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Earthquakes and geological structures of the St. Lawrence Rift System

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Abstract

The St. Lawrence Rift System (SLRS), which includes the Ottawa–Bonnechère and Saguenay grabens, is located well inside the North American plate. Most historic and the some 350 earthquakes recorded yearly occur in three main seismically active zones, namely Charlevoix (CSZ), Western Quebec (WQSZ), and Lower St. Lawrence (LSLSZ). Outside these areas, most of the Canadian Shield and bordering regions have had a very low level of earthquake activity. In the SLRS, moderate to large earthquakes (moment magnitude, **M**, 5.5 to 7) are known to have occurred since 1663, causing landslides and damage mostly to unreinforced masonry elements of buildings located on ground capable of amplifying ground motions. Most earthquakes in these seismic zones share common characteristics such as mid- to upper-crustal focal depths, no known surface ruptures, and proximity to SLRS faults. Variations also exist such as vast seismically active regions (WQSZ and LSLSZ), the presence of a large water body (CSZ and LSLSZ), and absence of SLRS faults near concentrations of earthquakes (WQSZ). The CSZ is the best studied seismic zone; there, earthquakes occur in the Canadian Shield, mostly in a 30 × 85 km rectangle elongated along the trend of the St. Lawrence River with local variations in focal depth distribution. Faults related to the SLRS and to a meteor impact structure exist, and earthquakes occur along the SLRS faults as well as in between these faults. Overall, the SLRS faults are probably reactivated by the larger earthquakes (**M** ≥ 4.5) of the twentieth century (CSZ in 1925; WQCSZ in 1935 and 1944; Saguenay in 1988) for which we have focal mechanisms. We propose that caution be exercised when linking historical events that have uncertain epicentres with SLRS faults. Similarly, SLRS faults should not necessarily be considered to be the reactivated structures for most small to moderate

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earthquakes (**M** < 4.5). A good example of this is the earthquakes of the WQSZ, which tend to concentrate in a well-defined NW–SE alignment with no obvious geological control, except perhaps a hypothetical hotspot track. Two local factors can lead to the occurrence of SLRS earthquakes: weak faults or enhanced stress levels. We propose that local conditions, concentrated in a few seismic zones, can alter these factors and lead to the occurrence of earthquakes, especially those with **M** < 4.5. At a continent-wide scale, the correlation between the SLRS and earthquakes is appealing. We suggest, however, that pre-existing faults related to the SLRS do not explain all features of the seismicity. Seismicity is concentrated in more active areas, some with conspicuous normal faults and some with suspected weakening mechanisms, such as intense prefracturing (e.g., due to a meteorite impact), the passage over a hotspot, or the presence of intrusions and lateral crustal density variations.

4.1 Introduction

The St. Lawrence Rift System (SLRS), which includes the Ottawa–Bonnechère and Saguenay grabens, is located well inside the North American plate (Figure 4.1). Its earthquake activity was first described in written accounts from the early 1600s and recorded by seismograph stations since the late 1800s. Table 4.1 provides an overview of the most important earthquakes of the SLRS together with a summary of the impact on the natural and manmade environments. Some prehistoric events have also been recognized. This knowledge defines seismic zones with sustained seismic activity (Charlevoix (CSZ), Western Quebec (WQSZ), Lower St. Lawrence (LSLSZ)) and others with a lower ("background") seismic activity. Outside these areas, most of the Canadian Shield and bordering regions have had a very low level of earthquake activity for hundreds of years. Since major population centres are located in the SLRS, including the metropolitan areas of Ottawa, Montreal, and Quebec City (Figure 4.1), its seismic hazard is of great interest for the protection of the public and infrastructures critical to the Canadian economy.

Through time, there have been multiple hypotheses to explain the earthquake activity. In the 1960s, the prevalent view was that the numerous epicentres lined up with Logan's Line, which is the main boundary between the Appalachian thrust belt to the southeast and the Precambrian Shield and overlying Ordovician sedimentary rocks to the northwest (Figure 4.1; Milne and Davenport, 1969). Following the work of Leblanc *et al.* (1973, 1977) it became obvious that earthquakes occurred on faults within the Precambrian Shield. The largest of these faults were normal faults created during the opening of the Iapetus Ocean (Kumarapeli, 1985). From the concentrations of hypocentres that were dipping similarly to the normal faults in the Charlevoix Seismic Zone, this hypothesis has become one of the two seismic zoning models in Eastern Canada (Adams *et al.*, 1995). In the seismicity-rift model, it was assumed that the mild activity outside the recognized active zones was due to

Table 4.1 *Damaging earthquakes of the St. Lawrence Rift System*

Event number	Date	Time (UT)	Region	Lat.	Long.	Mag.	Landslide	Description
1	1663–02–05	22:30	Charlevoix-Kamouraska, Quebec	47.6	−70.1	7	Yes	Epicentre most likely in the Charlevoix Seismic Zone, Quebec. Some damage to masonry in Quebec City, Trois-Rivières and Montréal. Landslides reported in the Charlevoix region and along the St. Lawrence Valley. Numerous aftershocks felt in Quebec City during the following months.
2	1732–09–16	16:00	Near Montreal, Quebec	45.5	−73.6	5.8	No	Probable epicentre near Montréal. Considerable damage in the city of Montréal where hundreds of chimneys were damaged and walls cracked.
3	1791–12–06	20:00	Charlevoix-Kamouraska, Quebec	47.4	−70.5	6	No	Damage to houses and churches in Charlevoix.
4	1860–10–17	11:15	Charlevoix-Kamouraska, Quebec	47.5	−70.1	6	No	Damage in the epicentral region on both shores of the St. Lawrence River.
5	1870–10–20	16:30	Charlevoix-Kamouraska, Quebec	47.4	−70.5	6½	Yes	Considerable damage to houses in Charlevoix along the South Shore of the St. Lawrence River. Damage to chimneys reported in lower town in Quebec City. Possible rock slide along the Saguenay River.

6	1925–03–01	02:19	Charlevoix-Kamouraska, Quebec	47.8	−69.8	6.2	Yes	The earthquake caused damage to unreinforced masonry (chimneys, walls) in the epicentral region on both shores of the St. Lawrence, and in Quebec City (including damage to port facilities), and in the Trois-Rivières region. Possible liquefaction in Charlevoix. Numerous felt aftershocks followed.
7	1935–11–01	06:03	Region of Témiscaming, Quebec	46.78	−79.07	6.1	Yes	The earthquake occurred approximately 10 km east of Témiscaming, Qc. Damaged chimneys were reported.
8	1944–09–05	04:38	Cornwall, Ontario-Massena, New York	44.96	−74.77	5.6	No	Considerable damage to unreinforced masonry in both Cornwall, Ontario and Massena, New York. About 2000 chimneys were damaged in the area.
9	1988–11–25	23:46	Saguenay Region, Quebec	48.12	−71.18	5.9	Yes	Laurentides Fauna Reserve, south of Saguenay (Chicoutimi), Quebec. Damage caused to unreinforced masonry along the Saguenay River, in Charlevoix, and along the St. Lawrence river up to Montreal. Liquefaction of soft soils in the epicentral region. Eleven cases of mass movements reported.
10	1990–10–19	07:01	Mont-Laurier, Quebec	46.47	−75.59	4.5	No	Some minor damage near the epicentre (cracked chimneys, water pipes broken).
11	1997–11–6	02:34	Cap-Rouge, QC	46.80	−71.42	4.9	No	Quebec City region. Minor damage reported in a school of the region.
12	2000–01–01	11:22	Kipawa, QC	46.84	−78.92	4.7	No	Some reports of minor damage in the epicentral region.
13	2010–06–23	17:41	Val-des-Bois, QC	45.88	−75.48	5.0	Yes	Some damage and landslides occurred in the epicentral region.

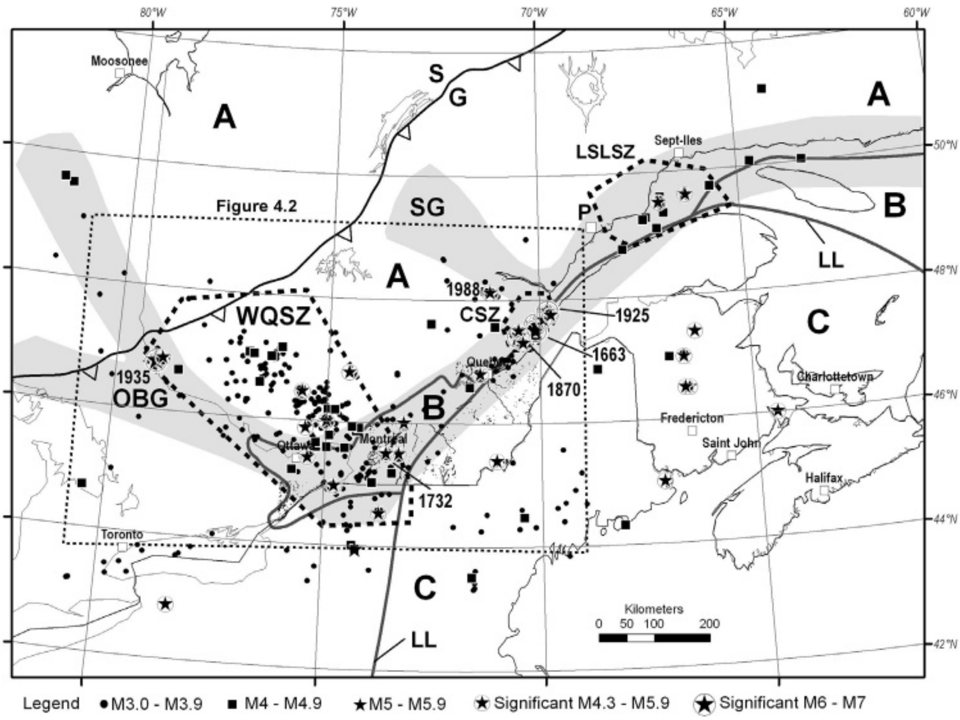

Figure 4.1 Seismicity of the St. Lawrence Rift System together with some geological features. The shaded area is the SLRS as we define it for discussion purposes. Earthquakes of magnitude larger than 3.0 recorded between 1980 and 2012 inclusively together with significant earthquakes of Eastern Canada are shown (Lamontagne *et al.*, 2008a). Seismic zones are: WQSZ, Western Quebec Seismic Zone; CSZ, Charlevoix Seismic Zone; and LSLSZ, Lower St. Lawrence Seismic Zone, as defined by Basham *et al.* (1982). Other acronyms are: OBG, Ottawa–Bonnechère Graben; SG, Saguenay Graben; A, Canadian Shield; B, St. Lawrence Platform; C, Appalachian Orogen; LL, Logan's Line; S, Superior Geological Province; G, Grenville Geological Province; P, Pessamit. The boundaries of Figure 4.2 are shown.

the short time window of observation (i.e., since the 1600s). It was also assumed that these semi-quiescent areas could become active if longer periods, such as tens of thousands of years, were considered. The other source model assumed that the seismic zones active in the past will be the likely sites of future sizeable earthquakes.

In this chapter, we describe the current knowledge of the seismically active areas of the SLRS (Figure 4.1) and examine possible contributing factors to the earthquake activity. We are going to use the expression "St. Lawrence Rift System (SLRS)" not necessarily to express a possible relationship between the earthquakes and the rift faults. We use SLRS as a convenient way to refer to the vast region along the St. Lawrence River, including the Ottawa and Saguenay river valleys.

4.2 Historical earthquakes and their impact

Although well inside the North American plate and remote from plate boundaries, the SLRS has been subject to damaging earthquakes larger than magnitude[1] 5 (Lamontagne *et al.*, 2008a; Cassidy *et al.*, 2010; Table 4.1). For the pre-twentieth-century events, magnitudes and locations are approximate due to the variable reliability of the sources of information: felt reports, detection completeness, and geographic distribution of the population. An example of this uncertainty is the debate over the magnitude of the 1663 Charlevoix earthquake (generally estimated at M 7; see Bent [2009] for a review). More recently, Ebel (2011) rates the same earthquake as moment magnitude (**M**) 7.5 $\pm$ 0.45 based on felt reports and reported falling of masonry. This is not just an academic debate, as it may appear: the preferred magnitude impacts seismic hazard estimates. The locations of some historical events are also uncertain. An example of this is the 1732 magnitude 5.8 Montreal earthquake, located either near Montreal (Leblanc, 1981) or in northern New York State (Figure 4.2; Gouin, 2001). These uncertainties in magnitudes and locations of historical earthquakes lead to discrepancies in the various earthquake catalogues (with impact on seismotectonic interpretation and seismic hazards). An example is the correlation of the 1732 earthquake with the rift faults around Montreal (Adams and Basham, 1989). Fortunately, most instrumentally recorded twentieth-century earthquakes have better defined magnitudes and epicentres (Bent, 2009).

Historically, only a few St. Lawrence Valley earthquakes had a geological impact such as rock falls, landslides, slumps, earth flows in clay deposits, ground cracking, lateral spreading, and liquefaction. No surface rupture has ever been reported. The St. Lawrence Valley (Figure 4.2) has thick sequences of post-glacial marine clay deposits and these have proven to be highly susceptible to earth flow under ground shaking. Of all SLRS earthquakes, the M 7 1663 earthquake had the largest geotechnical impact, causing landslides in the epicentral region (Filion *et al.*, 1991), and along the Saguenay and Saint-Maurice rivers, more than 200 km away (Legget and LaSalle, 1978; Desjardins, 1980). The 1663 event may have also produced a basin collapse in the Saguenay Fjord (Syvitski and Schafer, 1996). Some submarine landslides of the Saguenay and St. Lawrence rivers are also associated with some of these large Charlevoix earthquakes. An example is a landslide near Betsiamites on the Quebec North Shore, possibly caused by the 1663 Charlevoix earthquake, which first occurred on land and continued offshore beneath the St. Lawrence River (Cauchon-Voyer *et al.*, 2007). Another notable earthquake is the 1870 M 6½ CSZ earthquake (Figure 4.2), which had some dramatic consequences, including the collapse of buildings, liquefaction, and landslides in the epicentral region, and rock falls along the Saguenay River (Lamontagne *et al.*, 2007). A landslide, most probably linked with the earthquake, killed four people five days after the mainshock (Lamontagne *et al.*, 2007). More recently, the 1988 **M** 5.9 Saguenay earthquake (Figure 4.2) exemplifies the potential impact of a moderate eastern Canadian

[1] M is defined as the most widely accepted magnitude value (generally the moment magnitude or the felt area magnitude for historical earthquakes) for a given event.

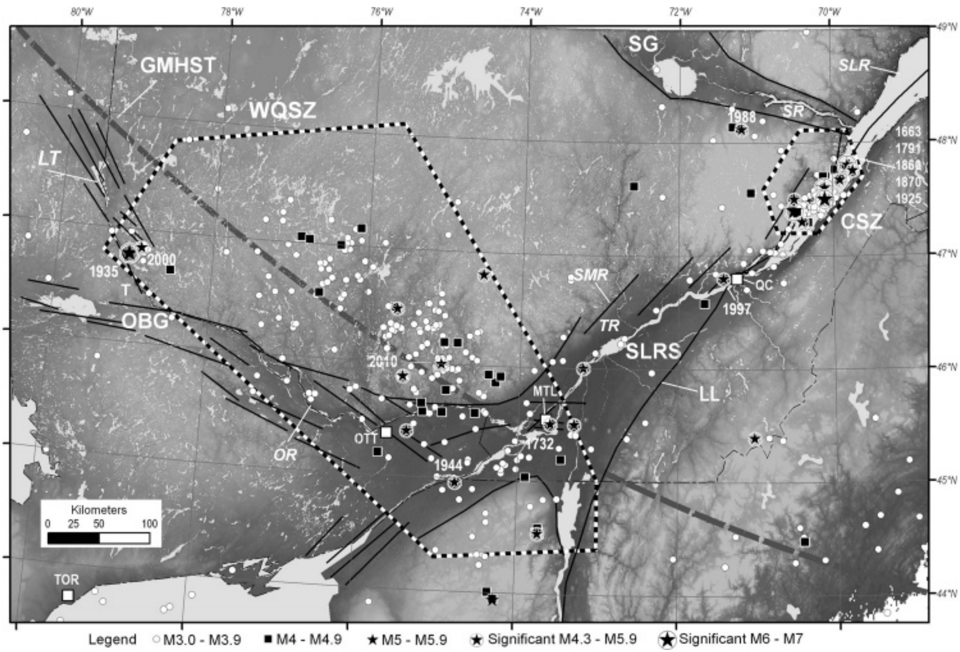

Figure 4.2 Seismicity of the SLRS in southern Quebec and eastern Ontario. Earthquakes of magnitude larger than 3.0 recorded between 1980 and 2012 inclusively together with significant earthquakes of Eastern Canada are shown (listed in Table 4.1; Lamontagne *et al*., 2008a). Seismic zones are: WQSZ, Western Quebec Seismic Zone; CSZ, Charlevoix Seismic Zone, as defined by Basham *et al*. (1982). Other acronyms are: GMHST, Great Meteor HotSpot Track (assumed); OBG, Ottawa–Bonnechère Graben; SG, Saguenay Graben; LL, Logan's Line. Place names are: TOR, Toronto; OTT, Ottawa; MTL, Montréal; QC, Quebec City; TR, Trois-Rivières; T, Temiscaming. Rivers and lakes mentioned in the text are: LT, Lake Temiscaming; OR, Ottawa River; SMR, Saint-Maurice River; SR, Saguenay River; SLR, St. Lawrence River. For colour version, see Plates section.

earthquake. Its effects included liquefaction and rock falls in the epicentral region and landslides as far as 200 km from the epicentre. At 30 km from the epicentre, for example, liquefaction caused extensive damage to local houses (Lefebvre *et al*., 1991; Boivin, 1992). There, evidence of older liquefaction and ground failure was found, possibly caused by older regional earthquakes (Tuttle *et al*., 1990). Other damage was seen as far as 170 km away from the epicentre where failures of railroad embankments were reported (Mitchell *et al*., 1990). Natural slope failures were also seen along the Saint-Maurice River (Lefebvre *et al*., 1992). For building design, this earthquake was the first **M** 6 eastern Canadian earthquake recorded by strong ground motion instruments and showed the high-frequency content of earthquakes there (Munro and North, 1989). In general, infrastructure in the SLRS is especially at risk where marine clays are found, because even moderate earthquakes can trigger earth flows. In 2010, for example, the **M** 5.0 Val-des-Bois earthquake caused two earth flows within a few kilometres of the epicentre (Perret *et al*., 2011).

Geological effects can provide evidence of prehistoric earthquakes. The area around Charlevoix is the site of numerous prehistoric earthquakes that caused submarine slumps and liquefaction (Doig, 1986, 1998; Tuttle *et al*., 1990; Filion *et al*., 1991; Ouellet, 1997; Tuttle and Atkinson, 2010; Locat, 2011). From their data, Tuttle and Atkinson (2010) could not document precise return periods. In the Ottawa River valley, landslide scars are interpreted as traces left by **M** > 6 prehistoric earthquakes (Aylsworth *et al*., 2000; Brooks, 2013). Between Quebec City and Trois-Rivières, where current seismicity is of very low level, no such prehistoric earthquake evidence was found (Tuttle and Atkinson, 2010). In Charlevoix, in the region covered by seismic surveys, no evidence for a coseismic rupture beneath the surface of the St. Lawrence River could be found (Lamontagne, 2002).

Earthquakes of the St. Lawrence valley also damaged man-made structures. In past earthquakes, damage was more common in buildings with unreinforced masonry elements (URM; i.e., masonry without steel reinforcement) that rest on thick clay deposits (Bruneau and Lamontagne, 1994). For the 1988 Saguenay earthquake, 95% of all damage was associated with soft soils (53% with clay, 24% on multi-layer, and 18% on sand). It was also found that damage to buildings built on sandy foundations was restricted to 150 km epicentral distance, whereas for clay foundations damage existed up to 350 km distance (Paultre *et al*., 1993). It is now recognized that urban and industrial developments in the SLRS have taken place on soils capable of amplifying earthquake ground motions. The presence of non-upgraded old buildings coupled with the population growth, particularly on sensitive soils, make earthquakes a significant natural hazard along the SLRS. Fortunately, despite the damage to buildings and earth movements, only two earthquake-caused deaths are known in Canada (Lamontagne, 2008).

In conclusion, the seismic risk in the SLRS is far from being negligible considering the seismic hazard, the low attenuation of Lg waves, the presence of unconsolidated deposits capable of amplifying ground motions and subject to mass movements, and a number of aging buildings built prior to modern building codes.

4.3 Seismic zones of the SLRS

Each year, approximately 350 earthquakes are recorded in the SLRS by the Canadian National Seismograph Network (CNSN). In the SLRS, the CNSN has had a detection completeness of about **M** 1.5 since around 1980. On average, of these 350 earthquakes, about one or two exceed **M** 3.5, ten will exceed **M** 3.0, and about twenty will be reported felt (**M** ≥ 2.0). A decade will, on average, include two events greater than **M** 4.5, which is the threshold of minor damage. On a yearly basis, most earthquakes occur in three main seismic zones, namely the Charlevoix, Western Quebec, and Lower St. Lawrence (Figure 4.1 and Figure 4.2).

4.3.1 Charlevoix

Based on historical and current earthquake rates, the Charlevoix Seismic Zone (CSZ) is the zone with the highest seismic hazard in continental Eastern Canada (Figure 4.3).

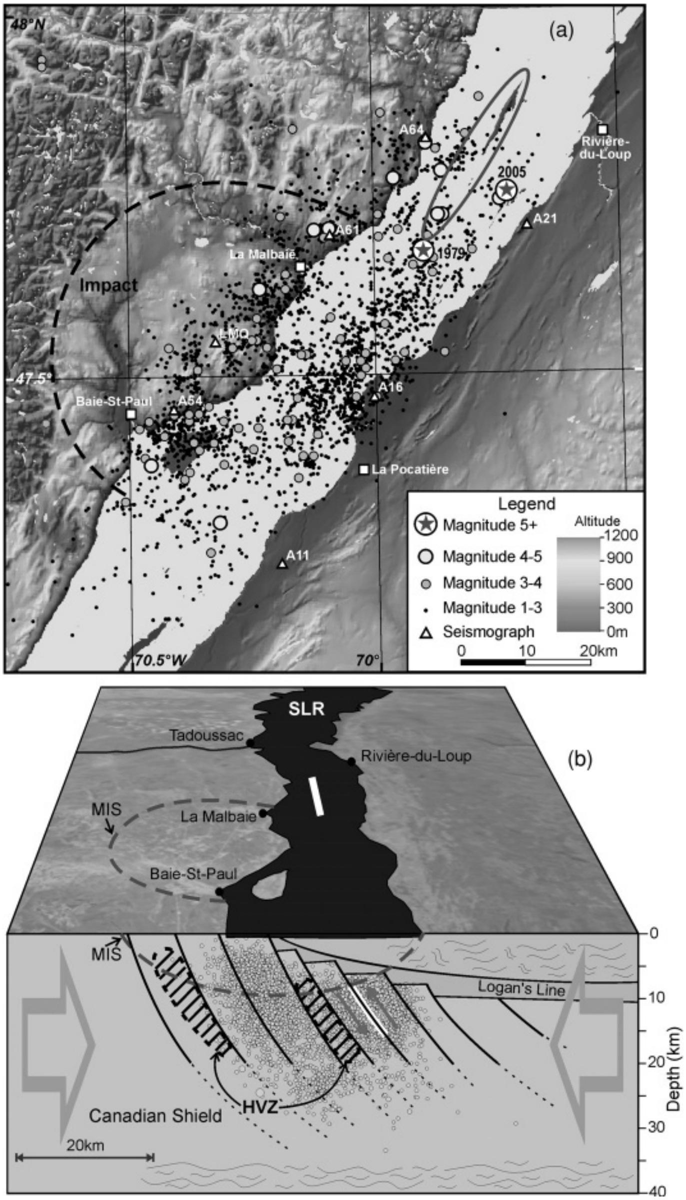

Figure 4.3 (a) Earthquakes of magnitude ≥1.0 recorded between 1978 and 2012 in the CSZ. The estimated position of the 1925 earthquake's epicentre and assumed fault plane orientation is shown as a red ellipse. The semi-circle of the Charlevoix meteorite impact crater is clearly visible on this image, where topography is shown using colours that range from blue for the lowest elevations to red for the highest ones. Since 1978, earthquakes have been monitored by a network of seven seismographs (white triangles). (b) Idealized cross-section perpendicular to the St. Lawrence River in the CSZ. The earthquake activity occurs in response to ridge push stresses (arrows) and is constrained to the Canadian Shield rocks, beneath the St. Lawrence River (SLR), St. Lawrence platform rocks (in grey), Logan's Line, and the Appalachian rocks. The 1925 earthquake ruptured along one of the large SLRS faults at focal depth of about 10 km at the extremity of the CSZ. A small proportion of earthquakes occur within the Charlevoix Meteor Impact Structure (MIS). Earthquakes tend to concentrate outside the zones of high velocities (Vlahovic *et al.*, 2003). For colour version, see Plates section.

Since the arrival of the first Europeans in the early 1600s, the CSZ has been subject to five earthquakes of magnitude 6 or larger: in 1663 (M 7), 1791 (M 6), 1860 (M 6), 1870 (M 6½), and 1925 (magnitude M_S 6.2 ± 0.3) (Lamontagne *et al*., 2008a). The earthquake potential of the area led the Government of Canada to conduct two field surveys that defined its main seismotectonic characteristics (Leblanc *et al*., 1973; Leblanc and Buchbinder, 1977). Earthquakes occur between the surface and 30 km depth, in the Precambrian Shield, which outcrops on the north shore of the St. Lawrence River or is found beneath Logan's Line and the Appalachian rocks (Figure 4.3). Hypocentres cluster along or between the mapped Iapetan faults (also called St. Lawrence paleo-rift faults). The largest earthquake of the twentieth century was the 1925 earthquake, and its focal mechanism has one nodal plane consistent with a reactivation of a southeast-dipping paleo-rift fault (Bent, 1992). The installation of a permanent seismograph network in 1978 (Figure 4.3) has helped to define additional characteristics of the area. The St.-Laurent fault, one of the major rift faults of the CSZ, was formed in the late Precambrian but was also active after the Devonian meteor impact (Rondot, 1979), probably during the early stages of the opening of the Atlantic Ocean in late Triassic–Jurassic times (Lemieux *et al*., 2003). This fault is not particularly active but appears to bound concentrations of hypocentres (Lamontagne, 1999).

Due to its concentration of earthquakes, the CSZ has been the focus of various geophysical studies (Buchbinder *et al*., 1988). Investigations of velocity structure include a seismic reflection–refraction survey (Lyons *et al*., 1980), microearthquake surveys (Leblanc *et al*., 1973; Leblanc and Buchbinder, 1977; Lamontagne and Ranalli, 1997), analysis of teleseismic events (Hearty *et al*., 1977), receiver function analysis (Cassidy, 1995), shear wave splitting and anisotropy studies (Buchbinder, 1989), and focal mechanisms of microearthquakes (Lamontagne, 1998; Bent *et al*., 2003). The concentration of earthquakes led to earthquake prediction studies in the late 1970s and early 1980s (Buchbinder *et al*., 1988). Roughly 80% of Charlevoix earthquakes occur in the depth range 5–15 km in Grenvillian basement rocks, with some as deep as 30 km (Figure 4.3b). Comparing this depth distribution to rheological models of the region, Lamontagne and Ranalli (1996) attribute earthquakes to faulting above the brittle–ductile transition to depths of at least 25 km. The reactivation of pre-existing faults could be due to high pore-fluid pressure at temperatures below the onset of ductility for hydrated feldspar at about 350 °C and/or a low coefficient of friction, possibly related to unhealed zones of intense fracturing. The distribution of spatially clustered earthquakes within the Charlevoix seismic zone indicates that very few earthquakes have occurred on the same fractures with similar focal mechanisms, implying that these fault zones occur in highly fractured rocks, especially those within the boundaries of the Devonian impact structure (Lamontagne and Ranalli, 1997; Figure 4.3). The hypocentre-velocity simultaneous inversion of local P and S waves produced a velocity model that revealed areas of high-velocity bodies at mid-crustal depths (Vlahovic *et al*., 2003; Figure 4.3b). These areas were interpreted to be stronger, more competent crust that separates CSZ earthquakes into two main bands elongated along the St. Lawrence River. Mazzotti and Townend (2010) noted that this seismic zone contains evidence for a local rotation of S_{Hmax} (direction of maximum horizontal stress axis) from the regionally NE–SW-oriented

$S_{Hmax.}$ This reorientation and the maximum plausible stress difference required to cause fault reactivation can be argued to support the presence of high pore-fluid pressures along faults (Lamontagne and Ranalli, 1996). On the other hand, Hurd and Zoback (2012) believe that high pore-fluid pressures are not necessary to explain the majority of preferred focal mechanisms in eastern North America.

There are indications that the region itself may have sub-areas with different rheological properties due to the presence of the meteor crater and its faults. Larger events concentrate at both ends of the seismic zone outside the meteor impact (Figure 4.3; Stevens, 1980; Lamontagne, 1999). Recently, geomechanical modelling showed that the weakening of the rift faults produces a stress increase in the region of the crater bounded by faults, leading to low-magnitude events within the crater and large events outside it (Baird *et al.*, 2009; 2010). If this hypothesis is true, the local seismicity would be caused by local conditions rather than by more regional conditions.

4.3.2 Lower St. Lawrence

Located in the estuary of the St. Lawrence River, the Lower St. Lawrence Seismic Zone (LSLSZ) experiences about 60 events with magnitude ≥2.0 yearly but, unlike the CSZ, only two earthquakes were near magnitude 5.0 in the last 100 years (in 1944 and in 1999; Lamontagne *et al.*, 2003; Figure 4.1). Most earthquakes occur under the St. Lawrence River with hypocentres in the Precambrian mid to upper crust, between 5 and 25 km depth, similar to the CSZ. Adams and Basham (1989) suggested that the seismicity was restricted to the regional normal faults, uniquely found offshore. Focal mechanisms show, however, variable fault plane orientations with most earthquakes clustering along or between the geologically mapped and geophysically inferred Iapetan faults (Lamontagne *et al.*, 2000). Mazzotti and Townend (2010) note that this seismic zone contains evidence for a local rotation of S_{Hmax} for the generally NE–SW-oriented $S_{Hmax.}$

4.3.3 Western Quebec

The Western Quebec Seismic Zone (WQSZ) includes the Ottawa Valley from Montreal to Temiscaming, as well as parts of the Laurentian Mountains (Basham *et al.*, 1982; Figure 4.2). This seismic zone includes historical earthquakes as large as **M** 6.2 and frequent low-level seismic activity. Earthquake epicentres define two sub-zones: a mildly active one along the Ottawa River, including a more active cluster in the Temiscaming region, and a more active one along the Montreal–Maniwaki axis.

The first band of seismicity, which parallels the Ottawa River, includes the epicentres of moderate earthquakes: **M** 6.2 near Temiscaming in 1935; an **M** 5.6 near Cornwall–Massena in 1944 and possibly a magnitude of about 5.8 near Montréal in 1732 (Figure 4.2; Leblanc, 1981). The diffuse group of earthquakes in the first band appear to correlate with a zone of Paleozoic or younger normal faults along the Ottawa River, called the Ottawa–Bonnechère graben (Forsyth, 1981). The Temiscaming area is more active than the rest of the Ottawa

River valley and was the site of the 1935 **M** 6.2 Temiscaming earthquake. In the year 2000, an **M** 4.7 earthquake reactivated one of the local faults (Bent *et al.*, 2002), adding support to the correlation between the earthquake epicentres and the normal faults of the Ottawa–Bonnechere graben (Adams and Vonk, 2009). These earthquakes appear to reactivate faults smaller than the conspicuous ones near Lake Temiscaming (Bent *et al.*, 2002).

The majority of WQSZ earthquakes, mostly smaller than **M** 4.5, occur in an elongated NW–SE zone within the Grenville Geological Province with most focal depths varying between 7 and 25 km (Lamontagne *et al.*, 1994; Ma and Atkinson, 2006). Although northwest-trending structural features are known, correlating these with the epicentral trend is uncertain because of the thrust sheets that make up the uppermost crust in this area of the Grenville Province. The mid-crustal hypocentral depths of many earthquakes, the east–west trend of the fault planes of some earthquakes, and variations in regional focal mechanisms all suggest reactivation of deep structural features that may not have a surface expression. It is possible that at the eastern end of this active band an anorthosite body may act as a stress concentrator (Lamontagne *et al.*, 1994).

4.3.4 Background seismicity

Outside the seismic zones described above, other areas along the SLRS are much less active (Figures 4.1 and 4.2). On a yearly basis, very few earthquakes are recorded in these areas and very few significant earthquakes are known (one notable exception being the 1988 **M** 5.9 Saguenay earthquake; Table 4.1). This earthquake occurred at lower-crustal depth (29 km) in the Canadian Shield, outside the seismic zones defined by the recurring activity (North *et al.*, 1989). Possibly due to its focal depth, very few aftershocks have been recorded and only four M $\geq$ 3 earthquakes were recorded there between 1988 and 2013. Despite the differences in orientation between the two nodal planes of the focal mechanism and the graben fault at the surface, the proximity of the earthquake epicentre to faults of the Saguenay graben suggested a relationship between the two (North *et al.*, 1989). Other authors warned of the difficulty in linking a lower-crustal depth earthquake with the numerous faults of the Precambrian basement that have unknown fault extensions at depth (Du Berger *et al.*, 1991). This earthquake suggests that moderate to large earthquakes can occur in areas outside the zones defined by recurring seismic activity and that many faults of the whole Precambrian Shield could be near failure. This suggestion is supported by the numerous cases of reservoir-triggered seismicity in areas almost devoid of naturally occurring earthquakes (Lamontagne *et al.*, 2008b). Another noteworthy event is the moderate 1997 **M** 4.9 Cap-Rouge earthquake, which occurred at a depth of 22 km with a focal mechanism consistent with the reactivation of a mapped SLRS fault (Table 4.1; Nadeau *et al.*, 1998).

4.4 The St. Lawrence Rift System

Kumarapeli and Saull (1966) proposed that the series of normal faults along the St. Lawrence valley indicated rifting similar to the East African rift system. Although it is now recognized that the St. Lawrence Rift System (SLRS) is not an active rift, this hypothesis provided a model that explained many of the geological characteristics of the area (including faults created in an extensional regime and carbonatite intrusions). Since that pioneering work, the normal faults of the St. Lawrence paleorift system have been described and mapped at the surface as well as in the subsurface (Du Berger *et al.*, 1991; Castonguay *et al.*, 2010; Tremblay and Roden-Tice, 2011). The SLRS is a half-graben created during the late Precambrian opening of the proto-Atlantic (Iapetus) Ocean (Kumarapeli, 1985). Many of the faults trend NE–SW and mark the boundary between the Grenville Province of the Canadian Shield to the northwest and the St. Lawrence Lowlands to the southeast. The grabens of Ottawa–Bonnechère and Saguenay River intersect the SLRS and are both interpreted as Iapetan failed arms (aulacogens; Kumarapeli, 1985; Tremblay and Roden-Tice, 2011). Faults related to the Ottawa–Bonnechère and Saguenay grabens trend mostly WNW–ESE.

The SLRS faults have been studied in the Montreal region (Rocher *et al.*, 2003) as well as in Charlevoix, Quebec City, and in the St. Lawrence estuary (Figure 4.2). SLRS faults consist of cohesive cataclastic rocks, with some major fault zones being marked by 10–20-metre-thick fault breccias, ultracataclasite, and foliated fault gouge (Tremblay and Roden-Tice, 2011). The rift system is a crustal-scale fault zone where reactivation occurred along Late Precambrian to early Paleozoic faults attributed to Iapetus rifting (Sanford, 1993). Studies of the age of formation and reactivation of these faults revealed various periods of reactivation (Rocher *et al.*, 2003; Faure *et al.*, 2006). Following the formation of the Appalachian orogen (started in the Ordovician) when these faults were under a compressive stress environment, rifting of the Atlantic Ocean–Labrador Sea during Mesozoic times possibly reactivated them (e.g., Kumarapeli and Saull, 1966; Carignan *et al.*, 1997). Apatite fission-track age discontinuities between the two sides of these faults are interpreted as the result of normal faulting at *c*. 200 Ma (Jurassic) followed by tectonic inversion about 150 Ma ago (Tremblay and Roden-Tice, 2011). The latter study provides support for Atlantic-related, extensional and compressive deformation within the interior of the Canadian Shield, more than 500 km west of the axis of the Mesozoic rift basins. The faults were most recently active during the Late-Triassic–Jurassic period, which corresponds to the creation of the current Atlantic Ocean with the separation of North America and Africa. The emplacement of igneous rocks of the Monteregian Hills occurred approximately at the same time, i.e., in early Cretaceous times (McHone and Butler, 1984; Foland *et al.,* 1986; Pe-Piper and Jansa, 1987). In the Montreal region, the E–W and SE–NW extensional tectonics are documented by brittle normal faults.

Beneath the Paleozoic and Appalachian cover, seismic reflection profiles in the St. Lawrence Lowlands have shown the upper crustal morphology of the faulted blocks (Thériault *et al.*, 2005; Castonguay *et al.*, 2010). SLRS faults exist at depth beneath the Appalachians, and their positions are revealed by their magnetic and gravity signatures (Charlevoix: Lamontagne, 1999; Lower St. Lawrence: Lamontagne *et al.*, 2003; St. Lawrence Lowlands: Lamontagne *et al.*, 2012). Although we know the positions of some of these rift faults beneath the sedimentary cover, our knowledge is far from complete. Based on geophysical information, Wheeler (1995, 1996) defined the northwestern

Table 4.2 *Main characteristics and differences of the earthquake zones of the SLRS (see Table 4.1 for a description of the earthquakes)*

Common characteristics of SLRS active areas	Differences/anomalies along the SLRS
99% of earthquakes occur at less than 25 km depth	Some larger events occur near or deeper than 25 km (1988 **M** 5.9 Saguenay, 29 km; 1997 Cap-Rouge **M** 4.7, 22 km; 2010 Val-des-Bois **M** 5.0, 22 km)
Earthquakes occur in areas where paleo-rift faults are present (CSZ, WQSZ, LSLSZ)	Part of the earthquake activity is also present where no SLRS faults are present (WQSZ)
No surface rupture known for any of the large events	
Proximity to a large water body (CSZ, LSLSZ) with tides	No large water bodies (WQSZ)
Earthquakes occur over a vast region (WQSZ, LSLSZ)	Concentrated activity (CSZ)
Protracted aftershock sequence for moderate earthquakes	Very few aftershocks (1988 Saguenay at 29 km focal depth)
Presence of a meteor impact structure (CSZ)	

and southeastern boundaries of the intact Iapetan margin and its potentially seismogenic faults for seismic zoning purposes.

A seismotectonic model of the earthquakes of the St. Lawrence Valley must consider the similarities and differences of the seismic and weakly seismic zones (Figure 4.2). Table 4.2 presents an overview of the similarities and differences between the various seismically active regions of the SLRS. In the 1970s, focal mechanisms showed that microearthquakes represented the reverse-faulting reactivation of pre-existing faults in the Precambrian basement (Leblanc *et al.*, 1973; Leblanc and Buchbinder, 1977). Later studies with larger magnitude earthquakes confirmed this (see Hurd and Zoback [2012] for a review). The hypothesis that connects the SLRS faults and the current seismicity was mainly derived from the concentration of CSZ earthquake hypocentres along southeast-dipping trends, similar to SLRS faults (Anglin and Buchbinder, 1981; Anglin, 1984). By extension outside the CSZ, it was hypothesized that all earthquakes along the St. Lawrence valley represented reactivation of similar faults. This SLRS connection was also made at the time when all worldwide M ≥ 7.0 continental earthquakes were shown to correlate with rifted cratonic areas (also called “extended crust”; Johnston and Kanter, 1990). This rift-faults–seismicity connection was also used by Adams and Basham (1989) in discussing the seismicity of the SLRS.

Outside the CSZ, the best correlation between the normal faults of the SLRS and seismicity can be found in the Temiscaming region, where the 1935 M 6.2 earthquake occurred (Figure 4.2). Near the epicentre of the earthquake, many Ottawa–Bonnechère graben faults have been mapped and many have conspicuous topographical signatures visible in remote-sensing imagery (Bent *et al.*, 2002). The focal mechanism of the 1935 earthquake (the largest event of the seismic zone) indicates thrust faulting on a moderately dipping northwest-striking plane, an orientation similar to the graben faults of the area (Bent, 1996). Similarly, in the Lower St. Lawrence Seismic Zone, SLRS faults are present and may be reactivated by the current seismic activity.

The rift–seismicity hypothesis has seismic hazard implications. The seismic provisions in the 2005 National Building Code of Canada use the larger of the ground motions derived from two seismic source zone models. The first model assumes that the historical earthquake clusters denote areas that will continue to be active. The second model assumes a common geological framework for the seismicity clusters (i.e., passive Paleozoic rift faults) and regroups them into large source zones (Adams *et al.*, 1995). In areas with historical damaging earthquakes, the hazard derived from the historical model dominates, whereas in areas without significant activity but where rift faults are present, the hazard is dominated by the geologically derived model (Adams and Halchuk, 2003).

4.5 The rift hypothesis and the SLRS: discussion and conclusions

Since the mid 1980s, the rift model has served as the main hypothesis that relates seismicity and faults along the SLRS. Although the rift model explains many characteristics of the SLRS seismicity, some questions remain. At the global level, the rift model was based on a correlation with "extended crust" that was correct for 100% of earthquakes of magnitude larger than 7.0; 60% between M 6 and 7; and 46% for earthquakes smaller than M 6 (Johnston and Kanter, 1990). Recently, however, Schulte and Mooney (2005) concluded that on a global scale the correlation of seismicity within stable continental regions (SCR) and ancient rifts has been overestimated in the past. They note that several apparently non-rifted crust areas have experienced multiple large ($\mathbf{M} \geq 6.0$) events. The rift model defined from a global perspective was extrapolated to the SLRS where only the 1663 Charlevoix shock is thought to have exceeded magnitude 7.0.[2] We note that the historical earthquake of 1663 has an uncertain epicentre location based on limited intensity information and is consequently only assumed to be along a rift fault. The rift hypothesis was extrapolated to smaller earthquakes in the magnitude 5.5 to 6.5 range, with most having occurred prior to instrumental recording, with uncertain magnitudes, epicentres, and unknown focal mechanisms. Based on these limitations, we raise questions about the applicability of the global rift model to SLRS earthquakes, especially for $\mathbf{M} < 4.5$ earthquakes.

Globally, we note that many recent earthquakes do not fit the assumed local seismotectonics model. Therefore, the correlation with any earthquake that occurs within the loose geographic boundaries of the SLRS with the faults of that system is not necessarily correct. In the case of the 1988 **M** 5.9 Saguenay earthquake, for example, despite the focal mechanism that did not match the orientation of the most conspicuous graben faults that were

[2] Although the 1663 event was not part of the database of Johnston and Kanter (1990), which only considered the last 200 years.

dipping away from the epicentre (Du Berger *et al*., 1991), this earthquake is now referred to a SLRS earthquake. To exemplify the difficulty in correlating faults and earthquakes, the 1982 **M** 5.8 Miramichi and the 1989 **M** 6.3 Ungava earthquakes, both shallow focus earthquakes, do not correlate with any fault of regional extent (Wetmiller *et al*., 1984; Lamontagne and Graham, 1993). These events are reminders that faults with dimensions necessary to generate an **M** 6 earthquake (about 10 km) exist almost everywhere in the Canadian Shield, even where rift faults are not present. Consequently, we conclude that smaller magnitude earthquakes (**M** < 4.5), which represent the vast majority of earthquakes in Eastern Canada, do not necessarily represent reactivated SLRS faults.

We believe that some global studies have partaken in the confusion because they did not look at the SLRS at the local level. For example, Schulte and Mooney (2005) used a continent-wide view to suggest a spatial correlation of the SLRS seismicity with the Appalachian front to the east (which is known to be inactive). They also correlate SLRS with the epicentre of the 1732 Montreal M 5.8 (which they rate as **M** 6.3) earthquake, which Gouin (2001) locates in Northern New York State from the same intensity data. We argue that the current earthquake database does not have a sufficiently large number of earthquakes with reliable epicentres to strongly support a correlation between earthquakes and the entire SLRS. Another example of a continental-scale view of seismicity is the suggested correlation between earthquakes and low-angle thrust faults ("sutures") of the Grenville Province (Van Lanen and Mooney, 2006). We note that numerous aseismic areas of the Grenville Province have these low-angle thrust faults. Recently, Mooney *et al*. (2012) suggested a correlation between crustal seismicity and younger lithosphere surrounding the ancient cratons, with aseismic zones corresponding to areas with thick (and cold) lithosphere.

One must note that the strongest support of a SLRS–seismicity connection is based on the locations of earthquake epicentres, and hypocentres in the CSZ, where SLRS faults are present. We believe that the regional-scale model, suggesting a seismic hazard based on the sole presence of SLRS faults, may not hold if one considers the local scale of the various seismic zones. In the CSZ, for example, the hypocentre–rift-fault connection suggested by Anglin (1984) is based on about 6 years of earthquake recording by a local network. Today, after some 35 years of recording, a more complex picture emerges, with numerous sub-zones, each one with its special focal depth distribution, level of activity, and focal mechanism complexity. There is some control by the SLRS faults, but they appear to bound seismically active blocks rather than being active themselves (Lamontagne, 1999).

Another problem with the rift fault hypothesis is that active regions are separated by weakly active areas (Figure 4.1). Tuttle and Atkinson (2010) did not find any evidence of Holocene earthquake activity in the Quebec City to Trois-Rivières segment, despite the presence of SLRS faults. These authors suggest that the seismic activity at Charlevoix may be localized. This suggests that local factors lead to seismicity in recognized seismic zones, not uniquely the presence of SLRS faults. In the WQSZ as well, the seismicity–rift-fault correlation does not explain that most earthquakes locate well to the north of the graben faults in a NW–SE alignment. In general, the Ottawa–Bonnechère graben is weakly seismic except near the 1935 Temiscaming epicentre (Figure 4.2), where only a small portion of the regional graben faults are active (Bent *et al*., 2002).

Our evidence in the SLRS suggests that the relation between earthquakes and rift faults breaks down for the lower magnitude earthquakes (probably smaller than **M** 6 and certainly less than **M** 4.5), where local stresses and/or fault weaknesses contribute to earthquake occurrences. This approach was also proposed on a global scale by Schulte and Mooney (2005) and Mooney *et al*. (2012). In the SLSR, and in the CSZ in particular, a study of focal mechanisms for magnitude 2 to 5 has shown that numerous fault orientations are reactivated in the smaller earthquakes, including mixed thrust–strike-slip events. When studied in detail, earthquakes in sub-areas of the CSZ do not occur on rift faults, or on some impact structure faults (Lamontagne and Ranalli, 1997). Baird *et al*. (2009, 2010) have emphasized the importance of local variations in stress level within the CSZ and of strength differences between the crater and the rift faults outside the crater. They showed that much of the background seismicity pattern can be explained by the intersection of weak faults of the St. Lawrence rift with the damage zone created by the Charlevoix impact. In terms of strain measurements, the first interpretations of crustal strain along the SLRS were supportive of a homogeneous stress system, but more recent results have shown that local stresses are anomalous in active areas such as the CSZ (Mazzotti and Townend, 2010). For smaller Eastern Canadian earthquakes, the direction of the maximum compressive stress varies, which suggests local perturbations in the stress field in contrast to studies that favour a regionally homogeneous eastern North American stress field (Zoback and Zoback, 1991).

Another possibility for the creation or reactivation of local faults is the role of sub-crustal processes, such as what was proposed for the NW–SE band of earthquakes of the WQSZ. Since no surface geological feature is seen along this trend, it was proposed that the seismicity could be due to an extension of the New England Seamount Chain track or the passage of this region over the Great Meteor hotspot between 140 and 120 million years ago (Figure 4.2; Sykes, 1978; Crough, 1981; Adams and Basham, 1989). Ma and Eaton (2007) have suggested that the passage over a hotspot, possibly imaged by lithospheric velocity anomalies at 200 km depth, may explain the enhanced level of seismicity either by thermal rejuvenation of pre-existing faults or by stress concentration caused by mid-crustal strength contrast between mafic and felsic rock. If this hypothesis is valid for the WQSZ, the hotspot would have generated crustal fractures (Adams and Basham, 1989) through the generation of thermo-elastic stress (e.g., Marret and Emerman, 1992). Two problems remain to be resolved before the hotspot–seismicity link is accepted in this area: the hotspot track has no geological expression (faulting, metamorphic grade changes, intrusives) in the WQSZ (Sleep, 1990) and there is very little seismic activity along most of the length of the hotspot track, including along the Monteregian Hills region.

Numerous hypotheses have been advanced to account for the local concentration of WQSZ earthquakes. For example, it was proposed that the NW–SE band of earthquakes in WQSZ represented crustal zones of weakness delineated by the drainage pattern (Goodacre *et al*., 1993). A possible correlation was also suggested with positive aeromagnetic anomalies and the Helikian (Mid-Proterozoic) paragneisses. The weakness of upper-crustal geological formations could lead to enhanced stresses at mid-crustal levels. Another hypothesis is that areas that are currently active represent aftershock zones of large historic or prehistoric earthquakes (Ebel *et al*., 2000).

If the SLRS connection only partially explains the St. Lawrence seismicity, what factors can render portions of the SLRS faults inherently weak and/or enhance the local stress levels?

Assuming that the level of tectonic stresses is everywhere similar in Eastern Canada, there are a few factors that can weaken the faults at depth and favour reactivation. In the brittle regime, the reactivation of a fault is controlled by the stress orientation, the friction coefficient on the fault, and the pore-fluid pressure acting on the fault interface. The reverse-faulting regime that prevails down to the mid-crust ($\sim$30 km) in Eastern Canada implies stress differences that are difficult to conceive. At 25 km, for example, a hydrostatic pore-fluid pressure ($\lambda = 0.4$) and a coefficient of friction of 0.75 imply a critical stress difference for sliding of about 1,200 MPa in an ideally oriented reactivated thrust fault. These high stress differences are about one order of magnitude larger than the upper limit of 100–200 MPa usually assumed for crustal stresses. This suggests that high pore-fluid pressure and/or low coefficient of friction must exist along reactivated faults to give rise to the SLRS mid-crustal seismicity (Lamontagne and Ranalli, 1996). Another factor is the orientation of the faults with respect to the maximum compressive stress (which is assumed to be sub-horizontal in Eastern Canada). Hence, only faults that strike nearly perpendicular to the maximum stress axis have a higher probability of being reactivated as reverse faults.

If, on the other hand, it is assumed that all faults are equally weak to the first order, then factors must be sought that can locally increase the stresses. Stress can concentrate in areas with lateral mass anomalies within the lithosphere (Goodacre and Hasegawa, 1980; Assameur and Mareschal, 1995). In the latter study, the fact that two out of three regions with the highest induced stress differences remain aseismic suggests that mass anomalies can favour but not control earthquake occurrences. Other models that have been suggested to explain the occurrence of seismicity within continental interiors include localized stress concentration around weak intrusions (Campbell, 1978), intersecting faults (Talwani, 1988; Gangopadhyay and Talwani, 2007) and ductile shear zones in the lower crust (Zoback, 1983). Some models relate seismicity to elevated temperatures at depth (Liu and Zoback, 1997). In these models, plate-driving forces are largely supported by the (seismogenic) upper crust; the lower crust is weakened as a result of higher temperatures and the total strength of the lithosphere is reduced. Other models refer to regional stress fields perturbed by forces associated with lithospheric flexure after deglaciation (Stein *et al*., 1979; Quinlan, 1984; Grollimund and Zoback, 2001). Based on strain determination for Eastern Canada, deglaciation is also used by James and Bent (1994) and Mazzotti and Townend (2010) to explain perturbed stress environments. A concentration of post-glacial rebound stresses can occur in local zones of weakness, perhaps containing low-friction faults (Mazzotti and Townend, 2010). This model suggests local sources of weakness to reconcile the apparent reorientation of maximum compressive stresses in the CSZ and in the LSLSZ.

The past few decades have brought to light numerous geological and geophysical characteristics of the SLRS and their links with seismicity. At a continent-wide scale, the correlation between the SLRS and earthquakes is appealing. When a more local scale is looked at or when smaller magnitude earthquakes are considered, questions arise and the picture becomes more complex. We conclude that the sole presence of pre-existing normal faults related to the SLRS does not explain all the features of the seismicity of Eastern Canada. Seismicity is concentrated in more active areas, some with conspicuous normal faults and some with suspected weakening mechanisms such as intense pre-fracturing (e.g., due to a meteorite impact), passage over a hotspot, or the presence of intrusive and lateral crustal density variations. In most cases, the superposition of the tectonic stress field and the relatively modest post-glacial rebound stresses is likely to play a role. Irrespective of the presence of rift faults, earthquakes are caused by dynamic instabilities along pre-existing faults caused by the inherent weakness of the faults, by the locally enhanced stress differences, or a combination of the two. A general unifying explanation is likely to be useful only at the large scale, and a detailed understanding of seismicity requires further work concentrating on local conditions.

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