Articles

The Seismicity and Seismotectonics of Canada East of the Cordillera [1]

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Introduction

Canada east of the Cordillera, extending north from the United States border to the Arctic Ocean, comprises about two-thirds of the stable craton of the North American plate. Much of this large area appears to be substantially aseismic, although it contains several zones of significant seismicity and a few other regions of lower-level seismicity. The seismicity of the southern part, together with the adjacent United States, was compiled comprehensively by Smith (1962, 1966), who collected earthquake reports compiled by others, analysed original records where possible, and decided on the best location and magnitude for each earthquake. Smith's maps have been widely used by others (*e.g.*, Yang and Aggarwal, 1981). Further earthquake analysis, including spatial distibution, recurrence rates, and relationship to geological structure, was made by Basham *et al.* (1979) and refined by Basham *et al.* (1982a). The latter paper, although the most thorough seismicity compilation to date, was compiled for an engineering seismic hazard study and is not widely known or circulated, although the conclusions were published (Basham *et al.*, 1985). Other recent reviews are Hasegawa *et al.* (1985), which deals with the crustal stresses driving the eastern Canadian seismicity, Hasegawa (1986), which discusses the seismotectonics of southeastern Canada, and Kumarapeli (1987), which discusses seismic zones in relation to ancient fault systems and post-glacial faulting.

Within the southern part of the continental region, seismicity is clustered in four zones. In three of these zones — (i) western Québec, which includes a band of earthquakes along the Ottawa River and a second band north of the river; (ii) Charlevoix, a repetitive source of large earthquakes and a continuous source of small earthquakes; and (iii) the lower St. Lawrence, a zone of mostly small earthquakes — most of the earthquakes are thrust events occurring at depths of 5 to 25 km within the Grenville cratonic basement, apparently chiefly through the reactivation of an Iapetan (late Proterozoic to Paleozoic) rift fault system along the St. Lawrence and Ottawa rivers (Kumarapeli, 1985). The fourth zone, the northern Appalachians, includes the Miramichi earthquakes of 1982 which were due to shallow (< 10 km) thrust faulting within a sheet of rocks that have been thrust over the older basement. From the Great Lakes west to the Cordillera, the most significant cluster of earthquakes is in Saskatchewan, as is adequately discussed by Horner and Hasegawa (1978).

Along the eastern margin of the continent (North Atlantic Ocean, Labrador Sea, Baffin Bay), the seismicity includes the 1929 M7.2 Grand Banks and 1933 M7.3 Baffin Bay earthquakes. These and smaller earthquakes appear to be concentrated at the ocean-continent transition, perhaps by reactivation of the Mesozoic rift faults created when the North Atlantic was formed. In the Labrador Sea, earthquakes also occur on the extinct spreading ridge and its associated transform faults.

The seismicity of northeastern (Arctic) Canada was poorly known until the installation of the standard seismograph stations in the early 1960s. After about 15 years of lower magnitude seismicity had been recorded, Basham *et al.* (1977) attempted the first comprehensive assessment of northern Canadian seismicity. Although a number of earthquakes in smaller sub-regions have since been described in more detail, as will be outlined below, much of our knowledge of the seismicity and seismotectonics of northern Canada has not advanced greatly beyond that available in 1977.

In the present paper, we discuss briefly the regional seismicity and relate it to the seismotectonics as it is currently understood. Developments to our understanding have happened faster in the more accessible southern part of eastern Canada monitored by digital seismographs, so that description is more extended. We begin in southern Canada, discussing each cluster of seismicity in turn and proceed from eastern Ontario, down the St. Lawrence River, to the earthquakes along the Atlantic margin, and the earthquakes of the Arctic, before summarizing our inferences about the causes of earthquakes in Canada east of the Cordillera.

Some of the ideas we support have previously been suggested by others, *e.g.*, Woollard (1969) and Kumarapeli and Saull (1966) for the activity of the St Lawrence Valley, and Sykes (1978) for the trend between Boston and Ottawa. We believe that many reached approximately the correct conclusions, but that our improved knowledge of the seismicity together with improvements to the understanding of the geological history of eastern Canada have allowed us to substantially refine the earlier ideas.

We now feel that all the earthquakes in Canada east of the Cordillera appear to be occurring within a regional stress field dominated by northeast to east compression, and that most large earthquakes have occurred near Paleozoic or younger rift structures that surround or break the integrity of the North American Craton.

Southeastern Canada

Western Québec. A significant cluster of earthquakes occurs in the Grenville Province of the Canadian Shield, predominantly in western Québec but extending across the Ottawa River into eastern Ontario and south into the Adirondack Mountains of New York State. An earthquake with magnitude about 6 occurred at or near Montréal in 1732 (Leblanc, 1981). During this century earth-

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SOUTHEASTERN EARTHQUAKES

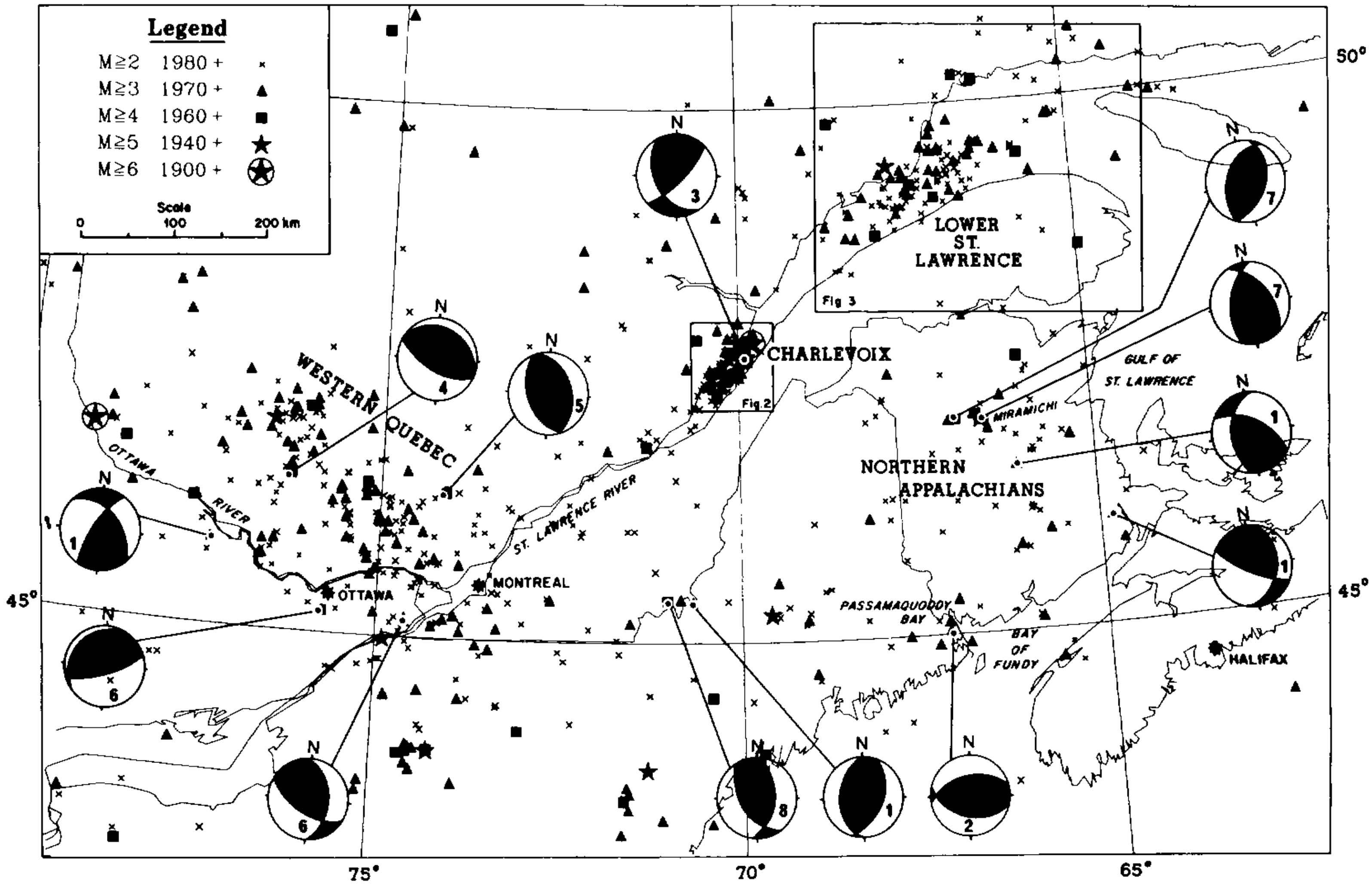

Figure 1 *Seismicity of southeastern Canada identifying the four most active zones. The map shows a selection of earthquakes prior to July 1987, as detailed in the legend, chosen to eliminate many of the older, very poorly located earthquakes. Focal mechanisms (lower hemisphere projections with compressional quadrants shaded) are from:* 1, Adams et al. *(1988); 2, Ebel (1985), as modified by Adams; 3, Hasegawa and Wetmiller (1980); 4, Horner* et al. *(1978); 5, Horner* et al. *(1979); 6, Wahlström (1987b); 7, Wetmiller* et al. *(1984); and 8, Yang and Aggarwal (1981).*

quakes of M6.2 occurred near Lake Timiskaming in 1935 and M5.6 near Cornwall, Ontario in 1944.

For the last twenty years, all earthquakes have been M4.3 or less and most have been located north of the Ottawa River. Significant recent earthquakes, most studied by monitoring aftershocks, include: near Maniwaki 1975 – M4.2 (Horner *et al.*, 1978); St. Donat 1978 – M4.1 (Horner *et al.*, 1979); Cornwall 1981 – M3.3 (Schlesinger-Miller *et al.*, 1983); Timiskaming 1982 – M4.3; and North Gower (near Ottawa) 1983 – M4.1 (Wahlström, 1987a). In an extension of the exposed Grenville Province into the Adirondacks of New York State, the largest recent earthquake was M5.2 at Goodnow in 1983.

In detail (Figure 1), the seismicity appears to occur in two bands, probably sufficiently distinct to be considered separate zones. The first band, trending slightly west of northwest, lies along the Ottawa River from Lake Timiskaming (near circled star at left side of Figure 1) to Ottawa and thence widens to extend southeast to Cornwall and east to Montréal. It includes the larger earthquakes near Timiskaming (1935, 1982), Rolphton (1963), North Gower (1983), Cornwall (1944, 1981) and Montréal (1732). The second band, containing more but smaller earthquakes, trends slightly north of northwest and extends from Montréal to the Baskatong Reservoir, about 200 km north of Ottawa. The last decade of monitoring by the Eastern Canada Telemetered Network (ECTN) shows the gap between the two bands is reasonably well defined at the northwestern end by an absence of even small earthquakes; however near the St. Lawrence River the two bands merge.

Field monitoring of aftershock sequences has provided good estimates for the focal depth of some earthquakes (Maniwaki, 17 km; St. Donat, 7 km; Cornwall, 16 km; North Gower, 12 km), and these together with some approximate depths computed from the digital ECTN suggest most earthquakes lie between 5 and 20 km, in the upper crust and within the Precambrian basement.

Focal mechanisms have been determined for about 40 earthquakes in the zone; the large number relative to the rest of eastern Canada reflecting the relatively dense network of digital seismographs. In addition to the four earthquakes just mentioned, Wahlström (1987b) and Adams *et al.* (1988) have derived mechanisms for smaller events, and a selection is shown on Figure 1. Almost all mechanisms have near-horizontal P-axes, and represent mainly thrust earthquakes. This evidence for high horizontal compression is confirmed by other evidence for regional stresses in eastern Canada (Hasegawa *et al.*, 1985; Adams, 1987, 1989). Although the regional stress field appears to have the compression axis in the northeast to east octant, the P-axes for some of the western Québec earthquakes are distinctly different. In a region about 100 km northeast of Ottawa, a patch contains six earthquakes with south-southeast trending P-axes, while outside the patch six earthquakes have northeast-trending P-axes. It appears that within the patch the principal and secondary horizontal stresses have become reversed relative to the regional field, perhaps suggesting the two stress magnitudes are not very different (Adams, 1989).

Forsyth (1981) has shown that the earthquakes in the first band, including the larger historical earthquakes, may be associated with a zone of normal faults along the Ottawa River that were formed in the late Proterozoic and were active into the Paleozoic. Because they offset the youngest bedrock (Ordovician near Ottawa and Silurian at Lake Timiskaming), it is not known when the extensional phase ceased. The normal faults form the Ottawa Graben, a rift structure with Moho expression (Mereu *et al.*, 1986) that has been termed by Kumarapeli (1985) an aulacogen or failed arm of the Iapetan rift system. The Iapetan rift system extends from the Ottawa Valley and from rifts beneath the Appalachian overthrust sheet at Lake Champlain, down the St. Lawrence River (Kumarapeli and Saull, 1966; Kumarapeli, 1985), and perhaps through southern Labrador to the Labrador Sea (Gower *et al.*, 1986). Related structures may extend from Timiskaming northwest to Kapuskasing, Ontario (Forsyth *et al.*, 1983). The main part of the rift — from Montréal to Sept Îles — is the ancient continental margin of North America and the Iapetus Ocean, and is marked by thinned Grenville crust cut by crustal-scale normal faults.

We prefer to call this rift structure the St. Lawrence Rift system because of its clear physiographic association with the St. Lawrence River and its tributaries. However this usage should not be confused with that of Kumarapeli and Saull (1966), who considered there to have been active extension and consequent volcanism on the structure in the Mesozoic. Although the last time of normal movement on these faults is not known, there is tenuous evidence that they might have been reactivated in the Mesozoic during the initial opening of the North Atlantic, as shown by Jurassic kimberlites in Ontario and New York State (Barnett *et al.*, 1984). However, if the rift faults underwent a major reactivation in the Mesozoic it is surprising that so little hard evidence has been found. We consider the rifting to be of Paleozoic age because we feel that it is the age of last significant extensional movement — not inception — that is important in determining contemporary seismic activity.

Regardless of age, recent focal mechanisms of moderate earthquakes (Adams *et al.*, 1988) and a study of seismicity near the 1935 Timiskaming Earthquake suggest the Ottawa Valley rift faults are seismically active. All the Timiskaming earthquakes have occurred in a zone 50 km long by 10 km wide which trends northwest parallel to the rift faults, and appear to have focal mechanisms consistent with thrust faulting on northwest-striking faults (Vonk and Adams, in prep). These observations support the NW-striking of the two possible thrust mechanisms derived for the mainshock by Ebel *et al.* (1986).

Assigning probable causative structures for the second band of seismicity is less easy. Forsyth (pers. comm., 1987) sees no evidence for any well-defined NW-trending features north of the Ottawa River, and earlier Forsyth (1981) suggested that the earthquakes there might be occurring on or between structural boundaries within the NE-trending Central Metasedimentary Belt. We suggest that the second band of seismicity is due to crustal fractures formed during the passage of North America over a hotspot between 140 and 120 million years ago (Crough, 1981). The younger path of the hotspot is marked by the Monteregian Hills, a line of early Cretaceous intrusions that lie mostly east-southeast of Montréal. In turn, these lie parallel to, but somewhat offset from, the still younger New England Seamount Chain (Sykes, 1978). Recent dating of the chain (Duncan, 1984) confirms Crough's inferrence that North America drifted over the hotspot at about 25 mm·yr^{-1}. The rate is consistent with the age of the Monteregian Hills, and of a northwest-trending kimberlite dyke at Kirkland Lake dated at 151 ± 8 Ma (Poole *et al.*, 1970), to the northwest of the seismicity in western Québec. Crough (1981) demonstrates that the passage of the hotspot caused a local uplift of the shield, resulting erosion of at least 1 km at Montréal and perhaps 6 to 7 km in New England. Although the hotspot may have been gaining strength when western Québec passed over it, we suggest that the Precambrian crust of western Québec was thermally stressed and fractured by differential uplift during the passage of the hotspot. This recent weakening of the North American craton has localized the the release of seismic energy today. Under New England, the plutonism in the White Mountains may have been sufficient to heal any deep-crustal fractures. Although central Québec and coastal Labrador are predicted to have passed over similar hotspots at about the same time as did western Québec (Duncan, 1984, fig. 3) there is no enhanced seismicity associated with these hotspot paths.

Charlevoix. The Charlevoix zone is historically the most active in eastern Canada with at least five earthquakes of magnitude 6 or greater (1663, 1791, 1860, 1870, and 1925), which are described by Smith (1962, 1966). Magnitudes for these events were reassessed by Basham *et al.* (1982a). Because of the historical seismicity, Charlevoix is thought to be the most probable site for a future large earthquake in eastern Canada, and so has been intensively studied (Buchbinder *et al.*, 1983, 1988).

Hypocentres located by a 6-station local network since the mid 1970s demonstrate that most earthquakes are confined to a zone that is about 80 km long by 35 km wide (Figure 2), mainly under the St. Lawrence River (Anglin, 1984). A relocation of some of the early instrumentally recorded earthquakes (Stevens, 1980) suggests that the larger earthquakes also occurred in this zone, perhaps initiating preferentially at the ends. The recent largest earthquakes are: 1939 – M5.6; 1952 – M5.2; 1979 – M5.0; and 1986 – M4.2.

Earthquake focal depths are well-determined for the past decade of micro-earthquakes, and are mostly between 5 and 25 km (Figure 2 inset). Paleozoic sediments that crop out on the south shore are only a few kilometres thick, so all the activity is occurring within the Precambrian basement (unshaded area on Figure 2 inset). Stereo plots (Anglin, 1984) demonstrate that most of the micro-earthquakes are occurring on northeast-striking planes that dip to the southeast. A projection of the hypocentres to the surface along the postulated faults (Figure 2) suggests the activity is confined between Paleozoic rift faults mapped on the north shore and a bathymetric feature near the river's south shore. Further, the earthquakes do not extend downriver beyond the cross-cutting Saguenay Graben faults (*e.g.*, Palissades Fault on Figure 2; Basham *et al.*, 1982b).

A number of earthquake focal mechanisms have been produced for Charlevoix. Field experiments in 1974 yielded six mechanisms (Leblanc and Buchbinder, 1977); Hasegawa and Wetmiller (1980) produced a mechanism for the 1979 M5.0 event (see Figure 1); Lamontagne (1987) derived a suite of composite mechanisms; and Adams *et al.* (1988) gave mechanisms of two recent events. No well-determined mechanism is available for the older large earthquakes,

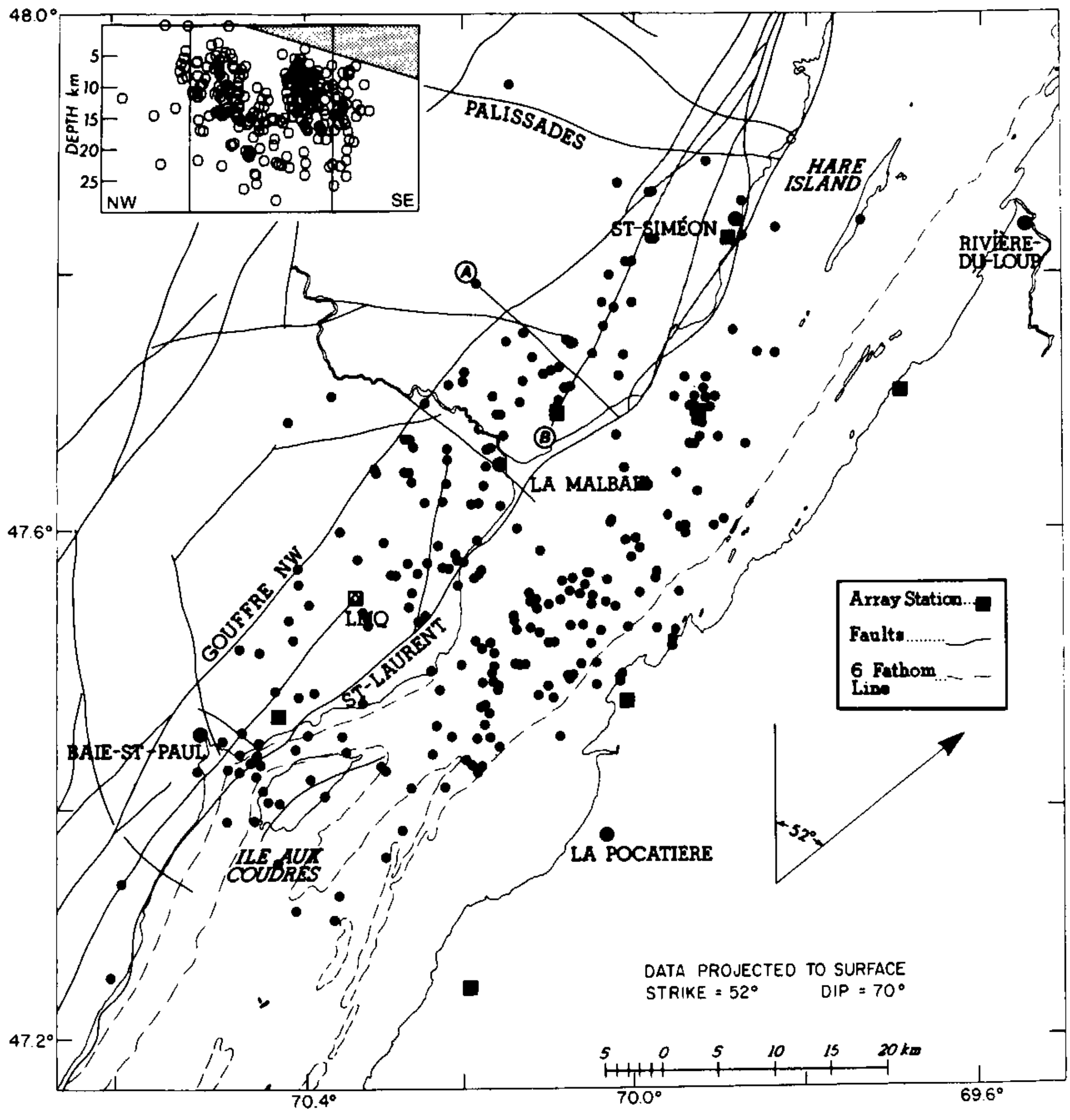

Figure 2 *Micro-earthquakes (1977-1983) in the Charlevoix area (Anglin, 1984). In order to show their relation to the mapped surface faults, the epicentres have been moved to the surface up the dip of the regional rift faults. The inset is a NW–SE cross-section of the hypocentres to show their depth distribution; all lie beneath the Paleozoic sedimentary wedge (shaded).*

although Ebel *et al.* (1986) suggest that both the 1925 and the 1939 earthquakes involved thrust faulting on either NE or NW striking planes.

The available mechanisms have not proved easy to interpret. In general, the derived P-axes lie in the east quadrant, and the mechanisms represent thrust or combination thrust/strike-slip faulting. No plane that might represent the rift faults is clearly identifiable in all of the mechanisms. At least some of the earthquakes appear to be occurring on small NW-trending transverse faults that offset the rift fault system (Figure 2).

Charlevoix, the Ottawa River, and the lower St. Lawrence seismic zones all lie along the St. Lawrence rift system. However, at Charlevoix the normal faults of the ancient rifted continental margin have been complicated by a late Devonian meteorite impact that caused ring faulting and distributed fracturing (between Baie-St-Paul and La Malbaie on Figure 2). Lamontagne's (1987) composite mechanisms suggest that micro-earthquakes within the impact structure have more varied mechanisms than those outside it, due to the extra structural complexity caused by the impact. Considerable attention is being paid as to whether the earthquakes at Charlevoix occur because of the associated impact structure (in which case the seismicity could be considered to be localized and unlikely to occur elsewhere) or whether the earthquakes just happened to be coincident with the impact (in which case other parts of the rift system could become similarly active). Because other meteorite craters in Canada are not seismically active, because the earthquakes at Charlevoix extend downriver beyond the impact structure, and because the hypocentral trends suggest reactivation of the rift faults, current opinion is that the impact structures are not the controlling factor in the seismicity.

Lower St. Lawrence. Similar to Charlevoix, the Lower St. Lawrence earthquakes also occur mainly under the St. Lawrence River and may involve reactivation of the Iapetan faults, though a correlation was previously made between the seismicity and gravitationally-induced stresses (Goodacre and Hasegawa, 1980). The record of felt earthquakes extends back at most 100 years in this sparsely populated area, and none is likely to have been much larger than M5. The largest earthquake with a well-determined magnitude is M4.8 in 1944. Epicentres located by the ECTN in the last five years lie almost exclusively under the river, which at this point is an estuary 50-100 km wide. Despite the lack of known larger earthquakes, the zone has magnitude 3 and 4 earthquakes as often as the more confined Charlevoix zone.

Early epicentre maps (Smith, 1962, 1966) show a scattering of earthquakes extending onto the north and south shore. Relocation of some of these early epicentres (Adams *et al.*, 1989) has affirmed that most actually occurred under the river (Figure 3). Half of the relocated epicentres lie just offshore but parallel to the northern shoreline, and are inferred to be occurring on offshore faults that have controlled the shape of the coastline. The only confirmed significant activity on the north shore involves a M4.1 earthquake in 1975 and its aftershocks induced by the filling of the Manic 3 reservoir, and a cluster of M3 events in 1966 spatially associated with the Manic 2 reservoir (Leblanc and Anglin, 1978).

Reliable estimates of earthquake depth have been obtained for two earthquakes recorded in 1983 and 1984 by the Yellowknife array, 30 degrees to the northwest. Phases interpreted as pP and sP gave hypocentral depths of 17 km and 19 km (Adams *et al.*, 1988). Other earthquakes for which approximate depths have been computed from the ECTN network lie mostly between 10 and 20 km. Like the Charlevoix zone, this places them within the Grenville basement and so beneath both the down-faulted sediments of the St. Lawrence platform and the overthrust sedimentary rocks of the Appalachian Orogen. The Manic earthquakes lie 50-100 km outside the rift system and, being induced earthquakes, are shallower than earthquakes under the St. Lawrence River.

Focal mechanisms have been derived for seven earthquakes (Adams *et al.*, 1988); they have P-axis orientations only slightly less variable than at Charlevoix, and also indicate mostly thrust faulting in response to compression from the east quadrant. Significantly, five of the mechanisms have a common plane that strikes parallel to the river and dips to the southeast; the remaining two represent thrust faulting on NW-trending planes. Taken together, the northeast-striking focal planes, the position of the larger earthquakes relative to the coastline, and the distribution of the smaller earthquake epicentres suggest that the Paleozoic rift faults are the chief active structure. Composite mechanisms of the Manic 3 induced earthquakes indicate thrust faulting on NW-trending planes (Leblanc and Anglin, 1978; Adams *et al.*, 1988), in agreement with the regional stress field.

Northern Appalachians. The northern Appalachian region, which includes most of New Brunswick and which extends into New England, is a zone of relatively uniform seismicity. Significant historical earthquakes include those near Passamaquoddy Bay on the New Brunswick/Maine border in 1817 (M4.8), 1869 (M5.7), and 1904 (M5.9), and near Moncton in 1855 (M5.2) (Leblanc and Burke, 1985).

Recent, well-located seismicity has occurred throughout New Brunswick (perhaps excluding the NW corner), the Bay of Fundy, and southwestern Nova Scotia. Significant recent events in the region include Québec-Maine border 1973 – M4.8 (Wetmiller, 1975; Yang and Aggarwal, 1981), Miramichi 1982 – M5.7, M5.1, M5.4, M5.0

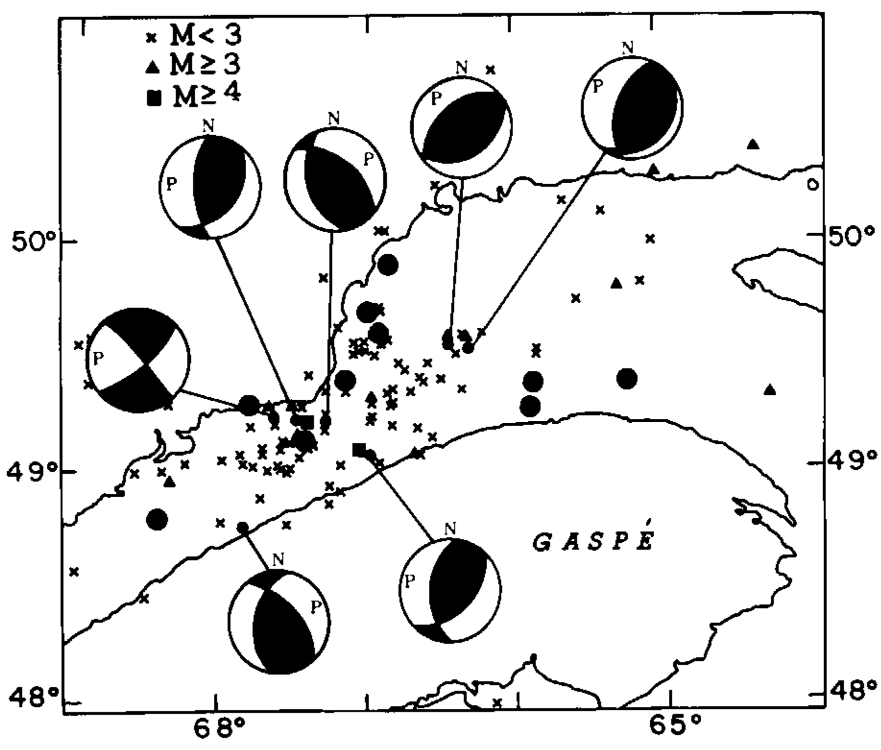

Figure 3 *Lower St. Lawrence seismicity from 1981-1987. Large circles represent relocated epicentres of some M3.5 to M4.8 earthquakes from the period 1944-1968. Focal mechanisms shown are those derived by Adams* et al. *(1988).*

(Wetmiller *et al.*, 1984), Trousers Lake 1982 – M4.7 (Wetmiller *et al.*, 1984), and a series of six M3 earthquakes near Passamaquoddy Bay in 1983-84. Within New Brunswick, apparent concentrations of earthquakes occur in the Miramichi Highlands and near Passamaquoddy Bay. Outside of the northern Appalachians zone, the general lack of recent seismicity in the rest of Nova Scotia, in Prince Edward Island, and in the Gulf of St. Lawrence has been confirmed for the historical record by searching old newspapers for felt earthquake reports (Ruffman and Peterson, 1988). There is also a lack of seismicity in the Gaspé, Québec and in the rest of the Appalachian fold belt through the island of Newfoundland.

Focal mechanisms are available for the Miramichi mainshock and several clusters of its aftershocks, and for the Trousers Lake earthquake (Wetmiller *et al.*, 1984), for two earthquakes in east-central New Brunswick (Adams *et al.*, 1988), for two earthquakes near the Québec-Maine border (Wetmiller, 1975; Yang and Aggarwal, 1981; Adams *et al.*, 1988), and for an earthquake at Passamaquoddy Bay (Ebel, 1985) (summarized on Figure 1). All represent dominantly thrust faulting, all but the one at Passamaquoddy Bay, in response to northeast- to east-directed compression.

The Miramichi earthquakes, because of their size and numerous aftershocks, have been unusually well studied. Even today (1988), six years after the mainshock sequence, M3 aftershocks are occurring every few months and recent aftershocks have been as large as M4.1. Field aftershock surveys in 1982 (Wetmiller *et al.*, 1984) established the general pattern of activity.

The 9 January 1982 mainshock, M5.7, has been interpreted as a thrust with rupture up-dip on a west-dipping plane. A magnitude 5.1 aftershock occurred 3.5 hours later, probably on the lower northern portion of this plane. On 11 January, a M5.4 aftershock ruptured (probably up-dip) a conjugate east-dipping plane and was followed by an intense aftershock sequence. Finally the 31 March M5.0 aftershock occurred as a repeat rupture on the upper northern portion of the west-dipping plane (Wetmiller *et al.*, 1984; Basham and Adams, 1984; Basham and Kind, 1986). Micro-earthquake monitoring for short periods in 1983 and 1985 confirmed that the epicentral region remained active and may have expanded slightly with time. Although considerable effort, including trenching, was made to locate a surface rupture, none was found, suggesting that the expected near-surface vertical offset of 0.3-0.5 m was distributed over a zone several hundred metres wide and not confined to a single plane.

Detailed geological and geophysical investigations have confirmed that the earthquakes all occurred within a single granodiorite pluton, but not on the obvious WNW-trending faults. No strong reason has emerged as to why the earthquakes occurred at the Miramichi site. If Miramichi-type earthquakes occurred regularly at the site, more surface evidence for thrusting, such as a degraded fault scarp, would be expected. Therefore, despite considerable effort, we do not understand why the earthquakes occurred where they did, and have no evidence that they occur often at the Miramichi site. Thus we must consider that Miramichi-sized earthquakes could occur anywhere in the northern Appalachians zone.

While aftershocks of the Miramichi and Trousers Lake earthquakes were shallow, all less than 9 km, earthquakes near Passamaquoddy Bay are deeper, perhaps 10-16 km, based on a 12 km depth from waveform modelling for a 1984 earthquake (Ebel, pers. comm., 1988) and computed depths for other earthquakes. Earthquakes in the Appalachian part of Canada may thus be significantly shallower than in the Grenville basement at Charlevoix and western Québec, perhaps reflecting a thinner brittle upper crust in the younger Appalachian belt (Hasegawa, 1986, fig. 9). In the Appalachians of New England, all earthquakes appear to be shallower than 13 km (Ebel, 1984). The shallow depths may indicate that the earthquakes are confined to the rocks above a shallow, sub-Appalachian detachment zone, though judging from the Québec-Maine seismic reflection line the detachment dips more steeply than in the southern United States and may not extend beneath southern New Brunswick (Spencer *et al.*, 1989). Northern New Brunswick seismicity may be compared with the central Virginia seismic zone where earthquakes occur above the detachment (Bollinger *et al.*, 1989), and it may follow the general style of Appalachian seismicity with earthquakes above the detachment to the east and below the detachment to the west of the crystalline overthrust sheet (Seeber and Armbruster, 1988), perhaps because, as at Charlevoix, the sedimentary rocks above the detachment are insufficiently brittle. Inhibition of seismicity in the basement beneath the eastern part of the overthrust sheet might be due to thermal or other effects. If a sub-Appalachian detachment were important for controlling Appalachian seismicity, it is possible that the Miramichi earthquakes nucleated at the base of the overthrust sheet and ruptured upward to the surface, while the Trousers Lake mainshock and aftershocks — all at essentially the same depth — may have occurred on the detachment itself.

Southeastern Continental Margin.
Although poorly monitored and little studied, the seismicity of the southeastern continental margin of Canada is clearly higher than that at many comparable passive margins, and higher even than the same margin off the eastern United States. The description below is condensed from Adams (1986), which relies heavily on data as yet unpublished.

About half of the earthquakes off the southeastern margin occur in the Laurentian Slope seismic zone, site of the M7.2 "Grand Banks" earthquake of 1929 (Figure 4). This earthquake caused a large submarine slump and consequent tsunami that killed 27 people on the south coast of Newfoundland. The larger historical earthquakes, and many recent smaller earthquakes, have been systematically relocated and found to lie within a 100 km E-W by 35 km N-S box at the mouth of the Laurentian Channel (Adams, 1986). In addition to the 1929 earthquake and its immediate aftershocks, there have been four M5 earthquakes since 1951, the most recent in 1975. The elongation of the seismic zone, and the location of the 1929 epicentre at the eastern end are consistent with Adams' 1986 hypothesis that current earthquakes represent belated aftershocks of the 1929 earthquake. If so, the earthquake appears to have ruptured westward along a fault about 70 km long.

Little is known about the source properties of these earthquakes. The 1975 earthquake appears to have occurred at a depth of 30 km, placing it in the upper mantle in this area of thinned continental crust, and appears to have involved thrusting on planes which, though not tightly constrained, probably strike roughly east-west, *i.e.*, in response to compression from the north or northeast (Hasegawa and Herrmann, 1989). The depth and mechanism of the 1929 earthquake are uncertain as the available seismograms are generally poor. Hasegawa and Kanamori (1987) have suggested that the long-period seismic source may have been the large submarine slump itself, although they admit that the slump may have been triggered by a "smaller" earthquake. Their explanation, if correct, has significant implications for the seismotectonic interpretations of large earthquakes on Canada's eastern margin, or indeed for other continental margins.

Our problems with the slump hypothesis are: (1) The fact that an epicentre can be computed for the event suggests that it was a sudden, high-energy event, and is not consistent with a slowly propagating slump. (2) The only energy in the slump mass is its potential energy, and it is not clear that the mass can travel downslope fast enough for its energy to appear in the seismic coda; (3) The slump mass required by the Hasegawa and Kanamori model is 5-10 × larger than that computed for the initial slump by Piper and Aksu (1987); (4) The computed intermediate-period body-wave and long-period surface-wave magnitudes are similar (Hasegawa and Kanamori, 1987, table 1), suggesting that the longer periods were not greatly augumented by low frequency slump-generated energy relative to shorter periods, and that much of the energy was released in the first 20-40 seconds; (5) The immediate aftershocks (up to magnitude 6) and the continuing seismicity are not easily explained; (6) The one earthquake of known depth (1975) is a mantle event; and (7) A magnitude 6-6.5 triggering earthquake is not ruled out by the data, particularily if its mechanism produces a S-wave node at Uccle and Kew.

Outside of the Laurentian Slope zone, scattered earthquakes occur northeast of Newfoundland and seaward of Nova Scotia. Some of the epicentres off northeast Newfoundland correlate with fracture zones (see Figure 6 and section on Labrador Sea seismicity below), and perhaps with a structural trend extending northeast from eastern Newfoundland to the continental margin, but others cannot yet be related to structure. South from the Laurentian Slope zone, a trend of seismicity of unknown origin occurs along the Laurentian Fan.

Off Nova Scotia, the earthquakes on Figure 4 suggest that the transition zone from oceanic to continental crust is active, though not currently at a very high level. None of these earthquakes has been larger than M4. Elsewhere, the transition is too poorly monitored to show if it is active at the low level we can detect off Nova Scotia.

The cause of earthquakes along the eastern margin and thus a rationale for their distribution are established in only a general way. Studies of stress directions from oil well breakout data (Podrouzek and Bell, 1985) confirm that the margin is subject to the same northeast-directed compression as the rest of eastern North America (Adams, 1989). While some studies have indicated local areas of near-surface faulting (*e.g.*, Orpheus Graben off Nova Scotia, Durling and Fader, 1986), many of the earthquakes along the margin probably occur near the ocean-continent transition on the deep-crustal rift faults formed during the opening of the Atlantic Ocean. Under the current northeast compression regime, these normal faults would be reactivated as thrust or strike-slip faults.

As the numbers, location, and nature of the offshore earthquakes are poorly understood, Basham and Adams (1983) produced a speculative model that suggests rare large earthquakes can occur along the whole margin, using the pervasive Mesozoic rift faults as the causative structure, with a rate of about one M7 earthquake per thousand years per thousand kilometres of margin. The model implies that inactive parts of the margin are quiescent but potentially seismogenic, and that recent earthquakes in active regions like the Laurentian Slope represent belated aftershocks that will diminish over the next century without producing another M7 earthquake. That such large earthquakes may have occurred in the past is suggested by prehistoric submarine slumps elsewhere along the margin (Piper *et al.*, 1985).

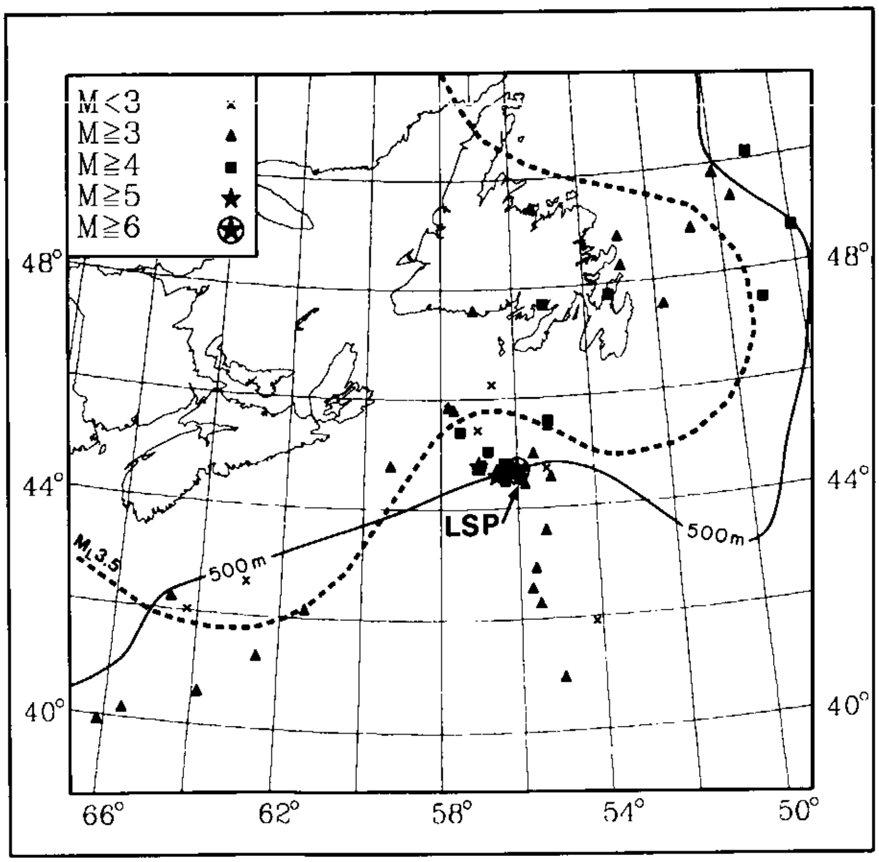

Figure 4 *Seismicity of the the southeastern margin (Adams and Wahlström, unpublished). Earthquakes in Nova Scotia and to the northwest are omitted. Inland of the striped line, M3.5 and larger earthquakes are thought to have been completely detected and located since 1983. Arrow marked "LSP" points to the many earthquakes near the epicentre of the 1929 "Grand Banks" earthquake. The 500 m isobath approximates the position of the ocean-continent transition.*

Southeastern Background Seismicity. Some scattered seismicity lies in less intense clusters outside of the seismic zones discussed above with reference to Figures 1 and 4, and is shown only on Figure 7. Here, we briefly mention these areas, again moving from west to east. The shield areas of Ontario (Basham and Cajka, 1985; Wetmiller and Cajka, 1989) and Québec show very low seismicity except for earthquakes near Cochrane and Iroquois Falls in northern Ontario, which may lie on an extension of the St. Lawrence rift towards Kapuskasing (Forsyth *et al.*, 1983). Further north, the earthquakes around and in James Bay have an unknown origin. A cluster of small earthquakes in the Burlington/Niagara Falls area of Ontario is poorly understood, but in part may represent very shallow stress release (Wetmiller, 1980), and in part may be related to seismicity that extends from Lake Ontario and Lake Erie to Ohio and into the central United States. Earthquakes near Québec City may lie on the St. Lawrence rift system. The earthquakes extending from Sept Îles across easternmost Québec and southern Labrador may lie on a strike-slip fault related to the St. Lawrence rift system (mapped by Gower *et al.*, 1986).

Northeastern Canada

We continue our discussion of eastern Canadian seismicity and seismotectonics by proceeding northward along the continental margin of the eastern Arctic, inland to the northeastern portion of the Canadian Shield, northward into the Arctic archipelago, and completing the picture with a brief discussion of the Arctic Ocean margin (Figure 5).

Labrador Sea. The seismicity of the Labrador Sea includes six earthquakes in the magnitude 5.0 to 5.6 range since 1934. The most recent moderate earthquake was M4.7 in 1986, which though 200 km offshore, was felt in Nain, Labrador. There are older reports of felt earthquakes from fishing villages along the Labrador coast as early as 1809 (Staveley and Adams, 1985). Currently, epicentres for these older events are assigned to the locations at which they were felt. However, 25 years of monitoring by the Canadian Network has provided no evidence that significant earthquakes are occurring onshore in this region. These older events likely occurred offshore, and may have been larger than hitherto thought (Basham and Adams, 1983).

The Labrador Sea was produced by rifting in the middle Cretaceous and sea-floor

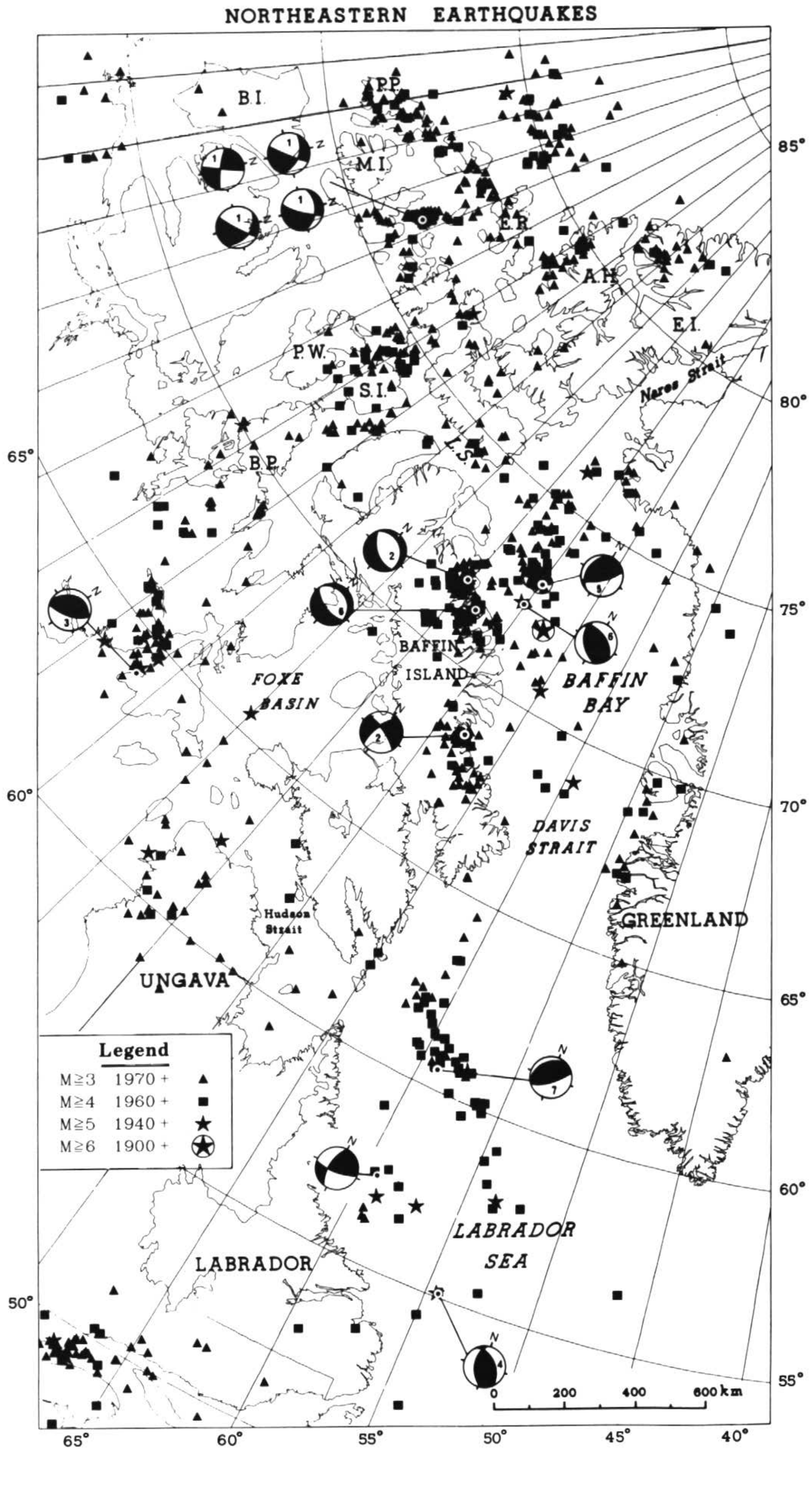

Figure 5 *Seismicity of northeastern Canada showing a selection of earthquakes prior to July 1987, as detailed in the legend. Focal mechanisms (lower hemisphere projections with compressional quadrants shaded) are from:*
1, *Hasegawa (1977);*
2, *Hashizume (1973);*
3, *Hashizume (1974);*
4, *Hashizume (1977);*
5, *Sleep et al. (1989);*
6, *Stein et al. (1979);*
7, *Sykes and Sbar (1974);*
(blank), Adams (unpublished).
Geographic features are identified by initials as follows:
AH, *Axel Heiberg Island;*
BI, *Banks Island;*
BP, *Boothia Peninsula;*
EI, *Ellesmere Island;*
ER, *Ellef Ringes Island;*
LS, *Lancaster Sound;*
MI, *Melville Island;*
PP, *Prince Patrick Island;*
PW, *Prince of Wales Island;*
SI, *Somerset Island.*

spreading between 95 and 50 m.y. ago. Srivastava and Tapscott (1986) have identified a central ridge and associated fracture zones from seismic and gravity profiles and linear magnetic anomalies. Simmons and Adams (1989) systematically relocated earthquake epicentres in the Labrador Sea and showed that seismicity can be associated with the extinct ridge and fracture zones (Figure 6), even though the spreading ridge has been extinct for 50 m.y. The apparent decrease in activity to the southeast along the ridge is due to an increase in the earthquake detection threshold away from station FRB on Baffin Island.

No earthquakes are known in the Labrador Sea between the Labrador Ridge and Greenland to the northeast (perhaps also a problem with detection threshold), but on the Canadian side a separate trend of seismicity occurs at the continental margin off Labrador. In general terms, we associate the earthquakes with pre-existing faults near the inactive ridge (a relatively young zone of weakness within the oceanic crust) and beneath the rifted continental margin. The best-studied ridge earthquake is the 1969 M5.0 (Sykes and Sbar, 1974) which represents compression along the ridge axis. On the slope, Hashizume (1977) determined a thrust mechanism for the magnitude 5.6, 1971 earthquake at a depth of 16 km implying compression normal to the margin, and Adams (unpublished data, 1987), a mechanism for the 1986, M4.7, implying compression along the margin (Figure 5).

North of the point where the ridge meets the continental margin, the Southeast Baffin Shelf shows a trend of small earthquakes along the former transform continental margin toward Davis Strait (Figure 5). Davis Strait itself appears to show little activity, at least at current detection levels (M about 3.5), until the southern portion of Baffin Bay is reached.

Baffin. The largest earthquake known to have occurred in northeastern Canada was the M7.3 event in Baffin Bay in 1933. This earthquake had aftershocks as large as M6.5, and M6 events have since occurred in the Bay in 1945, 1947 and 1957. Precise boundaries for the Baffin seismicity are difficult to define on the basis of geological and geophysical data, but there seems to be some evidence for a separation between the activity in the Bay (partly on oceanic and partly on transitional crust) and that on Baffin Island on true continental crust.

Jackson *et al.* (1979) found evidence for sea-floor spreading and an extinct spreading centre in the deep central region of Baffin Bay. Srivastava and Tapscott (1986) suggest rifting could have initiated in the latest Cretaceous and continued into the early Tertiary. However, there is now little or no seismic activity in the deep central Bay, and the seismicity is confined almost exclusively to the landward side of the 2000 m bathymetric contour in the northwestern segment of the Bay, which delimits the thick sedimentary sequence (Basham *et al.*, 1977). Sleep *et al.* (1989) have established a focal depth of 10 km for the 1933 event (a correction to the 65-km depth estimated by Stein *et al.*, 1979) and 24 km for a 1976, M5.4 further south in Baffin Bay.

Prior to 1960, only one earthquake is known to have occurred on Baffin Island, magnitude 5.5-6.0 in 1935, but with the development of the northern Canadian seismograph network in the 1960s the northeastern portion of the island was found to be highly active. The seismicity appears to be confined to the coastal region and does not occur much further inland than the heads of the fjords. The earthquakes are concentrated in the regions of Buchan Gulf and Home Bay with a gap between these two regions, but because of the short history and the swarm-like nature of the seismicity, this gap may be a zone of temporary quiescence. In any case, the tectonic significance of such gaps is unclear. All available evidence on focal depths suggests the earthquakes are shallow. Hashizume (1973) determined depths of 6-9 km for the 1970 – M4.4 and 1972 – M5.1 events, Liu and Kanamori (1980) a depth of 7 km for the 1963 – M6.1 event. A temporary deployment of ocean-bottom seismometers and land stations also established shallow focal depths on the island (Reid and Falconer, 1982).

Focal mechanisms are consistent with thrust faulting for Baffin Bay earthquakes of 1933 – M7.3 (Sleep *et al.*, 1989) and 1976 – M5.4 (Stein *et al.*, 1979). In contrast, the island earthquakes show normal faulting, which is rare for eastern Canada: 1963 – M6.2 (Stein *et al.*, 1979) and 1972 – M4.5 (Hashizume, 1973). On the basis of these focal mechanisms, Stein *et al.* (1979) proposed a model for the entire passive margin of eastern Canada with thrust faulting seaward of the 1000 m isobath and normal or strike-slip faulting for more landward events as due to Pleistocene deglaciation reactivating faults remaining from the rifting. Sleep *et al.* (1989) developed this model further for the Baffin region, suggesting that the recent removal of surface loads by glacial erosion and by deglaciation is the major source of local stress that produces flexure of the lithosphere, and is comparable in magnitude to and superimposed on the compressional stress in the North American plate due to Mid-Atlantic Ridge spreading. The spreading stress enhances the shallow tension produced by flexure within the continental lithosphere and shallow compression produced by flexure in the oceanic lithosphere. Quinlan (1984) has argued, however, that post-glacial rebound is capable of triggering earthquakes in pre-stressed regions but rarely capable of dictating the focal mechanism of these earthquakes.

The region surrounding the Baffin seismicity discussed above, and the continental margin of Greenland adjacent to Baffin Bay, have experienced low levels of both historical and recent seismicity. A trend of epicentres appears to extend westward from northern Baffin Bay into Lancaster Sound (Wetmiller and Forsyth, 1982), perhaps along the Cretaceous-Tertiary Lancaster Aulacogen (Kerr, 1980). Wetmiller and Forsyth (1982) have shown that Nares Strait, between Ellesmere Island and extreme northwestern Greenland, is currently aseismic. The region around Nares Strait marks the Tertiary collision of Greenland with North America, and deformation during that event may have sealed any old zones of weakness.

Boothia – Ungava. Basham *et al.* (1977) recognized an arcuate band of seismicity that extends southward from the Boothia Peninsula, across northern Hudson Bay, the Ungava Peninsula, and eastward through Hudson Strait, connecting with the northern end of the Labrador Sea. With the centre of post-glacial uplift over Foxe Basin and a high differential rate of uplift on northeastern Baffin Island, they speculated that the Baffin Island-Foxe Basin block is responding independently to post-glacial uplift (see also Hasegawa and Basham, 1989), and may be decoupled from the rest of the Shield to the southwest along the arc of seismicity. The correlation to geology is best at the north end where the seismicity lies along the Boothia Uplift from Somerset and Prince of Wales Islands northward, meeting the Sverdrup Basin in the region of Grinnell Peninsula on the northwest tip of Devon Island (Wetmiller and Forsyth, 1982). The Boothia Uplift, which has been geologically active from the Paleozoic to the Cretaceous (Okulitch *et al.*, 1986), was most recently active in a Cretaceous-Tertiary rifting episode in which north-south normal faults followed structural trends established earlier (Kerr, 1977, 1980). The Uplift continues to be seismically active at present, though Hashizume's (1974) focal mechanism of a M5.0 at 21 km depth on Southampton Island (Figure 5) represents thrust faulting in response to northeast compression rather than extension.

Sverdrup Basin. Most of the remaining seismicity in the Canadian Arctic archipelago can be spatially associated with the Sverdrup Basin, a 1000 km long northeast-southwest regional depression (between Melville and Ellef Ringnes Islands) in which the sedimentary rocks reach thicknesses of 10 km. The seismicity is characterized by intense low-magnitude swarms, such as that which occurred on Prince Patrick Island in 1965 (Smith *et al.*, 1968), by intense moderate-magnitude swarms such as that which started abruptly in 1972 in the Byam Martin Channel northeast of Melville Island (Basham *et al.*, 1977; Hasegawa, 1977), and by single earthquakes with only small aftershocks such as the M5.2 event on western

Axel Heiberg Island in 1975. Because of the swarm-like nature of the seismicity and the short instrumental observation period (25 years), it is unlikely that all potentially active regions of the basin have been identified.

A feature running through the basin that shows some geological and geophysical correlation with the highest levels of seismicity is the Gustaf-Lougheed Arch (Forsyth *et al.*, 1979; Basham *et al.*, 1982a). This arch is a structurally significant feature visible in the Bouguer anomaly contours that divides the western Sverdrup Basin into two separate sub-basins. Superimposed on the arch is a series of northeasterly trending minor magnetic highs reflecting mineralized faults or intrusive dykes.

The focal mechanisms for the four largest earthquakes in the Byam Martin Channel swarm (M5.1-5.7), while dominantly strike-slip, show some deviatoric tension at depths from 9 km (near the base of the sedimentary rocks) to 31 km (Hasegawa, 1977). This suggests that the fractures or dykes may be the loci of current activity, and there may remain a small component of tension similar to that responsible for the opening of the Arctic Ocean Basin in the early Cretaceous (Forsyth *et al.*, 1979).

Arctic Ocean Margin. The seismicity along the Arctic Ocean margin offshore of the Canadian Arctic archipelago is concentrated in distinct clusters in the Beaufort Sea and northwest of Ellef Ringnes Island, with only very scattered activity elsewhere (Basham *et al.*, 1977; Wetmiller and Forsyth, 1978; Hasegawa *et al.*, 1979; Wetmiller and Forsyth, 1982; Forsyth *et al.*, 1988). The largest earthquake, M6.5, occurred in the Beaufort Sea in 1920. Elsewhere along the margin there are no known events larger than about M5.5, and no focal depths or focal mechanisms are yet available.

The rifted margin was formed in early Cretaceous time, possibly when northern Alaska rotated anticlockwise away from Arctic Canada (Sweeney *et al.*, 1978). The ocean-continent transition is characterized by a zone of negative magnetic anomalies that extend from the Beaufort Sea to north of Ellesmere Island. A series of elliptical free-air

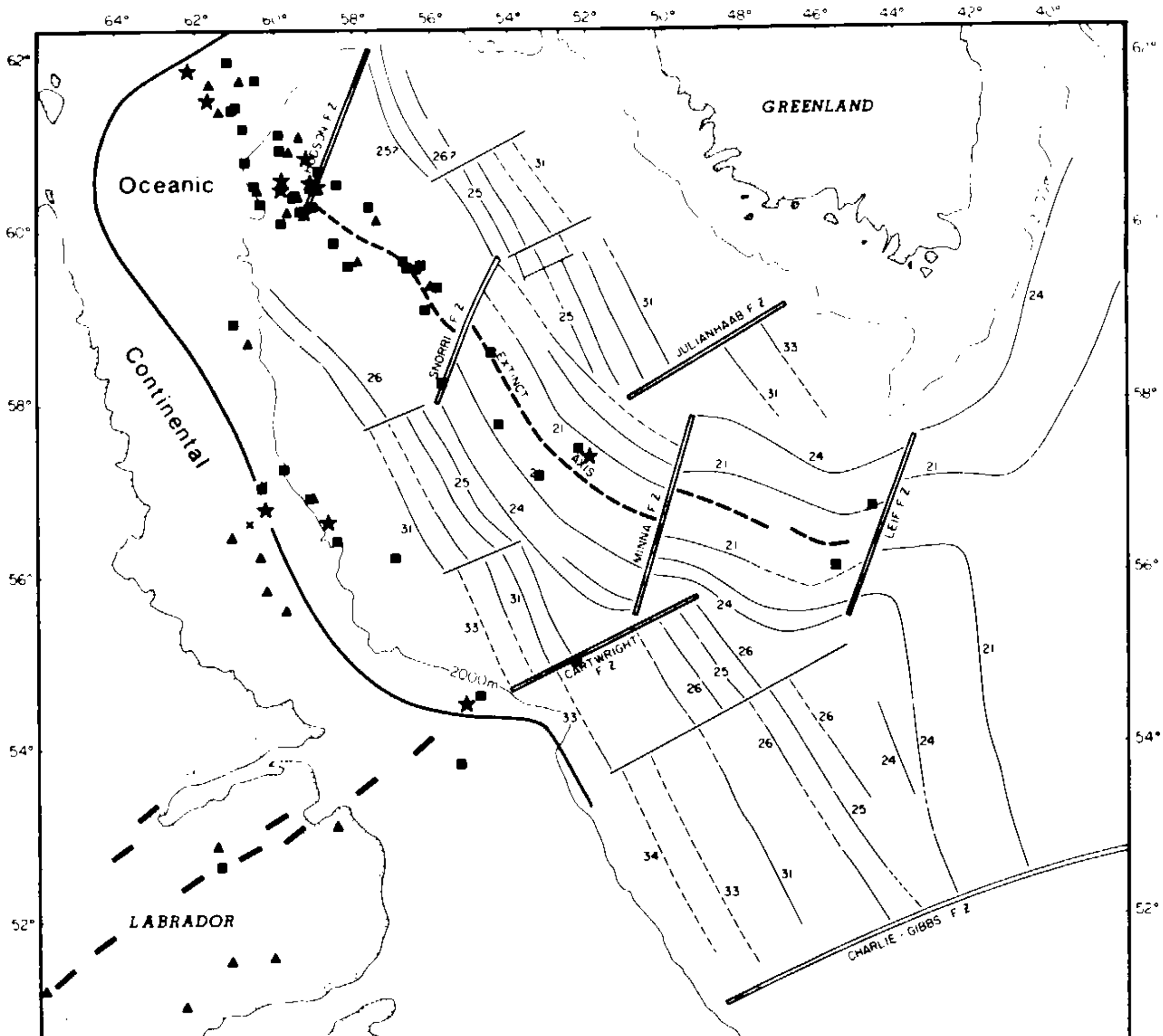

Figure 6 *Seismicity of the Labrador Sea (Simmons and Adams, 1989) showing relation of earthquakes to the extinct spreading ridge and transform faults (as taken from Srivastava and Tapscott, 1986), to the boundary between continental and oceanic crust, and to a zone of crustal weakness that extends across southern Labrador.*

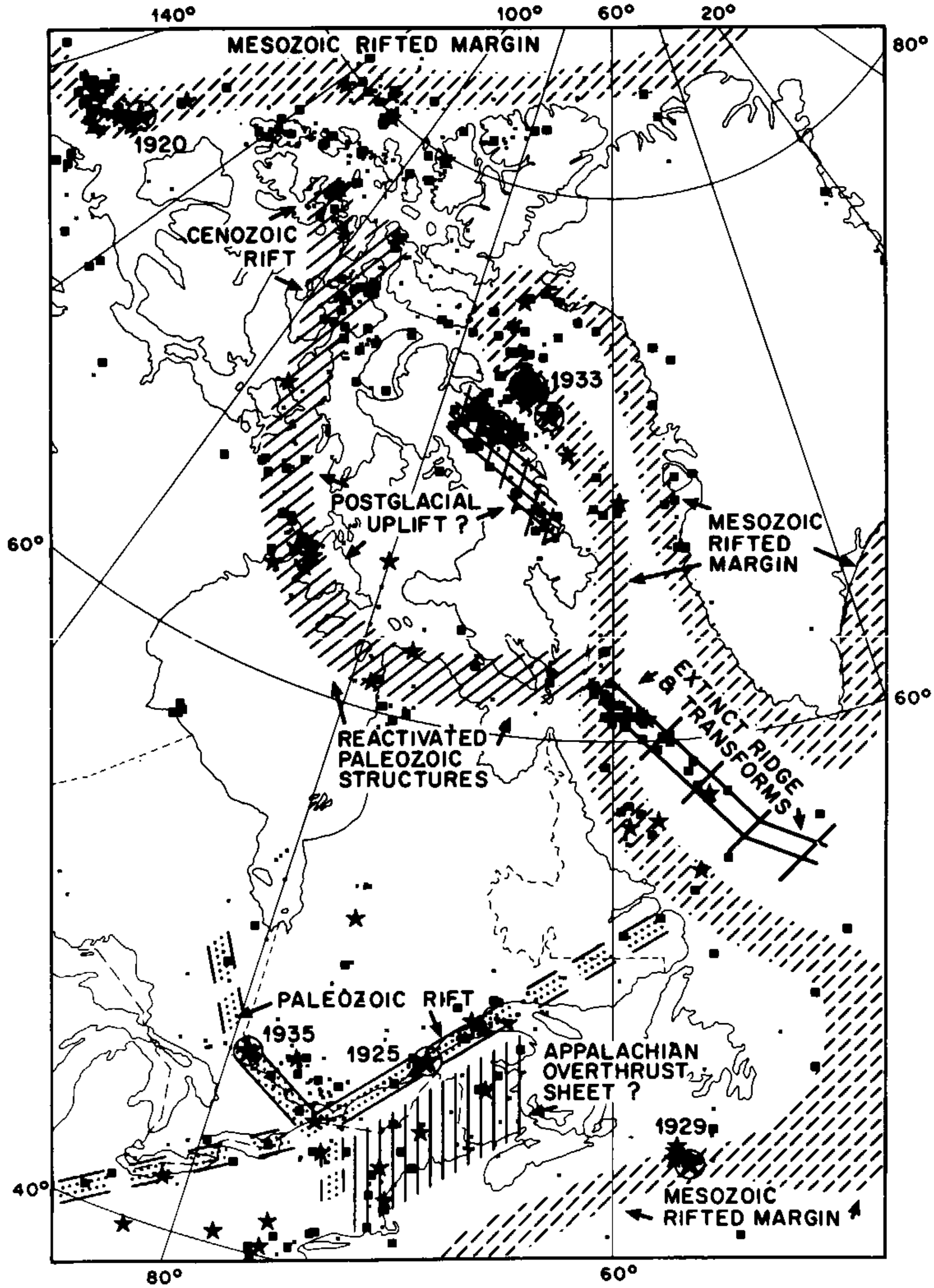

Figure 7 *Earthquakes of eastern Canada (M > 3 since 1970; M > 4 since 1960; M > 5 since 1940; M > 6 since 1900) together with an interpretative framework for the cause of the seismicity. Note that the extension of the Paleozoic rift structures through Lakes Ontario and Erie is speculative.*

gravity anomalies lie over major sediment accumulations near the shelf-slope break. Forsyth *et al.* (1988) have recently provided the following interpretation of the seismicity in terms of gravity and magnetic anomalies, bathymetry, and margin structures. The zone of rift faults separating continental and oceanic crust is inferred to lie immediately seaward of the magnetic lows. Although four major elliptical gravity anomalies lie along the margin, significant seismicity is associated with them only where the shelf break extends distinctly seaward of the magnetic low; *i.e.*, where the sediments have prograded over more oceanic crust. This suggests that earthquakes occur on the rift-related structures chiefly where the oceanic or transitional crust is loaded by sediments. Seismicity is much lower where the sediment is loading continental crust (*e.g.*, northwest of Banks Island). Perhaps for similar reasons, very little of the Beaufort seismicity extends landward of the gravity and magnetic anomalies, though the seismicity northwest of Ellef Ringnes Island extends onto the shelf and may connect with the seismicity of the Gustaf-Lougheed Arch discussed above.

Summary

We summarize our conclusions regarding the causes of seismicity in eastern Canada on Figure 7. Almost all the significant earthquakes in the continental part of southeastern Canada can be spatially and probably causally associated with the Paleozoic rift system along the St. Lawrence, the chief exceptions being the northern band of seismicity in western Québec (probably related to the hotspot trace) and the Appalachian seismicity southeast of the rift system. To a first approximation, the continent-wide stress field is uniform, represents compression from the east-northeast octant, and causes thrust or thrust/strike-slip earthquakes.

Along the eastern margin of the continent, the seismicity includes the 1929 M7.2 Grand Banks and 1933 M7.3 Baffin Bay earthquakes. These and smaller earthquakes concentrated near the ocean-continent transition appear to be thrust events ocurring through the reactivation of the Mesozoic rift faults formed during the opening of the Atlantic. In the mid Labrador Sea, earthquakes are associated with the extinct spreading ridge and its associated transform faults.

In Arctic Canada, continental earthquakes occur on Baffin Island, along an arcuate band between the Boothia and the Ungava peninsulas, and in the Sverdrup Basin. The Baffin and Boothia-Ungava earthquakes are spatially associated with steep gradients in the post-glacial uplift rate, suggesting that they may occur because of differential uplift. The Baffin Island earthquakes are unique in eastern Canada in that they represent normal faulting. The Sverdrup earthquakes represent strike-slip deformation beneath a thick accumulation of sediments. The passive Arctic Ocean margin has a rifted ocean-continent transition comparable to the Atlantic margin, but it appears to be seismically active mainly where it has been recently loaded by thick sediments.

It is therefore clear from the above discussion and Figure 7 that most of the larger earthquakes can be associated with the Paleozoic or younger rift systems that surround or break the integrity of the North American craton. By contrast, the largest earthquakes in the unbroken craton (in Canada, but outside the seismic zones discussed above) are probably not much larger than M5. Coppersmith *et al.* (1987) and Johnston (1989) have come to similar conclusions from a study of world-wide earthquakes in "stable continental interiors". They found that 71% of the seismicity of stable continental interiors was associated with imbedded continental rifts and continental passive margins (one-sided rifts). Further, all of their 17 $M \geq 7$ or larger earthquakes are strongly associated with the imbedded rifts or passive margins.

Both the Paleozoic and the Mesozoic rift systems in eastern Canada are continuous features, many thousands of kilometres long. Despite their continuity and the uniform stress field, the rift systems are only sporadically active, showing seismicity in clusters — as at Charlevoix and the lower St. Lawrence — or single large earthquakes — such as the "Grand Banks" earthquake.

For both rift systems, there is a lack of geological evidence for continual activity at present rates. We, like many others, have been confounded by the high levels of seismic activity at Charlevoix relative to other places in eastern Canada. At such high rates (M7 every few hundred years), the implied rates of geological deformation would amount to kilometres over a million years. Clearly kilometres of uplift (if due to thrusting), or even kilometres of strike-slip motion, would have been recognized at Charlevoix, had they occurred. That they have not, we infer must be due to intermittent activity at Charlevoix and at other parts of the rift system, perhaps with a time constant of thousands to tens of thousands of years.

For single large earthquakes such as the "Grand Banks" a similar argument can be made for intermittent activity on comparable time scales. This leads to the classic dilemma for hazard estimates that need to be made for low levels of probability: will the next large earthquake occur in a recognized seismic zone or not?

Note Added Following Review

After this paper was submitted to *Geoscience Canada*, and while it was under review, a magnitude 6 earthquake occurred on 25 November 1988 about 35 km south of Chicoutimi, Québec. This is being referred to as the "Saguenay earthquake" by the Geological Survey of Canada. The epicentre was at 48.12°N 71.18°W and the focal depth was about 30 km. The epicentre, if plotted on Figure 1, would be in a region where the selection criteria given in the legend shows no seismicity. Preliminary analysis indicates a thrust mechanism for the earthquake. We do not believe that this earthquake has invalidated any of the general conclusions of this paper; however, it clearly indicates a need for additions to the general framework for moderate to large earthquakes in eastern Canada. A preliminary report on the strong seismic ground motion has been prepared (Munro and North, 1988) and the first paper on the seismological analysis will be published in early 1989.

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