New Madrid GPS: Much Ado About Nothing?

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Two articles in this issue address the surprising and intriguing negative result that high-precision geodetic measurements find no compelling evidence for crustal motions in the New Madrid Seismic Zone of the central Unites States.

The geodetic measurements have drawn great interest because they add another puzzle to the many surrounding New Madrid. This area is the best known case of large earthquakes within the interior of the plates, in continental lithosphere. Such earthquakes are much rarer and release much less seismic energy than those at plate boundaries. Because idealized plates are perfectly rigid, these earthquakes demonstrate that deformation occurs within plates and provide a lower bound on their rigidity. Moreover, precisely because such earthquakes are rare, they pose a hazard to areas that are less well prepared for earthquakes than areas with more active ones. Assessing the hazard is further complicated by a growing sense that earthquakes within plates migrate among seismic zones that 'turn on,' remain active for some time, and then 'turn off.'

The potential hazard is illustrated by the large (magnitude 7) earthquakes that occurred in 1811 and 1812, causing shaking across much of the area. Houses collapsed in the tiny Mississippi River town of New Madrid, Mo., and minor damage occurred in St. Louis, Mo., Louisville, Ky., and Nashville, Tenn. The smaller earthquakes that continue today, which may be aftershocks of the 1811–1812 events, are more of a nuisance than a catastrophe. For example, the largest earthquake in the past century, the 1968 (magnitude 5.5) southern Illinois earthquake, was widely felt and caused damage but no fatalities. However, large earthquakes like those of 1811–1812 would be much more destructive. Paleoseismic data suggest that these have occurred about 500 years apart in the past 1000 years and hence may recur.

Surprisingly little is known about these earthquakes. It is not clear why they occur, when they started, when, if ever, they will recur, and how large a hazard they pose. As a result, researchers looked to the new tool of GPS geodesy for new insights and were surprised by the results [*Newman et al.*, 1999].

A GPS measurement yields a site's position to a precision of millimeters, so a series of measurements over time gives its velocity. This is typically plotted as a velocity vector from the site's position, with an error ellipse about the vector's head showing the uncertainty in velocity. Ideally, the ellipse is a small region about the vector's head, showing that the velocity is well constrained. This is far from the case for sites in the New Madrid zone (Figure 1, top). The site velocities shown, which are motions with respect to the rigid North American plate, are small—less than 2 millimeters per year—and generally within their error ellipses. Hence most sites show no motion significantly different from zero. In other words, the GPS data do not require that they be moving at all, and restrict any motion to being very slow. Moreover, the vectors do not show the spatially coherent pattern typically seen in deforming seismic zones.

The results are gratifying from the view of plate tectonics, in that they and sites elsewhere in eastern North America show that the plate is quite rigid, with the major deviation being vertical motion due to postglacial adjustment. Beyond this motion, there is no clear case for tectonic effects, in that the small motions could be a combination of observational and analysis errors, and small motions of the geodetic monuments. However, much faster motion had been expected because of the earthquakes. It had been suggested that the earthquakes of 1811–1812 were magnitude 8 events and occurred about every 500 years. If so, more than 5 millimeters per year of average motion during the interval between earthquakes would be needed to store up the slip for a future large earthquake (Figure 1, middle). Hence, the first inference from the slow motions was that typical large earthquakes in the area are smaller, magnitude 7, in accord with recent analysis of historic records of the intensity of shaking

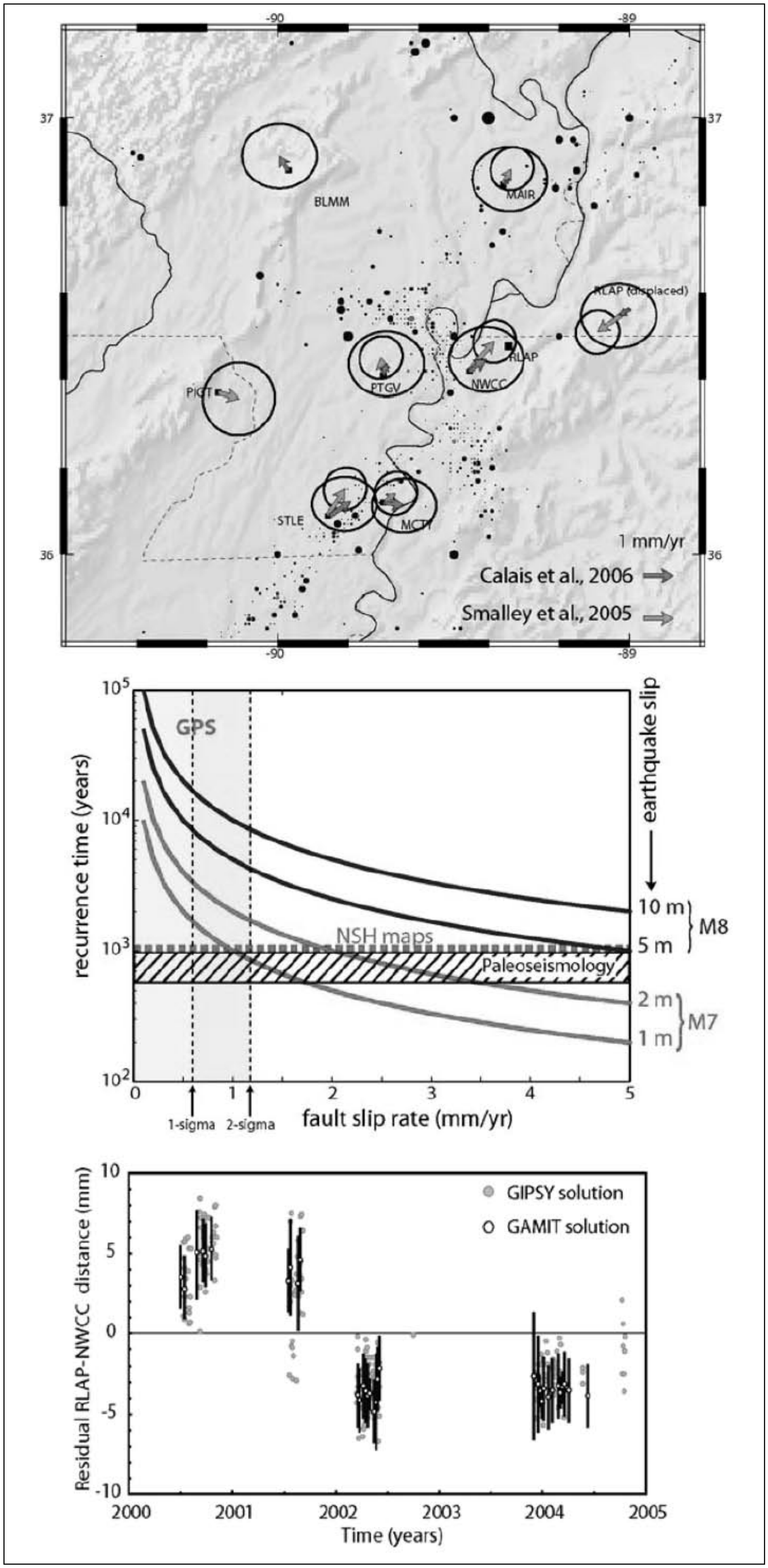

Fig. 1. GPS data and interpretations for the New Madrid Seismic Zone, after Newman et al. *[1999] and* Calais et al. *[2005, 2006]. (top) GPS site velocities and associated uncertainties (95% confidence ellipses) from two different analyses. Larger ellipses are from* Smalley et al. *[2005]. Site velocities within error ellipses show no statistically significant motion. RLAP motion is plotted displaced for clarity. Small circles show regional seismicity. (middle) Slip expected in a future earthquake assuming the geodetically observed fault slip rate accumulates over a given recurrence time. Curves show range for magnitude 7 and 8 earthquakes. Horizontal lines show recurrence times assumed in National Seismic Hazard (NSH) maps and inferred from paleoseismic studies. Vertical lines show maximum slip rate from GPS data. (bottom) Time series of estimates of the distance between sites RLAP and NWCC after removal of a mean distance, derived using two different GPS data analysis software packages. Note offset between 2001 and 2002.*

[*Hough et al.,* 2000]. Beyond this, debate has continued, focusing on two broad questions that are explored by the articles here.

The first question is whether the GPS data show any motion. The motions are so slow that minor differences in the length of data used, the processing method, or assumptions in the error analysis can lead to different interpretations (Figure 1, top). *Smalley et al.* [2005] conclude that significant motion occurs at two sites, RLAP and NWCC on opposite sides of the scarp thought to have been part of the fault break in 1811–1812. In contrast, *Calais et al.* [2005, 2006] find that none of the sites shows significant motion and that the inferred motion between RLAP and NWCC is due to a puzzling offset in the time series, suggestive of problems in the data or analysis rather than tectonic motion (Figure 1, bottom). *Newman* [this issue] explores a related issue, showing that reporting small motions as strains—differences between small motions at two sites divided by the distance between them—can be misleading.

The second question is what the small or nil motions imply for past and future earthquakes. Analyses to date have explored the implications of the slow motions for the time and magnitude of future large earthquakes. Another possibility is that the motions are nil, perhaps implying that the seismic zone is shutting down and will not generate future large earthquakes. Alternatively, *Rydelek* [this issue] shows that the motions may be transient effects from the 1811–1812 earthquakes and thus give no direct information about future earthquakes.

The two articles here illustrate the complexity of the issues, which will likely be debated for many years. GPS velocity issues will eventually be resolved because the precision of velocity estimates increases with time. Hence the estimated motion will either continue shrinking closer to zero or climb above the uncertainties to show significant motion. However, the tectonic issues and their implications for seismic hazard policy may take much longer—hundred of years or more—to resolve. This situation would have delighted Mark Twain, who piloted steamboats through the area only 50 years after the 1811–1812 earthquakes. In his words, "There is something fascinating about science. One gets such wholesale returns of conjecture out of such a trifling investment of fact."

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Earthquake Risk From Strain Rates on Slipping Faults

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Discrete geodetic measurements made near active faults may capture only small bits of a relatively complex field of deformation surrounding a fault, making it difficult to accurately describe the nature of ongoing activity along the fault. This difficulty is compounded when geodetic measurements are reported as strain rates, which involve differences in the displacement between two or more sites over time. As a result, very low displacement rates can be quoted as very high strain rates, which may lead to incorrectly inferring high seismic risk. As an example, I look at a recent deformation study across the New Madrid Seismic Zone (NMSZ). The NMSZ, located in east central United States and away from rapidly deforming plate boundaries, is best known for its series of three large earthquakes ($M > 7$) in the early 1800s, and continues today with numerous small earthquakes (Figure 1).

A recent study by *Smalley et al.* [2005] used two GPS sites within a larger continuous network to identify rapid strain accumulation across the central thrust segment of the NMSZ. Using the limited subset of data, *Smalley et al.* [2005] inferred that the strain rate measured at the NMSZ is comparable to rates along the San Andreas and other active faults, in apparent contrast to an earlier campaign GPS study that showed less than 2 millimeter per year of overall deformation [*Newman et al.*, 1999]. It has been argued that the new results do not require motion statistically different from zero, reflecting differences in processing techniques and random noise [*Calais et al.*, 2005]. However, here I consider a separate larger and more general issue. This is the validity of using simple strain measurements across a fault for implying earthquake recurrence and seismic hazards. The difficulty is that the small displacement rates can be quoted as high strain rates.

Of the eight sites in a network surrounding the NMSZ, *Smalley et al.* [2005] found that only two sites, which were 11 kilometers apart, showed potentially significant deformation of 2.7 ± 1.6 millimeters per year (Figure 1). Using the simple linear relation for strain rate,

$$\dot{\varepsilon} = \frac{\Delta \dot{u}}{l}$$

where $\Delta \dot{u}$ is the change in velocity over the distance l between measurements, they find a strain rate of approximately 10^{-7} per year. Unfortunately, this resultant strain rate alone does not yield useful information about the true strain accumulation across a slipping fault. That is because for a given

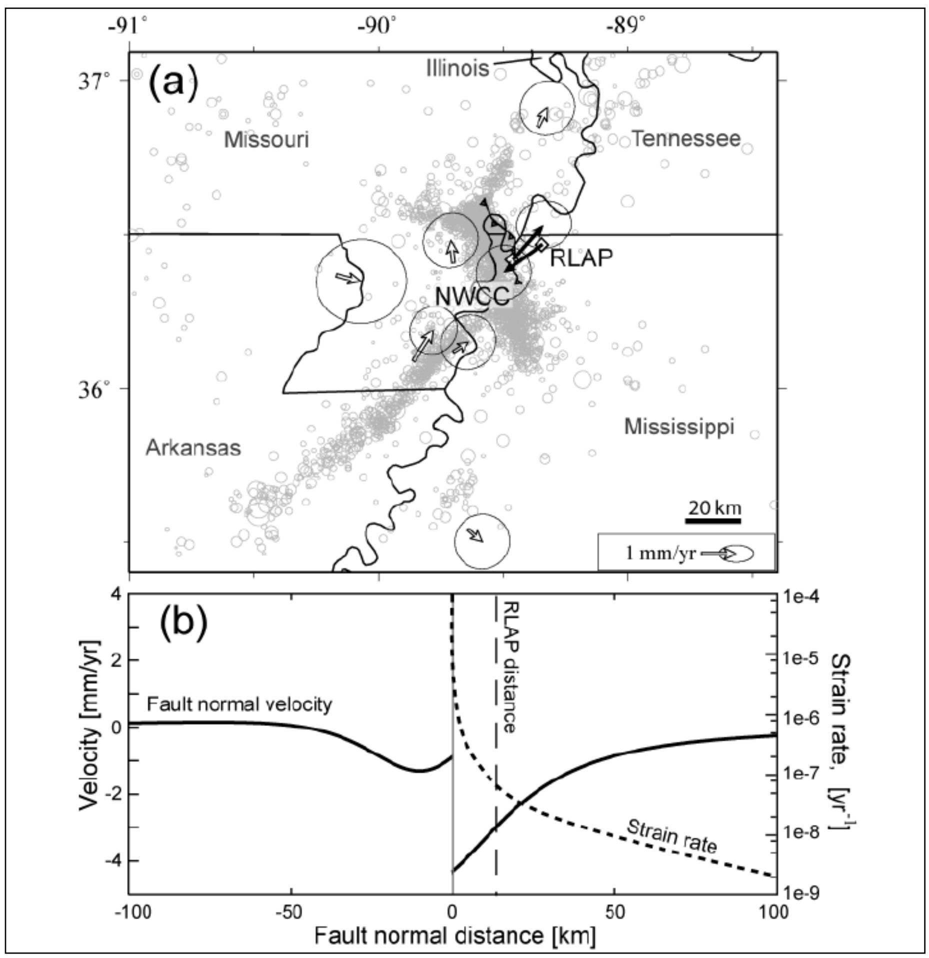

*Fig. 1. (a) New Madrid seismicity since 1974 (grey circles), GPS horizontal site velocities and 2σ errors (arrows and ellipses [*Smalley et al., *2005]), and approximate surface location of the Reelfoot thrust fault (thick toothed line). Solid arrows are site velocities nearest the thrust and were used to infer strain rates of 10^{-7} per year at the distance of site RLAP. (b) Predicted fault-normal displacements and strain rates (thick dark solid and dashed lines) for ongoing slip across a simple thrust assuming a 60-kilometer-long, 70° west dipping fault extending to 20 kilometer depth (approximating the Reelfoot thrust) using the analytic model of* Mansinha and Smylie *[1971]. Slip is scaled to approximate that reported by* Smalley et al. *[2005], 11 kilometers from the fault (thin vertical dashed line). The resulting strain rate is not constant, but instead increases rapidly near the fault.*

fault slip rate, the strain rate is directly dependent on the distance between the two sites across the fault. In the NMSZ, ongoing earthquake activity along the Reelfoot fault (Figure 1) suggests that this is the case near the two sites used in the Smalley paper.

To better illustrate the issue, the model curves in Figure 1 show fault-normal shortening and the resulting strain rate for a simplified fault analogous to the Reelfoot thrust. Slip is scaled to approximate the 10^{-7} per year strain rate at two sites about 5 kilometers from the fault, inferred by *Smalley et al.* [2005]. However, depending on changes in the measurement distance, strain rates decrease dramatically away from and increase rapidly very near the fault. Specifically, when measurements are made 100 kilometers from the fault, the resultant strain rate decreases by 2 orders of magnitude. However, as measurements are made right up to the fault, strain rates become infinitely large, as the distance between measurements goes to zero. Thus, it is clear that a direct comparison of strain rates alone from different fault systems is not useful for describing relative activity along faults, let alone their seismic hazard. It is better to either directly compare relative velocities at certain distances or compare the best models that describe such activity. In this case, the measured convergence suggests that there is active slip along the fault, consistent with ongoing microseismicity.

A significant consideration to be made here is whether or not ongoing active slip along a fault suggests increased seismic hazard, as has been suggested by *Smalley et al.* [2005]. It has generally been observed that along active faults, the regions that are not actively slipping are instead locked and thus able to build strain energy for possibly catastrophic release. These locked regions, identifiable by significant far-field and little to no near-field strain, have the greatest potential for moderate to large earthquakes [e.g., *Scholz,* 2002]. Thus, ongoing slip on a fault may suggest lower immediate danger because the fault is accumulating little if any strain energy for future events. In the NMSZ, if the relative motion is real, it may be due to postseismic relaxation within the asthenosphere due to past earthquakes [*Rydelek*, this issue], which causes transient motions near the fault rather than accumulating slip for future earthquakes.

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New Madrid Strain and Postseismic Transients

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A crucial issue for the assessment of earthquake hazard in the New Madrid Seismic Zone (NMSZ) of the central United States is whether the small motions inferred from geodetic measurements are actually the result of strain accumulation that will eventually be released in damaging earthquakes. The interpretation of these measurements has led to an ongoing debate over the associated seismic risk and hazard assessment in the NMSZ [*Zoback,* 1999; *Schweig et al.*, 1999; *Newman et al.*, 1999a, 1999b; *Stein et al.*, 2003]. The gist of the debate is whether or not models of high seismic hazard in this region are supported by the geodetic data and historic earthquake data.

A recent report by *Smalley et al.* [2005] on GPS measurements across the Reelfoot fault suggested a relatively high strain rate, of the order of 10^{-7} per year, comparable to that normally associated with convergence at plate boundaries. To some, these measurements seemingly ended the debate since they were taken to be the result of rapid strain accumulation that could unleash a large, devastating earthquake, which in turn prompted a general public warning from a top U.S. Federal Emergency Management Agency (FEMA) official [*Brown,* 2005]. Others, however, believe that the debate is not yet settled [*Calais et al.*, 2005]; it has been argued that these GPS measurements show no statistically significant motion and instead reveal a puzzling offset in one of the GPS time series [*Calais et al.*, 2006].

Smalley et al. [2005] offered several explanations for their observations, one of which was long-term postseismic relaxation following the 1811–1812 sequence of three large earthquakes that occurred in this seismic zone. Clearly, relaxation is fundamentally different from accumulation. Postseismic relaxation is due to the coupling of the rigid elastic crust to the underlying viscoelastic asthenosphere. An earthquake generates stresses that are relieved by both an immediate elastic response (coseismic effect) and a long-term viscoelastic relaxation (postseismic effect) that will persist for many decades because of the enormous value of the viscosity of the asthenosphere, on the order of 10^{20} pascal seconds. The long-term effects of postseismic relaxation were found to include migrations in seismicity [*Rydelek and Sacks,* 2001] and the triggering or inhibition of remote seismicity [*Rydelek and Sacks,* 1990; *Pollitz and Sacks,* 1997; *Rydelek and Sacks,* 2003].

It is odd that *Smalley et al.* [2005] did not pursue the possibility of postseismic effects, since it was previously shown by *Rydelek and Pollitz* [1994] that a large-magnitude strike-slip earthquake in 1811 along the Bootheel lineament in the NMSZ could generate a localized high rate of strain in the present day. This rate may resemble some features of the Reelfoot GPS measurements if that observation were indeed the result of steady strain deformation.

To investigate postseismic effects specific to a large thrust earthquake, model calculations [*Pollitz,* 1992] were done for a M_w = 7.8 event in 1812 along the Reelfoot fault shown in Figure 1 and with fault parameters given in Table 1. This earthquake was the last, and largest, of the three events that occurred in the winter of 1811–1812. A viscoelastic Earth model [*Rydelek and Pollitz,* 1994] believed to be appropriate for this region of the central United States was used, and the calculations were run to span the time interval 2000–2005, that is, results for 5 years of postseismic viscoelastic relaxation that correspond to the times and regions of the GPS measurements of *Smalley et al.* [2005].

Figure 1 shows the model results for the engineering strain $\gamma = \varepsilon_{EE} - \varepsilon_{NN}$, where ε_{EE} and ε_{NN} are the compressional components of the strain tensor. Calculated strain rates of order 10^{-7} per year are found in the vicinity of the Reelfoot fault, and the corresponding

Table 1. Fault model parameters for 1812 Reelfoot Earthquake. Uniform slip on the fault plane is assumed.

Latitude =	36.64°
Longitude =	-89.55°
Strike =	158°
Length =	58 km
Width =	27 km
Dip =	40°
Rake =	90°
Slip =	11.0 m
M_w =	7.8 (computed from fault parameters)

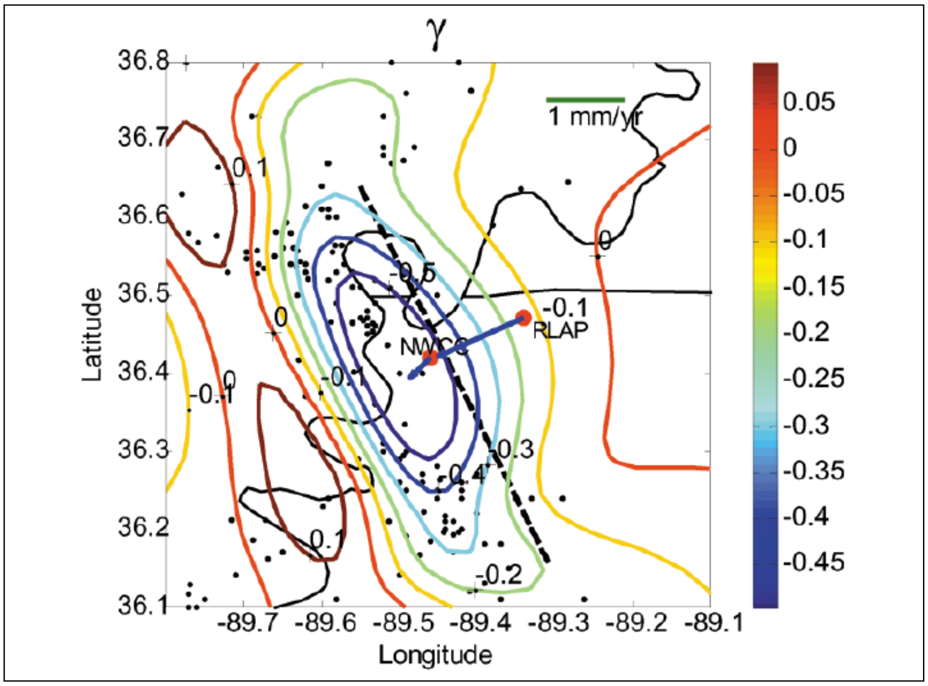

Fig. 1. Contours show the compressional components of the strain field from postseismic viscoelastic relaxation that would occur in 2000–2005 from a M_w = 7.8 thrust earthquake on the Reelfoot fault (dashed line) in 1812. Averaged over the 5-year span, the maximum strain rate is of order 10^{-7} per year on the hanging-wall side of the fault; the relative baseline shortening of GPS stations NWCC and RLAP is about 1 millimeter per year. The black dots are cataloged earthquakes with M ≥ 2.5 in the New Madrid Seismic Zone. Color scale is in units of microstrain.

convergence velocity of approximately 1 millimeter per year of the GPS locations across the fault is comparable to that reported.

Given the similarities between the deformation from the modeling of postseismic viscoelastic effects and the suggested scale of motions from the recent GPS measurements [*Smalley et al.,* 2005], it would seem premature to conclude that the apparent high rate of strain in this region is due entirely to accumulation, and therefore portends a significant hazard risk, until further data and analysis verify that this is not just a local effect of long-term postseismic relaxation. On the other hand, any offset in the GPS time series [*Calais et al.,* 2006] would be difficult to explain by either the steady accumulation of strain or the release of strain from long-term postseismic relaxation. Clearly, the modeling of postseismic viscoelastic effects and its interpretation may have important consequences for seismic hazard assessment in the central United States and should be considered in the unsettled and ongoing debate.

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