Rheological and Geological Constraints on the Earthquake Distribution in the Charlevoix Seismic Zone, Québec, Canada

by

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in partial fulfilment of the requirements for the
degree of Doctor of Philosophy

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Abstract

Located on the St. Lawrence River some 100 km downstream from Quebec City, the Charlevoix Seismic Zone (CSZ) is the most active seismic zone of eastern Canada with 5 historical earthquakes in the magnitude 6 to 7 range and continuous microearthquake activity. Between October 1977 and December 1997, some 1500 earthquakes with magnitudes between -1.0 and 5.0 were recorded by a local seismograph network. Epicentres define a 30 by 85 km ellipse with the long axis parallel to the river, with 99% of hypocentres shallower than 25 km depth ($D_{99\%}$). Earthquakes are not distributed uniformly across the seismic zone, but concentrate in groups separated by less active areas.

On the basis of the heat conduction equation and Grenvillian surface heat flow, the temperature at $D_{99\%}$ has been estimated to be between 215 and 355°C. These temperatures and the inferred quartz-depleted mid- and lower-crustal compositions imply a brittle-ductile transition deeper than 25 km. The quartz-depleted mid- to lower crust is supported by the high P-velocities revealed by seismic refraction surveys. This depth may represent the passage from velocity weakening to velocity strengthening if the onset of flow in hydrated feldspars occurs at temperatures in the 300-350°C range. With an assumed maximum crustal stress difference of about 200 MPa, fault reactivation at mid-crustal depth can occur with a low friction coefficient and/or with a high pore fluid pressure.

Remote sensing, magnetic, gravimetric and seismic reflection data are used to define the positions of geological faults. Most regional normal faults correspond to lineaments or geophysical trends generally parallel to the St. Lawrence River. Most regional faults bound active volumes while one correlates with earthquake hypocentres, including some magnitude > 4 events. Inside the impact crater, earthquakes are generally of smaller magnitude, possibly due to the high fracturing associated with the Charlevoix impact structure.

The orientations of the local stress and of the reactivated faults vary across the CSZ. The comparison of some 20 focal mechanisms (mostly reverse to reverse-oblique faulting) with earthquake groups and multiplets (found to represent less than 15% of the events between 1988 and 1997) suggest reactivated faults with various orientations, not necessarily parallel to the regional paleo-rift faults. Most microearthquakes appear to occur in highly fractured volumes, partly related to the impact structure. No surface rupture is found on the seismic reflection lines acquired on the St. Lawrence River.

Using local earthquake data, a pseudo-2D layered velocity model is proposed. The main upper crustal velocity perturbations are imaged, i.e. the 6 km thick Appalachian sequence (5.6 to 5.8 km/s P-velocity) and the Precambrian basement (6.2 to 6.6 km/s P-velocity). From V_p/V_s ratio information, the CSZ velocity structure differs from that of the neighbouring Laurentides Park region, where upper crustal basic rocks are found.

Even with the addition of all local stress contributors, the CSZ is not subject to substantially higher stress difference levels than the rest of Eastern Canada. Consequently, the anomalous CSZ earthquake activity must be due to inherent crustal weakness and/or high pore fluid pressure. Although the presence of a gouge may weaken some faults, the existence of pervasive high pore-fluid pressure coupled with a high degree of fracturing is the favoured explanation for the weak crust that gives rise to earthquake activity. A qualitative model is proposed where some rift faults act as conduits to crustal fluids under pressure, triggering earthquakes on these faults and in neighbouring fractured volumes.

Résumé

Située le long du fleuve Saint-Laurent à quelque 100 km en aval de Québec, la Zone sismique de Charlevoix (ZSC) est la plus active de l'est du Canada avec ses 5 séismes historiques de magnitude entre 6 et 7 et son activité micro-séismique continue. Entre octobre 1977 et décembre 1997, quelque 1500 tremblements de terre de magnitude variant entre -1.0 et 5.0 furent enregistrés par un réseau sismographique local. Les épicentres définissent une ellipse de 30 km par 85 km allongée parallèlement au fleuve, avec 99% des foyers moins profonds que 25 km ($D_{99\%}$). Les séismes ne sont pas distribués uniformément sur toute la superficie de la zone sismique, mais se concentrent en groupes séparés par des régions moins actives.

D'après l'équation de conduction de chaleur et d'un flux de chaleur à la surface typique du Grenville, la température à $D_{99\%}$ fut estimée entre 215 and 355°C. Ces températures et les compositions déficientes en quartz présumées de la croûte moyenne et profonde impliquent un passage fragile-ductile plus profond que 25 km. La composition de la croûte moyenne à inférieure est appuyée par les hautes vitesses sismiques révélées par les levés de sismique réfraction. Cette profondeur peut correspondre au passage de l'affaiblissement par vitesse (velocity weakening) au renforcement par vitesse (velocity strengthening) si le début du fluage des feldspaths hydratés se produit autour de 300-350°C. En présumant une différence maximale de contrainte d'environ 200 MPa pour la

croûte, la réactivation des failles dans la croûte moyenne peut se produire avec un bas coefficient de friction et/ou avec des pressions de fluide élevées.

Les données de télédétection, magnétiques, gravimétriques et de sismique-réflexion sont utilisées pour définir les positions des failles géologiques. La plupart des failles régionales correspondent à des linéaments ou à des orientations géophysiques généralement parallèles au fleuve Saint-Laurent. La plupart des failles régionales bordent des volumes actifs alors qu'une d'entre elles corrèle avec des foyers de séismes, incluant quelques événements de magnitude > 4 . À l'intérieur du cratère météoritique, les séismes sont généralement plus petits, possiblement reliés au haut niveau de fracturation associé à la structure météoritique de Charlevoix.

Les orientations des contraintes locales et des failles réactivées varient dans la ZSC. La comparaison de quelque 20 mécanismes au foyer (la plupart de faille inverse ou inverse-oblique) avec les groupes de séismes et multiplets (moins de 15% des séismes entre 1988 et 1997) suggère que des failles réactivées à orientations variées, pas toujours parallèles aux failles de paléo-rift régionales. La plupart des micro-séismes semblent se produire dans des volumes très fracturés, liés en partie à la structure météoritique. Aucune rupture en surface n'est évidente sur les profils de sismique-réflexion sur le fleuve Saint-Laurent.

À partir des données sismiques locales, un modèle multi-couches pseudo-2D est

proposé. Des perturbations de vitesse dans la croûte supérieure sont révélées, c'est-à-dire une séquence Appalachienne de 6 km d'épaisseur (vitesse de P entre 6.2 et 6.6 km/s) et le socle Précambrien (vitesse de P entre 6.2 et 6.6 km/s). À partir des rapports V_p/V_s , la structure de vitesse de la ZSC diffère de celle de la région voisine du Parc des Laurentides, où des roches basiques existent près de la surface.

Même en ajoutant les sources de contraintes locales, la ZSC n'est pas sujette à un niveau de contraintes différentielles tellement plus élevé que le reste de l'est du Canada. Par conséquent, l'activité sismique anormale de la ZSC doit être reliée à une faiblesse crustale particulière et/ou à des pressions de fluide élevées. Quoique la présence d'argile de faille pourrait affaiblir quelques failles, l'effet combiné des pressions de fluide élevées et d'un haut niveau de fracturation est l'explication avancée pour expliquer l'activité sismique dans la croûte fragile. Un modèle qualitatif est proposé dans lequel quelques failles de rift agissent comme conduits à des fluides crustaux sous pression, causant des séismes sur ces failles et dans les volumes fracturés adjacents.

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Original Contribution

This thesis contains only the results of the research conducted by the candidate, under the guidance of his supervisor. Most of the material presented in Chapter 2 has been published in Lamontagne and Ranalli (1996); similarly, most of the description of the earthquake activity near La Malbaie (Section 4.6.1.1) has been published in Lamontagne and Ranalli (1997). In both papers, the candidate is the senior author, and the supervisor's (second author) contribution has consisted of advice and cooperation in the writing of the paper. Due to an error in the computation of the stress difference, Figure 8B of Lamontagne and Ranalli (1996) should be replaced by Figure 2.9B.

A few sections of Chapter 3 result from collaborative work with scientists of the Geological Survey of Canada. The gravity and magnetic maps, and the Euler deconvolution solutions (Section 3.2.2.1), were produced by Dr. Pierre Keating. Reprocessing of the SOQUIP seismic line 13 was done by Gervais Perron (Section 3.2.2.3). Finally, the chromo-stereoscopic image of the Charlevoix region was produced by Thierry Toutin of the Canada Centre for Remote Sensing.

In Chapter 4, the focal mechanisms of Section 4.3 are described in Lamontagne (1998) and the 1996 Summer field work in Lamontagne et al. (1997).

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Chapter 1

The Charlevoix Seismic Zone: overview of the earthquake activity and the geological environment

1.1 Introduction

Located some 100 km downstream from Quebec City, the Charlevoix Seismic Zone (CSZ)¹ is the most seismically active region of eastern Canada (Figure 1.1). Historically, the zone has been subject to five earthquakes of magnitude 6 or larger: in 1663 (magnitude ~ 7.0); 1791 (magnitude ~ 6.0); 1860 (magnitude ~ 6.0); 1870 (magnitude ~ 6.5); and 1925 (M^2 6.2) (magnitudes as listed in Anglin et al., 1990; Bent, 1992). The earthquake potential of the CSZ led the Earth Physics Branch (EPB; now part of the Geological Survey of Canada: GSC) to conduct two field surveys, in 1970 and 1974 respectively (Leblanc et al., 1973; Leblanc and Buchbinder, 1977). The 63 hypocentres located during these two surveys clearly defined the boundaries of the CSZ, i.e. an active zone about 30 by 85 km, elongated along the St. Lawrence River, enclosing the towns of Baie-St-Paul, La Malbaie and La Pocatière (Figure 1.2). In 1977, the EPB deployed a seven-station local seismograph network centred on the active zone defined

¹ The name Charlevoix associated with the seismic zone is somewhat misleading: most of the earthquakes occur under the St. Lawrence River, between Charlevoix County on the north shore and Kamouraska County on the south shore. To consider this aspect, the CSZ is sometimes referred to as the Charlevoix-Kamouraska Seismic Zone. In this dissertation, we will use the traditional name CSZ of Basham et al. (1982).

² M is the moment magnitude, formerly listed as M_w in some publications.

by the field surveys. Between 1978 and 1997 inclusively, the network detected nearly 2200 local earthquakes, of which 54 exceeded Nuttli Magnitude (m_N) 3.0 (Figure 1.3). The current CSZ network detects more than 200 earthquakes per year. Based on historical and current earthquake rates, the CSZ is the highest seismic hazard zone in continental eastern Canada.

The hypocentres located ³ over the years have provided an insight into the CSZ seismotectonics. Most earthquakes cluster along or between the mapped Iapetan faults (also called St. Lawrence paleo-rift faults; Anglin, 1984; Adams et al., 1995). Under the north shore of the St. Lawrence River, for example, hypocentres define a SE steeply dipping plane. When projected to the surface along the trend and dip of that steeply dipping plane, the hypocentres are bounded by two Iapetan faults (the St-Laurent and the Gouffre NW faults; Figure 1.4). Based on this evidence, Anglin (1984) concluded that “almost all the activity in this section of the St. Lawrence River valley is related to preexisting steeply dipping faults that strike along the direction of the valley” (p. 602). Outside the CSZ, the Iapetan fault system extends along the St. Lawrence, Saguenay and Ottawa rivers. Therefore, in these areas, the earthquake hazard may be higher than suggested by the low level of historical seismicity. This geological control of earthquake occurrences has been included in the recent national seismic hazard maps of Canada

² In this dissertation, the term “located” refers to an event for which an epicentre (and often an hypocentre) can be determined. It differs from the terms “recorded” and “detected” earthquake, which represents an event recorded on one station or more, but not necessarily located. Since only located earthquakes are kept in the GSC database, the discussion will only refer to “located” earthquakes.

(Adams et al., 1996).

1.2 Objectives and outline of the dissertation

This dissertation uses newly available information to examine the factors that control the occurrence and spatial distribution of CSZ earthquakes. The current knowledge of CSZ seismotectonics is mainly based on the findings of Anglin (1984). Since then, the scientific database has increased considerably in quantity as well as quality. While Anglin (1984) relied on some 250 hypocentres recorded between October 1977 and May 1983, 1300 additional earthquakes were located between June 1983 and December 1997, using mostly three-component digital data. These high quality records can be used to study focal mechanisms, earthquake clusters, and crustal velocities. In addition, since the work by Anglin (1984), advances have occurred in several related fields: crustal processes leading to fault instabilities; earthquake clustering; and temperature distribution controlling rock rheology. Finally, faults, especially those under the St. Lawrence River, can now be inferred by improved techniques to analyze the gravity and magnetic fields, together with radar images and seismic profiles.

The dissertation is divided into six chapters. In Chapter 1, introductory information on the CSZ geology and earthquakes is presented. The section on seismicity covers the earthquake monitoring, locations, magnitude, frequency and distribution of events. In Chapter 2, factors that control earthquake occurrences (temperature distribution, rheology,

stress differences) are presented and applied to the CSZ. In Chapter 3, an update to structural geology of the CSZ is presented and compared with the seismicity. The section includes the description of new faults, mainly under the St. Lawrence River, as revealed with new data sets and techniques. In Chapter 4, focal mechanisms, earthquake groups and multiplets are integrated into a seismotectonic analysis of the CSZ. In Chapter 5, local earthquake data are used to invert local and regional velocity models. Finally, in Chapter 6, the factors leading to seismicity in the CSZ are discussed based on the results of the previous chapters.

1.3 Geology of the CSZ

Geomorphologically, the CSZ is an area of contrasting topography, with elevations varying between sea level and 1171 meters (Figure 1.5A). While the south shore is a gently rolling landscape, the north shore is a mixture of rugged highlands, plateaus and valleys, separated by dramatic changes in elevation (Figure 1.6). To the west and northwest of the CSZ, the Laurentian Plateau is an area of high elevation (up to 1171 m above sea level), cut by numerous steep valleys (Figure 1.5B). To the north and east, tectonic events and glacial erosion created a gentler relief. To the north of the CSZ, the plateau is cut by a series of EW normal faults related to the creation of the Saguenay Graben in late Precambrian to Ordovician (~700-500Ma). Along the St. Lawrence River, low relief generally corresponds to the hanging wall of normal faults of similar age. In addition, a Devonian meteorite impact (~350Ma) has shattered the plateau, creating a

semi-circular depression 56 km in diameter. The centre of the crater is a 768 m high central peak, Mont des Eboulements, which is surrounded by an interior plateau of up to 15 km radius, and by a peripheral depression of up to 27.5 km radius (Rondot, 1989). In the Quaternary, the multiple passages of the glaciers have preferentially eroded the shattered impact structure and the regional fault zones to the north and west of the CSZ. The St. Lawrence River covers about 25% of the surface of the CSZ. The deepest areas can be 150 m deep near the north shore. Southeast of Ile-aux-Coudres and Ile-aux-Lièvres, the river is generally shallow and the underwater relief very gentle.

Four main geological assemblages make up the geology of the CSZ: the Precambrian Shield of Grenvillian age, the Ordovician St. Lawrence platform, the Appalachian nappes and the Quaternary deposits (Rondot, 1979; Figure 1.7). The main lithologies of the Precambrian Shield are the charnockite-mangerite assemblages, meta-sediments, anorthosite intrusives and local gabbro dykes within fault zones. The charnockite-mangerite assemblages of the Laurentides Park Charnockite Complex cover most of the western part of the region. Locally, the complex is intruded by magmatic bodies, such as the St-Urbain Anorthosite. The Ordovician sedimentary rocks, mainly limestones, rest unconformably on the Precambrian basement, as thin erosion remnants on the north shore, and as thick sedimentary sequences under the St. Lawrence River. The Appalachian sequences are made up of Cambrian sedimentary rocks (sandstones, mudstones) thrust over the Precambrian and Ordovician rocks. The surface expression of this thrust plane is often referred to as Logan's Line. Finally, Quaternary deposits cover

the previous assemblages. On land, these Quaternary deposits are generally thin but pervasive sequences of glacial till and marine clays. Offshore, deposits hundreds of meters thick are found in elongated valleys filled with multi-episodic interglacial sedimentary sequences, covering more than 250,000 years of the St. Lawrence River's history (Occhietti et al., 1997).

All CSZ earthquakes occur within the Precambrian basement (Leblanc and Buchbinder, 1977). The seismogenic basement is cut by faults created during four major tectonic events: the Grenvillian collision (≈ 1100 to ≈ 900 Ma); the rifting episode related to the opening of the Iapetus Ocean (≈ 700 Ma); the Taconian reactivation of these faults at the closing of that ocean (≈ 450 Ma); and finally, a Devonian meteor impact (≈ 350 Ma; Rondot, 1979). In addition to being mapped in the field, most major faults correspond to strong lineaments in remote sensing imagery. Under the St. Lawrence River, geological structures in the seismogenic Precambrian basement are hidden by several kilometres of Appalachian nappes (Lyons et al., 1980) and hundreds of meters of Quaternary sediments. Chapter 3 describes how some fault positions can be determined with gravity, magnetics, remote sensing and seismic profiles.

1.4 Earthquakes of the CSZ

1.4.1 Earthquake monitoring

Between 1977 and 1998, the Charlevoix region has been the only eastern Canadian

seismic zone with a seismograph network sufficiently dense for routine hypocentre determinations. Since October 1977, the area has been monitored by a microseismic network of between 6 to 8 stations located on both shores of the St. Lawrence River (Figures 1.8 and 1.9). Between 1977 and 1988, the Charlevoix seismograph network consisted of a six to seven component vertical short-period analogue network. The signals from the stations were radio-transmitted to a central node where they were archived on a magnetic tape together with a time signal from a local clock. Weekly, the tapes were sent to the head office in Ottawa. Based on events seen on the LMQ analogue records, an analyst (generally F. Anglin of the EPB) would digitize the analogue data at 60 samples/s and print a selected time window of the data with a time scale of 1.5 cm/s (Anglin and Buchbinder, 1981; Figure 1.10). The Y-scale of the traces (gain) was adjusted to ease the phase picking on paper play-outs. Two stations completed the Charlevoix network, one analogue station on the north shore (LMQ) and one station on the south shore (at first, a low gain analogue station, POC, then, an Eastern Canadian Telemetered Network (ECTN) station, LPQ).

In November 1988, the Charlevoix local network became the Charlevoix Local Telemetered Network (CLTN), a digital three-component short-period array. The data, sampled at 80 Hz, were radio transmitted to a central node where timing was added. There, a detection algorithm analyzed each vertical component for seismic events, and, for each trigger, stored a time slice of the whole CLTN data. Daily, the data were transferred to Ottawa and analyzed with an interactive analysis package (Figure 1.10C).

Later, in January 1994, the station LMQ became a digital broadband Canadian National Seismograph Network (CNSN) station, with GPS timing and continuous archival of the data in Ottawa. Finally, in August 1994, each CLTN station was upgraded to a high gain short-period digital instrument with 100 Hz sampling rate and GPS timing. As with any CNSN station, an automatic trigger algorithm continuously scans the CLTN data for events of interest. To detect very small events, LMQ analogue seismograms are also examined. To complete the CSZ network, regional stations provide first motion data and peak amplitudes for magnitude calculations (Figure 1.11).

Between June and November 1996, additional stations were deployed to help define focal mechanisms of micro-earthquakes and to provide additional constraints on the simultaneous velocity inversion for this Ph.D. dissertation. Up to 8 additional sites were occupied by portable seismographs (Lamontagne et al., 1997; Figure 1.12). Six stations were digital three-component short-period instruments sampling at 200 Hz and two were analogue MEQ-800 instruments, used for picking first motions. The network detected some 120 events during the survey.

1.4.2 Hypocentre Locations

CSZ hypocentres are routinely located with direct P (Pg) and S (Sg) phases recorded by the local Charlevoix network. Phases from stations outside the CSZ are not used due to the poorly constrained regional velocity model. The GSC program “GRL”,

based on a grid search algorithm, locates hypocentres with a homogeneous half-space velocity model (6.2 km/s for Pg; 3.57 km/s for Sg). This velocity model does not consider the lower velocity Appalachian rocks, on which three stations are located. With a wedge of lower velocity material, most hypocentres shift by 0.3 to 1.5 km towards the southeast (Lamontagne, 1987). In this thesis, most hypocentral maps show earthquake locations computed with the homogeneous velocity model. The seismic velocity aspects are examined in more details in Chapter 5.

Most CSZ events locate within the seismograph network, with epicentral distances varying between 0 and about 80 km. The average Root Mean Square (RMS) for the solutions is about 0.08 s, with 90% of the RMS less than 0.12 s. These errors are due to the imprecision of phase picking and to the imperfect velocity model. The solutions have average formal errors of ± 0.6 km in epicentral location and ± 1 km in depth, for events within the network (Anglin, 1984). These errors can be larger for events outside the network.

The picking precision of Pg and Sg phases has changed over the years. In the location process, weights are assigned to the phases according to their quality (most Charlevoix stations are quality “A” (± 0.25 second; weight = 4), while some can be B (± 1 second; weight = 1) or C (± 4 seconds; weight = 0.25). Since the original analogue Charlevoix array had only vertical seismometers, it is probable that Sg phases were not as precise as they have been since November 1988, when the CLTN stations became

three-component. Since then, most Pg arrivals are picked on the vertical component, while most Sg phases are measured off the horizontal components.

1.4.3 Magnitudes for CSZ events

Since the beginning of operation of the Charlevoix local network, in October 1977, and December 1997, 2164 earthquakes were located in the CSZ. As for other eastern Canadian earthquakes, the magnitude scales M_L (Richter or local magnitude) and m_N (Nuttli) are used for CSZ earthquakes. Located CSZ earthquakes vary in size between magnitude M_L -0.5 and m_N 5.0. At the Geological Survey of Canada (GSC), the m_N and the M_L scales are used to measure eastern Canadian earthquakes. The following descriptions of the two scales are based on Drysdale et al. (1990).

Nuttli (1973) defines the m_{bLg} scale for two distance ranges: 50 to 400 km and 400 to 800 km. In eastern Canada, the first formula did not give acceptable results in its defined appropriate distance range. Thus, the m_N scale used by the GSC is an extrapolation of the second formula to epicentral distances between 50 and 400 km (R.J. Wetmiller, pers. comm.). The formula applied by the GSC is as follows

$$m_N = -0.1 + 1.66 \log_{10} D + \log_{10} \left(\frac{A}{|KT|} \right) \quad (1.1)$$

where D is the epicentral distance in km; A is the half peak-to-peak trace amplitude of

the S phase in millimetres; T is the period of the trace at the maximum; K is the trace magnification in thousands for that period.

The m_N scale is used for events recorded by stations beyond 50 km. On a given seismic trace, the peak-to-peak trace amplitude of the Lg wave is measured if not over attenuated. If recorded only within 50 km epicentral distance, the event is rated on the M_L scale. The M_L scale is defined as (Gutenberg and Richter, 1956)

$$M_L = \log_{10}\left(\frac{AK_w}{K}\right) - \log_{10} A_0(D) \quad (1.2)$$

where K is the trace magnification in thousands at period T, K_w is the magnification of the Wood-Anderson seismograph for that period, and $\log_{10} A_0(D)$ is the calibration function specified by Richter (1958). This scale was defined for California earthquakes, where the attenuation is higher than in the Canadian Shield. According to Leblanc et al. (1973), the M_L scale can be deficient at distances of less than 50 km, especially in respect to the attenuation of high frequency waves in the near-field, and the local variations of site amplification, radiation pattern and rock properties. Consequently, the M_L scale is used only as a relative measure of the size of CSZ earthquakes.

For the CSZ, the two magnitude scales are approximately related as (Figure 1.13)

$$m_N = 0.63 M_L + 1.41 \quad (1.3)$$

This relation was obtained by applying with robust statistics (Huber, 1981) to the data set described in the figure caption of Figure 1.13. This empirical relation remains approximate due to the large data scatter. Ideally, a uniform magnitude scale for the CSZ earthquakes could be defined with a spectral level analysis (moment magnitude), which we did not attempt in this study. Until then, predictions based on variations of b-values for micro-earthquakes cannot be properly evaluated.

For the time period 1970-1997, the number of located earthquakes is closely related to the number and sensitivity of the local stations (Figure 1.14). Prior to the deployment of the Charlevoix local network in October 1977, only a few tens of earthquakes could be located with the regional stations. The 1970 and 1974 field surveys revealed a high level of earthquake activity, undetected by the regional stations. With the local network, the annual number of located earthquakes increased to nearly 100 in 1978. At the same time, the magnitude threshold decreased from m_N 3.0-3.5 to m_N 2.0 (Figure 1.14).

As described above, two magnitude scales are used for CSZ events. Between 1977 and 1997, a gradual increase in the number of m_N magnitude events has occurred (Figure 1.15). This increase is closely related to changes in the number and the type of stations at regional distances. In the early 1980's, the m_N scale, extrapolated down to 50 km epicentral distance, started to be used routinely in eastern Canada. At the same time,

additional stations of the ECTN were installed within 200 km of the CSZ. These factors explain the increase in $m_N \geq 2.0$ events and the corresponding decrease in $M_L \geq 1.0$ events at around year 1980 (Figure 1.15). Starting in December 1988, the magnitude of smaller m_N events started to be calculated with the installation of the ECTN station Lac Daran (DAQ), located some 90 km west of the CSZ. In August 1994, the CLTN and station DAQ became part of the CNSN network. The continuous archiving of the data allowed the location of small events, previously undetectable, and the calculation of m_N magnitudes with DAQ data (Figure 1.15). These two factors led to an increase in the number of located events (ALL) and $m_N \geq 1.0$ ($M1+$) events. Finally, the 1996 June to November field survey caused a peak in the number of located events, largely due to location of many $M_L < 0$ events.

The initial year of complete reporting of different magnitude levels can be estimated. According to Basham et al. (1982), all $m_N \geq 2.8$ earthquakes have been located since 1968 (Table 1.1). In the period 1970-1976, however, a rather small number of earthquakes with $m_N \geq 3.0$ was recorded (three years out of six with no earthquakes; Figure 1.14). Two possibilities exist: a quiescent period, or a period with a detection threshold $> m_N 3.0$. The apparent increase in the number of $m_N \geq 3.0$ events after LMQ and the local network became operational (in November 1976) is more in line with the second interpretation. The location threshold for events in the 1968-1977 period is therefore conservatively taken to be $m_N 3.5$.

Table 1.1 Estimated first year of complete reporting of magnitude levels in the CSZ (Basham et al., 1982). The magnitudes are defined to one-tenth unit; each category includes earthquakes in a half-magnitude range; ex: magnitude 5.5 includes mag. 5.3- 5.7.

3.0	3.5	4.0	4.5	5.0	5.5	6.0	6.5	7.0
1968	1963	1937	1928	1920	1900	1800	1660	1660

For the period from October 1977 to December 1997, the magnitude of complete reporting is closely related to the early stages of the Charlevoix local network, between October 1977 and November 1988. During that period, the analogue records of station LMQ served as a visual detector of CSZ earthquakes, all located within 70 km epicentral distance. The detection of an event on the LMQ records occurred for an amplitude of about 2 mm (zero-to-peak). For comparison, on 19970217 at 19:47, an m_N 1.7 event, located 55 km from LMQ, gave a peak-to-peak amplitude of 6.7 mm on LMQ. From this, we can estimate that a 2-mm amplitude translates into a m_N 1.2 ($M_L \approx -0.3$). For the period 1977-1997, the location completeness is therefore conservatively taken to be m_N 1.5 (or M_L 0.2 based on Equation 1.3). After the conversion to the CNSN in August 1994, the location threshold is probably lower: during the 1996 field survey, most events missed by routine analysis, but detected by the temporary stations, were smaller than $M_L < 0.0$ (Lamontagne et al., 1997).

1.4.4 Frequency of occurrence of CSZ earthquakes

For the period 1977-1997, nearly 2000 earthquakes have been located in the CSZ

with magnitudes varying between M_L -0.5 and m_N 5.0 (Table 1.2). The distribution of earthquakes in time and space was examined interactively without finding any conclusive pattern. To illustrate this, the epicentre map was divided into sub-zones within which hypocentres showed similar characteristics (closely grouped hypocentres with similar depth distribution; Figure 1.16). While some zones show almost continuous activity (zone 10 for example), others have sporadic activity followed by quiescence (zones 3 and 15 for example; Figure 1.17). Magnitude $m_N \geq 4.0$ events do not seem to follow a definite pattern with other events in their surroundings; they can start, continue or end a period of increased activity.

An earthquake occurrence rate larger than 30 events per 30 days was thought to provide a warning for an imminent magnitude > 5.5 earthquake (Buchbinder et al., 1988). The activity plot by these authors was based on events detected on LMQ (and not necessarily located by the network). Since their data are lost and the exercise ended in the late 80's, I decided to produce a similar plot with the located events, keeping in mind that the local network capability controls the total numbers (Figure 1.18). During the time period 1977 to 1997, the number of located earthquakes varies between 0 and about 30 per 30-day period, similar to what was reported for the period 1977-1985. No rate increase is noticed prior to an event of magnitude $m_N \geq 4.0$.

Using the magnitude completeness and the M_L - m_N relation, the magnitude-recurrence curve was determined (Figure 1.19). For the period October 1977 to December 1997, the $\log(\text{cumulative rate})$ is a linear function of the magnitude, up to about

magnitude 4. The best estimate of the slope β is 2.14. At about magnitude > 4 , the slope changes, implying a higher rate for larger events than suggested by the low magnitude earthquakes. This may suggest that the magnitude > 4 events have overestimated magnitudes, or that there is a real change in occurrence rates at higher magnitudes.

Table 1.2 Number and yearly rate of located CSZ events by magnitude range for the period October 1977 to December 1997. Due to the M_L - m_N magnitude overlap, the numbers are the total number of earthquakes of both magnitude types. Consequently, some annual rate statistics were not computed where they are considered not representative (N.R.).

	All located events	$M \geq 0.0$ (M_L and m_N)	$M \geq 1.0$ (M_L and m_N)	$M \geq 2.0$ (mostly m_N)	$M \geq 3.0$ (m_N)	$M \geq 4.0$ (m_N)	$M \geq 5.0$ (m_N)
Number	2164	1766	838	318	54	8	1
Average Yearly Rate	N.R.	N.R.	N.R.	N.R.	2.7	0.4	N.R.

1.4.5 Spatial distribution and focal mechanisms of CSZ earthquakes

To define the characteristics of the earthquake activity of the CSZ, a hypocentre data set must be defined. From the 2164 earthquakes located between October 1977 and December 1997, some 1450 hypocentres were selected based on at least 5 local stations

with a minimum of 8 Pg and Sg phases (see Figure 1.3). This selection insures an acceptable precision to most hypocentre locations (same as in Anglin, 1984). Considering their distribution in respect to depth, two-thirds of Charlevoix earthquakes occur between 7 and 17 km (Figure 1.20), with more than 80% at less than 15 km. Interestingly, this is also the depth range where two large events occurred (1925 m_b 6.5: 10 ± 2 km; Bent, 1992; and 1979 m_{bLg} 5.0: 10 ± 2 km; Hasegawa and Wetmiller, 1980). The concentration of large earthquakes at 10 km depth may correspond to the strongest part of the crust (Wetmiller and Adams, 1990).

Across the CSZ, the earthquake depth distribution varies; shallow (<5 km) and deep (> 20 km) events, for example, concentrate in certain areas (Figure 1.21A-E). On a section perpendicular to the St. Lawrence River (Figure 1.22C), the hypocentres define two major groups, one that dips steeply to the southeast, and one located under the River without obvious alignment. The section parallel to the River shows the focal depth variations along that trend (Figure 1.21C). Events with magnitude $m_N \geq 4.0$ concentrate near the two zones of magnitude ≥ 4.0 for the period 1924-1978 (Stevens, 1980; Figure 1.23A). Earthquakes of smaller magnitude ($m_N \geq 3.0$ events; Figure 1.23B; $m_N \geq 2.0$ events; Figure 1.23C) do not cluster in one particular subarea.

CSZ focal mechanisms are quite variable in orientation and, to a lesser extent, faulting style (Figure 1.24). It is generally assumed that, on the average, most Charlevoix earthquakes occur as thrust events on preexisting SE steeply ($\geq 60^\circ$) dipping faults. These focal mechanisms, along with additional ones, are examined and discussed in Chapter 4.

Figure 1.1 Earthquakes in eastern Canada. Earthquakes were selected on a completeness basis to ensure a fair representation of the seismicity (historical and instrumentally recorded; Anglin et al., 1990). Damaging earthquakes, with magnitudes in brackets, include: 1663 and 1925 in Charlevoix; 1732 near Montréal; 1929 Grand Banks; 1935 Timiskaming; 1944 Cornwall-Massena; and 1988 Saguenay. Most eastern Canadian earthquakes occur in these seismic zones: Grand Banks; Western Québec; Lower St. Lawrence and Charlevoix.

Figure 1.2 Boundaries of the Charlevoix Seismic Zone (CSZ) after Basham et al. (1982). Some place names referred to in the text are shown.

Figure 1.3 Earthquake hypocenters of the CSZ for the period October 1977-December 1997. Only events with 8 phases from 5 stations are shown. Triangles are the stations of the Charlevoix network, Major faults (black lines) from Rondot (1979).

Figure 1.4 CSZ hypocentres (red circles) of the period October 1977 to May 1983 projected to the surface along a plane oriented N52°E and dipping 70°SE (from Anglin, 1984). Major faults (black lines) from Rondot (1979). Annotations A and B refer to faults that appear to bound the seismic activity according to Anglin (1984). The six-fathom line (in blue) appears to bound the earthquakes to the SE, while the Gouffre NW fault (in green) appears limits earthquakes to the NW. ©1987 Seismological Society of America. All rights reserved.

Figure 1.5 Topography of the CSZ with elevation in meters. (A) With major faults of Rondot (1979) and seismograph stations (triangles). (B) Chromo-stereoscopic image of the north shore of the CSZ integrating RADARSAT-SAR ortho-image with the digital elevation model. The colour range varies from -50 m (in blue) to 1082 m (in red) The texture of the land surface comes from the Radar data (image courtesy of T. Toutin, Canada Centre for Remote Sensing; Toutin and Rivard, 1997).

Figure 1.6 Scenery of the Charlevoix Seismic Zone. (A) Location map of the photos; (B) Gouffre River valley which represents a part of the peripheral graben of the impact structure; (C) Linear coastline of the north shore, created by normal faults; (D) The Malbaie River U-shaped valley, created by a fault and deeply eroded by glaciers; (E) View of the impact structure from La Malbaie and of its central peak, the Mont des Eboulements; (F) View of the impact structure from Baie-St-Paul. Ile-aux-Coudres to the left belongs to the Appalachian nappes.

South Shore Cap-a-l'Aigle La Malbaie Mont des Eboulements

(E)

Gouffre Valley Baie-St-Paul

Ile-aux-Coudres

(F)

Figure 1.7 Main geological features of the CSZ overlaying the elevation model. Tectonostratigraphic units are grouped as Precambrian units (dotted), Ordovician sedimentary rocks (bricks) and Appalachian rocks (slanted bricks). Logan's Line (LL) represents a structural discontinuity between the complexly deformed rocks to the southeast (the Appalachian nappes) and the rocks to the northwest (the Canadian Shield with its Ordovician sedimentary cover). Major fault positions (black lines) of the north shore are from Rondot (1979).

Figure 1.8 Time-history of the Charlevoix network between 1970 and 1998. Three field experiments took place: in 1970: Leblanc et al., 1973; in 1974: Leblanc and Buchbinder, 1977; and in 1996: Lamontagne et al., 1997. Acronyms used: (1) LPQ (La Pocatière), SHQ (Saint-Hilaire), LMQ (La Malbaie); and A56, co-located with LMQ, are seismograph stations; (2) CLTN: Charlevoix Local Telemetered Network; (3) CNSN: Canadian National Seismograph Network; (4) sp: short period; (5) bb: broad band; (6) CSZ network: Charlevoix Short-Period Local Network (6 stations).

Figure 1.9 Location of the permanent seismograph stations in the CSZ. Some stations were relocated over the first few years, without changing the general distribution of stations. Since 1985, the stations are: A11, A16, A21, A54, A61, A64. Other stations are: broadband (LMQ); digital short-period (LPQ); and analogue (POC; LMQ until 1993, SHQ).

Figure 1.10 Examples of how arrival times have been picked over the years. Traces of the 19840214 09:01 m_N 2.5 event from which Pg and Sg phases were picked during the period October 1977 to November 1988. (A) High gain display. The top and bottom X axes are second marks. The six traces correspond to stations A10; A16; A20; A64; A54 and A61. Arrival times were measured by interpolating between the two vertical 10 second lines. Since only the vertical traces existed, Pg phases were generally more evident than the Sg phases. (B) Low gain display. Some Sg phases were picked on this plot. (C) Example of a computer display used to pick Pg and Sg phases after 1993 (software “dan” by Nanometrix Inc). The icons (top) represent the various options to get, display or analyse data. The lower three traces are the three components of station A54 (Z: vertical; N: north-south; E: east-west). The time scale for these three traces can be seen at the bottom (40 seconds total). The top trace represents an enlarged view of 2 seconds of the trace A54..EHZ. Phases are picked in the zoom window, with adjustable X and Y scaling.

A

Figure 1.11 Regional stations surrounding the CSZ (triangles), digital (DPQ; DAQ) and analogue (CIQ; QCQ; SLQ).

Figure 1.12 Seismograph stations in the CSZ (triangles) during the 1996 Summer field survey: permanent (red); digital portable (blue); analogue (black); Lamontagne et al., 1997).

ML

Figure 1.13 Correlation between mN and ML magnitudes, computed with stations beyond 50 km epicentral distance for mN, and calculated from local Charlevoix stations at 25 to 50 km epicentral distance for ML. A total of 888 data points were used, which implies that many data points are superimposed especially between ML -0.5 and MN 1.0. The optimum curve is obtained with a linear regression routine.

Figure 1.14 Distribution of located earthquakes for the period October 1977 - December 1997 with respect to different magnitude thresholds: (top) for all events (ALL); $M \geq 0$; and $M \geq 1$; (bottom): for $M \geq 2$, $M \geq 3$ and $M \geq 4$ (circles). No distinction is made between the ML and mN magnitude scales. Major changes in the local network are shown in the upper diagram.

Figure 1.15 Number of events for various magnitude thresholds; (top): mN magnitude; (bottom) ML magnitude.

Figure 1.16 Sub-zones around earthquake clusters with similar depth distribution. These sub-zones are discussed in Chapter 4.

Figure 1.17 Time-history of the activity of the various sub-zones of Figure 1.16 for (A) the period October 1977 - December 1987; (B) The period January 1988 - December 1997.

B

1988-1997

Figure 1.18 A 30-day histogram of located earthquakes in the CSZ, between 1970 and 1997. Events of magnitude $m_N \geq 4.0$ are shown with stars. In March 1989, an earthquake doublet occurred (shown by a star at position $Y = 2$).

Figure 1.19 Magnitude-recurrence curves for the CSZ. Data points in blue are based on events before October 1977 that meet the magnitude completeness criteria of Table 1.1 (less than 300 points). Red dots represent the previous events, plus the events between October 1977 and December 1997 that meet the magnitude completeness criteria defined in Section 1.4.4. The blue and red curves represent the magnitude-recurrence curves (minimum, average and maximum) based on their respective data sets. In the upper right corner, the parameters of the recurrence curves in red are given: Mx : maximum magnitude; $Beta$: slope of the curve; $N5$: number of events at magnitude 5; N : total number of events; $N0$: number of events at magnitude 0.

Figure 1.20 Focal depth distribution of the CSZ earthquakes for the period October 1977- December 1997. Events are grouped in one kilometre intervals. Two styles of curves are drawn: one for the number of earthquakes at a given depth and one for the cumulative percentage. Two data sets are plotted, all events (blue) and events that met the magnitude completeness criteria (red).

Figure 1.21 CSZ hypocentres and stations (circles and triangles respectively) plotted on an elevation map of the CSZ: (A) 0 to 5 km focal depth; (B) 5 to 10 km depth; (C) 10 to 15 km depth; (D) 15 to 20 km depth; (E) deeper than 20 km. Circle colours represent the focal depth of the events.

Figure 1.22 CSZ hypocentres: (A) Hypocentres (circles) and stations (triangles) with the end points of the two cross-sections B-B' and C-C' plotted on an elevation map of the CSZ; (B) Projection of all hypocentres on a vertical cross-section perpendicular to the St. Lawrence River. The width of the active zones may represent the volume where fractures are reactivated as well as variations of fault trends. Red circles are the magnitude ≥ 4.0 earthquakes; (C) Projection of all hypocentres on a vertical plane parallel to the St. Lawrence River. Red circles are the magnitude ≥ 4.0 earthquakes.

Figure 1.23 Hypocentres by magnitude ranges. Hypocentres (circles) and stations (triangles) plotted for various magnitude ranges: (A) $mN \geq 4.0$ (the two large circles are the two areas of larger earthquakes of Stevens (1980); (B) $mN \geq 3.0$; (C) Magnitudes (mN and ML) ≥ 2.0 . Circle colours represent the focal depth of the events.

Figure 1.24 Lower hemisphere focal mechanisms in the CSZ. The zones of compressional readings are shaded. The maximum and minimum pressure axes are shown as P and T, respectively. The dates of the events and the magnitudes (M) of the two largest events are shown: (a) from Bent, 1992; (b) from Hasegawa and Wetmiller, 1980; (c) from Lamontagne, 1987; (d, e) from Wetmiller and Adams, 1990; (f, h, i, j, l, m) from Leblanc and Buchbinder, 1977; (k, n) from Adams et al, 1988; (g, o) from Adams et al., 1989.

Chapter 2

Thermal and rheological constraints on the spatial distribution of CSZ earthquakes

2.1 Introduction

Worldwide, numerous attempts have been made at explaining why, in a given region, crustal earthquakes do not occur beneath a certain depth. To be compared with the rheological characteristics of the crust, earthquake focal depths are evaluated in three different fashions in the literature: the depth where a decrease in the distribution of earthquakes occurs (D_{Dec} ; Meissner and Strehlau, 1982); the depth above which 80% of earthquakes occur ($D_{80\%}$; Smith and Bruhn, 1984; Doser and Kanamori, 1986; Banda and Cloetingh, 1992); and finally the maximum focal depth of earthquakes (named thereafter D_{Max} ; Chen and Molnar, 1983; Kuszniir and Park, 1984; Ranalli and Murphy, 1987; Chen, 1988). As described in Chapter 1, CSZ earthquakes occur exclusively in the mid- to upper crust, from the surface to a depth of 30 km, with D_{Dec} at about 13 km and $D_{80\%}$ at 15 km (Figure 2.1). In the CSZ, D_{Max} is at 29 km; however, only 1% of earthquakes occur between 25 and 29 km depth. We therefore adopt the depth of 25 km ($D_{99\%}$) as a representative value for the lower limit of the seismic crust. This should not be interpreted as a sign that earthquakes between 25 and 30 km are negligible; as a reminder the 25 November 1988 m_{BLg} 6.5 Saguenay earthquake occurred in the Grenville Province about 100 km west of the Charlevoix Seismic Zone, with an unusual focal depth of 29 km (North et al., 1989). The final discussion of this chapter will briefly examine the

significance of this earthquake, and of others in the 25 to 30 km range, in light of the Charlevoix modelling.

Rheological modelling can provide constraints to the earthquake depth distribution, from the seismogenic brittle layers at upper to mid crustal depth to the plastic regime in the mid- to lower crust. To model the factors that favour or inhibit the occurrence of earthquakes, lithology, structure, temperature and pressure, pore fluid pressure, stress and strain rate have to be considered (Ranalli, 1991). In this chapter, some important factors in modeling the brittle and ductile parts of the crust are reviewed and applied to the CSZ. Based on these results, the factors controlling the CSZ depth distribution are discussed.

2.2 Temperature

Crustal temperatures can be estimated assuming that conductive heat transfer is predominant in continental intraplate environments such as Eastern Canada. Steady state 1-D heat transfer in a medium with heat production $A(z)$ is used and can be described by a Poisson equation (see e.g. Chapman, 1986)

$$\frac{\delta^2 T}{\delta z^2} = -\frac{A}{k} \quad (2.1)$$

where T is temperature, z depth, A heat production rate, and k thermal conductivity. By double integration with respect to z , and by assuming q_T is the heat flow at the surface, one obtains for a layer of constant heat generation and thermal conductivity

$$T(z) = T_T + \frac{1}{k} (q_T z - \frac{A z^2}{2}) \quad (2.2)$$

where T_T is the temperature at the surface.

If a layer has thickness Δz , then, at the bottom of the layer, the temperature T_B and the heat flow q_B can be expressed in terms of the temperature and heat flow at the top of the layer (T_T , q_T) and properties (A, k) of the layer

$$T_B = T_T + \frac{1}{k} (q_T \Delta z - \frac{A \Delta z^2}{2}) \quad (2.3)$$

$$q_B = q_T - A \Delta z \quad (2.4)$$

Equations (2.3) and (2.4) can be used to calculate $T(z)$, resetting T_T and q_T at the top of each new layer with the values T_B and q_B solved for the bottom of the previous layer. This formula can be used for one-dimensional modelling of the geotherms where heat production is assumed constant within layers. Although one-dimensional modelling represents an oversimplification (Ranalli, 1991), the uneven data distribution for the Grenville Province restricts studies to the one-dimensional case.

There has been considerable debate as to how heat generation is distributed with depth. Another possibility is an exponential decrease of heat generation with depth

$$A(z) = A_0 \exp(-\frac{z}{D}) \quad (2.5)$$

where DA_0 represents the contribution from radiogenic heat sources in the upper crust, with D being related to the characteristic thickness of the heat generating layer. This approach is based on the known enrichment in K, Th and U of upper crustal rocks. The

exponential decrease model is not considered appropriate anymore; heat generation data from the 12 km deep Kola peninsula well support a step-like distribution rather than an exponential distribution (Drury, 1989). Thus, a step-like distribution is assumed in the following sections.

2.2.1 Surface heat flow

The most important factor in geotherm calculations is surface heat flow. At a given site, measurements are subject to experimental errors (generally considered small) and to near-surface perturbations of conductive heat transfer, such as ground water circulation. In continental environments, surface heat flow can be measured in different geological environments such as in deep wells in hard rock, in dry water wells and in lake bottom sediment columns. At a given site, the assigned heat flow value represents an average of values (temperature gradient and conductivity) from different depths. In general, the deeper measurements are the most reliable, being remote from near-surface perturbations caused by water circulation, topography variations or paleoclimatic disturbances. Due to these factors, it may be necessary to assume an uncertainty of at least 10% in heat flow measurements (Drury, 1987). In assigning a heat flow value to a given geological province, many assumptions have to be made: that quality of the data is high; that heat flow is relatively constant over the area (or that the mean and standard deviation of heat flow values provide a fair estimate of the real values); and that all geological settings of the province were properly sampled. In other words, a heat flow value can be considered typical of an area if heat flow measurements are reliable (i.e. the effects of near-surface perturbations are removed), numerous enough to be statistically

significant and spatially distributed over most sub-areas.

2.2.2 Heat production

Although the uncertainty in heat generation for a given sample is relatively small, variability of heat generation can be very high for a given rock type (Drury, 1987). Heat production studies provide estimates of various "typical" rock types that can be related to crustal levels. In general, upper crustal rocks have higher heat production values than the lower crust (generally higher than $0.4 \mu\text{W}/\text{m}^3$, and up to more than $5.0 \mu\text{W}/\text{m}^3$). This range is related to the more felsic composition of the upper crust with its higher concentration in radioactive elements. In most studies, upper crustal heat production is determined from values of surface rocks and from geological and geophysical knowledge of the area. In general, lower crustal heat production values are assumed to be in the $0.4 \pm 0.3 \mu\text{W}/\text{m}^3$ range, which corresponds to granulite facies rocks (Ashwal et al. 1987).

2.2.3 Thermal conductivity

Although thermal conductivity may vary from sample to sample, it is less variable than heat production and is generally assumed to be constant within a given crustal layer (generally between 2.5 and $3.0 \text{ W}/\text{m}\cdot\text{K}$). As will be examined in more detail in other sections, small variations in thermal conductivity have little effect on geotherms.

2.2.4 Thickness of layers

Geophysical information, such as seismic and gravity data, can be used to assess the thickness of the crustal layers. For geotherm calculations, typical values of heat generation and thermal conductivity can be associated with the assumed geological units. Major uncertainties arise when no geophysical profiles exist in a given area or when the depth extent of geological units cannot be adequately constrained. Due to spatial variations of tectonic history over any given geological province, extrapolations of thicknesses over the whole area is necessarily faced with large uncertainties. Generally, interpretations will rely on rather simple models with a few layers of different thermal characteristics. Fortunately, the few kilometres of uncertainty in layer positions do not have much of an impact on geotherm calculations (Chapman, 1986).

2.2.5 Mantle heat flow

Certain additional constraints govern the depth extrapolation of heat generation values. Surface heat flow represents the sum of the heat generated in the crust plus the contribution from the mantle. Although not directly measurable, mantle heat flow is generally assumed to represent between 1/3 and 2/3 of the surface heat flow. In addition, it is natural to assume that mantle heat flow is more or less constant beneath Precambrian shields. For the Superior Province for instance, Guillou et al. (1994), assume mantle heat flow to range between 10 and 14 mW/m². Jaupart et al., (1998) give a range of 7 to 15 mW/m² for the Canadian Shield as a whole. The crustal contribution represents the remaining part of the surface heat flow. With these constraints it is generally possible to

build plausible models of crustal heat flow.

2.3 The brittle (frictional) regime

From a rock mechanics point of view, earthquakes are dynamic slip instabilities usually occurring on preexisting sliding surfaces (Scholz, 1992). In a highly fractured intraplate environment subject to deviatoric stresses, such as the CSZ, rocks cross-cut by pre-existing fractures can slide abruptly, generating earthquakes. The process leading to fault reactivation is controlled by pressure, but is independent of rock type and temperature as a first approximation. To cause slip on a fracture (i.e. a fault in the case of an earthquake), the shear stress, induced by the existence of a tectonic stress field, has to overcome friction. Amonton's law (Jaeger and Cook, 1979) is an empirical formulation of the problem

$$| \tau | = \mu_o \sigma \quad (2.6)$$

where τ and σ are the shear and normal stress acting on the plane of fracture, and μ_o denotes the friction coefficient. The frictional law can also be written as the Navier-Coulomb criterion:

$$| \tau | = S_o + \mu_o (\sigma - P_f) \quad (2.7)$$

where S_o is the cohesion and P_f is the pore fluid pressure, assuming compressive stresses are positive.

Assuming $S_o = 0$ and $\mu_o = 0.75$, Sibson (1974) derived expressions of the critical

stress difference (strength) necessary to reactivate preexisting fracture planes favourably oriented with respect to the stress field, i.e. with an angle $\psi = (\frac{1}{2}) \tan^{-1} (1/\mu_o) \approx 26.5^\circ$ with the axis of maximum compression (σ_1)

$$\sigma_1 - \sigma_3 = \beta \rho g z (1 - \lambda) \quad (2.8)$$

where σ_3 is the minimum compressive stress, β is a factor related to the type of faulting (3 for thrust, 1.2 for strike-slip and 0.75 for normal faults), ρ the average density of the overlying rocks at depth z , g the acceleration of gravity, and λ the pore fluid factor ($\lambda = P_f / \rho g z$).

Expanding on Sibson's formulation, Yin (1993) derived more general expressions including variations in ψ , S_o , and μ_o , as well as fracturing of isotropic rocks (with S and μ , the cohesion and friction coefficient for unfractured rocks). For thrust faulting on a preexisting fracture, the expression of the critical stress difference is

$$(\sigma_1 - \sigma_3) = \frac{2\mu_o \rho g z (1 - \lambda) + 2S_o}{\sin 2\psi + \mu_o (\cos 2\psi - 1)} \quad (2.9)$$

Yin (1993) also derived expressions for the limiting angles for reactivation of preexisting faults, with $\phi_o = \tan^{-1}(\mu_o)$

$$\psi_{\min} = \frac{1}{2} \left\{ \sin^{-1} \left[\frac{(\mu_o^2 + 1) [(\mu_o \rho g z (1 - \lambda) + S_o) + (S\mu_o - S_o\mu)]}{(\mu_o^2 + 1) (\mu \rho g z (1 - \lambda) + S)} \right] - \phi_o \right\} \quad (2.10)$$

$$\psi_{\max} = \frac{1}{2} \left\{ \pi - \sin^{-1} \left[\frac{(\mu_o^2 + 1) [(\mu_o \rho g z (1 - \lambda) + S_o) + (S\mu_o - S_o\mu)]}{(\mu_o^2 + 1) (\mu \rho g z (1 - \lambda) + S)} \right] - \phi_o \right\} \quad (2.11)$$

If the angle between the maximum compressive axis and the fracture is outside these boundary values, a new fracture is created (if sufficient stress differences exist). Before applying these expressions to a crustal model of the CSZ, we review the main factors contained in Equations 2.9 to 2.11. Figure 2.2 shows graphically the parameters intervening in fault reactivation.

2.3.1 Crustal stress magnitude and orientation

Fault reactivation occurs when the deviatoric stresses are sufficiently large to make the shear stress overcome friction. Fault strength can be computed from equation 2.9; however, assessing the level of deviatoric stress in the crust is more difficult. In an intraplate environment such as the CSZ, there are many potential sources of deviatoric stress, some existing from plate-boundary processes and some of local origin (Hasegawa et al., 1985). In the North American plate, ridge push appears to be the main contributor to tectonic stress (Zoback, 1992a) with some contribution from basal drag. Local sources can also add to the ambient stress field. These sources are flexural stresses (such as glacial loading/unloading), lateral density contrasts, and crustal inhomogeneities (intrusions, structural boundaries, intersecting lineaments, impact craters) that can reorient and concentrate stresses. Studies of the local sources of stress have shown that glacial rebound stress are of the order of less than 5 MPa⁴ (Hasegawa et al., 1985). They could act as a trigger mechanism on optimally oriented pre-existing faults on the verge of failure (Wu and Hasegawa, 1996). Lateral density variations induce stresses that do not exceed

4

One MPa corresponds to 10 bars.

10 MPa in the CSZ (Assameur and Mareschal, 1995). Worldwide, direct measurements and modelling of tectonic stresses suggest that the maximum crustal stress difference has an upper limit of about 100-200 MPa (see Figure 2.3 with references).

To be reactivated, faults have to be favourably oriented with respect to the stress field. With all other conditions constant, a fault with an optimum orientation will necessitate a lower critical stress difference to cause slip. In the CSZ case, small scale fractures exist at all scales due to the multiple tectonic episodes of the region. Large scale features, on the other hand, concentrate along certain trends (see section 1.3).

2.3.2 Coefficient of friction and cohesion

The coefficient of friction μ (either for intact rock μ or for fractured rock μ_o) is defined as (neglecting cohesion)

$$\mu = \tau / \sigma \quad (2.12)$$

where τ and σ are the shear and normal stress acting on the surface during sliding (Byerlee, 1978). At low pressures (less than 5 MPa), experimental data show a wide scatter as friction relates mainly to surface roughness. At higher pressures, friction becomes independent of rock types and surface roughness. Byerlee (1978) obtained two expressions for the shear stress

$$\tau = 0.85 \sigma \quad (2.13)$$

for $\sigma_3 < 114$ MPa (or $z < 7$ km where σ_3 is vertical), and

$$\tau = 50 \text{ MPa} + 0.6 \sigma \quad (2.14)$$

for $200 \text{ MPa} < \sigma_3 < 2000 \text{ MPa}$ (or $z > 7 \text{ km}$ where σ_3 is vertical).

In most cases, μ varies between 0.5 and 0.8, with a generally accepted value of 0.75. Notable exceptions are fault gouges consisting of clay minerals that can have μ as low as 0.2. Lower values for the coefficient of friction widen the range of fault orientations that can be reactivated (see Equations 2.10 and 2.11). Finally, the cohesion of the surface (or cementation strength; Sibson, 1992) is the shear stress necessary to cause slip even in the absence of normal stress. It is generally of the order of 1 to 20 MPa for preexisting fractures (Etheridge, 1983).

Recent experiments have shown that a high pressure type of brittle fracture exists where the critical stress difference increases with pressure less steeply than in the frictional case (Shimada, 1993). At higher confining pressures, the high-pressure type of fracture occurs, since compressive strength becomes lower than the frictional strength. If confirmed by further experiments, these results may provide a mechanism by which the crust's strength may not reach values as high as those predicted by Byerlee's law at mid-crustal depth. At a certain depth, the crustal strength would increase, at a lower rate than near the surface, until the ductile field is reached.

2.3.3 Pore fluid pressure

Pore fluids can favour fracturing by lowering the normal stress acting on the fracture plane and/or by chemically weakening the fault surface (Locker, 1995). The

effective normal stress (σ_{eff}) is

$$\sigma_{eff} = \sigma - P_f \quad (2.15)$$

where P_f is the pore pressure. The pore fluid factor (λ) can be expressed as the ratio of pore fluid pressure to lithostatic pressure

$$\lambda = P_f / \rho g z \quad (2.16)$$

Absence of fluids implies $\lambda = 0$, $\lambda \approx 0.4$ (hydrostatic) means that the pores are connected from the surface to the depth considered and the pore fluid pressure is $P_f = \rho_w g z$, where ρ_w is the density of water; and a lithostatic pressure ($\lambda \approx 1.0$) implies that fluid pressure is equal to the overburden pressure. Suprahydrostatic pressures can occur by consolidation of sediments (in sedimentary basins) and by dehydration reactions in metamorphism (Scholz, 1990). The mechanism to sustain such pore fluid pressures, especially in areas without tectonic or volcanic activity, is still debated. There is geologic evidence that zones of suprahydrostatic pore fluid pressures are fairly common (Carter and Tsenn, 1987). As will be discussed below, this has implications for the nature of deformation at depth and for the magnitude of stress necessary to reactivate faults.

2.3.4 Density of overlying rocks

The average density of the overlying rock is used to compute the lithostatic pressure which, in the case of thrust faulting, is the minimum compressive stress (σ_3 vertical). We use an average density $\rho = 2.8 \text{ g}\cdot\text{cm}^{-3}$ for crustal rocks.

2.4 The ductile regime

The increase of temperature and pressure with depth affects the behaviour of rocks and fracture planes subject to deviatoric stress. When temperature and pressure are sufficiently high, minerals tend to flow instead of behaving in a brittle fashion. With increasing temperature, the whole rock can flow, leading to bulk flow (ductility or plasticity). Rock ductility can be modeled from extrapolations of experiments conducted at higher strain rates. The strength of the ductile part of the lithosphere is generally determined from the relation

$$\sigma_1 - \sigma_3 = \left(\frac{\dot{\epsilon}}{A}\right)^{\frac{1}{n}} \exp\left(\frac{Q}{nRT}\right) \quad (2.17)$$

where $\sigma_1 - \sigma_3$ is the creep strength (stress difference necessary to achieve a given steady-state strain rate $\dot{\epsilon}$, R is the gas constant, T the absolute temperature and A , n and Q are material parameters which are approximately stress- and temperature-independent in the range of interest (Kirby, 1983; Ranalli, 1995).

Although strain is generally localized in highly deformed areas such as shear zones of interplate regions, it is acceptable to use average strain rate in intraplate areas where deformation affects large areas. Whereas rates can be as high as 10^{-12} s^{-1} in regions such as the San Andreas fault, in an intraplate environment such as the CSZ, $\dot{\epsilon}$ is assumed to be in the 10^{-15} to 10^{-16} s^{-1} range (Hasegawa, 1986). For comparison, 10^{-16} s^{-1} corresponds to a yearly change of 3 mm over a horizontal distance of 1000 km. There are no direct strain rate measurements for the CSZ.

The material parameters A , n and Q are known for a selection of silicate materials (examples in Table 2.1; Carter and Tsenn, 1987; Kirby and Kronenberg, 1987). The experimental uncertainties of these creep materials correspond to a 10 to 20% uncertainty in the accuracy of the creep strength (Ranalli, 1991). The creep strength of silicate rocks is approximately inversely proportional to the silica content. In the lower crust, this translates into basic rocks being stronger than acidic to intermediate rocks. Water, even in trace amounts, can enhance rock ductility (Carter and Tsenn, 1987). Thus, a number of rocks have parameters measured as dry and wet. Because finer materials are generally weaker (since creep is effected by vacancy diffusion instead of dislocation motion), small grain sizes lower rock strength and can lead to linear rheology (Ranalli, 1991). If found in bands of regional extent, these finer materials can become shear zones by concentrating strain (Kirby and Kronenberg, 1987). Section 2.6 describes how the passage from the brittle to the ductile regime can be related to the earthquake activity.

2.5 Velocity weakening-velocity strengthening transition

When the crustal temperature reaches about 300°C, the frictional properties of fault zones change and a temperature-controlled sliding stability transition occurs (see e.g. Scholz, 1990). Unlike bulk flow which is related to whole rock properties, this transition impacts on the sliding properties of the fault surface. It was found experimentally that rock volumes subjected to sliding along a fracture plane, will react differently to a change of sliding velocity, depending on the temperature. At temperatures below 300°C, rocks

Table 2.1 Creep parameters of rocks used in this study

Material	A (MPa ⁻ⁿ s ⁻¹)	n	Q(kJ mol ⁻¹)	Reference
Adirondack Granulite	8.00e-03	3.1	243	Wilks and Carter (1990)
Orthopyroxenite, wet	6.30e-03	2.8±0.5	271±18	Ross and Nielsen (1978)
Pikwitonei Granulite	1.40e+04	4.2	445	Wilks and Carter (1990)
Websterite, dry	4.00e-07	4.3±0.6	326±19	Ave Lallemand (1978)
Websterite, wet	2.00e-05	5.3±0.4	382±27	Ave Lallemand (1978)
Clinopyroxenite, wet	1.50e+05	3.3±0.9	490±159	Boland and Tullis (1986)
Clinopyroxenite, dry	1.00e-05	5.3±1.1	380±30	Kirby and Kronenberg (1984)

will be in the velocity weakening mode, i.e., a change in the sliding velocity can lead to an unstable situation brought about by a decrease in the frictional shear strength. At higher temperature, velocity changes can lead to increased frictional shear strength, i.e. to velocity strengthening. The change from velocity weakening to velocity strengthening is possibly related to materials of the fault surface. For a quartzo-feldspatic crust, possible critical temperatures are 300°C and 450°C which correspond to the onsets of quartz and feldspar plasticity respectively (Scholz, 1988). Below the critical temperature, the fault is in the seismogenic velocity weakening field, while above it the fault is in the aseismic velocity strengthening field (Tse and Rice, 1986; Scholz, 1990). For granite, experiments with high fluid pressures show that the transition temperature is in the 350°C range (Blanpied et al., 1991).

2.6 Thermal and rheological controls on CSZ earthquakes

In most intraplate seismic areas of the world, the thickness of the seismogenic crustal layer correlates with surface heat flow (Sibson, 1982; Chen and Molnar, 1983; Chen, 1988). Geotherms control the maximum depth of crustal earthquakes (D_{Max}) either by a transition from brittleness to ductile flow (the brittle-ductile transition) or by a transition from velocity-weakening to velocity-strengthening fault behaviour. Both processes occur approximately at the same temperature.

In the first model, D_{Max} corresponds to the rheological change from a seismogenic brittle upper layer, modeled by the Navier-Coulomb criterion (Byerlee's law), to a ductile lower layer where strain is accommodated by aseismic creep (the brittle-ductile transition;

see e.g. Meissner and Strehlau, 1982; Sibson, 1982; Fadaie and Ranalli, 1990). The strength of the ductile part is modelled by an extrapolation to low strain rates of steady-state creep equations (Equation 2.17). This approach has limitations (cf. Rutter and Brodie, 1992), chief among which (in addition to uncertainties in thermal and rheological parameters) is the assumption that strain is uniform, while observation shows that strain is highly localized in deep crustal exposures. However, it provides a first-order estimate of the rheology of the crust with a general rule that the critical temperature for bulk flow is about one-half of the melting temperature.

In the second model, the lower limit of seismicity is caused by a modification of slip behaviour. D_{Max} corresponds to a temperature-controlled sliding stability transition rather than to a transition in bulk rheology (see e.g. Scholz, 1990). For a quartzofeldspatic crust, seismicity stops when a critical temperature of 300°C for quartz-rich rocks or 450°C for feldspar-rich rocks is reached (Scholz, 1988).

In most seismic regions, both the brittle-ductile and the velocity weakening-strengthening models fit as a first approximation (which implies that the difference between the two critical temperatures is not large). However, in some areas, such as northern Switzerland (Deichmann and Rybach, 1989) deep crustal earthquakes occur, apparently in the ductile zone.

The following sections examine the rheological controls on the Charlevoix earthquake depth distribution. First, our one-dimensional geothermal modelling is described. The computed geotherms are used to infer the transition depth to bulk flow or

to velocity strengthening. Finally, the frictional constraints imposed on seismicity in the mid- to lower crust are discussed.

2.6.1 Calculation of geotherms

Temperature is a critical factor in rheological models, but the uncertainties of geotherm calculations are often underestimated. Generally, these calculations use measured values of surface heat flow, and assumed values of thermal conductivity and heat production. This approach underestimates geotherm uncertainties. In the present analysis, a set of 22,000 geotherms is computed using wide ranges of values for all geothermal parameters. The calculation is based on one-dimensional heat conduction (equations 2.3 and 2.4). Fixed ranges of heat production and thermal conductivity within layers of known thicknesses are assumed.

The distribution of heat flow measurements is poor in the Grenville, with a mean value typical of Precambrian Shield regions of $41 \pm 10 \text{ mW/m}^2$ (mean value ± 1 standard deviation; Pinet et al., 1991; see Figure 2.4). There are only three surface heat flow determinations for the CSZ, one in the Appalachians (55 mW/m^2) and two in the Precambrian basement: 65 mW/m^2 (La Malbaie) and 57 mW/m^2 (Les Eboulements; Guillou-Frottier et al., 1995). The higher than average values for the two CSZ sites are due to higher heat production rates (Guillou-Frottier et al., 1995). Two series of calculations are made in the following: one with the $41 \pm 10 \text{ mW/m}^2$ mean value for the surface heat flow in the Grenville, and one with a surface heat flow of 60 mW/m^2 to consider the higher values measured in the CSZ (to be discussed later). The crust in the

Charlevoix region is 43 km thick with a discontinuity at 18 km depth (from seismic refraction; Lyons et al., 1980). On the basis of geological evidence (Emslie and Hunt, 1990), the upper layer is assumed to be more felsic than the lower one which is taken to be a basic residue of partial melting, rich in plagioclase and without quartz. This hypothesis is also supported by the rather high P-velocities (~ 6.7 - 6.8 km/s) measured below 18 km (Lyons et al., 1980; see Figure 5.23). In this two-layer crustal model, upper crustal heat production has been assumed to vary between 0.3 and $2.0 \mu\text{W}/\text{m}^3$ to include lateral heterogeneities such as granitic intrusives (high heat production) and large anorthositic bodies (low heat production). Lower crustal heat production has been taken to vary between 0.1 and $0.5 \mu\text{W}/\text{m}^3$ (granulite facies rocks; Ashwal et al., 1987). Thermal conductivity is assumed to vary between 2.3 and $2.7 \text{ W}/\text{m}^\circ\text{C}$, a range that covers most crustal rock types. Table 2.2 lists the ranges of values for thermal parameters. Three additional conditions are used: heat production in the upper crust has to be equal to or larger than that in the lower crust, as shown by measurements of representative rocks (Ashwal et al., 1987); mantle heat flow has to be in the 10 - $25 \text{ mW}/\text{m}^2$ range, which is plausible for the Precambrian Shield (it includes the 10 and $14 \text{ mW}/\text{m}^2$ assumed for the Superior Province by Guillou et al., 1994); and thermal conductivity in the upper crustal layer is higher than or equal to that of the lower crustal layer. Computations are performed assuming all possible combinations of surface heat flow values between 31 and $51 \text{ mW}/\text{m}^2$ (in $1 \text{ mW}/\text{m}^2$ increments), heat production (in $0.1 \mu\text{W}/\text{m}^3$ increments) and thermal conductivity (in $0.1 \text{ W}/\text{m}^\circ\text{C}$ increments) for the ranges specified above.

The results of 22,000 geotherm calculations can be summarized by five geotherms;

Table 2.2: Thickness and thermal parameters of crustal layers in the Charlevoix region

Layer	Thickness (km)	Parameter	Minimum	Average	Maximum
Surface		Surface heat flow (mW/m ²)	31.	41.	51.
Upper crust	18.	Heat production (μ W/m ³)	0.3	0.8	2.0
		Thermal conductivity (W/m°C)	2.3	2.7	3.1
Lower crust	25.	Heat production (μ W/m ³)	0.1	0.3	0.5
		Thermal conductivity (W/m°C)	2.3	2.5	2.7
Mantle		Mantle heat flow (mW/m ²)	10.		25.

Minimum; Minimum 90%; Average; Maximum 90%; and Maximum (Figure 2.5A). Minimum and Maximum are the absolute minimum and maximum values obtained for any combination of parameters. Average represents the average of all temperature calculations. MIN90% and MAX90% are the 5 and 95 percentiles, such that 90% of the iterations fall within them. At 25 km, 90% of all computed temperatures fall between 215 and 355°C (curves MIN90% and MAX90% of Figure 2.5A). We refer to this 90% range of values when discussing seismicity.

As already pointed out by Chapman (1986), surface heat flow is the most important uncertainty in thermal calculations. In Figure 2.5B, Q_{min} and Q_{max} denote the geotherms obtained by using the lowest and the highest values of surface heat flow and the average values of the other parameters (Table 2.2). Temperature variations due to other thermal factors are constrained within an approximate $\pm 50^\circ\text{C}$ band (arrows in Figure 2.5C). In the upper crust, variations are more influenced by heat production. While the range computed in this work is very wide (due to the inclusion of all plausible values of thermal parameters) and includes all previous values in the Precambrian Shield, the average temperature at 20 km ($T \approx 280^\circ\text{C}$) is very close to that estimated by Mareschal (1991) for the Charlevoix area. The Q_{min} - Q_{max} range also includes the most relevant geotherms of Chapman (1986) (Figure 2.5C). One-dimensional geotherm calculations that consider plausible ranges of thermal parameters clearly show the uncertainties of crustal temperatures at depth. To consider the possibility that the two surface heat flow measurements of Charlevoix are representative of a locally higher thermal regime, geotherm calculations were made with a 60 mW/m^2 surface heat flow. As expected,

higher crustal temperatures are obtained (Figure 2.6).

2.6.2 Ductile regime

Using equation 2.17 and the geotherms computed above, the creep strength can be calculated for given crustal material parameters. The lower crustal rocks used in this thesis are: Adirondack (felsic) and Pikwitonei (basic) granulites; dry and wet orthopyroxenites; dry and wet clinopyroxenites; and dry and wet websterites (rheological parameters given in Table 2.1). The Adirondack granulite has a much higher quartz content (21%) than plausible for the Charlevoix mid to lower crust and is only used as an upper felsic (soft) limit.

Results for three different geotherms (Minimum 90%; Average; Maximum 90%) are shown in Figures 2.7A, B and C. Geotherms lower than or equal to the Average produce a brittle crust down to 30 km or more. The maximum depth of earthquakes cannot be controlled by the brittle-ductile transition if the average Grenville geotherm applies to the Charlevoix region and the lower crust is basic. With the Maximum 90% geotherm, bulk flow may start at about 25 km, but only for the quartz-rich Adirondack Granulite. Lower strain rates could also raise the transition but the change is only a few kilometres for a decrease of two orders of magnitude. Passage to ductility at 25 km depth requires geotherms very close to the upper limit, or felsic rock composition (Adirondack granulite facies types). Ductility would be more probable with a wet lower crust as proposed by Hyndman and Shearer (1989). If the lower crust is dry as indicated by some interpretations of lower crustal petrology (Frost and Bucher, 1994), the crust would be

entirely within the brittle field.

Using the geotherm computed with a surface heat flow of 60 mW/m^2 , bulk flow was calculated for the same materials (Figure 2.8). As expected, the brittle-ductile transition is shallower for this warmer geotherm.

2.6.3 Change from velocity weakening to velocity strengthening

In the frictional model, the transition from velocity weakening to velocity strengthening behaviour marks the lower boundary of the seismic layer. Scholz's (1988) seismo-rheological model assumes that D_{Max} coincides with the onset of ductility of the most ductile rock component. Experimentally, 300°C marks the onset of quartz plasticity and 450°C corresponds to the onset of feldspar plasticity. Since the Charlevoix middle and lower crustal rocks are assumed not to contain quartz, the critical temperature should be the latter. The 450°C isotherm is found at about 33 km for the Maximum 90% and deeper for colder geotherms. A possible hydrolytic weakening of the feldspars could bring the critical transition temperature down to 350°C (Tullis and Yund, 1980). At 25 km, this temperature is reached by geotherms close to Maximum 90% (Figure 2.4A). Consequently, seismogenic control by onset of hydrated feldspar plasticity is possible with our warmest geotherms.

2.6.4 Maximum plausible stress difference

Another problem arises with the predicted stress differences necessary to cause frictional failure. The Coulomb-Navier failure criterion (with low pore fluid pressures and

a static coefficient of friction; Equation 2.8 and Figure 2.6) implies very high stress differences at mid- to lower crustal depths under dry conditions. 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 1200 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 discussed in Section 2.3.1.

Stress differences causing thrust fault reactivation can be computed as a function of depth, material parameters and fault orientation (Ranalli and Yin, 1990; Yin and Ranalli, 1992). The two parameters that can maintain $(\sigma_1 - \sigma_3)$ below 200 MPa in the mid- to lower crust are a high pore fluid pressure and/or a low coefficient of friction. Figure 2.9A shows the critical stress difference necessary to cause thrust faulting at a depth of 20 km, assuming a static friction coefficient of 0.75, as a function of the angle θ that the fault makes with the maximum compression direction and the pore fluid/overburden pressure ratio (λ). High pore-fluid pressures equal at least to 80% of overburden pressure must exist for $(\sigma_1 - \sigma_3)$ to be less than 200 MPa. However, the steeply-dipping rift faults ($\approx 70^\circ$) are not favourably oriented for thrust reactivation, assuming the sub-horizontal maximum stress direction that prevails in eastern North America (Zoback and Zoback, 1991), which may be slightly reoriented in the CSZ due to postglacial rebound (Wu, 1998). At a depth of 20 km, the limiting angle between pre-existing fault and maximum compression direction for reactivation as a thrust fault is 54° for any pore-fluid pressure below or equal to the lithostatic. An additional possibility is a reduced coefficient of friction. Figure 2.9B shows the maximum dip angle of a fault

with respect to the maximum compression direction assuming that the coefficient of friction is 0.3. Reactivation of faults dipping up to 72° is now possible, and the required pore-fluid pressures are considerably less. Assuming a 200 MPa maximum stress difference, it is possible to map the relationship between the pore-fluid pressure, the coefficient of friction and the angle between σ_1 and the fault surface (Figure 2.10). There is also the possibility that the paleo-rift faults become listric at depth. According to Figure 2.10B, fractures that make an angle of 40° with the axis of maximum compression can be reactivated with coefficients of friction in the 0.6-0.75 range. Although high pore-fluid pressures are still required ($\lambda \sim 0.8$), listric faults could be reactivated with standard values for the coefficient of friction. Finally, the high pressure type of fracture may also account for the maximum crustal strength at a level slightly above 200 MPa (Shimada, 1993; Figure 2.11).

2.7 Discussion and conclusions

In Charlevoix, $D_{99\%}$, $D_{80\%}$ and D_{Dec} do not appear to correspond to a transition to ductile bulk rock behaviour unless the geotherm is near the maximum possible value and/or the lower crust is quartz-rich. This conclusion is based on one-dimensional geothermal modelling of Grenville surface heat flow for a wide range of thermal parameters and on an examination of the rheology of materials likely to form the lower crust. None of the materials can be ductile within our 90% confidence limits for the geotherms. Only a high surface heat flow ($\geq 60 \text{ mW/m}^2$) can lead to a passage to ductility at $D_{99\%}$, with a quartz-rich mid- to lower crustal composition. However, the high heat flow value ($Q \sim 65 \text{ mW/m}^2$; Guillou-Frottier et al., 1995) appears to be affected by the high

thermal conductivity (J.-C. Mareschal, pers. comm.).

Therefore, although there is a global correlation between surface heat flow and maximum depth of crustal earthquakes (Sibson, 1982; Meissner and Strehlau, 1982; Chen and Molnar, 1983), a detailed analysis of the CSZ shows that the brittle-ductile transition is probably deeper than the $D_{99\%}$ of earthquakes. It should be noted that the present model uses a basic lower crust while most models assume a uniform felsic crust with the rheology of quartzite or granite. The present choice, on the other hand, is based on inferences on the composition of the mid- to lower crust of the Charlevoix region. Another difference between this approach and models of other regions is the consideration of temperature uncertainties. Most geotherm models assume fixed values for thermal parameters.

The passage from velocity weakening to velocity strengthening frictional behaviour may explain $D_{99\%}$. Feldspar only starts flowing at about 450°C, but this temperature can be lowered to $\approx 350\text{--}400^\circ\text{C}$ by the presence of H_2O (Tullis and Yund, 1980). Consequently, at 25 km depth, hydrated feldspar can possibly become ductile with the warmest geotherms (including the one with 60 mW/m² surface heat flow). Below $D_{99\%}$, the absence of three local instability factors may explain the quasi-complete absence of earthquakes. The available stress difference may not reach the frictional strength level. High pore-fluid pressures may exist only above $D_{99\%}$, possibly due to hydration reactions that eliminate any free water at temperatures over 250°C (Frost and Bucher, 1994). Finally, below $D_{99\%}$ fractures may be less numerous or more stable possibly due to healing. Figure 2.12 presents a synoptic model of seismogenic factors in this region.

Unless the high-pressure type of fracturing is validated by further testing, high pore fluid pressure and/or low coefficient of friction must exist not to exceed the maximum crustal stress difference level. Earthquakes occur where these conditions exist. The presence of deep crustal fluids and of high pore fluid pressures has gained much support in recent years (Hyndman and Shearer, 1989; Meissner and Wever, 1992). Examples for eastern Canada are numerous: Zoback (1992b) used a high pore-fluid pressure to explain the apparent mismatch between the regional compressive stress direction and the focal mechanism of the 1979 m_{bLg} 5.0 Charlevoix earthquake; shear wave splitting observed for Charlevoix earthquakes can be explained by saturated cracks (Buchbinder, 1989); in the Miramichi earthquake area in New Brunswick, near-lithostatic pore-fluid pressures can explain reactivation of high angle faults (Sibson, 1989). One open question, however, is how such high pore fluid pressures are created and maintained. In the CSZ, no active geological process, such as tectonism, sediment compaction or metamorphic dehydration, are likely to give rise to suprahydrostatic fluid pressures. In addition, it has been argued that at temperatures higher than 250°C, free water cannot exist due to a rapid consumption by hydration reactions (Frost and Bucher, 1994). Although the debate about the presence or absence of mid- to lower-crustal fluids is open, high pore-fluid pressures appear to solve the seismological problem of maximum possible stress difference at mid- and lower-crustal depth.

The alternative is a low friction coefficient. In a dry thrust environment with sub-horizontal maximum compressive stress, a low friction coefficient is necessary to reactivate steeply-dipping faults. This possibility was also suggested by Zoback (1992b).

Near the surface, low friction coefficients may correspond to thick fault gouges made up of clay minerals; at depth, however, these minerals lose their lubricating properties due to dehydration (Scholz, 1990). A low friction coefficient at mid-crustal depth may correspond to highly fractured zones. Variations in hypocentral depths along the trend of the St. Lawrence River, and variability in reactivated fault planes shown by focal mechanisms (Figure 1.26) indicate that Charlevoix earthquakes occur in zones of intense fracturing.

In terms of the seismic hazard estimate in the St. Lawrence valley, the presence of paleo-rift faults alone does not explain earthquake occurrence. Local fault conditions, such as transient enhanced pore-fluid pressures, leading to frictional instabilities should also be considered.

Worldwide, large intracontinental earthquakes tend to occur at the base of the seismogenic layer defined by micro-earthquakes, which represents the peak in the shear resistance of a rheological profile (Sibson, 1984). In contrast, the two largest CSZ earthquakes for which focal depths are known (1925 m_b 6.2: Bent, 1992; and 1979 m_{bLg} 5.0: Hasegawa and Wetmiller, 1980) occurred at about 10 km depth. In addition, the 5 to 15 km depth range also includes magnitude ≥ 3.0 earthquakes for the period 1977 to 1997. Based on our rheological modelling, the concentration of larger earthquakes between 5 and 15 km depth cannot be associated with the much deeper brittle-ductile transition. Again, local factors, such as high pore fluid pressures, degree of fracturing, or lower coefficient of friction, may explain the enhanced activity of this crustal level.

The m_{bLg} 6.5 1988 Saguenay earthquake occurred at 29 km depth, in the Grenville Province. Based on our rheological modelling, it probably occurred in the brittle crust, but much closer to the brittle-ductile transition, if a warm Grenville geotherm (Max90%) is considered (a situation similar to Figure 2.8). In this case, the earthquake may have occurred in the strongest part of the crust, as proposed by Sibson (1984). This would be consistent with the unusually high stress drop of this event (Atkinson and Boore, 1998).

If the two higher than average CSZ surface heat flow measurements of Guillou-Frottier et al. (1995) are representative of a regional anomaly in the Grenville, the $D_{99\%}$ may correspond to the brittle-ductile transition (assuming the warmest geotherm). In this case, many earthquake concentrations in the 20 to 30 km depth range could be seen as “sticking patches” in a surrounding ductile regime (e.g. Sibson, 1984). Interestingly, most earthquakes below 20 km depth occur in an elongated zone, that could be located between more ductile areas. It would be interesting to see if the next large CSZ earthquake nucleates at 20 to 30 km depth, near one of these locking points.

Although this study examines the case of the Charlevoix seismic zone, most questions raised in this analysis also apply to other seismically active areas of the continental crust. It appears that the onset of bulk flow is only one factor affecting the maximum depth of intraplate earthquakes. Charlevoix earthquakes are not controlled by the brittle-ductile transition. The probable composition of the mid- to lower crust and the surface heat flow show that rocks are still brittle at $D_{99\%}$. A passage from velocity-weakening to velocity-strengthening at 300-350°C may correspond to $D_{99\%}$, but only if this transition temperature applies to hydrated basic rocks, where feldspar is the most ductile

material. Some instability factors may explain the occurrence of earthquakes: high pore-fluid pressure and low friction coefficient, possibly related to unhealed zones of intense fracturing. Independently of the pore-fluid pressure, a low coefficient of friction may be another factor to account for the thrust reactivation of steeply-dipping faults.

Figure 2: Field data for the period October 1974 to September 1975. The plot shows the number of earthquakes (N) versus the magnitude (M). The data points are plotted for the period October 1974 to September 1975. The x-axis is labeled 'Number and percentage' and the y-axis is labeled 'Magnitude'.

Figure 2.1 Focal depth distribution of the Charlevoix earthquakes for the period October 1977 to December 1997. Events are grouped in one kilometre intervals. Two curves are drawn: one for absolute numbers (blue) and one for the cumulative percentage (red). See text for descriptions of cut-off depths.

Figure 2.2. Sketch representing the various factors intervening in fault reactivation.

Figure 2.3 Distribution of inferred maximum stress differences. The numbers refer to: 1) Ridge push: Zoback and Magee (1991); 2) Upper crustal amplification of ridge push: Kusznir and Bott (1977); Hasegawa et al. (1985); 3) Tensile fracturing : Etheridge (1983); 4) Earthquake stress drops in intraplate environments: Hasegawa et al. (1985); 5) Estimate along the San Andreas fault: Lachenbruch and Sass (1980); 6) Hydrofracturing-strain relief: McGarr and Gay (1978); 7) Composite estimates from the summation of eastern Canadian stress sources: Hasegawa et al. (1985); 8) Isostatic considerations: Molnar and Lyon-Caen (1988); 9) Mylonitization: Christie and Ord (1980); 10) Paleopiezometers: Kohlstedt and Weathers, 1980); 11) KTB borehole value measured at 6 km: Zoback et al. (1993); 12) Estimates along the Alpine fault of New Zealand: Scholz et al. (1979); 13) Mechanical twinning: Tullis (1980). Except for mechanical twinning, a 200 MPa is a conservative upper bound estimate of the maximum crustal stress difference.

Figure 2.4 Distribution of heat flow measurements in the Grenville Province, with circles proportional to the heat flow values, given in mW/m^2 (from Jessop et al., 1984; Drury, 1987; Pinet et al., 1991; Guillou-Fortier et al., 1995).

other

Figure 2.5 Computed geotherms for the Grenville region, (A) using parameters of Table 2.1. Five curves are shown: the absolute Maximum (MAX) and Minimum (MIN) geotherms; the Average (Av) geotherm which represent the average value of all geotherms at a given depth; and the MIN90% and MAX90% which enclose 90% of all geotherms. Ninety-nine percent of all earthquakes occur above $D_{99\%}$. (B) Temperature variations due to one parameter while others are kept at their average values of Table 2.1. Surface heat flow is the most important factor, while all others constrain geotherm fluctuations within 50°C when the average heat flow is used. (C) Comparisons of MIN90%, AV, and MAX90% geotherms with the geotherms of Chapman (1986) for a surface heat flow of 40 mW/m^2 (Ch40) and 50 mW/m^2 (Ch50). Both fall within the present 90% confidence margin.

Figure 2.6 Geotherms calculated with a surface heat flow of 60 mW/m² (Min 60: minimum; Ave60 average; Max60; maximum), compared with the minimum 90% and maximum 90% geotherms of Figure 2.5A.

Figure 2.7 Crustal strength as a function of depth. The brittle part is computed from the Ranalli and Yin (1990) formulation using $\mu = 0.75$, $S_0 = 10$ MPa, $\rho = 2.6 \text{ g}\cdot\text{cm}^{-3}$ and $\lambda = 0.0$ (dry) and 0.9. The ductile part is computed using equation 2.17, with rheological parameters given in the text, a strain rate of 10^{-16} s^{-1} , and three geotherms: A) Minimum 90% ; B) Average; C) Maximum 90%. The rocks used are: Adirondack Granulites (AG); wet Orthopyroxenite (wO); Pikwitonei Granulite (PG); wet Websterite (wW); and wet Clinopyroxenite (wC). Rocks that do not appear in the ductile part of the diagrams have a creep strength higher than 10000 MPa. MaxS placed at 200 MPa indicates the maximum crustal stress difference discussed in the text and interpreted from Figure 2.3.

Figure 2.8 Crustal strength as a function of depth, with same creep parameters as Figure 2.7, except for the geotherm which is the Max60 geotherm of Figure 2.6.

Figure 2.9 Stress difference (MPa) necessary to reactivate a fault located at a depth of 20 km for various angles between the maximum compressive stress axis and the fault and various ratios of pore fluid pressure λ : A) With a friction coefficient $\mu = 0.75$ (standard); B) with a low friction coefficient $\mu = 0.3$. In A), a maximum stress difference of 200 MPa can only be reached with very high λ . With a low friction coefficient, the maximum reactivation angle is increased and the necessary pore-fluid pressure is decreased.

Coefficient of friction = 0.75

Coefficient of Friction = 0.3

Figure 2.10 Map of the relation between pore-fluid pressure, coefficient of friction and reactivation angle, assuming a maximum stress difference of 200 MPa. A) Three dimensional view, with colours representing the pore fluid pressure ratio; B) a two dimensional view. Taking an angle of 30° as an example, one can see that the assumed value of 0.75 for the friction coefficient requires pore fluid pressures in the > 0.80 range. Similarly, if the angle between σ_1 and the fault plane is 70° , a < 0.35 friction coefficient is necessary to reactivate the fault. Areas in black are outside the range $0.0 \leq \lambda \leq 1.0$.

A

Angle of σ_1 with the fault plane.

Figure 2.11 Example of the impact of a change in fracture type for the rheological profile. At a given depth (10 km in this example), the brittle fracture changes from Byerlee's law (A), between the dry (black) and the wet (blue) curves, to a high pressure type of fracturing (A'). This change in fracture type maintains the strength near the maximum plausible crustal level (MaxS). Earthquakes can occur down to the ductile level without the need for very high stress difference.

Figure 2.12 Synoptic model of the factors controlling the earthquake occurrences in the CSZ. On a cross-section perpendicular to the St. Lawrence River, the earthquake distribution is shown, with D99% at 25 km depth. The major factor influencing the depth distribution is the temperature. Two possibilities exist: an average geotherm, where 300°C is reached at about 28 km, or the Max90% geotherm, where 300°C is reached at about 20 km. D99% may correlate with a number of factors: the passage to the velocity strengthening fault behaviour; a lessening of the pore-fluid pressure; or an increase in the friction coefficient. The brittle-ductile transition, however, lies below D99% with the average geotherm. In the brittle mid-crust, earthquakes occur in regions of enhanced pore-fluid pressures or of low coefficient of friction, possibly due to highly fractured zones.

Chapter 3

Geological faults in the CSZ and their correlation with earthquakes

3.1 Introduction

The relation between the earthquakes of the Charlevoix Seismic Zone (CSZ) and the St. Lawrence rift faults is a major element in the seismic zoning of Eastern Canada. It is generally assumed that CSZ earthquakes cluster along or between the St. Lawrence paleo-rift faults (also called Iapetan faults; Anglin, 1984; Adams et al., 1995). This relationship is obscured by the incomplete knowledge of geological faults, especially under the St. Lawrence River. In this chapter, the geological faults of the CSZ are described, in some cases for the first time, using geophysical, geological and remote sensing information. This analysis relies on new geoscientific information (gravity, radar images, seismic profiles) and more powerful analysis techniques of the potential fields. Based on this information, the spatial correlations between geological faults and earthquakes are reexamined. The chapter is divided in three parts. First, the geoscientific data sets and the interpretation methods are presented. Second, regional faults are described with newly-acquired remote sensing imagery, potential fields, seismics and geological mapping. Finally, the relationships between the earthquakes and faults of the CSZ are discussed. Figure 3.1 displays the main place names used in text.

3.2 Data sets and methodology

Three major types of information are used to refine knowledge of the positions of

geological faults: remote sensing imagery (airborne and satellite-borne radar); geophysical data sets (seismics, magnetics and gravity); and field mapping of the north shore (Roy, 1978; Rondot, 1979; 1989).

3.2.1 Remote sensing imagery

On land, remote sensing imagery, mainly radar, is central to analysis and interpretation. Passive remote sensing systems, such as SPOT or Landsat, record the energy reflected by the Earth's surface at frequencies roughly equivalent to those detected by our eyes. Landsat images of the region were examined but did not yield any structural information not already described by Rondot (1979). Radar sensors, on the other hand, are active systems that send a microwave pulse towards the Earth's surface and measure the amount of energy reflected (Figure 3.2). Unlike passive systems, radar is not affected by atmospheric conditions, such as darkness, clouds, rain, dust, or haze. For geological applications, the incidence angle of the beam enhances terrain topography, often related to structural lineaments.

Over the course of this study, radar coverage of the CSZ became available: Canadian airborne SAR (Synthetic Aperture Radar), the satellite-borne ERS-1 (European Remote Sensing Satellite), JERS-1 (Japanese Earth Resources Satellite) and Radarsat (Canada) (Figure 3.3), each one with its own look directions and incidence angles. Most images were corrected geometrically using ground reference points, an acceptable approximation for the low relief areas near the centre of the impact structure but approximate for the steep elevation gradients of the Laurentians. The Radarsat data were

ortho-corrected by the Canada Centre for Remote Sensing, resulting in accurate lineament orientations (T. Toutin, pers. comm.). The various radar images partly overlap over the crater.

On the radar images, the most striking regional feature is the impact crater. Inside the crater, the topography is generally subdued, due to the intense fracturing that favoured erosion. In the Laurentians, north-south lineaments are sub-parallel to the main direction of the glacier flow (Roy et al., 1993; De Sève et al., 1994), suggesting preferential glacial erosion along that trend. Other smaller lineaments match the regional lithologies.

3.2.2 Geophysical data sets

Geophysical data sets include the magnetic and Bouguer gravity fields and a series of seismic profiles acquired in the early 1970's, under the leadership of the Société Québécoise d'Initiatives Pétrolières (SOQUIP).

3.2.2.1 Magnetism

Regional magnetic maps show the spatial distribution of geological units with different magnetic susceptibilities. Sedimentary rocks have the lowest average susceptibility and basic igneous rocks, the highest (Telford et al., 1976). In the CSZ, these differences in magnetic susceptibilities result in the north shore having a highly variable magnetic field, due to the outcropping Precambrian rocks, and the south shore having a very smooth field due to the deep Precambrian basement under the almost magnetically transparent Appalachian nappes (Figure 3.4). The magnetically contrasting areas

correspond to variations in Precambrian lithologies and in Precambrian basement depth under the Appalachians. Steps in the Precambrian basement depth can correspond to normal faults. Our depth estimates are not dependent on the magnitude of the anomaly, but rather on the anomaly shape, especially the sharpness. An outcropping Precambrian unit will produce a sharp anomaly, whereas the same body will have a low frequency signature if buried under a few kilometres of Appalachian nappes.

With the Euler deconvolution technique, steep gradients in the potential fields can be interpreted as sources of various shapes and depth (Figure 3.5). Assuming a shape for the source of the magnetic anomaly, the Euler deconvolution provides the depth and the location of the source (Thomson, 1982; Reid et al., 1990). The technique is based on Euler's homogeneity equation

$$(x - x_0) \partial T / \partial x + (y - y_0) \partial T / \partial y + (z - z_0) \partial T / \partial z = N (B - T) \quad (3.1)$$

where (x_0, y_0, z_0) is the source position, whose total field (T) is detected at (x, y, z) , with a regional field value (B). The factor (N) is the order of homogeneity, also called the structural index. The structural index depends on the geometry of the causative body: for magnetics, a sub-vertical contact corresponds to $N = 0$; for gravity, $N = 1$. Other possible shapes for the magnetics are: sphere ($N=3$), vertical pipe ($N=2$), and dike ($N=1$). The method, explained in more detail in Reid et al. (1990), is as follows. First, the gradients $\partial T / \partial x$, $\partial T / \partial y$, $\partial T / \partial z$ are calculated from the gridded data. Second, a square window within the grids of gradient values is determined (10 X 10 for example). Third, for a given structural index, all points in the window are used to solve Euler's equation (3.1) for a

source position (x_0 , y_0 , z_0) and a background value B using least-squares estimates. The solution is kept if the depth uncertainty is less than a set proportion of the estimated depth. Four, the previous steps are repeated for all possible window positions over the map area. Five, maps of the solutions are produced with a symbol showing the solution's depths.

Euler deconvolution is very sensitive to the gradients of the field and is very useful to identify contacts, or body edges. Reid et al. (1990) provide examples of its use to delimit faults and geological units. With its down-faulted Precambrian basement under the St. Lawrence River, the CSZ geology is ideally suited for this method. The possibilities offered by the Euler solutions were demonstrated with the gravity field (Keating, 1998). Using the CSZ magnetic map, Euler solutions with $N=0$ provides an estimate of the location and depth of a sub-vertical contact separating lithologies with contrasted magnetic susceptibilities. Hence, where the Precambrian outcrops, Euler solutions should be shallow. Under the St. Lawrence River, along with the Precambrian basement, Euler solutions progressively deepen. Consequently, the Euler solutions provide estimates of the position and depth of the Precambrian contacts.

The Euler solutions for the CSZ are shown in Figure 3.5. As one can see, most of the shallow solutions are found on, or near, the north shore, where the Precambrian either outcrops, or is near the surface. The solutions can be used to map the depth variations of the Precambrian basement. From seismic reflection profiles, no Euler solutions exist where the Precambrian is deeper than about 6 km.

3.2.2.2 Gravity and rock densities

Bouguer anomaly maps represent the gravity field, which depends mainly on the crustal distribution of rock densities. Until 1994, the CSZ Bouguer anomaly map was based solely on 300 land measurements, leaving out the most seismogenic part of the CSZ, over the St. Lawrence River. Following a request from the Seismicity Section, the GSC gravity group completed the coverage in August 1994. In the CSZ, the St. Lawrence River is too shallow and narrow for a conventional ship gravity survey. Consequently, a LaCoste and Romberg dynamic gravity meter controlled by a portable computer was installed on a small boat (Figure 3.6; Lamontagne et al., 1995). Differential Global Positioning System (GPS) phase measurements were used to compensate uneven boat motions in the strong tidal currents. To minimize the corrections due to the rotation of the Earth, survey lines were mostly oriented north-south and east-west. Upstream from the CSZ, where navigation is constrained by water depth, the north and the south channels were surveyed. With this survey, the offshore coverage is now better than that on land (Figure 3.7A).

To model gravity anomalies, rock densities were measured on 288 samples from the rock collections of the GSC and the Québec Department of Natural Resources, and some hand specimens. Densities were determined using standard techniques for non-porous rocks (weight in air and water). Our results are comparable to other studies (Table 3.1; Figure 3.7C). The densities of the Grenville rocks show considerable scatter, due to the highly variable lithologies. Our mean density value for the Grenville, $2.71 \text{ g}\cdot\text{cm}^{-3}$, is somewhat lower than elsewhere in the Grenville ($2.74 \text{ g}\cdot\text{cm}^{-3}$, Keary and Thomas, 1979;

2.76 g·cm⁻³, Loncarevic et al., 1990; 2.85 g·cm⁻³, Feininger, 1993). The lower density may be due to the intense fracturing induced by the impact. The densities of the Ordovician limestones and the Appalachian nappes are clustered around a mean 2.65 g·cm⁻³. The density for the hard, compacted and boulder rich Quaternary sequence was established at 2.3 g·cm⁻³ (B. Long, pers. comm.). For modeling purposes, density contrasts with the Grenville were -0.15 g·cm⁻³ for the Ordovician-Appalachian rocks, and -0.5 g·cm⁻³ for the Quaternary sequence.

3.2.2.3 Seismic profiles

During the early 70's, SOQUIP lead a series of offshore seismic reflection surveys to define the hydrocarbon potential of eastern Québec (Figure 3.8). SOQUIP kindly provided paper copies of the profiles, and later on, the digital data from one seismic line for reprocessing (line 13). The number of folds generally determine the quality of the seismic profiles (Table 3.2). In most profiles, seismic energy trapped in the shallow water of the St. Lawrence River produced multiple reflections that masked deeper reflectors. Most profiles show sub-horizontal reflectors, some in the Quaternary deposits, some in the underlying Precambrian basement and in the Appalachian nappes. The Precambrian-Appalachian interface is generally difficult to identify, probably due to the small acoustic impedance contrast between the indurated Ordovician carbonates and the underlying Precambrian. In the figures and discussions, a seismic P-wave velocity of 6.0 km/s is used to convert time to depth (from seismic refraction results; Lyons et al., 1980). This

Table 3.1: Rock Densities

Rock units	Number of samples	Mean $\text{g}\cdot\text{cm}^{-3}$	Standard Deviation $\text{g}\cdot\text{cm}^{-3}$	Minimum $\text{g}\cdot\text{cm}^{-3}$	Maximum $\text{g}\cdot\text{cm}^{-3}$
GRENVILLE	205	2.71	0.16	2.52	4.05
Granite	10	2.66	0.05	2.61	2.78
Anorthosite	6	2.74	0.05	2.67	2.85
Charnockite-mangerite	10	2.75	0.10	2.64	2.81
Gabbro	6	2.95	0.19	2.66	3.23
Gneiss	35	2.77	0.13	2.64	3.14
APPALACHIANS	21	2.65	0.05	2.55	2.75
Ordo. LIMESTONES	31	2.67	0.04	2.54	2.73

Table 3.2 Seismic profiles of the CSZ

Line No.	Company-Date	No. of folds	Qual	Source	Appr. Length (km) ⁵	Remarks.
A	SOQUIP - 1971	12 (SW) 24 (NE)	A	Vibroseis	88	Line in two sections. One part inside the crater was never processed.
13	SOQUIP - 1971	24	B	Vibroseis	13	Line was reprocessed.
16	SOQUIP - 1971	12	B	Vibroseis	9	
17	SOQUIP - 1971	12	B	Vibroseis	8	
18	SOQUIP - 1971	12	B	Vibroseis	10	
21	SOQUIP - 1971	12	B	Vibroseis	9	
22	SOQUIP - 1971	12	A	Vibroseis	12	Clear and continuous PC-Appalachian boundary at 1.0 sec on most of the southern part of the profile.

⁵Length inside the map area.

23	SOQUIP - 1971	12	B	Vibroseis	13	Structures in the basement. The north channel shows discontinuous but poorly defined reflectors from 0.20 to 0.40 sec.
27A	CDRO- 1971.	1	C	Air Gun	14	
28	CDRO, 1971.	1	C	Air Gun	8	
30	CDRO, 1971.	1	B	Air Gun	16	
31	CDRO, 1971.	1	B	Air Gun	16	
35	CDRO, 1971.	1	B	Air Gun	19	
37	CDRO, 1971.	1	B	Air Gun	19	
42F	CDRO, 1971.	1	C	Air Gun	6	
42N	CDRO, 1971.	1	B	Air Gun	11	
130	SOQUIP-1972	6	A	Air Gun	15	
140-A	SOQUIP - 1972	6	B	Airgun	8	

Note: CDRO: Central-Del Rio Oils Ltd

velocity, acceptable for the Precambrian, is too high for the Appalachian (≈ 5.6 km/s) and Quaternary sequence (≈ 2 km/s). Thus, near surface reflectors in the Quaternary and Appalachian sequences are shallower than suggested by the depth values on the right of the profiles. The profiles are discussed in more detail below.

The marine vibroseis seismic line 13, shot across the St-Lawrence River in 1970, was reprocessed using up-to-date processing techniques. The initial seismic section showed up to 4 seconds (TWT) of data contaminated with water-bottom related multiples (energy bouncing between the bottom of the river, and/or within the sedimentary layers beneath, and the top of the water column). The seismic profile was reprocessed to image upper-crust and mid-crustal reflectors, which can be seismogenic faults, and to refine the regional geological model.

To maximize the possibility of finding deep reflectors, an extended correlation of the raw seismic records and the marine vibroseis sweep was performed, yielding a 13 second long data set (TWT). In theory, this procedure could image reflectors down to a depth of 30 km. At first, the processing sequence involved the extended correlation and integration of the survey geometry. The traces were inspected one by one to discard noisy recordings and to pick the first breaks. From a spectral analysis, most of the signal was contained between 8 and 55 Hz.

Minimum phase spiking deconvolution, followed by predictive deconvolution based on first break energy, were performed on shot gathers to compress the signals, level the spectrum and remove water-bottom related multiples. Velocity analysis turned out to be

the most important processing step. Semblance and constant velocity analyses were performed. Due to the presence of strong multiples, picks were made on constant velocity stacks. Best seismic velocities vary between 1500 m/s in the unconsolidated sediments to more than 5000 m/s in the Precambrian shield. The 24-fold data set was stacked and coherency filtering was applied. The resulting section is shown in Figure 3.13 G and H, and discussed in section 3.2.4.

3.2.3 Testing the methodology on the magnitude m_N 5.1 Cap-Rouge earthquake

On November 6, 1997 a magnitude m_N 5.1 earthquake occurred near Cap-Rouge, in the western suburbs of Quebec City (Nadeau et al., 1998). Since the depth to basement is well-assessed by high quality SOQUIP seismic profiles, this earthquake occurrence provided an opportunity to ground truth the methodology developed for Charlevoix. Geologically, the Quebec City region is very similar to the CSZ; it is an area where the Appalachian nappes were thrust over the rifted Precambrian.

Since the Appalachian nappes are nearly magnetically transparent, the magnetic field partly depends on the depth of the Precambrian Shield. Where the Precambrian Shield outcrops, such as to the North and West of Quebec City, the magnetic field is made up of very high frequency variations in its intensity. Where the Precambrian basement is deeper, such as under the south shore of the St. Lawrence River, the high frequency variations disappear, leaving a smooth field with low frequency variations. This change in wavelength of the magnetic anomalies is due to an abrupt change in the depth of the Precambrian rocks, most likely due to large throw on normal faults.

The correlation between the magnetic data and the normal faults inferred from the seismic profiles is good. These faults, which have a vertical throw of 2 km or more, are easily identified in the magnetic data. In general, they correspond to elongated magnetic anomalies whose response is slightly higher on the upthrow side than on the downthrow side. They trend at about N45°E.

The Cap Rouge epicentre is located on an elongated magnetic anomaly which is a normal fault of the Precambrian basement with vertical slip exceeding 2 km (Figure 3.9). If the faults dip steeply towards the southeast, as it is generally assumed for the St. Lawrence rift faults, the Neuville fault, located to the NW, may be the reactivated fault, an interpretation supported by the focal mechanism (Nadeau et al., 1998). The mapped Neuville fault is imaged in the magnetics and in the seismics, with a ≈ 2 km throw. The good correlation between the solutions of the Euler deconvolution and the seismics support the validity of our integrated approach to delineate normal faults in the CSZ, a region where seismic profiles are of lesser quality.

3.2.4 Profiles across the CSZ

To define the geological structures along the St. Lawrence River, four profiles were defined in the CSZ, each one with measured and computed Bouguer gravity anomalies, crustal density model, Euler solutions, structural interpretation and one or more nearby seismic profiles (Figure 3.10 to 3.14). Gravity modelling has shown that the Quaternary deposits controlled most of the high frequency variations, masking the signal from the Appalachians. The thickness of the Quaternary was resolved using the seismic

reflection profiles. The longer period variations of the gravity field are related, for the most part, to the depth of the Precambrian-Appalachian interface. Along the north shore of the St. Lawrence River, there appears to be a change in the density of the Precambrian basement, possibly related to the St-Laurent fault. For this reason, the depth of the Precambrian basement there was based on the Euler solutions. The structural interpretations include information from gravity, magnetics, and seismics and display major faults and structures, with approximate fault dips (sub-vertical) and basement depth.

The four models are presented in Figure 3.11-3.14. The Ile-aux-Coudres profile presents many complications, modeling Ile-aux-Coudres being one of them. Near the north shore, there is a sharp gradient in the gravity field, which is interpreted to be due to the combined Quaternary sequence, Ordovician-Appalachian sequence, and possible lower density of the Precambrian basement. Constraints for Ile-aux-Coudres are the depth-to-basement clearly seen on the land profile (Figure 3.11F), and the interpreted step-like structure southeast of the island (Figure 3.11G and H). The island appears to rest on an important normal fault, shown in the Euler solutions. South of the island, the Shield appears to deepen (gravity low with no gradient in the magnetics), while near the south shore, it becomes shallower (Euler solutions and gravity). These suggest the presence of an offshore graben. Under the south shore, there is a large step in the basement, interpreted from the gravity low and the lack of magnetic gradient. The modelled depth of 8 km is similar to that suggested by seismic refraction and reflection profiles (Lyons et al., 1980).

The crater profile presents characteristics similar to the Ile-aux-Coudres profile,

i.e., a central portion with no magnetic gradient, and a possible shallower basement towards the south shore. The gravity data suggest variations in the depth of the Precambrian basement under the St. Lawrence River.

The Line 13 profile appears to match a relatively smooth Appalachian-Precambrian interface for the central portion (Figure 3.13 A-D). This interpretation is supported by the reprocessed seismic line (Figure 3.13 H and I). There is a suggestion of a step fault, a few kilometres offshore from the north shore, that brings the Precambrian to a depth of about 2 km. Again, under the south shore, a large normal fault is interpreted from the gravity and the magnetics.

The Ile-aux-Lièvres profile is simpler than the previous profiles (Figure 3.14 A-D). A series of three steps in the Precambrian basement fits the gravity field and the magnetic gradient.

3.3 Geological faults of the CSZ

Based on the various geoscientific data sets, the main structural aspects of the CSZ can now be outlined. Numerous lineaments exist in the radar images and the geophysical data sets. On land, most faults are described in geological reports (Roy, 1978; Rondot, 1979; 1989), while offshore, most structures are interpreted for the first time. In this section, regional faults are discussed first, followed by the local structures, grouped by sub-areas with similar characteristics: northeast, central and southwest (Figure 3.15). Figure 3.16 shows the adopted fault model for the Precambrian structure of the CSZ.

3.3.1 Regional Faults

The CSZ is cut by the St. Lawrence paleo-rift system, a series of regional normal faults that run parallel to the St. Lawrence River (Kumarapeli, 1978; Rondot, 1979; also called Iapetan normal faults: Wheeler, 1995). Most faults that outcrop on the north shore of the St. Lawrence River, such as the St-Laurent and the Gouffre NW faults are described in geological reports (Rondot, 1979; 1989). Others lie beneath the St. Lawrence River where their existence is revealed by geophysics.

3.3.1.1 The St-Laurent fault

The St-Laurent fault is the most prominent CSZ structure in radar imagery and in the potential field maps. With a length exceeding 150 km and a vertical throw of 2 km (Rondot, 1970), it is a major fault of the St. Lawrence paleo-rift system. Created as an Iapetan fault, the St-Laurent fault was reactivated during the Taconian orogeny (Rondot, 1979). The St-Laurent fault is probably made up of multiple fault segments, as suggested by its jagged course between the Ile d'Orléans and the Saguenay River. The gravity and magnetics suggest three main subdivisions of the St-Laurent fault: a southwest and northeast areas separated by a middle portion centred on the impact structure.

In the central uplift of the impact structure, the St-Laurent fault is oriented N40°E and dips steeply (70°) towards the SE (Rondot, 1979). Here, the St-Laurent fault separates mangerite/charnockite assemblages to the NW from granodioritic to granitic migmatite assemblages to the SE. The lithological change is reflected in the magnetic field, where the generally high values to the NW abruptly change to low values SE of the fault. The

Bouguer gravity field gradually changes from high values to the NW (-30 mgal or more) to low values to the SE (-40 to -60 mgals). On land, the gravimetric low reflects solely the lower density rocks of the Precambrian basement, while offshore, it combines with the low densities of the platform sediments, the Appalachian sequences and the unconsolidated Quaternary sediments. The density change across the fault makes gravity modelling more difficult, as described above for the Ile-aux-Coudres profile.

To the SW of the impact, the St. Laurent fault is parallel to the north shore (oriented N40°E). It may line up with a series of Euler solutions in the magnetics under Ile d'Orléans. There, seismic reflection profiles indicate that, over a 10 km distance, a series of normal faults oriented north-south brings the depth to Precambrian basement from about 2.1 km to about 4.7 km (SOQUIP, pers. comm.).

To the NE of the impact, the St-Laurent fault is not as prominent in the geophysics. The St-Laurent fault never comes ashore there and its position is assumed to follow the north shore from the Gros Cap-à-l'Aigle to Tadoussac (Rondot, 1979). In contrast to the situation within the impact crater, the magnetics do not show any change across the fault. This can be explained by a smaller throw than to the SW or by a much thinner sedimentary cover. The fault may die out NE of the impact.

3.3.1.2 The Gouffre NW fault

The Gouffre NW fault (sometimes called Gouffre) is named after a small creek that flows into the Gouffre River near Baie-St-Paul. To the SW of the crater, this fault is a prominent lineament on the radar images and in the magnetics (Figures 3.17). Within the impact crater and to the northeast, the fault position is not obvious. Geologically, its position inside the impact is based on stratigraphic evidence that suggests that the fault existed at the time of deposition (Rondot, 1968). To the NE, its extension is based on lineaments with similar trends. Compared with the St-Laurent fault, the Gouffre NW fault appears to have a small throw with little or no geophysical signature inside the CSZ. Its position is approximate within the impact and uncertain to the NE.

3.3.1.3 The Charlevoix fault (new feature)

A few kilometres offshore from Ile d'Orléans to Tadoussac, a normal fault parallels the St-Laurent fault. The fault separates a 2-8 km wide Precambrian basement plateau from a deeper basement to the southeast. In the magnetics, the plateau corresponds to high frequency anomalies, while the deeper basement correspond to low frequency anomalies. The fault's gravity signature is less conspicuous due to the proximity of the St-Laurent fault and to the sediment-filled valleys. In the southwestern portion of the fault, the SE shallow-dipping Precambrian-Appalachian plateau is imaged at ≈ 3.0 km depth (Figure 3.18). Under Ile-aux-Coudres, the fault may also correspond to the truncated Precambrian reflector of Figure 3.11E. The position of the Charlevoix fault coincides with the overlying Logan's thrust, suggesting that the footwall may have acted as a buttress, or that

the fault was reactivated by the weight of the advancing nappes.

3.2.1.4 The South Shore fault (new feature)

Along the south shore of the St. Lawrence River, the gravity and the magnetic fields suggest a large normal fault in the Precambrian basement. On all profiles, gravity modelling suggests that the Precambrian depth drops from about 3-4 km to 7-8 km. The latter depth value corresponds to the seismic refraction results (Lyons et al., 1980). In the southern part of the fault, slightly outside the CSZ, the Bouguer anomaly low implies an even deeper basement than to the NE. The gravity suggests that this fault may terminate near Ile-aux-Grues.

3.3.2 Northeast zone

In the NE section of the CSZ, the NS trend of lineaments seen on land continues under the River. While the St-Laurent fault trends mostly NE, in this northern portion of the "Jacques Cartier block", a large proportion of lineaments trend NNE (Roy et al., 1993). Under the St. Lawrence River, a similar trend is found in the Euler solutions (Figure 3.5C). Most of these lineaments appear to relate to Precambrian lithologies.

South of Ile-aux-Lièvres and oriented parallel to the St. Lawrence River axis, a normal fault is suggested by the gravity and by the Euler solutions (Ile-aux-Lièvres fault). Gravity modelling (Figure 3.14A-D) suggests a 2 to 3 km throw for this fault, i.e. comparable to the St-Laurent fault. This is supported by the magnetics and the Euler solutions that get progressively deeper southeast of this fault. The fault runs from the

outer rim of the impact crater, near La Malbaie, and extends beyond the NE boundary of the CSZ.

The Palissades fault (PAL), the southernmost fault of the Saguenay graben, is supposed to correspond to the NE extension of the seismic zone (Anglin, 1984). This structure of regional extent can be seen on ERS-1 imagery to the NW, outside the seismic zone. The Palissades fault may continue offshore, as suggested by the Euler solutions and the gravity.

3.3.3 The impact structure faults

The 55 km diameter Charlevoix astrobleme of Devonian age (≈ 350 Ma; Rondot, 1979) is a "complex" impact structure characterized by a central uplift surrounded by an annular trough and a peripheral down-faulted structurally complex rim (Grieve, 1993). Across the structure, the nature and the extent of the block movements vary from upward movement in the central uplift to subsidence in the rim. The faults directly created by the impact are found near the central uplift, where the shock has been most intense, and where the vertical movement can reach 9 km (Robertson, 1975). This area is made up of a series of blocks separated by normal faults (Roy, 1978). In the rim, structures are a series of locally sub-parallel normal faults of about 11 km depth extent (Rondot, 1994; similar to the Siljan impact structure of Sweden; Juhlin and Pedersen, 1987). Whereas most rim faults are conspicuous in the radar images, inside the Charlevoix impact structure, lineaments are less clear (Figure 3.3D). This is possibly due to the erosion of highly fractured rocks. Beneath the zone affected by the impact, faults without surface

expressions probably exist.

Beneath the St. Lawrence River, geophysical data suggest structures possibly related to the impact. The impact structure deformation is illustrated by the SW part of seismic Line A (Figure 3.19). Moving towards the impact structure, a series of sub-horizontal reflectors is present until the 60 km radius line is reached. There, a small basin, possibly controlled by shallow faults, is found near the theoretical position of the peripheral graben.

The Île-aux-Coudres corresponds to an area of elevated basement, an interpretation supported by seismic reflection profiles and by higher magnetic and gravity values (Figure 3.11A-D). East of Île-aux-Coudres in the centre of the crater, the smooth magnetic field and the low gravity values suggest a deep Precambrian basement, which is interpreted as a graben structure on Figures 3.11A-D and 3.12A-D.

3.3.4 The SW section

To the southwest of Île-aux-Coudres, the gravity and the magnetics suggest a deep basin, probably controlled by a series of normal faults. Under Île-aux-Grues, where a gravimetric low is found, a SOQUIP seismic line suggests a Paleozoic-Precambrian interface at about 5.5 km depth. This basin is controlled by at least two normal faults, one oriented nearly E-W inferred from the seismics and one parallel to the north shore, inferred from the Euler solutions.

3.4 Faults and seismicity of the CSZ

This section examines the relationships between the geological faults described above and the CSZ earthquakes. Earthquake magnitudes and fault dimensions for intraplate earthquakes may be related through proposed scaling laws (Table 3.3). During the period 1977-1997, only eight earthquakes exceeded magnitude m_N 4.0. Thus, most CSZ earthquakes reactivated relatively small fault surfaces, that may not have a surface expression. As described below, regional faults bound highly seismically active blocks. Geographically, the level of activity varies: whereas the highly fractured central zone has frequent, but small magnitude, earthquakes, the SW and NE areas, with their long normal faults, produce infrequent, but larger, events.

Table 3.3 Fault dimensions for stable continental earthquakes of various moment magnitudes (M ; after Johnston, 1993)

M	Length (L)	Width (W)	slip (d)
0.0	5.8 m	5.8 m	1 mm
1.0	18 m	18 m	3 mm
2.0	58 m	58 m	9 mm
3.0	183 m	183 m	2.9 cm
4.0	577 m	577 m	9.1 cm
5.0	1.8 km	1.8 km	29 cm
6.0	8.0 km	4.1 km	91 cm
6.5	15 km	7.0 km	1.6 m
7.0	30 km	11 km	2.9 m

3.4.1 Faults of regional extent

The St-Laurent fault (SL) appears to bound the earthquake distribution, especially in the Les Eboulements-Cap-à-l'Aigle section. There, most shallow ($z < 10$ km) earthquakes locate to the NW of a south-east boundary, which is interpreted to be the sub-vertical SL (Figure 3.20). Below 13 km depth, most earthquakes occur SE of the fault in a central corridor bound by two weakly seismic areas. Near Ile-aux-Coudres, earthquakes are generally shallower than 10 km depth northwest of the fault, whereas to the SE, deeper earthquakes are found (down to 15 km depth). To the NE of Cap-à-l'Aigle, the St-Laurent fault does not appear to affect the earthquake distribution. There, another rift fault, the Rang-Ste-Mathilde fault, separates shallow activity to the west from deeper activity to the east.

To the NE, the Charlevoix Fault (CH) may be currently reactivated by the earthquake activity. The NE area is one of the two zones where magnitude ≥ 4.0 earthquakes in the period 1924-1978 occurred, including the M 6.2 1925 Charlevoix earthquake (Stevens, 1980). Under the river, most earthquakes occur within a river-parallel planar volume which includes most $m_N \geq 4.0$ CSZ events, including the magnitude 5.0 earthquake of 1979, and the recent October 28, 1997 m_N 4.7. A cross-section through the 1978-1997 hypocentres shows a remarkable alignment with a dip of about 50° (Figure 3.21), in agreement with the focal mechanisms and depths (10 km) of the 1925 M 6.2 and the 1979 m_N 5.0 earthquakes (Bent, 1992; Lamontagne, 1997). Inside the impact crater, the Charlevoix Fault may be a boundary to the earthquake activity rather than the active structure. Under Ile-aux-Coudres for example, events northwest of the fault are shallower than to the southeast.

To the SE, and possibly to the NE, regional faults act as boundaries. To the SE, the South Shore fault (SS) delimits the seismicity, without being active itself. To the NW, the Gouffre NW (GNW) fault has been proposed to limit the earthquake activity (Anglin, 1984). This correlation may only be apparent: the position of the Gouffre NW fault is approximate and hypocentres appear independent of a clear and linear NW boundary. Remote sensing images and aeromagnetics also suggest a NE lineament that lines up with a portion of the St-Laurent fault (named Crater Lineament, CRL; Figure 3.17).

3.4.2 The NE portion of the CSZ

In the NE, most earthquakes appear to correlate with faults. As described above, offshore hypocentres line up with the Charlevoix fault. In addition, this area also has multiple lineaments oriented NNE. It is possible that some faults with this orientation are reactivated (such as the 1989 doublet of Wetmiller and Adams, 1990). Anglin (1984) noticed that the "B" fault (Figure 1.4) matches numerous surface-projected hypocentres. This fault is one of many faults, parallel to the St. Lawrence River, seen in ERS-1 imagery. One of these lineaments could be the Rang-Ste-Mathilde fault (RSM) of Rondot (1989). Offshore, the Ile-aux-Lièvres fault (L) is not seismically active as it lies to the southeast of the activity previously linked with the Charlevoix fault. The NE extremity of the CSZ contains mainly deep events (> 20 km). The earthquake activity appears to terminate at the offshore extension of the Palissades fault (PAL).

3.4.3 The impact structure

The relationship between CSZ earthquakes and the impact structure remains

unclear. The majority of CSZ earthquakes locate within the volume of the impact structure defined by the 60 km diameter of the outer rim and the 10 km depth extent of the faults. Two focal mechanisms correspond to the trend of the rim faults (Lamontagne and Ranalli, 1997). This suggests that some impact-related faults are currently reactivated. One of the faults of the astrobleme ("A" fault; Figure 1.4) was thought to cause an offset to the seismicity (Anglin, 1984). Within the impact structure, clouds of earthquakes without any particular alignment are seen (except near St-Joseph-de-la-Rive, where micro-earthquakes define an alignment, which does not correlate with radar and geophysics lineaments). Under the river as well, the highest concentration of activity of the CSZ corresponds to the highly fractured graben, possibly created by the impact. It is possible that these highly fractured areas give rise to numerous low magnitude events with highly variable focal mechanisms. The central uplift of the astrobleme, where fracturing and faulting have been the most intense, however, is not more particularly active.

3.4.4 The SW section

Earthquakes occur very infrequently in this area and their correlation with faults is more difficult to establish. Since the area is outside the seismograph network, hypocentral depths are not as precise as in the rest of the CSZ. The area is important, however, since it is one of the two areas where magnitude ≥ 4.0 events of the period 1924-1978 have occurred (Stevens, 1980). Iapetan faults are known to exist there, with some possibly at high angle to the river axis. With conditions similar to the rest of the CSZ, the lower rate of earthquake occurrences is enigmatic. The normal fault near Ile-aux-Grues, or possibly the impact crater rim, may act as a boundary to the higher level of

activity to the NE.

3.5 Discussion

This analysis has brought to light the main structural patterns of the CSZ. In the NE area, the Precambrian basement is cut by a series of long normal faults parallel and at high angles to the river axis. In the centre, the impact crater controls the faults, at least down to about 10 km depth. Finally, in the SW zone, the Precambrian basement, deep under the Appalachians, is cut by normal faults parallel and at high angles to the river axis. The CSZ is a transition zone between the SW part, with its steeply-dipping faults with large normal throw, and a NE part, with its en echelon normal faults with progressive deepening of the basement under the Appalachians.

In Anglin (1984), CSZ earthquakes were correlated with five geological faults (Figure 1.4): St-Laurent; Gouffre NW; Palissades and faults "A" and "B". These correlations were established on the basis of projections of the hypocentre distribution to the surface along a trend (N52°E) and a dip (70°SE) that provided the tightest cluster of hypocentres. The main assumption was that most earthquakes occur on planar structures that cross the entire seismic zone. After some 20 years of micro-earthquake recording, it is now apparent that earthquakes do not concentrate only on planar structures. Locally, earthquakes define depth-limited volumes of enhanced activity. In this context, most large faults seem to bound these active volumes rather than being active themselves.

Three Iapetan faults bound seismically active sub-zones: the St-Laurent, South Shore and Charlevoix faults. These faults separate blocks possibly with different fracturing

levels. The foot-wall of the St-Laurent fault, for example, has most of the micro-seismic activity, whereas the hanging-wall has weaker and deeper activity. Conceptually, the foot-wall may be highly fractured, and/or subject to high pore-fluid pressures. Iapetan faults do not seem particularly active (for micro-earthquake activity): hypocentres do not concentrate along the St-Laurent and the South Shore faults. The best correlation between an Iapetan fault and earthquakes is found in the NE part of the zone. There, hypocentres define a narrow steeply-dipping volume, possibly the Charlevoix fault. Inside this volume, earthquakes may or may not occur directly on the main fault, as shown by the variability in the focal mechanisms.

Contrasting with its conspicuous surface expression, the SE-dipping St-Laurent fault itself is not particularly active. However, it bounds the earthquake activity over part of the CSZ. In the central part, most of the deep (> 20 km) events occur to the east of the St. Laurent fault. Our analysis suggests that the St-Laurent fault acts more like a boundary to the activity than as an active structure. Similarly, the South Shore fault bounds the seismicity to the SE. To the NW, a fault-controlled boundary is not as clear.

The relation between the Charlevoix impact structure and the current earthquake activity has been subject to much debate. On the one hand, those who disclaim any relationship observe that earthquakes do not distribute over the whole impact structure, but concentrate in zones parallel to the St. Lawrence paleo-rift faults (Adams and Basham, 1989). In addition, most impact structures are currently aseismic (Salomon and Duxbury, 1987). According to others, however, the structure might contribute to the inherent weakness of the region (Leblanc and Buchbinder, 1977; Lamontagne, 1987; Kumarapeli,

1987). "Impacting has undoubtedly complicated the pattern of potential failure in the area and consequently may have created an environment for stress concentration in an otherwise throughgoing Iapetan fault system" (Kumarapeli, 1987). At present, it is unclear why earthquakes do not occur over the whole impact structure. Instead, they appear to concentrate in two bands elongated along the St. Lawrence. Some earthquakes, however, appear to occur on faults, possibly created as Iapetan faults, but clearly reactivated by the impact (Lamontagne and Ranalli, 1997). Interestingly, most of the large events tend to concentrate at both ends of the CSZ, outside the impact structure. Within the impact structure, the highly fractured Grenville releases strain energy in small earthquakes only ($m_{bLg} \leq 3.5$), with highly variable fault plane orientations.

Figure 3.1 Location map of the CSZ surroundings and geographic names mentioned in the text.

Figure 3.2 Principle of the side-looking radar system (Radarsat in this case). Each beam is defined by the area covered and the level of detail (resolution) available)

Figure 3.3 Remote sensing radar images of the CSZ. On these various radar images, regional and local lineaments can be recognized: (A) Radarsat image of the St. Lawrence River between Quebec City and Tadoussac that shows the numerous lineaments in the Precambrian Shield, north of the river (courtesy of the Ministère des ressources naturelles du Québec). (B) JERS-1 image of the area to the west of the CSZ; (C) ERS-1 image of the CSZ; (D) Airborne SAR with West-facing direction of the beam; (E) Airborne SAR with South-facing direction of the beam; (F) Enlarged view of E that shows the high resolution of the airborne-radar system. (G) Chromo-stereoscopic image that integrates a Radarsat image (texture) and topographic information (colours). Courtesy of Thierry Toutin, Canada Centre for Remote Sensing.

B

25 0 25 50 Kilometers

C

25 0 25 50 Kilometers

D

25 0 25 50 Kilometers

E

25 0 25 50 Kilometers

F

13 0 13 26 Kilometers

G

Figure 3.4 CSZ total magnetic field (colours) with shorelines (black).

Figure 3.5 Solutions of the Euler deconvolution ($N = 0$) of the magnetic field. The circle sizes correspond to the relative depth of the solutions (shallow solutions are small); (A) All solutions; (B) Solutions between 0 and 500 m depth; (C) Between 0 and 1 km depth; (D) Between 1 to 2 km depth; (E) Between 2 to 3 km depth; (F) Between 3 to 4 km depth.

D

E

Figure 3.6 Photos taken during the August 1994 gravity survey: the LaCoste and Romberg dynamic gravity meter, with the portable computer and the GPS receiver (right), and the launch used on the St. Lawrence River (below)

Figure 3.7 Gravity and density data. (A) The gravity stations in the CSZ (crosses) superimposed on the Bouguer anomaly map; (B) The Bouguer anomaly map produced with standard contouring schemes; (C) Distribution of density values for the three main lithological groups of the CSZ.

B

Figure 3.8 Location map of the SOQUIP seismic reflection lines in the CSZ.

Figure 3.9 Euler solutions of the magnetic field in the Quebec City region (see description of solutions in Figure 3.5). The shore of the St. Lawrence River are shown as black lines. The epicentre of the November 1997 M 5.0 earthquake (star) lies on a NE-SW lineament (A). The Neuville fault (N), a paleo-rift fault dipping towards the SE may be the structure reactivated during the earthquake. Other magnetic lineaments (dashed blue lines) correspond to faults in the Precambrian basement evidenced in the SOQUIP seismic profiles.

Figure 3.10 Location of the four geophysical profiles discussed in text (red dotted lines) and presented in Figures 3.11 to 3.14. The yellow lines are the SOQUIP seismic profiles shown in the following figures.

Figure 3.11 Geophysical profile near Ile-aux-Coudres: (A) Bouguer anomaly (circles) and modelled field (red line); (B) Gravity model and density contrasts from the Precambrian; (C) Euler solutions within 5 km of the profile, plus locations of the seismic lines; (D) Structural model based on geophysical information; (E) Structural model with hypocentres located within 5 km on both sides of the profile; (F) Seismic profile on Ile-aux-Coudres with the red arrow indicating the surface of the Precambrian (SOQUIP, pers. comm.); (G) Seismic profile 140 on the St. Lawrence River; (H) Interpretation of the seismic profile.

F

H

NW

SE

Figure 3.12 Crater profile (A-E), with legend similar to 3.11 A-E; (F) Seismic line 30 over the St. Lawrence River (top) and interpretation (below). The interpreted Quaternary sequence is evident, while deeper reflectors are not continuous.

F

Figure 3.13 Line 13 profile (A-E) with legend similar to 3.11A-E; (F) Original SOQUIP profile with arrows showing interpreted reflectors in the Precambrian shield; (G) Reprocessed Line 13; (H) Interpretation of G. The Appalachian sequence (yellow), possibly cut across by two faults (red) appears to have a sub-horizontal interface with the Precambrian. A number of reflectors can be seen in the Precambrian. The attitude of these reflectors changes with depth. A normal fault may exist to the NE of the line (right side).

Seismic Line 13

A

B

C

D

G

Figure 3.14 Ile-aux-Lievres profile (A-E), with legend similar to 3.11 A-E; (F) Seismic line 35 (top) with interpretation (below).

Figure 3.15 The three main sub-regions discussed in text (NE, Impact crater (Central) and SW). Geological features presented in Figure 1.7 are also shown.

Figure 3.16 Interpreted structures of the CSZ. Acronyms used: PAL: Palissades fault; RSM: Rang Sainte-Mathilde fault; SL: Saint-Laurent fault; CH: Charlevoix fault; L: Lievres fault; SS South shore fault; G: postulated graben under the St. Lawrence River; CRL: crater fault; GNW: Gouffre NW fault.

Figure 3.17 Major faults of the CSZ, mapped (red), interpreted (dashed blue), and lineaments (dashed yellow) in the Radar chromo-stereoscopic image (courtesy of T. Toutin, Canada Centre for Remote Sensing).

Figure 3.18 SOQUIP seismic line 22. The position of the Precambrian-Appalachian interface (red arrow) is at about 3 km depth (SOQUIP, pers. comm.). The syncline to the left corresponds approximately with the border of the impact structure.

Figure 3.19 SOQUIP seismic line A, south of Ile-aux-Coudres.

Figure 3.20 Geological faults and CSZ hypocentres for the period October 1977 to December 1997.

Hypothetical
Rupture Plane

Figure 3.21 Vertical section across the 1979 M5 earthquake zone. Earthquakes are within 10 km of a line across the trend of the St. Lawrence River, and centred on the M5 event of 1979. Pure reverse-faulting on the best hypocentre alignment is shown.

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Chapter 4

Seismicity patterns from focal mechanisms, earthquake clusters, multiplet analysis, and the search for surface fault ruptures

4.1 Introduction

In the CSZ, most earthquakes are assumed to be spatially related to preexisting faults of the St. Lawrence Paleozoic rift system (Adams and Basham, 1989). This relationship is based on a projection of the hypocentres to the surface up the dip of the regional faults. When projected, the earthquakes are located between a paleo-rift fault and a possible fault-controlled bathymetric feature. In support of the paleo-rift model, some focal mechanisms have one nodal plane oriented similarly to the paleo-rift faults (Adams and Basham, 1989). It was recognized, however, that the highly variable focal mechanisms are not easy to interpret, possibly due to the complications introduced by the impact structure.

To evaluate the CSZ earthquake-fault connection, four sources of information are examined: possible surface ruptures, focal mechanisms, hypocentre alignments and earthquake multiplets. The best evidence of the earthquake-fault connection would be to find a surface rupture caused by some of the numerous $M > 6$ CSZ earthquakes. Section 4.2 describes the search for such a rupture under the St. Lawrence River. Focal mechanisms are a representation of the earthquake rupture process and, as such, constitute an essential source of seismotectonic information (Section 4.3). They provide the strikes

and dips of two nodal planes (one of which being the rupture plane), the trends and plunges of the P, T and B axes, and the nature of faulting that produced the earthquake. Fault zones can also be located from hypocentre groups and alignments (Section 4.4). Another source of information on fault orientations is provided by earthquake multiplets, which represent earthquakes with similar locations and focal mechanisms (Section 4.5).

In the CSZ, hypocentres are not distributed randomly. There are areas where earthquakes are shallow (< 10 km depth), intermediate (10-15 km), deep (> 15 km), or dispersed. To better define the local factors that determine the hypocentre distribution, earthquakes are divided into a series of sub-zones in Section 4.6. The seismotectonics of these sub-zones are examined using the available focal mechanisms, earthquake groups and multiplets. Based on the findings, regional seismicity aspects of the CSZ are discussed in Section 4.7. Finally, in Section 4.8, recommendations for future work are presented.

4.2 The search for surface fault ruptures

The best possible correlation between the CSZ earthquakes and the regional faults would be a surface rupture caused by one or by a series of magnitude > 6 earthquakes. In the CSZ, at least five earthquakes of this size are known to have occurred since the XVIIth century, and numerous prehistorical earthquakes are suggested by local landslides and lake disturbances (Filion et al., 1991; Doig, 1986). Despite these large earthquakes, no surface rupture has ever been reported in the geological maps (Rondot, 1979; 1989), in the historical accounts, or in the scientific reports on the 1925 earthquake (Hodgson, 1925; 1950). Based on the source parameters for a moment magnitude (**M**) 6.0 (Table 3.3)

and the focal depth (10 km; Bent, 1992), the 1925 earthquake is unlikely to have ruptured the surface. However, the recurrence of similar or larger events over thousands of years may have led to a surface rupture.

It can be argued that the likelihood of a surface rupture is greater under the St. Lawrence River where most magnitude > 4 earthquakes of the XXth Century occurred (Stevens, 1980; Chapter 1). Under the river, however, thick sequences of unconsolidated Quaternary deposits cover the basement. These sequences represent multiple glacial and inter-glacial periods covering more than 250,000 years (Occhietti et al., 1997). The seismic lines by the Société Québécoise d'exploration pétrolière (SOQUIP) suggest that some river-parallel valleys include up to 600 m of sediments. At present, the complete sequence has not been drilled and, consequently, the Quaternary age of the basal units can only be assumed. A model was defined for a surface rupture caused by a thrust earthquake occurring in the Precambrian basement (Figure 4.1). A rupture from the bedrock causes offsets or truncations of the basement-sediment interface, propagates into the sedimentary sequences, giving displaced, or truncated sub-horizontal horizons on seismic sections. A bathymetric feature can be evident in the depth soundings, such as those done during the 1994 gravity survey. To consider this model, a series of factors were defined (Table 4.1).

The SOQUIP seismic profiles, acquired in the early 70's (described in Chapter 3) can be used to infer the position of sub-river ruptures. Most profiles show continuous horizontal reflectors in the Quaternary deposits, in the underlying Precambrian basement and in the Appalachians (Figure 4.2). It is assumed that a vertical throw of 20 m (0.02 s offset) through the sediments would be evident on most SOQUIP lines. This corresponds

to the accumulated slip of seven shallow $M \geq 6.0$ earthquakes. In addition, a series of shallow high resolution seismic profiles were acquired jointly by the Université du Québec à Rimouski (UQAR) and the Université du Québec à Montréal (UQAM), with partial financial support by the GSC. More than 200 km of shallow seismic reflection profiles were acquired on the St. Lawrence and Saguenay rivers during the summers of 1995 to 1998. Up to 800 ms of analogue data were acquired (corresponding to about 800 m of penetration with a P-velocity of 2 km/s). The resolution power of the method is very high, and provides images of sedimentary facies within the Quaternary sequences.

Ruptures with meter-size slips should be seen. As of May 1998, the careful examination of the shallow seismic lines is ongoing at the UQAR (B. Long, pers. comm.). One of these lines (line 13) is shown in Figure 4.4B and is discussed below.

On most SOQUIP seismic sections, one or more continuous sub-horizontal reflectors is evident in the Quaternary sequences, in the basement or in the interpreted interface between the two (Figure 4.2). On some profiles of lesser quality, a few small, although ambiguous, offsets were seen southwest of the Ile-aux-Coudres and within the impact structure. Elsewhere, no unambiguous ruptures were found on any profile, and most areas have clear uninterrupted markers (Figure 4.3). The NE end of the CSZ, where most magnitude > 4 earthquakes concentrate, shows no evidence of a rupture on Lines 13, 35 and 37, including in the thick Quaternary sequence near the north shore (Figure 4.4A).

Table 4.1 Factors defining the nature of the potential ruptures

Nature of discontinuity	Quaternary	Precambrian Basement	Appalachian Basement	Interface Quaternary- Basement	Bathymetric Feature
Neotectonic Fault Rupture	Yes	Yes	Possible if SE of Logan's Line.	Yes	Possible
Syn-sedimentary Movement	Yes	No	No	No	Possible
Ancient and Inactive Fault	No	Possible	Possible	No	No

Yes: Should rupture the given layer.

No: Should not affect the given layer.

Similarly, at the SW end, where some magnitude > 4 earthquakes cluster, shallow seismic profiles do not show any abrupt disruptions that could be interpreted as a rupture. The seismogenic fault suggested by correlating surface-projected hypocentres and the six fathom isobath (Anglin, 1984), was not evident on the section that crosses the area (Figure 4.4A). The high resolution seismic line did not reveal any rupture. A series of reflectors can be seen near the North Shore, but none appear to be disrupted by a fault. Except for what appears to be a slump, most deformation appears to be related to compaction.

The analysis of some 200 km of SOQUIP seismic reflection lines did not reveal any unambiguous >20 m (≈ 20 ms) discontinuity (possibly a surface rupture), under the St. Lawrence River. High resolution seismic profiles support this finding. This implies that the offshore CSZ was probably not the site of a series of $M \geq 6$ shallow earthquakes in the Quaternary. These large earthquakes would have almost certainly produced a surface rupture detectable on the seismic profiles. Consequently, the CSZ earthquake activity may be relatively young, perhaps a few thousands of years old instead of hundreds of thousands of years. It is possible that the larger earthquakes occur at > 10 km depth, with the near-surface displacement being dissipated on a series of sub-parallel faults. The six-fathom bathymetry line, supposedly making a SE boundary to the seismicity (Anglin, 1984), is not linked with a deep seismogenic fault. Hence, the reference to “a possibly fault controlled fault evidenced in the bathymetry” (Anglin, 1984), should be dropped.

4.3 Focal mechanisms of CSZ earthquakes

4.3.1 Introduction

Traditionally, focal mechanisms of micro-earthquakes have been determined with P-wave first motions, due to the availability of the data (most stations are vertical component only) and to the simplicity of the method. Unfortunately for CSZ studies, events below magnitude m_N 4.0 rarely provide enough P-wave first motions for a well-constrained mechanism. With the present rate of $m_N \geq 4.0$ events (seven during the 1977-1997 period), decades would be needed to obtain a comprehensive focal mechanism database. To compute focal mechanisms of smaller earthquakes, two approaches are possible: adding S wave information and/or increasing the number of seismograph stations. Both solutions were attempted in this study. First, the information contained in the S wave was added. S_H first motions were preferred to S_v/P ratios, which are usually difficult to assess due to the imprecise onset of the S phases. S_H first motions are generally better defined and only require the conversion of the horizontal traces into the transverse component of the S wave. This rotation can be easily applied to the three-component Charlevoix data. Using P and S_H first motions, CSZ focal mechanisms can be computed for events larger than m_N 3.0. The second approach is to increase the number of seismograph stations, which was done during the 1996 summer field survey (Lamontagne et al., 1997). The following sections review the methodology and the results of these studies. Some re-computed solutions of previously published focal mechanisms are also given. More details on these aspects, including the full descriptions of the mechanisms, can be found in Lamontagne (1998).

4.3.2 Input data and methodology

In the CSZ, a focal mechanism can be determined if the event is sufficiently large (i.e. yields enough first-motion observations) and if three component data are available (i.e. to determine S_H first motions). Consequently, only events of magnitude $m_N \geq 3.0$ recorded on the three-component CSZ network were selected (between November 1988 and April 1997). Earthquake locations and focal mechanisms were calculated primarily with the data of the CSZ permanent network (LMQ; A11; A16; A21; A54; A61; A64; Figure 1.8). When available, first motions were added from five additional stations located within 150 km of the active zone (DAQ; SHQ; SLQ; CIQ; QCQ; Figures 1.11). During the period from mid-June to mid-November 1996, smaller events ($m_N \geq 2.0$) were also considered, with data from up to eight additional analogue and digital field seismographs (Lamontagne et al., 1997; Figure 1.12).

All hypocentre locations of this study were routinely computed with CLTN phases only, except between June to November 1996, when additional phases from field stations were added. Earthquakes were located using the "standard" GSC velocity model, which assumes a 36 km thick crust with 6.2 km/s Pg velocity and 3.57 km/s Sg velocity. The lateral velocity contrast introduced by the Appalachian nappes does not change significantly the takeoff angles and azimuths for south shore stations (Lamontagne, 1987).

P-first motions on the digital and analogue records were mostly read directly by the author. S_H first motions were read on rotated traces from the three-component stations. In general, S_H first motions were less consistent than P-first motions. For example, two

focal mechanisms had sufficient P first motions to give a reliable solution (events 960714 18:46 and 960924 23:41, both recorded during the 1996 summer field survey). In the first case, three S_H out of seven did not fit the solution, while two out of nine did not fit the second solution. From these results, it was decided that the P first motions were the primary constraints to the focal mechanisms, while S_H would further refine the solutions.

The mechanisms were determined with the program FOCMEC of Snoke et al. (1984). The program uses a grid search algorithm that finds all mechanisms that match a given number of P and S_H first motion misfits. The program was run interactively, using a one degree B-axis increment. In cases where too many solutions existed, two degree, then, five degree increments were used. The variations in the P, T and B axes were used to quantify the constraints on the nodal planes. The method was inspired by the work of Moustafa (1992). The P, T and B axes were treated as vectors for which average direction, plunge and length were computed. The average length is a measure of the dispersion of the axes, i.e. the best solutions have lengths close to 1. A quality factor was assigned to the solutions using the scheme shown in Table 4.2. The rating scheme agreed with a visual inspection of the focal mechanisms. In a few cases, the P, T and B axes were well constrained but the solutions were rejected (quality "X") based on the poor distribution of first motions. For every event, the description of the mechanism and three figures were provided: the focal mechanisms, the first motion picks and the epicentral map (Lamontagne, 1998; Appendix 1).

Table 4.2 Rating scheme of the focal mechanism

Quality	Considered as:	Average length P, T, B	Corresponding variation in degrees around average position	Maximum P errors (%)	Maximum S _H errors (%)
A	Reliable	> 0.994	+ - 10	<10	<50
B	Reliable	> 0.988	+ - 15	<25	<75
C	Weakly constrained	> 0.979	+ - 20	<50	100
X	Rejected	<= 0.979	> + - 20.	>50	100
O	Cannot be computed	-----	-----	-----	-----

4.3.3 Results

Some 20 new focal mechanisms were computed for the period November 1988 to December 1997. Out of the 27 magnitude m_N 3.0 events of that time period, 12 resulted in quality A and B mechanisms (Figure 4.5; Table 4.3). The addition of up to eight seismograph stations during the 1996 Summer field survey, provided data for 15 additional mechanisms, for earthquakes as small as m_N 2.0. Of these, eight were of quality A and B (Figure 4.6; Table 4.4). While most mechanisms were calculated for the first time, some had been calculated by other authors (referenced in Table 4.5) and are discussed below. Since Lamontagne (1998), a focal mechanism was computed for the 19971028 m_N 4.7 event (Appendix 1). This event was the largest CSZ earthquake since the 1979 m_N 5.0 earthquake. Although the oblique-slip mechanism is fairly well defined, it was quality C due to the poor constraints on the P, T, and B axes.

Some previously published focal mechanisms were re-examined. The six mechanisms of Leblanc and Buchbinder (1977) were re-computed using the original data. Two of the six mechanisms are well constrained (74/06/09 and 74/06/23; Figure 4.7), and similar to the solutions of Leblanc and Buchbinder (1977) and Adams et al. (1989). Two mechanisms have two main families of solutions (74/06/20 and 74/06/30) and the last two are very poorly constrained (74/07/02 and 74/07/13).

Due to its seismotectonic importance, the focal mechanism of the 19 August 1979, m_{blg} 5.0 earthquake was recomputed using the program FOCMEC. Since the original data set was lost, first motion data were assembled from GSC analogue and digital playouts,

Table 4.3: Focal mechanisms of earthquakes of magnitude 3.0 and larger

Q	Date	Time	Latitude	Long.	Depth	Mag	Sta/Pha	References:
	yyymmdd	hhmm ss	(°N)	(°W)	(km)			
A	890131	1439 48.20	47.4426	-70.6710	19.69	3.1m _N	7/012	
X	890309	0941 32.26	47.7171	-69.8569	10.52	4.3m _N	6/011	1 2 3
X	890311	0831 52.16	47.7182	-69.8699	10.41	4.4m _N	7/012	1 2 3
B	891013	1404 42.80	47.3926	-70.1330	22.74	3.2m _N	6/012	
A	891122	2302 51.72	47.4559	-70.3420	7.39	3.4m _N	7/014	
A	900303	0206 03.38	47.8559	-69.9765	20.85	3.6m _N	8/015	4
A	900313	1910 39.34	47.5338	-70.1366	15.38	3.2m _N	7/014	4
C	900421	0123 04.12	47.5532	-70.0698	9.56	3.1m _N	7/014	5
X	900423	0028 04.78	47.4143	-70.1787	8.04	3.0m _N	7/014	
A	901021	1338 43.20	47.3975	-70.3644	15.85	3.3m _N	7/013	5
X	901026	0913 51.51	47.5692	-69.9848	10.96	3.1m _N	6/012	
X	901106	1130 10.76	47.3943	-70.1506	14.19	3.4m _N	6/011	6
X	901218	0710 46.23	47.2627	-70.3359	9.38	3.3m _N	7/013	6
O	910703	0926 42.32	47.5290	-70.1464	18.44	3.0m _N	CLTN data not saved.	
B	911208	0300 30.12	47.7792	-69.8643	23.05	4.3m _N	7/013	
X	920310	0545 32.64	47.7167	-69.8574	9.96	3.3m _N	7/013	3
X	920501	0037 51.49	47.4463	-70.4069	2.67	3.2m _N	6/011	
A	930304	2202 21.84	47.5145	-70.3621	4.39	3.1m _N	6/011	
X	930807	2125 31.92	47.6681	-69.8893	7.75	3.1m _N	7/014	
O	931201	1247 15.89	47.4671	-70.1584	18.0g	3.5m _N	No CLTN data.	
X	931230	2301 47.69	47.4532	-70.3609	5.94	3.8m _N	6/012	
A	940925	0053 29.46	47.7518	-69.9612	12.18	4.3m _N	6/012	
B	941201	1302 47.14	47.4374	-70.3138	10.77	3.0m _N	5/009	
C	960512	1153 21.91	47.5161	-70.0281	14.82	3.1m _N	7/014	

Q	Date	Time	Latitude	Long.	Depth	Mag	Sta/Pha	References:
	yyymmdd	hhmm ss	(°N)	(°W)	(km)			

X	960607	0941	42.83	47.5299	-69.9417	13.32	3.1m _N	7/014
A	970110	1927	27.56	47.5094	-70.1965	17.06	3.2m _N	7/014
X	970114	0447	32.39	47.6574	-69.8765	14.96	3.1m _N	7/014

- 1- Drysdale et al. (1989);
- 2- Wetmiller and Adams (1990).
- 3- Li et al. (1995).
- 4- Drysdale et al. (1990);
- 5- Drysdale et al. (1991a).
- 6- Drysdale et al. (1991b).

Q: Quality.

A: Very good

B: Good

C: Fair

X: Rejected

O: Cannot be computed.

Total: 8 A; 2 B; 3 C; 12 X; 2 O

Table 4.4: Focal mechanisms of earthquakes recorded during the 1996 summer field survey

Q	Date	Time	Latitude	Longitude	Depth	Mag	Sta/Pha
	yyymmdd	hhmm ss	(°N)	(°W)	(km)		
A	960617	1118 30.66	47.5328	-70.1463	14.11	1.9m _N	8/016
X	960624	1311 16.55	47.4862	-70.1239	14.36	1.6m _N	7/014
X	960704	1227 07.48	47.6197	-70.1352	5.28	2.0m _N	7/014
A	960714	0715 02.89	47.4829	-70.0503	13.71	2.2m _N	7/014
A	960714	1846 49.22	47.6938	-69.9927	7.26	3.3m _N	7/013
C	960726	1438 45.14	47.6097	-69.9468	10.36	2.3m _N	7/014
X	960819	1706 09.66	47.3002	-70.2448	6.44	2.1m _N	11/022
C	960913	2355 35.92	47.5008	-70.2134	12.09	2.2m _N	11/022
B	960923	0526 54.22	47.6591	-69.8932	14.04	2.2m _N	12/024
A	960924	0644 45.55	47.5875	-70.1451	21.40	2.0m _N	12/024
A	960924	2341 02.88	47.5475	-70.2417	12.83	3.1m _N	11/021
C	960925	0834 24.87	47.8528	-69.7504	22.35	2.2m _N	9/018
C	961011	0228 50.34	47.4761	-70.0551	15.56	1.9m _N	11/022
A	961025	0947 24.43	47.4281	-70.3887	4.06	2.2m _N	13/025
A	961028	0245 39.23	47.5536	-70.0414	11.56	2.3m _N	12/024

Q: Quality.

A: Very good

B: Good

C: Fair

X: Rejected / 0: Cannot be computed.

Total: 7 A; 1 B; 4 C; 3 X.

Table 4.5 Other mechanisms. (1974; 1979 + events near La Malbaie)

Q	Date	Time	Latitude	Longitude	Depth	Mag	Sta/Pha
	yymmdd	hhmm ss	(°N)	(°W)	(km)		
B	19740609	2324	30.82	47.3431	-70.2393	10.12	0.6ML 16 / 27
X	19740620	1836	57.48	47.4041	-70.1802	17.30	1.5ML 18 / 22
A	19740623	1406	57.40	47.5127	-70.2144	14.95	0.4ML 16 / 28
X	19740630	1655	11.21	47.7155	-69.8409	15.54	2.0ML 20 / 27
X	19740702	0730	18.62	47.5641	-70.2271	4.43	0.2ML 13 / 22
X	19740714	0029	56.78	47.4919	-69.9720	12.71	0.5ML 13 / 24
B	19790819	2249	30.60	47.6720	-69.9010	10.00	5.0MN 8 / 13
X	19891208	1720	34.44	47.7009	-70.0644	10.37	2.6MN 6 / 12
C	19910723	0103	14.31	47.6869	-70.1031	11.31	1.9MN 7 / 12
B	19930330	0215	18.48	47.6844	-70.1040	4.36	1.8MN 7 / 14

Q: Quality.

A: Very good

B: Good

C: Fair

X: Rejected

0: Cannot be computed.

from paper playouts of Lamont-Doherty and Weston Observatory and from the ISC Bulletin for regional and teleseismic phases. A total of 45 P-first motions was gathered, slightly more than the 39 used in Hasegawa and Wetmiller (1980). Although both the original and the new solution show predominantly reverse faulting (Figure 4.8), the positions of the nodal planes are somewhat different. The new solution shows one plane steeply dipping to the west and one plane more shallowly dipping towards the southeast.

Some focal mechanisms of $m_N \geq 3.0$ events had already been published in the Canadian Earthquake Summaries (Drysdales et al., 1989; 1990; 1991a; 1991b). Four of the new mechanisms are similar to those previously calculated (890309; 890311; 900311; 900313) while four others are significantly different. The original solutions relied on P first motions and on Sv/P ratios. The differences arose due to the use of emergent P first motions at regional distances in the original solutions.

The four mechanisms of Li et al. (1995) were re-computed using P and S_H first motions. In Li et al. (1995), the plotting of the first motions for stations within 150 km was wrong. For direct P and S phases, the azimuths were not reversed, which imply that the mechanisms are questionable especially for the 901021 and 920310 solutions that depend heavily on the CLTN data. Three of the four mechanisms calculated by Li et al. (1995) did not meet our quality criteria and were rejected. Differences were mainly due to our conservative picking of the S_H first motions. Our quality "A" mechanism is sensibly different due to the differences in plotting the CLTN data.

For a sub-zone near station A61, Lamontagne and Ranalli (1997) published three

focal mechanisms for three small events (magnitude m_N 2.6: 891208; m_N 1.9: 910703 and m_N 2.6: 930330). While the general tectonic style is well defined for the three mechanisms, only the 930330 event is rated B.

4.3.4 Discussion of the focal mechanisms

With the current CSZ seismograph network, earthquake focal mechanisms can be computed for events as small as magnitude m_N 3.0, when additional S_H constraints are added. Approximately 50% of the mechanisms in that magnitude range can be considered reliable. The eight additional seismograph stations of the 1996 summer field survey added much needed P and S_H first motions. They allowed the calculation of focal mechanisms for events as small as magnitude m_N 2.0. Some published mechanisms were found to be poorly defined and should be discarded (Figure 4.9). More important, a total of 25 well-defined mechanisms is now available for the CSZ (Figure 4.10).

The faulting style that emerges from the focal mechanism study is mainly thrust to oblique-thrust faulting on fault planes with highly variable orientations. Out of 22 mechanisms of quality “A” and “B”, none was of a pure normal-faulting type. The nodal planes do not have any specific trend, only a weak concentration in the NE quadrant (Figure 4.11A). The stress system appears to consist of a shallow-dipping P axis with heterogeneous orientation (Figure 4.11B).

4.4 Hypocentre groups and alignments

4.4.1 Hypocentre groups

A series of earthquakes can reactivate various portions of a given fault. If sufficiently numerous, well-located hypocentres of these earthquakes can define the strike and the dip of the reactivated fault. Various methods have been proposed to compare hypocentre alignments with faults (e.g., Rieken and Thiessen, 1992; Jones and Stewart, 1997; Chapman et al., 1998). For this study, a simple method to define hypocentre alignments was developed. Given a group of hypocentres, each event is associated with its neighbour, located within a given distance. These associated hypocentres are grouped into strings of events (Figure 4.12). A string can be defined as a volume within which each event has at least one neighbour within the given radius. The analysis can define earthquake clusters, alignments, and concentration of activity.

As a first step, areas that include most CSZ events were defined. A data set that insures location completeness was used, i.e. events with $M_L \geq 0.2$ and $m_N \geq 1.5$ (as discussed in Section 1.7). The data set included some 1205 hypocentres of the period October 1977 to December 1997. To examine associations between events, a maximum inter-event distance of 2 km was selected, slightly larger than the uncertainty in relative hypocentre positions. Some 918 events (76%) have at least one neighbour within 2 km hypocentre distance, while 287 (24%) occurred as “isolated” events. Eleven groups include ten events or more, representing 54% of all CSZ events (Figure 4.13). Out of the eight magnitude ≥ 4.0 events of the 1977-1997 period, two occurred in the most active

string of events, one was within a 15-event string; three had between two and six events; and one was isolated. Except for the region surrounding the m_N 5.0 hypocentre, most magnitude ≥ 4.0 events occur in weakly active areas. About 30% of all CSZ events regroup into one group of hypocentres (Figure 4.14). This group of events defines a faulted volume under the St. Lawrence River with a length of about 40 km, a width that varies between 2 and 11 km, at depth between 6 and 18 km. Variations in strike and dip occur along the length of the fault zone: whereas it is well defined near the magnitude 5.0 earthquake (steeply dipping towards the SE; Figure 4.14D), it becomes progressively shallowly dipping (Figure 4.14C), and diffuse in the middle of the impact crater (Figure 4.14 A-B). Inside the group, various smaller alignments appear to exist. Other groups of events define a series of alignments either at an angle with the strike of the St. Lawrence River or as isolated clusters.

In order to delineate more detailed hypocentre concentration, a similar test was conducted for all CSZ hypocentres of the period 1977-1997, recorded on 5 stations with 8 phases, with a shorter inter-event distance (1.5 km). Out of some 1643 events, 517 (31%) had no associated neighbours. About 38% occurred in groups of 10 events or more. Compared with the 2 km groups, smaller groups become evident and display a variety of orientations.

4.4.2 Discussion

Over a 20 year period, approximately one-fourth of the CSZ events with $m_N \geq 1.5$ occurred as isolated events, i.e., without neighbouring events within a 2 km radius.

We can infer that stresses take a long time to rebuild along some faults and fractures. On the other hand, under the St. Lawrence River, one-third of all CSZ events, including some magnitude ≥ 4.0 events, occurred within a faulted volume of 50 km length, 10 km width, located between 5 and 20 km depth. Although the faulted zone is parallel to the St. Lawrence River, earthquakes define multiple orientations, especially towards the centre of the CSZ. Earthquakes surrounding the magnitude 5.0 event are discussed in detail in Section 4.6.3.

Grouped hypocentres can define a volume, which may be a fault or a weak zone where failure conditions exist. The orientation of these volumes can be compared with the focal mechanisms. Focal mechanisms of events within groups of 2 or more events are shown in Figure 4.15 (groups defined with a 1.5 km minimum inter-event distance). For most focal mechanisms, it is difficult to pick the best nodal plane by looking at the groups of events, either due to their clustered nature, to the small number of events or to the similar orientation of the two nodal planes of any given mechanism. In a few cases, the orientation of the group reflects one of the two nodal planes, providing a way to infer the actual fault trend. In other cases, however, the orientation of the whole group does not correspond to one fault plane, implying that the group defines a weak volume.

4.5 Earthquake multiplets

4.5.1 Introduction

In many seismic areas of the world, earthquakes occur in spatial clusters. Events

within these clusters (also called families or multiplets) display highly-correlated seismic traces, caused by repeated slip on the same fault planes. The identification of such clusters represents a real advance in seismology: what appeared as clouds of events can become hypocentres a few hundreds of meters apart (see, among many others, Deichmann and Garcia-Fernandez, 1992; Nadeau et al., 1994 and, for some Charlevoix multiplets, Li et al., 1995). Due to their spatial closeness, the hypocentres of a multiplet can be used to derive the source parameters of the main shock (the effects of the path, site and instrument response being removed; see e.g. Li et al., 1995) and to infer the orientation of the reactivated fault via precise hypocentre locations and possible comparisons with the nodal planes.

Earthquake multiplets were sought among the 915 hypocentres recorded with at least 8 phases from 5 seismograph stations between November 1988 and August 1997 (the data format prior to November 1988 is not currently readable). All existing time series were extracted from the archives of the GSC (a few files were missing before the beginning of continuous archiving, in August 1994). All time series files, originally in GSC MKII or SEED formats, were reformatted in SAC format. The original time series files and the SAC files now exist on one 1.2 Gbyte optical disk.

All available seismic traces were systematically cross-correlated to find earthquake multiplets. The procedure consisted of cross-correlating each of the three-component seismic traces for two events and calculating the normalized correlation coefficient. Two highly similar traces would yield a coefficient approaching one and

inversely, two dissimilar traces would give a coefficient approaching 0 (generally below 0.7). A time window was defined to include 0.5 s before the P onset and 0.5 s after the S onset. In most cases, the total trace duration varied between 2.5 and 6.5 s, depending on the hypocentral distance. Stations A61 and A54 were generally chosen due to their lower background noise levels. Figure 4.16 displays a series of traces for events that define a 9-event multiplet.

4.5.2 Results

Out of the 915 CSZ events tested, 170 (19%) turned out to have at least one associated event. These events constitute 67 multiplets, of which 51 are doublets, and five are five-tuplets or more (see Appendix 2 for complete list). During the time period considered, the majority of CSZ earthquakes (81%) occurred as isolated (i.e. non-multiplet) events. For events of the multiplets, the inter-event time varied between seconds and years, with 25 to 30% within the first 0.1 year (Figure 4.17). Spatially, about 90% of the events are within 2 km distance of the first hypocentre (Figure 4.17). Since two events of a multiplet must lie within a few hundred metres of each other (to have similar waveforms), the inter-event distance found is related to the precision of the routine hypocentre locations. When relocated with carefully re-picked phases, most events cluster within 1 km hypocentral distance. It was found that multiplets are more common in some areas than in others. The area near the 1979 magnitude 5.0 earthquake has the highest percentage with 38% of events that had at least one associated event (15 events out of 39). The most active areas of the CSZ, such as the Central and the Ile-aux-

Coudres sub-zones, do not have higher proportions of multiplets. This seems to imply that most events, even if spatially associated, are not related to the same fault plane, and consequently, reinforces the conclusion that seismicity occurs in highly fractured volumes.

Unlike grouped events which are only spatially associated, multiplets are events that occurred on the same fault plane or on nearby sub-parallel fault planes. If a focal mechanism exists, this fault plane is one of the two nodal planes. With high precision hypocentres, one can define which nodal plane is the actual fault plane. Ideally, one would have liked to find numerous multiplets with focal mechanisms, but only five multiplets included a quality A or B focal mechanism (Figure 4.18). Of these, only one had more than two associated events (911208 03:00 U.T m_N 4.3). This event is discussed in more detail in Section 4.6.3.2.

4.6 Integration of focal mechanisms, earthquake groups and multiplets

The previous four sections have provided the basis of a search for a correlation between earthquakes and faults in the CSZ. Without a surface rupture to directly correlate earthquakes and faults, 25 focal mechanisms can be compared with earthquake groups and multiplets. A similar approach was used in the eastern Tennessee seismic zone by Chapman et al. (1997).

To examine the local variations in CSZ seismotectonics, 19 sub-zones were defined on the basis of epicentral clustering and on common depth distribution (Figure

4.19). These sub-zones are re-grouped in the next three sections according to the value of the median of their respective focal depth (Table 4.6): shallow ($z < 10$ km; Section 4.6.1; Figure 4.20A); intermediate ($10 \leq z < 15$ km; Section 4.6.2; Figure 4.20B); and deep ($z \geq 15$ km; Section 4.6.3; Figure 4.20C). The weakly seismic sub-zones are examined separately (Section 4.6.4).

4.6.1 Sub-zones with shallow activity (A61, COU, EBO)

4.6.1.1 Earthquakes of the A61 sub-zone

A sub-zone near seismograph station A61 (thereafter named sub-zone A61) was defined for a pilot-test of our study of other sub-zones. The A61 sub-zone is a cluster of events within a few kilometres of station A61. The boundaries correspond to a change in focal depths to the southeast and to a decrease of earthquake activity elsewhere. The following section updates the description of the sub-zone given in detail in Lamontagne and Ranalli (1997).

The sub-zone presents many favourable characteristics for a detailed seismotectonic study: (1) good constraints on relative focal depths (probably better than ± 1 km due to the proximity of station A61); (2) a large proportion of shallow hypocentres (90% at less than 12 km depth) located onshore providing the potential for stronger correlations with remote sensing lineaments than in most of the CSZ; (3) location within the sub-zone of an onshore earthquake of magnitude $m_N \geq 4.0$ (most others $m_N \geq 4.0$ have occurred under the St. Lawrence River in areas known for their

Table 4.6 Statistics of the various sub-zones

Zone	Number of events ⁷	% of CSZ	Depth (km)		Magnitude		No. M4+	No. of Multiplets ⁶ (2) = doublets (3) = triplets, etc.
			Min	Max	Min	Max		
CSZ	1638	3.4	0.0	30.0	-1.5	5.0	8	
A61	88	5.4	0.0	18.2	-0.8	4.0	1	1(2); 1(3), 1(4)
COU	182	11.1	1.4	22.3	-0.8	3.8	0	10(2); 2(3)
EBO	53	3.2	2.1	21.6	-1.3	3.1	0	5(2)
MA5	83	5.0	2.0	16.2	-1.0	5.0	4	6(2); 1(3)
CEN	433	26.5	1.1	27.1	-1.0	3.6	0	11(2); 1(3); 3(4)
KAM	153	9.3	6.5	24.9	-1.0	3.1	0	6(2); 1(5)
POC	116	7.1	0.0	30.0	-1.1	3.4	0	1(2); 1(6); 1(9)

⁶

For November 1988-August 1997.

⁷

The seismicity numbers refer to events recorded between October 1977 and December 1997 on five stations with a minimum of 8 phases.

IRE	202	12.3	0.0	22.4	11.3	-1.5	3.7	0.6	0	7(2); 1(3); 3(4)
OIE	65	4.0	8.0	23.0	15.7	-1.0	3.9	0.4	0	0
DNE	91	5.6	6.4	27.5	21.6	-1.0	4.3	1.2	2	4(2); 1(5); 1(6)
SHI	56	3.4	1.4	28.5	10.7	-1.0	3.1	1.0	0	0
AIG	9	0.5	16.8	28.7	21.2	-0.3	2.7	1.2	0	0
SGA	13	0.8	6.1	17.7	11.4	-1.2	2.6	0.0	0	0
CGA	19	1.2	1.5	26.5	19.6	-0.6	2.5	0.8	0	0
NGA	13	0.8	7.2	23.0	14.2	-1.3	3.4	1.1	0	0
UPS	18	1.1	6.4	29.0	13.5	-0.2	2.2	1.4	0	0
DOW	4	0.2	18.6	24.0	22.9	0.1	2.4	1.6	0	
SOS	17	1.0	9.9	25.4	14.7	-0.5	2.4	0.5	0	
LOU	30	1.8	3.1	26.7	9.1	-0.6	4.2	0.8	1	

concentration of larger events); (4) two focal mechanisms available; (5) occurrence of faults related to both major structural systems, i.e. paleo-rift faults which are assumed to bound the earthquake activity (Anglin, 1984) and faults of the Devonian meteor impact structure.

A total of 88 seismic events were located in the A61 sub-zone, or about 5% of some 1644 CSZ hypocentres. In the sub-zone, earthquake epicentres can be divided into an eastern concentration (all located within 5 km of station A61), and a diffuse western group (Figure 4.21). In contrast with the CSZ as a whole, which shows a continuous increase of activity between 0 and 9 km depth, focal depths in the A61 sub-zone show a bimodal distribution which peaks at about 5 and 9 km (Figure 4.22). Compared with the activity of the CSZ, the 5 km group can be considered shallow, while the 9 km group is average. It is not known if this bimodal distribution is due to the short period of observation (i.e. the 8 km low might disappear over a longer time period) or to physical processes that cluster events at 5 and 9 km.

As a first step to evaluate clustering, routinely-determined hypocentre locations were compared to evaluate their spatial proximity. It was found that about 45% of events have at least one other hypocentre within 1.5 km. The rest, about 28%, occurred as isolated events. Ten clusters of two events or more are found, with one group of 31 events and one of nine. Surprisingly, the largest earthquake (m_N 4.0) has no event within 1.5 km of its focus, which implies that it was not preceded by foreshocks or followed by aftershocks.

Three multiplets were found in the sub-zone. In general, the normalized coefficients were lower for the vertical component than for the horizontal components, suggesting scattering of the incident Pg wave. The small magnitude of these events makes P arrivals emergent on most other stations. One quadruplet (891208; 891221; 940614; 960215 with a focal depth of 10 km) includes an event with a computed focal mechanism (event 891208). Routinely-computed hypocentres are very similar for these events. While most traces are consistent for the three events, some are different implying variations in source mechanisms. Although it can be assumed that most of these variations are due to variable focal mechanisms, the similarity of traces on some stations and differences on others suggests that velocity complexities near the hypocentres also play a role. Hence, for some azimuths, small changes in hypocentral locations can produce travel path changes. This quadruplet may have occurred near a lithological complexity, such as a geological contact or a wide fault zone. Traces are extremely similar for events of the triplet (dates of events: 930706; 950303 and 950820 with a focal depth of 9 km). Routine hypocentre locations also show that they are within 1 km of each other. One doublet was also found.

The multiplets represent a total of 9 events out of 47 events tested (or about 19% of the total). Thus, over the seven year period, only a minority of events occurred as consistent reactivations of the same fracture planes. In the A61 sub-zone, very few events could have occurred on the same fractures with similar focal mechanisms. A number of isolated events was found. The time intervals of up to 4.5 years between events of multiplets may indicate a slow strain build-up on a given fracture.

An interesting offshoot of the method described above is the discrimination between seismic events and quarry blasts, as previously applied in Mount St. Helens (Frémont and Malone, 1987). Three quarry blasts were detected by their highly similar waveforms (average normalized correlation coefficients greater than 0.9; Figure 4.23).

In the area surrounding the A61 sub-zone, a very dense system of faults of various orientations and nature exists (Figure 4.24), most of them corresponding to faults either mapped in the field or photo-interpreted by Rondot (1989). In the A61 sub-zone, the focal mechanisms of the two largest earthquakes (11 January 1986, m_N 4.0 and 18 March 1987, m_N 3.3) show mostly reverse faulting in agreement with most other solutions of the CSZ. In both cases, a nodal plane is sub-parallel to the conspicuous WNW-ESE set of lineaments (Figure 4.24). Thus, it is plausible that these earthquakes represent reactivations of this series of faults especially when one considers the shallow focal depths of these two events (in the 4-5 km range) and the well-constrained nodal planes based on first motions.

Three additional focal mechanisms have been computed for small magnitude events based on first motions and S_H polarities. These focal mechanisms are relatively well constrained and show strike-slip-reverse, reverse, and normal faulting respectively. While the reverse faulting type agrees with most mechanisms of the CSZ, dominant strike-slip components and normal faulting are less common. First motions on A61 indicate that the strike-slip component and/or normal faulting types are not unusual: many events had dilatational first motions on station A61. Dilatational first motions in

the centre of the focal sphere are inconsistent with pure reverse faulting mechanisms. In addition, a composite mechanism of all earthquakes of the sub-zone shows a large scatter in first motions incompatible with a unique focal mechanism for all events (78 out of 264 or 30% of all readings are misfits to the best solution). Faulting complexity appears to be present everywhere in the sub-zone; events located within a few kilometres of each other are found to have very different first motions and transverse S_H on A61.

For the period 1978-1982, hypocentres along the north shore of the St. Lawrence define steeply-dipping volumes bound by paleo-rift faults (Anglin, 1984). The present study suggests that within the A61 sub-zone, numerous preexisting fractures are being reactivated by local stress and/or strength variations. This WNW-ESE fault system is older than the paleo-rift faults (Rondot, 1979), but was reactivated by the post-impact readjustment as were many faults of the CSZ (Roy, 1978). These peripheral impact structure faults are part of an inward-dipping system of normal faults. In the A61 sub-zone, the south-dipping nodal planes should be the fault planes. Hence, it appears that some impact structure faults may be reactivated in the A61 sub-zone.

When earthquake concentrations are seen in a data set, the hypocentre precision is always questioned. The earthquake concentrations of this study are not an artifact of location uncertainty. In the A61 sub-zone, highly-correlated traces were found only for events that had routinely-determined hypocentres within 1 km. Thus, these hypocentres appear reliable, probably enough to examine possible correlations between earthquake alignments and lineaments.

Unlike in most of eastern Canada, focal mechanisms of the A61 sub-zone are not pure reverse faulting. While focal mechanisms of two earthquakes with magnitude larger than 3 show reverse faulting, smaller magnitude earthquakes indicate variations in the faulting style, evidenced by the mechanisms of some events (some strike-slip component and normal faulting), and by the rather inconsistent first motion distribution on A61 for the composite mechanism.

4.6.1.2 Ile-aux-Coudres sub-zone (COU)

The hypocentres of the Ile-aux-Coudres sub-zone occur from 2 km down to 22 km depth, with peaks at 5-7 km and at 10 km depth (Figure 4.20A). The four focal mechanisms display a mixture of pure reverse faulting on NE planes to strike slip mechanism (Figure 4.10). The area appears to be heterogeneous in terms of stress and reactivated fault planes. No nodal planes are oriented parallel to the NE trending paleo-rift faults with their steep dip to the SE. The deepening of the SE activity in respect to the NW may correspond to the Charlevoix fault, described in Chapter 3.

4.6.1.3 Les Eboulements sub-zone (EBO)

The Les Eboulements sub-zone is only mildly active (3% of CSZ) and is the shallowest sub-zone. The sub-zone includes one of the best epicentre alignment of the CSZ. Most events occur within a 4 km wide N-S corridor which does not correlate with any lineament in the Radar imagery. It is possible, however, that the earthquakes occur within a highly fractured layer, sub-parallel to the N-S magnetic lineaments to the North

of the sub-zone. The only focal mechanism has one nodal plane oriented N-S and steeply dipping to the west, consistent with a group of 10 earthquakes located within 1.5 km of each other. Hence, in this sub-zone, most earthquakes occur within a N-S volume, possibly lithologically controlled, which includes a fault with similar orientation. Earthquake groups suggest that this zone may continue under Ile-aux-Coudres.

4.6.2 Intermediate depth sub-zones (MA5, POC, IRE, CEN, KAM)

4.6.2.1 The Magnitude 5 sub-zone (MA5)

The Magnitude 5 sub-zone was named after the largest earthquake of the 1977-1997 period, the August 19, 1979 m_N 5.0 earthquake. The sub-zone is bound to the South by a cluster of earthquakes and by deeper activity elsewhere. With only 5% of the CSZ activity, the sub-zone is not the most active, but four of the eight $m_N \geq 4.0$ earthquakes of the 1977-1997 period occurred there. In addition, an anomalously high proportion of multiplets were found: 15 events out of 37 (40%). Four mechanisms are available, all for magnitude > 4.0 ; the m_N 5.0 event with its pure reverse faulting mechanism; a strike-slip solution for the 19971028 m_N 4.7 event; and two reverse faulting solutions for the m_N 4.3 and 4.4 doublet of March 1989. On a cross-section perpendicular to the River, the hypocentres define a steeply-dipping plane (Figure 4.14A).

Based on available information, it is likely that the 1925 M 6.2 earthquake occurred in this sub-zone. It is known that the earthquake occurred within the NE portion of the CSZ at a 10 km focal depth, that its magnitude defines fault dimensions of about

8 length by 4 km width, and that it was a pure-reverse faulting event on a plane oriented NNE to NE. The M5 sub-zone matches all these parameters: it lies within the epicentral zone of the 1925 earthquake, earthquakes concentrate at 10 km depth, and hypocentres define a plane with a trend and dimensions that match what is expected of the 1925 event. In addition, the sub-zone is subject to some of the largest earthquakes of the CSZ, and three out of four mechanisms have more-or-less the same reverse faulting mechanism as the 1925 event.

On October 28, 1997, a m_N 4.7 event occurred, the largest CSZ earthquake since the 19 October 1979 m_N 5.0 earthquake. The main shock was followed by a series of ten aftershocks (as of May 1st 1998; Table 4.7), all within 3 km epicentral distance of the main shock. Within the aftershock sequence, a doublet and a triplet were found. The largest aftershock of the sequence (magnitude m_N 3.2) occurred 90 seconds after the main shock. Its exact location in respect to the main shock is uncertain due to overlapping traces. Most of the other events were well recorded by the four nearest stations (A16, A21, A61 and A64), and were carefully picked to relocate the main shock and aftershocks (Figure 4.25). The aftershocks are all 0.9 to 3.4 km shallower than the main shock, and are well outside the dimensions of the rupture for an event of this size (in the 600 m range; Table 3.3). Interestingly, aftershocks of the nearby 1979 m_N 5.0 event were also shallower than the main shock, and also outside the immediate rupture area (Hasegawa and Wetmiller, 1980). A possible upward migration of fluids to trigger events outside the rupture zone, and possibly on other fault planes can be envisaged, especially if the surrounding volume is highly fractured.

Table 4.7 Aftershocks of the 19971028 event

Date	Time	Lat	Long	Depth	Mag
	(UT)			(km)	
97/10/28	11:44:18	47.67N	69.91W	11.3	4.7m _N
97/10/28	11:45:48	47.67N	69.91W	???	3.2m _N
97/10/28	15:14:10	47.67N	69.91W	9.4	1.5m _N
97/10/28	16:54:03	47.68N	69.91W	10.5	2.5m _N
97/10/29	13:25:15	47.67N	69.91W	8.2	1.3m _N
97/10/31	10:19:18	47.67N	69.91W	8.6	-1.0M _L
97/11/02	00:00:30	47.66N	69.92W	9.2	1.8m _N
97/11/28	10:56:35	47.68N	69.91W	9.1	2.4m _N
98/01/12	02:49:25	47.66N	69.90W	7.9	1.3m _N
98/03/10	13:28:43	47.68N	69.91W	10.4	1.0m _N
98/03/18	18:21:06	47.67N	69.92W	8.3	1.2m _N

4.6.2.2 The St-Irénée sub-zone (IRE)

The St-Irénée sub-zone is fairly active with 12% of the CSZ earthquakes. The earthquakes, however, are of low magnitude with only two events exceeding magnitude m_N 3.0. The area is bound to the SE by the St-Laurent fault. Only one focal mechanism exists for the sub-zone, displaying pure-reverse faulting on planes oriented ENE. About 20% of events (20 out of 98) were multiplets. The low magnitude of the events suggests a highly fractured area, possibly related to the meteor impact.

4.6.2.3 The Central (CEN) and Kamouraska (KAM) sub-zones

The CEN and KAM sub-zones group approximately 35% of all CSZ activity, with 25% only for CEN. Contrasting with this high level of earthquake occurrence, both areas did not witness any magnitude > 4.0 for the period 1924-1997. The CEN sub-zone is an area of diffuse activity, where earthquakes concentrate but do not define planar features (Figure 4.14 A-B). The KAM sub-zone, on the other hand, was defined primarily on the more homogeneous and slightly deeper depth distribution than the one of the CEN. The hypocentres of this sub-zone define a SE dipping plane, very similar to one of the nodal planes of the 1925 earthquake (Figure 4.14 C-D).

4.6.2.4 The La Pocatière (POC) sub-zone

The POC sub-zone, an area of mild activity, is made up of two main clusters of events. As opposed to the CEN sub-zone, it does not show one large hypocentre volume. It appears to mark a transition between the highly active central zone to a less active area

to the SW. A total of 17 events out of 66 were part of multiplets (26%), including a nine-event multiplet (the largest earthquake swarm found), and a six event multiplet. All events of these two multiplets were of small magnitude ($m_N \leq 2.2$), hence none has a focal mechanism.

4.6.3 Deep activity sub-zones (OIE, DNE)

4.6.3.1 The Cap-aux-Oies sub-zone (OIE)

The Cap-aux-Oies sub-zone lies along the trend of the midly active corridor that parallels most of the north shore of the St. Lawrence (see Section 4.6.4). The sub-zone has a radial trend from the central peak of the impact crater. The hypocentre groups suggest that the NE planes may be the reactivated structures. Four focal mechanisms exist for this sub-zone with two showing a reverse-oblique-slip mechanism on NE to SW oriented planes. No multiplets were found in this sub-zone.

4.6.3.2 The Deep Northeast sub-zone (DNE)

The Deep Northeast sub-zone includes a series of deep earthquakes located at the northeast extremity of the CSZ (Figure 4.26). The earthquakes are markedly deeper than in any other sub-zone of the CSZ. The depth of these earthquakes is not due to its location in respect to the seismograph network: the emergence angles on station A64 agree with deep focii. On land, the shallow depths to the West correspond to the Ste-Mathilde fault of Rondot (1989). Although most events occur as isolated earthquakes spread out over the whole sub-zone, there are a few clusters of activity, including some

multiplets that represent 25% of the events. The largest cluster started with the 19911208 m_N 4.1 earthquake, with an unusual strike-slip mechanism. It was followed by a series of aftershocks, including two within the first 6 hours, that constitutes a seven event multiplet. The Pg and Sg arrival times at four stations were carefully re-picked and the events were relocated using time corrections defined by the residuals of the main shock (as in Deichmann and Garcia-Fernandez, 1992). The hypocentres suggest that the NNE-SSW plane is the actual rupture. Two other mechanisms show pure-reverse faulting on NE oriented planes. These fault trends correspond to the numerous lineaments of the north shore.

4.6.4 Weakly seismic sub-zones (SHI, SGA, CGA, AIG, NGA, UPS, DOW, SOS, LOU)

A number of weakly active sub-zones exists in the CSZ, especially in a river-parallel corridor (SGA, CGA, AIG, NGA) or in the periphery of the more active zone (UPS, DOW, SOS, SHI). All together, only 11% of all CSZ earthquakes occur there, with most events representing isolated occurrences (no multiplets were found). Three focal mechanisms exist, showing mostly pure or oblique reverse faulting. At present, there is no evidence for a lithological control of the river-parallel corridor, but the NE boundaries of the SGA, CGA, AIG and NGA sub-zones correspond to changes in hypocentre density and depth, coincident with the position of the St-Laurent fault. The relative quiescence of this strong region may be due to an undetected time dependence of micro-earthquake activity (Wetmiller and Adams, 1990). More simply, the seismic

quiescence may be due to the absence of high pore fluid pressure or of gouge material with low coefficient of friction.

4.7 Discussion

In this chapter, four sources of information were examined to bring to light the relationships between CSZ earthquakes and faults. No surface rupture was found at the bottom of the St. Lawrence on the seismic profiles. Future work with high resolution profiles will bring a more definite answer on this. An absence of surface rupture implies a lower rate of activity during the past few thousands of years, or that earthquakes were too small and/or too deep.

For larger events of the CSZ, such as the 1925 M 6.2 and the 1979 m_N 5.0 earthquakes, focal mechanisms suggest that N to NE trending paleo-rift faults are reactivated in response to regional stresses. On the basis of earthquake depth, focal mechanisms, and hypocentre alignments, the area surrounding the 1979 m_N 5 earthquake may represent the source region of the 1925 event. This area has been subject to repeated $M > 4$ events and is a likely candidate of future large CSZ earthquakes. Although this area is the most likely to have a surface rupture, none was found on the seismic lines that cross it.

Concerning the possible correlations between seismic events and structural trends, some earthquakes do spatially relate to remotely-sensed and/or potential field lineaments, but not exclusively to paleo-rift faults. Although the A61 sub-zone is intersected by a

regional rift fault (the Rang-Ste-Mathilde fault of Rondot, 1989; Figure 4.24), nodal planes of the two largest earthquakes are not parallel to it. For lower magnitude earthquakes, smaller fractures with various orientations might be reactivated by local stresses, as shown by the variety of focal mechanisms and the scatter in P and T axes. In the A61 sub-zone, two of these events have clear orientations parallel to a WNW-ESE series of lineaments possibly reactivated by the Devonian meteor impact. Since not all CSZ impact structure faults are currently being reactivated, it is possibly the combination paleo-rift--impact faults that gives rise to the seismicity. The St-Laurent fault, evident in the remote sensing as well as in the Bouguer gravity anomaly, is a clear boundary to the seismicity of the sub-zones IRE and A61. Southeast of the fault, earthquakes are more dispersed. In general, earthquakes do not concentrate on this paleo-rift feature.

For small magnitude events, the varied faulting styles imply that local stress and/or strength variations control earthquake occurrence. This contradicts the general assumption that background seismicity corresponds to regional stress systems. In our opinion, earthquakes controlled by local conditions are more likely where resistance to sliding is low, for instance as a consequence of high background pore-fluid pressures or low friction coefficient. Interestingly, a high electrical conductivity zone has been detected by a magnetotelluric survey close to the eastern concentration of hypocentres of sub-zone A61 (Chouteau, 1985). This anomaly has been explained by the presence of water or solutions in a zone of high porosity at depth deeper than 1400 m. In the CSZ, two aftershock sequences may be explained by the migration of fluids away from the immediate rupture of the main shocks.

The complexity of CSZ seismicity can tentatively be explained by a combination of factors. In the A61 sub-zone, evidence of lithological complexity near faults is accompanied with variations in fault plane trends and faulting styles. This suggests that CSZ fault zones are irregular surfaces, surrounded by highly fractured rocks. These highly fractured zones respond primarily to regional stresses; however, for some smaller events, they may respond to local changes in stress and/or strength. The whole process can be enhanced, especially for deeper events, by high pore-fluid pressures as discussed in Chapter 2 and in Lamontagne and Ranalli (1996). The concentration of CSZ events near 10 km depth (Figure 4.20B) may indicate a highly stressed level or a frictionally weak level, possibly due to high pore-fluid pressures or to highly fractured rocks.

The small proportion of multiplets found in the CSZ (19%) contrasts, for example, with the Parkfield area of California where most earthquakes occur as clusters (63% for the period 1987-1992; Nadeau et al., 1995). This shows that stress takes a long time to replenish along most faults of the CSZ and that most sub-zones of the CSZ are intensely fractured.

Figure 4.1 Schematic cross-section of the type of rupture expected in the CSZ. Layers A, B, C, D represent sediments accumulated during and between glacial episodes. The rupture due to multiple reverse faulting events originating in the Precambrian basement propagates to the surface (possibly through the Appalachian sequence). The rupture reaches the thick sequence of sedimentary layers. Since more earthquakes ruptured the basal layers (A for example), the amount of slip is larger there than for the top layers. The rupture may reach the bottom of the St. Lawrence River, causing a bathymetric feature.

Figure 4.2 Example of a seismic section (SOQUIP line 22). The profile shows continuous, uninterrupted reflectors (layer in blue) that are interpreted as an absence of rupture. The interface between the Precambrian (P) and the Appalachians (A) is shown by an arrow.

Figure 4.3 Location map of the seismic lines acquired by SOQUIP and their partners. Lines in red show portions of profiles where sub-horizontal markers can be seen. These areas are interpreted to be free of > 20 m ruptures.

Figure 4.4 Seismic profiles that show the interpreted Quaternary sequence: (A) Enlarged view of the SOQUIP seismic lines near the North shore: Line 35 (top) and 37 (bottom). These two profiles cross the region with most $M > 4$ events. The sediment-filled valleys and their sub-horizontal reflectors can be seen. (B) The high resolution profile acquired along SOQUIP line 13. The reflector in red represents the bedrock. Reflectors in colour represent interpreted Quaternary layers that can be distinguished in the numerous multiples. The blue rectangle represents the area where the 6 fathom line lies. The line was proposed to represent a deep seismogenic fault (Anglin, 1984). No rupture can be seen within the rectangle.

Figure 4.5 Lower hemisphere focal mechanisms of the 12 quality A and B for events of magnitude $m_N \geq 3.0$. The maximum and minimum pressure axes are shown as (p) and (t), respectively and the null axis as (b). The dates of the events are shown above the mechanisms; (C): Compressional and (D) Dilatational first motions; (<) and (>) are the S_H first motions.

Figure 4.6. Focal mechanisms of the 8 quality A and B for 1996 summer events. Legend as in Figure 4.5

Figure 4.7 The two well-constrained focal mechanisms for the 1974 field survey (5 degree b-axis search). Legend as in Figure 4.5.

Figure 4.8 Focal mechanisms of the 1979, mN 5.0 earthquake: (A) as published in Hasegawa and Wetmiller (1980); (B) The revised version of Lamontagne (1998).

Figure 4.9 Revised focal mechanisms of Figure 1.26. Poorly constrained mechanisms of 1974 are crossed while the two acceptable ones are checked. Events 890309 and 890311 were found to have one poorly constrained nodal plane (shown as ?) and were rated "X". For the 1979 mN 5.0 event, the two nodal planes found in this study are shown on the previous solution. The other mechanisms were not reexamined.

Figure 4.10 Focal mechanisms of quality "A" and "B", including the mechanisms of the 1979 mN 5.0 event, the two 1974 events, and the event near La Malbaie. In addition, the 1925 focal mechanism (Bent, 1992) is shown in the upper right corner, in the general area where its epicentre was located by Stevens (1980). An arrow shows the location of a few mechanisms.

Figure 4.11 Results from the quality "A" and "B" events of Figure 4.8: (A) Nodal planes; (B) P, T, and B axes.

Figure 4.12 Method of defining earthquake groups. Hypocentres (dots) are grouped if located within a given distance (horizontal line) from each other. Although this figure shows epicentres for simplicity, the association is made according to the hypocentral distance. (A) and (B) are groups of events, whereas (C) and (D) are isolated events. The hypocentres of group (A) define a preferential orientation shown by the dotted line. This trend can be compared with the available focal mechanism of group (A).

Figure 4.13 Hypocentre groups with 10 events or more created with an inter-event distance of 2 km.

Figure 4.14 Vertical cross-sections across the largest group of hypocentres. The positions of the four cross-sections A, B, C, and D are shown in the epicentral map above.

Figure 4.15 Earthquake groups (1.5 km distance) with a focal mechanism (red dot). When possible, the mechanism is located as to line up one of the nodal planes with the orientation suggested by the group.

A54 Vert.

Figure 4.16 Traces of a nine-tuplet (earthquake swarm) recorded on stations A54 between July 1994 and October 1996; (A) Vertical component; (B) North-South component; (C) East-West component.

B

A54 N-S

C A54 E-W

Figure 4.17 Characteristics of the events of multiplets. (A) Time between the first event and the other members of the multiplets (in blue) and between events of the multiplets (in red). Events are grouped in 0.1 year time intervals. (B) Hypocentral distance between the first event and the other members of the multiplet. Events are grouped by 0.5 km distance, absolute distance in red; cumulative percentage in blue.

Figure 4.18 CSZ Multiplets of the period November 1988 to August 1997 (yellow); focal mechanism are centered on the event epicentre (red).

1- UPS	6- SGA	10- CEN	16- DNE
2- LOU	7- IRE	11- KAM	17- DOW
3- COU	8- OIE	12- A61	18- SOS
4- POC	9- CGA	13- AIG	
5- EBO	10- CEN	14- NGA	
		15- MA5	

Figure 4.19 Earthquake hypocenters and sub-zones. The sub-zone numbers refer to the acronyms listed below. Table 4.7 lists the earthquake statistics of the sub-zones.

Figure 4.20 Focal depth distribution for the various sub-zones. The number of earthquakes per 1 km depth intervals are presented as absolute numbers (thin lines) and cumulative percentage (thick lines). Subzones with earthquakes mostly: (A) Shallow; (B) Intermediate; (C) Deep.

Figure 4.21 Hypocentres of the A61 sub-zone in map view (top), W-E (below left) and N-S (below right) cross-sections. The map view presents focal mechanisms hypocentres with depth colour scheme similar to Figure 4.19. Red dots indicate the positions of events belonging to earthquake multiplets.

Figure 4.22 Compared depth distributions in absolute numbers and cumulative percentage between CSZ (in blue) and sub-zone A61 (in red). The curves show that the A61 sub-zone events are generally shallower than the CSZ earthquakes (maximum depths of 19 and 30 km respectively).

Figure 4.23 Vertical seismic traces of three quarry blasts recorded on station A61 and recognized by their highly similar traces.

Figure 4.24 Airborne Synthetic Aperture Radar (SAR) image of the A61 Sub-zone (green polygon) with earthquake epicentres of the period 1978-1995 (red dots). Lineaments oriented approximately E-W (shown by the dashed lines to the left) are sub-parallel to the strike of two nodal planes (blue lines) for the two largest events of the sub-zone. The WNW-ESE lineaments and their convergence towards the sub-zone are clearly seen. The Rang Ste-Mathilde fault (Rondot, 1979) is shown and is sub-parallel to one of the nodal planes of a focal mechanism.

Figure 4.25 The 19971028 aftershock sequence. Map view (above): epicentres of the 19971028 event (red dot), of its aftershocks and focal mechanism of the main shock. (Below): Cross-section perpendicular to one of the nodal planes that shows the shallower locations of the aftershocks.

Figure 4.26 Hypocentres of the DNE sub-zone. (Above) Epicentral map of the DNE sub-zone with focal mechanisms. (Below left) West-East cross-section; (Below right) North-South cross-section. Earthquake multiplets are shown in red on the cross-sections.

Chapter 5

Determination of the CSZ velocity structure from local earthquake data

5.1 Introduction

To locate an earthquake, one has to determine the optimum values of four earthquake parameters: the latitude, longitude, depth of the hypocentre, and the time of occurrence of the event. The best estimates of these parameters are those that minimize the differences between the observed and the calculated arrival times (i.e., the residuals) for each phase recorded at seismograph stations (Kissling, 1988). Residuals exist due to errors in the observed and the calculated (theoretical) arrival times. For the observed data, the errors come from the inaccuracy of the picked arrival times and of the time control of the recording system. The accuracy of calculated arrival times, on the other hand, depends mostly on how closely the velocity model resembles that of the real Earth, and to a smaller extent, on location errors of recorders, near-surface velocity perturbations and errors in the estimates of the hypocentral parameters.

With reliable observed data and good control on the sources of uncertainties, the residuals can be used to infer the velocity structure of the Earth. For an event, the residuals from all stations can be expressed in a RMS (Root Mean Square) sense as

$$RMS = \left(\frac{\sum_I (residual)^2 \times (weight)^2}{NP} \right)^{\frac{1}{2}} \quad \text{for } I = 1, NS \quad (5.1)$$

where the residual is the difference between the observed and the calculated arrival times, the weight is a quality factor assigned to a picked arrival time, NS is the number of stations, and NP is the number of phases used in the solution. The optimum velocity model minimizes the RMS of a series of events.

Extracting velocity information from microearthquakes is called simultaneous inversion of local earthquakes or local earthquake tomography (Thurber, 1993). In contrast with the tomographic inversion of controlled sources and teleseisms, the simultaneous inversion of local earthquakes determines the location and origin time of events, along with the velocity of the material between the source and the recording stations. With sufficient data, the inversion can define a local or a regional velocity model, either one-dimensional (1-D; homogeneous or layered) or three-dimensional (3-D; Kissling, 1988; Kissling et al., 1994).

In order to clarify seismotectonic properties, the main objective of this chapter is to infer the optimum CSZ velocity model contained in the local earthquake data of the period 1977-1997. The chapter is divided into ten sections. In Section 5.2, the previous studies of CSZ velocities are presented, along with the routine procedure for locating earthquakes. Section 5.3 discusses two possible sources of errors of the observed phase data, namely the time control of the seismograph network and the picking precision of Charlevoix phase data. The selected data used in the inversion are presented in Section 5.4. With this data set, optimum velocity models of the CSZ are presented: the average V_p/V_s ratio (Section 5.5); the homogeneous velocity model (Section 5.6); and two layered models (one for the north shore and one for lower velocity Appalachian sediments of the

south shore; Section 5.7). The impact of the velocity models on the earthquake locations is discussed in Section 5.8. An inversion of the Laurentides Park velocity model is presented in Section 5.9. Finally, the velocity information is discussed in Section 5.10.

5.2 Previous studies of the regional velocity structure and routine analysis of the CSZ data

Crustal velocity models of the Charlevoix-Saguenay area are shown in Figure 5.1. To the west of the CSZ, across the Grenville Front, P-wave upper crustal velocities in the 6.3-6.5 km/s range were determined, with mid to lower crustal values of about 6.55 km/s (Mereu and Jobidon, 1971; Berry and Fuchs, 1973). Later, these velocities were used to model the strong motion records of the 1988 m_{bLg} 6.5 Saguenay earthquake (Somerville et al., 1990). In the CSZ itself, a seismic refraction-reflection survey showed that the velocity structure differed between the two shores of the St. Lawrence River (Lyons et al., 1980). Logan's Line was modeled as a 20° SE dipping interface separating the Appalachian nappes and the Grenvillian rocks with respective P-velocities of 5.5 and 6.1 km/s. Under the north shore, velocities increase linearly from the surface to 18 km depth, where they jump from 6.6 to 6.8 km/s. Near surface velocities are lower inside the Charlevoix impact crater (6.08 ± 0.04 km/s) than in the north shore (6.27 ± 0.11 km/s), a fact possibly related to the intense fracturing. To the northeast of the crater, velocities progressively move towards the average value of 6.2 km/s. To the west of the crater, in the Laurentians, a higher apparent velocity of 6.48 km/s was found for one station.

The 35 microearthquakes recorded during the 1974 Charlevoix experiment were

inverted to obtain a crustal velocity model (Leblanc and Buchbinder, 1977). Within the P-velocity range of 5.6 to 6.6 km/s, the 6.2 km/s Pg gave the smallest RMS (with a 3.57 km/s Sg velocity). These velocities were later adopted for Charlevoix routine locations. During the same field survey, surface blasts of the north shore were best located with near surface P-velocities of 6.0 km/s.

In addition to the work of Lyons et al. (1980), the strong velocity contrast between the Grenville rocks and the overlying Appalachian sequence was also modelled with other data sets. Using locally recorded teleseisms, Hearty et al. (1977) modelled the interface as a 19° SE dipping plane separating the 6.1 km/s Grenville from the 5.4 km/s Appalachians. Using some local CSZ earthquakes, Lamontagne (1985) found that a 15° SE dipping boundary between the 6.2 km/s Grenville and the 5.5 km/s Appalachian rocks minimized the RMS. Although these experiments showed that this strong velocity contrast should be considered in CSZ locations, the homogeneous "standard" velocity model continued to be used for routine locations. It was assumed that the shift due to the Appalachian rocks would be approximately the same for all CSZ events, i.e., a few kilometres to the SE (Anglin and Buchbinder, 1980; Lamontagne, 1985).

Over the years, the Geological Survey of Canada (GSC) has used two programs to locate CSZ earthquakes. Between 1978 and 1993, the location program "LOC" (also named "CANCESS") was used. "LOC" uses a nonlinear regression technique, similar to the HYPO-series of programs (Drysdales et al., 1990). Since 1993, Charlevoix earthquakes are routinely located with the GSC program "GRL" which uses a grid search algorithm. The program calculates the RMS for a series of equally-spaced grid points in the origin

time, latitude, longitude and depth space. The grid is re-centred on the parameters with the minimum RMS, and the RMS are recomputed using a finer mesh. The grid search stops when the mesh is smaller than a predetermined precision for the origin time, latitude, longitude and depth. The program “GRL” currently accepts a homogeneous or a layered velocity model, with direct or refracted phases. In this thesis, the adopted approach combines the respective advantages of the two location programs. The program “GRL” is less susceptible to local minima due to its grid search approach. “LOC”, on the other hand, converges faster to the hypocentre than “GRL” (with a layered velocity model, it could be as much as 30 times faster). Although both “GRL” and “LOC” use layered velocity models, “LOC” can accept two layered velocity models to compute hypocentres. Thus, with “LOC”, a velocity structure can be associated with a group of stations. As described in Section 5.7, this property is central to the calculation of the north shore-south shore velocity contrast.

5.3 Sources of uncertainties in the observations

The timing accuracy of the Charlevoix Local Telemetered Network (CLTN) was briefly discussed in Section 1.4. For the period October 1977 to August 1994, time series data for all CLTN stations were time tagged at the concentrator site, providing uniform time across the network. During the period October 1977 to November 1988, two analogue (LMQ and POC) and one digital (LPQ) vertical sensors were in operation at various periods (Figure 1.7). As these stations were vertical component only, their lower precision data were not used to invert the velocity model.

For the post-August 1994 period, Global Positioning System (GPS) timing has been tagged along with the data at each site. This scheme ensures accurate timing at each site, completely independent of the transmission time to the central site. The only possible source of error is the drift of the internal clock that occurs when the signal from a GPS satellite is lost. This, fortunately, rarely occurs and is generally of very short duration (K. Beverly, GSC, pers. comm.). In January 1994, the station LMQ became a digital three-component broadband station, part of the Canadian National Seismograph Network. Accurate timing was provided by a GPS clock, which makes data from this station compatible with the CSZ local network. For station LMQ, however, the 40 Hz sampling rate (0.025 s/sample) makes the phase readings less precise than the 100 Hz (0.01 s/sample) for the local CSZ network.

In terms of station location accuracy, all CSZ stations, including those of the 1996 summer experiment, were located with GPS, accurate to within 25 m for latitude and longitude (G. Girouard, GSC, pers. comm.; Lamontagne et al., 1997). To be consistent with other locations of the network, elevations were scaled on 1/50,000 topographic maps.

In a simultaneous inversion study, the precision of the observed data also depends on the picking of the P and S phases. The picking precision can be evaluated by comparing the picks made by two analysts. To evaluate this precision, phase data from 21 events of the period 1988-1994, originally read by F. Anglin of the GSC, were compared with picks made by the author. The original data were read with the GSC analysis package "SAM" which offers a rotated trace option to ease Sg picking (F. Anglin, pers. comm.). The package "DAN" (Nanometrics Inc) was used for the re-reading.

This package is better at enlarging portions of a given trace. The two packages round off the time picks to the nearest hundred of second, implying that the measured time differences between the two picks are given in hundreds of seconds and not in number of samples (0.0125 s/sample for the 1988-1994 CLTN). In general, most phases are impulsive in nature, especially for sites at short epicentral distance or located on the north shore. There were, however, instances where background noise hampered precise phase picking.

The absolute values of the time differences are shown in Figure 5.2. In general, the two sets of P arrivals are more consistent than S arrivals; approximately 80% of phase readings for Pg are within 0.02 seconds whereas about 80% are within 0.05 s for Sg. The larger scatter of Sg picks may be partly due to shear wave splitting which creates smaller amplitude signals prior to the real S arrival (Buchbinder, 1985). Furthermore, the S phase can be emergent rather than impulsive, especially for smaller magnitude events, for larger epicentral distance, and for stations on the south shore which have a higher background noise. In addition, the S onset can be picked as a sharp increase in amplitude or as a weak emergence a few samples before. Compared with three component data, vertical traces do not allow the same precision level, especially for Sg arrivals at larger epicentral distances.

For the post-August 1994 period, the raw data were filtered with a Finite Impulse Response ("FIR") filter which gives rise to unwanted precursory effects prior to the P and S arrivals. In order to eliminate these artifacts, the time series data are routinely band-pass filtered (5-15 Hz pass band) at the analysis stage. This filtering removes the unwanted

ringing, while preserving the micro-earthquake signal. With this band-pass filter, the CLTN picks are within 0.01 s of those obtained with the “un-FIR filter” (M. Andrew, GSC, pers. comm.).

5.4 Selection of the earthquake data

The most reliable and consistent hypocentral data for the CSZ were recorded after the local network became a three-component array in November 1988. CSZ events with a minimum magnitude of m_N 1.5 were selected to insure conspicuous P and S arrivals on all CSZ stations. To obtain reliable depth control, selected events had to be located within the CLTN 6-station network, and recorded on at least 6 stations with a minimum of 12 seismic phases (P and S). Finally, only events with focal depth of 6 km and deeper were selected, to avoid the poorly converging solutions of some shallow events.

For the period November 1988 to December 1997, 171 events met the selection criteria (Figure 5.3A). Their depth distribution is very similar to that of all CSZ events recorded between 1978 and 1993 (Figure 5.3B). In order to account for the lower precision of S phase picks, P phases were weighted 4 (Quality “A” of the GSC scheme) whereas S phases were weighted 1 (Quality “B”). In the few cases where the residuals were higher than 0.4 s, a weight of 0.25 (Quality “C”) was assigned to the Sg pick. With this weighting scheme, the average RMS is 0.084 s, with 90% of RMS less than 0.12 s (Figure 5.3C).

5.5 V_p/V_s Ratio for the CSZ

The traditional method of computing the ratio of P to S wave velocities (V_p/V_s ratio) is the so-called Wadati plot (Wadati, 1933). On this type of graph, where the Sg-Pg travel time is plotted against the Pg travel time, the V_p/V_s ratio is simply equal to the slope of the best fitting line plus one. In the CSZ case, for given hypocentre and station, the Pg travel time is obtained by subtracting the origin time calculated with the “standard” GSC velocity model from the Pg arrival time. The S-P time is computed for each station by subtracting the Pg time from the Sg time. The best fitting line is obtained with a “robust” linear regression method (Huber, 1981). Based on some 1267 P and S phase picks from the data set discussed in Section 5.4, a V_p/V_s ratio of 1.74 is found (Figure 5.4), very close to the assumed value of the GSC “standard” velocity model (1.73, assuming a Poisson’s ratio of 0.25). Since the Pg travel time depends on the origin time computed using P and S phases, Wadati plots have some dependence on the S velocity used to derive the hypocentres. In the CSZ, this dependence is small, however, mainly due to the smaller weight assigned to S phases.

5.6 Homogeneous velocity model

The current “standard” homogeneous velocity model was established by Leblanc and Buchbinder (1977) using 35 micro-earthquakes. To test this model with a more complete data set, RMS values were computed for a series of Pg and Sg velocities. V_p was varied between 5.80 and 6.60 by 0.05 km/s increments, while Sg was computed from V_p using a V_p/V_s ratio of between 1.65 and 1.80 by 0.03 km/s increments. Velocities of

6.20 $\pm$ 0.10 km/s for Pg and 3.58 $\pm$ 0.06 km/s for Sg velocities gave the minimum RMS (Figure 5.5). These velocities correspond to the “standard” GSC velocity model, determined by Leblanc and Buchbinder (1977). The Vp/Vs ratio (1.73) is similar to the one determined with the Wadati plot of the previous section.

Although the homogenous velocity model is an acceptable representation, the 0.08s RMS is still higher than the picking precision of P and S phases. For the permanent stations, the Pg residuals, obtained with the “standard” velocity model, show a systematic bias (Figure 5.6). Whereas LMQ, A16 and A11 are more or less centred on 0.00 s, a bias can be observed for the stations A64 ($\approx$ -0.10 s), A61 ($\approx$ +0.05 s), A54 ($\approx$ +0.05 s) and A21 ($\approx$ +0.21 s). Thus, the average velocity to station A64 should be higher than the “standard” velocity model, whereas it should be lower for the three other stations. These results are in line with the seismic refraction results of Lyons et al. (1978), which showed a lower velocity for the south shore (A11 and A21) and for the crater (A54), and higher velocities for the NE portion of the CSZ, outside the crater (A64). To test these regional variations, the data set was subdivided into two groups: one for the southwesternmost stations (the “crater”: A11-A16-A61-A54) and one for the northeasternmost stations (“outside the crater”: A16-A21-A64-A61). Earthquakes in these two groups were relocated using only the phases from the four stations that defined the polygon, while only the residuals were computed for the two other stations. With this smaller number of stations, the RMS for the two groups were smaller for the “standard” model: 0.046 s for both groups (Figures 5.7). While the “standard” velocity model appears appropriate inside the crater, outside the crater slightly higher velocities provide the lowest RMS (about 6.3

km/s for Pg and 3.64 km/s for Sg). Higher velocities outside the crater were also suggested by Lyons et al. (1980). The Pg residuals for the two groups of stations also reveal a systematic bias: the velocity between the crater events and station A64 should be higher than 6.2 km/s (negative residuals), while velocities to stations A21, A11 and A54, should be somewhat lower (positive residuals; Figure 5.8). To consider this bias, station corrections were defined for the Pg and Sg phases (Table 5.1), bringing the RMS from 0.0838 down to 0.0719 s.

5.7 Calculation of layered velocity models for the north and the south shores

As the minimum RMS with a homogeneous velocity model remains higher than the reading uncertainties, layered velocity models were tested to further reduce the RMS. Since it is known that the largest velocity perturbations arise from the Appalachian rocks of the south shore, it was decided to use two velocity models for the CSZ, one for the north shore stations and one for the south shore stations. To test the validity of this approach, the velocity models of Lyons et al. (1980) were used as a first approximation to calculate travel times to the north shore and south shores (Figure 5.1; Table 5.2). With these models, the RMS was 0.071 s, similar to the best homogeneous model plus the station corrections. On the basis of these positive results, it was decided to further refine these models with the hypocentral data.

Table 5.1: Station corrections to be used with the standard velocity model

Station	Pg	Sg
	Station Correction	Station Correction
	(s)	(s)
A11	0.00	-0.15
A16	0.00	0.05
A21	-0.10	-0.15
A54	-0.05	0.00
A61	-0.05	0.00
A64	0.10	0.10
LMQ	-0.05	-0.05

Table 5.2 Velocities used to model the north and the south shore (based on Lyons et al., 1980)

Depth to top (km)	V _p	V _p
	North Shore	South Shore
0	6.4	5.5
6		6.5
8	6.5	6.6
14		
18	6.8	7.0
22		
29	6.9	7.1
37		

The best layered velocity models were sought by systematically computing the RMS for a suite of layered velocity models for the north and the south shores. A major constraint of the LOC program is the maximum number of layers (9). In order to prioritize the upper crustal velocity structure, the Appalachian sequence was defined as four 2 km thick layers, while the north shore had only one 8 km thick layer. The north shore and south shore models are the same below 8 km depth. Ranges of velocities and increments were defined for the eight layers of both models (Table 5.3), with the aim of defining the minimum RMS between areas of higher RMS. No velocity reversals with depth are allowed by the program LOC. This restriction is reasonable as a velocity increase with

Table 5.3 Range of P-wave velocities used in the 2-D velocity analysis

	North Shore		South Shore	
Depth to top (km)	Min Vp (km/s)	Max Vp (km/s)	Min Vp (km/s)	Max Vp (km/s)
0	6.1	6.4	5.5	5.9
2			5.6	5.9
4			5.6	6.0
6			6.0	6.4
8	6.1	6.4	Same as North Shore	
10	6.2	6.7		
12	6.2	6.7		
14	6.2	6.7		

depth is generally assumed for the crust. For a given layer, the S velocity was computed from the P velocity using a V_p/V_s ratio of 1.74, as suggested by the Wadati plot (Section 5.5). The RMS calculations were performed on nearly 28000 models which took 18 days to compute on an Axil 320 Sun workstation. Although this systematic search method is more computer intensive than an iterative method that converges to a minimum RMS, it tests all plausible velocity models.

The resulting velocity models for the north and the south shores are presented in Figure 5.9. The figures present the minimum RMS that can be obtained with a velocity

value at a given depth, providing a view of the velocity resolution. Using Figure 5.9B as an example, at 1 km depth, the best RMS that can be expected for a 5.9 km/s P-velocity is in the range between 0.061 s and 0.062 s, well above the best RMS value in the 0.059s range. Hence, for the 0 to 2 kilometer layer, the 5.5 to 5.8 km/s values are more likely. The velocity models that minimize the RMS to 0.0594 s are given in Table 5.4. With these velocity models, the P and S residuals become more tightly clustered at around 0.00s (Figure 5.10). The P residuals are clustered near 0.00 s, which implies that most of the homogeneous model P-velocity biases (Figure 5.6), are resolved with the layered velocity models. The S residuals are more dispersed, a combined consequence of their lower weight in the solutions, the lower precision of the observed data, and possibly, variations in the V_p/V_s ratio.

To a certain extent, the details of the resulting velocity models depend on the assumptions made to derive them. The two most important assumptions are the 0-8 km homogeneous velocity layer for the north shore, and the 1-D model under 8 km. Independent models for the north and the south shores could possibly define a layered structure for the north shore. Similarly, below 8 km, there could be lateral variations between the north and the south shore. These questions could be answered with future 3-D inversion.

Considering the picking precision of the S data, resolving the V_p/V_s ratio for each layer is not feasible. An attempt was made, however, to resolve the V_p/V_s ratio for the Appalachian and the Grenvillian sequences taken as two blocks. Using the minimum

Table 5.4 Velocity model that minimizes the RMS

Depth to top (km)	Vp North Shore (km/s)	Vp South Shore (km/s)
0	6.1	5.5
2		5.6
4		5.6
6		6.1
8	6.4	
10	6.4	
12	6.6	
14	6.6	
40	8.0	

RMS model of Table 5.4, V_p/V_s ratios for the two blocks were systematically tested between 1.70 to 1.85 by 0.01 increments; one block for the first four layers (0 to 8 km) of the south shore and one block for the remaining layers of both shores. The best V_p/V_s ratios were 1.72 for the 0 to 8 km south shore model, and 1.73 for the north shore model. With these slightly different V_p/V_s ratios, the RMS was lowered to 0.05844 s (from the original 0.05938 s). Hence, the average CSZ V_p/V_s ratio is close to the “standard” value of 1.73.

5.8 Impact on CSZ earthquake locations

Figures 5.11 and 5.12 compare the earthquake locations inferred here with the homogeneous “standard” velocity model. As expected, most earthquakes relocate to the SE due to the lower velocity Appalachian nappes. The average shift is about 1.5 km along an azimuth of 135° . About two thirds of the events are shallower than the original locations, but the average shift for the whole data set is close to 0 km. The epicentral shifts are relatively small (Figure 5.12). The general pattern of the CSZ earthquakes does not change with the improved velocity model.

5.9 Velocity model between the CSZ and station DAQ

One interesting question is how different the CSZ velocity structure is compared with the mostly aseismic Grenville Province. The seismograph station Lac Daran, Québec (DAQ), located some 100 km west of the CSZ (Figure 1.11), has recorded numerous CSZ events, as small as m_N 1.0. These phase data can be used to infer a 1-D velocity structure between the CSZ and station DAQ (thereafter named the Laurentides Park region). In addition to providing comparisons with the CSZ, the regional crustal velocity model could be used in future studies of the Saguenay earthquake zone.

A total of 96 events with $m_N \geq 2.0$ were chosen from the data set used in the CSZ inversion (Figure 5.13A). The hypocentral information (latitude, longitude, depth, origin time) was pegged at the values computed with the CSZ stations and the “standard” homogeneous velocity model. This data set has a RMS of 0.7883 s, due to rather high residuals for Pg (centred near -0.7 s) and Sg (centred near -0.3 s), which indicate higher

velocities than the “standard” velocity model (Figure 5.13B). DAQ phase readings are used to compute the RMS of the events, with a weighting scheme of 4, 1 and 0.25 for P, S and the absolute value of the residuals ≥ 1.0 s respectively. The Wadati plot indicates a V_p/V_s ratio of 1.786, with some scatter in the phase information (Figure 5.14). The best homogeneous velocity model gives a minimum RMS of 0.439, with P_g and S_g velocities of 6.45 and 3.69 km/s respectively (V_p/V_s ratio of 1.75; Figure 5.15). The best layered model gives a minimum RMS of 0.430 s, assuming a V_p/V_s ratio of 1.786 (Figure 5.16; Table 5.5). With this velocity model, P_g and S_g residuals are closer to 0.0 s, with some remaining events with residuals exceeding ± 0.5 s (Figure 5.17). Some events have anomalous P and S residuals, due possibly to differences in timing between the CLTN network (prior to August 1994) and station DAQ of the ECTN network. Two events, recognized as earthquake multiplets, had very different relative arrival times on DAQ. The timing difference between the CLTN and the ECTN cannot be measured directly.

To clarify whether the high P and S residuals could be due to timing, the layered velocity model was recomputed with events recorded after November 1995, i.e., after both the CLTN and station DAQ relied on GPS clocks. Events were selected with the same criteria as before, with additional events recorded between January and March 1998. The 44 event data set had a RMS of 0.644 s, with a large scatter of P and S residuals (Figure 5.18). The Wadati plot is less scattered and gives a V_p/V_s ratio of 1.89 (Figure 5.19). The homogeneous velocity model minimizes the RMS (0.239) with P_g and S_g velocities of 6.45 km/s and 3.57 km/s respectively (V_p/V_s ratio of 1.81; Figure 5.20). The best layered model has P-velocity of 6.0 km/s down to 6 km, and 6.6 km/s beneath 6 km (assuming

Table 5.5 Velocity model for the Laurentides Park region

Depth to top of layer (km)	Range of Vp values (km/s)		Optimum value (km/s)
	Minimum	Maximum	
0	6.0	6.6	5.9
2	6.0	6.6	6.1
4	6.0	6.6	6.2
6	6.2	6.7	6.3
8	6.2	6.7	6.4
10	6.3	6.7	6.7
14	6.3	6.7	6.7
18	6.3	6.8	6.7
40	8		

Vp/Vs of 1.81; Figure 5.21). The RMS is reduced to 0.238 s, slightly less than the homogeneous model. Although most of the P and S residuals are recentred near 0.00 s (Figure 5.22), high anomalous residuals remain, especially for the S phase. Except for one cluster of five events with high S residuals, most events with high P and/or S residuals are not clustered in one specific part of the CSZ. Since the timing uncertainty is not the source of the problem, variations in the Vp/Vs ratio must exist in the Laurentides Park

region. The positive S residuals suggest a lower-than-average S velocity, implying a V_p/V_s ratio > 1.81 .

The anomalous S residuals imply that the misfits of the 1988-1997 data set were not entirely due to bad time control of the ECTN and the CLTN. One reason for these anomalies can be that the geology between the CSZ and station DAQ is not homogeneous, but made up of various lithologies. For a given ray path, S residuals will vary according to the V_p/V_s ratio of the geological unit. Another possibility is that the P and the S phases are using different ray paths, possibly not the direct ones, due to the complex geology of the area. This could lead to anomalous S residuals.

5.10 Impact of the velocity models on the understanding of the CSZ

From the local hypocentre data, it has been possible to infer velocity information for the CSZ and the Laurentides Park region and to reduce significantly the RMS of the earthquake solutions. The current CSZ network is sufficient to invert a “pseudo-2-D” P-velocity model, i.e., two 1-D velocity models, one for the Appalachian sequence under the south shore, and one for the Grenville. For the south shore model, the 6 km thick Appalachian rocks have P-velocities in the 5.5-5.6 km/s range. For the Grenville, transitions in the P-wave velocities occur at 8 and 12 kilometres, with a passage from an average 6.2 km/s upper crustal to > 6.5 km/s mid-crustal velocities. Interestingly, this velocity transition also corresponds to the peak in the earthquake occurrence distribution, possibly related to a lithological change. In general, the velocity results are similar to those obtained in seismic refraction surveys (Figure 5.23). The layered velocity models

could be used as the starting model in future 3-D hypocentre-velocity inversion. Appendix 3 discusses how the current CSZ network could be upgraded to allow 3-D velocity determination.

Using the best layered velocity model, earthquake locations shift by an average 1.5 km to the southeast, with most hypocentral depths remaining within 2 km of the original positions. Although individual hypocentres may shift with the new velocity model, the hypocentre distribution at a whole may not be very different. The uncertainty in absolute hypocentre positions must be kept in mind while correlating earthquakes with lineaments or structures at depth.

In the Laurentides Park region, the P-velocity structure differs from the CSZ, especially in terms of the V_p/V_s ratio (1.81 and 1.73, respectively). Variations in V_p/V_s ratios are related to difference in pore fluid pressure, fabric and composition (Musacchio et al., 1997). High V_p/V_s ratios can correspond to high pore pressure, foliation parallel to the ray path or low quartz content. With high pore-fluid pressure, it is a decrease in S velocity, rather than an increase in P velocity, that increases the V_p/V_s ratio. Based on the homogeneous model for the Laurentides Park (Figure 5.20) however, we know that V_p is higher and V_s is similar to the CSZ. The second possible explanation, the parallel foliation, is also unlikely: constant foliation attitude is not realistic over some 100 km distance in the very geologically complex Laurentides Park region. The probable explanation lies in a difference in the mid crustal lithology. The Laurentides Park contains typical Grenvillian rocks, such as mangerite, charnockite and garnet gneiss. Similar high V_p/V_s ratios and high P-wave velocities are found in Grenvillian crust further south, in

the Southern Ontario-Northern New York region (Musacchio et al., 1997). Consequently, the Laurentides Park region appears to be a “standard” Grenvillian crust, with some spatial Vp/Vs variations as suggested by the misfit S residuals.

The CSZ, on the other hand, appears to differ from the Grenvillian crust found in the Southern Ontario-Northern New York region (average crustal Vp/Vs ratios: 1.73 for the Appalachians; 1.81 for the Grenville; Musacchio et al., 1997). While the CSZ Appalachian ratio is similar (1.72), the CSZ Grenville ratio is much lower than elsewhere (1.73). The reason for the CSZ average Vp/Vs ratio is not obvious. High pore fluid pressure can explain fault reactivation (Chapter 2), but its relationship with low Vp/Vs is unclear. High pore fluid pressure can be highly localized along fault zones. The Vp/Vs ratio, on the other hand, results from averaging over a whole region, including some inactive, and possibly dry, blocks. Unless pore fluids are pervasive, their presence may not be revealed by the current data. In some experiments with high pore pressure, the reduced effective pressure brings about an increase in the Vp/Vs ratio (increase in Poisson’s ratio; Spencer and Nur, 1976). The interrelations between Vp/Vs, lithology, pore geometry and pressure are not sufficiently well understood to be definite (Wenzel and Sandmeier, 1992). Lithologically, an average Vp/Vs ratio corresponds to granite at mid-crustal depth (Musacchio et al., 1997). This is incompatible with the outcropping CSZ lithologies (mangerites, charnockites and garnet gneisses) that have Vp/Vs ratios in the 1.78 to 1.81 range. The most likely explanation is that, with the CSZ data set, the more felsic upper crust ($Vp/Vs \approx 1.73$) is better sampled than the more basic lower crust. According to Figure 5.9A, velocities in the 6.5-6.6 km/s range, generally associated with

more basic rocks, are found below 12 km depth. Hence, most of the travel path, on which V_p/V_s is based, goes through more felsic rocks. In the Laurentides Park case, on the other hand, the basic rocks are shallower (below 6 km depth), and most of the travel paths take place within them. In the future, independent calculations of V_p and V_s at depth may validate this interpretation.

Figure 5.1 P-wave velocity models for the Charlevoix and Saguenay regions. The CSZ North and South shores are the velocity models obtained from seismic reflection-refraction surveys (Lyons et al., 1980). The "standard" velocity model is used for routine earthquake locations in eastern Canada, including the CSZ. The "standard" model assumes a V_p/V_s ratio of 1.73. The Saguenay model was defined by Somerville et al. (1990) on the seismic refraction survey described in Berry and Fuchs (1973).

Figure 5.2 Absolute time differences between Pg and Sg arrivals picked by F. Anglin of the GSC and the author for some 21 events. About 90% of the Pg picks and 80% of the Sg picks match within 0.05 s.

Figure 5.3 Characteristics of the events selected for the velocity inversion: (A) Epicentral map with black lines representing the faults of the Precambrian shield; (B) Depth distribution of the events by 1 km classes in absolute numbers (left) and in cumulative numbers (right). (C) RMS distribution of the selected events, calculated with the "standard" velocity model.

Figure 5.4 Wadati plot for the events selected.

Figure 5.5 RMS values for various combinations of V_p and V_s (scale in seconds). Areas in white represent V_p and V_s combinations that were not used.

Figure 5.6 Residuals for the various CSZ stations grouped by 0.05 s classes; (A) For P arrivals; (B) For S arrivals. As in similar graphs to follow, the value plotted represents the number of events that is less than or equal to the X value,. For example, the 80 value at 0.00 s for A16, means that 80 values are in the range -0.04 to 0.00 s range.

B

Figure 5.7 RMS values, as in Figure 5.5: (A) Using events within the Charlevoix crater, located only with data from stations A11, A16, A61 and A54; (B) Using events outside the crater, located only with data from stations A16, A21, A64 and A61.

B

Figure 5.8 P-wave residuals for events located inside (A64 and A21) and outside (A54 and A11) the impact structure.

Figure 5.9 Best RMS obtained as a function of P-wave velocity (scale in seconds) and depth of layers:
(A) North shore; (B) South shore.

Figure 5.10 Number of residuals grouped in 0.05 s classes for the CSZ stations: (A) P wave; (B) S wave.

Figure 5.11 Shifts between original event locations ($X = Y = 0$) and locations computed with the best layered model: (A) Epicentres; (B) Hypocentres along E-W vertical section. Blue crosses are shallower relocated hypocentres and red diamonds deeper ones.

Figure 5.12 Epicentral map showing shifts between original (yellow) and relocated (red) epicentres.

Figure 5.13 CSZ events recorded at station DAQ; (A) Hypocentres;
(B) Distribution of P and S residuals.

B

Figure 5.14 Wadati plot for the CSZ events recorded on DAQ.

Figure 5.15 V_p - V_s tests for the DAQ data with RMS values in seconds.

Figure 5.16 Best RMS for DAQ data obtained for given P-wave velocities and depth of layers.

Figure 5.17 Distribution of DAQ residuals for P and S waves grouped in 0.05 s classes.

Figure 5.18 Distribution of the P and S residuals grouped in 0.05 s classes for the CSZ earthquake data recorded on DAQ between November 1994 and March 1998.

Figure 5.19 Wadati plot for CSZ events recorded on DAQ between November 1995 and March 1998.

Figure 5.20 RMS calculated for crustal P and S velocities from CSZ events recorded on DAQ between November 1995 and March 1998.

Figure 5.21 Best RMS values at given depth and P velocities for a layered velocity model, using CSZ events recorded on DAQ between November 1995 and March 1998.

Figure 5.22 P and S residuals computed with the best layered velocity model, using CSZ events recorded on DAQ between November 1995 and March 1998.

Figure 5.23 Comparison of the computed 1-D velocity models with other models of the Charlevoix and Saguenay regions (North and South shores from Lyons et al., 1977; Saguénay from Somerville et al., 1990).

Chapter 6

Summary and discussion

6.1 Summary of the findings of this thesis

The CSZ is the most active seismic zone of eastern Canada. The region contains geological faults of the St. Lawrence paleo-rift system and of a Devonian impact structure. Between October 1977 and December 1997, monitoring of the earthquakes by a local network has defined a 30 by 85 km active zone of elliptical shape with major axis along the St. Lawrence River. Hypocentral depths are distributed from near the surface to 29 km, with about two thirds between 5 and 16 km. A prominent weakly seismic volume exists along the North Shore of the St. Lawrence River. Most shallow (< 5 km depth) and deep (> 20 km) events occur in specific subareas of the CSZ. Similarly to the pattern observed for the 1924-1978 period, 6 out of the 7 magnitude > 4 events in the period 1977-1997 occurred in the SW and NE ends of the CSZ. The improvement in the recording capabilities of the local seismograph network has brought the number of located earthquakes from about 100/year in the late 70's to more than 200/year since 1995. The number of located earthquakes varies between 0 and 30 per month, with no obvious correlation with magnitude > 4 occurrences, at either the regional or local level. Using microearthquakes recorded both on regional and local stations, an empirical relation

between the Nuttli magnitude (m_N) and the Richter magnitude (M_L , used for events only recorded locally) was defined: $m_N = 0.63 M_L + 1.41$ (Section 1.4.3). With a unified magnitude scale, the CSZ magnitude-recurrence curve was extended down to magnitude m_N 1.5, which is the estimated magnitude for completeness of reporting for the 1977-1997 period. On a magnitude recurrence curve, the rate of magnitude > 4 earthquakes exceeds the level suggested by the lower magnitude events recorded between 1977 and 1997. This may suggest that the magnitude > 4 events have overestimated magnitudes, or that there is a real change in occurrence rates at higher magnitudes.

On the basis of the heat conduction equation and Grenvillian surface heat flow, the temperature at the cutoff depth of CSZ earthquakes ($D_{99\%} = 25$ km), has been estimated to be between 215 and 355°C (Chapter 2). These temperatures and the inferred mafic mid- and lower-crustal compositions imply a brittle-ductile transition deeper than 25 km. The transition could occur at 25 km only for surface heat flow higher than the average Grenville value (in the 50 mW m⁻² range). Currently, two low quality measurements suggest such a high value. With the assumed absence of quartz at mid- and lower crustal depths, the earthquake cutoff depth may correspond to the passage from velocity strengthening to velocity weakening if the onset of flow of hydrated feldspars occurs at temperatures in the 300-350°C range. With an assumed maximum crustal stress difference of about 200 MPa, fault reactivation at mid-crustal depth can occur with a low friction coefficient and/or with a high pore fluid pressure, or with a change to a high-pressure type of fracturing (see also discussion in Section 6.2 in this chapter).

A series of normal faults associated with the St. Lawrence paleo-rift system breaks

the CSZ crust. On land, most of these faults correspond to lineaments in the remote sensing imagery. Some of these faults, but not all, have geophysical expressions in the magnetic and gravity fields (see discussion in Chapter 3). Offshore and under the Appalachian sequence, some regional-scale normal faults are revealed in the Precambrian basement, using magnetic, gravimetric and seismic reflection data. The structure of the CSZ was studied along four profiles perpendicular to the St. Lawrence River, revealing regional as well as local characteristics. Long normal faults, such as the St. Lawrence faults, and the newly-revealed Charlevoix and South Shore faults, cut across the whole CSZ. The Charlevoix fault is a few kilometres offshore and parallels the north shore. It may represent the structure reactivated by some of the largest CSZ earthquakes, including the 1925 M 6.2 earthquake. The South Shore fault appears to bound the seismic activity to the SE. The attitude of the Precambrian-Appalachian interface changes along the strike of the St. Lawrence River. The NE part of the CSZ, where most magnitude > 4 events occur, is an area with linear normal faults defining a series of steps in the Precambrian basement. The impact crater, on the other hand, coincides with an interpreted graben under the river, possibly created by the impact readjustment. The SW part, which is weakly seismic, is cut by numerous normal faults parallel and perpendicular to the river's trend. It is suggested that the Charlevoix fault could be one major reactivated fault, while the St-Laurent and South Shore faults bound highly fractured blocks. Reprocessing of a seismic line revealed deep reflectors in the Precambrian basement.

Although the CSZ has been subject to at least five earthquakes of magnitude 6 or greater, no surface rupture was found under the St. Lawrence River (see Chapter 4). A

systematic search for surface faulting was done on seismic reflection lines for oil exploration and for Quaternary mapping. Some 20 CSZ focal mechanisms for magnitude > 2 events show mostly reverse to reverse-oblique faulting in the CSZ, with no preferential orientation of the P-axis, but with a higher number of nodal planes in the NE quadrant. Local variations in the stress system appear to occur across the CSZ. Most earthquakes occur in clusters of two events or more, that may or may not have preferred orientation. One of these groups, located under the St. Lawrence, contains about one third of CSZ events, including the 1979 $m_N 5$ event. It defines one of the best alignments of hypocentres, possibly related to the newly-defined Charlevoix fault. Comparison of these groups with focal mechanisms suggest multiple orientations for the reactivated faults. For the period November 1988 to August 1997, cross-correlation of seismic traces has revealed that less than 15% of CSZ events are multiplets. Most of these events occurred within a few days of each other. Comparing earthquakes groups and multiplets with focal mechanisms has revealed some fault planes. Reactivated structures can be of local or regional importance, with orientations not necessarily sub-parallel to the Iapetan rift faults. Since most earthquakes are small, the actual dimension of the fracture, or reactivated fault surface, is small (generally in the tens of meters in radius). This aspect, together with the varied orientations of the fault planes, suggest that most earthquakes occur in highly fractured volumes. Study of the various sub-zones of the CSZ also suggest that faults with varied orientations can be reactivated, including some related to the impact crater.

The microearthquake data of the local Charlevoix network was used in an inversion procedure to determine the local velocity structure (Chapter 5). Using some 171

events recorded between November 1988 and December 1997, the best 1-D crustal model for the CSZ is very close to the “standard” GSC model, i.e., 6.2 km/s and 3.57 km/s P- and S-wave velocities. Station corrections, defined for the local network, imply lower velocities inside the impact crater. The data were also used to obtain a pseudo 2-D velocity model, i.e. two velocity models corresponding to the north and the south shores respectively. The model resolves the Appalachian sequence of the south shore (6 km thick sequence with velocities in the 5.6 to 5.8 km/s) and the Precambrian basement (increase in velocity between 8 to 14 km depth from 6.2 to 6.6 km/s). The best 2-D velocity model reduces significantly the RMS of the solutions. With the improved velocity model, locations of most CSZ events move towards the SE by about 1.5 km. Using CSZ events recorded on the regional station DAQ, a 1-D velocity model was defined for the Laurentides Park region, to the West of the CSZ. The model shows a gradual P-velocity increase from the surface to 10 km from about 5.9 to 6.7 km/s. The CSZ velocity structure is different from that of the Laurentides Park. While the V_p/V_s ratio is 1.73 for the CSZ, the V_p/V_s is higher (≈ 1.81) in the Laurentides Park. Lithology and possibly the degree of fracturing appear to be the main factor controlling these variations.

6.2 Why do earthquakes occur in the CSZ?

Over the last 30 years, a number of studies have proposed factors that may lead to CSZ earthquakes. The proposed causative factors can be grouped in four classes. First, some studies refer to the inherent geological weakness of the CSZ, either primarily due to the rift faults (Adams and Basham, 1991) or to the combined meteor impact-rift faults (Roy and Du Berger, 1983; Anglin, 1984; Lamontagne, 1987). A second group refers to

the combination of high stress difference and crustal weakness: either postglacial rebound-impact crater (Leblanc et al., 1973; Leblanc and Buchbinder, 1977; Anglin and Buchbinder, 1981); mafic intrusion-impact crater (Stevens, 1980); or stress concentration-impact crater (Lyons et al., 1980; Hasegawa and Wetmiller, 1980). A third group invokes mainly the effect of stress modifiers: postglacial rebound stress (that can favour thrust earthquakes in the CSZ: James and Bent, 1994; Wu and Hasegawa, 1996); crustal subsidence in the Laurentides Park (which is anomalous compared with the postglacial uplift witnessed in eastern Canada: Frost and Lilly, 1966⁸); or the effect of past major earthquakes (Bent, 1992; Ebel, 1998). Finally, some refer to possible anomalies in the rock properties such as high pore-fluid pressures at depth (Hasegawa, 1986; Sibson, 1989; Zoback, 1991; Lambert et al., in prep.) and/or low coefficient of friction (Lamontagne and Ranalli, 1996).

The Mohr diagram can be used as a graphic illustration of the factors leading to shear failure (Figure 6.1). Sliding occurs when the Mohr circle is tangent to the Coulomb envelope, i.e. when the shear stress exceeds the strength along the plane of failure. This instability can occur by changing the ambient stress field and/or the stability conditions of the pre-existing fracture. First, the Mohr circle (i.e. the stress conditions) can be brought closer to the failure envelope by modifying either its radius (i.e. the stress difference) or its position. The stress difference can be increased with a larger principal

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However, a recent re-evaluation of this subsidence has shown that it is within the error margins of measurements, and consequently of doubtful significance (T. Lambert, pers. comm.).

compressive stress (σ_1), or inversely, by a decrease in the least compressive stress (σ_3). The Mohr circle can approach failure (i.e. a shift to the left) with higher pore-fluid pressures, a process responsible for reservoir-induced seismicity. Secondly, fault reactivation can occur if failure conditions are modified. The failure envelope can be brought closer to the Mohr circle, by lowering the cohesion (S) of the pre-existing fractures, or by lowering the coefficient of friction (μ). In the following sections, I discuss separately the factors affecting the stress system, and those affecting the failure strength of rocks in the CSZ. A summary is given in Figure 6.2.

6.2.1 The stress system in the CSZ

In Eastern North America, the inferred maximum horizontal compressive stress axis correlates well with plate tectonics interpretations (Zoback and Zoback, 1991). Over most of the so-called mid-plate stress province, plate driving and resisting forces are primarily responsible for the predominantly ENE orientation of the maximum compressive stress axis. The reorientation in the CSZ may be due to glacioisostatic effects (Wu, 1998). The tectonic stresses can be amplified in the brittle upper crust leading to absolute levels in the 100 MPa range (Hasegawa et al., 1985). Thus, over most of eastern North America, the fundamental situation could be described by the Mohr diagram (Figure 6.1). The situation is stable with stress differences arising from the plate-wide forces which are generally too small to cause reactivation. Unstable situations, such as the one existing in the CSZ, can come about by adding local sources of stress to the ambient stress field (such as flexural stresses and lateral density contrasts/buoyancy forces) and/or by concentrating existing stresses.

Flexural stresses are caused by changes in the loads on or within the lithosphere (erosion, sedimentation, postglacial readjustment). Of these three, only postglacial rebound is significant in continental Eastern Canada. In areas formerly covered by continental glaciers, glacial rebound stresses can perturb the ambient stress field for thousands of years after glaciers have retreated. Interestingly, the change in the faulting style from reverse faulting in Eastern Canada to strike-slip faulting in the eastern United States occurs approximately at the limit of the former glacial maximum (Zoback, 1992; James and Bent, 1994). Consequently, it is evident that postglacial rebound stresses are perturbing the eastern Canadian stress field. The level of induced stresses, however, is low: the deviatoric stress field is increased only by about 5 to 10 MPa (Wu and Hasegawa, 1996). In the CSZ, the radial component of the postglacial stress field is compressive, oriented NW-SE and larger than the transverse component (James and Bent, 1994). Due to their orientation and small magnitude, postglacial stresses may contribute to thrust faulting in the St. Lawrence Valley (Hasegawa et al., 1985; Wu and Hasegawa, 1996; James and Bent, 1994). In other words, they can favour reverse faulting earthquakes, without constituting the primary cause.

Lateral mass anomalies within or just beneath the lithosphere are additional sources of stress. In the CSZ, upper crustal density variations exist between the Canadian Shield and the Appalachians (evident in Bouguer anomaly maps, e.g. Figure 3.7B). Two approaches have been used to model their impact on the local stresses. In the first one, an elastic model suggested near vertical maximum compressive stress corresponding to the negative gravity anomaly (which in the CSZ case, is approximately the St. Lawrence

River; Goodacre and Hasegawa, 1980⁹). In the second approach, an elastic multi-layer FE model was used to compute stresses induced by lateral density variations (Assameur and Mareschal, 1995). At 10 km depth, average stress difference can reach about 27 MPa in regions with the highest density contrasts (east of the Grenville Front, Thetford Mines and the lower St. Lawrence). Of these three areas, only the lower St. Lawrence is seismically active. In the CSZ, induced stress differences are about 10 MPa at 10 km depth. Assuming a 20 MPa horizontal compression oriented N60°E, the ambient stress difference increases to 24 MPa when the orientation of the local stress field is considered (Assameur and Mareschal, 1995). Other sources, such as crustal thinning and lithosphere thickening, should not apply to the relatively constant crustal thickness in the CSZ (as inferred from seismic refraction; Lyons et al., 1980). The fact that regions with the highest induced stress differences are mostly aseismic suggests that mass anomalies cannot be the main factor of the CSZ seismicity.

At least two factors can increase the crustal stress difference: inclusions of materials with elastic parameters different from the surroundings, and fault intersections. Stresses can concentrate around a weak inclusion in a rigid and brittle material. Weak hydrated materials such as serpentinized gabbros at mid-crustal depth are particularly good stress concentrators due to the high rigidity contrast they provide with the surrounding rocks (Campbell, 1978). Appropriate orientation of the long axis of the

7 In a thrust environment, additional vertical stress translates into more stable faults. In most papers, paired gravity low and high are usually referred to as a perturbing addition to the stress field, without mentioning the additional stability obtained in a reverse faulting environment. Consequently, the Canadian Shield-Appalachians boundary should not enhance earthquake probability in the CSZ.

intrusive with respect to the stress field and large size of the intrusive also enhance the effect. In Eastern Canada, it has been noticed that large intrusions such as the Morin anorthosite in the Western Quebec Seismic Zone and the Sept-Iles layered mafic intrusion in the Lower St. Lawrence Seismic Zone, are apparently aseismic areas in epicentral maps (Lamontagne et al., 1989; Lamontagne et al., 1994). In the CSZ, the massive St-Urbain anorthosite body does not seem to act as a stress concentrator; it is not seismic and it is not surrounded by earthquake activity. However, it is possible that a highly fractured volume (i.e. the volume beneath and around the impact crater) may have average elastic parameters lower than the surrounding, less fractured, rock. The analysis by Campbell (1978) shows that, if the body is elliptical in plan view with long axis perpendicular to the direction of maximum compressive stress, the largest stress differences occur just outside the ellipse along the prolongation of its long axis. This situation shows some similarity with the one in the CSZ, where the largest earthquakes occur at the two extremities of the area. The possible effect of contrast in elastic moduli cannot therefore be excluded.

Upper crustal stress concentration of 10 to 100% can arise due to the strength decrease of a localized area of the lower crust subject to ductile creep (Long and Zelt, 1991). This idea is somewhat different than the model of Hasegawa et al. (1985) which showed upper crustal stress amplification for plate-wide motions. In Chapter 2, the strength of a ductile layer was shown to vary with the lithology, the fluid content and the heat flow. In the CSZ, while lateral heat flow variations are unlikely, lateral changes of ductility in the lower crust cannot be excluded.

Stresses can also concentrate around intersecting faults (Talwani, 1988). In a rock subject to stress difference, stresses build up in fracture tips. Stress concentration is highest at the locked portions of intersecting faults, reaching eight times the ambient stress level (Talwani, 1988). The intersecting fault model could explain the high earthquake activity in the highly fractured zone within the impact structure. The intersection of these fractures could lead to enhanced stress levels, favouring ruptures, whereas areas with healed fractures would not give rise to the stress build-up necessary to cause an earthquake.

6.2.2 Physical conditions of the faults

On the whole, local sources of stress are relatively modest contributors ($\sigma_1 - \sigma_3 \leq 10\text{--}30$ MPa) to the CSZ stress field. Since these contributions are about the same everywhere in Eastern Canada, the CSZ faults must be inherently weak compared to aseismic faults elsewhere. This could be accomplished by either a decrease in friction, an increase in pore-fluid pressure, or a combination of both.

6.2.2.1 Coefficient of friction

From a fault mechanics point of view, a low friction coefficient decreases the fault strength, making it susceptible to reactivation at plausible stress difference values (≤ 200 MPa). As shown in Figure 2.9B, without near-lithostatic pore-fluid pressures, a fault with a coefficient of friction $\mu = 0.75$ is too strong to be reactivated in the current stress field. A low coefficient of friction (for instance, $\mu = 0.3$) makes the faults weaker, i.e. more susceptible to reactivation, and allows the reactivation of steeply-dipping faults ($\approx 70^\circ$,

as in the CSZ). Although the coefficient of friction is assumed to be decreasing with increasing depth as illustrated by the shallower slope of the Mohr envelope with increasing pressure (Figure 6.1), values of μ of less than 0.3 cannot be reached without a fault gouge consisting of hydrated clay minerals (Lockner, 1995).

Fault gouges are documented along regional faults of the CSZ. On the St-Laurent paleo-rift fault, for example, a fault gouge exists and was probably created in the normal faulting regime that prevailed in mid- to upper-Ordovician times (Rondot, 1979). The fault is a highly fractured zone, tens of meters wide, and includes breccias and diabase veins (Brassard, 1990). Down to at least 400 m (maximum depth of the well), the fractures are interconnected and linked to the surface, implying relative permeability. Alteration (silicification, chloritization and carbonatization) is present. In the Quebec City region, St. Lawrence rift faults show hydraulic breccias created during the opening of the Iapetus Ocean when relatively impermeable rocks of the Precambrian basement were shattered by the fluid pressure exceeding the rock strength (Lachapelle, 1993).

Although fault breccias are found along the St. Lawrence paleo-rift faults, the coefficient of friction on these faults is probably not sufficiently small as to bring the crustal strength in the 200 MPa range. A coefficient of friction of 0.3 corresponds to clays and other sheet silicates inter-layered with water, minerals that dehydrate and lose their lubricating properties at mid-crustal depth (Lockner, 1995; Scholz, 1990). Although the weakness of CSZ faults cannot be entirely due to the breccias, CSZ earthquake epicentres correspond to the locations of these breccias on the St. Lawrence rift system (Lachapelle, 1993). This may be a partial explanation of the confinement of CSZ earthquake clusters

along the St. Lawrence River, and not further inland in the Precambrian Shield. The fault gouge, mapped on numerous rift faults, may sufficiently lower the coefficient of friction to favour reactivation. It could, for example, lower it from high values found with rock on rock friction ($\mu \approx 0.85$) to lower values associated with smoother surfaces ($\mu \approx 0.5-0.6$).

6.2.2.2 Pore-fluid pressure

As discussed in Chapter 2, high pore-fluid pressures (near lithostatic level) can bring the crustal strength to within plausible levels (≤ 200 MPa), even with an average value for the coefficient of friction ($\mu \approx 0.75$; Figure 2.9B). Worldwide, high pore-fluid pressures appear to explain the relative crustal weakness of seismically active zones, the San Andreas fault being the most prominent example (Rice, 1992).

In the CSZ, evidences for crustal fluids exist¹⁰. In the late 70's, fluid flow along fault zones was suggested as an explanation for the temporal disappearance of a micro-gravity anomaly and a change in P-wave velocities (Buchbinder et al., 1988). The anisotropy of travel-time changes in the CSZ observed between 1979 and 1980 was explained by the closing of saturated cracks (Kirsch et al., 1987). Finally, near La Malbaie, magneto-telluric anomalies deeper than 1400 m can be explained by water or

⁸ Fluctuations of the water table creating hydraulic pulses were proposed as a means to trigger earthquakes in a prestressed environment at the limit of failure (Tsoflias et al., 1995). Since water levels were measured near Montreal, one may wonder if these measurements apply to the CSZ, some 300 km downstream, where the St. Lawrence River is 20 km wide. The diffusion of the water pressure fluctuations to mid-crustal depths implies an unlikely interconnection of the pores from the surface down to 20-25 km. In any case, the pore-fluid pressures could only be hydrostatic in such a model.

solutions in a zone of high porosity (Chouteau, 1985).

In intraplate environments, such as the CSZ, the sources of these near-lithostatic pore-fluid pressures at mid-crustal depths are subject to debate. In the upper crust, the recognized sources of high pore-fluid pressures are compaction of saturated sediments overlain by low-permeability rocks, and dehydration reactions in metamorphism (Scholz, 1990). These two mechanisms can hardly apply to mid-crustal depth in the CSZ: no sediments are present there and the absence of local tectonic activity does not support active metamorphism. According to some petrological arguments, lower crustal rocks are essentially dry due to hydration reactions that occur at temperatures above 250°C (Frost and Bucher, 1994). This opinion, however, is not shared by all (Hyndman and Shearer, 1989). We can speculate that fluids exist along fault zones, from evidence such as along the San Andreas fault (Sibson et al., 1988). Within the fault zone, high pore-fluid pressures exist, as opposed to the surrounding dry rocks. The source for these fluids could be the ductile roots of faults (Rice, 1992) or the lower crust (Sibson, 1992). For the CSZ mid-crust, two possibilities exist: fluids may be trapped by ancient tectonic processes (but the trapping of these fluids for millions of years remain enigmatic); or fluids may originate from below, possibly within the ductile layers of the lower crust (Rice, 1992) or the mantle (Sibson, 1989). Evidence for the deep origins of these fluids includes the distribution of CO₂-rich springs in present-day seismic belts (Irwin and Barnes, 1980) and the anomalously high proportion of gases associated with deep origin (such as helium). These anomalies are interpreted as chemicals from the mantle that percolate to the surface through the region's rift faults (from a yearly report of activities by the GHK Group,

1988).

Once present in the middle crust, fluid pressures must build up to reach near-lithostatic values to cause fault reactivation. This can be achieved in permeable volumes capped by impermeable material. The motion on the fault breaks the seal and allows fluids to migrate upwards, in a fault-valve fashion (Sibson, 1990). One can speculate that the paleo-rift faults of the St. Lawrence rift system represent the conduits to these fluids. As shown by the mapped hydraulic breccias, the Precambrian Shield (outside the meteor impact) is relatively impermeable and restricts fluid flow to these fault zones. In addition to increasing the pore pressure, the fluids can weaken fault zones by chemically corroding gouge material, lowering the coefficient of friction.

6.3 Discussion

The deviatoric stress field in the CSZ does not appear anomalously high compared to that of eastern North America. As elsewhere, the CSZ is subject to the mid-plate ambient stress field that provides a large proportion of the deviatoric stress. The CSZ stress difference may be marginally higher than elsewhere due to postglacial rebound, a factor contributing to reverse faulting (James and Bent, 1994). Mass anomalies and stress concentration factors (inclusions with different moduli and fault intersections) can modestly contribute to the stress field. Since most of these conditions exist throughout the Grenville Province, their impact in the CSZ can only be related to the inherent weakness of the mid- to upper crust.

Since the CSZ stress field is comparable to surrounding areas, the Charlevoix

earthquake activity is likely due to a local crustal weakness, caused by a low coefficient of friction and/or high pore-fluid pressures. A friction coefficient in the 0.2 to 0.3 range, necessary to bring the fault strength to plausible levels, is improbable due to the dehydration of gouge clay minerals at mid-crustal depths. The presence of a fault gouge may, however, bring the coefficient of friction towards the low end of the measured values ($\mu \approx 0.5-0.6$). High pore-fluid pressures appears as a likely explanation to many characteristics of the CSZ, such as the low crustal strength (Chapter 2), the sub-zones of enhanced activity (Chapter 4), the tendency of aftershocks to be shallow and outside the immediate rupture of the main shocks (Section 4.6.2.1), the variations in the stress field and in the reactivated fault orientations at the sub-zone level (Section 4.6), and the mismatch of the orientations of reactivated faults with respect to the mid-plate stress field (Zoback, 1992).

In order to take into account the seismic characteristics of the CSZ, any seismotectonic model must be centred on crustal weakness, created in all likelihood by high pore-fluid pressures. Such a model is proposed in Figure 6.3. First, fluids are assumed to exist along some of the major rift faults. The exact source of these fluids is uncertain but they could ascend from the mantle or possibly from the ductile lower crust (Chapter 2). One may speculate that the approximate linear zone defined by the 20 km depth earthquakes (Figure 1.21E) represents the mid-crustal conduit where fluids ascend to the surface. These fluids migrate to mid-crustal levels, where locally-sealed zones cause local over-pressures (Sibson et al., 1988). The over-pressures, mainly found near the major faults, can reach near-lithostatic levels, inducing fault weakening and favouring

reactivation. When an earthquake is triggered, the motion on the fault breaks the seal, allowing fluids to migrate upwards, in a fault-valve fashion (Sibson, 1990). If the shock is sufficiently large, aftershocks can be induced where the migrating fluids create instability conditions. Where the surrounding rocks are impermeable, due to fault healing for example, the fluids concentrate, and remain, along the rift faults. This situation appears to be found outside the impact crater, where hypocentres define a clear alignment, and where most magnitude > 4 earthquakes occur. Where highly fractured zones exist, fluids diffuse in all directions, locally increasing the pore-fluid pressure and eventually creating numerous but small earthquakes. This situation is found within the highly fractured zone inside the crater where hypocentre groups are diffuse, and active faults extremely variable in orientations (as shown by focal mechanisms, earthquake groups and multiplet analysis). Near the central peak of the impact crater, the dense network of intersecting faults can also amplify the local stress level favouring earthquake occurrences.

Outside the CSZ, the lack of seismicity could be due to a combination of factors such as the absence of fluids or of a conduit if fractures are sealed or without fault breccias. The absence of breccias can possibly explain the boundaries of the CSZ. To the NW of the CSZ, the stable craton was never reactivated after the Grenvillian orogeny, whereas the CSZ crust was reactivated in the Paleozoic. According to Wheeler (1996), the SE boundary of the seismicity corresponds to the transition from relatively intact Precambrian rocks to the NW to a zone with thinner Precambrian crust to the SE where faults have been healed. Outside the CSZ, but still along Iapetan rift faults, a much slower

rate of pore-fluid pressure build-up (and possibly a lower level of fracturing) could explain the rarity of earthquake occurrences. Using the CSZ model, fluid flow from the lower crust can explain the occurrence of two deep earthquakes: the 1988 Saguenay earthquake (29 km depth) and the 1997 Cap-Rouge earthquake (22 km depth).

The absence of any apparent CSZ surface rupture suggests that the current rates of seismic strain release are geologically recent. Had these rates existed over millions of years, evidence of this activity would have been mapped in the field and seen in the seismic reflection profiles. The earthquake activity likely started some thousands of years ago, possibly right after the last glaciation, when conditions were most favourable for reverse faulting in the CSZ (Wu and Hasegawa, 1996). In all likelihood, the current conditions that create CSZ earthquakes may persist for thousands of years in the future. The current stress field depends mainly on factors that change on geological time scales, such as plate motions, and to a lesser extent, postglacial rebound and other stress contributors that will continue to play a role for thousands of years to come.

Figure 6.2 Factors that can lead to earthquakes in the CSZ with the section numbers where they are discussed. 330

Figure 6.3 Schematic model of the CSZ earthquakes shown on a cross-section perpendicular to the St. Lawrence River. Major faults in the area are presented as permeable (solid black lines) or impermeable (i.e. healed; dotted lines). Fluids can move along the impermeable faults, represented as zones of enhanced fluid pressures (in blue). The origin of these fluids could be the ductile layers in the lower crust or the mantle. The fluids ascend towards the surface, triggering earthquakes where temporary seals halt their progression, causing local fluid-pressure build up. Near the surface, within the impact crater, the fluids migrate into highly fractured areas, causing repetitive, but small magnitude, earthquakes. See text for details.

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Appendix 1

Example of a focal mechanism calculation and plots

Example of the output provided for each focal mechanism in Lamontagne (1998). The 971028 11:44 U.T. event is used: 1) Text; A) The FOCMEC results in five focal spheres (Figure 12A). Starting in the upper left and going clockwise, the first sphere shows the series of solutions obtained with P and S_H first motions (entitled "*P, S_H and P-Nodal*"). The second shows the solutions obtained using the P first motions only (entitled "*Solution with P FM only*"). Unless well constrained with P first motions only (rather rare), the increment for the b axis is 5° . The lower right sphere shows the S_H first motions and the S_H "nodal" lines computed with the P and S_H solutions (entitled " *S_H motions and Lines*"). The lower left one shows all first motions (entitled "*First Motions Only*"). Finally, the central mechanism is the average solution using the P and S_H families of solutions (entitled "*Best Solution*"). One should also note that the exact position of the first motion on the focal sphere is not in the centre but in the lower left portion of the C and D first motion symbols; B) First motion picks; C) Location map of the earthquake (focal sphere) and background seismicity.

Earthquake of 19971028 1144 Magnitude 4.7MN

1- Earthquake information and list of first motions.

Date	Time	Latitude	Longitude	Depth	Mag	Sta/Pha
hhmm ss				(km)		
971028	1144	18.64	47.6721	-69.9051	11.42	4.7MN 6/012

COMMENTS:

CHARLEVOIX SEISMIC ZONE, QUE. Felt.

Felt in Charlevoix, Riviere-du-Loup, and in the Quebec City, Saguenay and Lac St-Jean regions.

18 km E from LA MALBAIE, QUE.

\$ LMQ; SLQ; QCQ down.

\$ CIQ FM from R. Du Berger.

\$ A11 starts with weak C, which is followed by an impulsive D.

\$ Ioseismals to come.

\$: REV is F. Revetta, Potsdam, N.Y.

\$ESDC SZ P XB1152 49.700 L

\$ARCE SZ P XB1152 58.300 L

\$LPAZ SZ P XB1154 51.300 L

\$DBIC SZ P XB1155 21.000 L

\$ZAL SZ P XB1156 9.800 L

\$ The Sg lines of some stations include the SH FM.

\$ The convention used is the Virginia Tech convention used in FOCMEC:

\$ <: SH first motion to left (back to event) impulsive

\$ >: SH first motion to right (back to event) impulsive

\$ Preliminary Focmec computed by ML.

\$ Note: most CLTN Sg FM are unclear.

Sta	Pha	Dist	Azim	Take-off	FM
		(km)	Angle		
A61	PG	14	280	-51	D
A61	SG	14	280	-51	>
A21	PG	17	78	-55	D
A64	PG	17	3	-56	D
A16	PG	24	199	-64	C
A16	SG	24	199	-64	>
SHQ	PG	38	259	-73	D
A54	PG	45	238	-76	D
A11	PG	53	205	-78	C
CIQ	PG	94	316	-83	C
DAQ	PG	105	288	-84	D
JOQ	PG	129	310	0	C
DPQ	PG	244	244	-87	D

Number of P first motions: 11

Number of SH first motions: 2

2- Most Representative Solution.

A total of 247 solutions were found using a b axis increment of 3 degree. The following parameters describe the most representative solution, which was derived from the average B, P, and T axes listed below.

dip,strike,rake	85.99	111.73	41.84			
dip,strike,rake	48.28	18.15	174.62	:auxiliary plane		
lower hem. trend, plunge of a,n			288.15	41.72	21.73	4.01
lower hem. trend & plunge of b			116.20	48.00		
lower hem. trend, plunge of p,t			237.24	24.90	343.65	31.33

Average B, P, and T axes:

	B	P	T
Mean trend:	N 137	N 246	N 348
Mean Plunge:	49	16	37
Vector Magnitude:	242	244	245
Mean length of resulting vector:	0.981	0.987	0.991

3- Misfits.

All 11 P first motions agreed with the solution.

Out of 2 SH first motions, a total of 1 did not fit the solutions (50/100).

Station Name	Number of misfits	Percentage of wrongs
A16	247	100%

4- Rating of the focal mechanism.

The solution has a C rating based on the poor constraints on the P, T and B axes (value of 0.986).

Appendix 1

A

19971028 11:44 U.T.

P, SH and P-Nodal

Solution with P FM only.

Best Solution

First Motions Only

SH Motions and Lines.

P Phases on Vertical Components

SH phases on Transverse Components

Appendix 2¹¹

List of earthquake multiplets by sub-zone

yyyyymmdd hhmm ss.ddd	Lat	Lon	Depth	Mag	Distance (km)
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Sub-zone DNE

19900907 0406 32.800	47.7910	-69.8170	21.87km	0.6ML	
19920113 0904 11.800	47.7900	-69.8190	22.91km	0.4ML	1.057
19920716 0328 06.100	47.7920	-69.8330	22.10km	0.8ML	1.225
19961230 0246 23.479	47.7985	-69.8120	21.62km	-0.2ML	0.947
19970329 0653 59.603	47.7897	-69.8172	21.78km	1.2MN	0.171
19911208 0300 30.994	47.7796	-69.8640	23.20km	4.3MN	
19911208 0631 04.400	47.7850	-69.8640	22.77km	0.9ML	0.738
19911208 0857 00.800	47.7850	-69.8640	22.56km	0.7ML	0.877
19911227 2324 37.400	47.7820	-69.8650	22.70km	1.3MN	0.572
19920218 0149 45.811	47.7946	-69.8454	23.34km	1.7MN	2.177
19951209 2205 37.713	47.7940	-69.8614	23.26km	1.6MN	1.613
19921011 1859 02.900	47.8140	-69.9480	20.30km	0.2ML	
19930216 0842 57.300	47.8200	-69.9690	19.52km	0.2ML	1.877
19940602 1445 37.118	47.7178	-69.9481	25.44km	2.4MN	
19950615 1559 17.215	47.7155	-69.9275	25.39km	2.7MN	1.567
19961228 1704 01.809	47.8257	-69.7970	21.81km	-0.2ML	
19970309 0655 11.523	47.8227	-69.7947	20.90km	-0.7ML	0.984
19970217 1947 43.077	47.8069	-69.7711	23.40km	1.7MN	
19970219 2056 24.113	47.8094	-69.7641	23.89km	-1.0ML	0.769

Sub-zone IRE

19891123 1308 56.900	47.5420	-70.2980	11.34km	0.4ML	
19960813 2237 10.375	47.5468	-70.2956	11.67km	1.2MN	0.653
19900429 0014 29.320	47.6185	-70.1676	11.80km	1.7MN	
19960403 0434 24.519	47.6174	-70.1687	11.34km	-0.5ML	0.483

11 Blank lines separate multiplets.

19950311	1119	04.600	47.5280	-70.3170	9.43km	0.5MN	
19951228	2139	01.771	47.5372	-70.3003	9.41km	-0.3ML	1.621
19950820	0416	10.309	47.4998	-70.2956	14.09km	0.8MN	
19951117	0757	08.213	47.5173	-70.2949	15.33km	1.2MN	2.307
19960105	0425	57.522	47.6003	-70.2264	6.88km	1.5MN	
19960105	0440	30.215	47.6017	-70.2251	5.85km	2.0MN	1.046
19961004	2216	37.536	47.5965	-70.2076	14.41km	1.3MN	
19970422	1100	45.076	47.6065	-70.2164	15.14km	2.4MN	1.485
19900618	0700	16.600	47.5870	-70.2440	12.69km	-0.2ML	
19961115	2329	01.620	47.5989	-70.2453	12.28km	-0.3ML	1.388
19921203	1002	07.911	47.6099	-70.1753	11.84km	2.0MN	
19951108	1646	05.796	47.6045	-70.1832	10.65km	-0.4ML	1.459
19960403	0434	24.519	47.6174	-70.1687	11.34km	-0.5ML	1.091
19960924	2341	02.882	47.5475	-70.2417	12.83km	3.1MN	
19961003	1024	46.893	47.5515	-70.2373	12.83km	1.3MN	0.554
19961003	1032	24.470	47.5442	-70.2396	13.14km	1.3MN	0.506

Sub-zone M5

19890309	0941	32.262	47.7170	-69.8568	10.51km	4.3MN	
19890310	0329	59.500	47.7160	-69.8540	9.23km	1.2MN	1.302
19890311	0831	52.168	47.7181	-69.8700	10.36km	4.4MN	1.009
19900305	0025	12.100	47.6540	-69.9020	7.60km	0.1ML	
19970316	1104	16.007	47.6608	-69.8939	8.61km	1.6MN	1.400
19910221	1702	48.400	47.7020	-69.8830	9.09km	0.9ML	
19940311	0836	33.510	47.6929	-69.8959	9.45km	1.7MN	1.446
19920311	0120	57.600	47.7180	-69.8570	9.45km	0.2ML	
19920314	2128	43.602	47.7170	-69.8570	9.77km	1.6MN	0.339
19930330	1235	28.300	47.6740	-69.9080	10.31km	0.0ML	
19930407	1402	29.500	47.6770	-69.9100	10.70km	1.3MN	0.535
19930807	2125	31.304	47.6654	-69.8857	4.87km	3.1MN	
19931013	0143	13.507	47.6808	-69.8784	13.96km	0.4ML	9.266
19950812	1742	55.813	47.6680	-69.9035	10.83km	2.5MN	

19950918 1609 47.386 47.6678 -69.9033 10.63km 2.6MN 0.202

Sub-zone A61

19891208 1720 34.446 47.7009 -70.0644 10.37km 2.6MN
 19891221 1211 28.000 47.7010 -70.0730 10.79km 0.2ML 0.770
 19940614 2339 58.000 47.7040 -70.0700 10.12km 1.4MN 0.598
 19960215 0005 39.828 47.7011 -70.0454 16.27km 0.0ML 6.070

19930706 0633 33.200 47.6950 -70.0730 9.02km 0.6ML
 19950303 2334 37.800 47.6960 -70.0750 9.73km -0.4ML 0.734
 19950820 0241 42.030 47.6946 -70.0687 9.03km 0.1ML 0.326

19960803 1827 03.119 47.6744 -70.1431 9.79km 1.9MN
 19960804 0918 55.206 47.6763 -70.1457 9.45km 0.9MN 0.445

Sub-zone CEN

19881220 0246 18.800 47.5310 -70.0840 8.81km 0.5ML
 19970130 2150 08.704 47.5299 -70.0695 10.36km -0.8ML 1.900

19890203 1355 39.800 47.5720 -70.0210 11.08km 1.1ML
 19910608 0338 49.800 47.5720 -70.0230 10.64km 0.4ML 0.465

19900522 1221 09.700 47.5350 -70.0290 7.45km 0.5ML
 19920130 1042 33.400 47.5389 -70.0349 5.99km 2.0MN 1.586
 19920130 1352 11.615 47.5403 -70.0277 7.20km 1.9MN 0.647
 19930801 1935 29.717 47.5355 -70.0390 6.69km 1.7MN 1.071

19901018 0603 28.467 47.4817 -70.0868 15.22km 2.8MN
 19920817 1255 26.600 47.4830 -70.0910 14.19km 0.4ML 1.087

19910711 0858 11.600 47.4730 -70.0400 7.27km 0.4ML
 19910711 1451 05.600 47.4750 -70.0400 7.20km 0.3ML 0.233

19911008 0411 02.800 47.5950 -69.9900 8.47km 0.2ML
 19920627 2343 25.900 47.5910 -69.9940 8.03km 0.5ML 0.694
 19940207 0609 01.442 47.5903 -70.0005 8.40km 2.0MN 0.949
 19970509 0906 28.311 47.5624 -69.9974 8.76km 1.3MN 3.677

19920404 1230 28.560 47.4325 -70.1733 18.92km 2.9MN
 19920420 0855 27.505 47.4333 -70.1748 18.90km 1.9MN 0.145

19920430 0401 41.000 47.4980 -70.0490 11.01km 0.3ML

19961222 0030 32.038 47.5059 -70.0637 10.83km 2.3MN 1.425

19920816 1851 25.300 47.5730 -70.0410 10.48km 0.4ML
 19960901 0517 18.200 47.5745 -70.0434 10.84km 1.4MN 0.436

19930605 0449 40.991 47.4931 -70.1126 11.03km 2.6MN
 19930606 0152 11.800 47.4940 -70.1130 10.41km -0.1ML 0.629

19950718 1306 38.413 47.5706 -69.9979 6.62km 1.9MN
 19950724 0328 01.314 47.5662 -69.9969 5.09km -0.3ML 1.608
 19951019 0316 07.603 47.5844 -70.0009 9.04km 2.1MN 2.874
 19970319 0953 12.399 47.5885 -70.0023 7.77km 1.5MN 2.322

19950719 0953 01.613 47.5675 -70.0007 13.66km -0.2ML
 19970401 1203 12.904 47.5652 -70.0017 13.81km 1.6MN 0.306

19950918 1322 03.731 47.4471 -70.1084 1.90km 0.6ML
 19970112 0523 24.741 47.4357 -70.0907 5.76km -0.7ML 4.276

19901221 0940 32.115 47.4583 -70.0715 8.52km 1.9MN
 19901221 0957 51.500 47.4560 -70.0680 8.75km 0.3ML 0.433

19910414 0508 35.300 47.5460 -70.0300 7.94km 0.1ML
 19930518 0017 41.713 47.5432 -70.0329 7.60km 2.4MN 0.510
 19940711 0935 41.143 47.5620 -70.0513 4.91km 2.8MN 3.862

Sub-zone COU

19920914 2038 14.800 47.4230 -70.4210 5.97km 0.8ML
 19970212 0541 31.702 47.4289 -70.4167 7.46km -0.2ML 1.660

19960824 0245 08.522 47.4013 -70.3200 11.47km 2.0MN
 19960824 0247 22.525 47.4017 -70.3196 11.63km 1.8MN 0.169

19961001 0049 03.173 47.4136 -70.3322 11.66km -0.6ML
 19961007 0950 11.871 47.4293 -70.3315 11.69km -0.4ML 1.746

19930814 0755 20.100 47.4170 -70.3380 10.29km -0.3ML
 19930814 1456 13.700 47.4170 -70.3320 9.28km 0.1ML 1.107
 19960518 0816 32.273 47.4208 -70.3143 9.84km 0.7MN 1.891

19910312 0657 08.946 47.3910 -70.4832 10.25km 1.9MN
 19950911 1410 21.046 47.3983 -70.4805 9.98km 1.7MN 0.879

19891122	2302	51.700	47.4558	-70.3430	7.61km	3.4MN	
19891231	0234	53.900	47.4510	-70.3430	7.63km	0.1ML	0.534
19900111	0715	12.208	47.4544	-70.3480	8.36km	1.8MN	0.854
19901026	1755	58.000	47.4290	-70.4090	6.86km	0.2ML	
19970212	0541	31.702	47.4289	-70.4167	7.46km	-0.2ML	0.835
19901021	1338	43.209	47.3975	-70.3640	15.82km	3.3MN	
19901102	0427	50.100	47.3970	-70.3610	14.93km	0.5ML	0.920
19960113	0732	50.103	47.4160	-70.4406	4.81km	1.9MN	
19960618	0144	24.387	47.4169	-70.4364	4.59km	0.8MN	0.399
19920311	1305	50.015	47.4578	-70.3570	7.03km	1.9MN	
19950502	1954	28.823	47.4493	-70.3766	7.72km	0.9ML	1.885
19920222	2155	32.800	47.4280	-70.3570	7.95km	0.3ML	
19920816	2106	58.000	47.4270	-70.3560	7.61km	0.2ML	0.366
19911102	0254	28.800	47.3850	-70.4410	9.24km	0.4ML	
19970409	0244	17.495	47.3785	-70.4647	11.35km	2.1MN	2.859

Sub-zone EBO

19900506	0204	43.500	47.5140	-70.3770	3.21km	0.9MN	
19940901	0444	43.300	47.5170	-70.3870	4.95km	0.5ML	1.925
19930417	0535	42.519	47.5293	-70.3672	7.55km	1.8MN	
19950401	0210	06.539	47.5435	-70.3978	6.33km	0.7MN	3.047
19950806	0436	28.823	47.5554	-70.3777	9.44km	0.8MN	
19961027	0647	37.598	47.5501	-70.3548	9.94km	-1.0ML	1.889
19951120	2111	25.907	47.4922	-70.4024	5.37km	1.0MN	
19960403	0123	52.212	47.4994	-70.3948	5.35km	1.4MN	0.984
19960518	0505	19.606	47.4687	-70.3891	5.82km	1.2MN	
19960530	0836	58.713	47.4632	-70.3824	5.54km	0.5ML	0.841

Sub-zone KAM

19900728	1437	10.000	47.6090	-69.9660	10.48km	0.4ML	
19920417	1330	24.600	47.6050	-69.9690	9.45km	0.6ML	1.144
19900916	2113	25.300	47.5630	-69.9580	11.04km	0.7ML	

19930823	0931	26.819	47.5603	-69.9423	11.89km	2.0MN	1.486
19901213	0350	53.100	47.5690	-69.9060	14.32km	0.4ML	
19911011	0659	02.600	47.5790	-69.9150	13.69km	0.3ML	1.446
19920526	0854	44.300	47.5940	-69.9270	11.64km	0.6ML	
19961224	1045	55.518	47.5904	-69.9320	11.99km	-1.0ML	0.651
19920609	1145	09.200	47.6050	-69.9210	12.20km	0.9ML	
19960920	1543	06.202	47.6065	-69.9199	12.24km	1.5MN	0.190
19910919	1106	53.170	47.5384	-69.9324	13.64km	2.6MN	
19920903	0704	54.400	47.5240	-69.9520	12.67km	0.5ML	2.383
19940910	2342	03.500	47.5270	-69.9520	12.60km	0.7ML	2.205
19960523	0939	27.217	47.5283	-69.9395	13.66km	1.7MN	1.243
19960607	0941	42.840	47.5299	-69.9417	13.31km	3.1MN	1.221
19950112	0324	08.120	47.6066	-69.9523	11.75km	2.4MN	
19960726	1438	45.149	47.6097	-69.9468	10.36km	2.3MN	1.491

Sub-zone POC

19900731	1510	35.938	47.3866	-70.2493	11.99km	0.6ML	
19921109	1757	13.600	47.3860	-70.2490	12.13km	0.1ML	0.157
19960205	0810	27.122	47.3849	-70.2561	13.47km	1.4MN	1.578
19960601	0201	42.306	47.3844	-70.2564	12.18km	-0.1ML	0.619
19960720	1649	21.345	47.3835	-70.2521	13.87km	0.2ML	1.923
19970315	1303	00.310	47.3853	-70.2520	12.38km	0.7MN	0.463
19940723	0734	25.500	47.3820	-70.1350	13.66km	1.4MN	
19940723	0946	41.208	47.3804	-70.1365	15.16km	2.0MN	1.515
19960625	0600	12.684	47.3778	-70.1384	14.57km	1.6MN	1.054
19960702	2359	31.804	47.3792	-70.1397	13.02km	1.0MN	0.795
19960803	0136	34.329	47.3812	-70.1387	12.94km	1.4MN	0.777
19960803	0159	20.822	47.3707	-70.1343	13.22km	0.9MN	1.332
19960814	2110	54.634	47.3816	-70.1371	13.90km	2.2MN	0.291
19961022	0247	22.775	47.3808	-70.1450	13.33km	1.7MN	0.835
19970101	1233	06.148	47.3783	-70.1342	15.55km	1.6MN	1.935
19891218	1455	35.616	47.3866	-70.1543	18.20km	2.8MN	
19970118	0149	25.479	47.3704	-70.1445	18.61km	-0.9ML	1.989

Appendix 3

Recommendations for future monitoring of the CSZ

The current CSZ seismograph network can be used to compute focal mechanisms for earthquakes as small as magnitude m_N 3.0. These solutions, however, are strongly dependent on the correctness of every first motion reading. In order to make the solutions more robust, the number of stations within 100 km of the seismic zone should be increased especially in the NE and SW quadrants. The 1996 summer field survey showed that with about six additional stations, focal mechanisms can be computed for events as small as magnitude 2.0. Clearly, additional CSZ stations are needed to derive well defined focal mechanisms from which seismo-tectonic conclusions can be drawn. The calculations of focal mechanisms with Sv/P ratios should be examined carefully to test its reliability. One could test how consistent the Sv/P ratios are in focal mechanisms well constrained with P first motions. For events below magnitude 4.0, P first motions are emergent on stations at regional distances and the velocity model between the CSZ and these stations is at best approximate. This causes problems in using Pn first motions to constrain CSZ mechanisms.

It is possible to resolve more accurately the relative arrival times at a station using the cross-correlation technique (Deichmann and Garcia-Fernandez, 1994; Li et al., 1995). The relative arrival times can be resolved at the sub-sample rate level. In the CSZ case, this method can only be applied to the data recorded between November 1988 and August 1994, i.e. when the three-component data was time-tagged at a central site. The current

scheme of GPS time tagging at the various CSZ sites does not provide the consistency required for such an analysis. In the future, the time series files for multiplets recorded between 1988 and 1994 could be correlated.

The data from the current seven local stations do not permit resolution of a 3-D velocity model, which could yield important seismotectonic information, such as lateral velocity contrast across geological faults. An estimate of the network requirements needed to infer a 3-D velocity model for the CSZ follows. The CSZ is a zone 40 km wide, 70 km long and 25 km deep. A velocity model with 2.5 km-sided blocks implies the determination of some 4480 blocks for P and for S velocities. The number of unknowns can be divided in half by assuming a V_p/V_s ratio. For each earthquake, four additional unknowns have to be defined: latitude, longitude, depth and origin time. The observed data, on the other hand, are the total number of stations, times the number of phases used (P and S) times the number of events. The minimum number of stations and events to solve a CSZ 3-D model can be approximated as follows

Number of observed data = number of unknowns

$$\text{Sta} * \text{Pha} * \text{Ev} = \text{NB} + 4 \text{ Ev}$$

Where Sta: Number of Stations; Pha: Number of Phases; Ev: Number of Events; and NB: Number of blocks.

Assuming Sta = 12; Pha = 2; and NB = 4680, the minimum number of events is 224. If one assumes an annual rate of 75 events with magnitude $m_N \geq 1.5$ (the minimum level for events to be recorded across the CSZ), then one would have to run the 12 station

network for a minimum of about 3 years. Naturally, additional stations mean a shorter time period to reach the objective, while providing a better sampling of the near surface layers. A better network also means more focal mechanisms and better hypocentre locations.

Therefore, the inversion of a local 3-D velocity model for the CSZ requires an expansion of the CSZ seismograph network from the current 7 stations, to 12 or more. During the 1996 summer experiment, up to 8 seismograph stations were operated simultaneously during a 6 month period (Lamontagne et al., 1997). A total of 142 events was recorded, of which 40 met the $m_N \geq 1.5$ selection criteria. The biggest drawback of the portable seismographs was the drift of the internal clocks between the weekly servicing. The drift was generally less than ± 0.05 s, which is still too high for local velocity inversion purposes. The expansion of the CSZ network should make use of portable seismographs with accurate timing. Fortunately, the next generation of portable seismographs will use continuous GPS timing. The joint inversion of the hypocentre-velocity structure is imperative to infer velocity variations and therefore imaging zones of possibly enhanced earthquake activity.