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THE -ENIGMA OF THE NEW MADRID EARTHQUAKES OF 1811-1812

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ABSTRACT

Continental North America's greatest earthquake sequence struck on the western frontier of the United States. The frontier was not then California but the valley of the continent's greatest river, the Mississippi, and the sequence was the New Madrid earthquakes of the winter of 1811-1812. Their described impacts on the land and the river were so dramatic as to produce widespread modem disbelief. However, geological, geophysical, and historical research, carried out mostly in the past two decades, has verified much in the historical accounts. The sequence included at least six (possibly nine) events of estimated moment magnitude **M ≥**. 7 and two of **M** 8. The faulting was in the intruded crust of a failed intracontinental rift, beneath the saturated alluvium of the river valley, and its violent shaking resulted in massive and extensive liquefaction. The largest earthquakes ruptured at least six (and possibly more than seven) intersecting fault segments, one of which broke the surface as a thrust fault that disrupted the bed of the Mississippi River in at least 2 (and possibly four) places.

... it is a riddle wrapped in a mystery inside an enigma.

Winston Churchill

1. INTRODUCTION

A sequence of powerful earthquakes struck the mid-Mississippi River Valley, central United States, in the winter of 1811-1812. The two largest probably exceeded the size of any continental western US earthquake. No fewer than 18 of these events were felt on the Atlantic seaboard or in Washington, DC (Nuttli 1987), at least 1000 km east, which implies moment magnitude **M** ≈ **6.0-6.5** (Table 1). Over time, this earthquake series has taken the name of the small riverboat town New Madrid, which lay at the heart of the epicentral zone and which in 1811 was the largest settlement on the river between St. Louis and Natchez. The name has proven apt, for New Madrid by happenstance marks the intersection of three of the six fault segments currently illuminated by microseismicity and believed to be rupture planes of the principal 1811-1812 earthquakes.

This review focuses on the 1811-1812 earthquakes themselves, their geophysical setting, and the factors that influenced their faulting dynamics and seismic moment release. The review is selective. For example, we do not assess the large literature on the seismic risk of the New Madrid seismic zone and the consequences of a repeat of an 1811 event, nor do we attempt to cover comprehensively the large body of geological and geophysical work in the New Madrid region that does not relate directly to the historical earthquakes. We justify this focus on the seismotectonics of the 1811-1812 earthquakes in terms of the uniqueness of the sequence: Globally it dominates all other documented earthquakes of stable continental regions (SCR) (Johnston et al 1994), a category of plate interiors that incorporates roughly 25% of all crust and fully two thirds of all continental crust. Why the New Madrid earthquakes are unique remains an enigma. Perhaps, given sufficient time, other stable continental plate interiors will experience earthquakes of the magnitude and numbers of the New Madrid sequence, although the worldwide historical record does not reveal a comparable sequence.

A comprehensive scientific assessment of the effects of the New Madrid earthquakes was not made until a century after their occurrence. Myron Fuller (1912) provides a thorough account of the geomorphic changes on the upper Mississippi Valley wrought by the earthquakes and a summary of the principal historical accounts. Placement of the earthquakes in the modern scientific framework of plate tectonics and seismic magnitude was achieved in the seminal papers by Burke & Dewey (1973), Ervin & McGinnis (1975), and Nuttli (1973). These papers initiated a 20-year period of concentrated research on the New Madrid seismic zone (Johnston & Shedlock 1992) that was spurred by the development of nuclear power generation in particular and seismic hazard concerns in general. Much of this work concerned the crustal structure in the vicinity of the 1811-1812 earthquakes, but there has been little additional

Table I The New Madrid earthquakes, 1811-1812

Event designator	Date	Time (local) (±30 min.)	m_{Lg} [a] 9	Estimated moment magnitude M: From m_{Lg} [b]	from isoseismal area [c]	M_0 release[e] (units of 10^{27} dyne cm)	(% total)
D1	16 Dec 1811	02:15	7.2-7.3	7.9	8. 1[c] ± 0.3	15.85	(44%)
D2	same	-03:00	6.2	6.2	6.6[d] ± 0.4 (A_{IV} radius)		
D3	same	07:15	5.5-6.0	5.8	5.9[d] + 0.55 (*felt* radius)		
D4	same	08:15	7.0	7.4	7.2[c] ± 0.3		
D5	same	10:00	-6.0	6.0	6.2[d] ± 0.55 (*felt* radius)		
D6	17 Dec 1811	12:00	6.6-6.8	7.0	7.1[d] ± 0.4 (A_{IV}, A_V radius)		
D7	16 Ian 1812	23:00	5.5-6.0	5.8	5.6d ± 0.55 (felt radius)		
		Total D1 sequence (m_{Lg} ≥ 5.0, **M** Z 4.7): 63 events				-18.0	(-50%)
J1	23 Jan 1812	09:00	7.1	7.6	7.8c ± 0.35.62 (16%)		
J2	same	23:00	-	-	5.5d ± 0.55 (*felt* radius)		
J3	27 Jan 1812	09:00	5.5-6.0	5.8	6.3d + 0.55 (*felt* radius)		
J4	04 Feb 1812	17:00	5.5-6.0	5.8	6.2d ± 0.55 (*felt* radius)		
		Total J1 sequence (m_{Lg} ≥ 5.0, M Z 4.7): 31 events				-5.7	(-16%)
F1	07 Feb 1812	03:45	7.3-7.4	8.0	8.0 [c] ± 0.3	11.22	(31%)
F2	same	20:00	5.5-6.0	5.8	6.3[d] ± 0.55 (*felt* radius)		
F3	same	22:40	6.6-6.8	7.0	7.0 [d] ± 0.55 (*felt* radius)		
F4	10 Feb 1812	16:00	6.2	6.2	6.5[d] ± 0.55 (*felt* radius)		
F5	11 Feb 1812	06:00	6.2	6.2	6.5[d] ± 0.55 (*felt* radius)		
		Total Fl sequence (m_{Lg} ≥ 5.0, M Z 4.7): 113 events				-12.1	(-34%)
Total 1811-1812 sequence:			>200 events, m_{Lg} 5.0, **M** ≥ 4.7:			-35.8 (**M** 8.3)	(100%)

[a] m_{Lg} estimated by intensity attenuation with distance by Nuttli (1973), Nuttli et al (1979), Street (1982), and Street & Nuttli (1984). [b] From the log (M_0)-m_{Lg} regression (period- 1 s) of Johnston (1996a). [c] From Johnston (1996c). [d] From log(M_0) regressions an isoseismal areas (Johnston 1996b); higher uncertainty results from using radii for equivalent areas. [e] Using $\mathbf{M} = (2/3) \log(M_0) - 10.7$ (Hanks & Kanamori 1979).

study of the events themselves, probably because of the dearth of quantitative information.

Constraints on the faulting that took place during the 1811-1812 earthquakes comes primarily from three sources: (a) historical accounts, including far-field intensity data and eyewitness reports from the epicentral zone (Figures 1 and 2), (b) seismological effects remaining from the earthquakes, such as preserved liquefaction features and present-day seismicity in the rupture zone (Figure 3), and (c) the physical structure of the crust of the 1811-1812 fault zone (Figure 4). Two additional and potentially powerful constraints are: (d) forward modeling of coseismic static strain fields to match with known topographic changes in the meisoseismal area and (e) geodetic/GPS measurement of the postseismic viscoelastic crustal relaxation still occurring today from the 1811-1812 faulting. Exploratory studies in both areas have been done (Gomberg 1992, Gomberg & Ellis 1994, Rydelek & Pollitz 1994), but the topographic and strain-rate databases must be considerably improved before they contribute significant constraints that rank with (a)-(c).

2. HISTORICAL SETTING

For the researcher trying to gain a modern understanding of the earthquakes, the timing and location of the New Madrid sequence is both fortuitous and frustrating. European settlement of the North American interior was well underway, and 1811 already fairly heavily traveled the Mississippi River. All river traffic was by unpowered flatboat, barge, or keelboat, but the first steamboat on the Mississippi River completed its maiden voyage from the Ohio River to New Orleans between the first principal earthquake on 16 December 1811 and the second on 23 January 1812. Settlements west of the Mississippi were so few that virtually all our information is limited to the river or points east (Figure 1). Had the New Madrid earthquakes occurred a century or so earlier they would have been included in the realm of paleoseismology; had they occurred a century or more later, millions of people would have been at risk, and abundant instrumental and macroseismic data would be available. However, they occurred in the transition, the crease in history, when the Mississippi River was for a brief period the western frontier of a new nation. Data useful for assessing the earthquakes in modem terms are available but fragmentary, and as a result legends and myths and scientific disbelief have proliferated concerning these events.

In order to gain an appreciation for the type of information available to us concerning the New Madrid events, consider the following descriptions of travelers caught in the massive liquefaction episodes that accompany major earthquakes in alluvial settings.

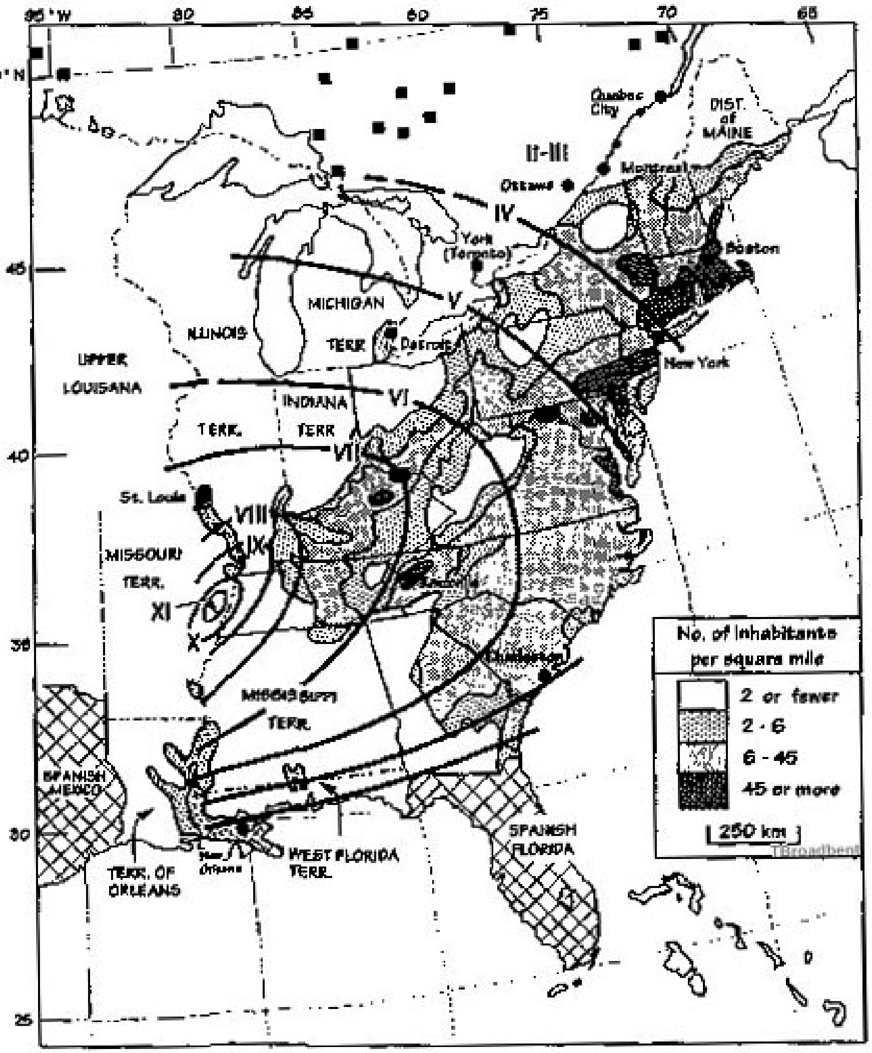


Figure 1 Historical setting in 1811-1812. States of the Union have continuous borders, territory and district borders have dash-dot borders, and Spanish possessions are cross-hatched. Modified Mercalli isoseismals of the 16 December 1811 mainshock (D 1) are from Stover & Coffman (1993), modified from Nuttli (1973). Population density for the US is for 1810 (Garrett 1988); in Canada, only principal settlements (dots) and trading posts *(squares)* are identified.

Figure 2 The Mississippi River in 1811-12. The river's precise course in 1811-12 is uncertain; this figure is based on maps from Cramer (1814), Wheeler & Rhea (1996), and US Geological Survey quadrangle maps. Island locations and numbers follow Cramer. See Table 2 for eyewitness references. (a) River locations of principal eyewitnesses to the earthquake. River width is exaggerated for clarity. Included for reference from Figures 3 and 4 are the seismogenic structures of the sequence. RS, Reelfoot scarp (dashed where inferred); RL, Reelfoot Lake; LP, Little Prairie; BFZ, Blytheville fault zone, BLS, Big Lake sunklands; SFS, St. Francis sunklands. Shaded area is eastern (nonfloodplain) upland. *(b)* The Kentucky (or New Madrid) bend of the Mississippi River in 1812, showing river locations of the principal eyewitnesses to the *FI* earthquake (all were in the Kentucky bend area). Present-day Reelfoot Lake is shown for reference (it was probably larger after the earthquake). Dash-dot line shows the 1995 river course (approximate centerline). Reelfoot scarp (from Van Arsdale et al 1995) has barbs on the hanging wall block and is dotted where inferred. Note that present-day New Madrid is - 2 km north of its 1812 location. Locations C-C' are profile endpoints for Figure 6.

I happened to be passing in its neighborhood where the principal shock took place... the water that had filled the lower cavities. . . rushed out in all quarters, bringing with it an enormous quantity of carbonized wood.. . which was ejected to the height of from ten to fifteen feet, and fell in a black shower, mixed with the sand which its rapid motion had forced along; at the same time, the roaring and whistling produced by the impetuosity of the air escaping from its confinement, seemed to increase the horrible disorder of the trees which everywhere encountered each other, being blown up cracking and splitting, and failing by thousands at a time. In the mean time, the surface was sinking and a black liquid was

Figure 3 Seismological setting of the New Madrid seismic zone. Plus (+) symbols are epicenters of magnitude (M_{bLg}, M_{coda}, or $M_{10\ Hz}$) ≤ 5.0 earthquakes located by seismic networks of St. Louis University and the Universities of Memphis and Kentucky from 1974-1993. Distribution of sandblow and fissuring liquefaction *(light shading)* from Obermeier (1989); trace of Bootheel lineament from Schweig & Marple (1991); Lake County uplift (LCU) from Russ (1982). Upland (nonfloodplain) areas are patterned. Open diamond symbols are sites of pre- 1811 paleoliquefaction features; see Section 6 and Figure 8 for label explanations and sources.

Figure 4 Structural setting of the New Madrid seismic zone. Epicenters and Bootheel lineament as in Figure 3. Blytheville arch boundaries *(outline with paired tics along rift axis)* from Hamilton & McKeown (1988). Margins of Reelfoot rift and flank plutons *(patterned shapes outside rift)* from Hildenbrand & Hendricks (1995); near-axial intrusive complexes *(patterned shapes within rift)* from Rhea & Wheeler (1994); New Madrid north fault (NN) from Zoback et al (1980); BP, Bloomfield pluton.

rising up to the belly of my horse, who stood motionless, struck with a panic of terror. . . . . These occurrences occupied nearly two minutes; the trees, shaken in their foundation, kept failing here and there, and the whole surface of the country remained covered with holes, which... resembled so many craters of volcanics...

... my [vehicle] suddenly began to rock in a most dangerous fashion ... As the rocking ceased, water spouts, hundreds of them throwing water and sand were to be observed on the whole face of the country, the sand forming miniature volcanoes, whilst the water spouted out of the craters; some of the spouts were quite six feet high... In a few minutes, on both sides of the road as far as the eye could see, was vast expanse of sand and water, water and sand. The road spouted water, and wide openings were to be seen across it ahead of me, then under me, and my [vehicle] sank while the water and sand bubbled, and spat and sucked till my axles were covered. 'Abandon ship' was quickly obeyed, and my man and I stepped into knee deep hot water and sand and made for shore.

... about sunrise another very severe one came on, attended with a perpendicular bouncing that cause the earth to open in many places... the deepest I saw was about twelve feet. The earth was, in the course of fifteen minutes after the shock... entirely inundated with water. The pressing of the earth, if the expression be allowable, caused the water to spout out of the pores of the earth, to the height of eight or ten feet! The agitation of the earth was so great that it was with difficulty any could stand on their feet, some could not. The air was strongly impregnated with a sulphurous smell.

The first and third of these graphic descriptions are of events in the New Madrid sequence (Bringer 1821, Fletcher 1812). [Bringer was not careful with dates, and it is unclear whether he was caught in a major aftershock or in the second main event of 23 January. Fletcher's account is from Little Prairie (Figure

2) and is of the major aftershock on 16 December.] 'Me second quotation, taken from Richter (1958), describes the great Bihar-Himalayan front earthquake of 1934, with an instrumentally determined moment magnitude of **M** 8.2. The main difference between this earthquake and numerous 1811-1812 accounts similar to those of Bringer is that the New Madrid earthquakes occurred in a densely forested region bisected by North America's largest river, and the effects on the trees and river were just as dramatic as the intense liquefaction phenomena.

A historical perspective is important for assessing both the near- and farfield effects of the New Madrid earthquakes. Figure 1 shows that the epicentral

region was on the forward edge of European settlement. Kentucky and Tennessee became states in 1792 and 1796, respectively; they were the only two states of the Union with territory in the meisoseismal zone, which lay primarily in what would become the states of Missouri (in 1821) and Arkansas (in 1836). The restricted population distribution (contoured in Figure 1) is a major problem in estimating the earthquakes' sizes from far-field isoseismal data (see Section 5).

Total population of settlements on the Mississippi River in the main disrupted zone-roughly from the mouth of the Ohio River to present-day Memphis, Tennessee-was less than 4000, with perhaps one half living in or near New Madrid (Penick 1981). (The number of Native Americans, although greater, is unknown.) Other estimates place New Madrid's population at only several hundred. Communications between the populous East and the river frontier were slow and unreliable. Not until months after the earthquakes did it become clear that all originated in the New Madrid area.

The earthquakes began with what was probably the largest shock of the entire series at 02:15 on 16 December 1811. (All times are local with probable ± 30 min uncertainty.) There were no known foreshocks. The mainshock isoseismals are shown in Figure 1. Note that no felt limit is included; given the early morning origin time and the limited population distribution, it has to date proven impossible to determine. Table I lists the major earthquakes of the ensuing sequence. The three principal shocks of 16 December 1811 (02:15), 23 January 1812 (09:00), and 7 February 1812 (03:45) are designated Dl, J1, and Fl. Major aftershocks are numbered sequentially after their mainshock. As Table I shows, a number of aftershocks were major earthquakes in their own right and thus cannot be neglected in fault rupture scenarios.

The main sequence duration spanned eight weeks, although the epicentral zone has remained active to the present and produced two additional large earthquakes in 1843 (**M** ~ 6.3) and 1895 (**M** ~ 6.6). Most authors (Nuttli 1973, 1983; Nuttli et al 1979; Street 1982; Street & Nuttli 1984) designate the third principal event F1 as the largest, in contrast with Johnston (1996c) and this review. It should be noted, however, that the magnitudes of the three principal events D1, J1, and F1 each lie within the uncertainty bounds of the other two, making their sizes all statistically equivalent, clustered about **M** 8.0 (Johnston 1996c). In addition to differences in size, variation in faulting mechanism and epicentral location can account for the reported differences in severity among D1, J1, and F1.

The most important of the meisoseismal historical accounts are geographically located in Figure 2 and summarized by principal event sequence and observed effects in Table 2. Nearly all are first-hand accounts taken largely from the compilations of Nuttli (1973), Street (1982,1984), and Street & Nuttli (1984). Many of the second-hand, summary accounts are cumulative for all the shocks and make it impossible to discriminate among D1, J1, and F1 effects, a problem that also frustrated Penick (1981).

There are, in addition, external factors that affect these surviving accounts from 1811-1812 that must be understood before informed interpretation is possible. For example, from Table 2 it appears that event J1 had no effect on the Mississippi River. In fact, after the mild weather during D1, severe cold weather iced over the Ohio River, and there was no riverboat traffic in the meisoseismal zone to report river effects for the J1 event. Likewise, the dearth of landslide reports in Table 2 is because the only steep slopes were the eastern bluffs of the Mississippi flood plain in Chickasaw Indian territory and Crowleys Ridge to the west of the flood plain in territory unsettled by Europeans. Only where the eastern bluffs approach the river (the series of Chickasaw bluffs, Figure 2a) were slope failures actually witnessed, though abundant landslide scars from 1811-1812 on both the bluffs (Jibson & Keefer 1988) and the ridge (Ding 1991) have been recently mapped. Finally, the only reports of waterfalls or rapids are for F1. We believe this is because only F1 involved thrust faulting that resulted in static offset and disruption of the riverbed (a possible exception is D1 aftershocks). This uniqueness of event F1 contributes an important constraint to the faulting scenarios we present in Section 7.

The remarkably few extant first-hand accounts from people who were caught in the 1811-1812 earthquakes are an irreplaceable resource. In this brief review, we cannot present the descriptions in detail but can only touch on highlights. For the D1 sequence the accounts from the small flotilla of flatboats, keelboats, and barges tied up for the night along the Mississippi River (north to south: La Roche 1927, Bedinger 1812, Pierce 1812, Davis 1812, and Bradbury 1817) make it clear that the most severe vibration and liquefaction was in the Little Prairie vicinity and that it was more intense for the D1 aftershocks (Table 1) than for D1 itself. For example, only Pierce and Bedinger recount large waterspouts on the river or explosive cratering. A large wave (and perhaps a temporary retrograde current) was noted only upriver of Little Prairie [Bedinger, Bryan (1848), and La Roche] and was followed by a rapidly rising river level and swifter current downstream from approximately Bedinger's location (Bedinger, Davis, Pierce, Bradbury). These latter events are consistent with the large volume of groundwater that must have been expelled by liquefaction. Tremendous noise, fissuring, splintering and toppled trees, and extensive caving of river banks were reported by all.

We shall apply these historical observations and others as constraints on plausible faulting scenarios for the D1, J1, and F1 sequences. However, other constraints come from the seismological and geophysical setting of the faulting and the size of the earthquakes, and we must therefore first examine the current scientific understanding of these aspects.

3. SEISMOLOGICAL SETTING

Earthquake triplets, with each principal shock generating its own aftershock series, are relatively common in intraplate settings and stand in contrast with most large earthquakes, with their classic mainshock-aftershock sequence. The New Madrid triplet, occurring within an eight-week period, is the largest known, but other recent sequences and their moment magnitudes include the following: 1990 Sudan (**M** 7.3, 6.7, 7.2 in 5 days); 1976-84 Gazli, Uzbekistan (**M** 6.8, 6.9, 6.9 in 8 years); 1985-88 Nahani, Canada (**M** 6.6, 6.8, 6.2 in 3 years); 1988 Tennant Creek, Australia (**M** 6.2, 6.3, 6.6 in 12 hours); and 1982 Miramichi, New Brunswick (**M** 5.6, 4.9, 5.0 in 3 days). All, including New Madrid (e.g. Ellis & Schweig 1995), are distinguished by the fact that each principal shock occurs on a different fault segment and that these segments exhibit complex intersecting patterns. Simple, throughgoing faulting for large earthquakes on well-developed faults may be the rule for interplate zones but apparently is less prevalent in continental plate interiors.

The present-day seismicity of the New Madrid seismic zone (NMSZ) (Figure 3) exhibits a complex pattern of intersecting linear segments in either map (epicentral) or cross-section (hypocentral) view (e.g. Chiu et al 1992, Himes et al 1988). A common assumption -one that we hold also- is that the current seismicity is illuminating the most active faults, i.e. those that ruptured in 1811-1812 and also in pre-1811 activity. The hypocenters are located with regional seismographic networks operated by three universities (Saint Louis, Memphis, and Kentucky). Within the NMSZ, which is defined by the epicenter alignments within 35.5-37.0'N and 89.2-90.5'W, the-epicentral uncertainty is < 1 km, but depths are poorly constrained. Since network monitoring began in 1974 more than 3000 NMSZ earthquakes have been located (Herrmann et al 1994), none of which reach moment magnitude **M** ≥ 5.0. A halo of seismicity ~ 100-200 km to the west and north of the NMSZ (mostly beyond the area of Figure 3) is present in both the historical- and network-era record and represents a seismic activity level significantly higher than the rest of the central United States.

The most extensive and dramatic present-day evidence that survives from the 1811-1812 ruptures is not the current seismicity but the sandblow and fissure liquefaction features preserved in the alluvial soils of the upper Mississippi embayment. A comprehensive mapping of these features was carried out originally by Fuller (1912) and then in much more detail by Obermeier (1988, 1989). Obermeier's zone, depicted in simplified form in Figure 3, encompasses only the areas of most intense liquefaction where 1% to > 25% of land surface is covered. Reports of significant outlying liquefaction episodes far removed from this zone include White County, Illinois, and a region near St. Louis, ~ 250-300 km north of the NMSZ (Berry 1908, Obermeier 1988); to the south to below the mouth of the Arkansas River (La Roche 1927), about 250 km from the NMSZ; and at Big Prairie (severe liquefaction), near present-day Helena, Arkansas, at > 100 km distance (Street & Nuttli 1984).

The area of Obermeier's severe liquefaction zone exceeds 10,000 km^2, and Fuller's zone, encompassing all types of ground failures, as modified by Street & Nuttli (1984), covers ~ 48,000 km^2 . These rank among the largest earthquake liquefaction and ground deformation fields ever documented, perhaps surpassed only by those produced by great Himalayan front earthquakes in the Ganges alluvial plain of India (Richter 1958). Moreover, the New Madrid liquefaction zone obviously was restricted by loess-mantled upland areas of low-to-negligible liquefaction potential to the west (Crowleys Ridge) and to the east (the river's eastern bluffs) (Figure 3). The New Madrid liquefaction zone encompasses all the linear epicentral segments of the NMSZ (except the eastern extreme of its central segment that extends beneath the loess bluffs), strongly suggesting that both the seismicity and the liquefaction are linked to the same set of source faults.

4. STRUCTURAL SETTING

Quite apart from the large earthquakes that struck there, the crust of the central Mississippi valley region is complex and challenging to geologists and geophysicists alike. The fact that this crustal volume produced a great earthquake sequence in historical times, however, provides additional impetus and support for the research. The complexity of the 1811-1812 sequence-a multiple-event seismic moment release rather than a simple **M** ~ 8.3 mainshock and aftershock sequence-most probably is a reflection of the heterogeneity of the crust in which the ruptures took place. Moreover, the multiple rupture pattern combined with the low finite displacement of the fault system as a whole is an indication that the fault zone is relatively young and will, over time, evolve into a simpler system (e.g. Schweig & Ellis 1994). There are a multitude of crustal structures at all scales in this failed continental rift setting. We are very selective in this section, emphasizing only those that played a prominent role in influencing the 1811-1812 fault ruptures.

Reetfoot-A Failed Rift

The Reelfoot rift (Ervin & McGinnis 1975), host structure to the New Madrid seismic zone, formed in Late Proterozoic to Early Cambrian times as an aulacogen or failed rift off the margin of the opening Iapetus Ocean, predecessor of the present-day Atlantic and Gulf of Mexico. Hence, a triple junction was involved. Burke & Dewey (1973) proposed a model with two Late Paleozoicto-Mesozoic triple junctions, one near present-day Jackson, Mississippi, the other near present-day Dallas. The work of Ervin & McGinnis, however, established a much more ancient rifting history. The triple-junction model has proven very useful: The successful arms of the junctions are the rift-developing oceans, and the failed arms are the Mississippi embayment and the southern Oklahoma aulacogen/Anadarko basin, respectively. The southern Oklahoma rift is the host structure of the Meers fault, which had a surface rupture - 1200 years ago, probably from an **M** ~ 7 earthquake (Crone & Luza 1990).

The work of Hildenbrand and others (Hildenbrand 1985, Hildenbrand & Hendricks 1995) has rather precisely delineated the boundaries of the Reelfoot rift of Ervin & McGinnis (1975) with potential field -mainly aeromagnetic- data. Ervin & McGinnis proposed a Late Precambrian triple junction at the intersection of the embayment axis and the Gulf coastal plain, a model similar to Burke & Dewey but 300-400 Ma earlier. Hendricks (1988) describes a Late Precambrian triple junction with the southern Oklahoma and Reelfoot as failed rifts branching off it; presumably this would then be a quadruple junction. Braile et al (1982, 1986) also proposed a quadruple junction more northerly than the others with all four arms failed.

More recently, Thomas (1991) has modeled the evolution of the southern North American margin with a quadruple junction in southern Arkansas similar to that of Hendricks (1988). In this model in the Early Cambrian, the southern Oklahoma and Reelfoot rifts are failed arms and the Ouachita rift and Alabama-Oklahoma transform are active boundaries. Hildenbrand & Hendricks (1995) have detected the relic AL-OK transform fault in gravity and magnetic data and identify it as the southern boundary of the Reelfoot rift. Its interpreted location is ~ 150 km southwest of the southwestern tip of the Blytheville arch. To the north, the Reelfoot rift probably merges with the east-west Rough Creek graben (Thomas 1991, Hildenbrand & Hendricks 1995). Although many details of evolution and configuration differ among the above models, the Reelfoot rift is a failed rift arm in all of them. Establishing the rifted character of the NMSZ crust is of fundamental importance to understanding the earthquake potential of the region because, worldwide, all large (**M** ;t 7) stable continental earthquakes occur in crust that has experienced such extensional tectonics (Johnston et al 1994).

Intrusions and the Blytheville Arch

As is commonplace in continental extension, the Reelfoot rift was the site of extensive magmatic activity during its development, although there is no firm evidence of volcanism during the initial (Late Precambrian to Cambrian) rifting stages (Hildenbrand 1985). The oldest intrusive activity probably was in the Late to Middle Paleozoic, although the large Bloomfield pluton (BP on Figure 4) may be Cambrian or older (Hildenbrand & Hendricks 1995). No intrusions younger than Late Cretaceous have been documented. There was extensive intrusive activity, especially along the northwest rift flank (Figure 4), but except possibly for the Bloomfield pluton, it had no influence on 1811-1812 faulting. The intra-rift intrusives, however, likely played a pivotal role in the observed complexity of the faulting sequence.

The axial rift intrusives in Figure 4 are those inferred from subdued gravity highs by Langenheim (1995), as interpreted by Rhea & Wheeler (1994). They have little or no magnetic signature, which suggests they are composed of dense, nonmagnetic rocks. Also within the rift, but for the sake of clarity omitted from Figure 4, are several large magnetic highs, also interpreted as intrusive complexes (Hildenbrand ' & Hendricks 1985) but with less expressive gravity signatures. A very simplistic explanation is that the axial, dense intrusive rocks are more felsic with fewer magnetic minerals than the more mafic magnetic intrusives. The latter's lack of gravity expression may be because they consist of multiple ring-dike complexes that lack an overall high-density contrast with the country rock of the rift's crust.

Thus intrusions are ubiquitous throughout the Reelfoot rift and its margins. Their emplacement commonly was controlled by preexisting structures, probably mainly faults, but the intrusions, in turn, almost certainly influence, perhaps control, subsequent faulting. It seems likely that, depending on composition and age, intrusions may be both a barrier to rupture nucleation and propagation, e.g. the massive Bloomfield pluton, or may enhance rupture nucleation and propagation as the axial intrusives appear to do (Figure 4) along the Blytheville arch. The Blytheville arch (Figures 2a and 4) was originally defined and mapped from industry seismic reflection profiles (Howe & Thomson 1984, Crone et al 1985). Its characteristic signature in these data is a strong upwarp of Paleozoic strata within a ~ 10-15-krn wide zone that widens to the northeast and is roughly centered on the axis of the Reelfoot rift. Flat-lying, continuous strata of Late Cretaceous and younger age overlying the upwarp and axial disruption of coherent reflectors in the crystalline basement beneath the upwarp also are characteristic. The borders of the Blytheville arch were subsequently refined by Hamilton & McKeown (1988), McKeown et al (1990), and McKeown & Diehl (1994). The cause of the upwarp of the Paleozoic reflectors remains a matter of debate, with Crone et a] (1985) favoring axial intrusions and McKeown et al (1990) arguing for diapiric action of less-dense sediments at the sediment-crystalline basement contact.

Langenheim (1995) interprets gravity data as indicating that nearly the entire arch is coincident with shallow intrusions, which would seem to support the intrusion mechanism. Figure 5a is a schematic depiction of this model. Arguing against the intrusion model, however, is the fact that not even subtle magnetic anomalies are present along the trend of the arch (Hildenbrand & Hendricks 1995). In addition to diapiric action, another way of getting the less-dense elastic sediments to intrude into the upper carbonate section and to produce a long, linear upwarp of strata is through a positive flower structure. Flower structures and associated folds have been recognized on seismic reflection lines 1 in strike-slip fault zones throughout the world (e.g. Sylvester 1988, Harding 1985). A positive flower structure, i.e. one that is arched along its length, is indicative of strike-slip faulting with a component of compression across the fault (Harding 1985). Thus, the Blytheville arch may have been formed during a period of transpressional strike-slip faulting along preexisting axial faults.

Regardless of which uplift mechanism is correct, they both include an axial fault zone at seismogenic depths within the crystalline crust. The axial fault zone is a first-order feature, perhaps the most important of any within the Reelfoot rift. Its axial location suggests it localized most of the extensional deformation of the rift's evolution. The initial nucleation of the 1811-1812 sequence (event D 1) was probably on the axial fault. As noted by Hamilton & McKeown (1988), the arch changes its fundamental character just at the southern boundary of the Missouri Bootheel (36'N). South of this point, the arch's upwarped reflectors are present and continuous on seismic reflection profiles; to the north they are absent, but the basement fault zone remains clear. Figure 4 shows that the intrusions based on gravity interpretations also deviate to the northwest from the rift axis and the seismicity at this point. Thus from the base of the Bootheel northeast to the Mississippi River the Blytheville arch apparently is not an arch but rather a major fault zone in the crystalline basement along the rift axis, one that lacks a gravity or magnetic signature of associated intrusions.

We believe that this fundamental change in the structure of the Blytheville arch had a major influence on the rupture history of the 1811-1812 earthquake sequence. Accordingly, we divide the arch into two different fault segments: the Blytheville arch (BA), southwest of the vicinity of present-day Blytheville (~ 36'N), and the Blytheville fault zone (BFZ), northeast of this point. The BA is more structurally complex and diffuse because it is also intruded by igneous rocks (Langenheim 1995) and has upwarped overlying sedimentary units. In Section 7 we examine how these structural differences might have affected the DI coseismic rupture, vis-à-vis the major D1 aftershocks. First we examine the BFZ more closely.

Seismogenic Faults

THE BLYTHEVILLE FAULTZONE (BFZ) The BFZ (Figures 2a and 9) is an on-trend continuation of the Blytheville arch to the northeast for - 55 km. Whether it is continuous with arch structures is not known, but a simple interpretation is that it is an unintruded extension of the axial fault zone. Alternatively, if the BA is a positive flower structure, the BFZ may represent a section that has not been under as much or as extended a period of compression and thus has not developed an arch. The lack of uplift of Paleozoic strata but the presence of disrupted basement reflectors is clear on seismic profiles (Hamilton & McKeown 1988) and is coincident with the concentrated zone of seismicity trending northeast across the river (e.g. Figure 4). The BFZ is important to fault-rupture scenarios (Section 7) because of its proximity to Little Prairie (Figure 2a), where some of the best historical accounts of the D1 sequence originate. An important constraint is that the major aftershock D4 and perhaps others were stronger than the D1 mainshock in Little Prairie.

It is possible that the BFZ and the southwestern extension of the Cottonwood Grove fault identified by river seismic profiling (Shedlock & Harding 1982) constitute a single fault system. The Cottonwood Grove fault trends subparallel to and ~ 4 km southeast of the BFZ. If the BFZ, the Cottonwood Grove fault, and the unnamed subparallel fault to the northwest of the Cottonwood Grove are all part of the same system, then the BFZ could continue and intersect the Reelfoot fault at the southwest end of Reelfoot Lake, extending its length to ~ 65 km.

THE REELFOOT FAULT (RF) The RIF is the deep, seismogenic fault that is expressed at the surface as Reelfoot scarp (Section 6). It is the only seismogenic fault in the NMSZ with clear surface expression. The RF has been mapped at the surface for ~ 32 km (Van Arsdale et al 1995), and segments of it have been imaged in shallow sediments (Woolery et al 1996) and in Paleozoic rocks to ~ 3 km depth with a 60 to 70° dip (Sexton & Jones 1986) using high-resolution seismic reflection profiling. Recurrent movement on the RF is indicated because older units exhibit greater (dip-slip) displacement than younger ones. Sexton & Jones estimated ~ 60 m of reverse offset of Late Paleozoic rocks but only ~ 15 m in for Late Eocene units. Below the sedimentary rocks of the Mississippi embayment, the RF has not been detected on seismic profiles (e.g. Zoback et al 1980), but Chiu et al (1992) have resolved a tabular zone of hypocenters dipping ~ 31° southwest to a depth of 12 to 14 km that may be the seismogenic expression of the RF in the crystalline basement (Figure 5b). The surface projection of this tabular zone would reach the surface approximately at the Reelfoot scarp if the 60-70° dip of Sexton & Jones is used at depths of ≤ 5 km.
An alternative interpretation to an arbitrary - 40' change in dip is depicted in Figure 6, which compares Reelfoot fault and s

Figure5 Schematic Reelfoot rift crustal profiles across the two principal seismogenic faults of the NMSZ. K, T, and Pz designate Cretaceous, Tertiary, and Paleozoic sedimentary units, respectively; pattern below Pz denotes crystalline basement. (a) NW to SE profile across Reelfoot rift at about Blytheville (Figures 3 and 4). Inferred axial fault zone from McKeown & Diehl (1994); depth extent of flank plutons and axial intrusions is highly uncertain. Depths and thicknesses of the sedimentary sequences and anomalous lower crust from data and sources in Rhea & Wheeler (1994) and Wheeler et al (1994). The quartz and feldspar brittle-plastic transitions (bpt) occur at ~ 300°C and ~ 450°C, respectively; the approximate indicated depth ranges depend on the rift's geothermal gradient (see text). *(b)* NE-SW profile across Reelfoot fault along profile C-C' (Figure *2b),* just north of Reelfoot Lake. The anomalous lower crust is thicker and the Moho deeper than for the axial fault zone of (a). Hypocenters from Chiu et al (1992); Reelfoot fault in the K and T layers from Sexton & Jones (1986); other profile data, including Pz uplift as the Pascola arch and depth of magnetic basement are from Rhea & Wheeler (1994) and Wheeler et al (1994). 0 and C in the Pz section refer to Ordovician and Cambrian units, respectively. The Reelfoot fault and scarp may be a splay reverse fault off the inferred master thrust fault that generated event Fl, or the dip of the thrust might steepen to become the Reelfoot fault in the K and T surface layer (see Figure 6).

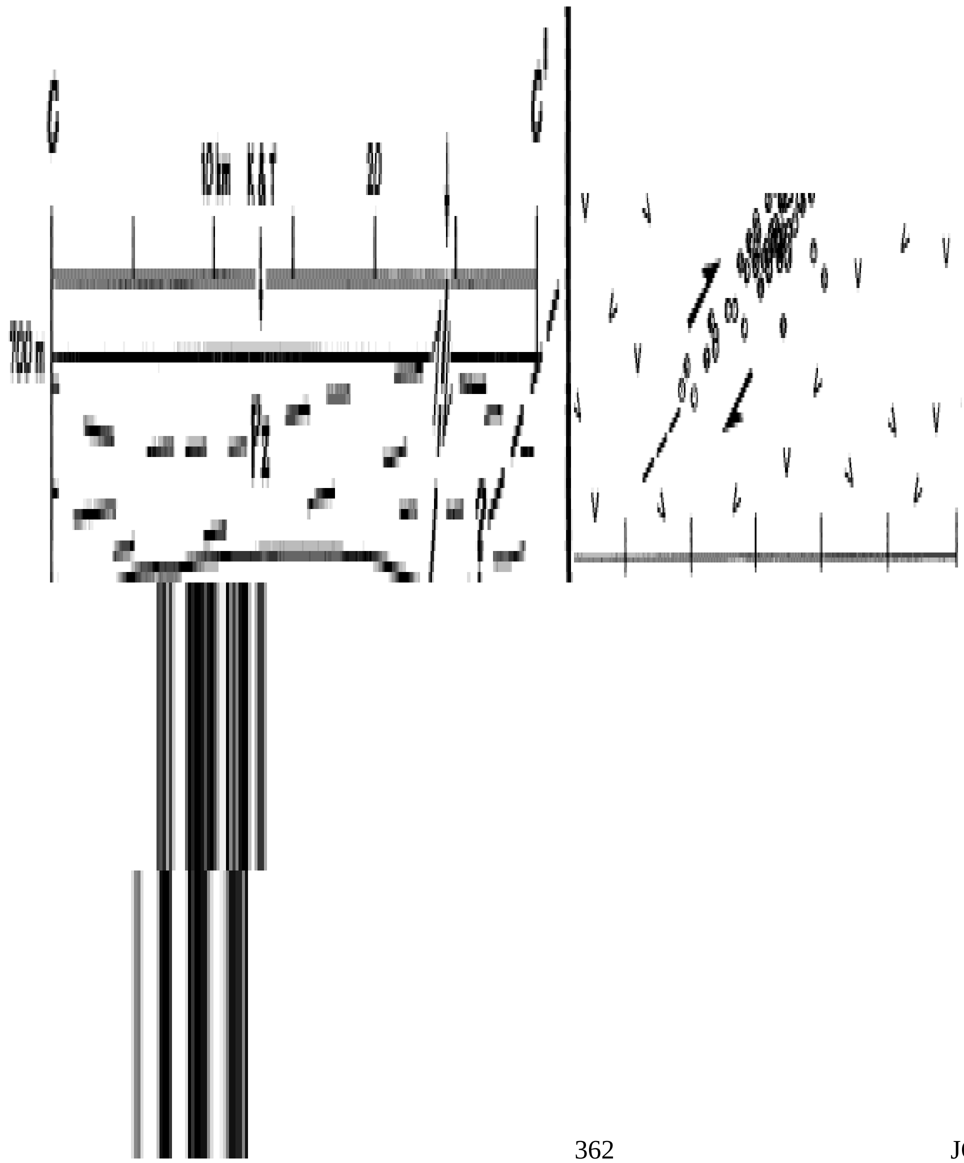
Reelfoot Fault scarp

Figure 6 (a) Topography of Reelfoot scarp from locations C to C′, Figure 2b. The elevation profile is shown relative to the inferred master thrust fault for event F1 in Figure 5b, but this crustal profile has no vertical exaggeration. There are no data that show Reelfoot fault through the Pz section; also no hypocenters are located there. (b) A similar geometry and relationship between elevation change and a shallow-dipping thrust rupture plane for the 1964 M 9.2 Alaska earthquake (modified from Plafker 1965). As in (a) most surface deformation is concentrated at the steeply dipping splay fault. The box labeled RF shows the size of the Reelfoot fault profile in (a) at this scale.

the great **M** 9.2 Alaska earthquake of 1964. Both examples have gently dipping thrust planes; for Alaska it is the top of the subducted Pacific plate; for New Madrid it is the alignment of hypocenters below 5 km. In Alaska, most of the vertical deformation was accommodated on a steeply dipping splay fault (or faults) off the master thrust at Montague Island. If similar faulting mechanics obtain in the NMSZ, then Reelfoot scarp would be a splay fault off the master Reelfoot thrust plane at depth. The splay fault concentrates the vertical deformation at the scarp rather than farther to the northeast along the shallowing master thrust. Furthermore, because they are mechanically more efficient, steep splay faults may be evolutionary in that they accommodate more and more of the vertical deformation with time as the master thrust matures through repeated faulting episodes. Applied to the New Madrid seismic zone, this model predicts that Reelfoot scarp will grow vertically in response to the regional approximately east-west compressional stress field at the expense of further elevating the eastern portion of the Lake County uplift.

The mapping of Reelfoot scarp through the Kentucky bend area of Figure 2b provides the single most stringent constraint to 1811-1812 fault rupture scenarios, for it restricts the Fl principal event to rupturing the RF segment. This can be stated unequivocally because of the precise dates and locations contained in the accounts of Speed (1812) and the "patron" (Shaler 1815), which place the first upstream waterfall on the Mississippi at island #10 (Figure 2b) in the early morning of 7 February 1812 (the date and time of FI). [The fact that river travelers used the island numbering of Cramer (1814) in their accounts greatly increases the scientific value of these historical records.] The location and existence of falls downstream of island #10 are much less certain in the historical accounts, but the reality of significant riverbed disruption in the Kentucky bend area of Figure 2b has been confirmed more than 150 years later by trenching and coring studies (Van Arsdale et al 1995) and both river and on-land seismic profiling (Odum et al 1996, Woolery et al 1996).

THE NEW MADRID NORTH FAULT (NN) This unnamed fault, which we call New Madrid north (NN), was identified in seismic-reflection profiles reported in Zoback et al (1980) and shown in Hamilton & Zoback (1982). Its length (minimum of ~ 30 km) and location (Figure 4) make NN of potential importance in fault rupture scenarios. It parallels but is slightly offset from the north northeast NMSZ seismicity segment that extends nearly to Illinois. Zoback et al (1980) describe it as having the largest apparent offset of reflectors (~ 35 m, east side down) in the area west of New Madrid; however, the Eocene Paleocene contact was not offset, indicating a Paleocene age for most of the dip-slip component.



As seen on Figure 4, NN could be regarded as a northern extension of the Bootheel lineament. Unlike the Bootheel lineament, however, NN traverses an area of medium-to-low liquefaction potential (Obermeier

1989); therefore the surface liquefaction features that clearly define the former are absent for the latter. If NN, which apparently is steeply dipping, was reactivated in 1811-1812 in a strike-slip sense, then the displacement would not show on the seismic profiles.

THE BOOTHEEL LWEAMENT (BL) The detailed surface trace of the northern Bootheel lineament forms a series of en echelon fissures, apparently formed during the 1811-1812 earthquakes (Schweig & Marple 1991, Schweig et al 1992a); a simplified continuous trace of these fissures is used in Figures 2a, 3, and 4. High-resolution seismic-reflection profiles across the lineament (Schweig et al 1992b, Sexton et al 1992) suggest that the fissures are the surface traces of a subsurface fault system that forms multiple flower structures. This deformation affects units as young as an Eocene/Quaternary unconformity, the youngest reflector visible on the profiles. Flower structures have been shown in experimental studies (e.g. Naylor et al 1986) to be cross sections through three-dimensional structures known as Riedel shears (Tchalenko 1970), which are common in developing strike-slip fault zones on many scales. Schweig & Ellis (1992) have argued that the en echelon fractures along the BL may also form distinct Riedel shears, perhaps indicating that the lineament is a newly developing strike-slip zone, the first stage of a straightening and smoothing of the irregularly segmented NMSZ. However, it is clear that the BL, is not a new fault, formed in 1811: The degree of structural deformation on all seismic reflection lines across the lineament is too great to attribute to a single rupture.

The BL does not exhibit nearly the level of current seismicity observed on other discussed fault segments, such as the Blytheville arch, the Blytheville fault zone, the Reelfoot fault, and perhaps the New Madrid north fault. The reason for the sparse seismicity is unknown, but we speculate that it may reflect the fact that the BL is younger and not as structurally developed as these other faults. However, because a small lingering doubt remains as to whether the BL is a seismogenic fault, we continue to refer to it as a lineament.

5. SIZE OF THE PRINCIPAL EVENTS

On the basis of the extensive macroseismic effects including massive liquefaction, fissuring, subsidence, or uplift of landforms; violent disturbance of the river; and destruction of extensive tracts of forests (Table 2), both Davison (1936) and Richter (1958) considered the principal New Madrid events to be "great" earthquakes. In Richter's case, this presumably means a Richter magnitude $\mathbf{M}_R \geq 8$, although he never explicitly assigned a magnitude value. In Davison's case, this put New Madrid in the company of the famous 1755 Lisbon, 1897 Assam, 1906 San Francisco, and 1891 Nobi earthquakes, among others. Both authors' estimates were based on the work of Fuller (1912) and consisted more of informed judgment than of quantitative analysis.

All useful analysis of the New Madrid earthquakes-from their recurrence intervals to dynamic fault-rupture modeling to present-day seismic hazard assessment-hinges on their size. In modern seismology "size" equates to the scalar seismic moment M_0, which defines moment magnitude $\mathbf{M} = (2/3) \log (M_0) - 10.7$ (Hanks & Kanamori 1979). [The symbol **M** conforms to the usage in the formal Hanks & Kanamori definition: for average stress drop, **M** is equivalent to the seismic energy magnitude *M*,,, of Kanamori (1977).] Previous magnitude estimates for the 1811-1812 events (Nuttli 1973, 1983; Nuttli et al 1979; Street 1982, Street & Nuttli 1984) were in terms of the short period magnitudes m_b or m_{bLg}, not easily related to scalar moment or moment magnitude.

There are a wide variety of analytical techniques to directly recover M_0 from instrumental seismic-wave data. For noninstrumental, historical events like New Madrid, a direct estimate of M_0 is possible only if the rupture is well expressed at the surface. But the scalar moment's physical definition ($M_0 = \mu dA$, where μ is the rigidity, *d* is the average rupture displacement, and A is the fault area) is dependent on both average slip and rupture dimensions. The New Madrid ruptures evidently were well masked or at least not preserved by the thick alluvial sediments of the Mississippi River Valley, except for the relatively short Reelfoot fault segment (see previous section).

Because of this lack of instrumental or surface-rupture data, the seismic moments of the New Madrid earthquakes must be estimated indirectly from historical accounts. Notwithstanding a sensitivity to source and path effects, the best type of data for this estimate is seismic intensities, which in North America are mostly standardized to the modified Mercalli (MMI) scale. Extent and severity of liquefaction (Youd et al 1989) and landsliding (Keefer 1984) can also provide useful constraints. All these effects were employed by Johnston (1996b,c) to obtain seismic moments with specified uncertainty bounds for the principal events D1, J1, and F1 and the largest aftershock D4 of the New Madrid sequence. This effort followed on the pioneering work of Nuttli (1973) with subsequent refinements (Nuttli 1983, Nuttli et al 1979, Street & Nuttli 1984) using intensity data to obtain the sizes of these events in terms of the short period regional magnitude m_{bLg} (see Table 1). However, because m_{bLg} has no direct relation to fault rupture parameters, size measured by M_0 or **M** is more useful for our purposes here.

Figure 7 The regressions of seismic moment (moment magnitude) on modified Mercalli felt isoseismal areas for stable continental regions (modified from Johnston 1996b). Large dots are the isoseismal areas for the DI New Madrid event (see Figure 1), reduced according to Johnston (1996c). Formal one-standard-deviation uncertainties in predicted log(MO) values are indicated as well as the best weighted average, M8.1. Horizontal shaded band at a radius of 100-150 km separates domains in which intensities (seismic-wave amplitudes) are controlled by the geometrical spreading of body waves and by anelastic attenuation of surface (Lg) waves. For very large earthquakes such as DI, are in the Lg domain.

Johnston (1996c) relied on a set of regressions of instrumental M_0 on isoseismal area (felt through VIII) developed for stable continental region earthquakes (Johnston 1996b, Johnston et al 1994). These regressions are summarized in Figure 7. To these basic relations, corrections were applied for North America's extremely low anelastic attenuation to the northeast of New Madrid and for the considerably higher attenuation to the west (Singh & Herrmann 1983). The isoseismal areas of event D1, calibrated in this manner, are added to Figure 7. With data weighting, the outer isoseismals have the lowest uncertainty and dominate the best weighted average M_{best} . Hence the low A_{IV} value more than compensates for the high A_V and A_{VII} values, yielding $\mathbf{M}_{best} = 8.1$. The one-standard-deviation uncertainty is ~ ± 0.3 **M** units. At this uncertainty level **M** 8 earthquakes can be clearly discriminated from **M** 7 or **M** 9 events if isoseismal data are adequate; discrimination from **M** 7.5 or **M** 8.5 is marginal. Similar **M** estimates 1, and D4 events using the isoseismal data of Street (1982) and Street & Nuttli (1984).

Three aspects of the size analysis in Figure 7 and results in Table 1 deserve further comment. Beyond an isoseismal area radius of ~100-150 km (shaded band, Figure 7), the anelastic attenuation of the continental *Lg* wave controls seismic ground motions; within ~ 100-150 km, geometrical spreading is the dominant control parameter (Frankel 1994, Hanks & Johnston 1992). The New Madrid D1 isoseismal areas are huge -among the largest known in the world- and even the outer radius for the onset of structural damage (MMI VIII) is in the regime of *Lg* attenuation, not geometrical spreading. This means that, for very large earthquakes (**M** ~ 7.5) such as event D1, damage areas in eastern North America may well be significantly larger than those in the West. Conversely, this may not be the case for earthquakes smaller than **M** 7.0-7.5 (but for a counterview see Bollinger et al 1993).

Second, there are insufficient data to determine a reliable felt area (MMI- I-III) for the New Madrid sequence principal shocks. From Figure I it is evident that a felt area determination would require a comprehensive examination of historical records from French and British Canada, Spanish Mexico (Texas), and Florida, and the unsettled but explored American West. Such a study has not been undertaken. Well-constrained felt limits could significantly reduce the uncertainties in the sizes of the principal 1811-1812 earthquakes.

Finally, from Table 1, the total seismic moment release of the 1811-1812 sequence is an estimated M_O(total) = 3.6 x 10^{28} dyne-cm, equivalent to a single great **M** 8.3 earthquake. How such a large moment release could occur on fault Segments of the dimensions defined by current seismicity and the Bootheel lineament becomes a question of primary importance. For example, if coseismic slip is limited to ≤ 10 m and occurs at depths of ≤ 20 km (depth maximum of current NMSZ hypocenters), a total minimum fault length exceeding 500 km is required for normal upper crustal rigidity. The total fault length of the 7 fault segments that we believe participated in the 1811-1812 ruptures (see Section 7) is ~ 350 km. This ~150 km discrepancy can be resolved in at least four ways: 1. Reduce the estimated seismic moments of the D1, J1, and F1 earthquakes (e.g. Gomberg 1992). 2. Allow unprecedentedly large coseismic fault displacements (d (15 m). 3. Allow rupture to depths significantly greater than the hypocentral depths of current seismicity. 4. Allow rupture on fault segments not identified by crustal structure studies or current seismicity.

A significant reduction in size of the 1811-1812 principal events leads to irresolvable difficulties in accounting for both the severe and extensive near-field effects and the great distances of far-field effects. Johnston (1996c) expressed the belief that the estimated seismic moments of events D1, D4, J1, and F1 (used in Table 1) were more likely to be in error by underestimating rather than overestimating size. The large displacement hypothesis also has numerous difficulties. Aside from exceeding known average displacements for all earthquakes except M > 9 subduction zone ruptures, it implies very high static stress and strain drops, which in turn imply relatively short rupture duration times. Multiple independent historical accounts in the near field indicate durations approaching one minute to several minutes, not seconds. Moreover, the abundant historical and physical evidence of massive liquefaction, which needs many vibrational cycles to develop fully, would argue for longer, not shorter, source time functions. The fourth, the unidentified coseismic fault segment(s) hypothesis, is a possibility that cannot be dismissed, but it amounts to special pleading with no supporting theory or data, and therefore we do not consider it further.

Johnston (1996c) has presented the case for the third possibility, that of deep coseismic rupture. The argument relies on the synoptic shear zone model for continental crust, developed most fully by Scholz (1988, 1990). This model applies to thick, "average" cratonic crust with abundant quartz and feldspar minerals. Earthquake nucleation is limited to crust cooler than the quartz brittle-plastic transition (bpt) temperature, - 300-350'C. However, fully plastic behavior does not occur until the feldspar bpt is reached, at - 4500C. Earthquakes cannot nucleate in the brittle-plastic transition zone between the quartz and feldspar bpts, but coseismic fault rupture can propagate there, driven by strain energy release at shallower depths in the seismogenic zone.

Analyses of the rupture characteristics of large earthquakes lend support to the synoptic shear zone model. For example, to model or explain the down-dip rupture extent of major decoupling earthquakes in subduction zones, both Tichelaar & Ruff (1993) and Hyndman & Wang (1995) incorporate a temperature-dependent transition zone, as proposed by Scholz (1988, 1990). Also, numerical analysis by Das (1983) predicts that the depths and magnitudes of rupture displacement in the plastic or semi-plastic transition zone will be significant for large earthquakes that rupture the full thickness of the fully brittle seismogenic zone. The depth extent of the transition zone will depend on the geothermal gradient; depth levels for the quartz and feldspar bpts for typical to slightly high geothermal gradients have been added to Figure 5a. In this interpretation, ruptures of events D1, J1, and F1 could extend down to depths exceeding 30 km and may have involved even the anomalous ultrarnafic crust of the Reelfoot rift. If rupture propagation to such depths can indeed occur, then model 3 above could explain the large seismic moments of the 1811-1812 sequence without the need to appeal to unknown active fault segments of the NMSZ.

6. PALEOSEISMOLOGY AND NEOTECTONICS

Was the New Madrid earthquake sequence of 1811-1812 a fluke of geological history, a one-time event? Or have large earthquakes occurred repeatedly in the NMSZ in the recent geological past? This question is important, not only for assessing the earthquake hazard of the New Madrid region, but also for understanding the long-term process of tectonic strain accumulation and release. To this point we have emphasized spatial relationships of structures within the NMS4 but because characterizing its temporal behavior is of equivalent importance, we now examine the existing evidence for previous seismic activity.

At first glance, the Mississippi River and its surrounding expanse of nearly level flood plains would seem to be evidence of a region that is tectonically and scismically dead. However, seismological, geodetic, and most paleoseismological data suggest a surprisingly short recurrence interval for major earthquakes in the NMS4 on the order of a thousand years or less, with deformation rates comparable to those at plate margins. If the largest of the 1811-1812 events each produced 8-10 m of slip (Johnston 1996c), then repeated earthquakes of such a magnitude would clearly result in major disruption of fluvial systems, as well as the landscape in general. Yet, other data, particularly the low regional relief, indicate that the rapid rates of crustal strain implied by such magnitudes and repeat times cannot have been maintained for geologically long periods of time (Schweig & Van Arsdale 1996).

One line of evidence for high strain rates and short recurrence intervals for large earthquakes is in the form of earthquake frequency-magnitude relationships. Johnston & Nava (1985) extrapolated the historical and instrumental record and determined that if a periodic seismic cycle is valid for the NMSZ, earthquakes of 1811-1812 magnitude should recur there every 550-1100 years on avenge, a repeat time frequently used in probabilistic seismic hazard analyses.

Liu et al (1992) reoccupied a 1950s triangulation network in the southern New Madrid seismic zone using the Global Positioning System. Their data indicate rapid crustal shear strain accumulation of the order of 10-7 /year, which results in 5-7 mm/year of right-lateral slip across the width of the network. At this rate of deformation, sufficient strain energy to produce an 1811-1812-type event could accumulate in 400-1100 years (Schweig & Ellis 1994).

Paleoseismological studies indicate similarly short recurrence intervals for earthquakes large enough to cause liquefaction or ground failure, although the magnitudes of prehistoric earthquakes have been difficult to assign. A variety of indicators of prehistoric seismicity are being used, including local and regional deformation, liquefaction, and dendrochronology.

Local Deformation: Reelfoot Scarp and Lake County Uplift

The Lake County uplift is a broad, low-amplitude anticline that lies within a left-stepping restraining bend in the NMSZ (Figure 3) (Russ 1982). Most of the current microseismicity in the central NMSZ underlies the Lake County uplift. This seismicity is attributed to strain release along a southwest-dipping reverse fault (RF) (see Section 4 and Figure 6) that underlies the Lake County uplift (Chiu et al 1992). As interpreted in Figure 6, RF reaches the surface at the base of the Reelfoot scarp, which also forms the western margin of Reelfoot Lake and part of the eastern margin of the uplift.

The Reelfoot scarp (Figure 2b) is locally 8 m high and has recently been mapped across the Mississippi River into Kentucky and Missouri (Van Arsdale et al 1995). The scarp is an east-facing monocline interpreted to be the eastern limb of a fault propagation fold (Kelson et al 1996, Van Arsdale et al 1994a). Trenches excavated across the scarp have revealed minor normal and reverse faults, assumed to be secondary faulting associated with surface folding. Detailed trench logs provide information on the geometry of scarp deformation and on the chronology of paleoseismic events (Kelson et al 1992, 1996; Russ 1979; Russ et al 1978). Trenching studies by Russ at the northern Reelfoot Lake site (near K in Figure 3), the pioneering paleoseismological work in the New Madrid region, resulted in the recognition of three faulting events on the Reelfoot scarp within the last 2000 years but was unable to date the two prehistoric events. Subsequently, Kelson et al (1992, 1996) presented additional trench evidence for an 1812 and two prehistoric faulting events (site K in Figures 3 and 8). Kelson et al (1996) estimate that the two prehistoric earthquakes occurred at approximately AD 900 and AD 1400. These results are consistent with Russ's work and for the first time directly link the scarp to deformation in the 1811-1812 earthquakes using geological evidence.

Liquefaction Studies

That there were prehistoric large earthquakes on the Reelfoot fault (RF) is not surprising; the associated scarp and the Lake County uplift are far too high to have been formed entirely in 1812. In the rest of the New Madrid seismic zone, however, other such fault-generated structures are not obvious. Sand blows, either surficial or buried, and the dikes of liquefied sand that feed them, are commonly well preserved in the geologic record of the New Madrid region (Saucier 1977). The potential of deposits of liquefied sand as indicators of prehistoric earthquakes has long been noted, but only in the past few years has the proper combination of investigators with geological, pedological, and perhaps most importantly, archeological skills come together to assess the ages of liquefaction deposits (Tuttle & Schweig 1996). In the New Madrid seismic zone, surficial sand blows are so large (commonly 1.0-1.5 m in thickness and 10-30 m in diameter) that they have been only slightly modified by plowing and are still easy to identify on aerial photographs and on the ground. Prominent liquefaction effects from the 1811-1812 earthquakes are present throughout the region of modern microseismicity (Figure 3; Obermeier 1989, Saucier 1977). Thus finding liquefaction is not a challenge, but finding the intruded sediments or liquefied sands in a datable context is difficult.

Saucier (1991) was the first to successfully use archeology to identify and date prehistoric liquefaction features in the NMSZ at the Towosahgy archeological site (S in Figures 3 and 8), located only 30 km northeast of New Madrid. He attributed these features to two pre- 1811 events, one between AD 539 and AD 911 and the other about 100 years before AD 539. Saucier (199 1) recognized dot die two events at Towosahgy could be the same prehistoric events found by Run (1979). Since Saucier's study, two additional paleoliquefaction sites have been found in the northern NMSZ (W and N on Figures 3 and 8). Site N is the northernmost paleoseismological site yet discovered in the region (Li et al 1994).

Most of the liquefaction evidence examined to date for recurrent earthquakes has been found in the southern New Madrid seismic zone, from the area between Blytheville, Arkansas, and Caruthersville (Little Prairie), Missouri; this is the locale of the Blytheville fault zone (BFZ) of Section 4 and Figure 2a where it adjoins the Blytheville arch (BA) and perhaps the nucleation zone of the D1 earthquake. At least five sites in this area contain confirmed prehistoric liquefw6on (sites within and just south of the Bootheel of

Missouri on Figure 3 and B2-86 on Figure 8). This area not only experienced intense liquefaction in 1811-1812 but has a rich and well-preserved archaeological history (e.g. Lafferty et al 1996, Morse & Morse 1983). The sites dated thus far are indicative of at least two liquefaction events that predate the 1811-1812 earthquakes. Their ages cluster around AD 800-1000 1996; Tuttle & Schweig 1995, 1996). There is also equivocal evidence for an earthquake between AD 1400 and AD 1600. The southernmost paleoliquefaction site is near the southwestern tip of the Blytheville arch (T on Figures 3 and 8).

Formation of Sunklands

An additional source of paleoseismological information is the "sunklands" (Fuller 1912) that lie above the northwestern flank of the Blytheville arch (Figure *2a*) and appear to be tectonic in origin. Fuller mapped numerous sunklands along the St. Francis and nearby rivers in Arkansas and Missouri. He believed that they formed during the New Madrid earthquakes. Indeed, several maps from the early- to mid- 1800s refer to some of the lakes associated with the sunklands by the name "Earthquake Lake." Two of the largest sunklands in northeastern Arkansas are Big Lake and Lake St. Francis (BLS and SFS on Figure 2a). King (1978) and Guccione et al (1993, 1994) cored the sediments in and adjacent to these lakes and conclude that they owe their present form to 1811-1812 deformation. Big Lake appears to have formed by a combination of subsidence and downstream uplift along the south-flowing Little River. Coring within Big Lake reveals two organic mats that apparently reflect 1811-1812 subsidence and a prehistoric subsidence event (Guccione et al 1993). Similarly,

Figure 8 Summary of paleoearthquake investigations in the New Madrid seismic zone. Vertical bars and lines show results from individual sites or studies, generally arranged from north on left to south on right (see Figure 3 for site locations). Thick horizontal line at top represents the 18111812 sequence. Horizontal shaded bands represent likely age ranges of earthquakes in the past 2000 years, with darker shading representing the more likely dates (see text). No implication of the magnitudes or areas affected by the earthquakes is inte (1994); S, Saucier (1991); K, Kel (1992, 1996); Sunklands, Guccione et al (1993), Guccione & Van Arsdale (1995); all others, Tuttle & Schweig (1996b), and Lafferty et al (1996). Sites Bl-B6 all cluster between Blytheville and Caruthersville (Figure 3).

Lake St. Francis formed because the St. Francis River was locally subsided and ponded by downstream uplift. At Lake St. Francis, however, there are core data to support four ponding events in the last 8000 years (Guccione et al 1994).

Another valuable paleoseismological tool in studying Reelfoot Lake and the other sunklands is dendrochronology. The dendrochronologic record of bald cypress tree rings in Reelfoot Lake dates back to AD 1677 (Stahle et al 1992). No pre- 1811-earthquake is supported by studies of this record. However, a tremendous post-1812 growth surge revealed in bald cypress tree rings in Reelfoot Lake does support the interpretation that Reelfoot Lake formed during 1812 coseismic uplift of the Lake County uplift and coincident ponding of the west-flowing Reel Foot River (Stahle et al 1992, Van Arsdale et al 1991). Additional dendrochronology at Reelfoot Lake h of t however, coring within the lake sediments has not revealed evidence for pre1811 lake formation (Valentine et al 1994).

Dendrochronologic studies of bald cypress trees at Big Lake and Lake St. Francis have been undertaken to assess the seismic histories of these areas (Van Arsdale et al 1994b). To date, no bald cypress that predate 1811 have been found at Big Lake, but numerous bald cypress at Lake St. Francis are older than 1811. These trees growth suppression between 1813 and 1840. Of particular interest is that the dendrochronologic record at Lake St. Francis extends to AD 1321 and that the only major tree ring growth anomaly is the 1813-1840 growth suppression (MK Cleaveland & DW Stahle, personal communication, 1995). This suggests that, if any large earthquakes did occur in the southern New Madrid seismic zone in the 490 years prior to 1811, they did not affect the bald cypress at Lake St. Francis. The opposing tree-ring growth patterns at Reelfoot (growth surge) and St. Francis (growth suppression), however, indicate that subsidence effects on bald cypress can be highly variable.

Summary of Paleoseismological Results

Figure 8 summarizes the published results of paleoseismology studies in the meizoseismal area of the 1811-1812 earthquakes. Not included is recent preliminary work indicating paleoearthquake liquefaction to the north and west of the seismic zone. The diagram shows the allowable ranges of earthquake events, and the ranges preferred by the individual investigators. The arguments for these preferences-archeological, pedological, and sedimentological --can be found in the individual references, and they vary in their persuasiveness. The simplest interpretation of these results is that in addition to the 1811-1812 events there were at least two strong ground-shaking earthquakes in the past 2000 years. Evidence for one of these, which likely occurred between AD 800 and AD 1000, is clear at Reelfoot scarp and north of New Madrid. There is also evidence for liquefaction of this age at the site T near Marked Tree, Arkansas (Figure 3), and possibly at site B6 near Blytheville, Arkansas- Evidence for a liquefaction-producing event between AD 1200 and AD 1400 is strong in the Blytheville area and may be present in the Reelfoot area and at the northernmost sites. Moreover, there is evidence in Figure 8, presently inconclusive, of liquefaction ages both younger than (AD 1600) and older than (prior to AD 600) these two age ranges.

Thus 1811 was not the first time in the Holocene that the New Madrid region experienced strong ground shaking. The data all are consistent with as few as two and as many as four earthquakes in the 2000 years prior to 1811. What were the magnitudes of the causative earthquakes? The only known post- 1812 earthquake in the New Madrid region large enough to have caused liquefaction is the M 6.6 1895 Charleston, Missouri, earthquake, which caused across (Obermeier 1988). Schweig & Ellis (1994) estimated that, when the severity of liquefaction is taken into account (Youd et al 1989), an M - 8 earthquake would be required for an event to have caused the liquefaction at sites S and the southern Blytheville sites, separated by about 100 km. If indeed an AD 900 earthquake is represented at the northernmost and southernmost sites (N and T, Figure 3), a multiple event scenario similar to the 1811-1812 sequence may be required.

7. FAULT RUPTURE SCENARIOS

The 1811-1812 New Madrid earthquake sequence has been described in numerous ways: by Mitchell (1815) in terms of a series of disconnected historical vignettes, by Fuller (1912) in terms of far-field intensities and near-field geomorphic effects, by Nuttli (1973) anf magnitude, and by Johnston (1996c) in terms of seismic moment. With the luxury of elapsed time, we now have the opportunity to integrate all these perspectives and attempt to delimit the faulting sequence that actually took place. The data at hand are still insufficient to uniquely specify the sequence and probably always will be so, but the information presented in this review does apply restrictions. In this section we identify those restrictions and use them to develop plausible faulting models of the 1811-1812 New Madrid earthquakes.

As if an earthquake triplet was not sufficiently complex, Street (1982) and Street & Nuttli (1984) demonstrated that some aftershocks of DI and FI were major (M $\geq$ 7) events in their own right (see Table 1), requiring the accommodation of nonnegligible fault areas. These authors, in fact, considered event D4 to be a fourth principal shock. Moreover, the Bootheel lineament investigations, beginning with Schweig &

Marple (1991), havsed the strong probability that fault segments not illuminated by present-day seismicity ruptured in the 1811-1812 sequence. The upshot of both these bodies of research is that a fault-rupture scenario for 1811-1812 must incorporate a minimum of seven fault segments (Figure 9) and six $\mathbf{M} \geq 7$ earthquakes (Table 1).

Our faulting scenario begins with the moment magnitudes of the 1811-1812 sequence listed in Table I * These yield a cumulative moment M_0 (equival.) = 3.6 x 10^{28} dyne-cm (M = 8.3). Whatever coseismic ruptures are adopted must sum to this cumulative M_0. In Section 5 we considered but rejected alternatives to this requirement; its adoption here means that ruptures extend into the crustal depth zone between the quartz and feldspar brittle-plastic transition temperatures (possibly 30 km or more). If then coseismic displacements are limited to ≤ 10 m to keep stress and strain drops reasonable, the fault lengths of the D1, Jl, and F1 principal events must be on the order of 110-150 km, 40-70 km, and 60-90 km, respectively.

Other general constraints are that coseismic ruptures must largely be contained within the intense liquefaction region described by Obermeier (1989)

(Figure 3) and must include all the fault segments illuminated by current seismicity. We do not adopt the more severe constraint that coseismic ruptures must be limited to these illuminated segments; hence the Bootheel lineament is not excluded from the rupture scenarios.

Figure 9 summarizes three possible fault-rupture scenarios for the 1811

1812 New Madrid sequence. Others are certainly possible; the ones highlighted here, however, mostly meet other constraints from the historical accounts and structural data previously examined. In rough priority of their importance and strength of evidence to support them, the historical constraints are:

1. The F1 rupture must include the Reelfoot fault (RF) segment.
2. The D1 rupture must include the Blytheville arch (BA) segment.
3. The Dl aftershocks were stronger than the D1 mainshock at Little Prairie.
4. Both the D1 and F1 principal events were larger than event J1.
5. The D1 and F1 aftershocks were major events, requiring tens of kilometers of fault length.

The candidate fault segments and their lengths are (Figure 9):
1. BA-Blytheville arch segment, axial fault, ~70 km;
2. BFZ-Blytheville fault zone segment, axial fault, ~ 55 km;
3. BL-northern Bootheel lineament, ~70 km;
4. NN-New Madrid north fault + seismicity continuation, - 60 km;
5. NW-New Madrid west seismicity trend, - 40 km;
6. RF-Reelfoot Fault, - 32 km; and
7. RS-Reelfoot south seismicity trend, - 35 km.

Figure 9 (a) **Fault segmentation of the NMSZ. Seismicity of the NMSZ, the Blytheville arch, and the Bootheel lineament/NN fault (left) yield the seven segments *(right)* identified as: BA, Blytheville arch; BFZ, Blytheville fault zone; BL, Bootheel lineament; NW, New Madrid west; NN, New Madrid north; RF, Reelfoot fault; RS, Reelfoot south. Segments NW and RS are defined solely from seismicity. (b) Possible fault rupture scenarios (S#l, S#2, S#3) for the 1811-1812 D1, J1, and F1 earthquake sequences, using the seven fault segments of** (a). **Based on historical and physical constraints (see text), the D1 principal event must rupture BA, and the F1 principal event must rupture RF in all scenarios. S#l is the favored scenario for reasons discussed in the text.**

5#3

Of the seven segment candidates, only one (RF) has structural expression at the surface. One other (BL) has secondary surface evidence in the form of aligned fissures and liquefaction features. Two segments, NW (strike slip) and RS (dip slip), are defined only by seismicity concentrations. At least five segments (BA, BFZ, BL, RF, NN) can be identified in the subsurface on seismic-reflection profiles, but their subsurface characteristics differ greatly. Some segments (RF, RS, BA) appear to lie wholly or partially within magmatic crustal intrusions. Two segments (NW, NN) are outside the Reelfoot graben, although probably still within its northwest margin zone.

Given the faulting constraints and candidate fault segments listed above, numerous faulting scenarios for 1811-1812 may be envisioned. We restrict our considerations to three (Figure 9), all of which satisfy the rupture size-fault length general constraint and all of the historical constraints (except that S#2 violates constraint number 3). In outline form for **M** ≥7, the scenarios are:

S#1: D1 on BA + BL. D4, D6 on BFZ
J1 on NN
F1 on RF + NW. F3 on RS;

S#2: D1 on BA + BFZ. D4, D6 on BL(?)
J1 on NN
F1 on RF + RS. F3 on NW; and

S#3: D1 on BA + BL. D4, D6 on BFZ
J1 on NW
FI on RF + RS. F3 on NN.

Because we intend S#I-S#3 to illustrate possibilities given the initial constraints of this section, we do not go into the pros and cons of each scenario except to point out several fault-mechanics reasons in favor of S#I. If event D1 ruptures BL, then the J1 event on NN becomes an almost on-strike extension of D1. If JI is dextral strike slip, then it and the D1 aftershocks on the BFZ would compressively load the left-step thrust fault RF for failure in the F1 event. No other scenario accommodates the logical sequence of loading fault segments and still satisfies the third historical constraint at Little Prairie.

8. CONCLUDING REMARKS AND A CONCLUDING SCENARIO

It is only by chance that the 1811-1812 New Madrid sequence, with recurrence time probably measured in centuries-to-millennia, coincided with the advancing western front of European settlement, a coincidence that (barely) places these events in the purview of historical analysis rather than paleoseismology. Contrast New Madrid with the most recent great Cascadia subduction zone earthquake of AD 1700, which occurred just 75 years prior to first European exploration of the northwest coast. Its size and characteristics-even its reality-were in the realm of northwest Indian legend or subject to the great uncertainties inevitable in paleoseismic analysis until K Satake (Satake et al 1996, Kerr 1995) identified and dated its tsunami in

Japanese historical records and was able to provide a fairly confident M ≥9 estimate based on wave amplitudes there.

In a similar vein we consider the most remarkable result to emerge from the past two decades of New Madrid research to be the conversion of many of the legends and myths (see Table 2) surrounding the 1811-1812 earthquake sequence to phenomena with a firm scientific basis. Consider the following.

- Extensive and intensive fissuring and sand/water fountaining. Confirmed by detailed mapping of Fuller (1912) initially, then by Obermeier (1988, 1989). Some of the features are huge with dike widths in meters and fissure lengths in kilometers.

- Creation of sunkland lakes, primarily Reelfoot in Tennessee; also St. Francis and Big Lake in Arkansas. Confirmed by dendrochronology (Stable et al 1992) and lakebed coring (Guccione et a] 1993, 1994).
-
- Waterfalls and barriers on the Mississippi River. Substantiated by seismic reflection profiling (Odum et al 1996, Woolery et al 1996) and neotectonic trenching and coring studies on Reelfoot scarp (Van Arsdale et al 1995).

- "Sunken" forests. An as-yet-undated example has been discovered in the Blytheville area, near the proposed D1 epicenter (MP Tuttle & ES Schweig, personal communication, 1995).

- Native American legends of previous large earthquakes (Fuller 1912). Substantial paleoliquefaction evidence now exists (see Section 6 and Figure 8) for at least two major pre- 1811 earthquakes in the New Madrid region.

To conclude our examination of this remarkable series of earthquakes on America's river frontier, we narrate what we believe happened in the winter of 1811 to 1812. This is the S#1 scenario of Section 7; its components range from confidently established to speculative. There are probably few earth scientists who will agree with all its aspects. It will be interesting to revisit this summary after a decade or two and see how well it stands the tests of time and additional research.

At 02:15 (local) on 16 December 1811, a mild (˚~ 45˚F) Indian Summer night, a great earthquake (D 1) nucleated on the axial fault of Reelfoot rift, near the intersection of the Blytheville arch (BA) and the Bootheel lineament (BL). It ruptured bilaterally, in dextral strike-slip motion, southwest along the arch and north-northeast along the less structurally developed BL for a total fault length of ~140 km. Average slip approached 10 meters. Seismic moment release exceeded 10^{28} dyne-cm, corresponding to about 8 x 10^{23} ergs of seismic strain energy release. Fissuring, fountaining, and other aspects of severe liquefaction of the saturated river flood plain were intense along the rupture length and up to about 50 km from it. The river towns Little Prairie, New Madrid, and Point Pleasant were shaken at MMI IX-XI. Intensities at the fourth Chickasaw bluff (future Memphis, Tennessee) reached at least MMI IX The seismic waves, principally L_g, were felt as far as 2000 km away (Quebec) and caused damage (MMI VII) over an area greater than 500,000 km^2. Major aftershocks (low **M** 7s) at 08:15 on 16 December and 12:00 on December 17 completed the rupture of the BA along the BFZ for ~ 60 km to the Mississippi River and produced the massive liquefaction that caused the abandonment of Little Prairie by its inhabitants. The Mississippi River was strongly affected by these earthquakes (but not to the degree of the F1 event). A large river tsunami or seiche was produced upriver in the New Madrid bend area, whereas downriver of Little Prairie, tremendous volumes of groundwater squeezed out by liquefaction drained into the river and caused a rapid rise in level and a much swifter current than normal. Certain reaches of the river from Little Prairie to the fourth bluffs were clogged with tree trunks, branches, and roots, some uprooted by vibration, liquefaction, and failing river banks and others brought from the riverbed to the surface by the intense, prolonged shaking or liquefaction of the riverbottom sediments.

For the next five weeks a vigorous aftershock sequence continued to shake the region. Extremely cold weather set in and froze the Ohio River so that by the third week of January 1812 there were few if any travelers on the Mississippi River. The D1 earthquake was very large, but it and its aftershocks released only ~ 50% of elastic strain energy stored in the Reelfoot rift crust. The mainshock rupture stopped not because it depleted all available strain energy but because structural barriers halted it-probably the termination of Blytheville arch to the south and possibly igneous intrusions to the north.

An additional ~16% of the available strain energy was released at ~ 09:00 on 23 January 1812 when the J1 principal shock ruptured the New Madrid north (NN) fault. Static strain from the D1 rupture had loaded NN with enough additional shear stress to induce failure. The right-lateral strike-slip rupture propagated northeastward on NN to near present-day Cario, Illinois-Charleston, Missouri. In New Madrid it was "as violent as the severest of the former ones" (Bryan 1848), but the reported far-field intensities nearly everywhere were distinctly less than those for D1 and Fl. The ~ 8 m of right-lateral strike slip in this **M** 7.8 earthquake formed a left stepover with the right-lateral D1 aftershock faulting on the BFZ, compressively loading the 32-km-long Reelfoot fault (RF) in the stepover zone.

The Ohio River icejam broke up at Louisville falls about the time of event JI [Nolte (1854) reports that earthquakes "loosened the ice"], and many boats that began the trip to New Orleans at the falls had reached New Madrid and tied up for the night of 6 February 1812. At 03:45 on 7 February the main dip-slip event of the entire sequence nucleated on the RF thrust fault plane in the left stepover that splays to the surface as Reelfoot scarp. The rupture was not contained by the RF, however, and continued onto the preexisting New Madrid west (NW) fault segment as a left-lateral strike-slip rupture. The NW segment developed perhaps as a deflection of the RF thrust around the dense, rigid Bloomfield pluton. Major aftershocks would extend the RF rupture to the southeast on the Reelfoot south (RS) segment, a separate, more steeply dipping fault plane. The FI sequence released the final third of the seismic strain energy available to drive the 1811-1812 earthquakes. The **M** 8.0 mainshock was probably a complex multiple event with both dip-slip and strike-slip subevents, averaging perhaps as much as 10 m displacement.

The RF thrust subevent of the F1 mainshock created one waterfall or rapids and two flow barriers on the Mississippi River's Kentucky bend; an additional falls may have formed on the bend's western limb by deformation in the hanging wall. The hanging wall of RF rose beneath the river during F1 from ~12 km to ~17 km upstream of New Madrid. This created an uplift that obstructed flow near island #10 and a downdrop falls or rapids downstream of the island. This combination was the most severe river disruption; it generated the great upstream wave and retrograde current so graphically described by Speed (1812) and the "patron" (Shatler 1815). Both travelers' flatboats would survive being swept over the downstream falls. The second intersection of RF with the river was immediately downstream of the town New Madrid (within 1 km). It uplifted the riverbed by one-to-several meters, accounting for the large wave and retrograde current at New Madrid, independently described by Bryan (1848) and Nolte (1854).

The riverbed from New Madrid to island #10 and the lakebed of to-be-formed Reelfoot Lake were on the footwall of a great thrust earthquake. Elastic rebound then accounts for their permanent subsidence by several meters relative to their pre-earthquake levels. Similarly, Reelfoot scarp's hanging wall was permanently uplifted. This combination of subsidence and uplift accelerated the town's takeover by the river and created the new lake.

Seven years later, on 16 June 1819, a small fort in remote western India would suffer a fate remarkably similar to New Madrid. Fort Sindree, built at sea level on a salt flat, was on the footwall of the **M** ~ 7.8 Kutch thrust earthquake, within several kilometers of its surface scarp, the *Allah Bund* or "wall of God." In stable continental regions, only New Madrid's D1 and F1 events were larger than Kutch. Oldham (1926) reports that the Allah Bund rose 6-7 m, the footwall dropped 3-4 m, and Fort Sindree was submerged to its turrets (see Figure 3-17 in Johnston et al 1994). Within the second decade of the nineteenth century a unique and common bond was established between a little town on the Mississippi River and a little fort on an Indian salt flat-a bond that none could have imagined beforehand.

The F1 principal event, with its coseismic faulting of the Mississippi riverbed on RF and large aftershocks on RS, was the culminating episode of the 1811-1812 New Madrid sequence. Evidently the huge reservoir of elastic strain energy was finally depleted. Just how and why and at what rate the tectonic strain accumulation took place-and is taking place-is currently unknown. Understanding this process will be the research frontier of the coming decade with global-position-system measurements and improved knowledge of crustal composition, structure, and rheology the principal weapons.

We have presented a faulting scenario for the 1811-1812 New Madrid earthquakes that is consistent with the available historical, geological, and geophysical evidence. This evidence - compiled primarily within the past two decades- to a large measure confirms the past anecdotal reports of the dramatic effects of the earthquakes on the land and the river of the central Mississippi Valley. The seismic moment release of this earthquake series probably equals or exceeds the total of the continental western United States in historic times. How this can be when the New Madrid seismic zone lies deep within the stable midplate crust of North America leads us back to Churchill's description of Russia: Despite all the research advances of the past several decades, much about the 1811-1812 New Madrid earthquakes remains a riddle wrapped in a
mystery inside an enigma.

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