–Richter recurrence relation; and c is a small constant, 0.05, to avoid singularity at $t = 0$. Average values of a and p, -1.76 and 1.07, respectively, for California aftershock sequences (Reasenberg and Jones, 1989) are appropriate for (i.e., insignificantly different from) North American intraplate earthquake aftershock sequences (Ebel *et al.*, 2000; Ebel, 2009; Fereidoni and Atkinson, 2014). Using the values listed above gives $t = 196$ yrs. To account for all modern seismicity, an **M** 7.9 earthquake would have had to occur in the early 1800s C.E., which is not plausible because there is no historical record of such an event.

Nevertheless, if a large earthquake had occurred shortly before European settlement in the ETSZ 500 yrs ago, some of the ongoing activity could be aftershocks. Rearranging equation (3) to solve for the expected rate of earthquakes:

$$N = 10^{a+b(M_{\max}-M)}(t+c)^{-p} \tag{4}$$

and using $t = 500$ yrs gives 0.9 **M** ≥ 3 and 7.6 **M** ≥ 2 events per year compared with observed rates of 2.4 and 21. Even for this maximum magnitude of and minimum time since a large mainshock in the ETSZ, only 37% of modern seismicity is readily attributed (following this deterministic methodology) to aftershocks. A more reasonable $M_{\max}$ estimated from the extent of currently active faults (**M** 7.45) yields 0.3 **M** ≥ 3 and 2.9 **M** ≥ 2, or 14% of the modern rates. Finally, if one relaxes the assertion that a large-magnitude event occurred on the very eve of European (and possibly Cherokee) settlement ~500 yrs ago, rates decrease further. An **M** 7.45 670 yrs ago or greater would account for less than 10% of modern seismicity, meaning that 90% of modern seismicity may reflect ongoing strain accrual.

For most the remainder of our analysis, we focus discussion on the most conservative scenario of an **M** 7.9 500 yrs ago. As a rule of thumb, however, we suggest that a more reasonable scenario of an **M** 7.45 or less that occurred longer ago would produce at most approximately one-fourth as many modern aftershocks.

PROBABILISTIC PALEOSEISMICITY

We next propagate uncertainty in each of the pertinent parameters—a, b, p, and $M_{\max}$—through the calculations outlined earlier. We sample parameter values from the appropriate distributions using a Monte Carlo approach and solve for t; 100,000 such realizations develop probabilistic distributions of the time since the mainshock that could account for all of the seismicity in the ETSZ. Using values given by Reasenberg and Jones (1989), a and p are drawn from normal distributions with means of -1.76 and 1.07 (standard deviations, σ, 0.07 and 0.03). Varying the cutoff threshold and resampling the seismicity catalog yields an approximately normal distribution of b with $\sigma = 0.03$. Finally, rupture length is fixed at 250 km, but we note the ± 0.24 magnitude unit uncertainty in length-to-magnitude scaling given by Wells and Coppersmith (1994).

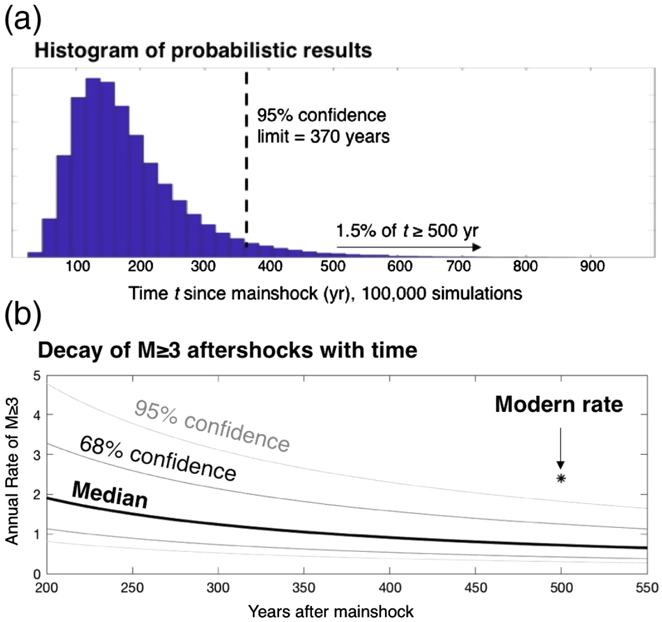
Selecting the four parameters in this way, each of 100,000 simulations solves equation (3) for the time t since the hypothetical mainshock if all modern earthquakes are aftershocks (Fig. 2a). Of these simulations, 95% require a major event in the past 370 yrs, and 98.5% require that the causative earthquake would have occurred in the past 500 yrs. We also solve equation (4) for the modern earthquake rates predicted by these simulations for a mainshock 500 yrs ago (Fig. 2b). The median rate for **M** 3 is 0.7 per year, 29% of the overall rate (13%–71%, 95% confidence interval; Fig. 2b). Again, a more reasonable estimate is roughly one-fourth of these rates, however.

STOCHASTIC (ETAS) MODELING

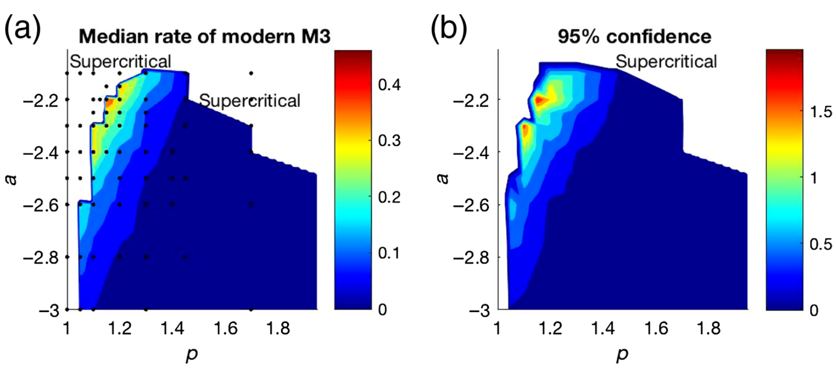
The most robust modeling of aftershocks accounts not only for the aftershocks of a mainshock but also for the aftershocks of those aftershocks and so on. The pertinent equations are similar in form to equation (4), but the parameters a and p are the direct—rather than indirect—Omori constants. For a given a, b, p, and $M_{\max}$, resulting rates during any time interval and in an aftershock–magnitude range cannot be computed deterministically. Instead, N in equation (4) is the expected value from a Poisson distribution, requiring the generation of stochastic catalogs.

Our approach is similar to that of Page and Hough (2014): For each a–p combination, we construct at least 100 synthetic catalogs and examine the distribution of modern rates of $M_L \geq 3$ ascribed to a 500-year-old mainshock and subsequent aftershock sequence. We exclude combinations of a, b, p, and $M_{\max}$ that create the nonphysical situation in which each earthquake triggers more than one aftershock on average, leading to an ever-increasing rate of seismicity (Sornette and Sornette, 1989; Page and Hough, 2014), a scenario referred to as supercritical. Unlike Page and Hough (2014), however, we additionally include uncertainty in mainshock magnitude and b-value using the distributions discussed earlier.

We generate 100 or more catalogs for each a–p combination as follows. An earthquake occurs at $t = 0$, with magnitude sampled from a normal distribution of 7.9 ± 0.24 and $b = 0.937 \pm 0.03$. For $t = 0, 1, \ldots, 499$ yrs, equation (4) calculates the expected rate of aftershocks, ranging from magnitude 1.5 to 8.5 (in 0.1 magnitude-unit increments) during each of the subsequent 500 yrs, and we sample the number of earthquakes for each year-magnitude combination from the attendant Poisson distributions. This catalog represents the direct aftershocks of the mainshock. This process is then repeated for the aftershocks that occur (i.e., Poisson-sample > 0) during year 0: Each of these earthquakes produces its own, 499-yr-long aftershock sequence. The number of these first-generation aftershocks for each year–magnitude combination is added to the number of direct aftershocks of the mainshock. Then we move to year 1 and model the summed first- and second-generation aftershocks therein and so on for the 500 yrs after the mainshock. In the end, each catalog contains an integer number of earthquakes each year at magnitudes between 1.5 and 8.5. The final step is to calculate the expected (noninteger) rate of events in the 501st year—representing the present—by loop-



▲ **Figure 2.** Results of probabilistic analysis. (a) Histogram of all values of t (see equations 3 and 4) for 100,000 Monte Carlo simulations of uncertainty in a, b, p, and M_{max}. With 95% confidence, a causative mainshock would be required in the past 370 yrs, which has not happened. Only 1.5% of the models feature a possible causative mainshock more than 500 yrs ago. (b) Probabilistic expectations of earthquake rates after an **M** ~ 7.9 mainshock. After 500 yrs, 0.7 **M** ≥ 3 events are expected annually (median value, shown as bold line), 29% of modern rates in the ETSZ. The color version of this figure is available only in the electronic edition.



▲ **Figure 3.** Results of epidemic-type aftershock sequence modeling. (a) Median annual rate of modern aftershocks $M_L \geq 3$. At least 100 catalogs were generated for each combination of a and p marked by a black dot. White region is supercritical. (b) 95% confidence limit of the annual rate of modern aftershocks $M_L \geq 3$. Even in this upper-bound scenario (**M** 7.9 500 yrs ago), no set of parameters reproduces current seismicity rates of 2.4 $M_L \geq 3$/year. The color version of this figure is available only in the electronic edition.

ing over time and magnitude, solving equation (4) for $M = 3$, and iteratively summing the expected rates.

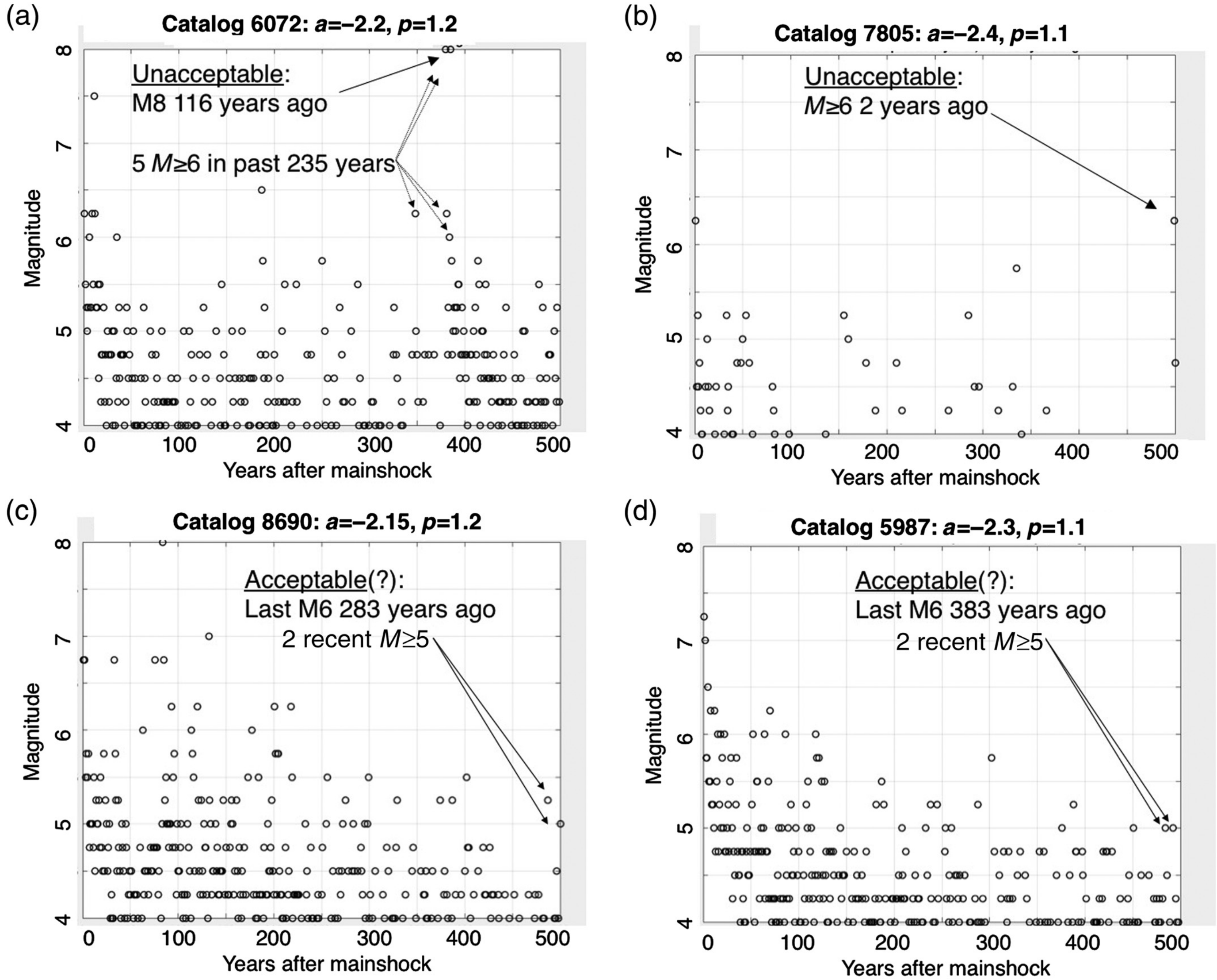
These catalogs place confidence limits on how much ETSZ seismicity could be an ongoing aftershock sequence. We do not impose any constraints other than that a, b, p, and M_{max} do not create a supercritical sequence. The most productive set of subcritical parameters (that with the highest median across all simulated catalogs) is $p = 1.15$, $a = -2.2$, so we focus discussion on these specific values, noting that other combinations produce fewer earthquakes (Fig. 3a). The median number of modern **M** ≥ 3 events across 150 catalogs with $p = 1.15$, $a = -2.2$ is 0.37/yr, with a 95% upper bound of 1.6/yr (i.e., 142 of 150 synthetic catalogs have lower associated rates; Fig. 3b). In fact, only 1 of the 150 catalogs would be expected to create more than 2.4 $M_L \geq 3$ events, giving a nominal probability that all modern ETSZ earthquakes are aftershocks of 0.67% even for this productive a–p combination. Other combinations are slightly more successful at explaining modern rates, but none has a nominal probability greater than 2.7%. Thus, to 97.3% confidence, no combination matches modern rates.

Of more than 11,000 synthetic catalogs for various a–p, only 22 produce modern seismicity rates. A subset of these catalogs is presented in Figure 4. Nearly all of the successful catalogs feature large-magnitude (≥ **M** 7) earthquakes in the past ~100 yrs (e.g., Fig. 4a) and/or **M** ~ 6 events in the past decades (e.g., Fig. 4b), each of which is inconsistent with the historical record. Disregarding such catalogs is important because most of the expected modern $M_L \geq 3$ events are simply direct or second-generation aftershocks of such implausible earthquakes. Only 2 catalogs of the 22 do not contain such events (Fig. 4c,d), but each has higher rates of **M** 5 events in the past decades than observed instrumentally (because none has been recorded). Not a single catalog of more than 11,000 can account for modern seismicity rates without invoking larger earthquakes than have been documented historically or instrumentally.

DISCUSSION AND CONCLUSION

Aftershocks from known or prehistoric earthquakes in the CEUS may complicate understanding of long-term strain accrual and therefore long-term seismic hazard. Separating such aftershocks could allow for both time-dependent and time-independent components of seismicity rate models (i.e., Gutenberg–Richter a-values) used for hazard assessment (e.g., Petersen *et al.*, 2014).

Currently, smoothed-gridded historical seismicity rates in the ETSZ are more than 10 times the average rate elsewhere in the CEUS (Petersen *et al.*, 2014). There are two simple explanations for this anomalous rate. A prehistoric mainshock could have created a long-lived aftershock sequence. If removing possible aftershocks still leaves anomalous rates, however, the ETSZ is more plausibly an area of localized, elevated release of long-term strain. We have shown that the probability that all ETSZ earthquakes—2.4 $M_L \geq 3$/yr—are merely aftershocks is, at most, 2.7%. Median rates from three different approaches are < 10%, 29%, and 37%, implying that the rate



▲ **Figure 4.** Example synthetic earthquake catalogs. (a,b) Examples of the 20 of 22 catalogs that account for the modern rate of 2.4 $M_L \geq 3$ per year but feature large earthquakes that are not documented historically. Two M_L 8 events did not occur in the late 1800s (a) in the ETSZ, nor did an M_L 6 event occur in the past few years (b). Many of the simulated modern $M_L \geq 3$ events are direct aftershocks of such events. (c,d) The two catalogs that reproduce modern earthquake rates without an **M** ≥ 6 event in the past 235 yrs. Each has higher rates of **M** 5 events in the past decades than the 0 recorded, however.

of release of long-term strain is at least six times as rapid in the ETSZ as the CEUS average.

These estimates are conservative for two main reasons. We allow that the causative earthquake ruptured the entire 250-km length of the ETSZ, comparable to the largest intraplate earthquake ever documented, even though no paleoseismic evidence and no topographic scarp are available to support this assertion and the fault that we have invoked is poorly oriented to slip in the modern stress field. A more logical bound on the mainshock magnitude is 7.45. Second, early—initially Spanish—Europeans were not met with accounts of a recent massive earthquake by early Cherokee inhabitants (Hancock, 2013), which makes its occurrence dubious or at least requires an older mainshock. Taking these two points into account, we have repeated the ETAS models for the average Omori parameters $a = -2.1, p = 1.35$ determined by Felzer *et al.* (2003) for California sequences, $M_{max} = 7.45$, and $t = 750$ yrs. The median rate of modern $M_L \geq 3$ events is 0.04/yr, with a 95% confidence of 0.21/yr, less than 10% of the modern seismicity rate. We therefore suggest that the rate of independent earthquakes in the ETSZ is more than nine times as great as the average in the CEUS.

This conclusion implies that the elevated seismic hazard modeled from *a*-values derived from modern earthquake rates in the ETSZ (Petersen *et al.*, 2014) represents time-independent hazard driven by long-term strain accrual rather than time-dependent aftershock moment release. Similar arguments have been presented for New Madrid (Page and Hough, 2014) and Charlevoix (Fereidoni and Atkinson, 2014), suggesting that the three most active seismic zones in central and eastern North America are not simply the ghosts of earlier earthquakes.

DATA AND RESOURCES

All data used in this paper came from published sources listed in the references.

ACKNOWLEDGMENTS

The authors thank two anonymous reviewers and Eastern Section Editor Martin Chapman for help in clarifying and strengthening this manuscript. The Colorado College Department of Geology funded WL through a visiting professorship.

REFERENCES

Bockholt, B. M., C. A. Langston, and M. Withers (2015). Local magnitude and anomalous amplitude distance decay in the eastern Tennessee seismic zone, *Seismol. Res. Lett.* **86,** no. 3, 1040–1050.

Boyd, O. S., R. Smalley Jr., and Y. Zeng (2015). Crustal deformation in the New Madrid seismic zone and the role of postseismic processes, *J. Geophys. Res.* **120,** no. 8, 5782–5803.

Chapman, M. C., C. A. Powell, G. Vlahovic, and M. S. Sibol (1997). The nature of faulting in eastern Tennessee inferred from a statistical analysis of focal mechanisms and epicenter locations, *Bull. Seismol. Soc. Am.* **87,** 1522–1536.

Clark, D., A. McPherson, and R. Van Dissen (2012). Long-term behaviour of Australian stable continental region (SCR) faults, *Tectonophysics* **566,** 1–30.

D'Agrella-Filho, M. S., E. Tohver, J. O. S. Santos, S.-A. Elming, R. I. F. Trindade, I. I. G. Pacca, and M. C. Geraldes (2008). Direct dating of paleomagnetic results from Precambrian sediments in the Amazon craton: Evidence for Grenvillian emplacement of exotic crust in SE Appalachians of North America, *Earth Planet. Sci. Lett.* **267,** nos. 1/2, 188–199.

Ebel, J. E. (2008). The importance of small earthquakes, *Seismol. Res. Lett.* **79,** no. 4, 491–493.

Ebel, J. E. (2009). Analysis of aftershock and foreshock activity in stable continental regions: Implications for aftershock forecasting and the hazard of strong earthquakes, *Seismol. Res. Lett.* **80,** no. 6, 1062–1068.

Ebel, J. E., K. P. Bonjer, and M. C. Oncescu (2000). Paleoseismicity: Seismicity evidence for past large earthquakes, *Seismol. Res. Lett.* **71,** no. 2, 283–294.

Felzer, K. R., R. E. Abercrombie, and G. Ekstrøm (2003). Secondary aftershocks and their importance for aftershock forecasting, *Bull. Seismol. Soc. Am.* **93,** no. 4, 1433–1448, doi: 10.1785/0120020229.

Fereidoni, A., and G. M. Atkinson (2014). Aftershock statistics for earthquakes in the St. Lawrence Valley, *Seismol. Res. Lett.* **85,** no. 5, 1125–1136.

Hancock, J. T. (2013). A world convulsed: Earthquakes, authority, and the making of nations in the War of 1812 era, *Doctoral Thesis*, University of North Carolina, Chapel Hill.

Hatcher, R. D., Jr., J. D. Vaughn, and S. F. Obermeier (2012). Large earthquake paleoseismology in the East Tennessee seismic zone: Results of an 18-month pilot study, in *Recent Advances in North American Paleoseismology and Neotectonics East of the Rockies*, R. T. Cox, O. L. Boyd, and J. Locat (Editors), Geological Society of America Special Papers, Vol. 493, 111–142.

King, E. R., and I. Zietz (1978). The New York–Alabama lineament: Geophysical evidence for a major crustal break in the basement beneath the Appalachian basin, *Geology* **6,** 312–318.

Levandowski, W., R. B. Herrmann, R. Briggs, O. S. Boyd, and R. Gold (2018). An revised stress map of the continental United States reveals evidence for heterogeneous intraplate stress, *Nature Geosci.* **11,** doi: 10.1038/s41561-018-0120-x.

Loewy, S. L., J. N. Connelly, I. W. D. Dalziel, and C. F. Gower (2003). Eastern Laurentia in Rodinia: Constraints from whole-rock Pb and U/Pb geochronology, *Tectonophysics* **375,** 169–197.

Omori, F. (1894). On the aftershocks of earthquakes, *J. Coll. Sci. Imp. Univ. Tokyo* **7,** 111–200.

Page, M. T., and S. E. Hough (2014). The New Madrid Seismic Zone: Not dead yet, *Science* **343,** 762–764.

Petersen, M. D., M. P. Moschetti, P. M. Powers, C. S. Mueller, K. M. Haller, A. D. Frankel, Y. Zeng, S. Rezaeian, S. C. Harmsen, O. S. Boyd, *et al.* (2014). Documentation for the 2014 update of the United States national seismic hazard maps, *U.S. Geol. Surv. Open-File Rept. 2014-1091*, doi: 10.3133/ofr20141091.

Reasenberg, P. A., and L. M. Jones (1989). Earthquake hazard after a mainshock in California, *Science* **243,** 1,173–1,176.

Sornette, A., and D. Sornette (1989). Renormalization of earthquake aftershocks, *Geophys. Res. Lett.* **26,** no. 13, 1981–1994.

Stein, S., and M. Liu (2009). Long aftershock sequences within continents and implications for earthquake hazard assessment, *Nature* **462,** no. 7269, 87.

Tohver, E., J. S. Bettencourt, R. Tosdal, K. Mezger, W. B. Leite, and B. L. Payolla (2004). Terrane transfer during the Grenville orogeny: Tracing the Amazonian ancestry of southern Appalachian basement through Pb and Nd isotopes, *Earth Planet. Sci. Lett.* **228,** nos. 1/2, 161–176.

Tuttle, M. P., E. S. Schweig III, J. Campbell, P. M. Thomas, J. D. Sims, and R. H. Lafferty III (2005). Evidence for New Madrid earthquakes in A.D. 300 and 2350 B.C, *Seismol. Res. Lett.* **76,** no. 4, 489–501, doi: 10.1785/gssrl.76.4.489.

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, *Bull. Seismol. Soc. Am.* **92,** no. 6, 2080–2089, doi: 10.1785/0120010227.

Warrell, K. F., R. T. Cox, R. D. Hatcher, J. D. Vaughn, and R. Counts (2017). Paleoseismic evidence for multiple $M_w \geq 6$ earthquakes in the Eastern Tennessee Seismic Zone during the Late Quaternary, *Bull. Seismol. Soc. Am.* **107,** no. 4, 1610–1624, doi: 10.1785/0120160161.

Wells, D. L., and K. J. Coppersmith (1994). New empirical relationships among magnitude, rupture length, rupture width, rupture area, and surface area, *Bull. Seismol. Soc. Am.* **84,** no. 4, 974–1002.

Zoback, M. L. (1992). Stress field constraints on intraplate seismicity in eastern North America, *J. Geophys. Res.* **97,** no. B8, 11,761–11,782.

Will Levandowski[1]
Department of Geology
Colorado College
14 East Cache La Poudre Street
Colorado Springs, Colorado 80903 U.S.A.
will.levandowski@tetratech.com

Christine A. Powell
Center for Earthquake Research and Information
The University of Memphis
Memphis, Tennessee 38152 U.S.A.

Published Online 7 November 2018

[1] Now at TetraTech, Inc., 1100 South McCaslin Boulevard, Suite #150, Superior, Colorado, U.S.A.