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NATURAL AND INDUCED SEISMICITY IN THE LAKE ERIE-LAKE ONTARIO REGION: REACTIVATION OF ANCIENT FAULTS WITH LITTLE NEOTECTONIC DISPLACEMENT

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ABSTRACT The two most prominent seismic zones in the Lake Erie-Lake Ontario region are associated with the Akron magnetic lineament and with the Clarendon-Linden fault. Both these features are recognized from geophysical data as regional basement structures related to the Grenville collisional orogen. Neotectonic displacement is not geologically evident, although Paleozoic reactivation is manifested by the Clarendon-Linden fault. We have sharpened the definition of seismic zones in the region by introducing newly discovered events, improving constraints on locations and size for many others, and omitting unreliable ones. This seismicity tends to occur on old faults with minor neotectonic displacements. Related conclusions are: 1) neotectonic surface displacement is not necessary for fault capability, 2) seismogenic faults may have geological and/or geophysical expressions, and 3) a stationary moment release at the historic level requires more capable faults than the ones active during the historic period. Waste fluid injection, oil recovery, and salt-brine recovery have been implicated in cases of induced seismicity in the study area and might have contributed a significant portion of the known earthquakes. Fluid is being injected into the basal platform formation at a depth of 1.8 km near Ashtabula, Ohio. In July 1987, about a year after the onset of injection, a m_{bLg}=3.8 main shock occurred within a 60 km wide area with no known prior seismicity. Aftershocks detected by a short-term local seismic network define a vertical left-lateral fault in the basement just below the platform rocks as close as 700 m from the injection well, probably a reactivated pre-existing fault. Subsequent seismicity suggests a westward migration by 5-10 km, possibly along the same fault.

RÉSUMÉ *Sismicité naturelle et provoquée dans la région des lacs Érié et Ontario: réactivation d'anciennes failles avec décalage néotectonique peu important*. Les deux principales zones sismiques de la région sont associées au linéament magnétique d'Akron et à la faille de Clarendon-Linden. Selon les données géophysiques, ces deux éléments sont reconnus comme étant des structures régionales du socle reliés à l'orogenèse de Grenville. On a précisé la détermination des zones sismiques dans la région en ajoutant les séismes découverts récemment, en précisant les lieux et les dimensions d'autres séismes et en éliminant ceux qui sont mal connus. La sismicité tend à se manifester dans d'anciennes failles avec peu de décalage néotectonique. On en conclut que 1) le décalage néotectonique superficiel ne met pas en cause la compétence d'une faille; 2) les failles d'origine sismique peuvent avoir des manifestations géologiques ou géophysiques; 3) un relâchement momentané localisé survenant au cours de la période historique requiert plus de compétence qu'en ont les failles actives au cours de cette période. L'injection de liquide, la récupération de pétrole ou de sel ont été impliquées dans les cas de sismicité provoquée dans la région et ont probablement causé une bonne partie des séismes connus. Près d'Ashtabula, en Ohio, on a injecté des liquides à une profondeur de 1,8 km à la base de la plate-forme. En juillet 1987, environ un an après le début des injections, un séisme de m_{bLg} = 3,8 s'est produit dans une aire de 60 km de superficie apparamment non sismique. Les répliques enregistrées par un réseau temporaire ont laissé voir une faille verticale à décrochement sénestre, immédiatement sous la plate-forme rocheuse, à près de 700 m du puits d'injection; il s'agit probablement d'une faille réactivée. Par la suite, la sismicité indique une migration de 5 à 10 km, probablement le long de la même faille.

ZUSAMMENFASSUNG *Natürliches und induziertes Auftreten von Erdbeben im Eriesee-Ontariosee-Gebiet: Rekativierung alter Verwerfungen mit geringer neotektonischer Verstellung.* Die zwei hauptsächlichen seismischen Zonen in der Eriesee-Ontariosee-Region werden mit dem magnetischen Lineament von Akron in Verbindung gebracht. Mittels geophysikalischer Daten erkennt man in diesen Bildungen regionale Untergrundstrukturen, die mit der Kollisions-Orogenese von Grenville verbunden sind. Wir haben die Definition der seismischen Zonen in dem Gebiet präzisiert, indem wir kürzlich entdeckte Erdbeben ergänzt und für viele andere die Definition der Ort und Große betreffenden Zwänge verbessert haben, und die Unzuverlässigen weggelassen haben. Diese Erdbeben haben die Tendenz, auf alten Verwerfungen mit geringen neotektonischen Verstellungen aufzutreten. Hieraus kann man schließen: 1) neotektonische Oberflächenverstellung ist keine Bedingung für Verwerfungsfähigkeit, 2) durch Erdbeben entstandene Verwerfungen können sich geologisch und/oder geophysikalisch ausdrücken und 3) eine ortsgebundene momentane Entlastung auf historischer Ebene erfordert mehr fähige Verwerfungen als die, welche während der historischen Periode aktiv waren. In den Fällen von induziertem Auftreten von Erdbeben im untersuchten Gebiet waren Zuführung von Flüssigkeit und öl- und Salzgewinnung beteiligt, und sie haben wohl einen bedeutenden Anteil der bekannten Erdbeben hervorgerufen. Bei Ashtabula, Ohio, hat man in einer Tiefe von 1.8 km Flüssigkeit an der Basis der Plattformbildung eingespritzt. Im Juli 1987, etwa ein Jahr nach Beginn der Einspritzungen, kam es zu einem neuen gewichtigen Beben von m_{bLg} = 3.8 innerhalb eines 60 km breiten Gebiets, in dem vorher kein Erdbeben bekannt war. Nachbeben, die von einem kurzfristigen örtlichen Erdbeben-Netzwerk registriert wurden, ergeben eine vertikale linksseitige Verwerfung im Untergrund unter der Felsenplattform, etwa 700 m vom Einspritzungsbohrloch entfernt; es handelt sich möglicherweise um eine reaktivierte, schon existierende Verwerfung. Darauffolgende Erdbeben weisen auf eine Wanderung westwärts von 5-10 km, möglicherweise längs derselben Verwerfung.

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INTRODUCTION

Intraplate neotectonics and seismogenesis in eastern North America (ENA) is problematic from both scientific and practical points of view. While the fundamental causes of seismicity are poorly understood, the level of seismicity is sufficiently high to cause a substantial hazard. Cost-effective hazard reduction measures depend on estimating realistically the distribution and the level of future seismicity. A reliable earthquake data base and some understanding of intracratonic neotectonic processes are fundamental for these assessments.

The characterization of future seismicity in areas of high deformation rates is generally based on the structural characteristics of active faults, on a geodetic measure of deformation rates and on a space-time sample of earthquakes. In intracratonic environments such as ENA, however, hazard assessments are based primarily on earthquake catalogs because other data pertinent to neotectonics and seismogenesis are scarce and often controversial. Even in areas with large historic earthquakes, major seismogenic structures and tectonic rates remain poorly resolved. Most large intracratonic earthquakes worldwide are spatially associated with preexisting rifts (Johnston, 1989). This correlation offers an opportunity to input, systematically, geologic data in charting sources of potentially damaging earthquakes. Nevertheless, seismicity is still the main observable which can offer insight into intracratonic neotectonics and a basis for hazard assessment.

Unfortunately, conclusions on future seismicity based solely on past seismicity can be grossly misleading, particularly if the available sample is short compared to repeat times of large earthquakes (*e.g.*, Veneziano and Van Dyck, 1985). Patterns of seismicity could change drastically after the occurrence of a large earthquake (*e.g.*, Seeber and Armbruster, 1987), in violation of the assumption of stationary seismic sources. More realistic models that allow temporal changes in seismicity (*e.g.*, Coppersmith and Youngs, 1989) are still poorly constrained because data on prehistoric earthquakes are sparse and intraplate tectonic processes are poorly understood. Finally, earthquake catalogs tend to be patchworks from different workers and contain diverse kinds of data; their reliability tends to vary greatly. Because the quality of a data set reflects the lowest quality in the set, quality-selection can be a very rewarding approach to improve resolution.

In this paper we attempt to characterize intracratonic seismogenesis in the region of ENA covering eastern Lake Erie, western Lake Ontario, and adjacent portions of Ohio, Pennsylvania, New York and Ontario (Fig. 1). In a two-step approach, we first examine the earthquake data base to improve constraints for a number of pre-instrumental events and to emphasize reliable and significant earthquakes. Then, we compare known structural features with seismicity patterns which are apparent in the revised earthquake data. Although correlation between structure and intracratonic earthquakes has been generally elusive, with improved resolution examples are now beginning to appear (*e.g.*, Braile *et al.*, 1982; Adams *et al.*, 1991; Dawers and Seeber, 1991). We are particularly interested in characterizing the kind of structures that may be reactivated in the current tectonic stress regime (*e.g.*, Sbar and Sykes, 1973; Zoback and Zoback, 1989) and to test the hypothesis that structures of any age may be pertinent to intracratonic seismogenesis.

The area of study (Fig. 1) is part of the ENA platform characterized by Paleozoic sedimentary rocks lying nearly flat on Precambrian (Grenville) basement. The thickness of these rocks vary from a few hundred meters north of Lake Ontario to several kilometers in the Appalachian foreland. The sedimentary rocks of the platform are being mined throughout the region for oil, gas and salt and are being used as a repository of waste. Some of these activities were already under way early this century. These engineering endeavors often involve pumping large volumes of fluids in or out of deep wells; such activities alter circulation, pressure, and chemistry of fluids in the sedimentary rocks and in the upper part of the basement at seismogenic depths. Artificial changes in the hydrology of the upper crust alter mechanical conditions and can trigger seismicity. The closer the environment is to failure in the natural state, the more likely that artificially induced changes will trigger earthquakes (Nicholson and Wesson, 1990).

Several recent cases of artificially triggered seismicity have been well documented in the study area. Seismicity has been associated with salt brine recovery since 1971 at Dale (Attica), western New York (Fletcher and Sykes, 1977), oil/gas recovery since 1979 in Gobles, southern Ontario (Mereu, 1986), and waste disposal since 1987 in Ashtabula, northeastern Ohio (Armbruster et *al.*, 1987). Several other less convincing instances of induced seismicity have also been reported (*e.g.*, Nicholson and Wesson, 1990). These include

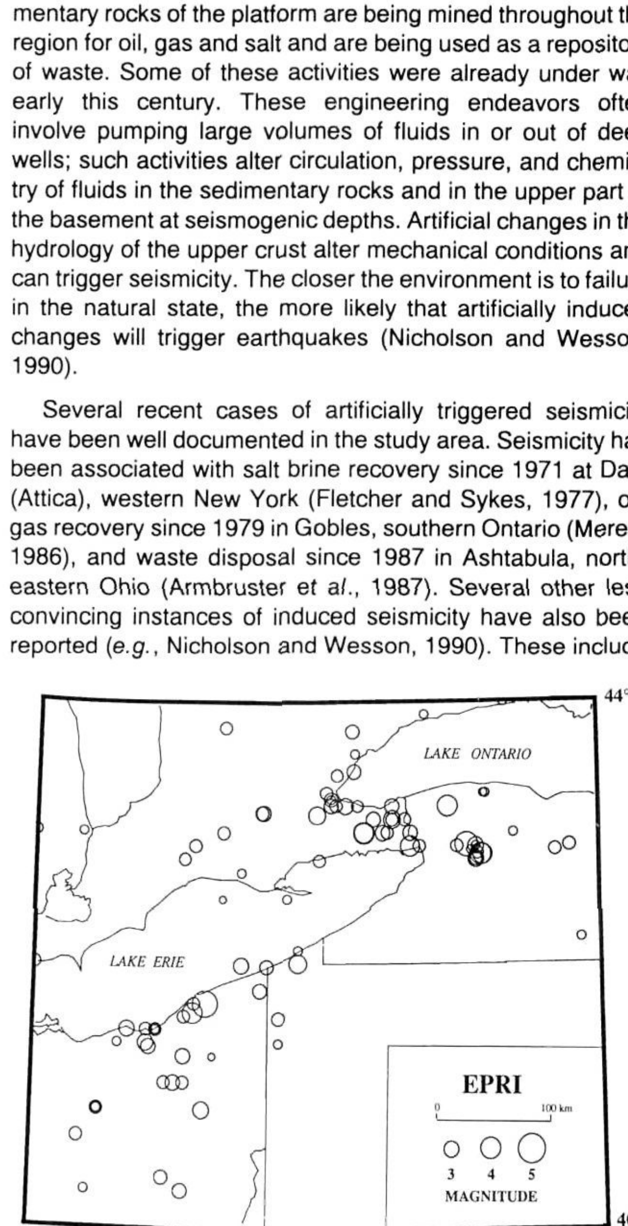


FIGURE 1. Epicenter map for the study area according to data available prior to our study. Included are all data available from the EPRI catalog for the historic period up to 1984 and from PDE 1985-1991.

Carte des épicentres de la région à l'étude établie à partir des données disponibles avant notre étude. Données du catalogue EPRI pour la période historique jusqu'à 1984 et de PDE pour 1985 à 1991.

recent well documented earthquakes where circumstances are ambiguous, such as the 1986 $m_{bLg} = 5.0$ Leroy earthquake, and earlier cases where engineering activities could have triggered seismicity, but available data are inconclusive. The 1929 $m_b = 5.2$ Attica earthquake may be an example of the latter. Confidence limits on its location (Dewey and Gordon, 1984) allow it to have originated near the Dale brine field which was already being hydraulically mined in 1929. Thus, it is possible that the two largest events in the study area were artificially triggered.

The overall space-time distribution of seismicity in the study area during the last half century may be biased by a substantial portion of artificially triggered earthquakes. Recognizing induced seismicity is important from the hazard viewpoint: first, because it offers the opportunity to reduce the hazard by reducing the seismicity; secondly, because it permits an unbiased record of natural seismicity; and thirdly because induced earthquakes occur in a partially controlled physical environment and offer a unique opportunity to explore the mechanical behavior of the upper crust.

EARTHQUAKE DATA

Maximizing spatial resolution and minimizing possible bias in the available earthquake data are essential first steps in the task of characterizing patterns of seismicity. In this paper, we reexamine archival sources, mostly newspapers, and recover new macroseismic data which improve constraints on source parameters and remove "noise" by identifying unreliable events. Our reexamination effort covers the period preceding seismic instrumentation as well as the early period of instrumentation because source parameters from early instrumental data can be quite misleading (Seeber and Armbruster, 1991) and constraints can be often improved with felt reports. By improving macroseismic coverage, magnitudes can be obtained from intensity distribution (*e.g.*, from felt area) rather than maximum intensity or from a single seismogram, which are notoriously unreliable measures of earthquake size. Finally, macroseismic data control the resolution of the data set as a whole because they tend to be less reliable than recent data from regional networks.

New macroseismic data obtained from archival searches caused substantial changes in the epicentral map (compare Figs. 1 and 2). Table I gives source parameters before and after modification brought about by this work. Three previously unknown, but significant ($M>3$), earthquakes in 1844, 1852 and 1855 were discovered, magnitudes and locations were revised for a substantial portion of the previously known earthquakes, and several events were recognized as explosions or as other circumstances leading to an earthquake report likely to be erroneous. Locations and magnitudes of earthquakes which occurred prior to the establishement of reliable seismic networks were derived systematically using MACRO (Figs. 3A-3L). MACRO is an algorithm, applied to earthquakes with felt reports from at least five different locations, that parametrizes intensity data in terms of location, magnitude, and the slope in a linear intensity-epicentral distance relationship (Armbruster and Seeber, 1987). Uncertainty is expressed as one-standard-deviation error bars; it depends on the number of available intensity points, on how well these points fit the model, and on a 'reading' error which we have assumed to be one intensity unit.

A large bias can result from an intensity coverage obtained with an incomplete search, particularly in areas where wide gaps in reporting are imposed by uninhabited regions, such as large bodies of water. A good example is the revision of source parameters for the 1857 earthquake. This event was originally located in northeastern Ohio, near Ashtabula (Fig. 3d). This location was assigned on the basis of felt reports exclusively from the southern shore of Lake Erie. Our archival search revealed previously unknown felt reports from the north shore of the lake. The areas and levels of intensity on the Canadian and US sides of the lake were similar. When the new data were included in a MACRO fit, the epicenter moved north by more than 50 km and a felt-area magnitude (M_{fa}) of 4.1 replaced a previous maximum intensity magnitude (M_{mi}) of 3.4 (Fig. 3d). This relocation placed the 1857 epicenter along the extrapolated northeast Ohio seismic zone and removed it from the Ashtabula area, where recent induced earthquakes are now the only known events (Fig. 4a).

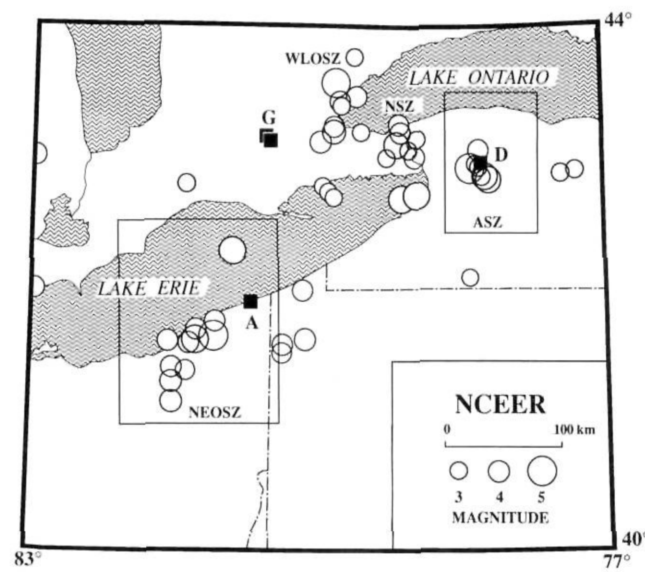


FIGURE 2. Epicenters of earthquakes listed in the NCEER-91 catalog (Seeber and Armbruster, 1991) with additions and modifications (see Table I). Events with $M \leq 3$ and/or reported felt only at one place and/or suspected to be non-earthquakes are excluded. Earthquakes inferred to be artificially triggered are represented with filled squares (A=Ashtabula; G=Gobles; D=Dale-Attica; ASZ=Attica seismic zone; WLOSZ=Western Lake Ontario seismic zone; NSZ=Niagara seismic zone; NEOSZ=Northeast Ohio seismic zone).

Épicentres des séismes compris dans le catalogue NCEER-91 (Seeber et Armbruster, 1991) avec des additions et des modifications (tabl. I). Les séismes de M ≤ 3 rapportés à un seul endroit ou mal identifiés sont exclus. Les séismes considérés comme ayant été provoqués sont représentés par des carrés noirs (A=Ashtabula; G=Gobles; D=Dale-Attica; ASZ= zone sismique d'Attica; WLOSZ= zone sismique de l'ouest du lac Ontario; NSZ= zone sismique de Niagara; NEOSZ= zone sismique du nord-est de l'Ohio).

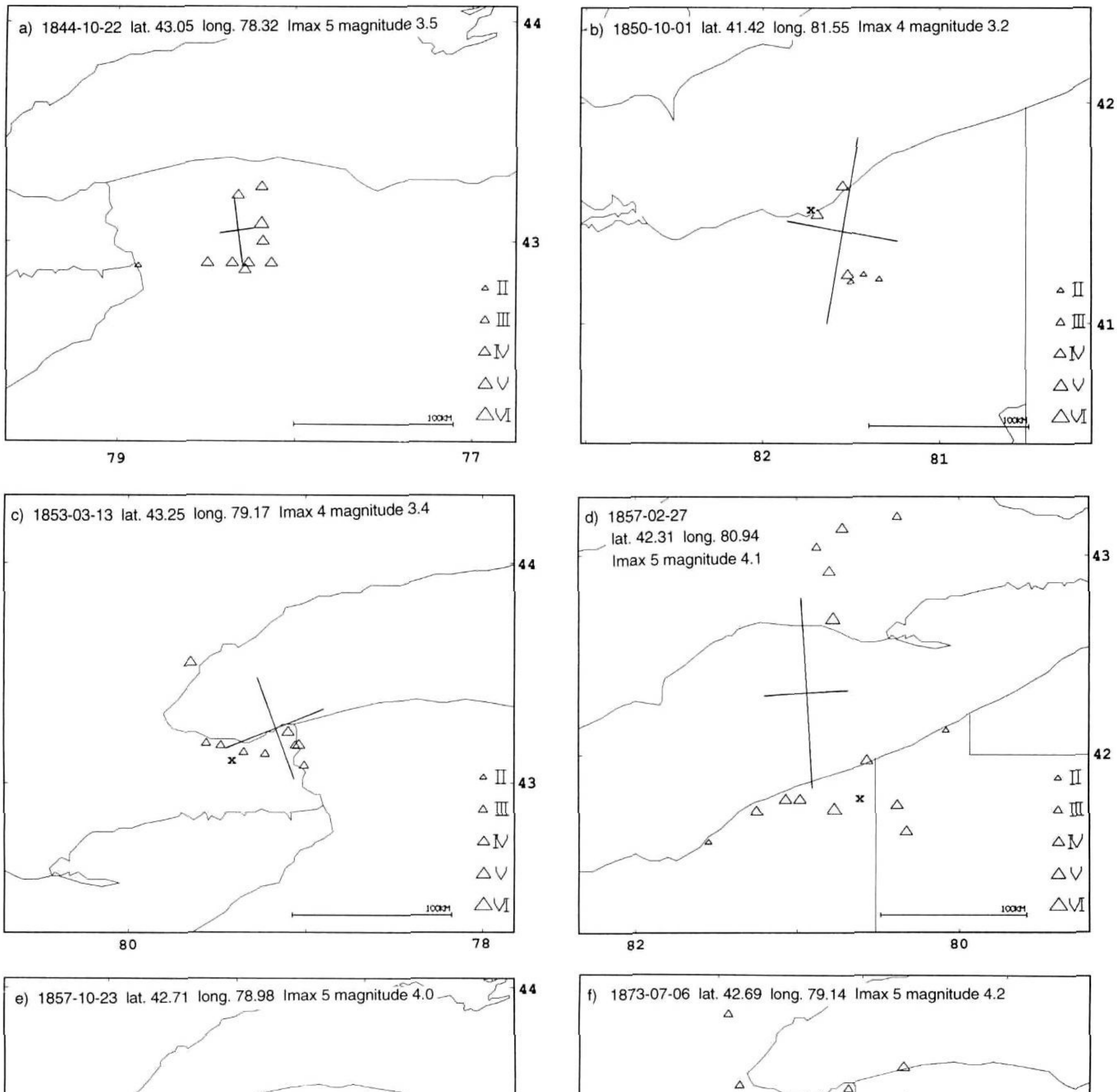
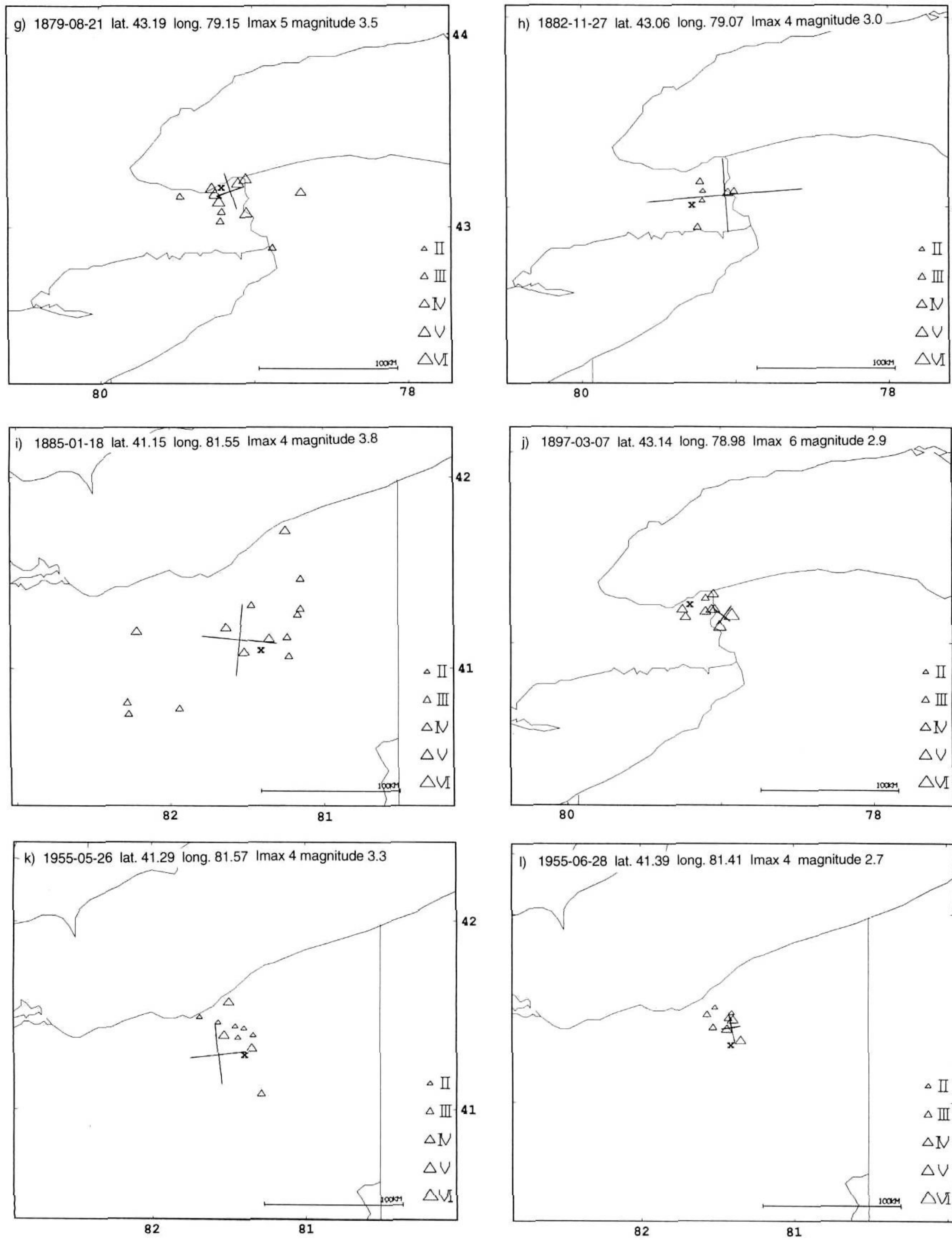


FIGURE 3. Relocated macroseismic epicenters for 12 historic earthquakes (a-l), each with at least five intensity data points (all available intensity data are plotted). Epicenters, magnitudes and error bars (one standard deviation) are obtained with the algorithm MACRO (see text). X's denote previous epicenters, as plotted in Figure 1. New source parameters are plotted in Figure 2.



Relocalisation de 12 épicentres macrosismiques (a-l), chacun ayant au moins cinq points de repères d'intensité (toutes les données ont été reportées). Les épicentres, les magnitudes et les marges d'erreur (un écart type) ont été déterminés à l'aide de l'algorythme MACRO (voir le texte). Les X localisent les épicentres antérieurs représentés à la figure 1. Les nouveaux paramètres apparaissent à la figure 2.

TABLE I

Changes in source parameters, including events interpreted as non-earthquakes

EPRI, 1987					THIS WORK					
Date	Lat.	Long.	M_{mi}	M_{fa}	Date	Lat.	Long.	M_{mi}	M_{fa}	Source
18230530	42.50	81.00	2.2		Doubtful report from only one location					3
	not known				18441022	43.05	78.32	3.5		1
18501001	41.50	81.70	3.1		18501001	41.42	81.55	3.2		2
	not known				18520915	41.63	80.17	3.7		1
18530313	43.10	79.40	3.4		18530313	43.25	79.17	3.4		1
	not known				18550917	42.08	78.43	3.0		1
18570301	41.80	80.60	3.4		18570227	42.31	80.94	4.1		2
18571023	43.20	78.60		4.5	18571023	42.71	78.98	4.0		2
18580410	41.70	81.30	3.1		18580410	41.70	81.30	3.3		3
18720723	41.40	82.10	2.5		7000 ton rock fall					3
18730706	43.00	79.50		4.6	18730706	42.69	79.14	4.2		2
18790821	43.20	79.20		3.5	18790821	43.19	79.15	3.5		2
18821127	43.00	79.25	3.1		18821127	43.06	79.07	3.0		2
18850118	41.10	81.40	3.1		18850118	41.15	81.55	3.8		2
18970307	43.10	79.20	3.1		18970307	43.14	78.98	2.9		1
19000409	41.40	81.80	3.7		Quarry blast					3
19060420	41.50	81.70	2.5		Felt only in part of Cleveland, two days after great California earthquake					1, 3
19060627	40.40	81.60	3.3		20 ton blast					3
19070412	41.50	81.70	1.9		Likely noise on seismographs in Cleveland: no felt reports					3
19280909	41.50	82.00		3.7	Three rumbles reported simultaneously with three 300 pound explosions					2
19290917	41.50	81.50	1.9		Felt only by one person					3
19300216	42.50	80.31	2.5		19300216	42.83	80.52	2.9		1, 3
19320122	41.10	81.50	3.5		Felt over a few adjacent streets and windows broken: not tectonic					1, 3
19341029	42.00	80.20	4.1		19341029	42.00	80.20		3.2	2
19390224	42.90	78.30	2.5		19390224	42.90	78.30		3.0	2
19511207	41.60	81.40	1.9		not on seismogram from Cleveland					3
19511222	41.60	81.40	1.9		not on seismogram from Cleveland					3
19540131	42.90	77.30	3.1		19540131	42.90	77.30		3.1	2
19550526	41.30	81.40	3.5		19550526	41.29	81.57		3.3	2
19550629	41.30	81.40	3.5		19550629	41.39	81.41		2.7	2
19580501	41.50	81.80	3.1		not on seismogram from Cleveland					3

1. New archival information
2. Seeber and Armbruster (1991)
3. Cleveland Electric Illuminating Company (1982)

SEISMIC ZONES AND PRE-EXISTING STRUCTURAL FEATURES

The data upgrade and quality selection processes leading from Figure 1 to Figure 2 have reduced the spatial scatter of epicenters. Excluding events for which the evidence of being artificially triggered is strong, most of the epicenters in Figure 2 are grouped in four distinct zones: the Northeastern Ohio seismic zone (NEOSZ); the Attica seismic zone (ASZ) in western New York; the Niagara seismic zone (NSZ) between Lake Erie and Lake Ontario; and the Western Lake Ontario seismic zone (WLOSZ) along the western shore of Lake Ontario. The first two seismic zones are correlated with major structural features, the Akron magnetic lineament and the Clarendon-Linden fault, and with the two largest known earthquakes in the study area, the $m_{bLg}=5.0$, 1986 Leroy earthquake and the $m_b=5.2$, 1929 Attica earthquake, respectively.

THE NORTHEAST OHIO SEISMIC ZONE (NEOSZ) AND THE AKRON LINEAMENT

Epicenters south of Lake Erie are concentrated along a 50+km long belt of seismicity striking northeast, the Northeast Ohio seismic zone (NEOSZ; Fig. 4a). Considering the large error bars of the macroseismic epicenters, a single fault could be the source of all the known $M \geq 3.0$ events, except the 1987 epicenter near Ashtabula believed to be associated with fluid injection and the 1991 epicenter near the

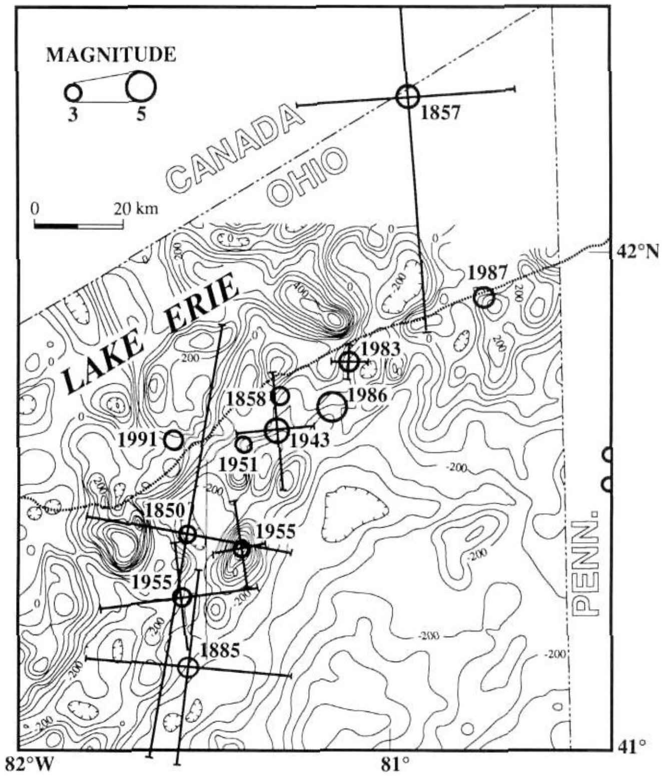


FIGURE 4a. The Akron lineament as represented by aeromagnetic data and M ≤ 3.0 epicenters. Macroseismic epicenters and related error bars have been determined with MACRO (see text). Error bars are smaller than the circles for most recent instrumental epicenters.

Le linéament d'Akron tracé à partir des données aéromagnétiques et des épicentres de M ≤ 3,0. Les épicentres macrosismiques et les marges d'erreur ont été déterminés à l'aide de MACRO (voir le texte). Les marges d'erreur sont plus petites que le cercle des épicentres déterminés très récemment.

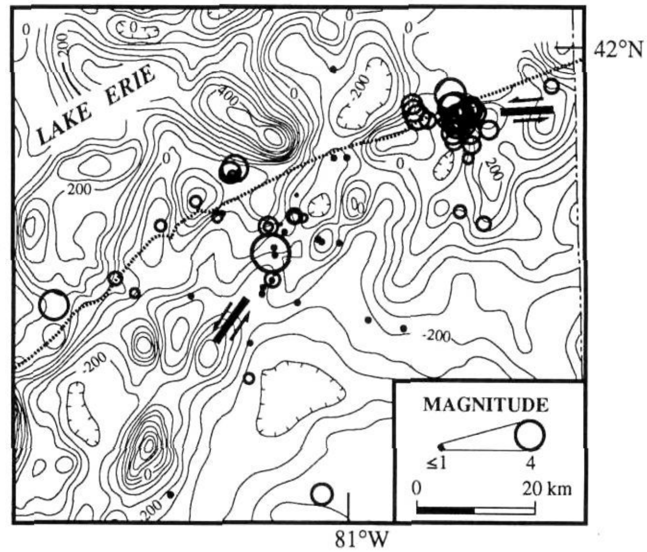


FIGURE 4b. 1986-1991 epicenters from the network operated by John Carroll University in Cleveland. Inferred rupture planes are shown for the 1986 Leroy earthquake (m_{bLg} = 5.0; Fig. 5) and for the 1987 Ashtabula earthquake sequence (m_{bLg} = 3.8; Fig. 7).

Les épicentres de séismes survenus de 1986 à 1991 déterminés par le réseau dirigé par la John Carroll University, à Cleveland. Les surfaces de rupture du séisme de Leroy (m_{bLg} = 5.0; fig. 5), en 1986, et de la succession de séismes d'Ashtabula (m_{bLg} = 3.8; fig. 7), en 1987, sont indiqués.

shore of Lake Erie. Some of the instrumental epicenters from recent M ≤ 3.0 events support this hypothesis, but others are broadly distributed (Fig. 4b).

Thus, the NEOSZ is elongated toward the northeast, but has a finite width. Such a belt of seismicity may be associated with a major fault and related secondary faults. This hypothesis is supported by detailed data from the 1986 m_{bLg} = 5.0 Leroy earthquake. The rupture of this earthquake, as inferred from the distribution of aftershocks and focal mechanisms, is vertical and strikes northeast (Fig. 5; also Nicholson *et al.*, 1988). The right-lateral slip on this plane is consistent with the east-northeast direction of the regional maximum compressive stress (*e.g.*, Zoback and Zoback, 1989). Small earthquakes since the Leroy sequence tend to locate along strike of the inferred rupture (compare Figs. 4b and 5) and near a set of deep injection wells for the disposal of hazardous waste fluid 12 km north of the Leroy epicenter. Some of this seismicity, including possibly the 1986 main shock, may have been triggered by the injection (Nicholson and Wesson, 1990).

The NEOSZ correlates spatially with the prominent Akron magnetic lineament (Fig. 4; Seeber and Armbruster, 1989) which coincides with a portion of the Akron Magnetic Boundary (Forsyth *et al.*, 1988, 1993). This feature represents a boundary between provinces with distinct magnetic signatures, high susceptibility and high relief on the west, low susceptibility and low relief on the east, probably reflecting different lithologies in the Precambrian basement (Lidiak *et al.*, 1985; Lucius and von Frese, 1988). A first-order structural boundary, interpreted from reflection data (discussed below), seems to be associated with this feature. Henceforth, "Akron lineament" will refer to the feature as expressed by the combined magnetic and seismic data sets.

In contrast with the prominent basement feature manifested by aeromagnetic data and by the NEOSZ, no structure corresponding to the Akron lineament has been detected in the overlying Paleozoic rocks. Correlation of abundant well data (*e.g.* Hansen, pers. comm.) and reflection profiles (Pratt *et al.*, 1989) have so far failed to detect vertical separation of stratigraphy in the ≈ 2 km thick platform cover rocks that may be associated with a fault along the lineament. A vertical separation of only a few tens of meters may go undetected and horizontal offsets cannot be resolved at all by these methods. A topographic lineament, likely to result from a fault with significant offsets of any kind, is not observed. Thus, a Precambrian age feature, expressed by the magnetic lineament, is correlated with the NEOSZ, which is a manifestation of ongoing deformation. Yet, no corresponding structure has been detected in the Paleozoic cover rocks. The lack of an observed structure limits the possible accumulated displacement on any single fault associated with the seismicity, not only during the neotectonic regime, but also from the early Paleozoic.

THE ATTICA, NY SEISMIC ZONE (ASZ) AND THE CLARENDON-LINDEN FAULT

The Clarendon-Linden fault was first postulated from surface data (Chadwick, 1919) and then characterized as a ≈10 km wide zone of offset bedding from subsurface data (Van Tyne, 1975). It is a regional feature, traced from the south shore of Lake Ontario southward for at least 100 km (Fig. 6). Yet, the vertical displacement on the main strand, which has a steep easterly dip, is only about 30 m. This feature is recognized as a fault below the Silurian Salina (evaporite) formation and as a broad monocline in the overlaying formations. This fold constrains the accumulated slip on the blind fault to be primarily dip-slip reverse. The folding is post-Devonian based on the youngest deformed rocks. Abrupt thickening of Ordovician strata east of the fault zone suggests an earlier phase of activity where the structure operated as a growth (normal) fault (Van Tyne, 1975). The complex development of the Clarendon-Linden structure is likely to extend back to Grenville time (Fakundiny *et al.*, 1978). The structure, as mapped in the Paleozoic platform formations, coincides

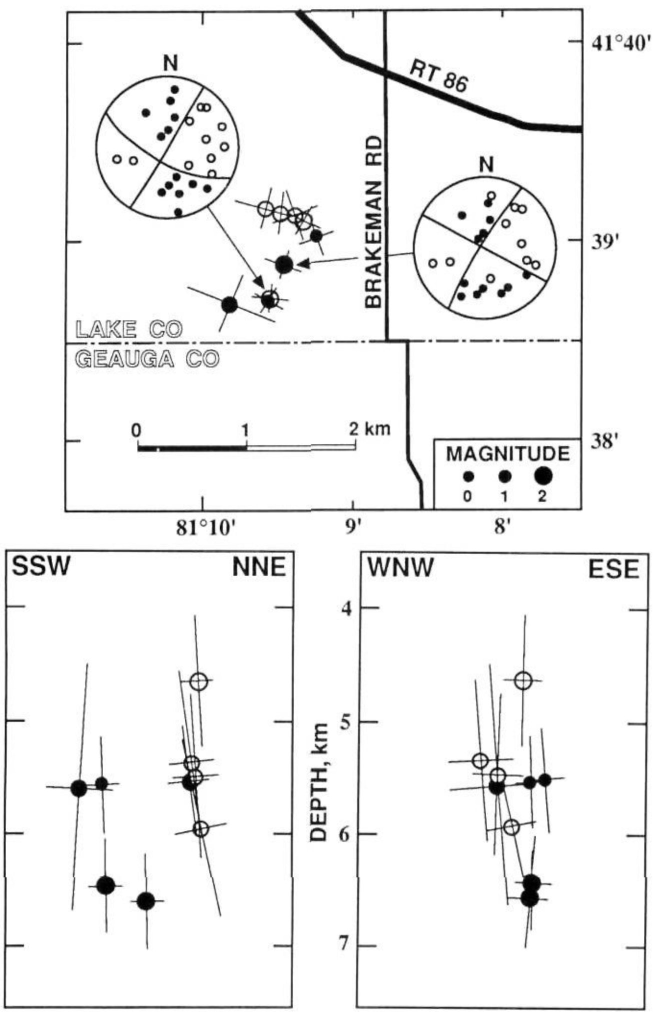


FIGURE 5. Map and two perpendicular vertical sections of the aftershock hypocenters of the 1986 m_{bLg}=5.0 Leroy, Ohio earthquake as located with data from a temporary local network. Early aftershocks suggest a vertical rupture about 1 km across and striking northeast. The depth range of the aftershocks places the rupture in the Precambrian basement, significantly below the unconformity at the base of the Paleozoic sedimentary rocks (from Seeber and Armbruster, 1989).

Carte des hypocentres et de deux coupes verticales perpendiculaires des répliques du séisme de Leroy (m_{bLg}=5,0), en 1986, déterminés par un réseau local temporaire. Les premières répliques indiquent une rupture verticale d'environ 1 km vers le nord-est. L'amplitude de la profondeur des répliques localise la rupture dans le socle précambrien, très au-dessous de la discordance à la base des roches sédimentaires du Paléozoïque (de Seeber et Armbruster, 1989).

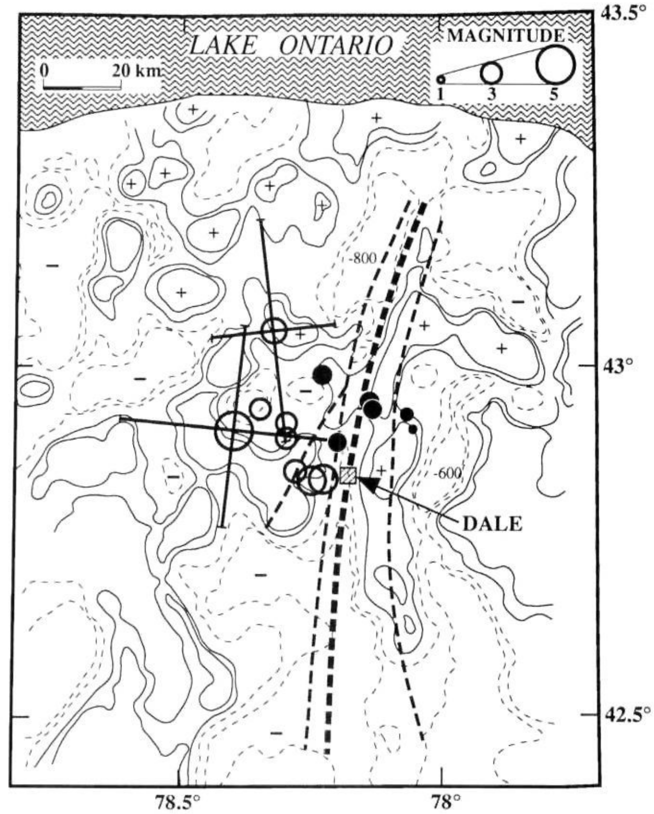


FIGURE 6. The Clarendon-Linden fault as mapped in Paleozoic sedimentary rocks (Van Tyne, 1975); aeromagnetic data (Hildebrand and Kucks, 1984), and epicenters in the Attica area of western New York; same epicenters as in Figure 2 (open circles) plus regional network data from the Lamont-Doherty Observatory in New York and the Geologic Survey of Canada (filled circles). The 1929, m_b=5.2 epicenter (to the south) and the 1844, M=3.5 epicenter (to the north) are the only ones with error bars (80% confidence, from Dewey and Gordon, 1984; and 75% confidence, from MACRO, respectively). These confidence limits allow them to be on the Clarendon-Linden fault zone. All other epicenters in this figure are similarly or better constrained and could also be in the fault zone.

La faille de Clarendon-Linden dans les roches sédimentaires du Paléozoïque (Van Tyne, 1975), les données aéromagnétiques (Hildebrand et Kucks, 1984) et les épicentres de la région d'Attica, à l'ouest de l'État de New York (les mêmes qu'à la fig. 2; cercles vides) ainsi que les données du Lamont-Doherty Observatory de New York et de la Commission géologie du Canada (cercles noirs). Les épicentres de 1929 (m_b=5,2; vers le sud) et de 1844 (M=3,5; vers le nord) sont les seuls comprenant des marges d'erreur (fiabilité à 80% de Dewey et Gordon, 1984, et fiabilité à 75% de MACRO, respectivement). Ces taux de fiabilité leur permettent de faire partie de la zone de Clarendon-Linden. Tous les autres épicentres sont aussi bien ou mieux déterminés et pourraient aussi faire partie de la zone de failles.

spatially with an aeromagnetic feature (Fig. 6). This magnetic feature can be traced beyond the mapped portion of the corresponding brittle fault zone, both northward into Lake Ontario (Hutchinson *et al.*, 1979) and southward to form a continental scale feature named the Amish anomaly (Culotta *et al.*, 1990). This linear magnetic structure is east of, and sub-parallel to, the Akron lineament.

Seismicity in the ASZ does not define a large elongated source zone correlated with the Clarendon-Linden fault as it does along the Akron lineament (Fig. 6). However, the association of earthquakes and the Clarendon-Linden fault seems very likely according to the high resolution data for induced earthquakes near Dale (Fletcher and Sykes, 1977). These earthquakes were concentrated in the immediate vicinity of the main strand of the Clarendon-Linden fault and were triggered only by injection at wells nearest the fault. Location uncertainties allow the rest of the earthquakes in the Attica area (Fig. 6) to originate from this fault zone as well. The Clarendon-Linden fault was named as the possible source of the $m_b = 5.2$, 1929 Attica earthquake in newspaper articles a few weeks after the event (Attica News, 29 August 1929). The 90% confidence limit of the instrumental epicenter reaches out to the fault (Dewey and Gordon, 1984). Intensity data for some of the aftershocks suggest a source within the fault zone. Both focal mechanisms for the two earthquakes in 1966 and 1967 (the two most southerly epicenters in Fig. 6) show steeply-dipping planes striking parallel to the Clarendon-Linden fault (Herrmann, 1978).

The epicenters in the ASZ are clustered in the vicinity of Dale, where fluid has been injected in the subsurface for brine recovery since late in the 19th century (Fig. 6). It is possible, therefore, that some of the known earthquakes prior to the well-documented induced seismicity in the early 1970's (Fletcher and Sykes, 1977), were also induced, including the $m_b = 5.2$, 1929 event (*e.g.*, Nicholson and Wesson, 1990). The newly discovered 1844 earthquake (Fig. 3a), however, preceeds injection operations and is clearly natural. Thus, the ASZ is likely to include both induced and natural seismicity.

THE WESTERN LAKE ONTARIO SEISMIC ZONE (WLOSZ)

A broad area of seismicity is centered between Lake Erie and Lake Ontario. The epicenters in this area are concentrated near the shores of both lakes and along the Niagara Gorge. This area is characterized by a high density of population and industrial plants; the possibility that some of the epicenters may reflect earthquakes triggered by engineering activities, such as the injection of fluid in deep wells, needs to be considered case by case. This task is not within the scope of this paper. Some of the earthquakes in this area occurred in the 1800's (*e.g.*, 1857, 1873, 1879, 1882, and 1897 in Fig. 3) and are clearly not induced.

The revised data in Figure 2 suggest two distinct seismic zones, one centered in the Niagara area, the NSZ, the other along the western shore of Lake Ontario, the WLOSZ (see also Mohajer, 1987, 1993). The WLOSZ appears to be elongated in a north-northeasterly direction, parallel to the regional basement trends, and to structures associated with the other seismic zones in the region. The Niagara-Pickering magnetic lineament (Wallach, 1990) is a prominent northeast-trending feature in this region; it is a possible extension of the Akron magnetic lineament (*e.g.*, Forsyth *et al.*, 1993) and is associated with brittle faulting in both the basement and Paleozoic cover rocks (Wallach, pers. comm.). In view of the evidence linking the Akron lineament with seismogenesis, the Niagara-Pickering magnetic lineament has been suggested as a possible source of earthquakes in the WLOSZ (Wallach, 1990; Mohajer, 1993).

THE NIAGARA SEISMIC ZONE (NSZ)

Several earthquakes occurred close to the Niagara River between Lake Ontario and Lake Erie (Fig. 2). Three of these have well-defined intensity patterns characterized by small intensity areas but relatively high maximum intensities (Figs. 3g, 3h, 3j). These intensity patterns suggest shallow hypocentral depths. The very rapid erosion of the Niagara River bed by the waterfall is forming a gorge downstream of the falls. The falls have been migrating upstream at about one meter per year (Van Diver, 1985, p. 45). This process is unloading the crust differentially and is expected to generate large topographic stresses in the vicinity of the gorge (*e.g.*, Huntoon and Elston, 1979). This stress may be responsible for shallow earthquakes near the gorge. Similar differential unloading as generated by the removal of rock at large quarries is believed to trigger earthquakes (Pomeroy *et al.*, 1976). However, not all the earthquakes in the vicinity of the Niagara Gorge could be accounted for by unloading stresses. Prominent among them are the 1857 (Fig. 3e) and the 1873 (Fig. 3f) events. They are located below Lake Erie, well to the south of the gorge, and they are not likely to be particularly shallow because they have a broad intensity field. In terms of source and cause, these events are probably distinct from the shallow seismicity near the Niagara Gorge.

GRENVILLE-AGE SUTURES ASSOCIATED WITH CURRENT SEISMICITY?

Culotta *et al.* (1990) correlated from the Adirondacks to Tennessee, a pair of oppositely dipping crustal-scale shear zones they recognize on several latitudinal reflection profiles. This continent-wide basement feature strike northeast, sub-parallel to the Grenville front. The east-dipping shear zone, on the west side of the pair, is associated with the west-verging collisional Grenville front. The shear zone on the east side of the pair, the Coshocton zone, dips to the west and was interpreted by Culotta *et al.* (1990) as another collisional suture, but with vergence to the east. They identified the Amish magnetic anomaly, their version of the New York-Alabama lineament of King and Zietz (1978), as the eastern boundary of the Coshocton zone. They subscribed to the King and Zietz (1978) hypothesis that prominent linear anomalies with regional scope, such as the Amish lineament, may mark terrane boundaries. The Clarendon-Linden fault was correlated with the portion of the Amish anomaly in western New York (Culotta *et al.*, 1990) and was thought to represent the reactivation of a basement structure (Hutchinson *et al.*, 1979; Fakundiny *et al.*, 1978).

Another prominent magnetic boundary (high to low intensity west to east) correlates with the western limit of the Coshocton zone in the COCORP Ohio profile (*e.g.*, Fig. 3 of

Culotta *et al.*, 1990). This step in the magnetic intensity profile is similar to the change in intensity across the Akron magnetic lineament and is along strike of it, about 100 km southwest of Figure 4. Thus, the Akron lineament may mark the western border of the Coshocton zone of west-dipping crustal reflectors. The Akron lineament, the Niagara-Pickering magnetic lineament (Wallach, 1990), and the Central Metasedimentary Belt boundary zone (*e.g.*, Forsyth *et al.*, 1988) have been considered as possible expression of the same regional basement structure (*e.g.*, Wallach, 1990). Such a regional continuity of the magnetic feature associated with the Coshocton zone is consistent with the interpretation of this zone as a continental scale suture (Culotta *et al.*, 1990).

The seismicity in our study area is generally associated with the Coshocton zone and, in particular, with the western (the Akron lineament) and eastern (Clarendon-Linden fault and Amish anomaly) boundaries of this zone. The Western Ontario seismic zone is aligned with an aeromagnetic structure adjacent to and west of the Niagara-Pickering magnetic lineament and may be in a similar relation to the Coshocton zone as the Akron lineament. The association between seismicity and prominent Precambrian structures persists west of the study area, where the Anna seismic zone is spatially associated with the Grenville Front (*e.g.*, Braile *et al.*, 1982).

In summary, the most prominent seismic zones in the study area appear to be correlated with the two most prominent Grenville-age structural boundaries as interpreted from geophysical data. This correlation is expressed spatially in map view; the data available are limited and a comparison between fault geometries inferred from the two types of data is premature. In any case, the Precambrian suture zones are broad features exhumed to their original ductile deformation levels. The geometry of brittle faulting within or near these zones may have a complex relationship with the geometry of the pre-existing ductile shear.

AN EARTHQUAKE SWARM TRIGGERED BY FLUID INJECTION IN ASHTABULA, OHIO: COHERENT SLIP ON A SINGLE FAULT

Changes in ambient mechanical conditions brought about by human activities can trigger earthquakes (*e.g.*, Simpson, 1986; Evans, 1988). A natural precondition for failure can account for the relatively small size of these changes. In most cases of triggered seismicity fluid conditions have been altered, either directly by fluid flow through wells or indirectly by the effect of a surface load. Human management of fluid circulation can add or remove loads and change pore pressure and differential stress — both well known factors affecting the likelihood of rock failure. Changes in the rate of fluid flow can also affect the rate of fault slip. The injection or removal of fluid by deep wells is particularly effective in causing locally high flow rates. This flow may be concentrated along faults where fracturing is intense and permeability is high. This condition may lead to accelerated pressure-solution creep causing raising stress and seismogenic ruptures on adjacent locked portions of the fault (*e.g.*, Chester *et al.*, 1993). Fluids are being extracted from, or injected into, the platform rocks of eastern North America by deep wells for a variety of purposes, including oil/gas extraction, salt mining, and waste disposal. Induced seismicity has been reported in the study area from each of these types of activities (*e.g.*, Mereu *et al.*, 1986; Fletcher and Sykes, 1977; Armbruster *et al.*, 1987; Nicholson and Wesson, 1990).

The 13 July 1987, $m_{bLg} = 3.8$ Ashtabula earthquake (Fig. 4) is one of the clearest examples of a macroseismic event triggered by fluid injection in a deep waste disposal well (Nicholson and Wesson, 1990; Armbruster *et al.*, 1987). This event generated relatively abundant aftershocks, which we locally monitored for ten days beginning two days after the main shock. The aftershocks, which we refer to as the Ashtabula sequence, occurred near an injection well that had been in operation for only a year, since July 1986. The zone of injection for this, and for other deep waste-disposal wells in northeastern Ohio, is in the Mt Simon Formation, a permeable sandstone at the base of the sedimentary rocks. This well is less than 1 km from the July '87 hypocenters (Fig. 7). No prior earthquakes are known to be located within 30 km of Ashtabula. The close spatial and temporal correlation between the injection well and seismicity is strong evidence for causality.

The 36 hypocenters of the Ashtabula sequence, all of which are well determined, cluster in a narrow east-striking vertical zone about 1.5 km long and extending from a depth of 1.7 km to 3.5 km (Fig. 7). The combined hypocentral and first-motion data delineate a vertical, active fault, which we name here the Ashtabula fault. First-motions fit a well-constrained composite fault-plane solution characterized by a left-lateral nodal plane that matches closely the attitude of the hypocentral zone (Fig. 8), and is consistent with the east-northeast direction of the greatest principal horizontal compressive stress (Zoback and Zoback, 1989). The Appalachian Plateau in northeastern Ohio is characterized by sub-horizontal Paleozoic sediments (4.5 km/sec) above Precambrian basement (6.0 km/sec). The depth of the unconformity in the epicentral area is 1.8 km. Thus, seismicity is concentrated near the top of the basement; considering location uncertainties (Fig. 7) most, and possibly all, of the July 1987 hypocenters were located below the unconformity.

In an isotropic medium, mechanical changes induced by injection are expected to decrease rapidly away from the injection point. Since the Ashtabula fault is located 0.7 km from this point, a distance clearly exceeding the confidence limits of the locations (Fig. 7), it is likely to be a reactivated pre-existing fault rather than a new fault, which may be expected to radiate from the well. No evidence for such a fault has yet been reported from geologic data. A reflection survey for the purpose of studying the structural environment of the well apparently did not resolve the seismogenic fault (Ashtabula Star Beacon, 9/24/92). According to the same report, the well operators concluded from the reflection data that "fault lines are not located near the area and the earthquakes are not related to the operation of the injection well".

If the preliminary results of the reflection study, as reported by the press, are confirmed, the failure of the reflection data to exhibit the fault may place a low limit on the maximum possible vertical accumulated displacement on the fault. The

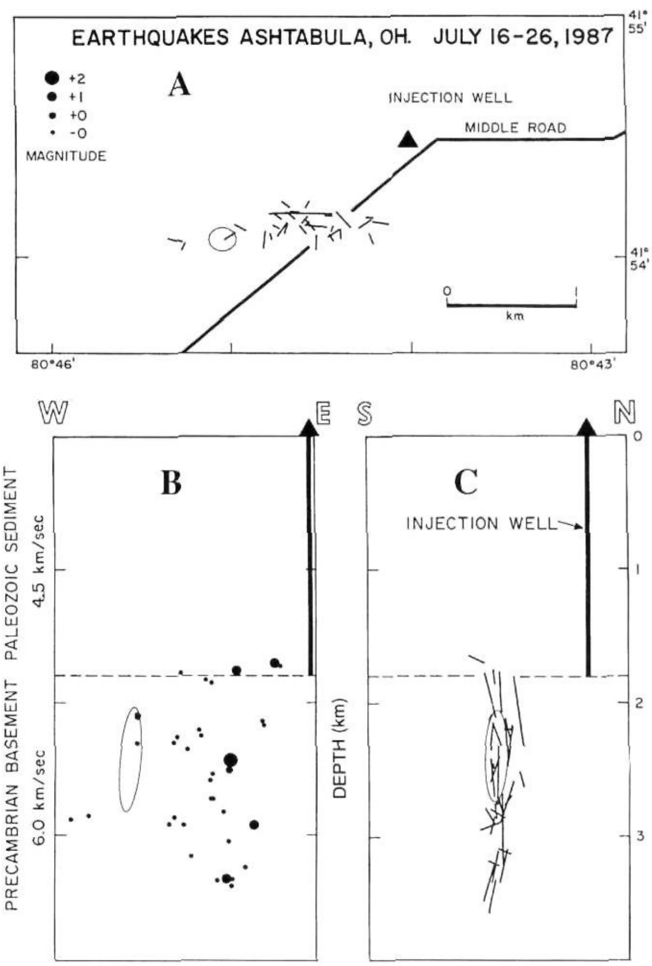


FIGURE 7. Map and vertical sections of the seismicity in Ashtabula following the June 1987, $m_{bLg}=3.8$ event, the first known earthquake within 30 km. The position of the class 1 injection well operated by Reserve Environmental Services is indicated by the triangle and the vertical bar. This facility pumps waste fluids into the Mount Simon, the basal formation just above the unconformity between Precambrian basement and platform rocks at a depth of 1.8 km (dashed line). The 35 hypocenters (obtained with the two-layer 1-D velocity model shown in B) were quality selected from a total of 54 events recorded by a temporary network operated for ten days (Fig. 10). Line segments in A and C connect two hypocenters obtained for each of the events with two independent sets of phase readings; they are a realistic indication of confidence limits. The HYPOINVERSE error ellipse represents an event recorded in August 1989 during a 5-day re-deployment of the local network; it is representative of formal confidence limits for the other hypocenters. Note that the size of the 1987 (main shock $m_{bLg}=3.8$) and 1986 (main shock $m_{bLg}=5.0$; Fig. 5) aftershock zones are similar.

Carte et coupes de la sismicité à Ashtabula, après le séisme de $m_{bLg}=3,8$ survenu en juin 1987, le premier séisme connu à l'intérieur de 30 km. L'emplacement du puits d'injection (Reserve Environmental Services) est indiqué par un triangle ou une ligne verticale. On y injecte des liquides usés dans le mont Simon, la formation de base juste au-dessus de la discordance entre le socle précambrien et la plateforme rocheuse, à une profondeur de 1,8 km (tireté). Les 35 hypocentres (déterminés à l'aide du modèle illustré en B) ont été choisis à partir d'un ensemble de 54 séismes enregistrés pendant une période de 10 jours (fig. 10). Les segments en A et C relient deux hypocentres déterminés à chacun des séismes à deux ensembles indépendants de relevés de phases; ils fournissent une bonne indication des limites de fiabilité. L'ellipse d'erreur représente un séisme enregistré en août 1989 par le réseau local temporaire; elle est représentative des limites de fiabilité pour les autres hypocentres. Noter que la dimension de la zone des répliques de 1987 (séisme de $m_{bLg}=3,8$) et de 1986 (séisme de $m_{bLg}=5,0$) est semblable.

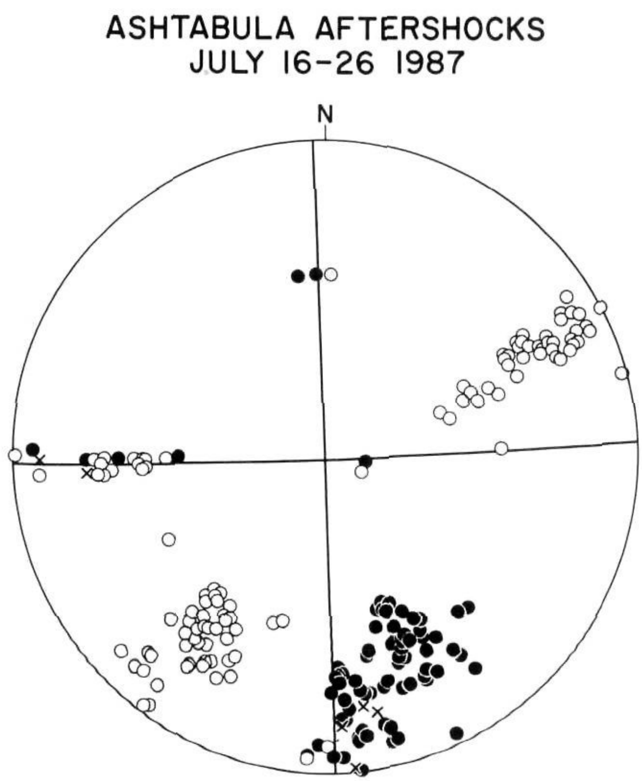


FIGURE 8. Composite focal mechanism for the events shown in Figure 7. Filled dots, open dots and crosses are compressions, dilatations and nodal first-motions, respectively. The proposed mechanism is tightly constrained by the first-motions and is consistent with all of the data (after allowing for some scatter). The east-west left-lateral nodal plane is parallel to the plane defined by the hypocenter distribution in Figure 7 and is chosen as the fault plane.

Mécanisme au foyer composite des répliques représentés à la figure 7. Les cercles noirs, les cercles vides et les croix représentent respectivement les compressions, les dilatations et les points nodaux des premiers mouvements. Le mécanisme proposé est étroitement déterminé par les premiers mouvements et est compatible avec toutes les données. Le plan nodal est-ouest est parallèle au plan défini par la répartition des hypocentres de la figure 7 et constitue le plan de faille.

Ashtabula fault is currently active in a strike-slip mode; this deformation would not contribute to vertical offset, which is usually necessary to image a vertical fault on a reflection profile. In general, lack of evidence for seismogenic faults on reflection profiles is not surprising considering, for example, similar negative results in the source areas of the 1886 Charleston SC and the 1811-12 New Madrid earthquakes (*e.g.*, Hamilton *et al.*, 1983; Crone, 1992), the largest known earthquakes in eastern US. Data from reflection profiles cannot be trusted to delineate all faults that may be seismogenic and therefore cannot provide evidence against the possibility or the occurrence of induced seismicity.

Following the 1987 sequence, earthquakes have continued to occur in Ashtabula; the latest felt earthquakes were in March 1992 (Fig. 9). These earthquakes have not been monitored with local instruments and constraints on their locations are much poorer than for the ones in 1987. Nevertheless, all the available data combined, including felt reports for the 1992 earthquakes (Fig. 10), suggest that: 1) the seismicity may be originating from an east-west zone 5-10 km long and possibly from the same fault identified from the 1987 data; and 2) the seismicity is migrating westward away from the well. The hypothesis of a rapid westward migration of seismicity along a single fault is consistent with the concept that fluid injected into the basal Paleozoic sandstone has penetrated the basement and is flowing near the unconformity along a high-permeability pre-existing fault. This and other hypotheses relevant to induced seismicity in general as well as to the specifics of Ashtabula could be tested by locally monitoring the seismicity and by modeling the pressure history at the well (*e.g.*, Nicholson and Wesson, 1990).

The possible concentration of Ashtabula seismicity on a single basement fault has important implications for earthquake hazard assessment. First, the combined data from 1987 (Figs. 7 and 8) and from 1992 (Fig. 10) are consistent with a single active fault that may be 5-10 km long and is slipping coherently in response to the regional stress field. Conceivably, such a fault could rupture in a single event and generate an earthquake in the M=5-6 range. Secondly, if the fluid is migrating away from the injection area by a quasi 1-dimensional flow along the upper reaches of a steep basement fault, then effects of the injection, including induced seismicity, may occur at distances from the well much larger than for two-dimensional flow (Nicholson and Wesson, 1990). Thirdly, induced slip on the vertical basement fault may extend into the overlying sedimentary rock locally increasing permeability. The opening of fractures caused by the deformation may breech flat-laying aquicludes and may increase the risk of contamination of shallow aquifers by the injected waste fluid.

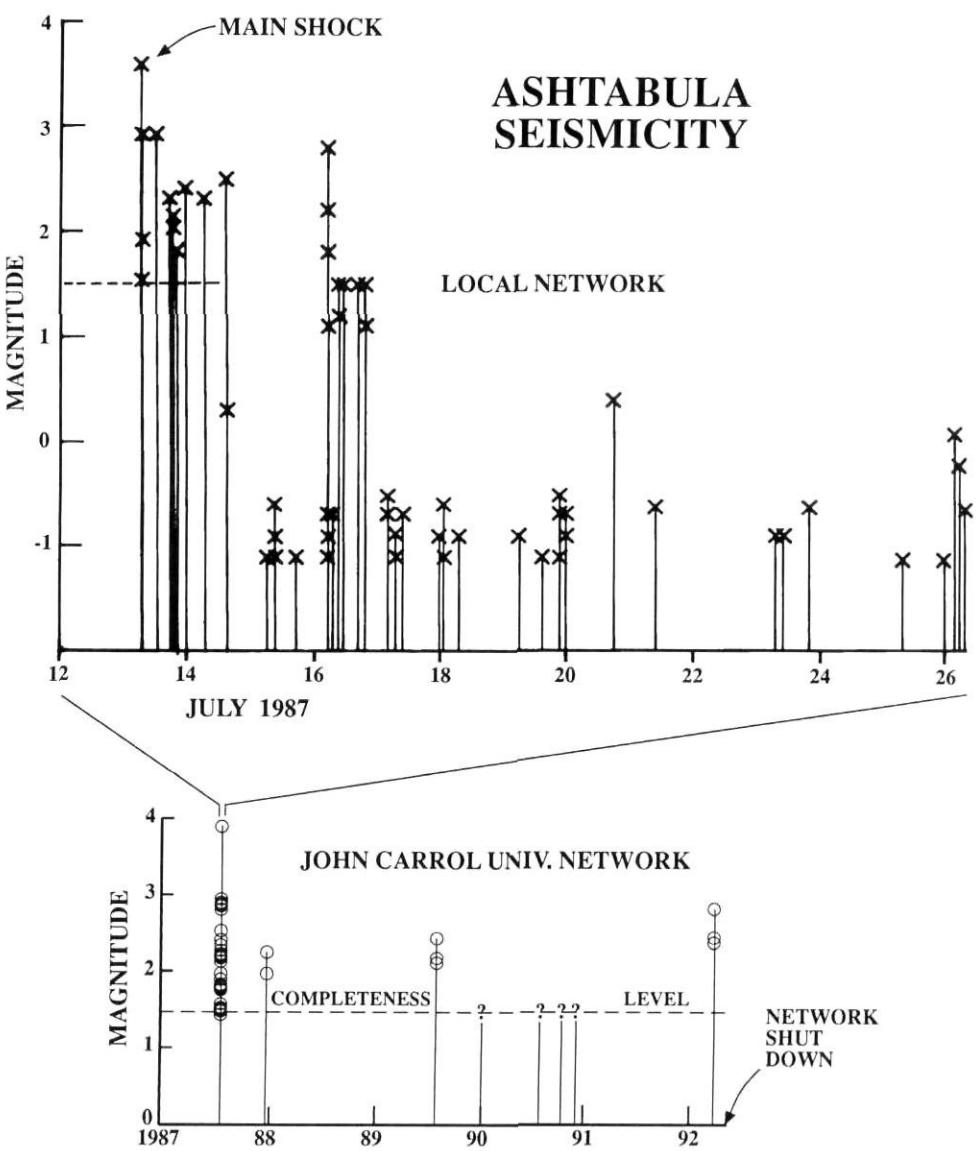


FIGURE 9. Histograms of seismicity at Ashtabula. Above: data from the local network deployed following the first known and largest earthquake so far at Ashtabula, the m_{bLg}=3.8, July 13, 1987 event; Below: data from the John Carroll University network which was centered east of Cleveland and west of Ashtabula. Magnitude threshold for detection by this network at Ashtabula (M≈1.5) is similar to the felt threshold (M≈2.0).

Histogrammes de la sismicité à Ashtabula. En haut: données du réseau mis en place à la suite du premier et plus important séisme connu à Ashtabula (m_{bLg}=3.8), survenu le 13 juillet 1987. En bas: données du réseau régional de la John Carroll University, centré à l'est de Cleveland et à l'ouest d'Ashtabula. La limite de détection par ce réseau (M≈1,5) est semblable à la limite de perception (M≈2,0).

The limited earthquake data available in Ashtabula were collected fortuitously; nevertheless, since 1987 they have made a strong case for triggered seismicity at that site (*e.g.*, Nicholson and Wesson, 1990). Yet, claims to the contrary were made repeatedly by the operator of the injection well (*e.g.*, Yogi V. Chokshi of Reserve Environmental Services as quoted in the Ashtabula Star Beacon, 9/24/92) and no earthquake monitoring program was undertaken while the injection operation has continued. Less obvious cases of induced seismicity could easily go unnoticed, particularly if they occurred in conjunction with unwillingness to address openly such phenomena. A concern for adverse publicity and potential liabilities by enterprises operating facilities that affect underground fluid circulation may be responsible for the apparently low level of interest in studies of induced seismicity in Ashtabula and elsewhere. The resulting lack of data may also be a cause for underestimating the level of induced seismicity and, consequently, for overestimating the level of natural seismicity in the affected areas.

DISCUSSION

SEISMICITY IN THE VICINITY OF A PRE-EXISTING FAULT – IS REACTIVATION THE ONLY OPTION?

Intracratonic seismogenesis, as represented by the data in our study area, seems to be generally correlated with pre-existing ductile features in the basement. In the simplest interpretation, seismicity may indicate a reactivation of Precambrian faults in the neotectonic stress regime. Alternatively, it may reflect small volumes of stress and/or weakness in the vicinity of pre-existing faults which, however, may themselves not be reactivated. Petrology, fracture concentration, fluid flow, fluid chemistry and other factors pertinent to the mechanical behavior of the rock are often different in the neighborhood of faults, both active and inactive ones, than in the surrounding country rock. Typically, these differences are such that pressure solution and/or other processes leading to permanent aseismic deformation in the upper crust would be enhanced in the neighborhood of faults (*e.g.*,

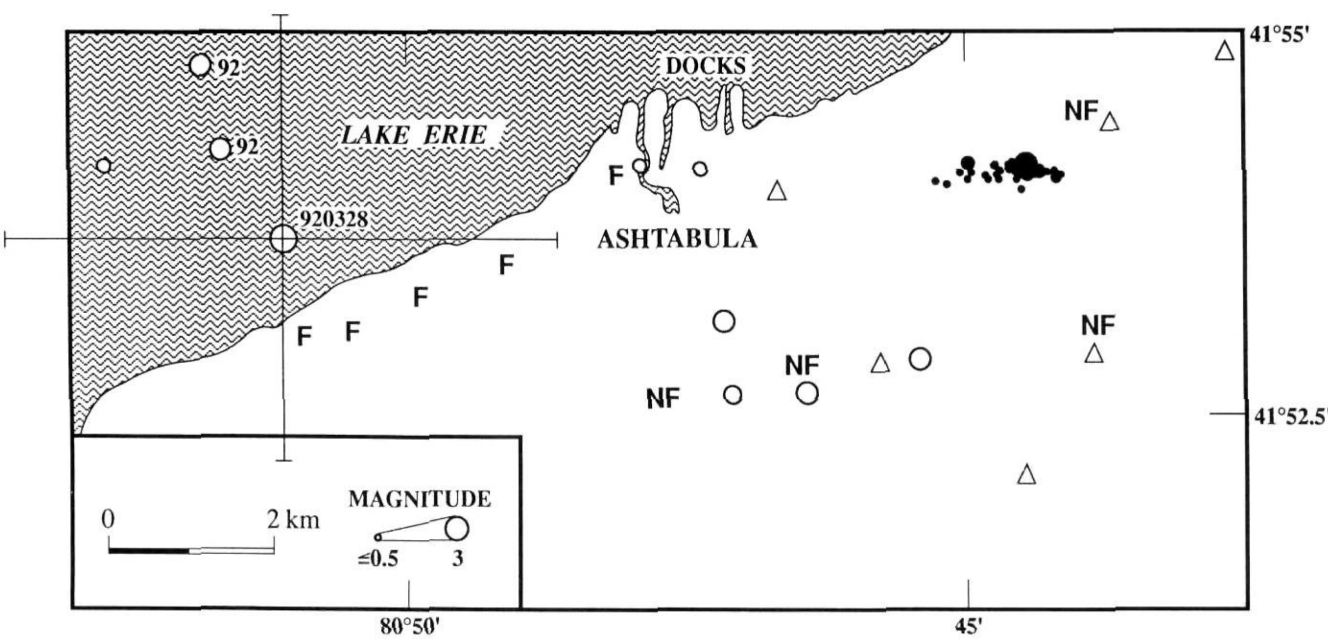


FIGURE 10. Map of epicenters in the Ashtabula area and locations of felt (F) and not felt (NF) reports for the M=2.9, March 28, 1992 event. Filled circles denote epicenters from the local network (triangles) operated in July 1987 (the same data as in Fig. 7). Open circles denote epicenters from the John Carroll University regional network (all stations are south and west of the area shown) for the period following the temporary local network (August 1987-March 1992). Average 1-sigma error bar (shown on 92/03/28 epicenter) is deduced by assuming pre-local network epicenters to be located at the source defined by the black epicenters. Thus confidence limits allow all events to be located on a westward extrapolation of the fault delineated by the filled epicenters. But the 1992 epicenters are significantly to the west of most of the previous epicenters. A similar conclusion is required by the distribution of felt reports for the M=2.9, March 28, 1992 event. Felt reports for the other two events in 1992 are similarly distributed.

Carte des épicentres de la région d'Ashtabula et localisation des rapports de perception (F) et et de non-perception (NF) du séisme de M= 2,9, du 28 mars 1992 (site et marge d'erreur). Les cercles noirs identifient les épicentres déterminés par le réseau local en juillet 1987 (même données qu'à la fig. 7). Les cercles vides identifient les épicentres enregistrés par le réseau régional de la John Carroll University (les stations étant toutes situées au sud et à l'est de la région) après l'arrêt du réseau local (août 1987-mars 1992). La moyenne de 1 sigma de marge d'erreur a été établie en présumant que les épicentres des séismes survenus avant la mise en marche du réseau local sont situés aux foyers identifiés par les épicentres noirs. Ainsi, les limites de fiabilité permettent l'inclusion de tous les séismes dans le prolongement vers l'ouest extrapolé à partir de la faille délimitée par les cercles noirs. Toutefois, les épicentres de 1992 sont situés très à l'ouest de la plupart des autres épicentres. On doit également faire la même extrapolation à partir de la répartition des rapports de perception du séisme de M=2,9, survenu le 28 mars 1992. Les rapports de perception pour les deux autres séismes de 1992 sont répartis de façon semblable.

Chester *et al.*, 1993). Volumes yielding, aseismically, to ambient stress may cause stress concentrations in adjacent rocks and lead to brittle failures. Such volumes of permanent deformation, however, could be sufficiently isolated from each other so as not to interact and bring the pre-existing fault, as a whole, significantly closer to failure.

Seismicity spatially associated with a large pre-existing fault is consistent with the hypothesis of a large possible earthquake. This hypothesis is substantiated if the fault is reactivated and a single rupture over the entire fault is possible. If, however, seismicity is the consequence of the growth of stress around localized volumes of aseismic deformation, the maximum magnitude of such sources could reflect the size of such anomalies and be much smaller than the magnitude expected from a rupture of the entire fault. For a given level of historic low-magnitude seismicity, the long-term moment-release and fault-slip rates are also expected to be larger in the case where coherent motion does occasionally occur on a large fault patch than in the case where seismicity stems solely from localized stress concentrations or weaknesses.

SMALL OFFSETS ON NEOTECTONIC FAULTS VERSUS HIGH MOMENT RELEASE RATES

Small offsets accumulated during the neotectonic regime seem to characterize seismogenic faults that are known to have generated large intracratonic earthquakes. The two sources of the largest known earthquakes in the eastern US, in 1811-12 New Madrid, MO and in 1886 Charleston, SC, are in areas covered by flat-lying Mesozoic and younger sedimentary rocks that show no evidence of accumulated fault offsets larger than a few tens of meters (*e.g.*, Hamilton *et al.*, 1983; Crone, 1992; Luzietti *et al.*, 1992). Several dip-slip earthquakes in Australia have ruptured the surface along faults that have no geomorphic or paleoseismic evidence of prior activity (Crone *et al.*, 1992). The Meers fault in Oklahoma ruptured the surface in one or more large prehistoric earthquakes, but again it shows no evidence of significant accumulated neotectonic displacement (Ramelli *et al.*, 1987). Finally, the 1989, M_s=6.3 Ungava earthquake in northern Québec was also associated with surface rupture controlled by an Archean age fault exhumed to its ductile deformation level. No brittle shear failure apparently preceded the 1989 rupture (Adams *et al.*, 1991). In general, the lack of evidence for substantial accumulated neotectonic displacements on structures that have produced large intracratonic earthquakes suggests that the long-term return rate of such earthquakes from these sources is very low. Furthermore, if the historic rate of large earthquakes is representative of the neotectonic regime, the known sources may be only a small portion of the total number of sources contributing to the long term rate of large earthquakes.

Thus, the absence of evidence for significant accumulated neotectonic offsets on faults that appear to be seismogenic, such as the Akron lineament and the Clarendon-Linden fault, do not rule out the possibility of large earthquakes from these faults. Geologic constraints on the long-term behavior of a fault can be used to estimate an upper limit to the number of such hypothetical earthquakes. For example, only a few tens of $M \approx 7$ earthquakes are permissible on a 50 km long fault with a maximum neotectonic offset of 100 m. These earthquakes would be expected to occur in less than 1 m.y., a period 1 to 2 orders of magnitude shorter than the duration of the current neotectonic regime (which may be as long as 100 m.y. or more), if such a fault produced seismicity at a rate similar to the Clarendon-Linden fault or the Akron lineament, with a typical frequency-magnitude distribution and a maximum magnitude $M \approx 7$. Alternatively, the maximum magnitude on the hypothetical fault would have to be similar to the maximum historic magnitude in the area of this study ($M \approx 5$), in order for the displacement on the fault to remain small over a period of accumulation comparable to the duration of the intracratonic regime. A low-magnitude-limit, however, cannot be a general characteristic of earthquake sources in eastern North America, if these sources are to account for the overall historic rate of large earthquakes.

Under the assumption that the pattern of seismicity is stationary, the long-term rate of $M \geqslant 7$ main shocks in eastern North America would be the same as the historic rate of about two per century (*i.e.*, 1811, 1886, 1929, and 1933). If the maximum number of repeated $M \geqslant 7$ ruptures that would generate cumulative fault offsets invisible to standard seismic reflection imaging is only on the order of several tens of meters, then many thousands of distinct faults would be needed to account for the seismicity during a neotectonic period of several tens of m.y. At an average density, many tens of these faults would be expected in the area of this study. Such a high density of potentially active faults capable of large earthquakes may not be acceptable because the faults currently active seem to correlate with prominent pre-existing structures, which are limited in number. Then, non-stationary seismic sources may be the way to account for the historic rate of moment release and the small accumulated fault offsets; the long-term neotectonic moment release rate may be lower than the historic rate. Induced seismicity may have inflated the latter portion of the historic record. Furthermore, the temporal proximity of our observation window to a major deglaciation event that caused a drastic redistribution of mass on the earth's surface could be responsible for an unusually high rate of seismicity in the Holocene (*e.g.*, Hasegawa and Basham, 1989).

Turning the above argument around, the seven known or suspected sources of $M \geqslant 7$ earthquakes in eastern North America (Meers fault; St Lawrence Valley; New Madrid, MO; Charleston, SC; Baffin Bay; Grand Banks; Wabash Valley) could continue to produce $M \geqslant 7$ events at the current rate (a few tens of events) for only several millennia, if the net offset on each of the main faults in these sources was to remain consistent with current observations. Whether or not the current seismicity rate is representative of the long-term rate, we will probably experience the current rate of seismicity for some time and the next fault producing a $M \geqslant 7$ earthquake in eastern North America may well be in a source zone not known to have generated such an earthquake before.

CONCLUSIONS

1. Two of the most prominent seismic zones in the Lake Erie-Lake Ontario area are associated with the Akron lineament and the Clarendon-Linden fault. These features are delineated by regional northeast striking aeromagnetic anomalies believed to reflect major Grenville-age structural boundaries in the basement. No Paleozoic or younger fault is yet recognized along the Akron lineament. Post-Grenville tectonism along the Clarendon-Linden fault is manifested by two distinct phases of brittle deformation during the early Paleozoic, and by the current seismicity. The large size of these pre-existing structures suggests upper magnitude limits higher than the historic $m_{bLg} = 5.0$ and $m_b = 5.2$ related to these structures, respectively.

2. In a pattern common to most faults associated with large intracratonic earthquakes worldwide, Paleozoic and younger maximum possible displacements accumulated on the Clarendon-Linden fault and the Akron lineament are small. These structures could not account for many large earthquake ruptures (unless the sense of motion can be reversed during the neotectonic regime).

3. In general, the small accumulated offsets on seismogenic intracratonic faults, including sources of known large earthquakes, is in apparent conflict with the observed moment release rate or with the rate of $M \geqslant 7$ earthquakes. These geologic and seismologic data can be reconciled by either allowing for many potentially active faults, many more than the ones that generated large earthquakes during the historic period, or by permitting the overall rate of moment release to vary in time, *i.e.*, by a non-stationary seismicity. The latter seems a more acceptable solution in view of the association of known earthquake sources, such as the Akron lineament and the Clarendon-Linden fault, with major pre-existing structures.

4. Seismicity near Ashtabula defines a vertical, east-west, left-lateral fault in the basement, just below the platform rocks. This seismicity began in 1987, one year after the onset of fluid waste injection in a well which penetrates the rock formation to a depth just above the 1.8 km deep basement and is located less than 1 km from the seismicity. From 1987 to 1992, the seismicity appears to have migrated away from the well out to a distance of 5-10 km. All well constrained hypocenters fall on the same vertical planar zone, less than 0.25 km wide, which is close to the well, but does not radiate from it. This zone is inferred to represent a pre-existing basement fault brought to failure by the fluid flow and/or increased pore pressure induced by fluid injection. Fluid diffusion from the well may be controlled by high permeability along the active fault. As injection at the Ashtabula well continues, there is an urgent need to monitor the behavior of the active fault by recording earthquakes with a local seismic network.

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