
27 Mar 87 Main Shock

- M ≥ 3.0
- 2.0 ≤ M < 3.0
- 0.0 ≤ M < 2.0
- M < 0.0

Ground Fissures Location

Strong Motion Instrument

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OF 10 JUNE 1987


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ABSTRACT

The June 10 southeastern Illinois earthquake was the 11th largest earthquake felt in the central U.S. during this century. Source parameters of the main shock were estimated from an analysis of surface-wave amplitude spectra. The source that best fit the observed data has focal depth of 10 ± 1 km; mechanism with strike = 40.6° ± 5.9°, dip = 76.2° ± 5.6°, slip = 159.7° ± 6.0°; tension and pressure axes of (T) trend = 357°, plunge = 24°, (P) trend = 89°, plunge = 4°; and a seismic moment of 3.1 * 10^26 dyne-cm.

With the combined efforts of six institutions, a 24-station analog microearthquake network was deployed around the main shock epicenter. One hundred eighty-five aftershocks were recorded in the first week of monitoring, providing 144 hypocenter determinations. A subset of 51 well recorded events was used for joint relocation and calculation of station corrections for the stations within 100 km of the main shock epicenter. Joint hypocenter-locations differ only slightly from the original locations. The spatial distribution of well located aftershocks indicates that the rupture was confined to a pencil-like zone within the Precambrian basement, extending from 7 to 11 km depth.

INTRODUCTION

The magnitude 4.9 (PDE) southeastern Illinois earthquake of 10 June 1987 was the largest event to occur in Illinois since the 5.5 m6 quake on 9 November 1968. The June 10 main shock was large enough to be felt in 21 states and southern Canada with maximum intensity VI in the epicentral area (PDE), and was the 11th largest event felt in the central U.S. during this century (Nuttli, 1983). It occurred less than 27 km from a permanent seismograph station of the Central Mississippi Valley Seismic Network operated by St. Louis University. Using local and regional data, a hypocenter was calculated by SLU at 38.71°N, 87.95°W, but had a poorly constrained focal depth of 5 km (Stauder et al., 1987). Using data from 12 short-period vertical records from WWSSN and Canadian Network stations, a mBtg = 5.2 magnitude was calculated for the main shock.

In order to understand the tectonic significance of the June 10 event, source parameters were estimated for the main shock. In addition, we present hypocentral relocations for 144 recorded aftershocks. The relocation work is presented in the second half of this paper.

SEISMIC HISTORY

The June 10 earthquake occurred in the Wabash River Valley of southern Illinois. This zone is noted for its moderate level of seismicity, and as such is usually defined to be a seismic source zone for ground motion hazard analysis (e.g. Barstow et al., 1981). During the past 30 years, the two largest earthquakes in the Midwest occurred within this zone, and not in the more active New Madrid Seismic Zone to the southwest. The historical seismicity from 1800 - 1974 for the region is plotted in Figure 1. No prominent trends are apparent at least partially because of the uncertain locations of many events in the nineteenth century.

Since 1974, a regional microearthquake network has been operated to the southwest by Saint Louis University. Instrumentally located seismicity in the Wabash Valley region is shown in Figure 9. The dense cluster of events at 38.71° N and 87.95° W is the 10 June earthquake sequence. The instrumental seismicity data suggest the existence of a zone of seismicity in southern Illinois subparallel to the Wabash River and 25 km to the northwest. Gordon (1988) also showed, in a JHD relocation of all earthquakes (mBtg ≥ 3.0) occurring between 1931 and 1980 in the Wabash zone, that the seismicity did not correlate with the area traversed by the Wabash Valley fault system, but was displaced to the northwest.

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Fig. 1. Historical seismicity in the region centered on the lower Wabash River Valley from Nuttall (1983). One hundred sixty-seven events plotted. Map symbol sizes are proportional to magnitude. The June 10 main shock is shown by the star.

**MAIN SHOCK PARAMETERS**

The focal mechanism for the main shock was determined by making combined use of P-wave first motions and surface-wave amplitude spectra. Sixty-one clear, impulsive P-wave first motions were obtained from local and regional stations. Incident angles were calculated by the computer program FASTHYPO (Herrmann, 1979a) using a source region specific modification of the SLU network UPLANDS earth model (Stauder et al., 1987). Because there can be a 180° ambiguity in focal mechanisms calculated from surface-wave amplitude spectra, the P-wave first motions were used to constrain the surface wave mechanism.

Vertical long-period Rayleigh-wave data in the 6 to 40 second period range were obtained from 19 stations: SCH, OTT, MNT, STJ, WES, SCP, PAL, BLA, SHA, JCT, LUB, BKS, COR, NEW, PNT, SES, EDM, FFC, and YKC. In addition, long-period Love-wave data in the 7 to 40 second period range were collected from 10 stations: STJ, WES, SCP, PAL, SHA, BKS, PNT, EDM, FFC, and YKC. The data set consisted of 225 Rayleigh-wave and 113 Love-wave spectral amplitude-period data points. Surface-wave focal mechanism analysis was performed using the techniques outlined by Herrmann (1979b). A similar analysis was performed by Herrmann et al. (1982) for the Sharsburg, Kentucky, earthquake of 1980. The spectra were corrected for anelastic attenuation and for geometrical spreading. Data from stations with epicentral distances ≥ 3500 km were not used to avoid a possible biasing of spectral amplitudes due to possible errors in the anelastic attenuation coefficients (e.g., Herrmann, 1979b; Herrmann, 1982).

A systematic search for the focal mechanism and seismic moment which best fit the observed amplitude-spectra radiation patterns was next performed. Identical searches were made over a range of focal depths between 2 and 20 kilometers. For each depth, theoretical Rayleigh- and Love-wave radiation patterns were calculated from a mechanism defined by a set of strike, dip and slip angles. The theoretical patterns were calculated using the Central U. S. Earth Model (Herrmann, 1986). A two-degree increment in the slip, dip and strike angles was used in the fine search.

For each mechanism, the theoretical Rayleigh- and Love-wave radiation patterns were compared to the observed spectral-amplitude data to obtain the correlation coefficients RR (for Rayleigh) and RL (for Love). In addition, two independent estimates of the seismic moment MR (for Rayleigh) and ML (for Love) were computed using the independent Rayleigh- and Love-wave data, respectively. The multiplicative product of RR, RL, and the ratio of ML/MR or MR/ML, (if ML > MR), gave an estimate of the goodness of fit for each mechanism. This product was largest for a source depth of 10 ± 1 km. The multiplicative product was also used as a weighting function to give a weighted average solution for each depth. For the 10 km source depth, the mechanism has nodal planes (1) strike = 40.6°± 5.9°, dip = 76.2°± 5.6°, slip = 159.7°± 6.0°; (2) strike = 135.6°± 5.3°, dip = 70.3°± 6.0°, slip = 14.6°± 5.7°. These give tension and pressure axes of (1) trend = 357°, plunge = 24°, (P) trend = 89°, plunge = 4°. The estimated seismic moment for this mechanism is 3.1 * 10^23 dyne·cm. A lower hemisphere equal-area plot of the focal mechanism for the weighted solution together with the observed P-wave first motions is shown in Figure 2.

The weighted average surface-wave solution is inconsistent with only 7% of the observed P-wave first motions. Langer et al. (1987), taking care to eliminate arrivals between 10 and 20 degrees, independently calculated a similar focal mechanism using only regional and teleseismic P-wave first motion data.

Figure 3 shows the degree of fit between the theoretical surface-wave radiation patterns and the observed data. These patterns were generated with a source step dislocation of 3.1 * 10^23 dyne·cm, the weighted average mechanism, and a source depth of 10 km. For this mechanism, RR equals 0.756 and RL equals 0.707. One can see that there is 180° symmetry in both the Rayleigh- and Love-wave radiation patterns. Even though the largest azimuthal gap in the Rayleigh-wave data is 80°, due to this symmetry the largest actual gap in the data is only 34°. Likewise, for the Love-wave data the largest azimuthal and largest actual gaps are 98° and 58°, respectively. This means that the Rayleigh-wave data sampled 81% of the pattern, while the Love-wave data sampled 68%.

A lower hemisphere equal-area plot of the focal mechanism for the weighted average solution is shown in Figure 4. The mechanism is plotted together with
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The time history of the sequence follows the "Omori's Law" power law decay of aftershocks with time, with the four largest aftershocks, \( M_B \approx 2.3 \) occurring at 1, 14, 17, and 284 hours after the main shock, respectively. Over 50 percent of the recorded aftershocks occurred within 48 hours of the main shock. This rapid decay emphasizes the importance for rapid deployment of portable networks following significant earthquakes in the eastern and central U.S.

Temporary Array

Immediately following the main shock, field teams from six institutions (St. Louis University, Indiana University, the Indiana Geological Survey, Memphis State University (CERI), Purdue University, and the U. S. Geological Survey) began preparations to deploy a variety of portable seismographs to record aftershocks of this event. The St. Louis and Indiana groups deployed a total of eight microearthquake recorders within 8 hours after the main shock. The five St. Louis stations were located 25 km to the northeast of the main shock epicenter to augment the permanent

Fig. 2. Focal mechanism for 10 June 1987, \( m_B = 4.9 \) event. The focal mechanism is the weighted average surface-wave solution. The northeast nodal plane has strike = 40.6° ± 5.9°, dip = 76.2° ± 5.6°, and slip = 159.7° ± 6.0°. The tension and pressure axes are (T) trend = 357°, plunge = 24°, (P) trend = 89°, plunge = 4°. Also plotted are the 61 P-wave first motions from local and regional stations. Compressional first motions shown as circles, dilatational ones as triangles. The plot is a lower hemisphere equal-area projection.

the planes for mechanisms ± 1 standard deviation in strike, dip, and slip angles away from the average solution. Using the most recent central U.S. scaling laws, for a source with \( M_0 = 3.1 \times 10^{23} \) dyne-cm, one would obtain a duration of 1.1 sec, a fault length = fault width of 2.3 km, stress drop of \( \Delta \sigma = 62 \) bars, and an average fault displacement of \( \bar{u} = 18 \) cm (Nuttli et al., 1989).

AFTERSHOCK SEQUENCE

The June 10 earthquake was followed by a notable aftershock sequence. The permanent Central Mississippi Valley Seismic Network, which includes nine stations in southern Indiana and Illinois, was augmented by a 24-station temporary network surrounding the main shock. The outstanding seismograph coverage made this aftershock sequence one of the best recorded in the eastern and central U.S. The temporary network recorded 185 events, ranging in magnitude from approximately -1.5 to 2.8. Of these events, 144 proved to be locatable, with ten having sufficient size to be detected and recorded digitally by the permanent network. The hypocentral parameters for all relocated aftershocks are listed in the Appendix.

Fig. 3. Theoretical surface-wave radiation patterns plotted with observed spectral amplitude-period values for 6 selected periods. Patterns calculated using a source depth of 10 km and the weighted average surface-wave focal mechanism. The correlation coefficients between the observed and theoretical are for Love (RL) = 0.787, for Rayleigh (RR) = 0.756. The Love- and Rayleigh-wave data are shown in the three upper and three lower plots, respectively. All patterns are scaled for a seismic moment of 3.1 \( \times 10^{23} \) dyne-cm. The scale bars show relative size for comparison between periods. The period, \( T \), of each pattern is given in seconds at the bottom of each plot.
Fig. 4. Focal mechanism calculated from the weighted average surface-wave solution with mechanisms for ± 1 standard deviation in strike, dip, and slip angles away from the average solution.

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arrivals.

Carver et al. (1987) pointed out a further complication in the arrival time picks. Their analysis of the digital recordings showed that the vertical seismograms contained what they interpreted as a P to SV converted phase. This phase, which could easily be misidentified as S on the vertical analog recorder seismograms, precedes the true S phase by a few tenths of a second on most records. S-wave picks made from the analog records are thus rendered considerably less reliable due to possible misidentification. Unfortunately, the depths of most of the events from the sequence are poorly constrained without S phases. To try to compensate for this problem, we consistently picked the larger S phase following the intermediate converted phase. In addition, all S phases were assigned less weight to minimize the impact of potential errors in picking S. This gives a solution consistent with a potential error in the picks of a few tenths of a second. Nonetheless, the reader should recognize that earthquake depths estimated for the aftershock data could be subject to significant errors related to the possible misidentification of the S arrival.

Estimation of Station Corrections

To better constrain the hypocenters of the aftershocks, we performed a joint location of selected events. Of the 144 locatable aftershocks, we selected a subset of 51 well recorded earthquakes (Figure 5a) with a total of 862 P-arrival measurements. Using the "Separated Earthquake Location" (SEL) procedure described by Pavlis and Hokanson (1985), P-wave station corrections were calculated for the 51 stations located within 100 km of the main shock epicenter. The SEL technique is founded on Jordan and Sverdrup's (1981) hypocentral decomposition theorem. This theorem states that joint location of a small group of earthquakes is fundamentally ambiguous because the group as a whole is subject to a bias that is impossible to determine from the arrival time data alone. This bias is present, though not explicitly accounted for, in all joint hypocenter location methods (e.g., JHD of Douglas, 1967). The unique feature of the SEL method is that it allows one to independently specify the bias term (i.e., the term that controls the position of a group of events as a whole), and extract from the data only the part of the solution that is actually constrained by the data (i.e., the relative position of events within a group).

In our case we have little a priori information about the bias term. Therefore, we chose to derive this term in an indirect way from the data. That is, we initially located all 51 events using both P and S wave picks with no station corrections. We fixed the locations of all 51 events at these positions and calculated average P wave residuals for each station. These averages were then used as what Pavlis and Hokanson call a "reference solution" and as a set of starting station corrections. This reference solution applies a "bias" that keeps the group near the initial location, and permits the SEL program to simultaneously solve for relative locations of events and a set of refined station corrections. The final SEL locations for the selected aftershocks are shown in Figure 5b. These final loca-
Fig. 5. Relocation of selected aftershocks by the Separated Earthquake Location (SEL) method. (A) Preliminary single-event locations used as input data for the SEL program. Open circles show preliminary locations of all aftershocks; filled circles indicate subset selected for use with SEL. (B) Final location estimates of selected subset from SEL. Distances and depths are in km along an east-west profile. Projection is looking north.

Locations differ only slightly from the original locations in spite of a 33 percent reduction in average RMS residual for the group. Much of that reduction resulted from application of station corrections to the recording stations. The joint locations fit the arrival time data to within measurement precision, suggesting that the remaining scatter in the data is the result of inaccuracies in timing and picking of arrival time data. One minor drawback of the Pavlis and Hokanson technique is that the program currently only estimates station corrections for P waves. We therefore derived the S wave corrections for each station by multiplying the P delays by an average Vp/Vs ratio (1.738) determined for the whole aftershock data set.

Final Locations

Earthquake locations were obtained using the single-event location program HYPOELIPSE (Lahr, 1984), using both P and S arrival times. We chose the modified UPLANDS crustal model that has been used for permanent network locations in the area (Stauder et al., 1987). This earth model consists of a four-layer crust over a two-layer upper mantle. Because of the size and geometry of the temporary network, the choice of velocity model is not especially influential in controlling the final aftershock locations. The station corrections derived from the joint location program (obtained using the same velocity model) were used for the single-event locations. We used a distance weighting scheme that linearly reduced the weight of arrival times from full weight (at 30 km distance) to zero weight (at 200 km distance). We also applied a truncation scheme that removed arrival time data whose residual exceeded 2.0 sec. This removes only the extreme outliers. Smaller errors can be difficult to detect with small networks, and this approach assumes that such errors are accounted for through appropriate grading criteria.

The final locations were graded into four categories of relative location accuracy. Although the divisions between categories are arbitrary, they provide a means to discriminate the best constrained locations for evaluation of the spatial distribution of aftershocks. The quality of the aftershock data permitted application of extremely stringent grading criteria. For instance, in order for earthquake locations to be given "A" quality, they required 15 arrival time readings, including 6 S-arrivals, an RMS residual less than 0.2 sec, an azimuthal gap less than 90 degrees, a minimum epicenter-to-station distance less than 5 km, and the maximum estimated error ellipsoid axis (usually vertical) less than 2.0 km. The average RMS residual for the final locations is 0.10 sec, suggesting that we have succeeded in matching theoretical and observed arrival time data for the entire aftershock data set to within measurement precision estimated for each arrival time pick.
The final earthquake locations are presented in map view in Figure 6 and in cross section in Figure 7. The best constrained events (shaded and filled circles in Figures 6 and 7) fell into an extremely tight cluster, approximately 1 km in diameter, offset from the main shock epicenter (probably due to main shock mislocation) by about 1.5 km. The poorly constrained locations (crosses in Figures 6 and 7) were distributed more diffusely about the cluster, which suggests that this scatter is largely a result of location errors. No "A" or "B" quality hypocenters were located more than 2.5 km from the centroid of the cluster.

In cross section (Figure 7) the well located events appear to form a narrow pencil-like zone, extending from about 7 to 11 km in depth. The elongate, vertical aftershock distribution is unlikely to be an artifact of the misidentification of the S phase discussed above. The multiple event relocations of the selected subset (Figure 5b), using P-arrival data only, show the same pattern. These locations are undisturbed by the possible misidentification of the S phase times determined from the analog records. On the other hand, the apparent length of this zone is due at least in part to larger errors in hypocentral depth (vertical error ellipsoid axes are typically 3-5 times larger than horizontal axes). Nonetheless, the best located events (filled circles) suggest a depth range of at least 4 km. For the "A" grade events, the largest axis of the error ellipsoid (typically near-vertical) averaged ± 1.68 km. A more conservative estimate of the 95 percent confidence interval of the hypocentral depth was determined by the technique described by Rowlett and Forsyth (1984). This statistical approach suggests that the vertical errors may exceed ± 4 km. If this is the case, it is feasible that the vertical appearance of the aftershock zone may be an artifact of the large vertical errors in hypocentral determination, and the actual source area of the aftershocks could be much smaller than it appears.

Using either approach to estimate the error in hypocentral depth, the overall depth range of the aftershocks shown in Figure 8 clearly indicates that the
rupture took place within the Precambrian basement, 4-6 km below the Phanerozoic sediment cover (Hamburger and Rupp, 1988). The depth range for this sequence is typical for intraplate earthquakes (Chen and Molnar, 1983), but appears to be considerably shallower than the larger ($m_s = 5.5$) 1968 southern Illinois earthquake (Stauder and Nuttil, 1970; Herrmann, 1973; Chen and Molnar, 1983).

DISCUSSION

Since 1838, thirty-eight earthquakes with MM Intensities $\geq$ IV have occurred within 100 km of the June 10 epicenter (Nuttil, 1983). Nine of these have occurred during the last 30 years, when instrumental locations were possible. Focal mechanisms, using surface-wave techniques, have been calculated for only four of the nine events. These mechanisms are listed in Table 1, and are plotted in Figure 9 with instrumentally located seismicity from 1974 to 1987 for the region.

Street (1976) and Street et al. (1974) have reported additional mechanisms in this region obtained from first- and second-phase $P$ arrivals, however, Herrmann and Canas (1978) argued that mechanisms obtained from second-phase arrivals could have large errors because the secondary arrivals are not refracted arrivals but rather supercritically reflected, and would have undergone a phase change. For this reason, Street's mechanisms were not used in this study.
Comparing the mechanisms listed in Table 1 and plotted in Figure 9, one can see that the mechanism of the June 10 main shock is very similar to one reported for a 3 April 1974 event located 16 km to the southwest (Hermann, 1979b). Both events are strike-slip with a small component of dip-slip. These mechanisms contrast with that of the larger (mb = 5.5) 9 November 1968 earthquake located 100 km to the southwest at 22 km depth. That event is reverse dip-slip (Stauder and Nuttli, 1970; Hermann, 1973). All four mechanisms have pressure axes oriented east-west, but the tension axis for the November 9 event is nearly-vertical, while the other mechanisms have tension axes nearly horizontal, in a north-south orientation.

Unfortunately, the aftershock distribution provides little insight into the source mechanism of the June 10 event. The epicentral map (Figure 6) does not suggest lineations along either of the main shock nodal plane directions. The cross sectional views (Figure 7) are suggestive of a near-vertical source zone, with a range in depth of at least 2-4 km. This distribution suggests that the earthquake took place along a narrow, vertically inclined fault or fault intersection located within the Precambrian basement rocks.

Two hypotheses have been suggested to explain the tectonic significance of this sequence. Hamburger and Rupp (1988) have argued that the sequence could be associated with the La Salle Anticlinal Belt, a possible fault-cored fold structure extending to the northwest from the main shock epicenter. An alternative interpretation is that the sequence could instead be related to the reactivation of northeasterly trending basement graben faults such as those seen by Sexton et al. (1986) in seismic reflection and potential field profiles 50 km to the south of the main shock epicenter. Even deeper basement faults have been reported in more recent reflection profiles collected 30 km south of the 10 June sequence. Pratt et al. (1989) interpret the large offsets in the sub-basement to be faulted Proterozoic layered sequences. These offsets are at 10 km depth and approximately 45 km southwest of the main shock, along the same trend as the April 1974 event.

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