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Intraplate Triggered Earthquakes: Observations and Interpretation

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Abstract We present evidence that at least two of the three 1811–1812 New Madrid, central United States, mainshocks and the 1886 Charleston, South Carolina, earthquake triggered earthquakes at regional distances. In addition to previously published evidence for triggered earthquakes in the northern Kentucky/southern Ohio region in 1812, we present evidence suggesting that triggered events might have occurred in the Wabash Valley, to the south of the New Madrid Seismic Zone, and near Charleston, South Carolina. We also discuss evidence that earthquakes might have been triggered in northern Kentucky within seconds of the passage of surface waves from the 23 January 1812 New Madrid mainshock. After the 1886 Charleston earthquake, accounts suggest that triggered events occurred near Moodus, Connecticut, and in southern Indiana. Notwithstanding the uncertainty associated with analysis of historical accounts, there is evidence that at least three out of the four known M_W 7 earthquakes in the central and eastern United States seem to have triggered earthquakes at distances beyond the typically assumed aftershock zone of 1–2 mainshock fault lengths. We explore the possibility that remotely triggered earthquakes might be common in low-strain-rate regions. We suggest that in a low-strain-rate environment, permanent, nonelastic deformation might play a more important role in stress accumulation than it does in interplate crust. Using a simple model incorporating elastic and anelastic strain release, we show that, for realistic parameter values, faults in intraplate crust remain close to their failure stress for a longer part of the earthquake cycle than do faults in high-strain-rate regions. Our results further suggest that remotely triggered earthquakes occur preferentially in regions of recent and/or future seismic activity, which suggests that faults are at a critical stress state in only some areas. Remotely triggered earthquakes may thus serve as beacons that identify regions of long-lived stress concentration.

Introduction

According to modern analyses, central and eastern North America has experienced five earthquakes with $M_W \geq 7$ over the historic record: the three principal 1811–1812 New Madrid mainshocks (Nuttli, 1973; Street, 1982, 1984; Johnston, 1996a,b; Johnston and Schweig, 1996; Hough *et al.*, 2000); the 1886 Charleston, South Carolina, earthquake (Dutton, 1889; Bollinger, 1977; Johnston, 1996b); and the 1929 Grand Banks event (Bent, 1995). An additional large earthquake, thought to have been upward of M_W 7, occurred near Charlevoix, Canada, in 1663 (Smith, 1962), but accounts of this event are especially sparse. Of these, instrumental records are available only for the Grand Banks event. The magnitudes for the other earthquakes are estimated principally from the shaking effects as documented by contemporary accounts and are not precisely constrained. There is, however, little doubt that they were large ($M_W > 7$) earthquakes. The largest New Madrid event produced surface rupture or flexure that disrupted the Mississippi River in several places (e.g., Odum *et al.*, 1998; Mueller *et al.*, 1999). Substantial surface deformation, possibly reflecting primary surface rupture, has also been inferred for the 1886 Charleston earthquake based on systematic shortening of railroad tracks (Seeber and Armbruster, 1987; A. Johnston, personal comm., 2000).

Several studies have shown that recent large ($M_W \geq 7$) earthquakes in seismically active regions were associated with remotely triggered events at regional distances (Hill *et al.*, 1993; Bodin and Gomberg, 1994). Although there can be ambiguity between remotely triggered earthquakes and aftershocks, the former are generally considered to be events that are more than 1–2 fault lengths away from the mainshock. The documented triggered earthquakes have occurred preferentially, although not exclusively, in active volcanic and/or geothermal regions. In the limited time since triggered earthquakes were first broadly recognized by the seismological community in 1992, no intraplate events have

been observed to generate triggered earthquakes. However, this absence of observations could merely reflect the slow rate at which substantial intraplate events occur. In central and eastern North America, only a handful of events with magnitudes close to (or larger than) M_W 6 have occurred since 1886, including earthquakes in Charleston, Missouri (1895) (Johnston, 1996b); Charlevoix, Quebec (1925) (e.g., Bent, 1992); Timiskaming, Quebec (1935) (Hodgson, 1936); Miramichi, New Brunswick (1982) (e.g., Basham and Kind, 1986); Saguenay, Quebec (1988) (Duberger *et al.*, 1991); and Ungava (1989) (e.g., Bent, 1994). Results from interplate regions, however, suggest that remote triggering is associated with earthquakes of M_W 7 and larger (Gomberg and Davis, 1996). In light of data limitations that will likely continue for many years to come, the historical accounts of preinstrumental earthquakes are a valuable resource.

An earlier study re-examined historic accounts from the New Madrid sequence and concluded that two of the three mainshocks triggered at least three substantial earthquakes well outside the New Madrid Seismic Zone, most likely in northern Kentucky/southern Ohio (Hough, 2001). The first occurred approximately 4 days after the second New Madrid mainshock on 23 January 1812. The second and third triggered events occurred approximately 20 and 22 hr after the 7 February 1812 mainshock. The largest of these triggered events is estimated to have had a magnitude in the low to mid M 5 range. Additionally, Hough and Martin (2002) presented evidence that a large (M 6) aftershock on 17 December 1811 occurred well south of the New Madrid Seismic Zone. (Throughout this article we use M as a generic term for magnitude, unless values are given as moment magnitudes, M_W. In eastern North America reported local magnitude is generally $m_b(Lg)$ [Kim, 1998]).

In this article we explore further interpretation of the results presented by Hough et al. (2000) and Hough and Martin (2002) as well as additional observations from the New Madrid sequence and from the 1886 Charleston, South Carolina, sequence. We discuss these results in light of recent suggestions that the crust in intraplate regions is characterized by a critical stress state (e.g., Townend and Zoback, 2000).

The observations discussed in this article are gleaned from two sources: anecdotal accounts of the New Madrid sequence and the National Center for Earthquake Engineering Research (NCEER) catalog (Armbruster and Seeber, 1992). Accounts from the New Madrid sequence have been published previously by Street (1982) and are being made available online as part of the New Madrid compendium (www.ceri.memphis.edu/compendium). Although the analysis of historic accounts raises issues regarding reliability and precision, the value of such work has been demonstrated by innumerable careful investigations of important historic earthquakes. A detailed discussion of the issues associated with such investigations was presented by Hough (2000).

Observations: The 1811–1812 New Madrid Sequence

The New Madrid sequence was centered in the so-called boot-heel region of Missouri, in the south-central United States (Fig. 1; Table 1). The town for which the sequence was named, New Madrid, was one of the earliest American settlements along the Mississippi River. Modern estimates of the magnitudes of the three principal events range from the low to mid M_W 7's (Hough *et al.*, 2000) to over M_W 8 (Johnston, 1996b). The three principal events occurred on 16 December 1811, 23 January 1812, and 7 February 1812 (hereafter, NM1, NM2, and NM3, respectively). Although the precise magnitudes remain in question, the three principal events were clearly very large earthquakes.

Triggered Earthquakes in Northern Kentucky

Hough (2001) presented evidence that NM2 and NM3 were followed by remotely triggered earthquakes in the northern Kentucky/southern Ohio region. These events occurred at approximately 8:45 a.m. local time (LT) on 27 January 1812, 8:30 p.m. (LT) on 7 February 1812, and 10:40 p.m. (LT) on 7 February 1812 (hereafter, NM2-A, NM3-A, and NM3-B, respectively). The events are inferred to have been triggered dynamically for two reasons: (1) the distribution of their felt reports strongly suggests an epicenter well outside the New Madrid region, and (2) the qualitative description of the ground motions from the events is very suggestive of local, moderate events. Shaking from NM2-A is generally described as having been "violent," "severe," or "smart," suggestive of high-frequency shaking and a short duration. Although such descriptive terms can be difficult to interpret in general, in this case the key point is that the shaking was described as being violent, severe, or smart relative to the mainshock ground motions.

Jared Brooks of Louisville, Kentucky, described NM3-A as having been "violent in the first degree, but of too short duration to do much injury." Brooks further described NM3-B as "violent in the second degree, quickly strengthening to tremendous" (McMurtrie, 1839; a complete copy of Brooks' notes can also be found at http://pasadena.wr.usgs.gov/office/hough/Brooks.html). These accounts show that untrained but perceptive observers are capable of distinguishing ground motions from moderate, local events from those produced by large earthquakes at regional distances. However, because different types of ground motion can produce the same modified Mercalli intensity (MMI), it is important to consider the original accounts rather than the interpreted MMI values. (Virtually all of the accounts of the New Madrid sequence discussed in this article are available in the compilation of Street [1982].)

Several modern, instrumental studies of triggered earthquakes have shown that, in at least some cases, the earliest triggered events occur within seconds after the passage of the S/surface waves (e.g., Hill *et al.*, 1993; Power *et al.*, 2001; Hough and Kanamori, 2002). Such events would be

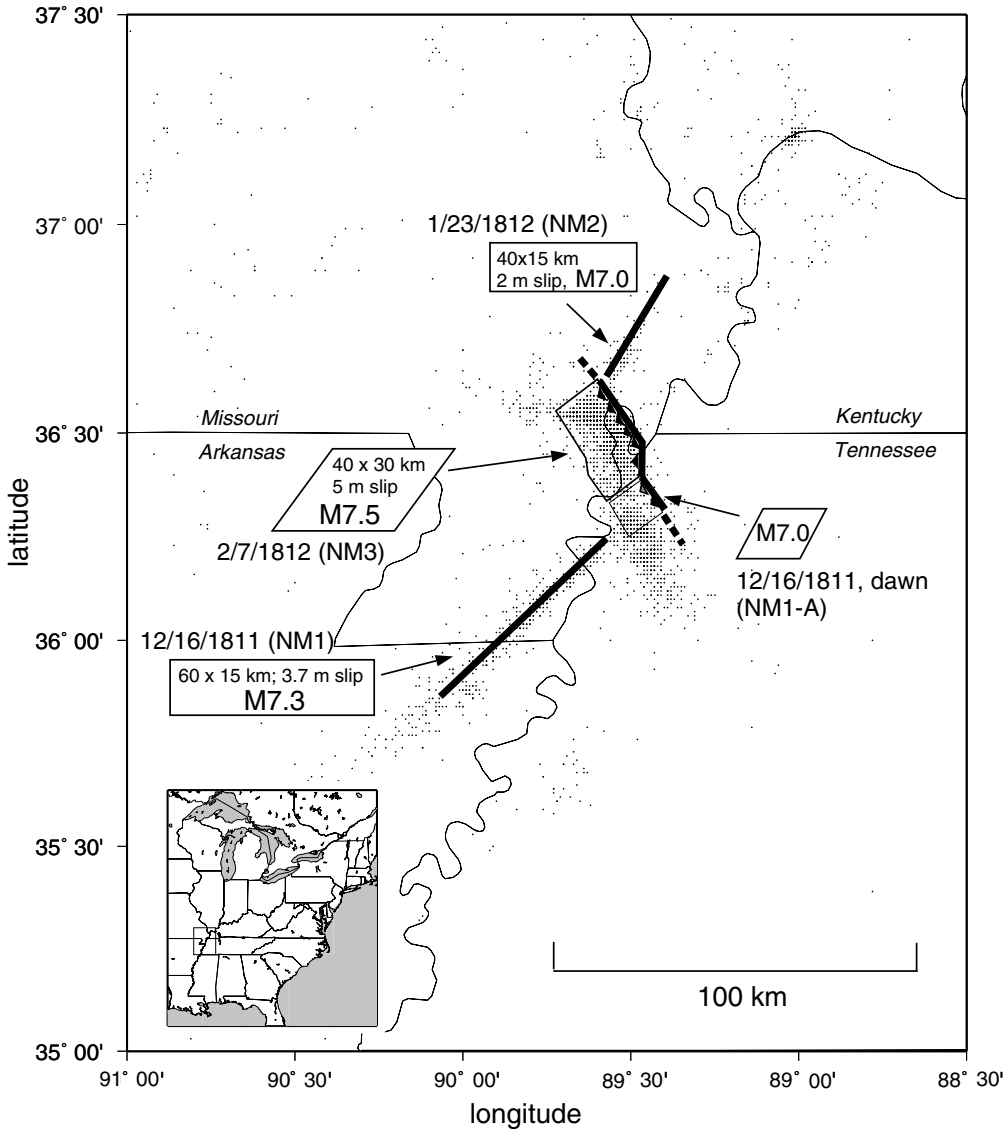

Figure 1. Map showing location of the New Madrid Seismic Zone as illuminated by microseismicity between 1974 and 1996. Locations are from the New Madrid catalog (see Taylor *et al.*, 1991) and are reported only to the nearest hundredth degree. Proposed fault ruptures of the three principal 1811–1812 mainshocks are shown (schematically) (after Johnston and Schweig, [1996], as modified by Hough *et al.* [2000] and Hough *et al.* [2002]). The rupture scenario proposed in this study for the NM1-A aftershock is also shown (Hough and Martin, 2001). Solid line with dashed ends shows inferred location of Reelfoot fault (after Odum *et al.*, 1998).

nearly impossible to identify from noninstrumental data. However, one intriguing account of the NM2 event from Newport, Kentucky (over 500 km from New Madrid), includes the remark that "there was one [shock] which ended with so severe a jolt, that I could scarcely keep my feet." A culminating, strong jolt is not commonly described in accounts of the New Madrid mainshocks, nor is it consistent with expectations for ground motions from a large central U.S. earthquake felt at regional distances. Typically, such ground motions are dominated by the crustal *Lg* waves, which can be felt as a distinct arrival but which also tend to have a prolonged duration. On the other hand, Figure 2 presents an example of a regional strong motion recording of the M_W 7.1 Hector Mine earthquake of 16 October 1999, in which a local $M_W \approx 4.5$ triggered event is clearly evident (Hough and Kanamori, 2002). Were such ground motions to be described qualitatively, one imagines an account very similar to that of NM2 from Newport.

Three other accounts of the NM2 mainshock are found to describe distinct episodes of shaking within a span of minutes. The accounts are from Frankfort, Kentucky; Hodgenville, Kentucky; and Cincinnati, Ohio (Table 2). Al-

Table 1
New Madrid Sequence: Mainshocks, Principal Aftershock, and Triggered Events

Event	Year	Month	Day	Local Time (hh:mm)	Longitude*	Latitude*	M_w†	Ref.‡
NM1	1811	12	16	02:15	−90.00	36.00	7.2–7.3	H00
NM1-A	1811	12	16	07:15	−89.50	36.25	≈7.0	H02
NM2	1812	1	23	08:45	−89.67	36.58	7.0	H00
NM2-A	1812	1	27	09:00	−84.02	38.94	NE	H01
NM3	1812	2	7	03:45	−89.60	36.35	7.4–7.5	H00
NM3-A	1812	2	7	20:30	−84.02	38.94	≈4.5	H01
NM3-B	1812	2	7	22:40	−84.02	38.94	5.0–5.5	H01

*Crudely estimated longitude and latitude in decimal-degrees north and west.
†Range of inferred moment magnitude from previous studies and this study. NE, no estimate.
‡Reference for published magnitude results: H00, Hough *et al.*, (2000); H01, Hough (2001); H02, Hough and Martin (2002).

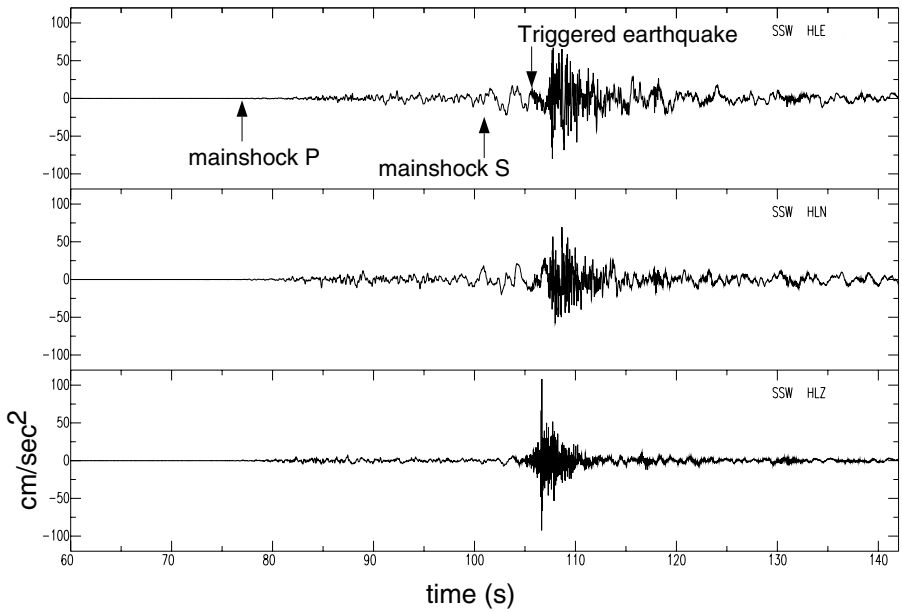

Figure 2. Three components of strong motion data (acceleration in centimeters per second squared) recorded at station SSW (just south of the Salton Sea, southern California) for the Hector Mine mainshock. At this station a moderate remotely triggered earthquake is apparent even without filtering. (Hough and Kanamori, 2001).

Table 2
Accounts from 9:00 a.m. Local Time 23 January 1812 (15 min after NM2)

Location	Longitude	Latitude	Report
Newport, Kentucky	−84.49	39.09	Shaking ended with severe jolt
Frankfort, Kentucky	−84.87	38.19	Followed in a few minutes by "another less violent"
Hodgenville, Kentucky	−85.74	37.57	Followed by another shock that "lasted a minute or two"
Cincinnati, Ohio	−84.52	39.16	"Two or three distinct exascerbations" over 4–5 min

Accounts are contained in compilation of Street [1984] except for that from Cincinnati, which is from Drake [1815].

though it is possible that distinct *S* or *Lg* arrivals were mistaken for distinct earthquakes, we note that the four accounts in Table 2 are the only ones, from a total of 59, that imply distinct events, and they are all clustered within a few hundred kilometers of each other in northern Kentucky.

Interpretation of historic accounts always involves an element of judgement and inferences that are difficult to quantify in precise terms. Here again, however, conclusions can be drawn from the descriptions of earthquakes relative to other events felt by the same observers. For example, the

Newport, Kentucky, account suggests that it was unusual (relative to the ground motions from other large earthquakes in the sequence) that NM2 ended with a strong jolt. The accounts of distinct shocks were also made by observers who had felt, and made no similar observation about, other strong earthquakes. Finally, with any historic earthquake the distribution of felt effects provides *prima facie* evidence that helps constrain the location of any historic earthquake. For both the triggered earthquakes discussed by Hough (2001) and the possible early triggered earthquake discussed here, available accounts strongly suggest locations in the northern Kentucky region.

Taken together, we conclude that the accounts discussed here provide inconclusive but intriguing evidence that moderate remotely triggered earthquakes may have occurred within seconds to minutes of the passage of surface waves from NM2.

Triggered Earthquakes in South Carolina?

Similarly inconclusive but interesting accounts are available from Charleston, South Carolina, where the three principal New Madrid mainshocks were widely felt, even though the city is approximately 970 km from New Madrid. Ground motions from NM1 stopped pendulum clocks and caused suspended objects to swing (see compilation of Street [1982]). Shaking from the NM3 mainshock was stronger than that of the first two mainshocks. One observer wrote, "The whole house rocked like a ship, and we were for at least a minute and an half under the awful impression that we should be buried in its ruins, or swallowed up in the earth . . . " (Street, 1982).

Two reports from the compilation of Street (1982) are suggestive of local earthquakes. One is an account of NM3. A letter from Charleston printed in the *Philadelphia Daily Advertiser* ends with the observation that "no noise was heard by us until after the motion ceased, when a roaring like the troubled sea, occasioned by a momentary purturbation of its waters upon the breakers, was heard very plainly." Although one might appeal to a mechanical source of the noise, small earthquakes in eastern North America are very commonly described in similar terms (and reported as heard rather than felt). This account is thus somewhat similar to the Newport, Kentucky, account of NM2 in its suggestion of especially high frequency ground motions occurring at or near the end of the perceptible shaking.

Secondly, an account published in the *New York Spectator* described earthquakes being felt in Charleston, South Carolina, in the days following the 7 February 1812 event. On 11 February, a shock is described at "24 minutes past 6" that "continued about 30 seconds." The account further states, "A tremulous motion of the earth was distinctly felt through the whole of [February 11th]. Light pendulous articles vibrated frequently." But a high level of activity is not suggested in accounts from other regions, including those from a couple of individuals who recorded every felt event. In Cincinnati, Ohio, Daniel Drake (1815) did report a felt event near 6:00 a.m. (LT) on 11 February 1812, but his account includes no other events through the rest of that day. A second individual, Jared Brooks, kept thorough records of the New Madrid events in Louisville, Kentucky, even rigging up suspended pendulums to sense vibrations (see McMurtrie, 1839). His account also includes an event near 6:00 a.m. (LT) on February 11 and a couple of additional events that day, but he does not describe ongoing activity throughout the course of the day.

The direct evidence for triggered earthquakes in or near Charleston, South Carolina, is less definitive than that for triggered earthquakes in the Ohio Valley, discussed in detail by Hough (2001). However, locally triggered events would help explain why the "mainshock" ground motions were so high at a distance of almost 1000 km. Although site response on coastal plains sediments would have amplified shaking to some extent, the ground motions in South Carolina were more severe than at other coastal sites. In the town of Columbia, in central South Carolina, the February mainshock damaged plaster walls and cracked chimneys.

Taken together, we conclude that the accounts are suggestive of more events and more high-frequency energy than can be accounted for by large events known to have happened in either the New Madrid or the northern Kentucky region. As discussed, one account is consistent with triggered events occurring in the Charleston region within a few minutes to a few days of NM3. The events described by the *New York Spectator* account occurred 4 days after this same mainshock. These delays are very similar to those of other documented cases of remotely triggered earthquakes (e.g., Hill *et al.*, 1993; Brodsky *et al.*, 2000).

Triggered Earthquakes in the Wabash Valley?

A final suggestion of remote triggering following the New Madrid mainshocks is perhaps found in the account of Drake (1815). In his summary observations, Drake wrote, "After the second year of their duration, [they seem] to have ascended the Mississippi to the Ohio, and then advanced up that river about 100 miles, to the United States' Saline [River]; at which place shocks have been felt almost every day for nearly two years." Although this later activity could have been independent of the New Madrid sequence, felt earthquakes are rare in the midcontinent, and a casual link is suggested. Figure 3 shows the locations of the Wabash Valley site as well as other sites where historical accounts suggest that triggered earthquakes occurred, albeit with considerable delay, following the New Madrid mainshocks.

Observations: The 1886 Charleston, South Carolina, Sequence

By virtue of the greater American population density by the end of the nineteenth century, considerably more information is available for the 1886 Charleston, South Carolina, sequence than for the New Madrid events. Previous studies have documented the occurrence of three small to moderate

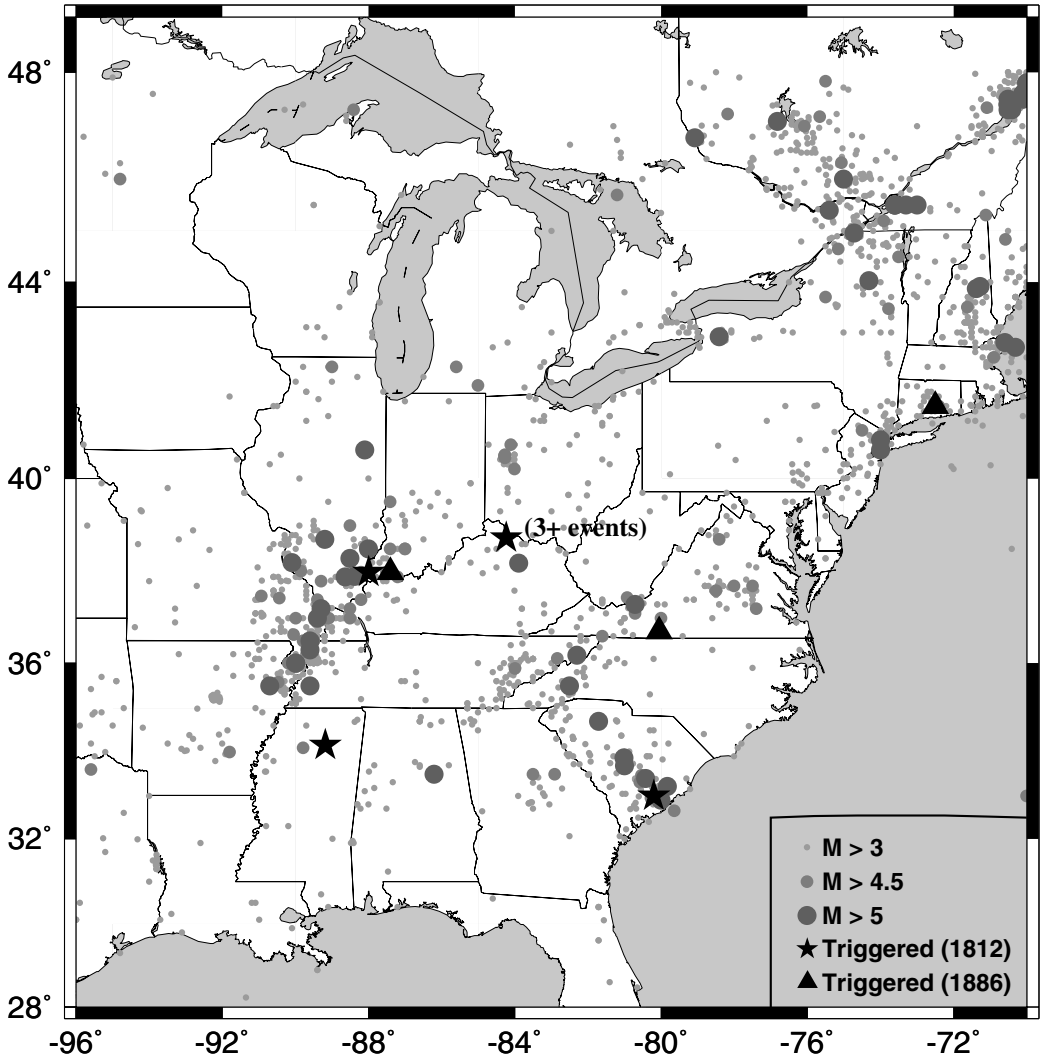

Figure 3. Location of suggested remotely triggered earthquakes generated by the 1811–1812 New Madrid mainshocks (black stars) and the 1886 Charleston, South Carolina, mainshocks (black triangles). Also shown are events in the entire NCEER catalog (gray circles). The NCEER catalog includes events as early as 1627, although its completeness level changes significantly with time.

foreshocks prior to the 1 September 1886 mainshock, as well as dozens of aftershocks (Seeber and Armbruster, 1987).

Four days after the Charleston event, an earthquake occurred well north of South Carolina. Based on its felt reports, it was estimated to be M 3.3, located at 41.5° N −72.5° W, near Moodus, Connecticut (Fig. 3) (see Ebel *et al.*, 1982; Armbruster and Seeber, 1992). In this case, the evidence for the earthquake is fairly clear; the question is whether or not it was triggered by the Charleston earthquake. To answer this question, we consider the observed rate of earthquakes in the central and eastern United States as inferred from the NCEER catalog (Armbruster and Seeber, 1992). On average, approximately 10–15 M 3.3 events occur in this region per year. The odds of an event occurring in a given 4-day window by random chance are thus on the order of 10%–16%.

A search of historic newspapers from Indiana also revealed evidence for an event on 7 September 1886 that was felt at three towns in Indiana separated by over 80 km (Fig. 3). This extent is only slightly smaller than the felt area of an M 3.9 event that occurred in Indiana on December 7 2000 (see http://pasadena.wr.usgs.gov/shake.cus; Wald *et al.*, 1999). Approximately 8–10 events of M 3.6 or larger occur per year in the central and eastern United States. The odds of experiencing such an event in a given week are also on the order of 15%. For both this event and the one near Moodus, the statistical significance thus cannot be proven at better than an 85%–90% confidence level. However, again, we conclude that a causal relationship is suggested.

There is no evidence for larger (M >4) remotely triggered earthquakes outside of South Carolina following the Charleston earthquake, but the "widespread burst of seismicity" identified by Seeber and Armbruster (1987) provides

a measure of evidence for substantial activity well beyond the dimensions of a traditional aftershock zone.

Interpretation

Interpretation of historical accounts is inevitably fraught with a certain degree of uncertainty. However, we have presented evidence suggesting that two of the three principal New Madrid mainshocks triggered local earthquakes near northern Kentucky, the Charleston, South Carolina, area; and in southern Illinois within the Wabash Valley. Following the 1886 Charleston earthquake, available accounts suggest that triggered earthquakes occurred in New England; near Moodus, Connecticut, and in southern Indiana.

The evidence for some of the remotely triggered earthquakes is clearly better than the evidence for others. Of the events discussed, there is compelling evidence for remotely triggered earthquakes in northern Kentucky in 1812; near Moodus, Connecticut, in 1886; and in the Wabash Valley region in 1886. There is also fairly compelling evidence, discussed by Hough and Martin (2002), that the substantial 17 December 1811 aftershock occurred to the south of the New Madrid Seismic Zone. The conclusion that triggered earthquakes occur in zones of persistent seismic activity is therefore supported by the best data as well as by the more speculative results.

Given the established rate of background seismicity in the entire central and eastern United States, it is relatively easy to demonstrate the high likelihood that felt events that occur within a few days are causally related to the preceding large mainshocks. $M_W \geq 5$ events are expected to occur only once every 10–100 yr (Frankel *et al.*, 1996). $M_W \geq 4$ events are expected only about once a year, on average. The odds of experiencing an $M_W \geq 5$ event (anywhere in the central and eastern United States) by random chance in any given 3-day period are on the order of 1 in 1000. If the time window is extended to a full week, the odds are still on the order of only 1 in 500. The odds of smaller events (M 3–4) occurring by random chance in a 3-day window are also low.

The occurrence of remotely triggered earthquakes in intraplate crust is perhaps not surprising. The dynamic stress change associated with surface waves from large earthquakes is thought to trigger earthquakes on faults that are close to failure, essentially "advancing the clock" by suddenly introducing additional stress (Gomberg and Davis, 1996). If anything, dynamic stress changes are expected to be larger at regional distances than those generated by large earthquakes in interplate crust because of the lower attenuation in intraplate regions. But with a lower rate of strain accumulation, the same dynamic stress will potentially represent a much bigger "clock advance."

Additionally, Seeber (2000) has suggested that in low-strain-rate environments, nonelastic permanent deformation might play a more significant role in the cycle of stress accumulation and failure than it does in more active regions. We now explore a quantitative standard rheology model to further investigate the possibility that permanent deformation can help explain our result that remotely triggered earthquakes occur (apparently) commonly in intraplate crust. Considering an overall regional (input) strain rate $d\varepsilon/dt$ given by

$$d\varepsilon/dt = d\varepsilon_e/dt + d\varepsilon_p/dt, \tag{1}$$

where the two terms on the right side of the equation represent elastic and permanent strain rates, respectively, we explore the role of the nonelastic term by assuming that the elastic strain is governed by Hooke's law and the nonelastic term by power-law creep:

$$d\varepsilon/dt = Ed\sigma/dt + C\sigma^k, \tag{2}$$

where E is the reciprocal of Young's modulus, σ is stress, and C and k are constants. For a given strain rate, the effect of a positive, nonzero C will clearly be to slow the rate of stress accumulation. Heuristically this is expected to be more important in a low-strain-rate, intraplate environment than in an interplate region. To further illustrate this point, we make the simplifying assumption $k = 1$, which reduces the second term in equation (2) to simple viscous flow. In this case, C is simply the reciprocal of viscosity and equation (2) has the solution

$$\sigma = S' - [(S' - \sigma_0)\exp(-(C/E))t], \tag{3}$$

where S' is $(1/C)\, d\varepsilon/dt$ and σ_0 is the stress at the start of the earthquake cycle. Making the final simplifying assumption that $\sigma_0 = 0$, one obtains

$$\sigma = S'(1 - \exp(-(C/E)t)). \tag{4}$$

Using this equation, one can investigate the effect of a nonzero C term for differing values of strain rate. These calculations are meant to be only illustrative, as actual aseismic deformation processes in the Earth are complex and will be governed by both equations and parameters that are difficult to estimate.

However, it is instructive to explore the behavior of equation (4) assuming nominal values for all of the terms. Assuming a value for E of 10^{-11} Pa^{-1}, we start by calculating $\sigma(t)$ for an interplate strain rate of 1×10^{-6} yr^{-1} given C values of 10^{-22} and 10^{-23}. This range of C values is intermediate between that predicted from the viscosity of granite, which is approximately 10^{18}–10^{19} Pa sec and that of the shallow mantle (10^{23}–10^{24} Pa sec). As shown in Figure 4, the smaller C term implies a negligible effect of nonelastic deformation, in which case the solution reduces to the elastic case. The larger value of C, however, has a more pronounced effect. If one repeats the calculations for a strain-rate value 1–2 orders of magnitude smaller (5×10^{-8} yr^{-1}), one obtains qualitatively similiar results (Fig. 4). However, if the failure stress in both cases is roughly comparable (for illus-

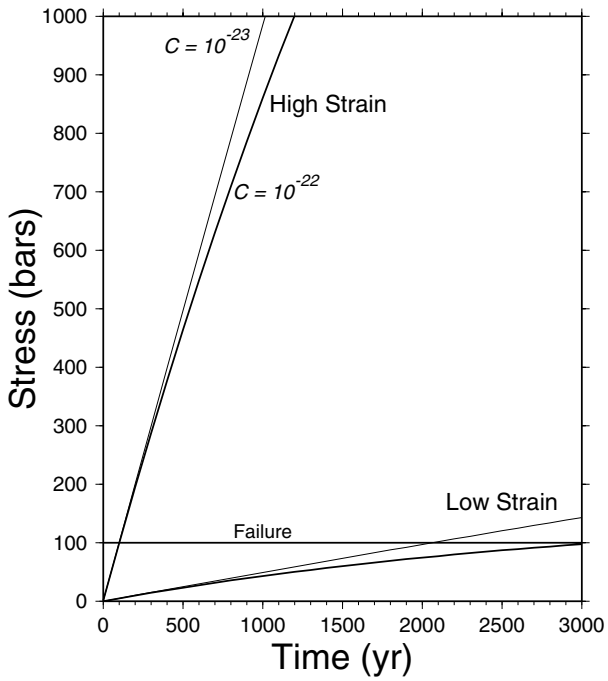

Figure 4. Top set of curves shows solution of equation (4) for a strain rate of 10^{-6} yr^{-1} and C values of 10^{-23} (solid line) and 10^{-22} (dashed line). Bottom set shows results for the same two C values but for a strain rate of 5×10^{-8} yr^{-1}. Horizontal line indicates failure stress, assumed to be 100 bars.

tration, approximately 100 bars), then nonelastic deformation would play a very different role in the two cases. In the high-strain-rate case failure stress is reached centuries before nonelastic deformation becomes significant, while in the lower strain-rate case nonelastic deformation affects stress accumulation for a long time before the failure stress is achieved.

Again, these calculations are intended for illustration only, as they represent both gross simplifications of the physics and gross approximations of parameters. Although it would be possible to more fully explore both full solutions to equation (2), and the dependence of solutions on variations in the parameters, we do not mean to emphasize the particular process of power-law creep, which may or may not represent the actual aseismic deformation processes that occur in the Earth. Rather, these equations, which represent a standard Maxwell elastoviscous rheology, are used to explore the effect that aseismic deformation processes will have on stress accumulation. Even simple qualitative considerations show that, given a constant process of strain accumulation, aseismic deformation will slow the accumulation of stress available to drive earthquakes. The effect of aseismic deformation will clearly also be more important in low-strain-rate environments, where the repeat time of earthquakes is much longer.

These calculations do suggest that, for a reasonable model and choice of parameters, aseismic deformation can have an important effect on the stress accumulation cycle. In particular, the calculations illustrate that even a small amount of nonelastic deformation could play an important role in the very slow stress accumulation in intraplate crust. For a given failure level there is, in fact, a strain rate sufficiently low that all strain will be accommodated by permanent deformation. In general, the results suggest that intraplate faults might remain close to their failure stress for a much longer part of the earthquake cycle than faults in interplate crust, thus making them potentially more responsive to dynamic stress changes associated with large earthquakes.

Discussion and Conclusions

A number of recent studies have documented remotely triggered earthquakes following mainshocks of $M_W \approx 7$ and greater, including the 1992 Landers, California, and 1999 Izmit, Turkey, events (Hill *et al.*, 1993; Gomberg and Davis, 1996; Bodin and Gomberg, 1994; Brodsky *et al.*, 2000; Gomberg *et al.*, 2001). Strains of 10^{-5} to 10^{-6} (at frequencies of 0.1–0.5 Hz) have been shown to be sufficient to trigger earthquakes at regional (500–1000 km) distances in interplate regions (e.g., Gomberg and Davis, 1996). Previous results also predict that, even if the New Madrid mainshocks were no larger than M_W 7.0–7.3, they would have produced dynamic strains in this range at distances of 500–1000 km (Gomberg and Davis, 1996).

In some respects, remotely triggered earthquakes may seem less probable in intraplate regions. For one thing, one would not necessarily expect triggering threshholds to be the same in intraplate regions. Previous studies of triggered events in California have concluded that triggering occurs preferentially in regions of active volcanic and/or geothermal activity, either because of the presence of fluids (Linde *et al.*, 1994; Sturtevant *et al.*, 1996; Brodsky *et al.*, 1998) or because faulting in extensional regions occurs at low stresses (Hough and Kanamori, 2002). Neither condition is expected to be found in central or eastern North America. On the contrary, several lines of evidence suggest that earthquakes in intraplate environments are associated with high stress drop values (e.g., Scholz *et al.*, 1986).

Nonetheless, our results suggest that remotely triggered earthquakes occur commonly following large (M_W upward of 7) earthquakes in central and eastern North America. Although it remains to be seen whether this is true in stable continental regions worldwide, we conclude that, in a low-strain-rate environment, a dynamic stress change can represent a large clock advance. We have also shown that aseismic deformation can act to increase the length of time that faults are close to failure in low-strain-rate environments.

Although these considerations are expected to be valid in general for intraplate crust, the list of potential sites of possible remotely triggered events presented in this study is striking in one respect: they are all sites known to have been seismically active during historic times. The NCEER catalog

includes only 70 $M \geq 5$ events between the late 1660s and 1985, yet there is at least one of these events within 100 km of four of the five (inferred) triggered event locations (Fig. 3). The remaining location, Moodus, Connecticut, has been the site of microearthquake activity in recent years and experienced an event of estimated M 4.5 in 1791 (Armbruster and Seeber, 1992).

The other regions are generally recognized to be loci of continued seismic activity: coastal South Carolina, southern Illinois, and Moodus, Connecticut, in particular. Although northern Kentucky is not known for high levels of earthquake activity, it was the site of the M 5.1 Sharpsburg, Kentucky, earthquake of 1980 (Mauk *et al.*, 1982).

It has been suggested that intraplate crust is characterized by a critical stress state (see Townend and Zoback [2000] for a summary). According to this model, stress in the crust is pervasively close to the failure stress of faults, although stress will still be accumulated and released over the course of a seismic cycle. However, the observations presented in this study suggest that triggered earthquakes do not occur with a random spatial distribution in central and eastern North America, but rather cluster near locations of past (and perhaps future) activity. This perhaps implies that the critical stress model is appropriate for limited regions within intraplate crust, regions in which local stress perturbations are present.

Moreover, if triggered earthquakes occur preferentially in regions where larger past events have occurred, this suggests that critically stressed regions of intraplate crust are not relaxed by a single large event. This conclusion is conceptually consistent with the specific mechanical model recently presented by Kenner and Segall (2000) for the New Madrid Seismic Zone, a model in which relaxation is accommodated by a series of large, quasiperiodic earthquakes that continue for thousands of years. In the model of Kenner and Segall (2000), the prolonged sequence of earthquakes is explained as a consequence of feedback between a fault in the upper crust and a viscoelastic lower crust through which the driving tractions are transmitted. This contrasts with plate boundary regions, where plate motions provide driving tractions that act on faults through the elastic upper crust and are relieved by large earthquakes with no feedback (with the driving tractions).

We suggest that our results provide evidence that, in a qualitative sense, the model of Kenner and Segall (2000) is applicable for intraplate earthquakes; that is, while earthquakes in other intraplate regions might not be related to the same tectonic setting as found at New Madrid, they are likely to reflect similarly local stress perturbations rather than stresses resulting from far-field plate motions. Remotely triggered earthquakes may thus serve as effective beacons that identify locations where such local stress concentrations exist.

The hypotheses presented in this article could be tested with additional data, but appropriate data are limited given the slow rate of large earthquakes in intraplate regions. Historic accounts of the 1811–1812 New Madrid and 1886 Charleston earthquakes have been studied in considerable detail already. We consider it unlikely that compelling evidence for other triggered earthquakes would be found, although accounts from the Charleston earthquake have not, to our knowledge, been revisited in detail since remotely triggered earthquakes were first recognized in the early 1990s. It would be possible to search modern catalogs for evidence of remotely triggered earthquakes following moderate earthquakes such as the 1989 M 6 Saguenay, Quebec, earthquake. It might also be possible to investigate historic accounts of large earthquakes in other intraplate regions worldwide, in places such as India. However, demonstrating a causal link between mainshocks and remotely triggered earthquakes requires catalogs of sufficient quality to establish the background rate of small to moderate events. In many parts of the world, this presents an additional limitation. Nonetheless, careful reinterpretation of historic accounts of earthquakes in other regions might yield evidence for larger remotely triggered earthquakes.

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