Stress evolution and seismicity in the central-eastern United States: Insights from geodynamic modeling

Qingsong Li
Mian Liu
Qie Zhang
Eric Sandvol

Department of Geological Sciences, University of Missouri, Columbia, Missouri 65211, USA

ABSTRACT

Although the central and eastern United States is in the interior of the presumably stable North American plate, seismicity there is widespread, and its causes remain uncertain. Here, we explore the evolution of stress and strain energy in intraplate seismic zones and contrast it with that in interplate seismic zones using simple viscoelastic finite-element models. We find that large intraplate earthquakes can significantly increase Coulomb stress and strain energy in the surrounding crust. The inherited strain energy may dominate the local strain energy budget for thousands of years following main shocks, in contrast to interplate seismic zones, where strain energy is dominated by tectonic loading. We show that strain energy buildup from the 1811–1812 large events in the New Madrid seismic zone may explain some of the moderate-sized earthquakes in this region since 1812 and that the inherited strain energy is capable of producing some damaging earthquakes (M >6) today in southern Illinois and eastern Arkansas, even in the absence of local loading. Without local loading, however, the New Madrid seismic zone would have remained in a stress shadow where stress has not been fully restored from the 1811–1812 events. We also derived a Pn velocity map of the central and eastern United States using available seismic data, the results do not support the NMSZ being a zone of thermal weakening. We simulated the long-term Coulomb stress in the central and eastern United States. The predicted high Coulomb stress concentrates near the margins of the North American tectosphere, correlating spatially with most seismicity in the central and eastern United States.

Keywords: NMSZ, intraplate earthquakes, stress, seismicity, modeling.

INTRODUCTION

Plate tectonic theory provides a successful geodynamic framework for understanding the majority of earthquakes that occur along plate-boundary zones. However, it offers no ready explanation for earthquakes in the presumably rigid plate interiors. One such region is the central-eastern United States, defined broadly as the region of the continental United States east of the Rocky Mountains. Although the central and eastern United States is in the middle of the North America plate where Cenozoic crustal deformation is minimal, both historic earthquakes and instrumentally recorded earthquakes are abundant. Major seismic zones in this area include the following (Dewey et al., 1989) (Fig. 1):
1. The New Madrid seismic zone and Mississippi Embayment, which was the site of the famous 1811–1812 large earthquakes. The magnitudes of the largest three events were Mw 7–7.5 (Hough et al., 2000). Paleoseismic results indicate that major earthquakes occurred ca. 900 and 1400 A.D. (Kelson et al., 1996; Tuttle et al., 2002). Modern instrumentation has recorded thousands of events since 1977.

2. Southern Valley and Ridge, also referred to as the eastern Tennessee seismic zone, where modern seismicity is concentrated beneath the Valley and Ridge Province, near the western edge of the Appalachians. The largest historical event here was the M 5.8 Giles County, Virginia, earthquake of 31 May 1897 (Nuttli et al., 1979).

3. South Carolina seismic zone, where the best-known event was a destructive (M ~6.5–7.0) event near Charleston, South Carolina, on 31 August 1886 (Nuttli et al., 1979). Paleoseismic studies indicate at least two prehistoric earthquakes in the past 3000 yr (Obermeier et al., 1985; Talwani and Cox, 1985).

4. New England and the St. Lawrence River valley: Earthquake epicenters in central New England, upstate New York, and adjacent Canada form a northwest-trending belt of seismicity, sometimes called the Boston-Ottawa zone (Diment et al., 1972; Sbar and Sykes, 1973). The largest historic earthquake in the U.S. part of this zone was probably the M ~6 Cape Ann, Massachusetts, earthquake of 1755 (Street and Lacroix, 1979). Further north in the St. Lawrence River valley, numerous events with M 6–7 have been recorded.

Despite intensive studies, the mechanics of earthquakes in the central and eastern United States remain poorly understood. Some workers have suggested that these seismic zones occur within ancient rifts, thus proposing crustal weakness as the main cause of these earthquakes (e.g., Johnston and Kanter, 1990; Johnston, 1996). Others have suggested stress concentration by various factors, including regional and local crustal structures, as the main cause (e.g., Grana and Richardson, 1996; Stuart et al., 1997; Kenner and Segall, 2000; Grollimund and Zoback, 2000).
2001; Pollitz et al., 2001a). Although intraplate earthquakes are commonly believed to differ fundamentally from interplate earthquakes, their differences in dynamics are not clear. In this study, we first explore the basic mechanics of intraplate seismic zones and compare them to the mechanics of interplate seismic zones. We then apply the results to investigate seismicity in the New Madrid seismic zone. Finally, we present a regional geodynamic model of the central and eastern United States constrained by the lithospheric structure based on seismic studies by others and our Pn tomography.

THE MECHANICS OF INTRAPLATE VERSUS INTERPLATE SEISMIC ZONES

We developed three-dimensional viscoelastic models to explore the differences in stress evolution between intraplate and interplate seismic zones. To illustrate the basic physics, we kept the models relatively simple. We considered two contrasting properties of these zones: (1) intraplate seismic zones are of finite length, surrounded by strong ambient crust, whereas interplate seismic zones are effectively infinitely long; and (2) tectonic loading for intraplate seismic zones is applied at far-field plate boundaries and typically produce low strain rates, whereas interplate seismic zones are loaded directly by relative motions of tectonic plates, causing relatively high strain rates (Fig. 2). The model rheology is a viscoelastic (Maxwellian) medium. The model domain is 500 km × 500 km. Both models include a 20-km-thick stiff upper crust and a ductile lower crust. The viscosity for the upper crust is $8 \times 10^{23}$ Pa s, which makes it essentially elastic for the time scales considered here (thousands of years). For the lower crust, a range of viscosity values ($1.0 \times 10^{19}$ to $1.0 \times 10^{21}$ Pa s) was explored for the effects of postseismic relaxation. For the intraplate model (Fig. 2A), a 150-km-long fault zone was used to simulate a finite seismic zone. The boundary conditions included

Figure 2. Finite-element models for intraplate (A) and interplate (B) seismic fault zones (in dark shading). Points M and N are where stress evolution is shown in Figure 4, and points O, P, and Q are where stress evolution is shown in Figure 7.
0.5 mm/yr compression imposed on the two sides of the model domain. The resulting strain rate is \( \sim 2.0 \times 10^{-9} \) yr\(^{-1}\), which is close to the upper bound for the central and eastern United States (Newman et al., 1999; Gan and Prescott, 2001). Young’s modulus and the Poisson’s ratio are 8.75 \times 10^{10} \) Pa and 0.25, respectively, for the entire crust (Turcotte and Schubert, 1982). For the interplate model, the fault zone cuts across the entire model domain (Fig. 2B), and the boundary condition is 10 mm/yr on both sides, causing an average slip rate of \( \sim 28 \) mm/yr along the fault zone in the model, which is similar to that on the San Andreas fault (Bennett et al., 2004; Becker et al., 2005). In both models, the fault zones are represented by special elements, on which we simulate earthquakes by using instant plastic strain to lower the stress to below the yield strength (Li et al., 2005).

### Intraplate Seismic Zones

Many seismic zones in the central and eastern United States are marked by past large earthquakes, the initial triggering mechanisms of which are uncertain. Here, we focus on stress evolution in the seismic zones following a large earthquake.

Figure 3 shows the calculated evolution of the Coulomb stress following a large intraplate earthquake. The Coulomb stress on a plane is defined as

\[
\sigma_f = \tau_{\beta} - \mu \sigma_{\beta},
\]

where \( \tau_{\beta} \) is the shear stress on the plane, \( \sigma_{\beta} \) is the normal stress, and \( \mu \) is the effective coefficient of friction (King et al., 1994). Outside the main fault zone, we calculated the optimal Coulomb stress, which is the stress on planes optimally orientated for failure (King et al., 1994). We assumed that, initially, the stress in the upper crust is close to the yield strength, a condition that might be applicable to many continental interiors (Townend and Zoback, 2000; Zoback et al., 2002) and is consistent with the widely scattered seismicity in and around the seismic zones in the central and eastern United States (Fig. 1). The model started with a large earthquake, simulated by a 7.5 m sudden slip across the entire fault plane. This event is equivalent to an \( M_w \sim 8.0 \) earthquake, which caused \( \sim 5 \) MPa stress drop within the fault zone. Coseismic stress release from the fault zone migrates to the tip regions of the fault zone and loads the lower crust below the fault zone. Postseismic viscous relaxation in the lower crust then causes the stress to reaccumulate within the upper crust, mainly near the tips of the fault zone. Similar results have been reported in previous viscoelastic models (Freed and Lin, 2001; Pollitz et al., 2001b).

Note that 200 yr after the main earthquake, the fault zone remains in a stress shadow where the stress relieved during the earthquake has not been fully restored. This is mainly due to slow tectonic loading, the effects of which are insignificant over 200 yr (compare Figures 3B and 3C). In this model, we assumed a complete healing of the fault zone, such that the yield strength returned to the original level immediately following the main shock. If the fault zone were unhealed or partially healed, stress reaccumulation within the fault zone would be even slower.

In addition to the rate of tectonic loading, postseismic stress evolution depends on the rheology of the lithosphere, especially the lower crust. Figure 4 shows the effects of lower-crustal viscosity on the modeled stress evolution. A less viscous lower crust causes more rapid viscous relaxation and stress reloading in the upper crust. However, without fast tectonic loading from the far field, the total amount of stress restoration within the fault zone is

![Figure 3. Predicted Coulomb stress change following a large earthquake in the intraplate seismic zone. (A) Coseismic stress change. (B) The sum of coseismic and postseismic (200 yr) stress change. (C) Same as B but without boundary loading. The bottom panels are depth sections, with 200% vertical exaggeration. The white lines (map view) and black frames (depth section) show the ruptured fault zone. Values were calculated assuming a viscosity of \( 10^{19} \) Pa s for the lower crust.](image-url)
largely determined by the stress relieved from the earthquake. For the viscosity range typical of the lower crust ($10^{21}$ Pa s), viscous relaxation and far-field loading cannot fully restore stress in the fault zone thousands of years after the major earthquake. This may be a fundamental difference between intraplate and interplate seismic zones; the latter are directly loaded by plate motions at high rates, and a ruptured fault segment can also be influenced by earthquakes on nearby fault segments (see next section).

There is a migration and accumulation of strain energy associated with the stress evolution, defined as

$$E = \frac{1}{2} \sigma'_{ij} \varepsilon''_{ij},$$

where $\sigma'_{ij}$ and $\varepsilon''_{ij}$ are the deviatoric stress and strain tensors, respectively, using the Einstein summation convention for indexes $i$ and $j$.

Figure 5 shows the coseismic and postseismic changes of strain energy. Similar to the stress change (Fig. 3), most increase of strain energy is near the tips of the fault zone. Because of the slow tectonic loading, much of the strain energy is inherited from the main shock. It would take thousands of years for the far-field tectonic loading to accumulate a comparable amount of strain energy.

**Interplate Seismic Zones**

We may better appreciate the stress and strain energy evolution in intraplate seismic zones by contrasting them with interplate seismic zones (Fig. 2B). Some of the processes are similar.

An interplate earthquake relieves stress to the lower crust and the tips of the ruptured fault segment. Viscous relaxation in the lower crust further loads the upper crust, similar to what occurs in intraplate seismic zones. However, the high strain rates associated with plate motions restore stress in the ruptured segment more quickly than in intraplate fault zones. The essentially infinitely long fault zone also confines earthquakes to largely be within, and migrate along, the fault zone (Fig. 6A). Usually, other segments rupture before an earthquake repeats on the same segment (Fig. 6B). Thus, for each segment of the ruptured fault zone, postseismic stress recovery may be affected by three major factors: tectonic loading, viscous relaxation, and stress migration from nearby earthquakes. Figure 7 illustrates such stress evolution at three neighboring points in the fault zone. An earthquake at one of these points causes an instant stress drop. Postseismic stress restoration at the ruptured segment is fast within the first few tens of years because of both tectonic loading and viscous relaxation in the lower crust, transferring stress to the upper crust. A period of roughly steady-state stress buildup follows, resulting from tectonic loading. Sudden stress jumps may occur when a nearby segment ruptures, which may trigger a new earthquake. Such dynamic behavior has been reported in many interplate seismic zones, including the San Andreas fault (Stein et al., 1997; Rydelek and Sacks, 2001; Lin and Stein, 2004).

We have shown that, in an intraplate seismic zone, strain energy released from a large earthquake will migrate to the surrounding regions and dominate the local strain energy budget for thousands of years (Fig. 5). This is generally not true for...
interplate seismic zones, where tectonic loading dominates the strain energy budget. Figure 8 shows one selected episode of the interplate model experiment. Here, postseismic energy evolution is influenced by strain energy migration from the ruptured segment, viscoelastic reloading, and tectonic loading. Figures 8C and 8D compare strain energy in the model crust 200 yr after the earthquake, with and without tectonic loading. Clearly, tectonic loading dominates the strain energy evolution. To show the cumulative strain energy produced from tectonic loading, we artificially prohibited earthquakes for the period shown here. In reality, strain energy will be modulated by ruptures of other segments of the fault zone (Figs. 6 and 7).

STRESS EVOLUTION AND SEISMICITY IN THE NEW MADRID SEISMIC ZONE

In this section, we apply the model results to the New Madrid seismic zone, perhaps the best known seismic zone in the central and eastern United States. At least three large earthquakes occurred here within three months in the winter of 1811–1812. The magnitudes of these events have been estimated to be 7–7.5 (Hough et al., 2000). Since then, a dozen or so major events (M 5–6) have occurred in the New Madrid seismic zone and surrounding regions, and modern instruments have recorded thousands of small events in the past few decades (Fig. 9).

The New Madrid seismic zone fault zone is generally delineated by seismicity. Only one segment of the fault system, the NW-trending Reelfoot fault, has clear surface expression. Other parts of the New Madrid fault zone, including the southwestern segments (Cottonwood Grove fault) and the northeastern segment (New Madrid North fault), are inferred mainly from reflection seismic and aeromagnetic data, and seismicity (Hildenbrand and Hendricks, 1995; Johnston and Schweig, 1996). The Reelfoot fault is a reverse fault, whereas the southwestern and northeastern segments are inferred to be right-lateral faults from morphologic and geologic features (Gomberg, 1993). These faults are believed to be within a failed rift system formed in Late Proterozoic to Early Cambrian times (Ervin and McGinnis, 1975). Herein, we use “New Madrid seismic zone” when referring to the geographic region of concentrated seismicity, and “New Madrid fault zone” when referring to these fault structures.

The results in the previous section indicated that following the 1811–1812 large earthquakes, the New Madrid fault zone would still remain in a stress shadow with Coulomb stress lower than the pre–1811–1812 level; this condition would be unfavorable for repetition of large earthquakes. This has important implications for assessing earthquake hazards in the New Madrid seismic zone. Here, we further explore this issue with a more realistic model (Fig. 9). The New Madrid fault zones are represented by two vertical strike-slip branches that approximate the northeastern and southwestern fault segments, connected by a NW-trending reverse fault that dips 45° SW, based on inferred geometry of the Reelfoot fault (Chiu et al., 1992; Mueller and Pujol, 2001). The compressive stresses across the North American plate were simulated by applying a 0.5 mm/yr velocity boundary condition on the eastern and western edges of the model domain (Fig. 9). This produced a strain rate of $-2 \times 10^{-9} \text{yr}^{-1}$ within the model domain, which is likely the upper bound of internal deformation rate within the North American plate based on global positioning system (GPS) and seismological data (Anderson, 1986; Newman et al., 1999; Zoback et al., 2002). Other model parameters, including the initial conditions and rheological structures, are similar to those in Figure 2A.

Previous work has concluded that three large earthquakes occurred on the Reelfoot fault and the southwestern branch of the New Madrid seismic zone (the Cottonwood Grove fault) between December 1881 and February 1882 (Johnston, 1996; Johnston and Schweig, 1996). Some recent studies have suggested that
there may have been four large events in the 1811–1812 sequence of events (Hough et al., 2000), and at least one of the main shocks may have been outside the New Madrid seismic zone (Mueller et al., 2004; Hough et al., 2005). Because our focus here is the long-term effects of the large 1811–1812 events, we ignore the detailed rupture sequences and treat these large events as having occurred simultaneously along the entire fault zones. This was simulated with ~5 m instant slip along the model fault zones, resulting in a Coulomb stress drop of 5 MPa within the fault zones, as estimated by Hough et al. (2000). Figure 10 shows the calculated Coulomb stress evolution following the 1811–1812 events. In the upper crust, the increases in maximum stress occur near the NE and SW ends of the New Madrid fault zones. Conversely, stress decreases within the New Madrid fault zones and along a broad zone extending roughly NNW-SSE across the New Madrid seismic zone. This general pattern is similar to that in Figure 3.
Figure 6 (continued). (B) Depth sections of the predicted stress evolution shown in A. The labels of the panels correspond to the map-view panels in A. The white lines (map view) and black frames (depth sections) mark the fault segments that have ruptured during the 20 yr time interval.

Figure 7. Predicted Coulomb stress evolution at three points on the interplate seismic fault zone (see Fig. 2B). At each point, the cycle of stress evolution includes four stages: (1) stress drop during an earthquake; (2) accelerated stress restoration because of viscous relaxation of the lower crust; (3) steady stress increase mainly from tectonic loading; and (4) stress jump due to triggering effect of nearby earthquakes.
Each of the large 1811–1812 events was followed by numerous large aftershocks (M > 6.0) (Johnston and Schweig, 1996), and since 1812, a dozen or so moderate-sized events (M > 5) have occurred in the New Madrid seismic zone and surrounding regions (Fig. 9). Although not all these events were included in the calculation, their effects have likely been minor in terms of energy release. This can be seen from the stress perturbation by two of the largest earthquakes in the New Madrid seismic zone region since 1812: the 1895 Charleston, Missouri, earthquake (M = 5.9) and the 1843 Marked Tree, Arkansas, earthquake (M = 6.0) (Stover and Coffman, 1993) (Fig. 10). The results show some local stress changes near the epicenters of these events, but the general stress pattern remains dominated by the 1811–1812 large events, leaving the New Madrid seismic zone in a stress shadow where stress has not reached the pre–1811–1812 level. The largest Coulomb stress increases are in southern Illinois and eastern Arkansas. Interestingly, these are located where many of the major earthquakes (M > 5) since 1812 have occurred (Fig. 10). The spatial correlation is not perfect because seismicity is controlled by both stress and crustal strength, but the lateral variations of strength in the ambient crust are not included in the model. Thus, the clustering of moderate earthquakes in southern Illinois and western Indiana may be attributable to both the increased Coulomb stress and the relatively weak crust in the Wabash Valley seismic zone, and seismicity near the Missouri-Illinois border, where the Coulomb stress actually decreased following the 1811–1812 large events, may indicate weakness in both crust and the uppermost mantle (see following). If one of the 1811–1812 main shocks occurred outside of the New Madrid seismic zone, as suggested by Mueller et al. (2004) and Hough et al. (2005), the predicted stress field may differ somewhat from that in Figure 10.

The predicted stress evolution is consistent with seismic energy release in the New Madrid seismic zone and surrounding regions following the 1811–1812 large events. Figure 11A shows the calculated seismic energy release based on historic and modern earthquake data from the National Earthquake Information Center (NEIC) catalog (http://neic.usgs.gov/neis/epic/epic.html). We used the Gutenberg-Richter formula (Lay and Wallace, 1995), and approximated all magnitudes as Ms. The spatial pattern is dominated by a dozen moderate-sized events (M > 5) since 1812, especially the two M ~ 6 events near the NE and SW tips of the New Madrid seismic zone (Fig. 9). Figure 11B shows the excess strain energy, calculated by assigning a strain change in each element, if needed, to bring the deviatoric stress below the yield strength of the crust during a time step. A vertical integration of the product of such strain changes and stress gives the total excess strain energy accumulated over a single time step at a given place. The spatial pattern of the calculated excess strain energy is consistent with the seismic energy release in the past two centuries (Fig. 11A), but the magnitude of the excess strain energy is two to three orders higher, presumably because not all energy has been released via earthquakes. The relation between strain energy before the large earthquakes, the energy released during them, and the fraction of energy radiated as seismic waves remains unclear (Kanamori, 1978). It is possible that the fraction of the energy release that is radiated as seismic waves (seismic efficiency) is
only ~10% (Lockner and Okubo, 1983). Multiplication of the estimated seismic energy release (Fig. 11A) by a factor of 10 provides an estimate of total energy released by earthquakes. Subtracting it from the excess strain energy in Figure 11B gives the residual strain energy, some of which may be released by future earthquakes. The partition between seismic and aseismic energy release is uncertain—estimations range from 2% to 80% (Ward, 1998). Figure 11C shows the estimated seismic energy in the New Madrid seismic zone region assuming that 10% of the total excess strain energy will be released in future earthquakes. This energy is capable of producing a number of Mw 6–7 earthquakes in southern Illinois and eastern Arkansas.
The basic mechanics illustrated by the simple model of intraplate seismic zones (Fig. 2A) thus appear to apply to the New Madrid seismic zone. Without some kind of local loading, the New Madrid fault zone is expected to remain in a stress shadow today, and the repetition of large earthquakes within the New Madrid fault zones would be unlikely in the next few hundred years. On the other hand, much of the strain energy released by the 1811–1812 events has migrated to southern Illinois and eastern Arkansas, where a number of moderate earthquakes have occurred since 1812. Based on this model, the residual strain energy in these regions, even without additional contribution from local loading, is capable of producing damaging earthquakes.

LITHOSPHERIC STRUCTURE AND SEISMICITY IN THE CENTRAL AND EASTERN UNITED STATES

So far our discussion of intraplate earthquakes has focused on postseismic evolution after a large earthquake. Given the low strain rates in the central and eastern United States and most other stable continents, it remains unclear what caused these large earthquakes in the first place. It has been suggested that most intraplate earthquakes, especially the large events (Mw >6.0), occur in ancient rift zones (Johnston and Kanter, 1990). This is true for the New Madrid seismic zone, which is within the Mesozoic Reelfoot rift system (Ervin and McGinnis, 1975). Most hypotheses of local loading mechanisms responsible for the large earthquakes...
in the New Madrid seismic zone are based on inferred properties of the rift, including the sinking of an intrusive mafic body in the rift (Grana and Richardson, 1996; Pollitz et al., 2001a), detachment faulting at the base of the rift (Stuart et al., 1997), and an unspecified sudden weakening of the lower crust (Kenner and Segall, 2000). However, Figure 12 shows that not all seismic zones in the central and eastern United States are associated with rifts, and not all rifts are seismically active. One notable example is the Mid-Continent Rift, one of the most prominent rift systems in the central and eastern United States that has been essentially aseismic in historic times. On the other hand, earthquakes in the central and eastern United States appear to concentrate along the margins of the seismologically inferred North American craton, or the “tectosphere” (Jordan, 1979), which is defined by abnormally thick lithosphere.

Stress Field in the Central and Eastern United States

Could the lithosphere-tectosphere transition zone concentrate stresses and thus contribute to seismicity in the central and eastern United States? To address this question, we developed a finite element model for the central and eastern United States (Fig. 13). To simulate the long-term stress pattern, we treated the lithosphere as a power-law fluid continuum with a relatively high viscosity \(10^{24} \text{ Pa s}\), underlain by a viscous asthenosphere with a lower viscosity \(10^{21} \text{ Pa s}\). The thickness of the model lithosphere was based on seismologically derived thermal lithosphere thickness (Goes and van der Lee, 2002). The bottom of the model domain was a free slip boundary. The model domain was loaded on both sides by a 30 MPa compressive stress oriented N60°E, the direction of maximum tectonic compression for the central and eastern United States (Zoback and Zoback, 1989). The calculated Coulomb stress was concentrated in the zones of relatively thin lithosphere, around the margin of the North American tectosphere and under the Mississippi Embayment (Fig. 14). The regions of high Coulomb stress showed a strong spatial correlation with seismic zones in the central and eastern United States, suggesting that the lateral heterogeneity of lithospheric structures is an important factor for seismicity in the central and eastern United States.

Pn Tomography of the Central and Eastern United States

The calculated high Coulomb stress in the Mississippi Embayment results from relatively thin lithosphere inferred from low Vs velocities (Goes and van der Lee, 2002) (Fig. 12), which relate to heat-flow anomalies in the New Madrid seismic zone region (Liu and Zoback, 1997). To refine the uppermost mantle velocity structure beneath this area, we derived a preliminary Pn velocity map (Fig. 15). Pn is a leaky mode guided wave that travels primarily through the uppermost mantle and is therefore most sensitive to seismic velocity fluctuations there. Pn tomography has become a common method of exploring the lithospheric mantle velocity structure (Hearn et al., 1994). This method commonly uses a least-squares algorithm (Paige and Saunders, 1982) to iteratively solve for all event-station pairs to obtain slowness, anisotropy, and station and event delays. The method includes damping parameters on both velocity and anisotropy to regularize the solution and reduce noise artifacts. P-wave traveltime residuals (<8 s) from sources at 2° to 14° are inverted for uppermost mantle velocity. A straight-line fit for the initial traveltime residuals versus distance gives an apparent Pn velocity of 8.1 km/s for the study area.
To map the Pn velocity structure in the central and eastern United States, we collected ~10,000 Pn traveltimes from International Seismological Centre (ISC), NEIC, and 750 handpicked arrivals from both permanent and temporary stations throughout the central and eastern United States. To compensate for the relatively small numbers of ray paths, we used a relatively large cell size in our model parameterization (0.5° × 0.5°). Overall, we have a relatively high density of ray paths within the active seismic zones in the central and eastern United States and lower ray coverage to the west of the Great Lakes and along the southern coastline of the United States (Fig. 15A).

We found a first-order agreement between the NA00 model (Goes and van der Lee, 2002) and our Pn tomographic velocity model (Fig. 15B). However, we also observed interesting small-
scale heterogeneity, such as the high velocities (~8.25 km/s) beneath both the New Madrid Seismic Zone and the Eastern Tennessee Seismic Zone. A low velocity zone (~8.0 km/s) is found in the western Ohio as well. The lithospheric mantle velocities within the North American shield are consistent with the high S-wave velocities measured at 100 km depth. Our results also show relatively low velocities (~8.05 km/s) in the northern and southern Appalachians.

The primary difference between our P-wave velocity measurements and the surface wave velocities (Goes and van der Lee, 2002) are in the Eastern Tennessee seismic zone, where we found a region of relatively high velocity that is not apparent in the NA00 model. A viscosity contrast and hence a change of lithospheric mantle properties here may concentrate stress and thus help to explain the Eastern Tennessee seismic zone seismicity.

DISCUSSION

One major result from this study is that the strain energy inherited from large intraplate earthquakes may dominate the local strain energy budget for hundreds to thousands of years. This result is expected, given the generally low strain rates in stable continents, including the North American plate interior (Dixon et al., 1996; Gan and Prescott, 2001). Applied to the New Madrid seismic zone, we have shown that the predicted spatial pattern and values of the stress and strain energy buildup following the 1811–1812 large events may explain the occurrence of many moderate-sized earthquakes in areas surrounding the New Madrid seismic zone since 1812. Some of these events may be viewed as aftershocks, as slow loading usually causes a long duration of aftershocks (Stein and Newman, 2004). Many of these events occurred outside the New Madrid seismic zone and were triggered or even directly produced (in terms of energy source) by the main events. Furthermore, we have shown that, after large earthquakes, intraplate seismic zones tend to stay in a stress shadow where full stress restoration may take thousands of years, longer than predictions based solely on regional strain rate estimates. This is because seismic zones within a stable continent are of finite length and are surrounded by relatively strong crust. As long as deviatoric stresses are supported by the ambient crust, little stress is available to reload the fault zones. This result is consistent with geodetic measurements in the New Madrid seismic zone and surrounding regions, which show that the current strain rates are very slow (0 ± 2 mm/yr) (Newman et al., 1999; Gan and Prescott, 2001), rather than 5–8 mm/yr reported earlier (Liu et al., 1992). More recent GPS data have confirmed the low strain rate around the New Madrid seismic zone (Smalley et al., 2005a); whether or not higher strain rates within the fault zone can be detected from present GPS data is debatable (Calais et al., 2005; Smalley et al., 2005b).

These results do not contradict the present rate of seismicity in the New Madrid seismic zone. Although thousands of events have been recorded in the New Madrid seismic zone in recent
decades, most are small (M < 4) and thus release little energy. No major (M > 5) events have occurred within the New Madrid fault zone since 1812, and the two largest events in the past two centuries, the 1895 Charleston, Missouri, earthquake (M = 5.9) and the 1843 Marked Tree, Arkansas, earthquake (M = 6.0), occurred near the tip of the inferred New Madrid fault zones, which is consistent with the model prediction.

However, the model results are inconsistent with paleoseismic data indicating that at least two more events similar to the 1811–1812 large events occurred in the New Madrid seismic zone, around A.D. 900 and 1400 (Kelson et al., 1996; Tuttle et al., 2002). Given the difficulties in determining the size and location of paleoearthquakes from liquefaction data, it is not surprising that some conclusions drawn from paleoseismic data are questionable. Newman et al. (1999), for instance, argued that the size of these paleoevents may have been overestimated: these may have been M ~ 7, rather than M ~ 8, events, which would be more in line with the new estimates for the 1811–1812 events (Hough et al., 2000). However, our model shows that repetition of even M ~ 7 events every few hundred years would be difficult in the New Madrid fault zone. Thus, explanations of paleoseismic data require local loading. Various local loading mechanisms have been proposed, including sinking of a “mafi c pillow” within the Reelfoot rift (Grana and Richardson, 1996; Pollitz et al., 2001a). We have avoided including these models in our calculations because of the large uncertainties of these models. Because seismic activity in the New Madrid seismic zone likely started in the Holocene (Pratt, 1994; Schweig and Ellis, 1994; Van Arsdale, 2000), any local loading mechanism must also explain why it started in the Holocene. Stress triggering associated with glacial isostatic adjustment (GIA) provides some interesting possible causes. James and Bent (1994) and Wu and Johnston (2000) concluded that GIA may be significant for seismicity in the St. Lawrence valley but not for the more distant New Madrid seismic zone, because GIA predicts predominately thrust faulting in the New Madrid seismic zone, not strike-slip faulting as expected, and the stress change would be too small (0.01 MPa). Hough et al. (2000) suggested that the main mechanism in the New Madrid seismic zone is actually reverse faulting. Grollimund and Zoback (2001) showed that GIA could have caused three orders of seis-

Figure 15 (on this and following page). (A) Ray coverage for Pn paths between 2¢ and 14¢ distance in the central and eastern United States. Approximately 10,000 raypaths are used in the model.
mic strain rate increase in the vicinity of the New Madrid seismic zone if there were a weak zone there. Refined imaging of crustal and lithospheric structures under the New Madrid seismic zone and other seismic zones in the central and eastern United States would help to test potential local loading mechanisms.

The stress field in the central and eastern United States is characterized by a nearly horizontal, NE- to E-striking axis of maximum compressive stress (Sbar and Sykes, 1973; Herrmann, 1979; Zoback and Zoback, 1989). The uniformity of stress-tensor orientation over a broad area of the central and eastern United States suggests that the stress field arises from forces that drive or resist plate motions (Richardson and Solomon, 1979; Zoback and Zoback, 1989). Given the rather uniform far-field stresses and the stability of the plate interior, crustal weakness, often found in ancient rift zones, is commonly related to intraplate earthquakes (Johnston and Kanter, 1990; Johnston and Schweig, 1996). This seems true in the central United States, especially in the Mississippi Embayment (Fig. 12), but not in the eastern United States, where seismic zones seem to be spatially associated with ancient faults that developed when the eastern United States was near plate boundaries (Dewey et al., 1989), or faults that may be related to transform fracture zones in the Atlantic Ocean floor (Sykes, 1978). Whereas these seismic zones may be associated with different structural causes, we suggest that there may be a common and deep cause for most of the seismicity in the central and eastern United States: the transition zone between the thick North American tectosphere and the surrounding lithosphere. Our calculations show that such lateral heterogeneity of lithospheric structure could concentrate stress near the margins of the tectosphere, and the predicted regions of high stresses have a strong correlation with seismicity in the central and eastern United States. However, our regional model does not include lateral heterogeneity of crustal structure, which would further affect stress distribution and seismicity. Further testing of the causal relationship between lithospheric structures and seismicity must await more detailed data about crustal and lithospheric structures of the central and eastern United States.

Figure 15 (continued). (B) Preliminary Pn tomographic map for the central and eastern United States. The minimum hit count for plotting is 3. Black circles are the same earthquake epicenters shown in Figure 9.
CONCLUSIONS

1. Intraplate seismic zones tend to remain in a Coulomb stress shadow for thousands of years following large earthquakes. The slow far-field tectonic loading rates and the relatively strong ambient crust make stress reaccumulation within intraplate fault zones difficult, unless there are some local loading mechanisms. On the other hand, a significant amount of the stress relieved from large intraplate earthquakes, and the associated strain energy, may migrate to and be trapped within the neighboring crust, mainly near the tips of the fault zones. Such inherited strain energy may dominate the strain energy budget in the intraplate fault zone and surrounding regions for hundreds to thousands of years, and it can produce aftershocks hundreds of years after the main shocks. These behaviors are fundamentally different from interplate seismic zones, which are constantly loaded by plate motions.

2. The 1811–1812 large earthquakes in the New Madrid seismic zone caused significant build up of Coulomb stress and strain energy in the surrounding regions, mainly southern Illinois and eastern Arkansas. Many of the moderate-sized earthquakes (M > 5) in these regions since 1812 may have been triggered or facilitated by stress and strain energy inherited from the 1811–1812 large events. The residual strain energy from the 1811–1812 main shocks is capable of producing some damaging (M > 6) earthquakes in areas surrounding the New Madrid seismic zone today, even in the absence of local loading. Conversely, the New Madrid fault zones should remain in a stress shadow, and thousands of years may be needed to restore the stress to the pre-1811–1812 level. Thus, some local loading mechanism must be active if numerous large events similar to the 1811–1812 events have occurred in the fault zones during the Holocene, as suggested by paleoseismic data. Although a number of local loading mechanisms have been proposed, more studies, including refined imaging of the crustal and lithospheric structures in the New Madrid seismic zone region, are needed to test these hypotheses.

3. Seismicity in the central and eastern United States shows a strong spatial correlation with the margins of the North American tectosphere, consistent with our model prediction of high Coulomb stress in the tectosphere-lithosphere transition zones. In the New Madrid seismic zone, the seismicity seems to be related to an abnormally thin lithosphere under the Mississippi Embayment. Again, further imaging of the crustal and lithospheric structure will help to address the cause of seismicity in the New Madrid seismic zone and other seismic zones in the central and eastern United States.

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