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Seismicity of the Southern Great Lakes: Revised Earthquake Hypocenters and Possible Tectonic Controls

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Abstract Using data from 27 seismograph stations for the period 1990–2001, we have relocated 106 hypocenters of earthquakes with magnitudes from 0.9 to 5.4 in the region of the southern Great Lakes. Two complementary methods were used for relocation: a conventional least-squares approach (Lienert and Havskov, 1995) and joint hypocentral determination (Pujol, 2000). These two methods yielded mutually consistent spatial patterns of seismicity with an average difference of 3.7 km in epicentral locations and 1.1 km in focal depths. We show that the hypocenter locations are not very sensitive to realistic uncertainties in 1D crustal velocity. Our sharpened definition of zones of seismicity delineates several clusters beneath Lake Ontario, around Niagara Falls, and near the south shore of Lake Erie. These seismicity zones appear to correlate with areas where the regional magnetic data exhibit prominent short-wavelength (<5 km) linear anomalies. The magnetic anomalies are associated with basement structures that formed during the Precambrian (Mesoproterozoic) Grenville orogen. Both the seismicity and magnetic anomalies exhibit statistically significant preferred orientations at N40°E–N45°E, but the correlation of the earthquake clusters with specific aeromagnetic lineaments remains uncertain. Three preliminary focal mechanisms of earthquakes with magnitudes m_N 3.1 to 3.8 show unusual normal faulting, with nodal planes in almost the same direction as the magnetic trends, N42°E–N52°E. Proximity of the earthquake clusters to large bodies of water, coupled with colinearity with magnetic anomaly trends, suggests that both surface water and pre-existing basement structures may play significant roles in controlling intraplate seismicity in the southern Great Lakes region.

Introduction

The southern Great Lakes area (41.0°–44.5° N, 78.0°–82.0° W) is a major economic and industrial hub and home to more than 20 million people in Canada and the U.S. Although not as seismically active as other areas of eastern North America, this region has experienced recent moderate earthquakes, such as the 1998 m_N 5.4 (m_b *Lg* 5.2) Pymatuning earthquake (Armbruster *et al.*, 1998). (Here and later in this article the magnitude m_N is used, as defined in Nuttli [1973].) The nature of this seismic activity and its relationship (if any) to zones of weakness associated with ancient fault zones is poorly known (Adams and Basham, 1991). This uncertainty, and uncertainties in the magnitude-recurrence parameters for local sources, contributes significantly to the overall uncertainty in local hazard assessments (e.g., Geomatrix Consultants, 1997b; Adams *et al.*, 1999).

Historical data about the earthquakes in the Lake Ontario–Lake Erie region date back to 1823, but modern instrumental recordings are very sparse before 1970. From 1823 to the end of 1990, 131 events were documented in this region, with a maximum magnitude m_N of 5.2 (the Attica earthquake in 1929). As pointed out by Stevens (1994), the density of population in the southern Great Lakes region since at least the mid-1800s ensures that no earthquakes with a magnitude larger than 6 could have gone unnoticed. Based on the available data about the historical seismicity, previous authors have proposed a few localized seismicity clusters or zones (Fig. 1). These zones include the Burlington-Niagara Falls cluster (Adams and Basham, 1991) or Western Lake Ontario Seismic Zone (Seeber and Armbruster, 1993); the Attica cluster, as a part of the western New York State zone (Ebel and Kafka, 1991; Seeber and Armbruster, 1993); the Niagara Seismic Zone (Seeber and Armbruster, 1993); and a cluster of events south of Lake Erie on the Ohio-Pennsylvania border (Northeast Ohio seismic zone) (Seeber and Armbuster, 1993). Some of the earthquakes in this area are thought to have been induced by human activities, such as deep fluid injection and active solution in salt-mining operations (e.g., Mereu *et al.*, 1986; Nicholson *et al.*, 1988; Nicholson and Wesson, 1990). Overall, the seismicity of the region of Lake Ontario has been characterized by Stevens

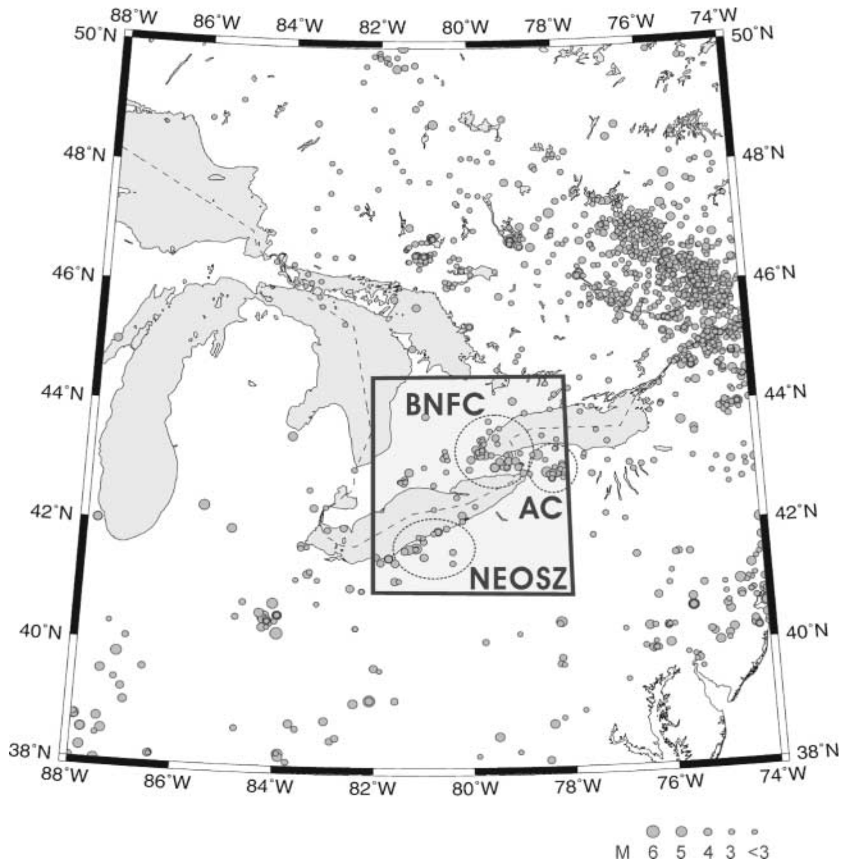


Figure 1. Historical seismicity (1737–1990) in the Great Lakes area. Ovals are seismicity clusters: BNFC, Burlington-Niagara Falls cluster (Adams and Basham, 1991); AC, Attica cluster (Ebel and Kafka, 1991); NEOSZ, Northeast Ohio seismic zone (Seeber and Armbuster, 1993). Earthquake catalogs of GSC and NEIC were used to make this figure. All known mining events and quarry blasts are removed from the catalog. The shaded rectangle shows the study area. Dashed line through the lakes is the U.S.-Canada border. This figure and all other map figures are plotted with GMT software (Wessel and Smith, 1991).

(1994) as "intermittent scattered," although other workers have attributed greater significance to apparent linear trends of seismicity in the historical catalog (e.g., Mohajer *et al.*, 1992; Mohajer, 1993, 1995a,b; Wallach *et al.*, 1998).

Since 1990, the University of Western Ontario has operated a seismograph network (Southern Ontario Seismic Network [SOSN]) for the purpose of obtaining information on the seismicity and seismic hazards in the general vicinity of a number of large nuclear power stations in southern Ontario. During the past decade, SOSN has recorded more than 100 events in the southern Great Lakes region, most of them small ($m_N < 3$). An improved monitoring capability has substantially reduced hypocentral location uncertainties for earthquakes within the study region, revealing at least one linear trend of epicenters near the west end of Lake Ontario (Mereu *et al.*, 2002). Whereas less well constrained historical data (largely based on felt reports) place almost all events outside of the lake, the emerging pattern of seismicity clearly shows that southern Ontario events are concentrated mainly beneath Lake Ontario or in adjacent areas to the south (Mereu et al., 2002).

The first objective of the present work is to establish better constraints on epicenter locations and focal depth estimations, since these are among the key parameters for earthquake hazard models. Our second objective is to test the hypothesis that local seismicity is, in part, controlled by pre-existing basement structures. Our primary approach has been to sharpen the definition of zones of seismicity by meticulous event relocation using two complementary techniques. Because uncertainties in crustal velocities can affect event locations, a comparison between hypocenter locations using three different 1D velocity models has been made. One of our relocation methods was the joint hypocenter determination (Pujol, 1988, 2000), which accommodates velocity model uncertainty through a station-correction term. Our revised epicenter maps provide the basis for delineation of empirical event clusters. To examine the statistical significance of apparent correlations between magnetic-anomaly trends and empirical earthquake clusters, we use a Monte Carlo method (Lutz, 1986). We also compare focal-plane solutions for three of the strongest events in our catalog (m_N 3.1 to 3.8) with the defined linear trends.

The seismicity of the western Lake Ontario area using data from 11 stations of the newly constructed SOSN was originally presented by Mereu *et al.* (2002). In that article, attention was focused almost entirely on locating the western Lake Ontario earthquakes using only *Pg* and *Sg* arrival times from that network. The analysis of Mereu *et al.* (2002) is expanded in this article to include (1) a much broader areal coverage, (2) data sets not only from SOSN but also from the Geological Survey of Canada (GSC) and the National Earthquake Information Center (NEIC), (3) independently repicked phases, (4) *Pn* and *Sn* waves, (5) a comparison of different velocity models and inversion techniques to test the robustness of the results, (6) spatial statistical analysis of the seismicity and magnetic anomaly trends, and (7) focal-plane solutions for three events beneath Lake Ontario.

Tectonic Setting

The study area of Lakes Ontario and Erie forms part of the seismically less active ("stable") continental interior (Fenton and Adams, 1997). Geologically, it is situated in the Interior Platform (Fig. 2), a region characterized by relatively thin (<1.0 km), flat-lying calcareous Paleozoic strata between the Appalachian Orogen and the Grenville Province of the Canadian Shield (Williams, 1984; Williams *et al.*, 1991). This region is underlain by a broad northeast-trending basement arch, comprising the Algonquin and Findlay Arches, separating the Appalachian (Allegheny) and Michigan sedimentary basins (Easton and Carter, 1995). The Frontenac Arch and Adirondack Mountains bound the area on the northeast and east, respectively.

The Precambrian rocks that underlie the sedimentary platform are part of the Grenville Province of the Canadian Shield. These rocks recorded tectonic processes that occurred along the ancient convergent continental margin of the protocontinent Laurentia during the interval 1.3–1.0 Ga (Rivers, 1997). The Grenville Province contains both parautochthonous, reworked material from the Laurentian margin and allochthonous crustal blocks that were subjected to high-grade metamorphism and penetrative deformation during the Grenville Orogeny (Rivers, 1997). The Central Metasedimentary Belt Boundary Zone (CMBBZ) represents a first-order subdivision of the Grenville Province (Fig. 2) that separates parautochthonous domains of the Central Gneiss Belt from allochthonous domains of the Composite Arc Belt (Carr *et al.*, 2000). The CMBBZ developed within the lithologically diverse rocks of the Composite Arc Belt. CMBBZ has a clear structural fabric that is subparallel to the bounding shear zones formed by ductile deformation concentrated within marble units between more massive intrusive bodies (White *at al.*, 2000). The southward extrapolation of the trace of the CMBBZ beneath the platform sediments is not certain, with some authors following a linear magnetic anomaly beneath western Lake Ontario (Easton and Carter, 1995; Forsyth *et al.*, 1994) and others placing it more to the west (Lidiak and Hinze, 1993).

To the north and east of the study area, the Early Paleozoic Iapetan rifted margin is prominently expressed. The Ottawa-Bonnechere graben (Fig. 2) is interpreted by some as a failed rift arm associated with the establishment of this passive margin. It represents a type of intracratonic rift system that is often associated with enhanced earthquake activity in stable continental areas (Dix *et al.*, 1997). According to one of the two current seismic hazard models used in Canada (Adams *et al.*, 1999), the region containing the Ottawa-Bonnechere graben is classified as a geological

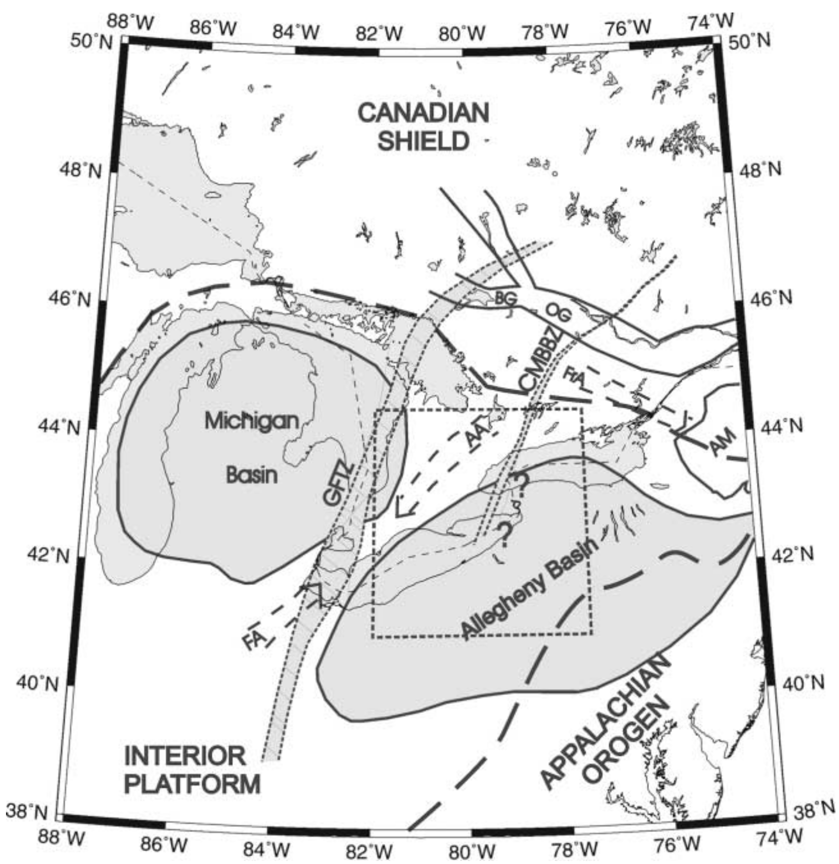


Figure 2. Tectonic elements in the Great Lakes area. CMBBZ, Central Metasedimentary Belt Boundary Zone; GFTZ, Grenville Front Tectonic Zone; FA, Findlay Arch; AA, Algonquin Arch; FrA, Frontenac Arch; OG, Ottawa Graben; BG, Bonnechere Graben; AM, Adirondack Mountains (from Williams *et al.*, 1991; Martini and Bowlby, 1991; Easton and Carter, 1992; Eyles *et al.*, 1997; Daneshfar and Benn, 2002). The dashed rectangle shows the study area.

source zone for earthquakes (Failed Iapetan Rift Zone). Regions containing older rifts, such as the Precambrian midcontinent rift, as well as other parts of the Canadian Shield that have not been subjected to rifting and extension (including the present study area), are not explicitly designated as geological source zones for strong earthquakes in current hazard models.

It is assumed that a relatively uniform stress field characterizes eastern North America, with nearly horizontal compression along the northeast–southwest azimuth (Adams, 1989a; Zoback, 1992). The source of this uniformly oriented stress environment is probably related to plate-driving forces (e.g., Adams and Bell, 1991; Richardson and Reding, 1991; Zoback and Zoback, 1991). A change in the stress regime from thrust faulting (in southeastern Canada) to strike-slip (in the eastern United States) is approximately coincident with the U.S.–Canada border (Zoback, 1992). This transition lies near the southern terminus of the Laurentide Ice sheet, which retreated locally around 12,000 years ago after cyclical episodes of advances and retreats during the past million years (Eyles *et al.*, 1997). It has been suggested that some of the earthquake activity could be the result of stress perturbations induced by postglacial rebound (Zoback, 1992; James and Bent, 1994). Numerous pop-ups, joint sets, and deformation features throughout the area are surface manifestations of the postglacial rebound process (Adams, 1989b; Godin *et al.*, 2002). Both tectonic stresses and glacial rebound stresses appear to be necessary to explain the distribution and style of contemporary earthquake activity in former glaciated areas of eastern Canada (Stewart *et al.*, 2000).

Seismological Data

The present study is based on 106 events recorded between the end of 1990 and 2001 and located in the area between latitudes 41.0° N and 45.0° N and longitudes 78.0° W and 82.0° W. The stations used in this study are indicated in Figure 3. We compiled arrival-time information from the Canadian National Database (the GSC database), the NEIC, U.S. station bulletins, and the University of Western Ontario (UWO) bulletin. Most of the data in the compiled bulletins come from three-component SOSN digital stations. There are a few arrival times from stations of the UWO local seismic network (DLA, LDN, and ELF), which we picked from analog seismograms. All digital stations have Global Positioning System (GPS) timing. The analog stations have had GPS timing since 1996. Between 1990 and 1996 a Geostationary Operational Environment Satellite (GOES) clock was used in these stations. There is a small inaccuracy due to the data transmission along telephone lines that is not taken into account. Technically the smallest uncertainties for *P*- and *S*-arrival picks for all stations are 0.01 sec, but in many cases they are up to 0.1 sec because of the presence of noise. In cases where significant arrival-time residuals were apparent after a first-pass analysis of more than 500 local and regional phases, we revised the picks from the short-period SOSN and UWO local networks. For each event some trial-and-error work was necessary to ensure that *Pg*, *Pn*, *Sg*, and *Sn* phases were correctly identified, especially for epicentral distances between 150 and 190 km. These phases are important because they contribute to reduction in focal-depth uncertainty, since the tradeoffs that affect the crustal phases (*Pg* and *Sg*) are opposite in sign from those that affect the upper-mantle-refracted phases (*Pn* and *Sn*). The final identification of these phases was based on minimization of the root-mean-square (rms) residual error between observed and calculated arrival times.

The studied events are almost all small. The largest event has a magnitude m_N of 5.4 (25 September 1998). The next largest event has a magnitude m_N of 4.4 (26 January 2001); there are 11 events with magnitudes m_N in the range 3.0–3.8; 58 events in the magnitude m_N range 2.0–2.9; 33 events in the m_N range 1.0–1.9; and 2 events with magnitudes m_N 1.0. A Gutenberg-Richter plot of cumulative earthquake frequency versus magnitude (Fig. 4) suggests that the overall magnitude threshold for the studied area during the period 1990–2001 was about 2.0, based on the change in slope. Away from Lake Ontario, however, the magnitude threshold is probably somewhat higher ($m_N \sim 2.5$).

There are only 12 events recorded by less than 10 stations, and a few events were recorded by more than 20 stations. On average, for every event there are 17 readings for the phases *P* (*Pg* and/or *Pn*) and *S* (*Sg* and/or *Sn*). The epicentral distances in the compiled bulletin data vary between 3 and 1800 km. Approximately 76% of events have recordings at less than 350-km distance. For about 30% of all events, there is at least one station at an epicentral distance of less than 25 km. The azimuthal gap for all events is between 58° and 347°. Due to the geometry of the network, the events near the western part of Lake Ontario have a smaller azimuthal gap (up to 284°) than the events south of Lake Erie.

Methods and Results for Hypocenter Locations

Two complementary algorithms are used for hypocenter location: (1) the HYPOCENTER program included in the SEISAN package (Earthquake Analysis Software, Havskov and Ottemöller, 2003) and (2) joint hypocenter determination (JHD). Each of these methods uses both *P*- and *S*-wave arrival-time information. The HYPOCENTER program (Lienert and Havskov, 1995) calculates the location and origin time independently for each event by tracing rays through the velocity model and adjusting earthquake parameters using a conventional least-squares approach. The method uses both direct and refracted waves and the arrival times are weighted according to their accuracy and epicentral distance. The JHD method (Pujol, 1988, 2000) iteratively determines the hypocenter location, origin time, and station corrections by operating simultaneously on all events from a specified cluster. The JHD method accounts for uncertain-

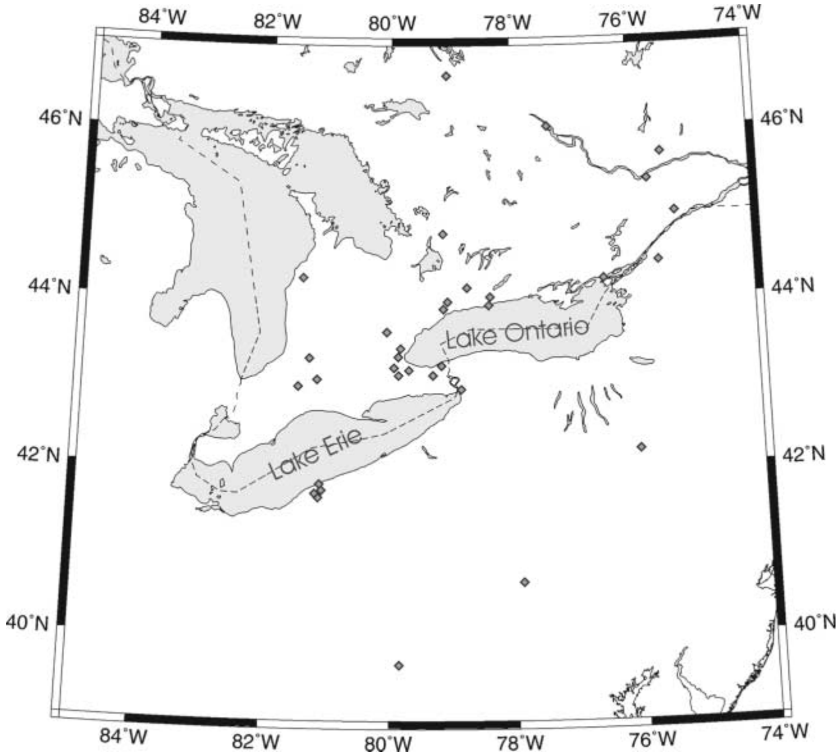


Figure 3. Map of seismic stations that recorded more than one event from the lakes Ontario-Erie region during 1991 through 2001. The dashed line through the lakes is the U.S.-Canada border.

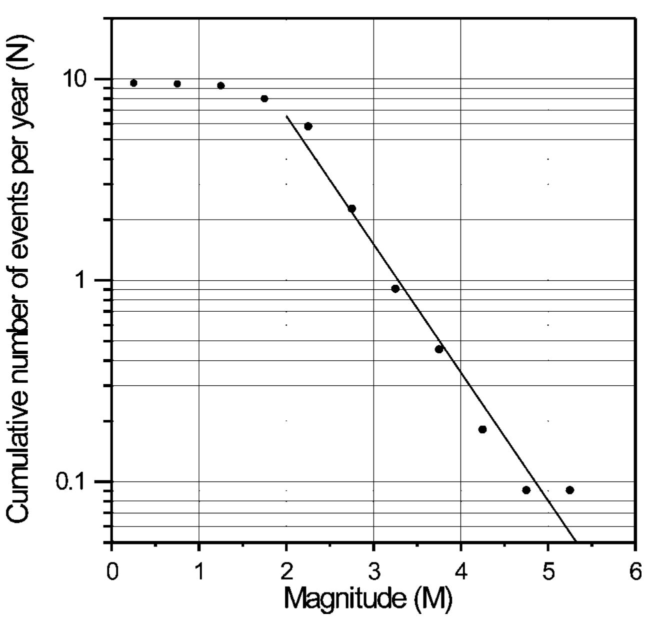


Figure 4. Gutenberg-Richter graph of the cumulative earthquake frequency in the study area per year for the period 1991–2001. The fitted line is obtained by the least-squares method. Its equation is: $N = -0.64*m_N + 2.09$. The coefficient of determination is $R^2 = 0.97$.

ties in velocity in the vicinity of each station in the station-correction term. Due to this advantage JHD provides a better map of the relative distribution of earthquake epicenters within a cluster of events. In some cases, however, there is an ambiguity in the absolute position of the hypocenters from JHD analysis. Some events are rejected during the iterative process based on a decreasing standard-deviation cutoff criterion (see below). Thus, it is useful to have results from both techniques to serve as a comparison.

For both of the methods, we use a simple 1D velocity model. More realistic 2D or 3D velocity models are not feasible at the present time because of the lack of constraints in the study area. To assess the sensitivity of the results to velocity uncertainties, three different 1D velocity models (Fig. 5) are tested with the hypocenter estimation by method 1. These models are based on the nearest available crustal seismic refraction study (Winardhi and Mereu, 1997) and are similar to the crustal model derived for North America by Perry *et al.* (2002).

Method 1: Single Event Location

In our initial pass, preliminary locations for all events were calculated using the compiled raw arrival-time data, the computer program HYPOCENTER (Lienert and Havskov, 1995) and the default velocity model (Fig. 5). After this, all arrival times with residuals larger then 1.0 sec (about 20% of all phases) were repicked from digital and a few analog seismograms. Weights for all phases were assigned based on the quality of picks according to the following scheme: a weight of 1 was assigned to a very impulsive pick;

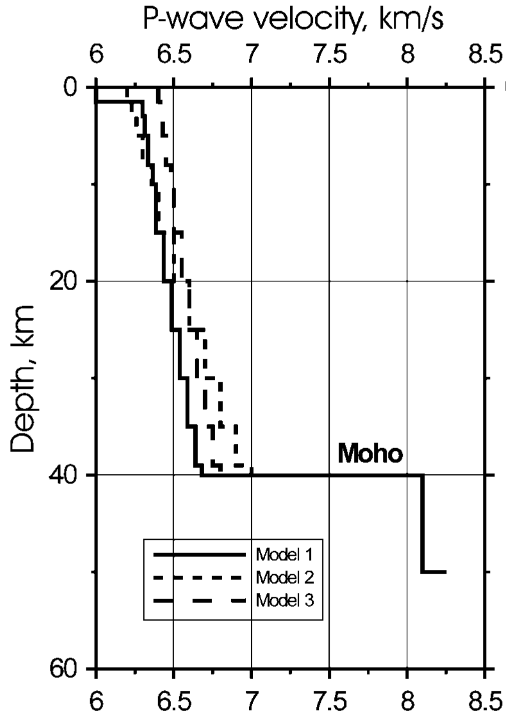


Figure 5. Velocity models used for hypocenter location (see text for details). The final results with the HYPOCENTER program and with the JHD program are obtained with model 1.

a weight of 0.75 was assigned to a clear pick with background noise; a weight of 0.5 was assigned to a more ambiguous pick with some noise; 0.25 was assigned to an ambiguous pick on a very noisy record; and 0 was assigned to a doubtful pick. An additional weighting term was assigned according to epicentral distance. This weight had a value of unity for distances $<$ 250 km, decreasing linearly to zero between 250 and 350 km. Revised hypocenters and origin times of 106 selected events were then calculated using the HYPOCENTER program (10 stations on average). GSC catalog locations were used as initial hypocenters. None of the location parameters, including the depth, were fixed in the relocation.

We found that the rms misfit of hypocenter locations for the whole data set varies between 0.34 sec (for model 1) and 0.38 sec (for model 3). Although there are some differences in epicenter locations, in general, the overall spatial distribution of epicenters remains the same for all three models. The average distance between epicenters obtained with models 1 and 2 was 1.56 km, and with models 1 and 3 the average distance was 2.54 km. The variance in depth is more significant, however. The average difference between depth estimations using models 1 and 2 was 2.7 km, and using models 1 and model 3 the average distance was 3.0 km. Taken together, there are only five hypocenters with depths below 20 km, with a maximum depth of 30 km. The depth of the two deepest events ($\sim$30 km) is resolved poorly. Model 1 produced the smallest rms misfit and was chosen as the default model used to calculate the results presented below.

The revised epicenters obtained by method 1 (Fig. 6) fall within two main zones. The first is located around western Lake Ontario (region 1, herein named "western Lake Ontario-Niagara seismic zone"), and the second is located in the south-central part of Lake Erie (region 2, herein named the "Ohio-Pennsylvania seismic zone"). The western Lake Ontario-Niagara seismic zone contains 64 events and the Ohio-Pennsylvania zone contains 26 events, whereas the remaining 16 events are scattered over a large area.

The rms arrival-time residuals are shown in Figure 7. The majority of the residuals are $<$0.70 sec, with an average value of 0.30 sec for the western Lake Ontario-Niagara seismic zone and an average value of 0.47 sec otherwise. The estimated depth distribution of earthquake hypocenters is summarized in Figure 8. All events are shallow, with a maximum estimated focal depth of about 19 km for the western Lake Ontario-Niagara seismic zone and about 24 km for the Ohio-Pennsylvania seismic zone. The number of events gradually decreases with increasing depth, but specific details of the focal depth distributions are not considered to be significant. The median depth calculated by the HYPOCENTER program is 6.4 km for western Lake Ontario-Niagara seismic zone and 5.6 km for the Ohio-Pennsylvania seismic zone.

We have compared the new hypocenter locations with hypocenter locations previously obtained by Mereu *et al.* (2002) using a different location algorithm (HYPO-UWO) and a different set of arrival-time data. The average difference between locations obtained by the HYPOCENTER program and the previous hypocenters is 2.5 km (based on 55 event locations). For 48 well recorded events the average difference is 1.6 km, and it is less than 2 km for 69% of the events and less than 1 km for 33% of the events. Depths, estimated by the HYPOCENTER program and HYPO-UWO, all fall within mutual error limits.

Method 2: Joint Hypocenter Determination

JHD obtains hypocentral parameters by solving a large system of linear equations. Various algorithms have been proposed to handle the inversion step. In this study we used the formulation of Pavlis and Booker (1983), as modified by Pujol (1988, 2000). This method uses singular-value decomposition (SVD) to invert the JHD system of equations in an iterative fashion. To satisfy the assumption that the path component is invariant and to have comparatively good azimuthal coverage, we applied this method to only 60 events from the western Lake Ontario-Niagara seismic zone, recorded at the 13 closest stations. Data are restricted to *Pg* and *Sg* arrivals when they are first arrivals. We selected standard parameters for the inversion, with five iterations in total, an ICONDN (the ratio of the largest to the smallest singular values) of 2000, a maximum allowed condition number for the matrix of partial derivatives for the individual earthquakes locations of 200, and a least-squares starting

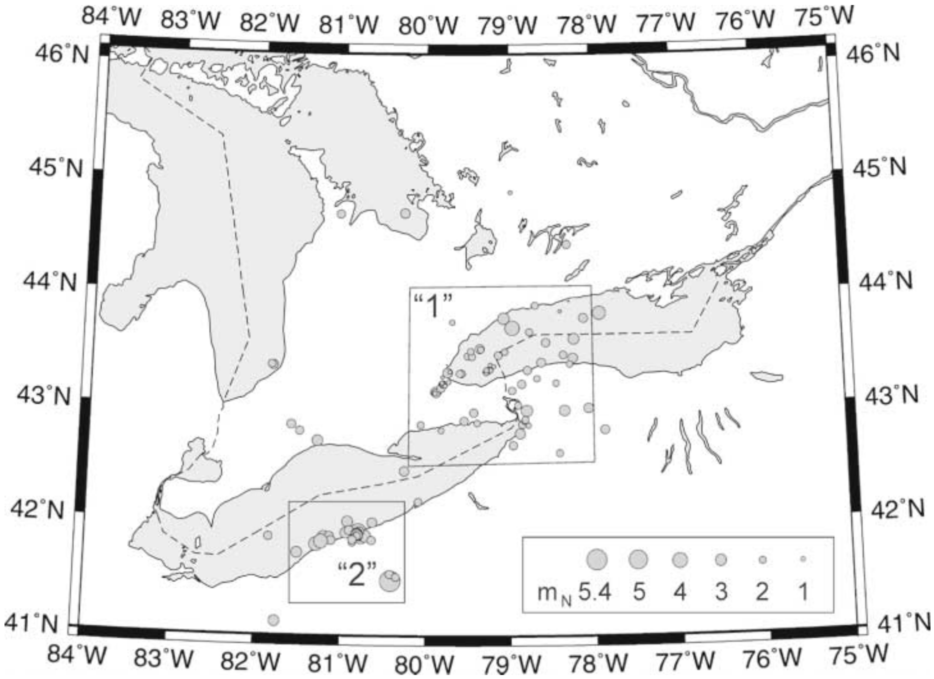


Figure 6. Epicenter map obtained using the single-event location program HYPOCENTER with the default velocity model. 1, western Lake Ontario-Niagara seismic zone; 2, Ohio-Pennsylvania seismic zone.

damping factor of 0.2. The cutoff standard deviations (S.D.) for each of the five iterations were assigned the values 2.0, 1.5, 1.0, 0.9, and 0.8 sec. There were 48 events left after the fifth iteration. The ratio of maximum and minimum singular values for this iteration was 824. According to the program manual, this ratio indicates that there were no numerical problems in the computation of the generalized inverse. The sum of the stations corrections is of order 10^{-4} sec.

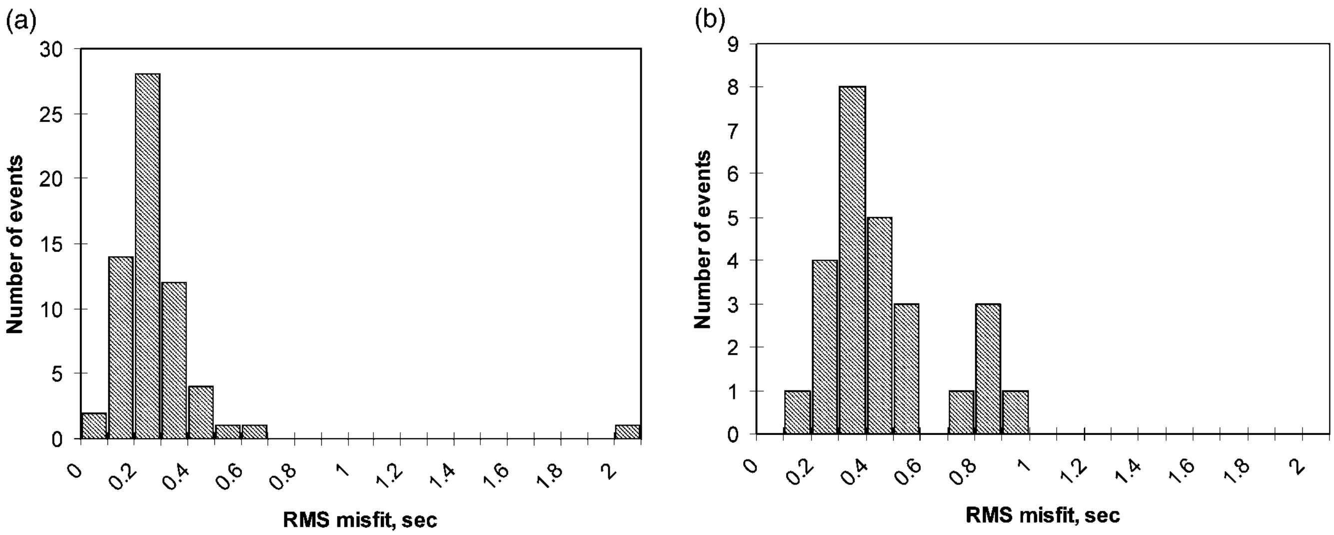
Figure 9 shows an epicenter map for the events relocated with the JHD program, compared with those determined using method 1. In general, the overall distribution of events from this technique is very similar to the distribution from method 1, adding confidence that the event locations are reliable. The average distance between epicenters for the same events using two different methods is 1.8 km, with a maximum discrepancy of 6.2 km. The rms arrival-time residuals of method 2 (Fig. 10) is in the interval 0.11 to 0.45 sec. The average rms for the hypocenters obtained by JHD is 0.27 ± 0.10 sec. For comparison the rms arrival-time residuals of method 1 for the same events range between 0.02 and 0.59 sec, with an average value of 0.26 ± 0.11 sec.

The focal depth distribution obtained by the JHD technique (Fig. 11) is similar to that obtained by method 1 (Fig. 8a). The average difference between depth estimations using method 1 and method 2 is 0.8 km with a maximum of 3.6 km. Although the median depth obtained by method 1 is 6.8 km, by method 2 (JHD) this depth is 7.4 km. In the JHD approach, the maximum depth is 19.7 km, which is not significantly different from the maximum depth of 18.7 km obtained for western Lake Ontario using method 1. We can be confident on the basis of these results that most of the earthquakes in western Lake Ontario are shallower than 20 km.

Fault-Plane Solutions

Focal mechanisms were obtained for three Lake Ontario earthquakes with magnitudes $m_N > 3$ (Table 1), using the program FOCMEC of Snoke *et al.* (1984) (from the SEISAN software package). This program uses a grid-search algorithm over the whole focal sphere (in this study on a 2° interval). Only polarities of *P* waves were used for the calculations. For each event, between 12 and 15 polarities (*Pg* and *Pn*) were picked from digital seismograms, all but one of which were from short-period stations. The solutions (Table 2) are median solutions from a set of possible solutions obtained by the program. Note that due to the network configuration, the azimuthal coverage is not optimal (Fig. 12). Takeoff angles are very restricted: around 50° for *Pn* and between 90° and 105° for *Pg*. As a result the solutions are not unique. For example, one event yielded two completely different sets of focal mechanisms.

The focal mechanism for the first event, 25 December 1998, is well constrained and it is of dip-slip type with *T* and *P* axes with intermediate plunge. The more probable fault plane is almost vertical, with strike northwest–southeast. (The other possible plane is almost horizontal.) The



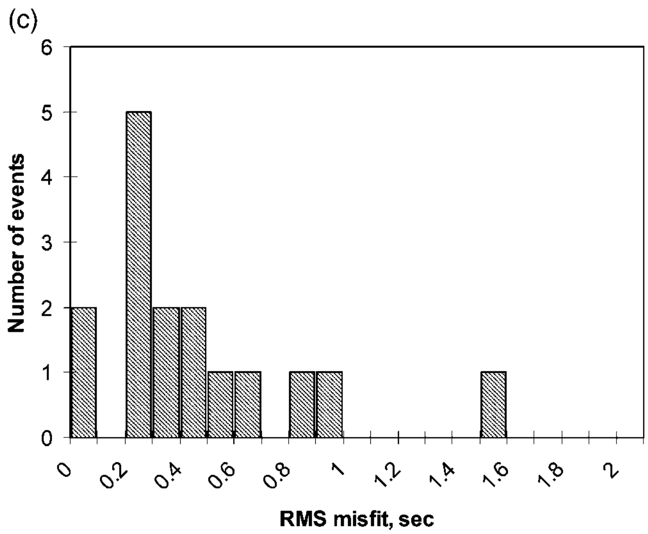


Figure 7. rms arrival time residuals, obtained using the program HYPOCENTER for the western Lake Ontario-Niagara seismic zone (a), the Ohio-Pennsylvania seismic zone (b), and all other events (c).

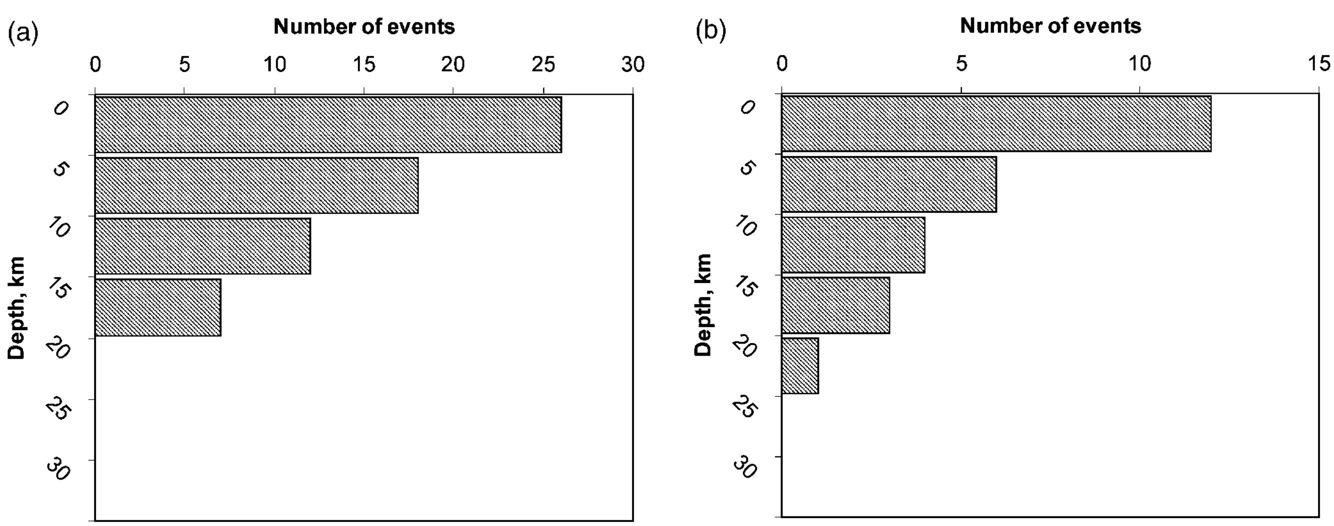


Figure 8. Earthquake focal depth distribution determined using the program HYPOCENTER for the western Lake Ontario-Niagara seismic zone (a) and the Ohio-Pennsylvania seismic zone (b).

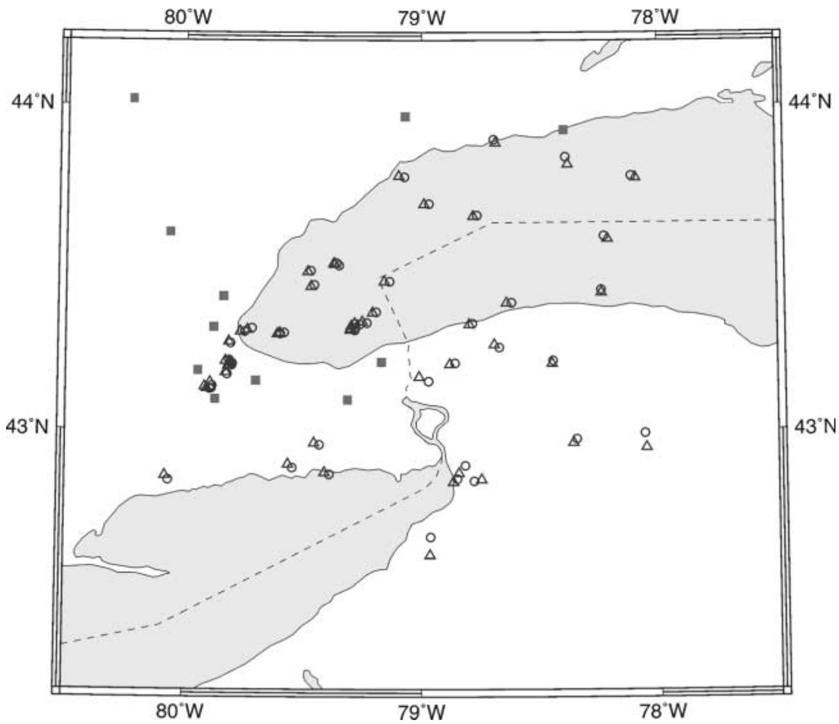


Figure 9. Comparison between the epicenter locations for cluster 1, obtained using the program HYPOCENTER (circles) and the JHD technique (triangles). The seismic station locations are shown with black squares.

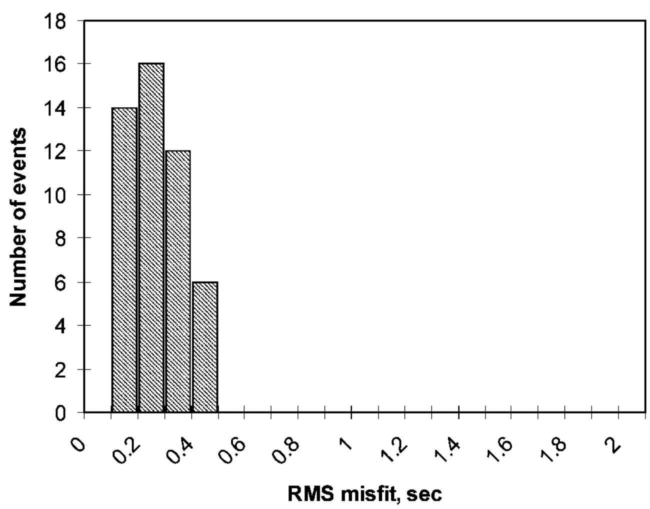


Figure 10. rms arrival time residuals distribution obtained using the JHD technique for the western Lake Ontario-Niagara seismic zone.

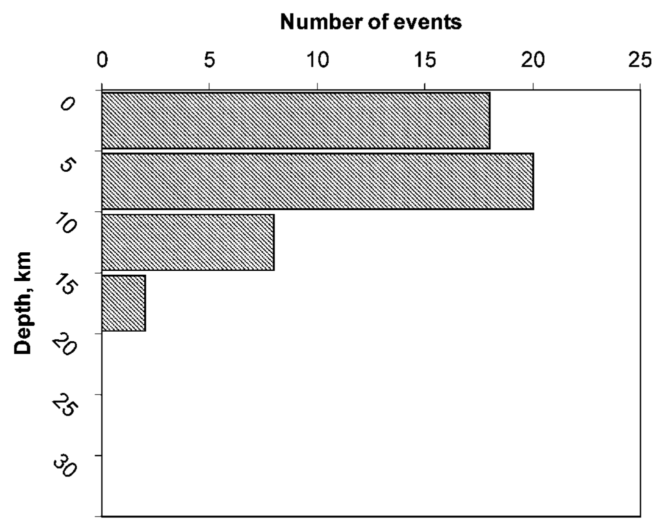


Figure 11. Earthquake focal depth distribution obtained using the JHD technique for the western Lake Ontario-Niagara seismic zone.

focal mechanism for the second event, 26 November 1999, is not very well defined. The solutions group into two different sets. One set of solutions corresponds to oblique thrust faulting with *P* axes closer to the horizontal plane and intermediate plunge of the *T* axis. For this set of solutions, one of the possible fault planes strikes northeast–southwest, whereas the other strikes north-northwest–south-southeast. The other set of solutions corresponds to oblique normal faulting. In this case the *T* axis is almost horizontal and the *P* axis has an intermediate plunge. The first possible fault plane for the second set of events strikes northeast–southwest (almost coincident with the first plane of the first set of solutions), whereas the other plane strikes northwest–southeast. The focal mechanism of the third event, 24 May 2000, is comparatively well defined. It is of oblique normal type with the *T* axis closer to the horizontal plane and intermediate *P* axis. The focal planes are oriented northeast–southwest and north–south, respectively.

Table 1
Parameters of Earthquakes with Focal Mechanisms

No.	Date (mm/dd/yy)	m_N	Time	Latitude (N)	Longitude (W)	Depth (km)
1	12/25/98	3.6	13:30:26.2	43.82	77.87	13.0
2	11/26/99	3.8	22:33:01.6	43.70	78.97	9.2
3	05/24/00	3.1	10:22:46.8	43.78	79.01	5.0

Table 2
Fault-Plane Solutions for Three Events in Lake Ontario

				Plane 1			Plane 2			*P* axes		*T* axes	
No.	No. of Polarities	No. of Bad Polarities	Type of solution	Strike	Dip	Rake	Strike	Dip	Rake	Trend	Plunge	Trend	Plunge
1	12	1	DS	232 ± 5	84 ± 4	−84 ± 4	7 ± 35	8 ± 4	−135 ± 30	149 ± 5	51 ± 4	317 ± 9	39 ± 2
2	15	0	OT	228 ± 1	38 ± 5	145 ± 7	347 ± 5	69 ± 6	58 ± 5	101 ± 5	18 ± 6	218 ± 8	54 ± 4
	15	0	ON	227 ± 2	60 ± 15	−133 ± 15	109 ± 8	50 ± 13	−40 ± 14	78 ± 15	56 ± 7	346 ± 10	1 + 8
3	14	2	ON	42 ± 6	24 ± 14	−54 ± 8	183 ± 6	71 ± 12	−105 ± 8	71 ± 15	62 ± 9	284 ± 10	24 ± 13

The numbers in the table correspond to the numbers in Table 1. DS, dip-slip faulting; OT, oblique thrust faulting; ON, oblique normal faulting.

Trend Analysis of Magnetic Anomalies and Seismicity

Although many authors have proposed causal links between magnetic lineaments and seismicity in western Lake Ontario (e.g., Mohajer *et al.*, 1992; Wallach *et al.*, 1998; Boyce and Morrice, 2002), previous studies have been primarily qualitative in nature. Here, we have applied a statistical approach to attempt to test the hypotheses that seismicity patterns in western Lake Ontario and Lake Erie are (1) clustered, (2) exhibit a preferred orientation, and (3) can be correlated to magnetic anomaly trends.

Before addressing the question of whether seismicity patterns show a preferred orientation, it is necessary to establish that earthquake epicenters can be considered as clustered in a statistically significant sense. A clustered point distribution is one in which the points (epicenters) tend to be concentrated within localized regions of the attribute space (Swan and Sandilands, 1995), in this case, geographic position at the surface. We applied the χ^2 test for randomness (Swan and Sandilands, 1995, p. 272) to make this determination. This simple test is performed by subdividing the area of interest into square cells, counting the number of points (epicenters) that fall within each cell, and constructing a histogram. A statistically random spatial distribution of point data will theoretically yield a histogram with a Poisson's distribution, whereas a pseudoregular (ordered) distribution is expected to yield a roughly Gaussian distribution. Spatial distributions that are clustered at the scale of the cell dimensions are expected to show significant deviations from either of these. Figure 13 shows the histogram that we obtained using a 30 km × 30 km box-shaped sample. The histogram shows no resemblance to a Gaussian distribution and, relative to a Poisson's distribution, the histogram shows a deficiency of cells with one or two events and a surplus of cells with zero and three or more events. Essentially the same histogram shapes were obtained using different box dimensions in the range of 20–50 km. We conclude that earthquake epicenters in the region for the period 1990–2001 are spatially clustered with a length scale of 20–50 km. Because $\chi^2/\chi^2_{crit} = 8.93$ (chi-square statistic), this result has a very high degree of statistical significance.

To test the hypothesis that the epicentral data show a preferred orientation, we used the azimuthal histogram analysis method of Lutz (1986). In this method, azimuths are computed for line segments connecting all possible pairs of points. The histogram produced by this method (Fig. 14) contains a pronounced peak at N40°E, indicating a strong tendency for pairs of points to have this orientation. Much of this trend, however, may simply reflect the presence of distinct earthquake clusters near the south shore of Lake Erie and western Lake Ontario. The centers of these clusters form an azimuth of approximately N40°E; this spatial relationship may be unrelated to any intrinsic preferred orientation. To correct for this potential bias, we computed confidence limits using 100 histograms constructed from random point distributions constrained to lie within the area of investigation (Lutz, 1986). Each of these random point distributions was a realization of an event distribution with the same spatial probability density function as that of the observed data (i.e., the number of events in each 30 km × 30 km cell matched the observations). This statistical test is much stronger than a purely random earthquake distribution. It shows that, even after removal of any potential bias resulting from the geographic locations of the main earthquake clusters, there is a statistically significant (at the 90% confidence limit) preferred orientation of earthquake epicenters between N35°E and N40°E.

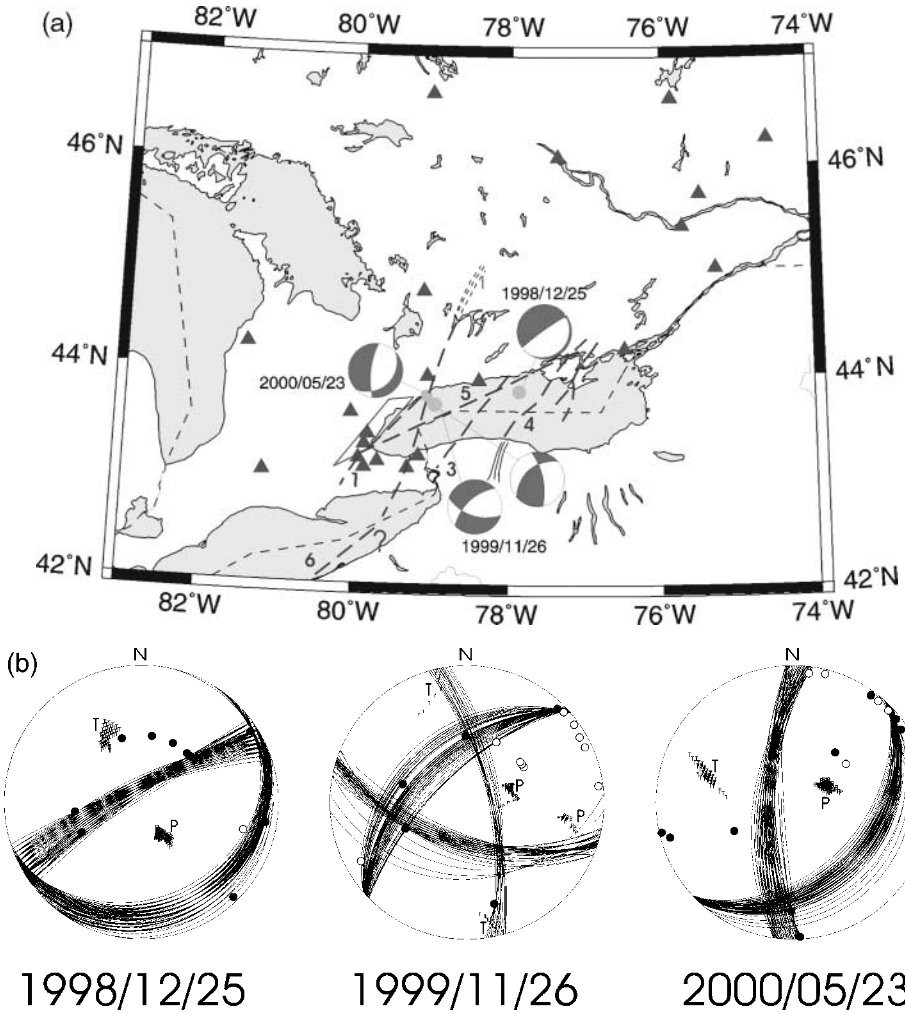


Figure 12. Focal mechanisms for three earthquakes with epicenters in Lake Ontario: 25 December 1998 (m_N 3.6), 26 November 1999 (m_N 3.8), and 25 May 2000 (m_N 3.1). For the second event there are two possible types of solutions and they are both shown. (a) Map of the events with fault plane solutions and beach-ball diagrams for the three events. (b) *P*-wave polarities used for calculating the focal mechanisms and all possible solutions obtained by 2° increment in search. The seismic station locations are shown with black triangles.

Magnetic anomalies in this region primarily reflect lithotectonic and structural features in the Precambrian basement underlying the lake waters and the Phanerozoic sedimentary rocks (Ouassaa and Forsyth, 2002). The presence of faults and lithotectonic features can be highlighted by the magnetic field. If these features are being reactivated by the current stress field, they should have a trend similar to that of the earthquake clusters. We have applied the azimuthal histogram technique to the magnetic anomaly map by defining sets of point data from aeromagnetic local maxima and minima. For completeness, we incorporated the most recent (1999) aeromagnetic data (Kiss and Coyle, 2000) into our analysis. The azimuthal histograms (Fig. 15) resolve a broad trend between N20°E and N45°E in the case of the magnetic maxima, and two distinct trends at N15°E–N25°E and N35°E–N45°E in the case of minima. Note that apparent magnetic anomaly trends at 0° and 90° (Fig. 15) are strongly contaminated by grid artifacts.

The preferred azimuthal orientation of seismicity data (N35°–40°E) falls within a dominant magnetic anomaly trend. This similarity in trend directions is the first statistical evidence in support of a link between some basement structural elements and patterns of seismicity. It should be cautioned, however, that the magnetic data also contain prominent trends with other orientations that do not appear to be associated with seismicity. This fact is most likely because only a subset of the magnetically discernible basement structural elements are favorably oriented for failure within the present-day stress regime.

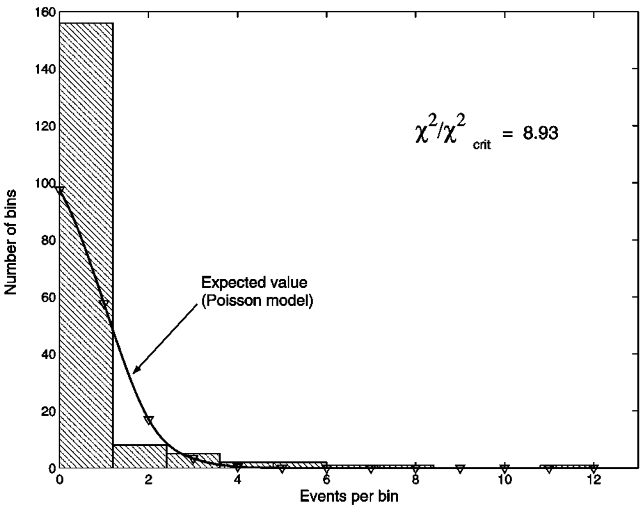


Figure 13. χ^2 test for randomness, based on point count within a 30 × 30 km sliding box. See text for details.

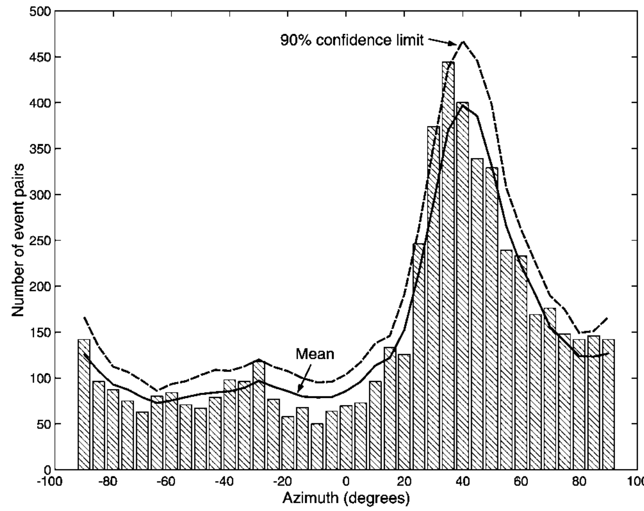


Figure 14. Azimuthal histogram of earthquake epicenters. The 90% confidence limit was calculated by a Monte Carlo simulation method (Lutz, 1986) that accounts for any bias due to the geographical distribution of earthquake clusters. The azimuth bins are centered on integer multiples of 5°.

Empirical Clusters

We have established that both the clustering of earthquakes and preferred cluster orientation of N40°E are statistically significant. Here, we describe these clusters in more detail.

We recognize five distinct clusters of seismicity (Fig. 16). Three of these clusters (A, B, and C) are located in western Lake Ontario and are characterized by an apparent strike direction of approximately N48°E. Starting from the west, the A cluster (or Hamilton-Burlington cluster; Mereu *et al.* 2002), contains 15 events and is defined by a 40-km-long, north-northeast-oriented line of epicenters. Cluster B (about 25 km long) and cluster C (about 35 km long) lie between Toronto and St. Catherines. All three clusters occur in the general proximity of previously defined lineaments but do not coincide precisely with them. A fourth more irregular cluster (D) is evident in the vicinity of Niagara Falls. It is oriented northeast–southwest near the east end of Lake Erie but changes its orientation to northwest–southeast farther north. The total length of this cluster is about 55 km. Finally, cluster E appears to be present in the southern middle part of Lake Erie. This cluster trends northeast–southwest, following the lakeshore for a distance of about 60 km.

To assess whether hypocenters along clusters A, B, and C have a preferred dip direction, we have constructed vertical cross sections (Fig. 17) using results from the HYPOCENTER program. Although resolution is limited by depth uncertainties, the HYPOCENTER results suggest that these events may have occurred along planes that are near vertical. On Profile I two parallel planes are apparent with a separation of 10 km, corresponding to clusters A and B on the epicenter map on Figure 16. On Profile II two planes are apparent with separation of approximately 70 km. They correspond to clusters A and D on the epicenter map on Figure 16.

Discussion

On the basis of linear trends in magnetic data or linear physiographic features, previous authors have proposed numerous lineaments transecting the study area. The seismogenic significance of these linear features, however, is not well established. For a complete description of proposed lineaments, the reader is referred to Geomatrix Consultants (1997a) and Boyce and Morris (2002). We limit the following discussion to a few of these features (Fig. 18) that appear, on the basis of their spatial location, to be most relevant to the seismicity trends evident from the epicenters of recent earthquakes.

1. The Toronto-Hamilton seismic zone (Mohajer, 1993) is associated with a magnetic lineament extending from Burlington to Toronto, offshore of and parallel to the northeast-trending coastline of Lake Ontario (Thomas *et al.*, 1993). A second, parallel magnetic lineament bounds the zone 20 km to the west. The zone bounded by both lineaments was dubbed the Toronto-Hamilton seismic zone by Mohajer (1993). Our cluster A is adjacent to Mohajer's Toronto-Hamilton seismic zone, but is offset by about 7–8 km to the southeast. Our cluster B is parallel to the Toronto-Hamilton seismic zone but shifted to the southeast by about 30 km.
2. The Niagara-Pickering Linear Zone (Wallach and Mohajer, 1990) is a 30-km-wide trend defined by a linear magnetic anomaly, a less well defined Bouguer gravity anomaly, and linear physiographic features. Some authors (Easton and Carter, 1995; Forsyth *et al.*, 1994) interpret the Niagara-Pickering Linear Zone as the southward projection of the CMBBZ, whereas others place the

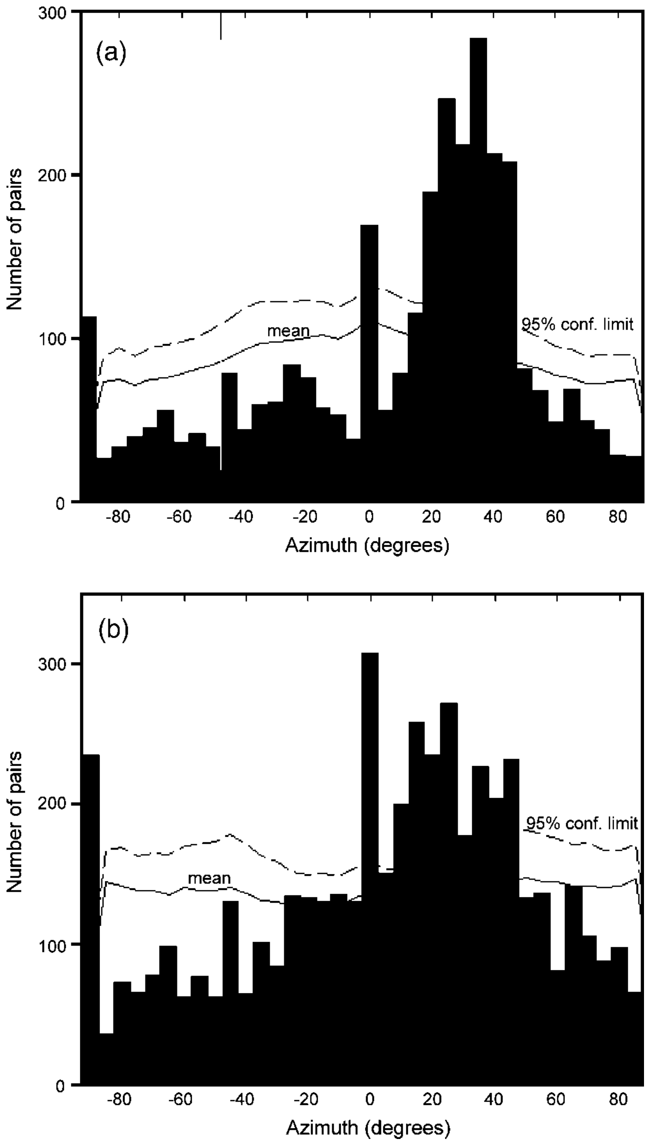


Figure 15. (a) Azimuthal histogram of aeromagnetic maxima using the method of Lutz (1986). (b) Azimuthal histogram of aeromagnetic minima.

CMBBZ farther to the west (Lidiak and Hinze, 1993; O'Dowd *et al.*, 2003). Our clusters A and B cross the Niagara-Pickering Linear Zone. Their trend is rotated in relation to the Niagara-Pickering Linear Zone by about 30°.

3. The Wilson-Port Hope Magnetic Lineament (Mohajer, 1993; McQuest Marine, 1995) is a northeast-trending magnetic lineament, from Port Hope, Ontario, to Wilson, New York. It is approximately 80 km long and is subparallel to the Niagara-Pickering Linear Zone and the Clarendon-Linden fault system. Our cluster C is parallel to the Clarendon-Linden fault system but slightly shifted to southeast by about 4 km.
4. The Clarendon-Linden Fault System is a set of northeast-trending faults that are connected to the strongest event in the area: the Attica earthquake in New York state (m_b 5.2) in 1929 (Seeber and Armbruster, 1993). Since then, a few other events with magnitudes from 2.7 to 4.7 have occurred along this system (Fletcher and Sykes, 1977). Geological evidence shows that the Clarendon-Linden Fault System has experienced a combination of dip-slip and strike-slip motion (Jacobi and Fountain, 1995). After high-pressure injection for hydraulic mining of salt at Dale, New York, induced seismicity was triggered along this system in 1970 (Nicholson and Wesson, 1990). During the recording period considered here, little seismic activity occurred along the Clarendon-Linden Fault System, with only two events near its southern end.
5. The Hamilton-Presqu'ile Lineament (McFall and Allam, 1991; Ontario Geological Survey, 1991) is a subtle aeromagnetic lineament that extends in the east-northeast direction for approximately 200 km from near the western shore of Lake Ontario. It lies parallel to a possible southwest extension of the St. Lawrence rift zone, first suggested by Adams and Basham (1989). Clusters A, B, and C are situated close to the Hamilton-Presqu'ile Lineament, but neither the orientation of clusters nor the hypothetical line connecting clusters A and B are parallel to this lineament.

The southern Great Lakes region falls within a transition from thrust-dominated seismicity to the north and strike-slip-dominated seismicity to the south. Although to the north and east of the study area (between the western Quebec seismic zone and the northeastern United States) the transition is very sharp (Du *et al.*, 2003), along the western Lake Ontario zone we have mixed solutions (oblique slip). The focal mechanisms obtained here show a predominantly oblique normal type of faulting. There is a previously obtained focal mechanism in Lake Ontario that is also a normal type, 23 July 1988 (Adams *et al.*, 1989; Reinecker *et al.*, 2003). In the recently published article by Bent *et al.* (2003) there are focal mechanisms for two of the events we have studied here, 26 November 1999 and 24 May 2000. For the first event their single solution is almost identical with our second set of possible solutions for this event, e.g., it is of the oblique normal faulting type. The focal mechanism of their second event is of the thrust type and differs significantly from ours. We note, however, that this solution is of quality C in the list of solutions of Bent *et al.* (2003).

The three focal mechanisms obtained here are weakly constrained. This variability probably reflects large uncertainties due to the small magnitude of the events, which reduce the number of observations that are available to constrain the solutions. Finally, we note that many of the solutions contain nodal planes that strike in a direction (42° to 52°) consistent with the orientation of earthquake clusters and the dominant trend of the magnetic anomalies.

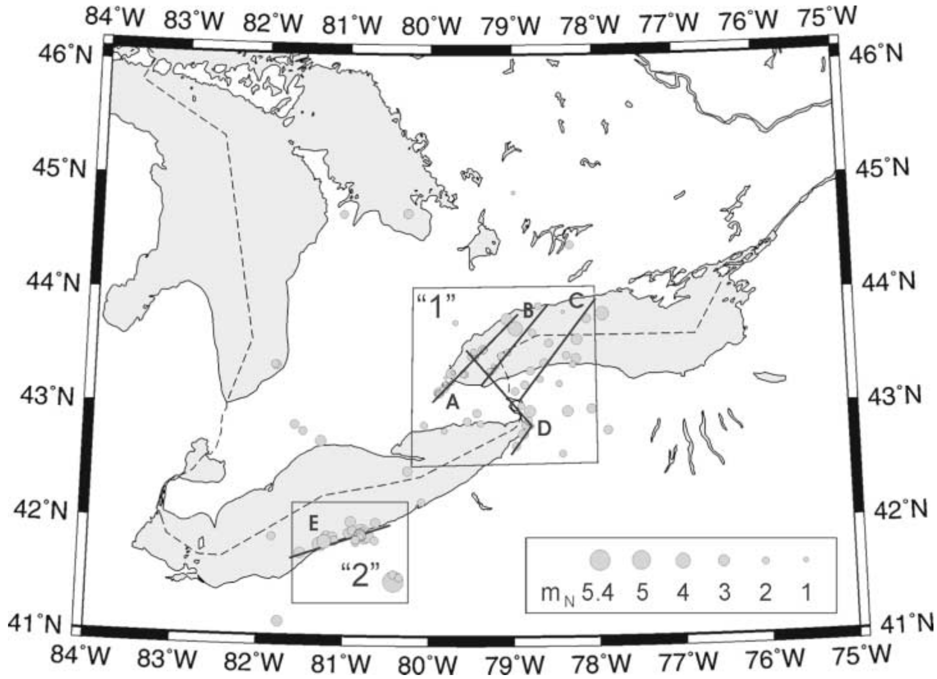


Figure 16. Defined linear trends of clusters in western Lake Ontario-Niagara seismic zone ("1") and Ohio-Pennsylvania seismic zone ("2"). Heavy lines indicate hypothetical linear zones and capital letters indicate clusters, defined in the present work. The event locations were obtained using the program HYPOCENTER with the default velocity model.

Conclusions

Revised hypocenter locations have been estimated for earthquakes in the southern Great Lakes region for the period 1990–2001 using two different methods: single-event location using the program HYPOCENTER (106 events) and joint hypocenter determination (49 events). Comparison of the results from this study, as well as 55 independent hypocenter locations of Mereu *et al.* (2002), reveals very good agreement. All events are confined to the midupper crust above 30 km, and most events occur above 20 km.

The epicenters are concentrated beneath Lake Ontario and Lake Erie. The epicenters exhibit statistically significant clustering and a northeast alignment, at 95% confidence level. Aeromagnetic data which provide the best available proxy for basement structural fabric, define two statistically significant trends at the 95% confidence level: north-northeast and northeast. We recognize five distinct clusters, four in the western Lake Ontario-Niagara Falls seismic zone) and one in the Ohio-Pennsylvania seismic zone. Available preliminary focal mechanisms are not well constrained. They show oblique slip with some possible fault planes that are consistent with the orientation of basement structures observed in the magnetic data.

The inferred spatial correlations between modern seismicity and magnetic anomalies and the proximity of seismicity to lakes suggest that seismicity in the region tends to be localized mostly in areas where pre-existing tectonic structures are favorably oriented with respect to the present-day stress field and water is present at the surface. It should be cautioned that, although previous authors have proposed a large number of basement lineaments, based on aeromagnetic and gravity data, our results show that there is no evidence for seismic activity along the majority of the proposed lineaments. Taken together, the data suggest that both fluids and basement structures (in particular, those that are favorably oriented with respect to the ambient stress field) influence the location of earthquakes in this region.

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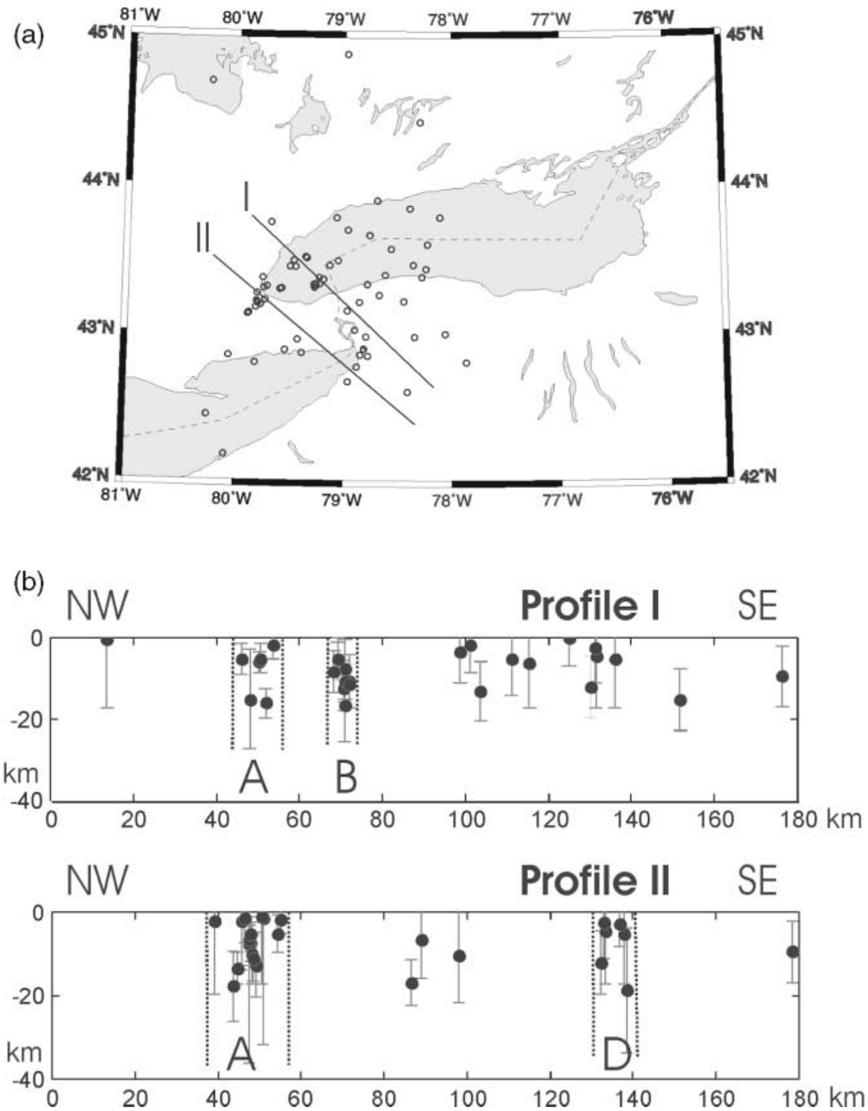

Figure 17. Vertical cross sections. (a) Map, showing the cross-section locations. (b) Cross sections using hypocenters obtained using the program HYPOCENTER. Dotted lines outline the separate seismic clusters. The calculated depth errors are shown. The apparent vertical planes are indicated with the same letters as the corresponding clusters on Figure 16.

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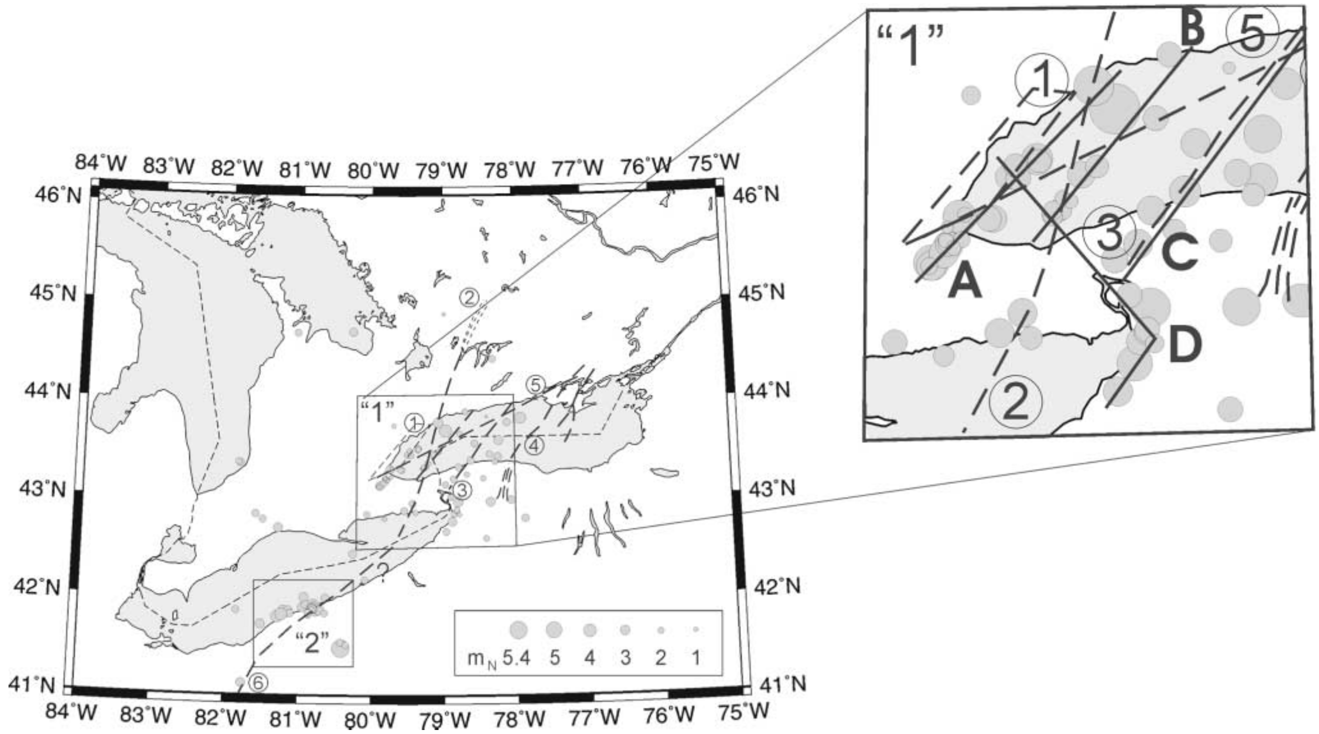

Figure 18. Comparison of earthquake epicenters obtained by using the program HYPOCENTER with previously defined and postulated lineaments and faults (according to Mohajer *et al.*, 1992; McQuest Marine, 1995; Geomatrix, 1997a). 1, Toronto-Hamilton Seismic Zone; 2, Niagara-Pickering Linear Zone; 3, Wilson-Port Hope Magnetic Lineament; 4, Clarendon-Linden Fault System; 5, Hamilton-Presqu'ile Fault; 6, Akron Magnetic Anomaly. On the right side the magnified western Lake Ontario-Niagara Zone is shown. Additionally, the linear trends of clusters defined in this study (A, B, C, and D) are given for comparison with the previously defined lineaments.

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