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A Fluid-Injection-Triggered Earthquake Sequence in Ashtabula, Ohio: Implications for Seismogenesis in Stable Continental Regions

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Abstract A persistent earthquake sequence in northeast Ohio includes many distinct fore–main–aftershock subsequences, illuminates two faults, and was triggered by fluid injection. The first known earthquake from within 30 km of Ashtabula was an M_{blg} 3.8 mainshock that shook the downtown area in 1987. Seismicity has continued at an average of about one felt event per year. The largest magnitude so far, M_{blg} 4.3, caused slight damage (modified Mercalli intensity VI) on 26 January 2001. The latest subsequence started in July 2003 with an M_{blg} 2.6 event. Accurate hypocenters and focal mechanisms are available from three local seismograph deployments in 1987, 2001, 2003 and from regional broadband seismograms. These hypocenters are in the Precambrian basement, 0–2 km below the 1.8-km-deep Paleozoic unconformity, and illuminate two distinct planar east–west–striking sources zones 4 km apart, one in 1987 about 1.5 km long, the other in 2001 and 2003 about 5 km long. We interpret them as steep subparallel faults slipping left laterally in the current regime. Like many of the faults that ruptured in hazardous stable continental region (SCR) earthquakes, these faults were previously unknown and probably have small post-Precambrian displacements. The 1987 source was active a year after onset of class 1 fluid injection only 0.7 km north of the fault. The second fault, 5 km south of the injection well, became active in 2000, while the 1987 source was inactive. The well injected about 164 m^3/day of waste fluid into the 1.8-km-deep basal sandstone with about 100 bars of wellhead pressure from May 1986 to June 1994. An annular high pore-pressure anomaly is expected to expand along this hydraulically confined horizon at the top of the basement, even after injection ends and pressure drops near the well. Over 16 years, seismicity has shifted southward from $\leq$1 to 5–8 km from the point of injection. It seems to initiate when and where a significant pore-pressure rise intersects pre-existing faults close to failure and to be turned off when pressure starts dropping back. The largest earthquakes postdated the end of injection at both Ashtabula and at the Rocky Mountain Arsenal near Denver, Colorado. Anthropogenic earthquake hazard may thus persist after the causative activity has ceased but can generally be closely monitored. High-stress and low-strain rates in SCRs can account for a larger proportion of triggered earthquakes in the eastern United States and other SCRs than in active regions.

Introduction

The town of Ashtabula, along the southern shore of Lake Erie (Figs. 1 and 2), has experienced eight distinct episodes of felt earthquakes since 1987. The largest event occurred in 2001 and caused light damage. When contrasted with the lack of prior known seismicity in that area of northeastern Ohio, the proximity in time and space of this seismicity to an injection well was strong evidence that increased pore pressure triggered the seismicity (Nicholson and Wesson, 1990; Seeber and Armbruster, 1993). Accurate hypocenters illuminate two subparallel faults, one active in 1987, the other in 2001. Temporal coverage by local seismic stations is sorely incomplete, but available data are consistent with progressive migration of seismicity away from the well. When viewed against the low-seismicity background (Fig. 1), this seismicity appears intense and tightly confined and can be reasonably assigned to a single sequence of causally related earthquakes. The well-known precedent of earthquakes triggered by underground fluid injection in Denver,

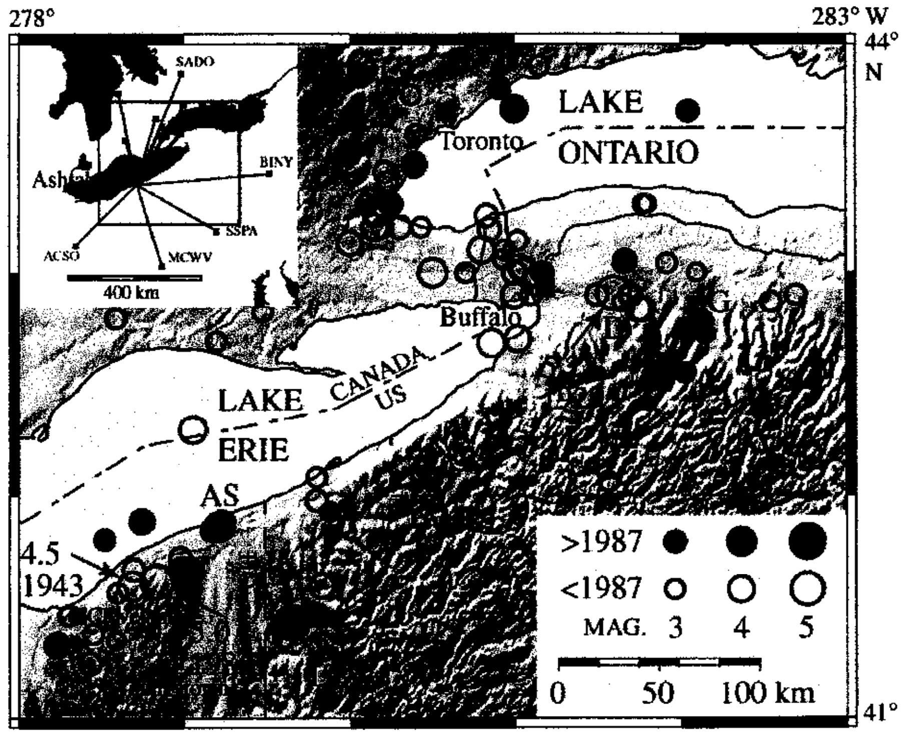

Figure 1. Regional seismicity in the Lake Erie–Lake Ontario region (Seeber and Armbruster, 1991). Filled symbols are for epicenters after the onset of seismicity in Ashtabula (AS) in 1987. Other known sources of seismicity triggered by engineering activities are near D, Dale; AV, Avoca; and G, Genesseo. The 1929 *M* 5.2 and the 1986 *M* 5.0 earthquakes may also have been triggered by humans (Nicholson and Wesson, 1990; Seeber and Armbruster, 1993). The inset shows stations used in the relocation of the 26 January 2001 mainshock.

Colorado, lends support to this hypothesis. Seismicity triggered by that injection continued at least 6 years and produced the third $M \geq 5$ earthquake 21 months after the end of injection (Healy *et al.*, 1968).

The sequence in Ashtabula is still ongoing as of July 2003 and has immediate implications for hazard in that metropolitan area of 50,000 people. It is one of many examples of seismicity triggered by engineering activities in stable continental regions (SCRs) (e.g., Fig. 1) and underscores characteristics of this kind of seismicity that makes it particularly dangerous: shallow and close to a human population (McGarr *et al.*, 2002). Perturbations may persist and may continue to evolve after anthropogenic forcing has ceased, particularly ones that involve pore-pressure gradients and interstitial fluid flow (Healy *et al.*, 1968; Seeber *et al.*, 1998; McGarr *et al.*, 2002). As much as 10^6 m^3 of fluid may be injected into a waste disposal well over several years. The resulting high pore-pressure anomaly can be expected to spread and may reach 10 km or more from the well, depending on hydrologic and geometric parameters (Hsieh and Bredehoeft, 1981; Nicholson *et al.*, 1988; Nicholson and Wesson, 1990). After the end of injection, fluid pressure will drop near the well, but the anomaly will continue to spread, causing the pressure to rise progressively further from the well. The amplitude of this perturbation and the potential for triggered earthquakes is expected to decay, but it could remain significant for a long time. Multiyear delays between external forcing and triggered earthquakes have been documented for natural causes (e.g., Seeber and Armbruster, 2000) as well as for different types of anthropogenic causes (McGarr *et al.*, 2002). One reason the spreading pore-pressure anomaly can remain significant to large distances and for long times is that stress changes as small as 0.1 bar are sufficient to trigger or inhibit earthquakes (Reasenberg and Simpson, 1992). This work adds to previous conclusions that human forcing of mechanical and hydrologic conditions in the upper crust may have long-lasting undesirable effects. These perturbations, therefore, are analogous to other types of pollution (Seeber, 2002b).

The main purpose of this work is to trace the development of the earthquake sequence in Ashtabula, Ohio, and to characterize the source faults using high-resolution hypocenters and focal mechanisms. The results provide the basis for a detailed comparison of the space-time history of the earthquake sequence with fluid injection at the disposal well.

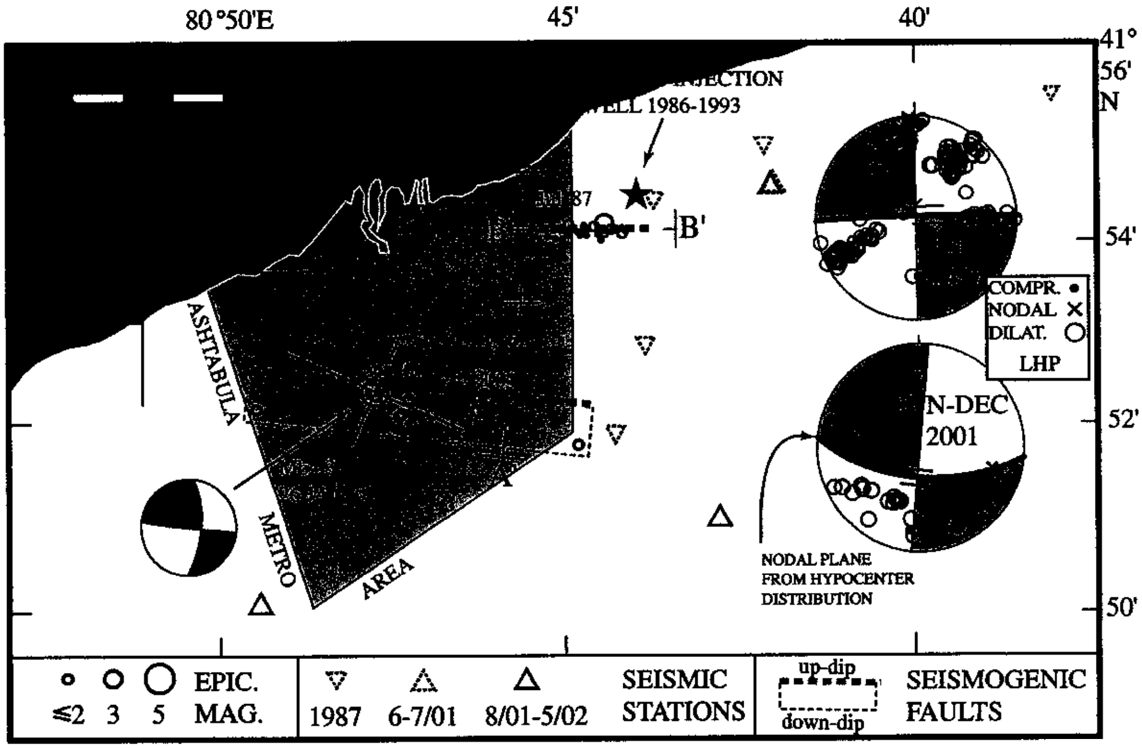

Figure 2. Accurate hypocenters and first motions in Ashtabula, Ohio, from two short-term deployments of portable seismographs. Data from 1987 illuminated a vertical east–west-striking left-lateral fault in the basement (Seeber and Armbruster, 1993). This activity was 0.7–2.0 km from a waste-disposal well (star) and started 1 year after the onset of injection. Several episodes of felt earthquakes during the following years were not monitored by local instruments. An M_{blg} 4.3 mainshock on 26 January 2001 caused light damage (MMI VI). The focal mechanism (Du *et al.*, 2003) and epicenter of this event were obtained from regional waveforms. Another fore–main–aftershock subsequence during June 2001 was captured with a local network. These data (see also Fig. 3) illuminate another fault (thick line is fault trace at unconformity) similar to the one in 1987, but 4 km south. The January mainshock is probably also from this source. The two dotted first motions are from the latest and westernmost hypocenter and are inconsistent with the composite focal mechanism.

Earthquakes in Ashtabula, Northeast Ohio

Seismotectonic Setting

Ashtabula is on the North American platform. Subhorizontal Paleozoic strata attest to the long-term tectonic stability of this continental region (SCR). These strata overlie the 1-billion-year-old Grenville basement with a sharp unconformity, which is 1.8 km deep in Ashtabula. Some faults in the Paleozoic section have been reported by gas recovery operators in the Astabula area and are being exploited as permeability anomalies. These faults are small, with displacements typically less than 10 m, and are of unknown postdepositional age (M. Hansen, personal comm., 2001). Their significance in terms of current tectonics and seismogenesis is unknown. No evidence has so far been reported in the Paleozoic platform rocks of the basement faults illuminated by hypocenters. Ashtabula is also in an area of low historic seismicity (Fig. 1).

Onset of Seismicity

In 1987 Ashtabula experienced an M 3.8 (U.S. Geological Survey Monthly List, Preliminary Determination of Epicenters, 1987) event. Using aftershock hypocenters from a 2-week deployment of analog seismographs, Seeber and Armbruster (1993) showed that the seismicity originated in the Precambrian basement just below the 1.8-km-thick platform rocks. All of the 36 reliable 1987 hypocenters (Table 1) were located on a ≈ 1.5- by 1.3-km planar vertical patch whose thickness of 1/4 km probably reflected the resolution limit of the locations. When combined with first motions, these hypocenters delineated an active, east–west-striking, vertical fault (Figs. 2 and 3). Left-lateral motion on this fault was consistent with the regional northeast–east-northeast direction of the maximum horizontal stress (Zoback and Zoback, 1989). After the relatively small mainshock in 1987, activity decreased rapidly, but it did not end. Earthquakes in

Table 1
Accurate Hypocenters Determined from Local Stations in Ashtabula

Date	Time		Latitude		Longitude		Depth	M_{bLg}	N	Gap	MD	RMS
(yymmdd)	(hhmm)	(sec)	(Deg)	(Min)	(Deg)	(Min)	(km)			(deg)	(km)	(sec)
870716	1143	7.45	41	54.06	80	44.44	2.00	+1.5	6	151	2.3	0.02
870716	1836	8.84	41	54.11	80	44.54	2.65	+1.1	6	156	2.5	0.01
870717	406	1.23	41	54.06	80	44.81	2.67	−0.6	6	155	2.5	0.02
870717	406	10.89	41	54.06	80	44.68	3.04	−0.5	6	154	2.5	0.02
870717	406	20.85	41	54.05	80	44.74	2.58	−0.7	6	154	2.5	0.02
870717	633	25.07	41	54.09	80	44.31	3.08	−1.1	6	152	2.4	0.03
870717	648	48.40	41	54.05	80	44.37	3.05	−0.9	6	150	2.3	0.02
870717	923	42.43	41	54.05	80	44.44	2.88	−0.9	6	150	2.3	0.01
870717	949	35.29	41	54.05	80	44.52	2.99	−0.7	6	152	2.4	0.02
870717	2354	15.94	41	54.12	80	44.54	2.50	−0.9	6	156	2.5	0.02
870718	158	46.15	41	54.05	80	44.72	2.65	−1.1	6	154	2.5	0.02
870718	626	15.18	41	54.11	80	44.74	2.21	−0.9	6	157	2.6	0.02
870719	602	39.46	41	54.00	80	45.17	2.30	−0.9	6	153	2.1	0.02
870719	1540	0.10	41	54.04	80	44.47	2.56	+0.8	6	150	2.3	0.01
870719	2124	14.94	41	54.16	80	44.68	2.00	−0.7	6	160	2.6	0.03
870719	2137	22.36	41	53.97	80	44.50	2.72	−1.1	6	147	2.2	0.02
870719	2138	47.72	41	54.03	80	44.57	2.42	−0.5	6	151	2.4	0.02
870719	2224	0.01	41	54.09	80	44.51	2.19	−0.7	6	154	2.4	0.02
870721	939	32.62	41	54.02	80	45.26	2.53	−0.6	6	156	2.0	0.01
870723	629	44.63	41	54.09	80	44.63	2.23	−0.9	7	154	1.4	0.01
870723	827	41.12	41	54.14	80	44.55	2.48	−0.9	7	158	1.2	0.02
870723	1846	59.99	41	54.13	80	44.62	2.34	−0.6	7	158	1.4	0.02
870725	2225	51.05	41	54.10	80	44.74	2.16	−1.1	7	157	1.6	0.02
870726	305	12.70	41	54.04	80	44.97	2.19	+0.1	7	155	1.9	0.02
870726	530	35.71	41	54.06	80	44.94	2.25	−0.2	7	156	1.9	0.02
870726	711	6.44	41	54.04	80	44.73	1.84	−0.6	7	151	1.6	0.02
890805	0054	46.96	41	54.06	80	45.05	2.41	+1.0	7	158	2.0	0.02
010603	942	8.18	41	52.29	80	47.29	2.02	+1.7	8	222	2.5	0.004
010603	2236	46.98	41	52.19	80	46.24	2.26	+3.0	8	253	1.5	0.002
010603	2247	45.58	41	52.17	80	46.17	2.21	+1.9	8	257	1.5	0.002
010604	147	47.84	41	52.17	80	46.22	2.11	+1.8	6	255	1.6	0.001
010604	643	59.15	41	52.16	80	46.34	2.16	+0.9	6	251	1.7	0.002
010605	827	16.02	41	52.18	80	46.14	2.17	+2.4	8	258	1.5	0.004
010606	346	35.41	41	52.17	80	46.09	2.18	+1.3	5	261	1.5	0.003
010618	558	44.91	41	52.10	80	47.43	2.27	+1.1	6	233	2.8	0.004
010619	1009	56.80	41	52.13	80	47.25	2.50	+1.5	8	180	2.6	0.004
010626	2031	58.49	41	52.34	80	46.02	2.92	—	6	226	2.8	0.052*
010806	2333	44.95	41	51.79	80	44.85	3.15	+1.5	7	169	2.4	0.008
010822	1218	57.49	41	52.13	80	47.39	2.62	+0.7	6	231	2.8	0.003
010923	446	19.03	41	52.12	80	46.94	1.75	+0.5	5	247	2.3	0.007
011130	1830	57.50	41	52.15	80	49.52	3.21	+1.7	6	279	3.2	0.011
030722	1554	01.24	41	52.32	80	46.17	2.53	+2.1	8	111	1.8	0.007
010126	0303	21.50	41	52.32	80	47.76	3.75	4.3†				

*Not shown in Figures 2 and 3.

†Joint-hypocenter location of the 26 January 2001 M_{bLg} 4.3 mainshock using as master the 3 June 2001 hypocenter. Location uncertainty: vertical ±1.5 km; horizontal maximum ±1.9 km, azimuth 114°; minimum ±1.2 km.

N is the number of phase picks; Gap is the largest angle uncovered by stations as seen from epicenter; MD is the epicentral distance to closest station; and RMS is the root mean square observed − calculated arrival times.

Ashtabula continued to be felt and to be recorded by regional stations much after the temporary instrumental deployment (Fig. 4).

Seismicity Triggered by Fluid Injection

A year prior to the 1987 mainshock, a disposal well had started injecting fluid in the basal formation (Mt. Simon sandstone) 0.7 km north of the active fault (Figs. 2 and 3). A relatively uniform flow rate of about 164 m^3/day and a wellhead pressure close to 100 bars, the maximum permitted, had forced into the rock about 60×10^3 m^3 of waste fluid at the time of the first mainshock (Nicholson and Wesson, 1990). The total injected surface volume during the 8-year lifetime of the well was about 0.34×10^6 m^3 (well data from H. Gerrish, 2003). The toxicity of the fluid injected and stringent requirements about hydraulic isolation from potable aquifers characterizes this as a class 1 well (www.epa.gov/safewater; November 2002). The temporal and spatial

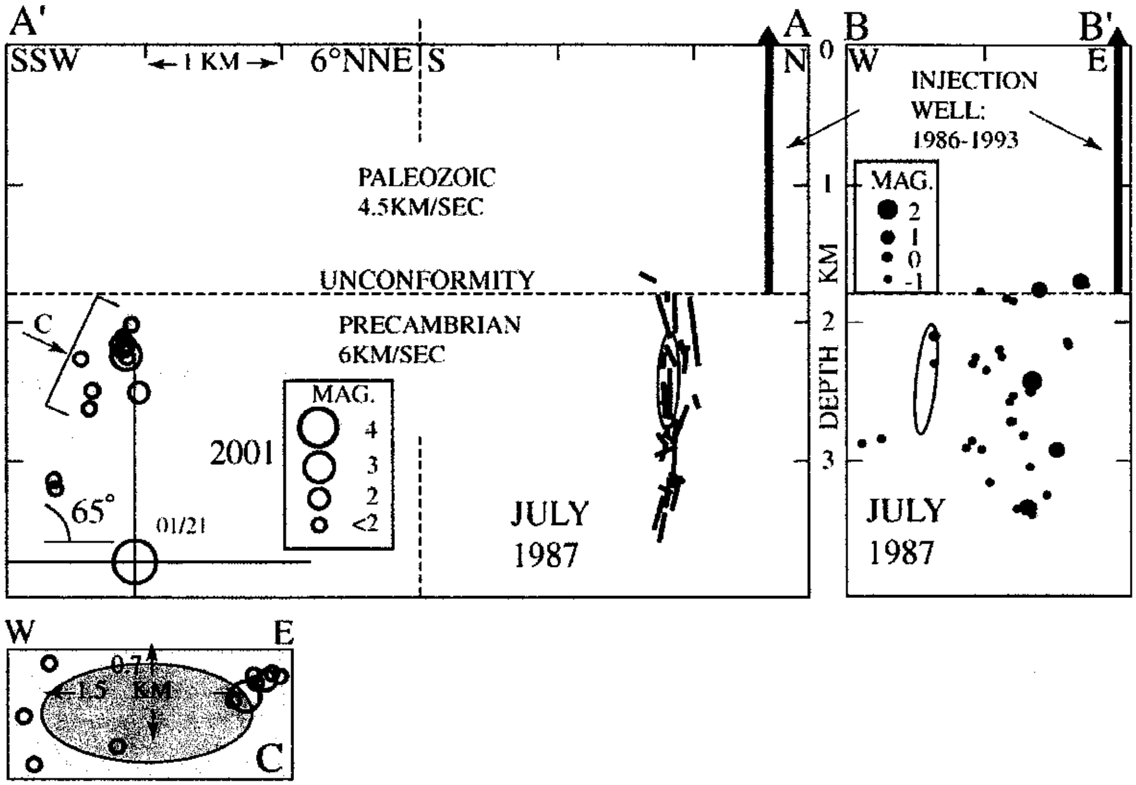

Figure 3. Ashtabula hypocenters constrained by local networks (except for the 26 January 2001 mainshock). A and B are vertical sections (located in Fig. 2), and C is an oblique section (located in A) parallel to the interpreted fault and showing the central cluster in 2001 and the elliptical rupture assumed for the 26 January mainshock. Segments connect 1987 hypocenters of the same event obtained with independent phase readings on analog recordings (Seeber and Armbruster, 1993). The HYPOINVERSE error ellipse is for an event recorded by a short instrumental deployment during August 1989. Higher resolution digital data in 2001 allow even smaller location uncertainties (<100 m, or about the size of the smallest symbol). The well was used for injecting toxic fluid waste (class 1) in the Mt. Simon formation at the base of the Paleozoic section from 1986 to 1994.

proximity between injection and earthquake generation, particularly when compared to the regional quiescence that preceded it (Fig. 1), provides evidence that the injection triggered the seismicity (Nicholson and Wesson, 1990; Seeber and Armbruster, 1993). Fluid injection ceased in June 1994, but seismicity continued. A burst of seismicity in 1995 included felt earthquakes. No earthquakes were reported during the next 5 years, but small earthquakes may have gone unnoticed because instrumental coverage was poor during this period (Fig. 4). This apparent temporary quiescence, however, may mark the end of seismicity from the fault close to the well. Preliminary modeling indicates that pore pressure begins to drop at that fault about a year after the well was shut down in 1994, but it does not reach a maximum at the more distant fault until 2000 (H. Gerrish, personal comm., 2003).

Two Subsequences during the First Half of 2001

Seismicity returned in 2000 and increased dramatically in 2001, producing the largest event so far. After a felt M_{blg} 2.6 foreshock on 20 January 2001, an M_{blg} 4.3 mainshock caused slight damage (modified Mercalli intensity [MMI] VI) on 26 January. A focal mechanism obtained by modeling regional waveforms of this earthquake (Du *et al.*, 2003) is consistent with composite focal mechanisms from locally recorded earthquakes (Fig. 2). Our location of the 26 January mainshock (discussed later) rules out a source for this event on the fault active in 1987 and is consistent with a source on the fault defined by seismicity that occurred during the second half of 2001 (Fig. 2). Thus all the 2000 and later seismicity is likely to originate from that source.

On 3 June 2001, an M_{blg} 3.0 earthquake was felt in Ashtabula. Two days earlier, we had installed four portable

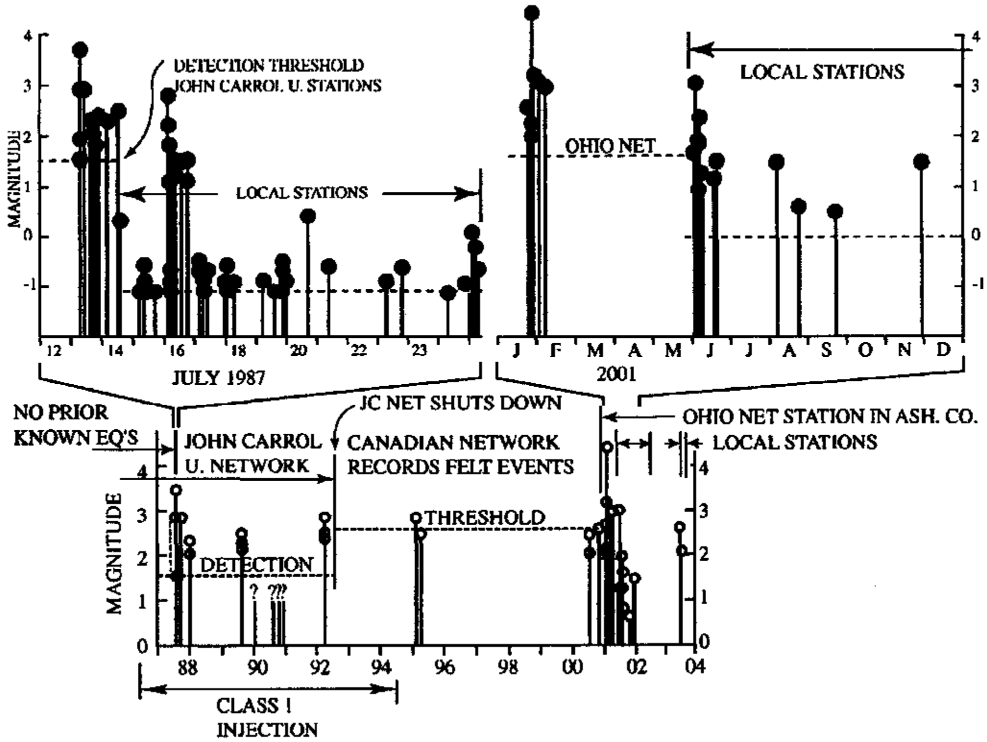

Figure 4. Temporal distribution of known seismicity in Ashtabula. The closest epicenter prior to 1987 is about 30 km from Ashtabula. The detection threshold drops about 2 orders of magnitude when local networks are in place. Monitoring continued until May 2002 without detecting other local events. The magnitudes of small earthquakes recorded only by the local analog recorders in 1987 are probably underestimated.

digital seismographs within 5 km of the source. An additional instrument was installed a week later. We recorded not only the M 3.0 mainshock, but also a foreshock and several aftershocks. About one earthquake per month was recorded during the latter half of 2001. No earthquakes were detected afterward up to May 2002, when our second instrumental deployment ended. On 17 July 2003, an $M_{b_{Lg}}$ 2.5 earthquake was felt in Ashtabula and was located by the Ohio Geologic Survey near the 2001 source. Our third instrumental deployment started 21 July 2003 and captured an M 2.1 event on July 22. The hypocenter is within the preliminary location uncertainty from the 2001 source (Fig. 3). The complex 16-year-long earthquake sequence in Ashtabula is still continuing 9 years after the cessation of class 1 fluid injection (Fig. 4).

Analysis

The June–December 2001 data from our small but strategically located network permit accurate locations (Table 1). Ray paths are short (2–5 km), attenuation is low (Frankel *et al.*, 1996), and seismic waves are rich in high-frequency energy. Sampling rates are high (200 samples/sec), and relative phase arrival times can be picked with high precision by matching waveforms (Fig. 5). Independent phase picks generally differ by less than 0.01 sec, which translates into uncertainties of relative locations $\leq$0.1 km. Independent phase readings from smoke-paper records in 1987 yielded somewhat larger, but still remarkably small, uncertainties (Fig. 3). Ray paths are mostly through horizontally stratified platform rock with velocities inferred from the log of the Ashtabula well (C. Nicholson, personal comm., 1988). Thus lateral velocity changes are probably small and locations are expected to be reliable absolutely, not just relative to each other.

Thirteen hypocenters have been determined from the June–December 2001 data of the local network. These hypocenters are scattered over 7 km east–west (Fig. 2), but they closely ($\pm$0.1 km) fit a plane striking 96° and dipping 65° south (Fig. 3). This is also a nodal plane in a composite solution for the same events and is very similar to one of the nodal planes for the 26 January mainshock (Fig. 2). In the simplest interpretation, this plane represents the source fault for the seismicity in 2001. A preliminary hypocenter from an ongoing third local seismograph deployment starting July 2003 is also from the same source. This fault resembles the source of the 1987 seismicity in having a nearly east–west

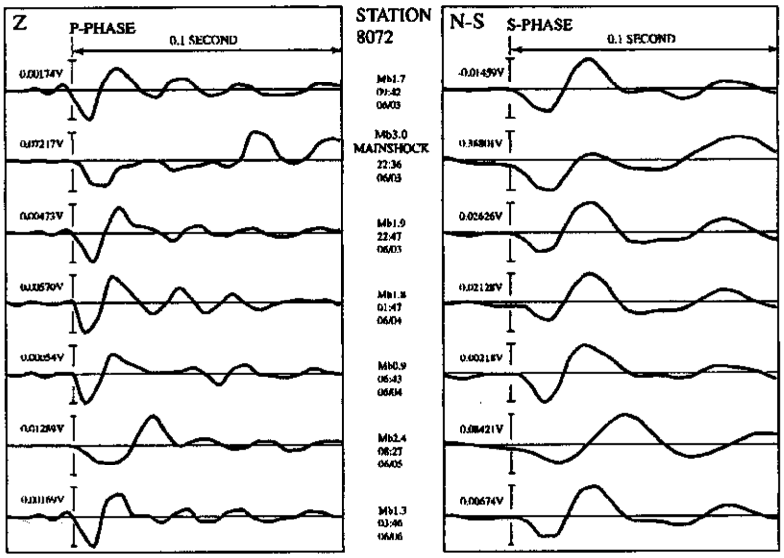

Figure 5. Example of digital seismograms (200 samples/sec) from the June 2001 subsequence in Ashtabula, Ohio. They are amplitude normalized (maximum amplitude in volts) and are aligned according to the phase picks. Waveforms are very similar, except two. The M_b 3.0 mainshock stands out because of relatively low frequency energy (~10 Hz). The aftershock on June 5 exhibits relatively slow rise time in both P and S phases.

strike, left-lateral slip, and in being located in the basement just below the unconformity. These subparallel faults are 0.7 and 4.5 km south of the well, respectively, and are 4 km apart. They account for all the accurately located earthquakes in 1987, 2001, and 2003, including the 26 January mainshock (Figs. 2 and 3). Unfortunately, all other earthquakes from August 1987 to 2001 (Fig. 4) have poor location constraints and cannot be assigned to either of these or other possible sources below Ashtabula. The M 2.9 on 28 March 1992 was located significantly to the west of the 1987 and 2001 sources but could originate from either of the active faults (Fig. 2) (Seeber and Armbruster, 1993).

The 26 January 2001 M_{blg} 4.3 Hypocenter from Regional Waveforms

Both $M_b \geq 3$ Ashtabula events in 2001 were recorded by a set of regional stations in the distance range 150–500 km, including five with broadband instruments (SADO, BINY, SSPA, MCWV, and ACSO) (Fig. 1). The seismograms from these events are remarkably similar (Fig. 6), suggesting similar source location and fault kinematics. The M_{blg} 3.0 June event was also recorded locally and accurately located. We thus located the M_{blg} 4.3 January event relative to the June 3 hypocenter. Accurate differential Pn arrival times were obtained by seeking maxima in waveform correlation. Additional constraints were obtained from Sn-Pn times derived from cross-correlating waveforms. The June event was then used as the master in a joint hypocenter determination with the January event. Results show that the southern fault is the likely source of the January mainshock (Table 1; Figs. 2 and 3).

Seismogenic Faults versus Mainshock Ruptures

Accurate aftershock hypocenters are often found to outline the mainshock rupture, even for small mainshocks (e.g.,

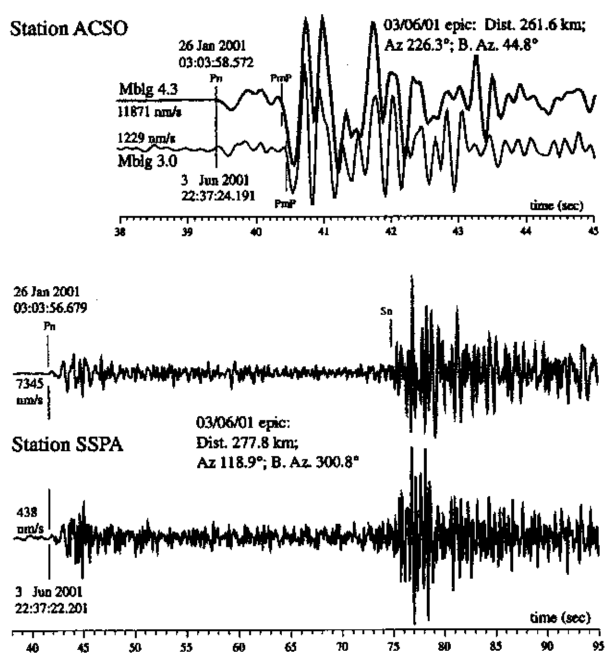

Figure 6. Comparison of vertical seismograms from the mainshocks of the January and June 2001 subsequences recorded at stations ACSO and SSPA. The seismograms are aligned where the cross correlation is at a maximum. Uncertainty in relative P-wave arrival time is <0.1 sec. Note the 0.05-sec difference in PmP-Pn time for the two events recorded at ACSO. This suggests that the larger event in January is about 1 km deeper than the June event.

Hough and Seeber, 1991). Excluding two isolated events that occurred late in the sequence (Table 1), 2001 hypocenters outline a fault patch about 1.5 km along strike and 0.7 km along dip (Fig. 3C). Assuming the January 2001 M_{blg} 4.3 mainshock had an elliptical rupture with axes $2r_1 = 1.5$ km and $2r_2 = 0.7$ km, we calculate a stress drop of $\Delta s =$ 21 bars from the 2001 mainshock moment $M_o = 8 \times 10^{21}$ dyne cm (Du *et al.*, 2003). This stress drop is low, but within the typical range, and it is similar to the one of the very shallow 1994 quarry-triggered earthquakes in Cacoosing, Pennsylvania (Seeber *et al.*, 1998). In contrast, the 1987 hypocenter patch, 1.0×1.6 km, is large compared to the M_{blg} 3.8 event in 1987 and is likely to extend far beyond the rupture of that mainshock. In agreement with this hypothesis, the "aftershocks" are not clustered along the edges of the patch but are scattered over it. In time, they cluster in two subsequences against a steady background of very small events (Fig. 4). This evidence suggests that the 1987 seismogenic patch was the portion of the fault activated by the high pore-pressure transient caused by the injection, rather than marking any single rupture.

Discussion

Injection-Related Seismicity in Ashtabula Compared

The time history of waste fluid injection and related seismicity at the Rocky Mountain Arsenal near Denver, Colorado, is similar to that at Ashtabula. The Denver injection was in 3.8-km-deep fractured basement and enjoyed relatively compliant hydrologic conditions. The total fluid injected near Denver was about 0.6×10^6 m^3, nearly twice as much as in Ashtabula. Although the wellhead pressure was higher in Ashtabula (about 100 bars) than in Denver (0–72 bars), the average flow was about 5 times lower in Ashtabula. Seismicity near Denver followed the onset of injection almost immediately, continued after the end of injection for at least 6 years, reached as far as 6 km from the well, and produced three $M_b \geq 5.0$ earthquakes, including one 21 months after injection ceased. This seismicity was shown to be triggered by the high pore-pressure front migrating outward from the well (Healy *et al.*, 1968; Hsieh and Bredehoeft, 1981).

Preliminary hydraulic modeling (H. Gerrish, personal comm., 2003) suggests that pressure on the fault 0.7 km from the well rose rapidly during the first year and reached near-maximum values at the time of the first mainshock, July 2003. In 1995, 1 year after the end of injection, the expanding decompression zone is expected to reach that fault. The burst of seismicity in 1995 (Fig. 4) was probably the last significant seismicity from this fault. Fourteen years after the onset of injection, the crest in the pore-pressure rise is expected to reach 5 km from the well, near the fault that became active at that time and continues to be active in 2003. The prominent quiescence from 1995 to 2000 (Fig. 4) is relevant to the lack of hypocenters between those faults and suggests that no other favorably oriented fault was available between them. Generally, then, precise hypocenters and preliminary modeling support the hypothesis that earthquakes are triggered on pre-existing faults by injection-related pore-pressure rise and are turned off when the pressure begins dropping back, according to the Coulomb fault failure criterion (e.g., Stein and Lisowski, 1983). This scenario is consistent with hydrologic modeling of the Calhio Chemicals, Inc. waste disposal wells, which are 50 km west-southwest of Ashtabula and have similar geologic, hydrologic, and engineering characteristics as well as similar time-space relation to seismicity. The Calhio Chemicals, Inc. wells were injected from 1972 to 1985 and may have triggered the M_b 5.0, MMI VII January 1986 earthquake 12 km south of the wells, 14 years after the onset of injection (Fig. 1) (Nicholson *et al.*, 1988; Seeber and Armbruster, 1993). During a similar time lag after the onset of injection, the Ashtabula sequence reached as far as 9 km from the well (Fig. 2). Given the time histories of injections, the pressure perturbation may still be significant at 10 km from the well if the flow is confined within an aquifer with two- or one-dimensional geometry (Hsieh and Bredehoeft, 1981; Nicholson and Wesson, 1990). Two-dimensional flow is expected to characterize fluid waste disposal into the aquifer confined at the base of the platform sequence. Permeability at this horizon may depend not only on the basal formation (Mt. Simon sandstone) but also on a fractured and weathered surface at the top of the basement. Fracture density at the unconformity may be further enhanced along the intersection of steep faults leading to linear zones of enhanced permeability. The overall behavior of the pressure front is therefore likely to reflect a combination of two- and one-dimensional flow paths.

Are All Ashtabula Earthquakes Related? How?

The spatial and temporal clustering of the seismicity in Ashtabula (seven distinct episodes of felt earthquakes in 14 years within a $\approx$4-km radius; Figs. 3 and 4) stands in sharp contrast to the low regional historic seismicity ($\approx$30 km from Ashtabula to the closest epicenter for at least a century before 1987; Fig. 1). Given such a background, from 1987 to 2001 seismicity was persistent and concentrated (Fig. 4), as would be expected in a single sequence. Locations exclusively from regional data would be much fewer and probably indistinguishable. Conceivably, however, two independent earthquake sequences could occur in close proximity. Assuming that these two sequences could occur any time within 100 years and in an area with a radius of 30 km, the minimum space-time without felt earthquakes around Ashtabula prior to 1987, the likelihood that they would occur within a 4-km radius and 10 years is very low: $\pi(4 \text{ km})^2(10 \text{ yr})/\pi(30 \text{ km})^2(100 \text{ yr}) = 1/562$. The proximity between these two hypothetical sequences would not be as astonishing, however, if they were related to separate instances of fluid injection, which of course are not randomly distributed. This hypothesis would require a second injection site in the vicinity of the 2001 hypocenters (Fig. 2). Ashtabula County and many other areas of the Appalachian Plateau have been exploited for gas since early in the twentieth century. Often gas extraction involves flooding, or nontoxic brine injection in wells that are designated class 2 (www.epa.gov/safewater; November 2002). These wells have less stringent requirements of hydrologic confinement from potable aquifers and are typically less deep than class 1 wells. A rise in the price of natural gas might have recently accelerated this activity. It is unclear what role these wells are playing in the seismicity of the Appalachian Plateau. Available evidence does not seem to implicate class 2 wells in the Ashtabula seismicity for the following reasons (information on gas fields personally communicated by M. Hansen, 2001): (1) the extent of gas extraction in time and space is much larger than the extent of the seismicity; (2) no well is active within 5 km of the 2001 source; and (3) class 2 wells in the Ashtabula area generally penetrate only the upper part of the Paleozoic section, which is hydraulically and mechanically isolated from the lower section and from the basement (Evans, 1987).

A single sequence of causally related earthquakes is thus the more likely hypothesis for the seismicity in Ashtabula. Within this hypothesis, the 2001 seismicity could be triggered by earlier fluid-injection-related earthquakes in a domino effect. This, however, seems also unlikely, given the small size of these earthquakes ($M_{b_{lg}} \leq 3.8$) and the distance between sources (4 km). We conclude that the 1987–2001 seismicity in Ashtabula is probably all related to the 1986–1994 injection episode.

Fault Reactivation

In Ashtabula, all accurately located earthquakes originated from two parallel faults. These data cover only two short windows of the active period, but they are inconsistent with a zone of diffuse small faults and joints, such as the one proposed to control both fluid flow and seismicity in the Denver sequence (Hsieh and Bredehoeft, 1981). The two distinct sources may be the only significant pre-existing basement faults oriented favorably in the current stress field and within reach of the pore-pressure anomaly. Their spacing may be representative of potentially seismogenic faults in SCRs (Seeber *et al.*, 1996). These faults, however, are not associated with significant vertical offsets of the platform and were unknown to local geologists. A reflection survey revealed no fault in the vicinity of the well (Reserve Environmental Services, Inc., Ashtabula Ohio, quoted in the *Star Beacon*, 24 September 1992). Current strike-slip motion (Fig. 2) is not expected to offset the horizontal reflectors, but these markers are Paleozoic and the negative result suggests that these active faults are either very young or very slow.

The hypothesis that pre-existing rift-related faults are the sources of the largest SCR earthquakes (e.g., Johnston, 1989) has been a guiding principle in hazard assessment. Our results in Ashtabula suggest pre-existing structures play also a role in nonrifted areas of SCR. Most surface-rupturing SCR events have occurred on relatively small faults with subtle structural features that had not been recognized prior to the rupture. The 1990 rupture in Ungava, Quebec, displayed no evidence of prior brittle faulting (Adams *et al.*, 1991). The source fault of the 1993 Killari, central India, event had accumulated a maximum displacement of 6 m since the Cretaceous (Seeber *et al.*, 1996). The 1994 Cacoosing Valley, Pennsylvania, earthquake ruptured a fault delineated by aftershocks as close as half a kilometer from the surface, but its only structural expression is a joint set (Seeber *et al.*, 1998). Thus the potential for destructive earthquakes is not necessarily diminished by the absence of a clear geologic signal for the Ashtabula faults.

Triggered Earthquakes: A Characteristic of Stable Continental Regions?

Seismicity in SCRs is relatively low yet poses substantial risk (e.g., Frankel *et al.*, 1996). Stress is generally uniformly oriented over large SCRs and is measured consistently close to failure (Evans, 1987, 1989; Zoback and Zoback, 1997). Thus SCRs, like rapidly deforming regions, seem to be kept at a critical state by deformation and brittle failures. Deformation rates in SCRs, however, are several orders of magnitude lower than at typical plate boundaries and are generally below the resolution limit of both geologic and geodetic measurements (Argus and Gordon, 1996). Earthquakes show that deformation is occurring in SCRs and that it is consistent with the regional stress state (Sbar and Sykes, 1973; Anderson, 1986).

High-stress and low-strain conditions in SCRs are manifested by a relative abundance of triggered earthquakes. Earthquake–earthquake triggering is manifested by nonrandom distributions, such as prolonged sequences of earthquakes in specific hot spots, or by bursts of distant events following a large earthquake. These groupings could derive from static (e.g., Stein and Lisowski, 1983; Kenner and Segall, 2000) or dynamic (Gomberg, 1996; Hough, 2001; Hough *et al.*, 2003) earthquake-induced changes in stress, respectively. In a steady-state regime, these naturally triggered earthquakes are a stable component of natural seismicity, which is controlled by tectonic strain rate. In contrast, seismicity triggered by human activities, or anthropogenic seismicity (Evans, 1987; McGarr *et al.*, 2002), is a recent phenomenon, which will require millennia to reach equilibrium with the very low strain rate typical of SCRs (Seeber, 2002a). Thus, anthropogenic seismicity has had no significant effect yet on the rate of natural SCR seismicity and is therefore added onto it.

The ratio of anthropogenic to natural earthquakes is expected to be much higher in SCRs than in active areas because the level of natural seismicity is proportional to strain rate, while the likelihood that an anthropogenic perturbation brings a fault to failure depends on stress level, as long as the perturbation is short relative to the loading cycle (Seeber, 2002a). From this consideration alone, therefore, the number of triggered earthquakes in active areas and SCRs should be similar, in contrast to vastly different levels of natural seismicity. The potential for anthropogenic earthquakes may actually be higher in SCRs (Seeber, 2002b; Hough *et al.*, 2003). One reason is that the upper few kilometers of SCR crust tends to be more competent than in active areas. This difference in the mechanical properties is manifested by distinct depth distributions of seismicity. While the upper few kilometers of the crust nucleates many of the large SCR earthquakes (e.g., Choy and Bowman, 1990; Adams *et al.*, 1991; Seeber *et al.*, 1996), the same depth range in active areas generally produces relatively little seismicity. Thus, anthropogenic perturbations, which are typically significant only in the upper few kilometers, may be more likely to find mature conditions for a damaging earthquake in an SCR than along a plate boundary.

The Ashtabula sequence is continuing after more than 15 years and includes several mainshock–aftershock subsequences, which are typically only a few days long. These distinct timescales suggest distinct processes responsible for interdepence between earthquakes. While the short aftershock sequences are likely to stem from typical earthquake–earthquake interactions, the multiyear anthropogenic perturbation of pore pressure is thought to be responsible for the long-term sequence. Long-lasting perturbation in fluid pressure and flow may also be induced by large earthquakes or other natural processes and may play a role in natural seismicity as well (Muir-Wood and King, 1993). Rather than permanent features of the tectonic regime, prominent spatial clusters of SCR seismicity may instead be long-lasting temporary sequences of related events (Seeber and Armbruster, 1989; Ebel *et al.*, 2000). The Ashtabula sequence could therefore serve as an analog for long-term natural sequences similarly controlled by multiple time constants.

Triggered Earthquakes and Hazard

The occurrence of earthquakes can be affected by small stress changes of as little as 0.1 bar (Reasenberg and Simpson, 1992; Seeber and Armbruster, 2000). In SCRs, where strain rate and natural seismicity are low and stress is high, seismicity may be particularly prone to triggering by nontectonic factors, such as human activities (Simpson, 1986; McGarr *et al.*, 2001; Nicholson and Wesson, 1990; Seeber, 2001). This hypothesis is borne out by high-resolution data on SCR earthquakes worldwide. For example, a substantial portion, about 1/3, of the earthquake sequences in the northeastern United States monitored by the Lamont Cooperative Seismic Network were triggered by human activities (e.g., Table 2). Recent human-triggered seismicity in the eastern United States appears to be comparable to natural seismicity in both the number of earthquakes and their size distribution, but the common assumption is that an earthquake is natural unless proven otherwise. Generally, engineering activities keep expanding and anthropogenic earthquakes may still be on the rise. This class of earthquakes contributes significantly to the overall hazard in SCRs. More importantly, this hazard is generally concentrated in small areas and specific times that can be precisely identified from known mechanical pollution. This pollution needs to be systematically monitored and considered as a distinct element in hazard calculations (Seeber, 2002b).

A Strategy for SCR Seismometry

High-resolution earthquake data are critical to understand and monitor earthquake triggering. Such data require dense seismograph coverage. Over large SCRs, dense coverage can only be accomplished by temporary targeted deployments of portable seismographs. Aftershock surveys rely on rapid deployment to exploit the high data return in the immediate aftermath of a mainshock, but they tend to be too temporary. Their data often resolve first-order geometry and kinematics of the source, but rarely the temporal structure. SCR seismometry needs to take stock of possible fundamental differences between aftershocks and complex long-term earthquake sequences. In order to discriminate among these processes, temporary instrumental deployments need more staying power. Experiences in Ashtabula and elsewhere in the eastern United States suggest that sequence monitoring may offer unique scientific returns. Sequences

Table 2
Earthquakes thought to be Anthropogenic

Earthquakes Known to be Anthropogenic
Dale, New York, 1971, M <2 (hydraulic salt mining) (Fletcher and Sykes, 1977)
Wappingers Falls, New York, 1974, M 3.0 (quarry unloading) (Pomeroy *et al.*, 1976)
Mineville, New York, 1983–1990, M_c 3.0 (flooding of deep mine) (J. G. Armbruster, unpublished data, 1990)
Ashtabula, Ohio, 1987–2003, M_{blg} 3.8–4.5–3.0, MMI VI (deep fluid injection) (Seeber and Armbruster, 1993; this article)
Cacoosing, Pennsylvania, 1993–2000, M_{blg} 4.6, MMI VI–VII (quarry unloading) (Seeber *et al.*, 1998)
Earthquakes Suspected of Being Anthropogenic
Avoca, New York, 2001, M_{blg} 2.9 (deep fluid injection) (J. G. Armbruster, unpublished data, 2001)
Leroy, Ohio, 1986, M_{blg} 5.0, MMI VII (deep fluid injection) (Nicholson and Wesson, 1990; Seeber and Armbruster, 1993)
Attica, New York, 1929, M_b 5.2, MMI VIII (hydraulic salt mining) (Seeber and Armbruster, 1993)
Genesseo, New York, 1994, M 3.6 (mine collapse) (L. Seeber and J. G. Armbruster, unpublished report/proposal to Akzo Salt, Inc., 1994)
Florida, New York, 2003 M 2.3 (quarry unloading)

Basis: seismological observations in New York, New Jersey, Maryland, Delaware, Pennsylvania, and eastern Ohio (the area monitored by the Lamont Cooperative Seismic Network).

could be monitored by portable networks that are kept permanently deployed, rather than on stand-by waiting for a mainshock. Sequence monitoring can be a component of regional monitoring and needs to be incorporated into the design of the Advanced National Seismic System for the eastern United States.

Conclusions

1. The first known earthquake in the Ashtabula, Ohio, area was felt in 1987. At least seven episodes of felt earthquakes occurred in the next 16 years, including an event that caused slight damage in 2001. All these earthquakes originated from a relatively small area (≈10 km wide) and are likely to form a single sequence of causally related earthquakes.
2. The Ashtabula earthquake sequence is closely associated with injection of waste fluid in the basal Paleozoic formation (about 0.3×10^6 m^3 with a wellhead pressure of 100 bars from 1986 to 1994). This anomaly is expected to expand from the injection point two dimensionally along the basal Paleozoic formation and to continue to do so even after the end of injection. At that time, however, pore pressure is expected to drop near the well. Felt earthquakes started in 1987, 1 year after the onset of injection. At that time earthquakes were located 0.7–2.0 km to from the injection site. Seismicity continued and in 2001, 5.5 years after the end of injection, hypocenters were then 5–9 km from the injection site. The only known episode of seismicity in Ashtabula, therefore, is not only closely associated with the 1986–1994 class 1 injection, but the pattern of accurate hypocenters is consistent with the one expected for the high pore-pressure anomaly spreading from the injection site. We consider this correlation strong evidence that the seismicity was triggered by the injection.
3. All 42 accurate hypocenters from the Ashtabula sequence are associated with one or the other of two faults defined by this seismicity. The structures are steep, strike nearly east–west, slip left-laterally, are 4 km apart, and are interpreted as reactivated pre-existing faults. Both seismogenic fault patches are centered in the uppermost portion of the Precambrian basement and are bound up-dip at the 1.8-km-deep unconformity between basement and Paleozoic platform rocks. One patch was active in 1987 with an M_{blg} 3.0, the other in 2001 with an M_{blg} 4.3 and again in 2003. Many other earthquakes in the same sequence were not constrained by local stations, but the temporal distribution is consistent with all 1987–1995 and 2000–2003 earthquakes originating from the 1987 and 2001 sources, respectively, in the context of triggering by an expanding high pore-pressure anomaly. Quiescence from 1995 to 2000 is interpreted as lack of favorably oriented faults between the two sources.

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References

Adams, J., R. J. Wetmiller, H. S. Hasegawa, and J. Drysdale (1991). The first surface faulting from an intraplate earthquake in North America, *Nature* **352,** 617–619.

Anderson, J. G. (1986). Seismic strain rates in the central and eastern U.S., *Bull. Seism. Soc. Am.* **76,** 273–290.

Argus, D. F., and R. G. Gordon (1996). Test of the rigid plate hypothesis and bound on intraplate deformation using geodetic data from Very Long Baseline Interferometry, *J. Geophys. Res.* **101,** 13,555–13,572.

Choy, G. L., and J. R. Bowman (1990). Rupture process of a multiple main shock sequence: analysis of teleseismic, local, and field observations of the Tennant Creek, Australia earthquakes of January 22, 1988, *J. Geophys. Res.* **95,** 6867–6882.

Du, W.-X., W.-Y. Kim, and L. R. Sykes (2003). Earthquake source parameters and state of stress for northeastern United States and southeastern Canada from analysis of regional seismograms, *Bull. Seism. Soc. Am.* **93,** 1663–1648.

Ebel, J. E., K.-P. Bonjer, and M. C. Oncescu (2000). Paleoseismicty: seismicity evidence for past large earthquakes, *Seism. Res. Lett.* **71,** 283–294.

Evans, K. F. (1987). Assessing regional potential for induced seismicity, National Center for Earthquake Engineering Research, Buffalo New York, Technical Report, NCEER87-0025, 30 pp.

Evans, K. F. (1989). Appalachian stress study, regional scale stress variations and their relation to structure and contemporary tectonics, *J. Geophys. Res.* **94,** 17,619–17,645.

Fletcher, J. B., and L. R. Sykes (1977). Earthquakes related to hydraulic mining and natural seismic activity in western New York State, *J. Geophys. Res.* **82,** 3767–3780.

Frankel, A., C. Mueller, T. Barnhard, D. Perkins, E. V. Leyendecker, N. Dickman, S. Hanson, and M. Hopper (1996). National Seismic Hazard Maps: documentation June 1966, *U.S. Geol. Surv. Open-File Rept. 96-532,* 70 pp.

Gomberg, J. (1996). Strain changes and triggered seismicity following the M_w 7.3 Landers, California earthquake, *J. Geophys. Res.* **101,** 751–764.

Healy, J. H., W. W. Rubey, D. T. Griggs, and C. B. Raleigh (1968). The Denver earthquakes, *Science* **161,** 1301–1310.

Hough, S. E. (2001). Triggered earthquakes and the 1811–1812 New Madrid, central U.S. earthquake sequence, *Bull. Seism. Soc. Am.* **91,** 1574–1581.

Hough S. E., and L. Seeber (1991). Seismological constraints on the source properties of m_b = 4.0 1985 Ardsley New York earthquake: a characteristic rupture? *J. Geophys. Res.* **96,** 18,183–18,195.

Hough S. E., L. Seeber, and J. G. Armbruster (2003). Intraplate triggered earthquakes: observations and interpretations, *Bull Seism. Soc. Am.* **93**, 2212–2221.

Hsieh, P. A., and J. D. Bredehoeft (1981). A reservoir analysis of the Denver earthquakes: a case of induced seismicity, *J. Geophys. Res.* **86**, 903–920.

Johnston, A. C. (1989). The seismicity of stable continental interiors, in *Earthquakes at North-Atlantic Passive Margins*, S. Gregersen and P. W. Basham (Editors), Kluwer, Hingham, Massachusetts, 299–327.

Kenner, S. J., and P. Segall (2000). A mechanical model for intraplate earthquakes: application to the New Madrid seismic zone, *Science* **289**, 2329–2332.

McGarr, A., D. Simpson, and L. Seeber (2002). Case histories of induced and triggered seismicity, in *International Handbook of Earthquake and Engineering Seismology*, W. H. K. Lee, P. C. Jennings, H. Kanamori, and C. Kisslinger (Editors), Academic Press, 647–664.

Muir-Wood, R., and G. C. P. King (1993). Hydrological signature of earthquake strain, *J. Geophys. Res.* **98**, 22,035–22,068.

Nicholson, C., and R. L. Wesson (1990). Earthquake hazard associated with deep well injection: a report to the U.S. Environmental Protection Agency, *U.S. Geol. Surv. Bull. 1951*, 74 pp.

Nicholson C., E. Roeloffs, and R. L. Wesson (1988). The northeastern Ohio earthquake of 31 January 1986: was it induced? *Bull Seism. Soc. Am.* **78**, 188–217.

Pomeroy, P. W., D. W. Simpson, and M. L. Sbar (1976). Earthquakes triggered by surface quarrying—The Wappingers Falls, New York sequence of June, 1974, *Bull. Seism. Soc. Am.* **66**, 685–700.

Reasenberg, P. A., and R. W. Simpson (1992). Response of regional seismicity to the static stress change produced by the Loma Prieta earthquake, *Science* **255**, 1687–1690.

Sbar, M. L., and R. L. Sykes (1973). Contemporary compressive stress and seismicity in eastern North America: an example of intraplate tectonics, *Geol. Soc. Am. Bull.* **84**, 1861–1882.

Seeber, L. (2002a). Triggered earthquakes and hazard in stable continental regions, report for the U.S. Army Engineers, Waterways Experiment Station, Vicksburg, Mississippi, 40 pp.

Seeber, L. (2002b). Mechanical pollution, *Seism. Res. Lett.* **73**, 315–317.

Seeber, L., and J. G. Armbruster (1989). Low-displacement seismogenic faults and nonstationary seismicity in the eastern United States, *Ann. N. Y. Acad. Sci.* **558**, 21–39.

Seeber, L., and J. G. Armbruster (1991). The NCEER91 earthquake catalog: improved intensity-based magnitudes, and recurrence relations for U.S. earthquakes east of New Madrid, National Center of Earthquake Engineering, Technical Report 91-0021, 98 pp.

Seeber, L., and J. G. Armbruster (1993). Natural and induced seismicity in the Erie–Ontario region: reactivation of ancient faults with little neotectonic displacement, *Géogr. Phys. Q.* **47**, 363–378.

Seeber, L., and J. G. Armbruster (2000). Earthquakes as beacons of stress change, *Nature* **407**, 69–72.

Seeber, L., J. G. Armbruster, W.-Y. Kim, C. Scharnberger, and N. Barstow (1998). The 1994 Cacoosing Valley earthquake near Reading, Pennsylvania: a shallow rupture triggered by quarry unloading, *J. Geophys. Res.* **103**, 24,505–24,521.

Seeber, L., G. Ekstrom, S. K. Jain, C. V. Murty, N. Chandak, and J. G. Armbruster (1996). The 1993 Killari earthquake in central India: a new fault in Mesozoic basalt flows? *J. Geophys. Res.* **101**, 8543–8560.

Simpson, D. W. (1986). Triggered earthquakes, *Ann. Rev. Earth Planet. Sci.* **14**, 21–42.

Stein, R. S., and M. Lisowski (1983). The 1979 Homestead Valley earthquake sequence, California: control of aftershock and postseismic deformation, *J. Geophys. Res.* **88**, 6477–6490.

Zoback, M. L., and M. D. Zoback (1989). Tectonic stress field of the continental United States, in *Geophysical Framework of the Continental United States*, L. Parkiser and W. Mooney (Editors), *Geol. Soc. Am. Memoir*, Vol. 172, 523–539.

Zoback, M. L., and M. D. Zoback (1997). Crustal stress and intraplate deformation, *Hallesches Geowiss. Jahrb.* **15**, 116–123.

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