bearing frontal appendages; (ii) apparent absence of other jointed appendages including antennae; (iii) large, stalked, dorsally set eyes; (iv) a mouth set ventrally, distant from the anterior margin; (v) a series of ventrally based, segmentally arranged, imbricating, paddle-shaped, movable lateral flaps; (vi) comb-like gills positioned between the flaps; (vii) two exsagittal rows of segmentally repeated ventral or internal structures of unknown function, preserved as black or light-reflective patches; (viii) a telson-like structure composed of two or three pairs of large flaps; and (ix) furcae, at least in some forms. *Kerygmachela* shares most of these characters, besides general similarities including overall appearance and inferred predatory mode. Features (i) through (ix) are regarded here as synapomorphies indicating a monophyletic origin for the forms. Some of the structures are considerably modified in *Opabinia*: There are five eyes rather than two, the lateral flaps usually imbricate in a reversed direction, and the single preoral appendage is apparently formed by fusion of the pair present in anomalocaridids (*14*, *28*)—an intermediate stage may be represented by *Kerygmachela*.

These differences, and, for example, the sclerotized mouth apparatus in anomalocaridids, indicate considerable evolution between the common ancestor and its descendants. The assignment (*24*) of *Kerygmachela* to the lobopods is regarded here as erroneous, and the concomitant suggestion of lobopod legs in *Opabinia* is incompatible with the Chengjiang anomalocaridid evidence, showing the fibrous nature of these structures [character (vii) above]. We therefore reject assignment (*24*) of these taxa to the Lobopodia. We regard the anomalocaridids, *Kerygmachela*, and *Opabinia* as representing a group of phylum-level rank, and propose an unnamed phylum-level taxon, defined by characters (i) through (ix).

Several features indicate affinities of the group to accepted arthropods: the presence of a tough exoskeleton, growth by moulting, true segmentation, comb-like gills, and pivot joints in the appendages. The (super)-phylum Arthropoda also embraces the groups from the dismantled (*29*) Uniramia, and evidently also the Onychophora, after recent molecular work (*30*). If a superphylum level is used for this revived arthropod concept, the group is a phylum, whereas if the phylum level is chosen, the group is a subphylum of the Arthropoda.

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Reconciling Short Recurrence Intervals with Minor Deformation in the New Madrid Seismic Zone

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At least three great earthquakes occurred in the New Madrid seismic zone in 1811 and 1812. Estimates of present-day strain rates suggest that such events may have a repeat time of 1000 years or less. Paleoseismological data also indicate that earthquakes large enough to cause soil liquefaction have occurred several times in the past 5000 years. However, pervasive crustal deformation expected from such a high frequency of large earthquakes is not observed. This suggests that the seismic zone is a young feature, possibly as young as several tens of thousands of years old and no more than a few million years old.

Over the past decade, conflicting evidence has been mounting regarding the recurrence intervals of large earthquakes in the New Madrid seismic zone of the Central United States, the site of at least three great earthquakes in 1811 and 1812. Seismological, geodetic, and some paleoseismological data suggest a relatively short recurrence interval, on the order of 1000 years or less, and deformation rates comparable with those at plate margins. Yet, other data indicate that these rapid strain rates cannot have been constant for geologically long periods of time.

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Seismological evidence for a short recurrence interval is in the form of earthquake frequency-magnitude relations. Johnston and Nava (*1*) analyzed the historical and instrumental record and determined that earthquakes of surface-wave magnitude, M_S, ≥ 8.3 should recur every 550 to 1100 years on average in the New Madrid seismic zone.

Liu *et al.* (*2*) reoccupied a 1950s triangulation network in the southern New Madrid seismic zone using the global positioning system (GPS). Their data indicate unexpectedly rapid crustal shear strain accumulation on the order of 10^{-7} per year, which results in 5 to 7 mm/year of right-lateral slip over the width of the network. At this rate of deformation, enough strain energy to produce an 1811–1812–type event could accumulate in 400 to 1100 years (*3*).

Paleoseismological studies indicate similarly short recurrence intervals for earthquakes large enough to cause liquefaction or ground failure (Figs. 1 and 2). In the 1970s, Russ *et al.* (*4*) excavated a trench across the Reelfoot scarp in northwestern Tennessee (site B in Figs. 1 and 2) and found evidence that two events strong enough to form surface faulting and liquefaction [body-wave magnitude, m_b, ≥5.5 (*5*)] occurred in the past 2000 years and before 1811. This suggests a maximum recurrence interval of about 900 years. A more recent trenching study across the scarp by Kelson *et al.* (*6*) supports Russ's results (site C). They found evidence for an earthquake that occurred between A.D. 1310 and A.D. 1540 as well as equivocal evidence for an event before A.D. 900.

Other evidence for paleoearthquakes comes from an archaeological site near East Prairie, Missouri (site A). At this site, about 35 km northeast of Reelfoot scarp, Saucier (*7*) attributed paleoliquefaction features to two pre-1811 events, one between A.D. 539 and A.D. 911 and the other about 100 years before A.D. 539.

Tuttle *et al.* (*8*) found liquefaction evidence at site D strongly suggestive of at least two pre-1811 earthquakes in the last 5000 years. Additionally, at site E, a pre-1811 liquefaction event is archaeologically constrained to be about 1000 years old, and there is some evidence for an older event of undetermined age.

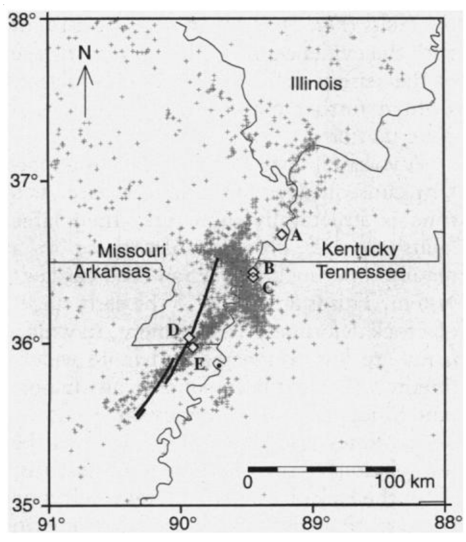

Fig. 1. New Madrid seismic zone. Diamonds at A, B, C, D, and E are the paleoseismology sites discussed in the text; the bold northeast-trending lines mark the Bootheel lineament; and seismicity from July 1974 to December 1991 is shown in the background by cross symbols.

Available age data (*4, 6–8*) are consistent with at least two pre-1811 events having occurred in the last 2000 years (Fig. 2). The youngest pre-1811 events at sites A and C, however, are not the same age (within one standard error), but both sites record an event predating A.D. 991. Thus, at least three pre-1811 events capable of inducing liquefaction have occurred in the past 2000 years.

Although these paleoseismological studies appear to support a recurrence interval of 1000 years or less, the magnitudes of the causative earthquakes are difficult to determine. The only known post-1812 earthquake in the New Madrid region large enough to have caused liquefaction is the 1895 Charleston, Missouri, earthquake of m_b = 6.2 or moment magnitude (**M**) of 6.8 (*9*), which caused liquefaction over an area 16 km across (*10*). Saucier used empirical relations between earthquake magnitude and maximum distance from the source of significant liquefaction (*11*) to conservatively estimate m_b = 6.2 as the minimum-magnitude earthquake that could be responsible for liquefaction at both sites A and B, separated by a distance of about 35 km. With this same method, if liquefaction from the same earthquake is being sampled at sites A and E, separated by about 100 km, a minimum magnitude would be about **M** = 6.9. However, the paleoliquefaction at these sites is quite intense; consideration of the liquefaction severity index (*12*) indicates a magnitude of at least **M** = 8.0.

Although seismological, geodetic, and paleoearthquake studies are consistent with a 1000-year recurrence interval for great earthquakes, other data argue against such a short interval or, alternatively, that such high rates must be a short-term phenomenon, perhaps due to postseismic relaxation from the 1811–1812 events (*13*). An **M** = 8.0 New Madrid earthquake can be expected to produce about 8 m of slip (*14*). For example, with a repeat time of 1000 years over a period of 20 million years, 20,000 earthquakes should have occurred with a cumulative slip of 160 km. Even if the slip is distributed across the late Precambrian to early Paleozoic Reelfoot rift, which bounds much of the New Madrid seismicity, such large deformation should have a significant topographic signature and should have left a clear record in the near-surface stratigraphic section. Such a record is not observed in the topography or the subsurface.

Dozens of high-resolution seismic reflection surveys have been completed in the past 3 years across areas of the New Madrid seismic zone considered most likely to display significant Quaternary deformation (*15*). None of these studies has shown the Quaternary section to be more than gently warped, and only minor post-Cretaceous faulting is evident. Even if all of the faulting is strike-slip, pervasive disruption of the section should be evident. Yet, the few seismic lines that extend outside the areas of predicted deformation show a relatively undisturbed Tertiary section. Thus, it cannot be argued that the large-magnitude slip is distributed across the rift, nor has a concentrated zone of large, young strike-slip faulting been found. In addition, aeromagnetic and gravity surveys suggest that several Paleozoic or older upper crustal features that cross the northern arm of New Madrid seismicity (Fig. 1) have cumulative offsets of less than 10 km (*16*).

The length of the New Madrid seismic zone, and hence the probable maximum length of any single fault, is about 250 km. Cumulative displacement-length ratios along geological (that is, finite displacement) faults are no more than about 10^{-1} (*17*). Thus, it is unlikely that there is more than about 25 km of cumulative displacement, and there is probably less than a total of 12 km of displacement, because none of the linear arms of seismicity (Fig. 1) is longer than 120 km.

The pattern of microseismicity also argues against the New Madrid seismic zone being a mature, well-developed fault system. It is likely that the zone represents a right-lateral strike-slip fault system with a restraining left step (*18, 19*). Such restraining steps are the types of discontinuities

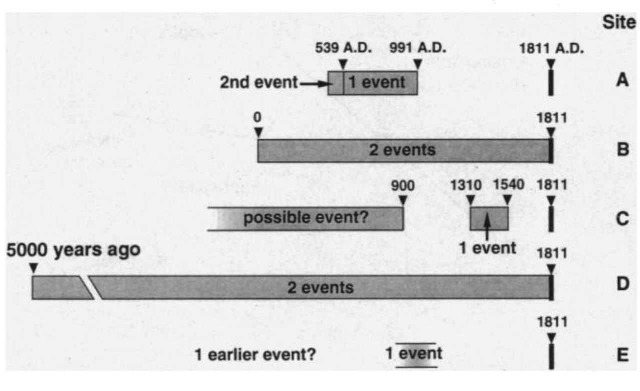

Fig. 2. Paleoseismology studies indicate at least three significant prehistoric earthquakes in the New Madrid seismic zone. All sites show evidence of 1811–1812 earthquakes, shown as a bold line on the right. Age increases to the left. Locations of the sites are shown in Fig. 1.

that most severely impede motion along strike-slip faults (*20*). The step on the New Madrid seismic zone is 33 km wide, which cannot be maintained with repeated slip over long periods of time and must be inherently unstable (*21*). Eventually, a linking fault would have to form between the two existing strike-slip faults to smooth the fault zone (Fig. 3) (*22*), a phenomenon commonly observed in analog models of strike-slip faults of low finite displacement (*23*). The Bootheel lineament of southeastern Missouri and northeastern Arkansas (*24*), a collection of en echelon fissures that apparently formed during the 1811–1812 events, may be such a linking fault. Results of recent seismic reflection studies have been interpreted to show the lineament to be underlain by a complex zone of strike-slip deformation consisting of multiple flower structures (*15*). Additionally, the distribution of seismicity is diffuse in comparison to well-developed strike-slip faults such as the San Andreas, which implies a relatively low total displacement (*25*).

Taken together, these observations argue that the total displacement across the New Madrid seismic zone may be as low as a few kilometers; total displacement is almost certainly less than 25 km. How do we reconcile these observations with the evidence for apparent high rates of strain throughout the Holocene?

There are at least three possible interpretations of the observations of short recurrence intervals and high strain rates but apparently youthful fault geometry and lack of major post-Cretaceous deformation. One is that the seismological and geodetic evidence has been misinterpreted and represents, for example, post-1812 strain relaxation and that the paleoseismological studies cited above reflect moderate local earthquakes. Liu *et al.* (*2*) considered this, but concluded that most of the relaxation occurred during the first two decades after the 1811–1812 earthquakes.

Work by Wesnousky and Leffler (*26*) supports the idea that the evidence for pre-1811 earthquakes represents local events. Wesnousky and Leffler did not find unequivocal evidence of pre-1811 earthquake-induced liquefaction in the southern New Madrid seismic zone in spite of reconnaissance of tens of kilometers of drainage ditches in an area with abundant 1811–1812 liquefaction. They concluded that earthquakes similar in size to the great New Madrid earthquakes probably have not occurred in this area during the 5000 to 10,000 years before 1811. Other studies also have failed to find evidence for pre-1811 liquefaction (*27*). None of these studies, however, have precluded the possibility that evidence for paleoearthquakes is not as widespread as evidence for the 1811–1812 earthquakes or that the evidence has been obscured by later earthquake deformation or post-earthquake soil-forming processes. And, in fact, more recent studies have uncovered evidence of paleoliquefaction in the southern New Madrid seismic zone (*8*). Also, geodetic work in the central New Madrid seismic zone (*28*) shows horizontal shear strains similar to those seen farther south (*2*) but not significantly different from zero. Although the former study does not support high strain rates, it does not preclude them.

A second possibility is that seismic activity in the New Madrid region is cyclic or irregular. The geological and geodetic observations that suggest relatively short recurrence intervals may, for example, reflect a time of high, but geologically temporary, pore-fluid pressure (*29*). Bollinger and Wheeler (*30*) suggested that seismic activity in the southern Appalachians may be controlled by the successive failure of a network of faults distributed uniformly through the stable crust. Failure of such a network may give rise to a cyclic or dual-poissonian type of distribution (*31*).

A third scenario is that the New Madrid seismic zone is a geologically young feature that has been active for only the last few tens of thousands to hundreds of thousands of years. This may be argued both from field observations and from a consideration of driving forces.

If we assume that displacement is 25 km across the New Madrid seismic zone and that the recurrence interval for an event with 8 m of displacement is 1000 years, then the seismic zone is no more than about 3.1 million years old. We emphasize that 25 km is a maximum displacement, so 3.1 million years is a maximum age. Observation of an irregular fault geometry associated with an unstable restraining step, a series of en echelon and discontinuous lineaments that may define the position of a youthful linking fault, and the general absence of significant near-surface faulting or topography indicate that the displacement is likely much smaller and the age much younger. The driving force for deformation may be derived from the motion of the North American (NAM) plate (*32*). It has been suggested that the motion of NAM, relative to the underlying mantle, changed through an angle of 60° at some time between 11 million and 3 million years ago (Ma) (*33*). Before and during the change in plate direction, the stress field resolved onto the New Madrid strike-slip faults would have been unsuitable for right-lateral rupture of the faults (Fig. 3). Though this is consistent with the evidence for a relatively young age of the seismic zone, it remains speculative without further constraints on the age of plate motion.

A youthful New Madrid zone has important consequences for seismic hazard. If a zone is structurally immature, then large faults are less ordered than those in a mature zone, such as the San Andreas fault system. This is analogous to the early stages of a rock deformation experiment, in which faults are less connected and more widely oriented. Thus, it becomes critically important to identify the relatively large seismic source zones (faults) because these will be less obvious than along mature systems. Also, the lack of a single, discrete fault zone suggests that the region is deforming in three dimensions. Rock deformation experiments suggest that three-dimensional strain may result in temporally variable stress fields (*34*) whose character may be important to recognizing which regional fault may be the next to sustain a large

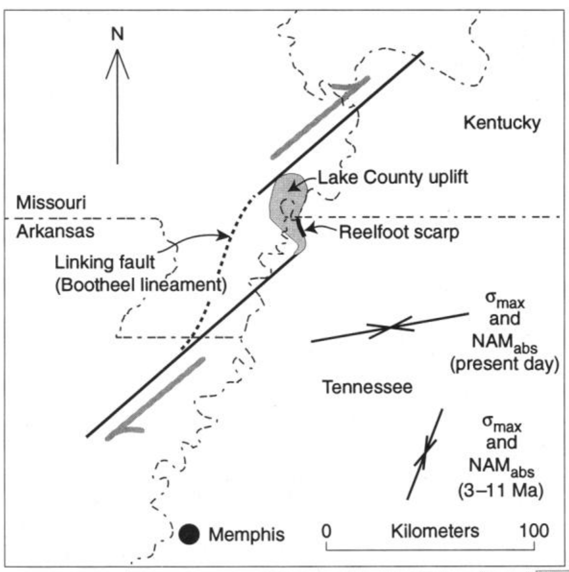

Fig. 3. Schematic view of faults inferred from seismicity and surface features in New Madrid seismic zone. The right-lateral sense of displacement across the main faults is supported by the existence of a region of active uplift in the left step-over (Lake County uplift), which currently has a structural relief of 10 m above the local floodplain. The southeastern edge of Lake County uplift is also marked by an active fault (Reelfoot scarp) that may also be the surface expression of a thrust fault inferred from recent microseismic studies (*35*). Arrows show plate motions (NAM_{abs}) and presumed maximum horizontal stress directions (σ_{max}) at New Madrid for the periods 0 to 3 Ma and 3 to 11 Ma (*33*).

earthquake. Therefore, it is important to maintain a quasi-continuous and dense measure of the regional strain field.

We conclude that active deformation in the New Madrid region may be as young as several tens of thousands of years old and certainly no more than a few millions of years old. This conclusion does not rely on the offset of any particular structure within the seismic zone. We thus reconcile the short recurrence interval with lack of deformation by suggesting that the New Madrid seismic zone is a relatively young feature.

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Mechanism of Catalytic Oxygenation of Alkanes by Halogenated Iron Porphyrins

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Halogenation of an iron porphyrin causes severe saddling of the macrocyclic structure and a large positive shift in the iron(III)/(II) redox couple. Although perhalogenated iron(II) porphyrins such as Fe(TFPPBr$_8$) [H$_2$TFPPBr$_8$, β-octabromo-tetrakis(pentafluorophenyl)-porphyrin] are relatively resistant to autoxidation, they rapidly reduce alkyl hydroperoxides. These and related reactivity studies suggest that catalysis of alkane oxygenation by Fe(TFPPBr$_8$)Cl occurs through a radical-chain mechanism in which the radicals are generated by oxidation and reduction of alkyl hydroperoxides.

Iron complexes of halogenated porphyrins, such as 2,3,7,8,12,13,17,18-octabromo-5,10,15,20-tetrakis(pentafluorophenyl)porphyrinato-iron(III) chloride [Fe(TFPPBr$_8$)-Cl], are remarkably active catalysts for hydroxylation of light alkanes by O_2, under mild conditions (25° to 60°C, 4 to 8 atm O_2) (*1*, *2*). Among the possible mechanisms considered, Lyons and Ellis have offered the suggestion that the active oxidant could be an iron-oxo intermediate, (TFPPBr$_8$)Fe^{IV}=O, produced through homolysis of the peroxobridged dimer formed by the reaction of Fe^{II}(TFPPBr$_8$) with O_2 (*1*). Although mechanistically related to cytochrome P-450 hydroxylations (*3*), this system has the advantage of not requiring a co-reductant. To assess the viability of direct generation of an active hydroxylating species from Fe^{II} and O_2, we examined the electrochemical properties, reactivities, and structures of Fe$^{III/II}$(TFPPBr$_8$) complexes.

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The spectroelectrochemistry of Fe-(TFPPBr$_8$)Cl (Fig. 1) shows an isosbestic (at 457 and 580 nm) transformation between Fe^{III} (402, 442, and 560 nm) and Fe^{II} (478 and 598 nm). The high Fe$^{III/II}$ reduction potential (0.31 V versus AgCl/Ag in 1 M KCl) suggests that Fe^{II}(TFPPBr$_8$) is strongly stabilized relative to other Fe^{II} porphyrins (*4*). Indeed, Fe^{II}(TFPPBr$_8$) is inert for many hours in the presence of an O_2 partial pres-