Slow Deformation and Lower Seismic Hazard at the New Madrid Seismic Zone

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Global Positioning System (GPS) measurements across the New Madrid seismic zone (NMSZ) in the central United States show little, if any, motion. These data are consistent with platewide continuous GPS data away from the NMSZ, which show no motion within uncertainties. Both these data and the frequency-magnitude relation for seismicity imply that had the largest shocks in the series of earthquakes that occurred in 1811 and 1812 been magnitude 8, their recurrence interval should well exceed 2500 years, longer than has been assumed. Alternatively, the largest 1811 and 1812 earthquakes and those in the paleoseismic record may have been much smaller than typically assumed. Hence, the hazard posed by great earthquakes in the NMSZ appears to be overestimated.

During the years 1811 and 1812 a long sequence of earthquakes (*1*), including three large shocks thought to have had surface wave magnitudes near or exceeding 8 (*2*), was centered near New Madrid, Missouri. The region remains active, with small to moderate (magnitude less than 5) earthquakes concentrated along planar segments (Fig. 1) presumed to reflect subsurface faults (*3*). The few available focal mechanisms suggest that right-lateral strike-slip occurs on two NE-SW–striking, subvertical faults, whereas thrusting occurs on a shallowly SW-dipping fault that forms part of the left step between the strike-slip faults (*4*). The seismicity may reflect reactivation of Paleozoic faults (*5*) by platewide or local stresses (*6*).

We use local and platewide GPS data to assess the deformation and draw inferences about the recurrence of large earthquakes. In the local study, we monitor a network of geodetic monuments including sites near the seismicity, in unconsolidated alluvium and semiconsolidated bedrock of the Mississippi embayment, and in bedrock outside the embayment (*7*). The network was surveyed in November 1991, October 1993, and October 1997 (*8*). Horizontal velocities obtained from least-squares fits to the site positions (*9*) are shown in Fig. 1, after removal of the motion of the North America plate (*10*). Almost all residual velocities are small and within the 2σ error ellipses estimated from the repeatability of daily site positions, scaled to include the effects of time-correlated errors (*11*).

We examined the data using a geometry in which the motion expected between large earthquakes should be evident, assuming strain accumulation on a locked right-lateral strike-slip fault. Site motions were projected along N42°E, the approximate strike of the strike-slip faults. We then removed the mean and compared the data with the predicted velocity profile for a locked vertical strike-slip fault, a steep gradient near the fault that asymptotically approaches the far-field interseismic rate away from the fault (Fig. 2) (*12*). The assumed locking depth affects only the shape of the steep central part of the profile. We used a locking depth of 25 km and considered rates between 6 and −6 mm/year (negative rates denote left-lateral motion) to find misfit as a function of rate. To estimate uncertainties in the best fitting rate, we rescaled the misfit to have reduced $\chi^2 = 1$ at the minimum and used a χ^2 test to find 95% (2σ) confidence limits.

Using all sites yields a best fitting rate of -0.2 ± 2.4 mm/year. Dividing these into near-field (within the embayment) and intermediate-field (primarily hard rock) sites yields 0.6 ± 3.2 and -0.9 ± 2.2 mm/year, respectively. None of these values, including the slightly higher one in the near-field where noise might be highest, differs significantly from zero. Because the rates are low, even their small uncertainties permit a range of interpretations (they are consistent with both 0 and 2 mm/year at 2σ). Although velocity uncertainties will be reduced by longer observation periods, the present data show little of the steady fault-parallel far-field motion which would be expected before future strike-slip earthquakes (*13*). However, the slightly higher near-field values may reflect a small tectonic signal, which could be more complex due to the details of the fault geometry, including the thrust segment (*7*).

Similar results emerge from analysis of continuously recording GPS sites away from the New Madrid seismic zone (NMSZ), in the presumably stable portion of North America (Fig. 3) (*14*). Modeling stable North America as a single rigid plate fits the site velocities well, with a mean residual of 1.0 mm/year. The misfits are not significantly reduced by assuming separate blocks east and west of the NMSZ (Fig. 3, inset), and the predicted motion across the NMSZ is indistinguishable from zero.

These results have implications for earthquake recurrence. It has been assumed that

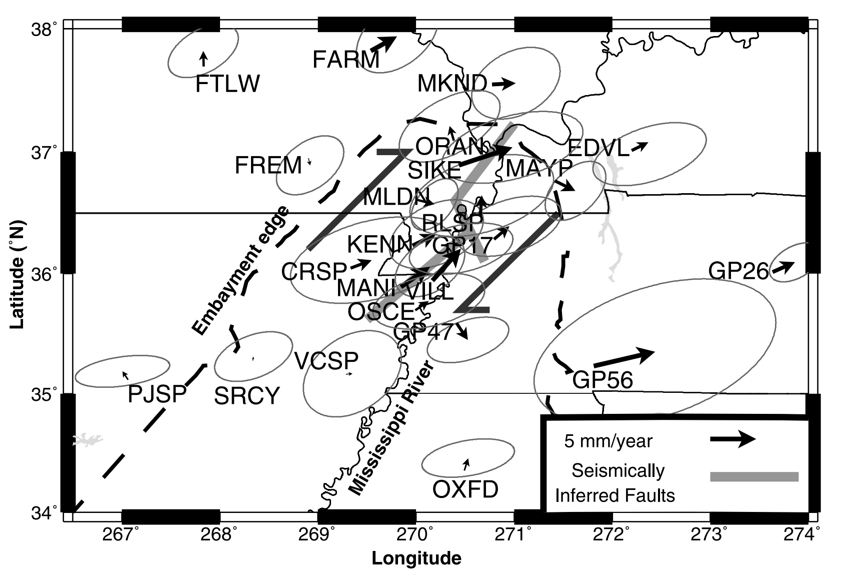

Fig. 1. Residual horizontal site velocities (1997 to 1991) for the New Madrid GPS network, after removal of the motion of the North American plate. Velocities are small, within 2σ error ellipses.

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the 1811–1812 earthquakes were magnitude 8 events, with 5 to 10 m of horizontal slip (*2*). On the basis of that assumption, earlier geodetic results that found surprisingly rapid near-field strain accumulation, about one-third to two-thirds of that for the San Andreas fault system, were interpreted as consistent with 500- to 800-year recurrence for such great earthquakes (*15*). Our lower estimate of far-field interseismic motion, less than 2 mm/year, implies a recurrence period for such earthquakes well exceeding 2500 years (Fig. 4A). This period is a minimum because it is calculated for the maximum rate (at 2σ) rather than the best fitting near-zero value, which predicts a much longer period, and is based on the assumption that all interseismic motion will be released seismically. Because deformation rates here are less than a few millimeters per year, compared with tens of millimeters per year on plate boundaries, the recurrence times for similar large earthquakes must be correspondingly longer.

A similar conclusion emerges from estimating the recurrence of future larger earthquakes from the observed rate of smaller earthquakes (Fig. 4B) (*16–18*). From recent (1974 to 1998) seismicity, when the earthquake catalog should be most complete and the magnitudes have been seismologically determined, we expect magnitude 7 and 8 earthquakes on average every 1700 and 15,000 years, respectively. Similar values (1000 and 13,000 years) are predicted from consideration of the post-1816 seismicity. These estimates are consistent with our GPS results but exceed those of an earlier study (*19*) because of different assumptions about the magnitudes of the larger earthquakes.

Significantly shorter recurrence intervals of 400 to 600 years have been estimated from the geological record (*20*). However, the paleoseismology can be reconciled with the geodesy and frequency-magnitude relation. One possibility is that the 1811–1812 and earlier large earthquakes in the paleoseismic record were significantly smaller than previously assumed, perhaps magnitude 7 with slip of about 1 to 2 m. If so, 1 to 2 mm/year of interseismic motion

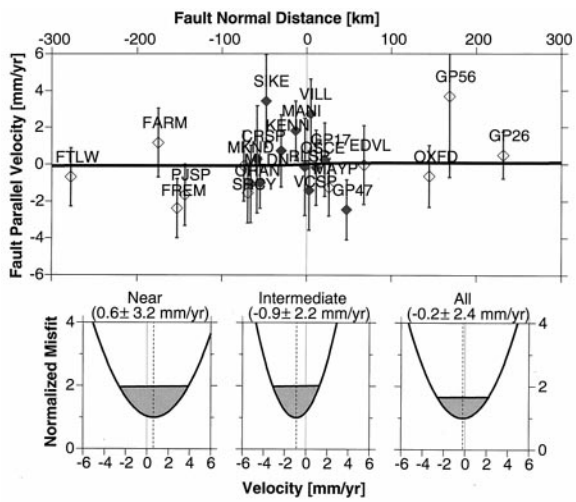

Fig. 2. (**Top**) Profile of site velocities with 1σ error bars, parallel to the approximate strike of the major strike-slip faults in the NMSZ. Solid and open symbols denote near- and intermediate-field sites. Mean velocity is removed. Also shown is a best-fitting model profile for a locked vertical strike-slip fault driven by interseismic motion. (**Bottom**) Misfit as a function of interseismic rate is shown with 2σ ranges shaded. None of the best-fitting rates differs significantly from zero.

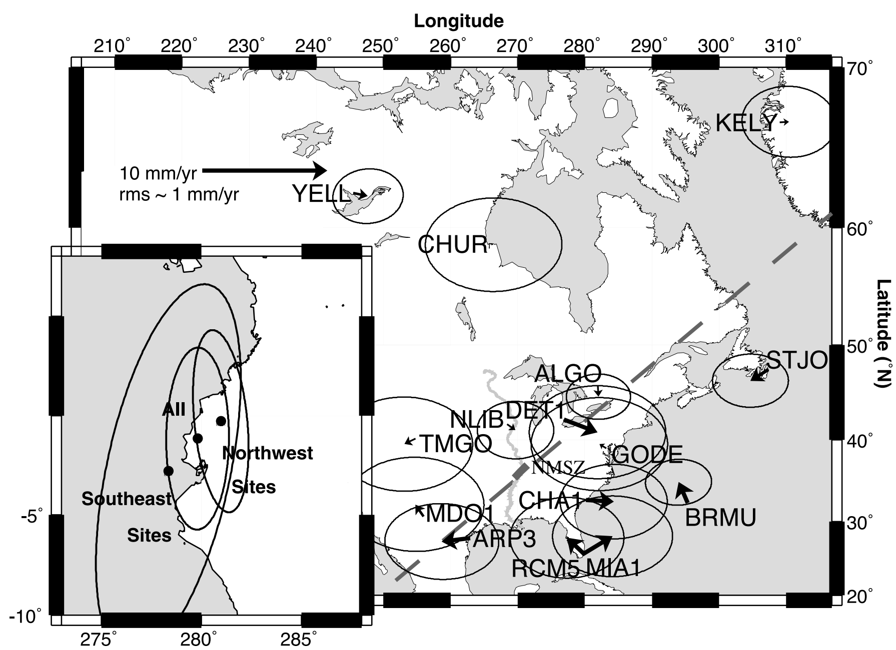

Fig. 3. Locations of continuously recording GPS sites used to estimate an Euler vector for the presumably stable portion of North America. For each, the misfit between the observed velocity and that predicted for a single plate is shown. (**Inset**) Euler poles for the eastern and western subsets of the sites (divided by dashed line in main panel) compared with that for the entire set. Because the poles for the east and west data overlap at 95% confidence, the platewide GPS data show no resolvable motion across the NMSZ.

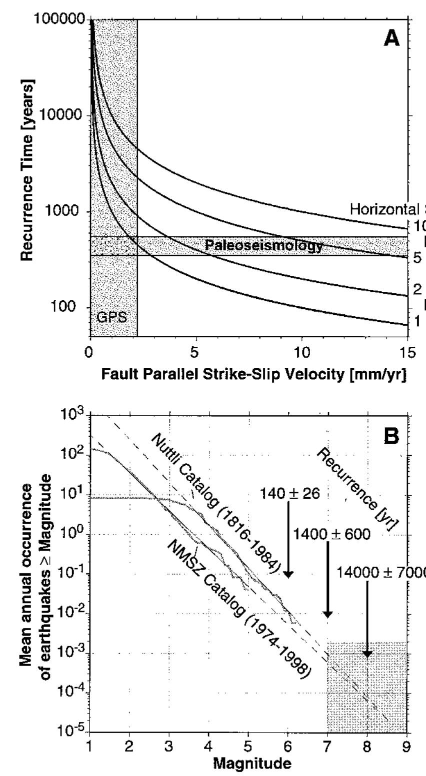

Fig. 4. (**A**) Relation between interseismic motion and the expected recurrence of large New Madrid earthquakes. For an assumed horizontal slip in 1811–1812 of 5 to 10 m, the geodetically observed interseismic motion of less than 2 mm/year implies recurrence times greater than 2500 years. Also shown are recurrence estimates from paleoseismic studies. The paleoseismic and geodetic data are jointly consistent with slip in 1811–1812 being about 1 m, corresponding to a magnitude 7 earthquake. (**B**) Earthquake frequency-magnitude data for the NMSZ. Both the recent and historic (1816 to 1984) data have slopes close to 1 and predict a recurrence interval exceeding 1000 years for magnitude 7 earthquakes and 10,000 years for magnitude 8 earthquakes. Estimates are shown with 2σ uncertainties.

would correspond to a 500- to 2000-year recurrence (Fig. 4A), consistent with the recurrence for earthquakes of this size from the frequency-magnitude relation. Although this magnitude is smaller than what was inferred from the felt area of the 1811–1812 earthquakes and the spatial extent of the paleoseismic deformation, both techniques have considerable uncertainties in estimating earthquake magnitude. Hence, we feel that smaller 1811–1812 earthquakes can reconcile the different techniques, given uncertainties.

It is also possible that 1811–1812–style earthquakes may never recur. If more accurate future surveys continue to find essentially no interseismic slip, we may be near the end of a seismic sequence. It has been suggested that because topography in the New Madrid region is quite subdued, the NMSZ is a feature no older than a few million years and perhaps as young as several thousand years (*21*). Therefore, New Madrid seismicity might be a transient feature, the present locus of intraplate strain release that migrates with time between fossil weak zones.

Although much remains to be learned about this intriguing example of intraplate tectonics, the present GPS data imply that 1811–1812–size earthquakes are either much smaller or far less frequent than previously assumed. In either case, it seems that the hazard from great earthquakes in the New Madrid zone has been significantly overestimated. Hence, predicted ground motions used in building design there, such as the National Seismic Hazard Maps (*22*) that presently show the seismic hazard there exceeding that in California, should be reduced.

References and Notes

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2. A. Johnston, *Geophys. J. Int.* **126**, 314 (1996);
3. L. Himes, W. Stauder, R. Herrmann, *Seism. Res. Lett.* **59**, 123 (1988); J. Chiu, A. Johnston, Y. Yang, *ibid.* **63**, 375 (1992).
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6. M. L. Zoback and M. D. Zoback, *Science* **213**, 96 (1981); J. Granga and R. Richardson, *J. Geophys. Res.* **101**, 5445 (1996).
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8. The sites occupied are listed in (*7*), except CRVL, which was buried by the construction of a baseball field. The 1997 survey was similar to that in 1993 except for shorter (2 day) site occupations.
9. GPS data were analyzed at the University of Miami following the procedures of T. Dixon *et al.* [*J. Geophys. Res.* **102**, 12017 (1997)]. We used high-precision nonfiducial satellite orbits and the Jet Propulsion Laboratory GIPSY analysis software to estimate site velocities in the ITRF-96 reference frame. Site velocities were estimated from least-squares fits to daily positions, weighted by errors following the method of Mao *et al.* (*11*). The 1991 data have higher root mean square (rms) scatters compared with 1993 and later data. Scaling the 1991 position errors with the correlations between weighted rms and white and colored noise (*11*) de-weights them compared with later data.
10. We subtracted motion predicted by an Euler vector in ITRF-96 for stable North America (1.16°S, 80.2°W, 0.193° per million years) determined by inversion of GPS data from 16 continuous stations (*11*). This subtraction removes most network-wide motion, although a small (but not statistically significant) eastward drift remains. Whether this drift is tectonically significant is unclear; it has no effect on our velocity gradient analysis (Fig. 2) because a mean velocity is also removed.
11. Until recently, most GPS studies assumed that geodetic noise is uncorrelated in time. Recent studies show that GPS noise is time-correlated and that assuming noise is uncorrelated can underestimate velocity errors by up to an order of magnitude [A. Mao, C. Harrison, T. Dixon, *J. Geophys. Res.* **104**, 2797 (1999)].
12. J. Savage and R. Burford, *J. Geophys. Res.* **78**, 832 (1973).
13. We compared geodetic data to a model in which steady far-field motion loads a fault before earthquake rupture. This model is commonly used for plate boundaries, where space geodetic data show rates of plate motion consistent with those over millions of years, indicating that steady far-field motions give rise to episodic earthquakes [R. Gordon and S. Stein, *Science* **256**, 333 (1992); S. Stein, in *Space Geodesy and Geodynamics*, *Geodynamics Ser. 23*, D. Smith and D. Turcotte, Eds. (American Geophysical Union, Washington, DC, 1993), pp. 5–20]. Similarly, space geodetic data (for example, Fig. 3) show that plates thought to have been rigid on geological time scales are quite rigid on decadal scales. Hence, application of these ideas of steady motion to intraplate settings seems plausible but has not been demonstrated. We did not consider strain transients after the 1811–1812 earthquakes [P. Rydelek and F. Pollitz, *Geophys. Res. Lett.* **21**, 2303 (1994)], which predict motions larger than interseismic motion. Similarly, our geodetic approach implicitly focuses on motions due to platewide rather than locally derived stresses (*6*).
14. T. Dixon, A. Mao, S. Stein, *Geophys. Res. Lett.* **23**, 3035 (1996). These results are updated here with additional sites [A. Mao, thesis, University of Miami (1998)].
15. L. Liu, M. Zoback, and P. Segall [*Science* **257**, 1666 (1992)] used GPS to remeasure monuments previously measured by triangulation and reported rapid strain accumulation in the southern NMSZ corresponding to 5 to 7 mm/year of slip. A similar study across the northern NMSZ found strain rates indistinguishable from zero [R. Snay, J. Ni, H. Neugebauer, in *U.S. Geol. Surv. Prof. Pap. 1538-F* (1994), pp. F1–F6]. Our earlier study (*7*), based on the first two GPS occupations of presumably more stable monuments, found a far-field rate of 3 ± 3 mm/year (limits are from the approach used in Fig. 2), indistinguishable from zero at 2σ. Hence, improved geodetic techniques and longer measurements generally reveal successively slower motion, presumably because the far-field velocity is small (or zero). Because the data have uncertainties, the first two measurements typically overestimate the velocity, which successive measurements better approximate. Unless the uncertainties are well understood, the estimated velocity may appear unduly significant (*11*). For New Madrid, the older triangulation data were presumably less accurate than GPS because of limitations of the technique, possibly compounded by instability of shallow-rooted triangulation pillars. Moreover, for low–strain rate areas a few measurements can change triangulation results significantly [R. Snay, *J. Geophys. Res.* **91**, 12695 (1986)]. Our GPS surveys use deeper rooted and presumably more stable monuments and extend outside the embayment, but geodetic GPS technology was still immature in 1991. By 1993, improved GPS receivers and an improved network of global tracking sites yielded better data, as determined by better repeatability of site positions between successive days. GPS velocity analysis is also less sensitive to site-specific errors than the triangulation analysis. Therefore, we consider the GPS results here more accurate than earlier surveys.
16. Earthquake populations approximately follow log $N = a - bM$, where N is the number of earthquakes (total or annual) whose magnitude exceeds M. Because b is about 1, earthquakes of a given size are about one-tenth as numerous as those one magnitude unit smaller. Although body and surface wave magnitudes do not exceed about 6.4 and 8.4, respectively, earthquake catalogs typically show b to be about 1 because body wave magnitudes are reported for small earthquakes and surface wave magnitudes are reported for large earthquakes [E. Okal and B. Romanowicz, *Phys. Earth. Planet. Int.* **87**, 55 (1994)]. Hence, if the linear frequency-magnitude relation is used, magnitudes above 6.4 should be treated as surface wave magnitudes.
17. Extrapolation of the recurrence of larger earthquakes from the rate of smaller earthquakes, used because of the limited data available for earthquakes before this century, faces various uncertainties. M. Stirling, S. Wesnousky, and K. Shimazaki [*Geophys. J. Int.* **124**, 883 (1996)] found that such extrapolation overestimates recurrence times inferred from geological data, whereas E. Triep and L. Sykes [*J. Geophys. Res.* **102**, 9923 (1997)] found that this extrapolation underpredicts recurrence times for large intracontinental earthquakes.
18. For 1974 to 1998, the New Madrid catalog (http://elwe.ceri.memphis.edu/~seisadm) of earthquakes with seismologically determined magnitudes yields a and b values of 3.446 ± 0.041 (1σ) and 0.954 ± 0.013. For 1816 to 1984 (beginning at 1816 excludes major aftershocks), Nuttli's catalog (http://www.eas.slu.edu/Earthquake_Center), in which body wave magnitudes before 1964 are typically inferred from the reported shaking, yields a and b values of 4.537 ± 0.105 and 1.079 ± 0.022. Combining the two lines predicts recurrence times (with 2σ uncertainties) for magnitude 7 and 8 earthquakes of about 1400 ± 600 and 14,000 ± 7000 years. These values seem plausible: since 1816, there have been 16 earthquakes with magnitude greater than 5 (about a 10-year recurrence), and two with magnitude greater than 6 (about a 100-year recurrence), so magnitude 7 and 8 earthquakes should have about 1000- and 10,000-year recurrences, respectively. These estimates do not depend on whether the seismogenic stresses are local or platewide in origin (*5*, *13*).
19. A. Johnston and S. Nava [*J. Geophys. Res.* **90**, 6737 (1985)] inferred 550- to 1100-year recurrence for earthquakes with surface wave magnitude (M_s) > 8.3. This frequent recurrence results from treating magnitude 7 earthquakes as body wave magnitude 7, and equating them to surface wave magnitude 8.3. Recent results (*16*) indicate that such earthquakes are better treated as surface wave magnitude 7.
20. S. Wesnousky and L. Leffler, *Bull. Seismol. Soc. Am.* **82**, 1756 (1992); M. Tuttle and E. Schweig, *Geology* **23**, 253 (1995); K. Kelson *et al.*, *J. Geophys. Res.* **101**, 6151 (1996).
21. E. Schweig and M. Ellis, *Science* **264**, 1308 (1994).
22. A. Frankel *et al.*, *National Seismic Hazard Maps Documentation: U.S. Geol. Surv. Open-File Rep. 96-532* (1996). These maps assume that a magnitude 8 earthquake occurs every 1000 years at New Madrid, so the predicted peak ground acceleration expected in 50 years at 2% probability for the NMSZ exceeds that in San Francisco, and the predicted highest acceleration (exceeding 1.2*g*) area for the NMSZ is larger than for Los Angeles or San Francisco.

23. Supported by NASA grants NAGW-2522 and NAG5-6685. We thank R. Graves, K. Wellington, C. Bunker, K. Nolan, and C. Griffen for field assistance in the third survey; F. Farina for assistance with data processing; E. Okal and K. Mueller for helpful discussions; R. Snay and an anonymous reviewer for useful reviews; and University Navstar Consortium for technical support.

21 December 1998; accepted 1 March 1999