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Further studies on the 1988 MW 5.9 Saguenay, Quebec, earthquake sequence

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2 **earthquake sequence**

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Abstract:

Many small earthquakes occur annually in Eastern Canada, but moderate to strong earthquakes are infrequent. The 25 November 1988 M_W 5.9 Saguenay mainshock remains the largest earthquake in the last 80 years in eastern North America. In this article, some aspects of that earthquake sequence were re-analyzed using several modern methods. The regional depth phase modeling procedure was used to refine the focal depths for the foreshock, the aftershocks and other $m_N \geq 2.5$ regional earthquakes. The hypocenters of ten earthquakes were relocated using hypoDD. The spatial distribution of eight relocated hypocenters defines the rupture plane of the mainshock. The moment tensor for the mainshock was retrieved using three-component long-period surface wave records at station HRV (Harvard seismograph station) with additional constraints from P -wave polarities. One nodal plane is conclusively identified to be close to the rupture plane, and its strike is similar to the trend of the south wall of the Saguenay Graben. Based on the consistency between the strike of the nodal plane and the trend of the Graben, as well as the deep focal depth distribution, we suggest that the Saguenay earthquake sequence is related to the reactivation of one of the faults of the Saguenay Graben.

Key words: Saguenay Graben, 1988 Saguenay earthquake, hypoDD aftershock relocation, moment tensor, rupture plane, depth phase.

Introduction

The largest earthquake in Eastern Canada in the last 80 years occurred near Saguenay, Quebec, on 25 November 1988. The earthquake's location in a weakly-active seismic area, its mid- to lower crustal focal depth, and the large amount of high-frequency energy released were unusual. Figure 1 shows the seismicity surrounding the 1988 Saguenay M_W 5.9 earthquake and the extent of the Saguenay Graben. The earthquake occurred in the Grenville geological province of the Canadian Shield. Its epicentre is within the Jacques Cartier Bloc, an area wedged between the Saguenay Graben and the St. Lawrence Rift system (Du Berger et al., 1991). The area comprises mid-crustal metamorphic and igneous rocks such as mangerites and gneisses of Grenvillian age. Some major tectonic events occurred there, including continental collision, opening and closing of the Iapetus Ocean and the opening of the current Atlantic Ocean (Du Berger et al., 1991). The epicenter lay between the Saguenay Graben and the St. Lawrence River and was 75 km from the outer boundary of the Charlevoix Seismic Zone (CSZ). The CSZ was historically a source of five earthquakes of magnitude between 6 and 7 (North et al., 1989), and is one of the most seismically active zones in eastern North America. Due to the earthquake history and the presence of the faults of the St. Lawrence Rift System, both regions have significant hazard in the seismic zoning maps (Halchuk et al., 2015). An improved understanding of the 1988 Saguenay earthquake sequence may provide help to assess the seismic hazard in those regions.

The mainshock and its aftershock sequence were described elsewhere (North et al., 1989; Du Berger et al., 1991; Haddon, 1995; Boore and Atkinson, 1992; Somerville et al., 1990). The mainshock was preceded by a foreshock of m_N 4.6 on 23 November 1988 at 09:11:27.3 UT (m_N - Nuttli magnitude, Nuttli, 1973), which was strongly felt by the vast majority of the inhabitants of the Saguenay and Lac St. Jean areas. Since it occurred in a seismically inactive area, this event caused the Geological Survey of Canada (GSC) to deploy analog field seismographs. Three MEQ-800 portable seismographs were deployed within 10 km of the epicenter of the foreshock within one day of the event (North et al., 1989). With these field instruments and the

Charlevoix Local Telemetered Network (CLTN) located about 90 km to the southeast, the Saguenay epicentral area was relatively well monitored during the mainshock and aftershocks.

The focal depth distribution defines the depth extent and thickness of the seismogenic part of the crust. This information is useful for seismic hazard assessment. For small earthquakes, focal depths are generally computed using the arrival times of the P_g and S_g phases that are recorded at field stations within a few tens of kilometers. This approach was used for numerous aftershocks that occurred in 1988 and 1989 when field stations were in operation.

By the fall of 1989, however, the field stations in the epicentral region were removed and only one vertical short-period digital station remained within 80 km of the main shock. For events that occurred after the fall of 1989, earthquake epicenters were computed using data from that station plus the CLTN stations, the focal depth could not be reliably determined and a default focal depth of 18 km was assigned in the GSC earthquake catalog.

The regional depth phase modeling procedure (RDPM) is an alternative method to determine focal depths for moderate and small earthquakes. The errors in the interpreted focal depths are small if the identifications of the depth phase and its reference phase are correct. The procedure to use regional depth phases, such as sPg , $sPmP$, and sPn , to determine focal depths is described in Ma and Atkinson (2006), Ma (2010), and Ma and Motazedian (2016).

CLTN records of the sPg phase generated by most aftershocks with a magnitude of $m_N \geq 2.5$ and the sPg records of the foreshocks were used for this article. The waveforms for the events of $m_N \geq 2.5$ were retrieved, and a technique outlined by Ma and Atkinson (2006) and Ma (2010) was used to model the regional depth phase for 40 events. Focal depth solutions for 33 events were successfully determined.

The rupture plane of the mainshock shows the active fault beneath the surface in the epicentral region. A focal mechanism or a moment tensor solution revealed two nodal planes. The nodal plane that represented the rupture plane could be identified using the spatial distribution of the aftershocks if the hypocenter errors are small.

110 We achieved small hypocenter errors by relocating the hypocenters using the
111 hypoDD method (Waldhauser and Ellsworth, 2000; Waldhauser, 2001). This method
112 reduced the impact of path effects and increased the relative accuracy of hypocenter
113 locations. Ma and Eaton (2011) made a small revision to the technique to overcome
114 the trade-off between the focal depth and the epicenter.

115 The modeling of long-period surface waveforms can provide reliable nodal plane
116 solutions. Unfortunately, the only available regional broadband records generated by
117 the Saguenay mainshock were the three-component records at HRV. A surface
118 waveform modeling method and some P -phase first motions were used to determine a
119 moment tensor solution for the mainshock.

120 In the following sections, we outline the methods used in our work, and present
121 the focal depth solutions obtained using the RDPM procedure, the hypocenter
122 relocations for 10 events that occurred between 23 November 1988 and 18 April 1989,
123 and the moment tensor solution retrieved by modeling the long-period surface
124 waveforms and P -phase polarities.

125 The new findings, especially the rupture plane, identified from the two nodal
126 planes, may improve our understanding of this earthquake sequence and of intraplate
127 earthquakes in Eastern Canada.

128

129 **Methods**

130 In this article three methods were used. They are introduced here to make our
131 results easily understood.

132 **(1) *the regional depth phase modeling***

133 The focal depths of aftershocks can be jointly estimated by a conventional
134 locating method using the arrival times of the Pg and Sg phases (P and S phases
135 traveling in the upper crust) recorded at close stations. However, the uncertainty in a
136 focal depth solution may be too large to be used to determine the rupture plane for the
137 mainshock. This is mainly due to the trade-off among the focal depth, the epicenter
138 and the origin time.

139 In eastern Canada, regional depth phases sPg , $sPmP$, and sPn are often observed.
 140 Figure 2 provides an example. Ma and Eaton (2011) showed the sketch travel paths of
 141 the depth phase sPg , $sPmP$, and sPn , as well as their reference phases Pg , PmP , and
 142 Pn . The travel path of a depth phase and that of its reference phase have a large
 143 difference at the source region. At a seismic station, the travel time difference is
 144 mainly generated by the focal depth, while the epicentral distance have a small
 145 contribution, so the travel time difference is not sensitive to the epicentral distance,
 146 and can be used to determine focal depth.

147 The procedure that uses a depth phase to determine focal depth is to search a
 148 depth that can be used to generate a time difference between a synthetic depth phase
 149 and its reference phase, and this time difference is close to the time difference
 150 between the observed depth phase and its reference phase.

151 Figure 3 shows an example of the modeling of the depth phase sPg at station A64.
 152 In Figure 3 (a) the top trace (17.2) was generated using a depth of 17.2 km, the crustal
 153 velocity model in Figure 4, and the moment tensor solution we obtained for the 1988
 154 Saguenay mainshock. Other synthetic traces were generated with an increment of 0.1
 155 km. The trace indicated with A64/hhz is the record. The modeled depth is 17.5 km.
 156 The modeled focal depth will change if the velocity model used to generate the
 157 synthetic traces is modified. The synthetic traces in Figure 3 (b) were generated using
 158 a crustal model in which the velocities are 4% slower than those in Figure 4, the
 159 modeled focal depth using the same record is 16.7 km. This depth is 4.57% shallower
 160 than 17.5 km.

161 When the time difference between sPg and Pg is used to determine focal depth,
 162 PmP phase is not relevant. However, when depth phase $sPmP$ is used to determine a
 163 focal depth, PmP should be taken as its reference phase. To understand how the PmP
 164 phase varies with velocity structure and azimuth, we have conducted a series of
 165 experiments and the results are shown in Figures 5 and 6. Here we briefly discuss its
 166 features. As PmP is a reflected phase at the Moho interface, the velocity contrast there
 167 plays a crucial role on the amplitude of PmP recorded at a seismic station. Figure 5
 168 shows that when the P wave velocity ratio is 0.89 (7.1/8.0) the PmP amplitude is

169 much stronger than that when the ratio is 0.95 (7.1/7.5).

170 At the same distance, the amplitude of PmP changes with station azimuth. Figure
171 6 shows 14 traces generated at 14 azimuths with epicentral distance 100 km. Those
172 traces were generated using the focal mechanism in Figure 15. At the nodal plane
173 strike directions (160° and 299°) the amplitudes of PmP are not weak. The weak
174 amplitudes occurred at directions somewhere between the strike and the dip directions.
175 The reason is that the take-off angles of those PmP waves were not close to the two
176 dip angles, they did not travel along the two nodal planes.

177 The amplitude of PmP and the time difference between PmP and Pg also change
178 with the distance at which they are recorded (Figure 7).

179

180 **(2) hypoDD algorithm**

181 The hypoDD algorithm (Waldhauser and Ellsworth, 2000; Waldhauser, 2001) is
182 used to relocate hypocentres within earthquake clusters. In the original hypoDD
183 program, for each event, four corrections corresponding to the longitude, latitude,
184 focal depth, and the origin time need to be solved. These four corrections come from
185 trade-off during the equation solving process, usually the obtained focal depth
186 solution having the largest uncertainty. To reduce the uncertainty in the focal depth,
187 the original hypoDD program was revised (Ma and Eaton, 2011). In the revised
188 version the focal depth values are fixed to the values obtained using the RDPM
189 procedure or to the trusted catalog values.

190 **(3) The method used to retrieve a moment tensor**

191 An earthquake source can be described using a moment tensor. A moment tensor
192 is a 3×3 matrix. In linear algebra a complex matrix can be expressed by summing
193 several simple, independent matrices. An arbitrary moment tensor is expressed by
194 summing six different constant moment tensors (Kikuchi and Kanamori, 1991):

$$195 \quad M = \sum_{n=1}^6 a_n M_n \quad (1)$$

196 where M_1 to M_5 indicate different constant shear dislocation sources, M_6 indicates an

197 explosion source, and a_n ($n = 1, 2, \dots, 6$) are summation coefficients.

198 The relationship between the moment tensor and the coefficients is as follows:

$$199 \quad M = \begin{bmatrix} a_2 - a_5 + a_6 & a_1 & a_4 \\ a_1 & -a_2 + a_6 & a_3 \\ a_4 & a_3 & a_5 + a_6 \end{bmatrix} \quad (2)$$

200 Given an earthquake hypocenter and earth model, at a specific seismic station
201 each of the above constant sources can be used to generate synthetic seismograms,
202 called Green's functions. We assume the three component Green's functions generated
203 by M_n are $U_n(t)$, $V_n(t)$, and $W_n(t)$. The synthetic three component seismograms at the
204 station generated by the moment tensor can be calculated using the following
205 formulas:

$$206 \quad U(t) = \sum_{n=1}^6 a_n U_n(t)$$

$$207 \quad V(t) = \sum_{n=1}^6 a_n V_n(t) \quad (3)$$

$$208 \quad W(t) = \sum_{n=1}^6 a_n W_n(t)$$

209 Once the 6 coefficients are obtained, the moment tensor can be calculated using
210 formula (2).

211 Both an objective function and an inversion method need to be determined to fit
212 waveforms to invert a moment tensor. One objective function is a combination of the
213 correlation between the observed waveforms and the synthetic waveforms and the
214 P -phase polarity signs at the stations surrounding the earthquake epicenter. The
215 objective function used in a computer program (Yao, personal communication, 1995)
216 is as follows:

$$217 \quad S = \sum_{j=1}^{J_0} \left(1 - \frac{\sum_i X_j(t_i) \cdot O_j(t_i)}{\sqrt{\sum_i X_j^2(t_i) \cdot \sum_i O_j^2(t_i)}} \right)^2 + \sum_{k=1}^{K_0} |op(k) \times tp(k)| \times w(k) \quad (4)$$

218 where subscript j ($j = 1, 2, \dots, J_0$; J_0 is the total number of the records) is an ordinal
219 number of the digital records used in the inversion; subscript i ($i = 1, 2, \dots, I_0$; I_0 is

the total number of the data points) is the data point index in the record or the synthetic waveform; $O_j(t_i)$ is a segment of a digital record; $X_j(t_i)$ is a segment of a synthetic seismogram corresponding to $O_j(t_i)$; $op(k)$ is the observed P -phase onset polarity sign (+1 or -1) at station k ($k = 1, 2, \dots, K_0$; K_0 is the total number of the stations at which the observed and calculated polarities are opposite. This number is counted during an iteration) and $tp(k)$ is the theoretical P -phase onset polarity sign (-1 or +1) at the same station. The term $tp(k)$ can be calculated using formula 4.91 in Aki and Richards (1980), and $w(k)$ is a weighting factor assigned to station k .

The moment tensor inversion was performed using a revised simulated annealing method (Liu et al., 1995; Ji, 1994). If one moment tensor within a set of trial moment tensors can bring S in formula (4) to its minimum value, that moment tensor is considered to be the solution. The $w(k)$ values in the inversion program for all stations were assigned to 1/3.

Focal depth distribution in the 1988 Saguenay earthquake epicenter region

Regional depth phase identification

To correctly understand, identify, and use depth phases, we generated a set of synthetic seismograms to show the key features that depth phases have when distance changes. Figure 7 shows those key features. One key feature is that the sPmP phase is weak within about 100 km ($\sim$ one degree), while sPg is weak beyond 1.1 degree. Another feature is that the time difference between a depth phase and its reference phase changes slightly with distance. In Figure 7, the dashed line indicating the Pg phase is almost parallel to the dashed line indicating the sPg phase. These features also exist along the waveform records.

Figure 2 shows the vertical records generated at six stations by the m_N 3.3 event on 26 December 2013 (No. 31 in Table 1). With the six P -wave onsets aligned, the time differences between depth phase sPg and its reference phase Pg are almost equal, while the largest difference in the epicentral distances is 31.6 km (114.1 – 82.5). As

the sPmP phase did not develop well at about 100 km, the phase in Figure 2, indicated with sPg, must be depth phase sPg. It is impossible for the indicated sPg to be PmP as the difference between Pg and PmP along a trace noticeably changes when the epicentral distance becomes longer (Figure 7).

Focal depths for earthquakes with $m_N \geq 2.5$ in the epicenter region

We retrieved the waveform records from the GSC earthquake database for about 40 earthquakes with $m_N \geq 2.5$ that occurred between November 1988 and June 2016 in the region surrounding the Saguenay mainshock (latitudes 47° N to 49° N; longitudes 69.5° W to 73° W). Not all waveforms were archived because the CLTN stations used a trigger algorithm. We analyzed the available waveform records and found 33 events that had depth phase records. Using the depth phase modeling procedure, we retrieved focal depths for the 33 events. The focal depth solutions are listed in Table 1 (column H).

We divided the 33 events into two groups according to the event occurrence time. The foreshock, mainshock, and aftershocks occurred before May 1989, the first 11 events in the table, compose the first group. In this group, the average focal depth in column H is 27.3 km. The remaining 22 events compose the second group. In this group, the focal depths are notably shallower: the average focal depth in column H is 17.4 km.

In the first group, the GSC catalog provided the focal depths (column GSC). After the three field stations were removed in the autumn of 1989, most events in the second group were assigned a default focal depth of 18 km.

The epicenters of the 33 events in Table 1 are plotted on Figure 8 (a). The solid circles (the epicenters) with deep color in the middle (the epicenter area of the mainshock) are the events in the first group. The other solid circles (the epicenters of the events in the second group) are scattered in and around the mainshock.

To observe the possible change of the focal depth with time (vertical seismicity), we plotted Figure 8 (b). The height of a vertical line segment represents the focal depth. The vertical line segment with two arrows indicated with number 1 shows that

the focal depths in group 1 are between 22 km and 33 km. While the vertical line segment with two arrows indicated with number 2 shows that most events in group 2 occurred between 10 km and 25 km; some events are even shallower than 10 km. If we assume that the seismic activity in the epicentral region is related to the mainshock, the change in the hypocentral distribution over time may suggest a stress migration or redistribution.

The errors in routinely computed focal depths may be large

As mentioned the error in focal depth obtained by a conventional method may be large. Here we will prove this point by comparing focal depths of two earthquakes obtained by RDPM and provided by GSC. Based on the definition of a depth phase, we know that similar time differences between the sPg and Pg phases at the same station represent similar focal depths for the aftershocks of the same mainshock because the distances between the station and the aftershocks are similar. Figure 9 shows two vertical records at station A61, generated by the m_N 4.1 aftershock occurred on 26 November 1988 at 03:38 and the m_N 2.9 aftershock occurred on the same day at 04:54 (Table 1, No. 3 and 6, respectively). The time differences between the sPg and Pg phases were almost equal, implying that these two events had similar focal depths. The time differences between the Sg and Pg phases were also almost equal, implying that these two events also had very similar epicentral distances.

Figure 10 shows the focal depth phase modeling for the m_N 4.1 aftershock. The modeled focal depth was about 22.0 km. The m_N 2.9 aftershock was assigned the same focal depth as the two time differences between the sPg and Pg phases are almost equal.

According to the GSC database, the focal depths for the events on the 26 November 1988 at 03:38 (m_N 4.1) and at 04:54 (m_N 2.9) were 26.4 km and 22.3 km (events 3 and 6 in Table 1), respectively. Our focal depth solution for the first event is not close to 26.4 km, but it is close to the reported focal depth of 22.3 km for the second event. This comparison shows that the error in the focal depth solution provided by GSC for the event No. 3 reaches more than 4 km.

309

310 ***The errors in focal depth obtained by depth phase modeling***

311 Any factor that is involved in synthetic seismogram generation can cause errors in
312 the focal depth solutions obtained by the depth phase modeling procedure. These
313 factors include the crustal velocity model, the Moho depth and other strong interfaces
314 within the crust, epicentral distance, and focal mechanism. Ma and Atkinson (2006)
315 and Ma (2010) discussed these factors. The key factor related to the errors in the
316 modeled depth is the crustal velocity model.

317 The Western Quebec Seismic Zone (WQSZ) is active in eastern North America,
318 and the crustal structures are relatively well known. Mereu et al. (1986) conducted a
319 major long-range seismic refraction and wide-angle reflection experiment across the
320 Grenville province of the Canadian Shield and obtained some crustal models. After
321 those crustal models were studied and modified manually, one crustal model was
322 obtained (Figure 4). When Motazedian et al. (2013) studied the crustal velocity models
323 using Rayleigh wave dispersion data in northeastern North America, and found the
324 velocity model in Figure 4 was faster in the middle and upper crust than those models
325 obtained using the Rayleigh wave dispersion data on the northern side of the St
326 Lawrence River.

327 As the Moho interface in the crustal models obtained by the Rayleigh wave
328 dispersion data are not well confined, and a good Moho interface is necessary to
329 generate a clear sPmP phase for depth phase modeling; consequently, the crustal model
330 in Figure 4 is set to be the default model in the computer program.

331 Because both the Saguenay earthquake epicenter region and the WQSZ are on the
332 northern side of the St Lawrence River, the crustal structures in the two regions have
333 similarities. When the crustal model in Figure 4 is used in the Saguenay region, the
334 error in the obtained focal depth is not large.

335 Based on the results of Motazedian, et al. (2013), the average S-wave velocity
336 value obtained by the Rayleigh wave dispersion data within 30 km of the upper and
337 middle crust on the northern side of the St. Lawrence River is 3.61 km/s; within this 30
338 km, the average S-wave velocity value in the default model is 3.76 km/s (Figure 4). The

339 difference between the two average S-wave velocities is 0.15 km/s or ~4% of the
 340 assumed V_s value. It implies an average of 4.57 reduction in the modeled focal depth.

341 When performing the depth phase modeling, the error in the arrival time
 342 measurements can easily cause a depth error of a couple of kilometers. To reduce the
 343 measurement error, we use the peak times to align phases.

344

345 **HypoDD relocation of 10 earthquakes of $m_N \geq 2.5$ that occurred between 23**
 346 **November 1988 and 18 April 1989**

347 If we assume that most of the immediate aftershocks occurred on or near the
 348 rupture plane of the mainshock, then we can use the spatial distribution of the
 349 aftershock hypocenters to determine the orientation of the rupture plane, or help us
 350 identify one nodal plane that is close the rupture plane. To use the hypocentral
 351 distribution to determine the fault plane, the errors in the hypocenters need to be small
 352 and consistent.

353 Many factors influence the accuracy of earthquake hypocenters that are
 354 determined using the arrival times of P - and S -waves. One key consideration is the
 355 crustal structures through which the P - and S -waves propagate. The hypoDD
 356 algorithm (Waldhauser and Ellsworth, 2000; Waldhauser, 2001) can cancel the path
 357 effects to some extent; thereby increase the accuracy of the relative hypocentral
 358 locations.

359 We chose a total of 13 pick files in which the arrival times of P_g and S_g phases
 360 are recorded for the foreshock, mainshock, and aftershocks with $m_N \geq 2.5$, occurred
 361 before May 1989 in the epicentral region ($48.0^\circ\text{N} \sim 48.15^\circ\text{N}$; $71.0^\circ\text{W} \sim 71.25^\circ\text{W}$).
 362 Then we analyzed the files and found that three files did not have a sufficient number
 363 of P_g and S_g arrival time records for the hypoDD inversion; consequently, we
 364 removed those three files. The remaining 10 events with a sufficient number of P_g and
 365 S_g arrival time records are listed in Table 2.

366 Using the 10 pick files, we measured the P_g and S_g arrival times at the
 367 Charlevoix station array. Most measurements only have a very subtle difference
 368 compared to those in the pick files; only a notable difference was found at the S_g

369 arrival time at station A61 in the pick file for the foreshock. Using our measurements,
370 we revised the 10 pick files. The three local stations, CQ1, CQ2, and CQ3, are analog
371 stations and for this reason only the arrival times could not be checked. The onsets of
372 *Pg* and *Sg* should have been clear and easy to measure correctly at these three local
373 stations, and so we trusted the measurements in the pick files.

374 We retrieved *Pg* and *Sg* arrival times from the pick files and relocated the 10
375 events in Table 2 using the revised hypoDD method (Ma and Eaton, 2011). The
376 relocation procedure was performed using the crustal model shown in Figure 4. All 10
377 events were successfully relocated using the gradient algorithm Sparse Linear
378 Equations and Least Square (LSQR) (Paige and Saunders, 1982); the results are listed
379 in Table 3.

380 The solutions were related to the damping factor value in the LSQR program input
381 data. Ma and Eaton (2011) performed a test using different factor values on the same
382 data set for an earthquake group. When the factor value in the program input file was
383 10, the relocated epicenters formed a small cluster; however, when the factor value
384 was 90, the relocated epicenters were close to the initial values. For the 10 events, the
385 solutions were similar when the factor values were 60, 70, and 80. When the factor
386 value was 50 or smaller, only some of the 10 events could be relocated. The solutions
387 in Table 3 were obtained using a factor value of 70.

388 Panel (a) in Figure 11 shows the distribution of the stations that were used for the
389 hypoDD inversion, panel (b) shows the relocated epicenter distribution, panel (c)
390 shows the hypocenter projection on vertical profile A-A', and panel (d) shows the
391 hypocenter projection on profile B-B'. The strike of the A-A' profile is at azimuth 68°
392 and that of the B-B' profile is at azimuth 158°.

393 In Figure 11 (b) the epicenters form a linear trend. When the epicenters for the
394 same 10 events in the GSC catalogue were plotted, the linear trend does not exist.
395 Figure 12 shows the comparison between the relocated epicenters and those in the
396 GSC catalogue.

397 Figure 13 demonstrates how a possible rupture plane for the mainshock was
398 retrieved. This figure is a 3D distribution of the 10 hypocenters, viewed at a

399 north-westerly direction. The 26 November 1988 at 08:13 m_N 2.5 aftershock and the
 400 19 January 1989 m_N 3.6 aftershock (No. 7 and 9) did not match the linear trend. These
 401 two aftershocks possibly occurred along another fault, and so they were removed,
 402 leaving the other eight events to simulate a rupture plane. Figure 14 shows two
 403 projections of the fitted spatial plane. The strike of this plane is at azimuth 158° (or
 404 338°) and the dip angle is 72° .

405 The sizes of the solid circles in Figures 13 and 14 are proportional to the
 406 earthquake magnitudes. The deep blue colors of the circles show the early events; red
 407 colors show the later events.

408 The columns indicated with e_EW and e_NS in Table 3 show the errors for the
 409 latitude and longitude of the relocated earthquakes, respectively. They are the output
 410 from the hypoDD program. Those values can help us determine a proper factor value.
 411 The error values are also able to help us judge the used data quality for a specific
 412 earthquake. For example, if the error values for an earthquake are much larger than
 413 those for other earthquakes, the measured P and S arrival times may not be correct.

415 **Moment tensor solution for the 1988 M_W 5.9 Saguenay mainshock**

416 In 1988, the seismic station coverage was not ideal. Authors obtained their moment
 417 tensor or focal mechanism solutions based on a limited data set. The first 4 sets of the
 418 solutions in Table 4 were those believed to be the best based on the available data set.
 419 However, a reasonable moment tensor solution can also be obtained if the technique
 420 introduced in the method section is used for the Saguenay mainshock because the
 421 technique can combine waveform fitting and P -wave polarity information.

422 Waveform records for the Saguenay mainshock were retrieved from IRIS and
 423 analyzed. It was found that the Rayleigh waveform records at station HRV were
 424 excellent for the moment tensor inversion. Short period waveform records were
 425 retrieved from the GSC waveform archive database, and the P -wave polarities were
 426 measured. The waveform records at HRV and the polarities measured at Canadian
 427 stations were used as input for the moment tensor inversion method introduced above,
 428 and several moment tensor solutions were obtained. Figure 15(a) shows one of the

429 moment tensor solutions.

430 Two nodal planes can be retrieved from a moment tensor solution. The spatial
431 distribution of the immediate aftershocks is often used to select the rupture plane from
432 the two nodal planes. Figure 14 shows that those eight events were distributed on a
433 plane with a strike in the northwesterly direction and a dip in the northeasterly direction.
434 Comparing Figure 14 (b) to nodal planes in Figure 15 (a), it was inferred that nodal
435 plane *p*₂ was close to the rupture plane of the mainshock. Figure 15 shows a revised
436 version from the report draft by Ma and Adams (2002).

437 Other authors performed moment tensor and focal mechanism inversions for the
438 Saguenay mainshock using different datasets and methods. Table 4 lists the
439 parameters of the nodal plane 2 obtained by different authors. All the solutions show
440 that the Saguenay earthquake is a thrust event. Compared to other solutions, our strike
441 azimuth was rotated counterclockwise by about 20°, while the dip and the slip angles
442 are similar.

443 As the solutions in Table 4 came from different datasets, and those datasets were
444 very limited as the seismic station coverage in 1988 was not good, we do not have
445 criteria to judge which solution is the best. Ma and Adams (2002) made waveform
446 comparisons between the observed and synthetic Rayleigh waves at station HRV. The
447 synthetic seismograms were generated using focal mechanism solutions obtained by
448 different authors. If the shapes and amplitudes of the observed and the synthetic
449 Rayleigh waves are similar, the moment tensor solution used to generate the synthetic
450 waves is correct, but the solution contains an error. When the maximum amplitude
451 ratios of the observed and synthetic waveforms at three components equal 1 and the
452 cross-correlations also equal 1, the moment tensor solution is perfect. However, this
453 ideal case cannot be achieved since the solution has errors.

454 The column indicated with σ in Table 4 shows the ratio variance relative to 1. To
455 some extent this variance value shows the error inherent in the solution. The smallest
456 variance value was obtained from the solution by North et al. (1989). As the variance
457 values are not available at other seismic stations surrounding the epicenter, we still
458 cannot judge which solution is the best.

459 A prominent fault marks the southern boundary of the Graben and is assumed to be
 460 dipping to the north. Since the mainshock locates to the south of this fault, it cannot have
 461 reactivated it (Figure 1). The south wall is represented by the Lake Kenogami lineament,
 462 which trends at 105° (Du Berger et al. 1991; the trend along the southern edge near the
 463 1988 earthquake epicenter in Figure 1 is also at about 105°). This trend agrees with the
 464 opposite direction of the strike we obtained ($299^\circ - 180^\circ$). Due to the consistency
 465 among the strike of nodal plane 2, the trend of the south wall of the Graben, and the
 466 deep focal depth distribution, we deduce that the Saguenay earthquake sequence
 467 possibly shows the reactivation of a deep hidden fault south of the Graben.

468 **Summary and discussion**

469 The 25 November 1988 M_W 5.9 Saguenay earthquake was an important event in
 470 eastern North American. Although the earthquake sequence is the subject of many
 471 studies, some issues remain mysterious, including the large amount of high-frequency
 472 energy, the unusually deep focal depth, and the small number of aftershocks mainly
 473 with low magnitudes (e.g. North et al., 1989). We did not try to solve the mysteries;
 474 instead, we used several modern techniques to further study the source parameters.

475 We first used the RDPM to retrieve focal depths for the foreshock, aftershocks,
 476 and the earthquakes with magnitudes $m_N \geq 2.5$ surrounding the mainshock. For 11
 477 events, we compared the focal depth values obtained using RDPM to those
 478 determined by GSC using Pg and Sg arrival times recorded at nearby stations (Table
 479 1); we found that for some earthquakes, the two focal depth values are similar while
 480 for others, they are different, sometimes significantly such as event No. 3 in Table 1.

481 The major part of this article is to identify the rupture plane using the distribution
 482 of the hypocenters. Previously, North, et al. (1989) tried to determine the rupture
 483 plane of the mainshock with the distribution of 55 aftershocks obtained using the joint
 484 hypocenter determination method (Douglas, 1967). They projected the 55 hypocenters
 485 onto two profiles striking at azimuths 55° and 118° , respectively, and found that there
 486 was a distinct linear trend of events that included the main shock, and this trend was
 487 aligned roughly parallel to the 67° dip of the nodal plane. They believed that this zone

488 may be the best indication of the mainshock rupture plane in the aftershock data
489 although the evidence was not conclusive. We looked at the right panel in their Figure
490 3 and found there was a hypocenter trend running through the mainshock, but the
491 group that included the largest aftershock (No. 3 in Table 1) was off trend. The largest
492 aftershock, main shock, and other smaller aftershocks seemed to form a triangle in
493 their Figure 3. The focal depth of the largest aftershock was 24.8 km in their Table 1.
494 If this depth were moved up 3 km to our solution (22 km), their linear trend would be
495 conclusive. The linear trend in Figure 13 and Figure 14 (b) was clear and conclusive.
496 The rupture plane of the mainshock was close to the fitted spatial plane or close to
497 nodal plane 2 in Figure 15 (a). The calculated strike of the fitted plane in Figure 14 (a)
498 is at azimuth 158° (or 338°) and the dip angle is 72° (Figure 14 (b). These two
499 parameters better match those of North, et al. (1989) than those by other authors.

500 As previously mentioned, in 1988 the broadband seismic station coverage was not
501 ideal. Clear long period surface wave regional records were available only at HRV.
502 There were regional short period seismic stations surrounding Saguenay, but those
503 short period waveform records could not be modeled. However, P-wave polarities
504 could be measured at those short period stations. To obtain a reliable focal mechanism
505 solution from P-wave polarities alone is not easy. One reason is that, at the stations
506 along the nodal plane region, positive or negative P-wave polarities are often hard to
507 determine.

508 We used 3-component Rayleigh waves at HRV and 17 polarities to retrieve the
509 moment tensor for the Saguenay mainshock. Figure 15 (a) shows that most positive
510 polarities fall in the dilatational quadrants, with only a few mismatches. Figure 15 (b)
511 shows that the waveform similarity between the observed and the synthetic is good
512 and the maximum amplitude ratios (1.16, 1.02, and 0.82) of the observed and the
513 synthetic waveforms at tangential, radial, and vertical component are all reasonable.
514 These criteria show that our moment tensor solution for the mainshock is acceptable,
515 and the error is not large.

516

517 Data and Resources

518 Seismograms were obtained from the Geological Survey of Canada (GSC) at
519 <http://earthquakescanada.nrcan.gc.ca/index-eng.php> (last accessed 30 August 2016),
520 and from the Incorporated Research Institutions for Seismology (IRIS) at
521 <http://www.iris.edu/hq/> (last accessed 30 August 2016). The earthquake catalog and
522 the pick files were also retrieved from the GSC. The extent of the Saguenay Graben
523 was taken from the presentation by Thériault, et al. (2005) at
524 [http://sigpeg.mrn.gouv.qc.ca/gpg/images/collectionInterne/presentation_orale/14-Pres-](http://sigpeg.mrn.gouv.qc.ca/gpg/images/collectionInterne/presentation_orale/14-PresentationACG-AMC2005.pdf)
525 [entationACG-AMC2005.pdf](http://sigpeg.mrn.gouv.qc.ca/gpg/images/collectionInterne/presentation_orale/14-PresentationACG-AMC2005.pdf).

526

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535 retrieve the waveform records and pick files from the GSC database. The waveform
536 records were processed using SAC2000, *redseed*, and *geotool* programs. Some figures
537 were prepared using MATLAB and Generic Mapping Tools (Wessel and Smith,
538 1998).

539

540

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Figure 1. Map of the region surrounding the Charlevoix Seismic Zone and the epicenter of the 1988 M_w 5.9 Saguenay earthquake (1988). Colours indicate the elevation of the ground. The earthquake hypocentres (circles) are events with $m_N \geq 2.8$ that occurred between 1985 and 2017 inclusively, with symbol sizes related to the magnitude and the colour to the focal depth. The dashed lines show the major faults that define the Saguenay Graben (Du Berger et al., 1991). The eastern extent of the southern boundary is approximate. CNSN stations are marked by squares; while the locations of the Quebec City and Saguenay community are marked by diamonds.

Figure 2. Vertical component records at Charlevoix seismic stations A54, A61, A64, A16, A11, and A21, generated by aftershock 31 in Table 1. The time at the left side of each trace is the record start time. There is only a small difference between the lengths of the line segments indicated with δt . Each record's epicentral distance is indicated along the respective trace. The difference between the closest epicentral distance and the furthest is ~ 32 km, but there is only a subtle difference in the time durations between the sPg and Pg phases when the top and bottom traces are compared. This is a unique feature of depth phase and can be used for depth phase identification.

Figure 3. (a) Regional depth phase sPg modeling at station A64 (epicentral distance ~ 96 km). Traces 1, 2, 3, 4, 5, 6, and 7 were generated with depths 17.2, 17.3, 17.4, 17.5, 17.6, 17.7, and 17.8 km, respectively. The trace indicated with A64/hhz is the P -wave segment recorded at A64, generated by event 31 in Table 1. The used crust model is in Figure 4. The time difference between sPg and Pg along trace 4 best matches that recorded. The modeled focal depth for this event is about 17.5 km. **(b)** Synthetic traces generated using a crustal model 4% slower than the model in Figure 4 at depths 16.5, 16.6, 16.7, 16.8, 16.9, and 17.0 km, and at the same distance. The time difference between sPg and Pg along trace 3 best matches that recorded. The modeled focal depth for the same event is 16.7 km.

Figure 4. Crustal velocity models used to generate synthetic seismograms.

Figure 5. Synthetic seismograms generated at depths 22, 23, 24, 25, 26, and 27 km, and distance 100 km, using the same crustal model (Figure 4), but a different P wave velocity beneath the Moho. The S wave velocity and the density beneath the Moho were also adjusted, accordingly. The P wave velocity in the layer above the Moho is 7.1 km/s; the velocities beneath the Moho are 7.5 km/s, 7.7 km/s, and 8.0 km/s, respectively. The PmP amplitudes along the top six traces are weak; while the amplitudes along the bottom six traces are strong.

668
669 **Figure 6.** Synthetic seismic traces generated at 14 azimuths; using depth 22 km,
670 distance 100 km; and crustal model in Figure 4, and focal mechanism in Figure 14, in
671 which the strike of the nodal plane 1 is 160°, the dip direction is 250°; the strike of the
672 nodal plane 2 is 299°, the dip direction is 29°. The amplitudes of the PmP phase
673 change with azimuth. Weak amplitudes occurred at directions somewhere between the
674 strike and the dip directions.

675
676 **Figure 7.** Synthetic regional phases. The top trace 0.91 was generated using a distance
677 0.91°, depth 22.1 km, and the crustal velocity model in Figure 4. Other traces were
678 generated with the same depth and velocity model, and a distance increment of 0.1°.
679 At the distance 0.91°, sPmP is not well developed, while sPg is. The sPmP phase is
680 clear after 2.0°. The time difference between PmP and Pg changes with distance.
681 Between 1.5 ~ 2.0°, Pg and PmP, sPg and sPmP are hard to separate. The time
682 difference between sPg and Pg slightly changes with distance. The time difference
683 between sPmP and PmP only very slightly changes with distance. The Pn phase
684 appears at 1.31° and is weak. The time differences between Pn and Pg, as well as Pn
685 and PmP increase with distance.

686
687 **Figure 8.** (a) Epicenter distribution in the 1988 Saguenay earthquake area and its
688 vicinity. The focal depth of each earthquake is indicated by the color of the solid circle
689 that represents the earthquake. The color palette at the right shows the depth value
690 associated with each color. The number near a circle is the year in which the
691 earthquake occurred. (b) The focal depth distribution with time. The length of each
692 vertical solid line segment shows the focal depth for an event of group 1 in Table 1,
693 while the length of each dashed vertical line segment shows the focal depth in group 2.
694 The position of a vertical line segment along the horizontal axis shows the time of the
695 event occurrence. The date and the time of each event of Table 1 were converted to
696 days counted from the origin of the foreshock.

697
698 **Figure 9.** Full waveform records generated by aftershocks 3 and 6 in Table 1. The
699 time differences between phase pairs sPg and Pg, Sg and Pg are almost identical along
700 these two traces. The time at the left side of each trace is the start time of the original
701 record.

702
703 **Figure 10.** Regional depth phase sPg modeling at station A61 (distance ~ 83 km).
704 Each synthetic trace was generated using the depth indicated at the left. The bottom
705 trace is the P-wave segment recorded at A61, generated by event 3 in Table 1. The
706 modeled focal depth for this event is about 22.0 km.

707
708 **Figure 11.** Projections of the 10 relocated events obtained using the crustal model in
709 Figure 4. (a) The distribution of the stations used; (b) the distribution of the 10
710 epicenters; (c) the projection of the 10 hypocenters on vertical profile A-A' at azimuth
711 68°, shown in (b); and (d) the projection of the 10 hypocenters on vertical profile B-B'

at azimuth 158° . The arrows point to the aftershocks 7 and 9 in Table 3.

Figure 12. Epicentre distributions. The solid circles show those relocated 10 events, while the diamonds show the same 10 events in the GSC catalogue.

Figure 13. Spatial distribution of the 10 relocated events. The frame was manually adjusted to show the linear trend of the hypocenters, looked at a north-west direction. The colors of the solid circles show the time progression of the events. The numbers 1 to 10 are the event index in Table 2 and Table 3. The size of each solid circle is proportional to the magnitude of the event depicted by the circle. The m_N 2.5 aftershock on 26 November 1988 and the m_N 3.6 aftershock on 19 January 1989 (No. 7 and 9, respectively) did not fit the trend. The symbol Lat means latitude; while Lon means longitude.

Figure 14. Projections of the fitted rupture plane for the mainshock using the hypocenters of the eight relocated earthquakes in Table 3 (No. 7 and 9 were removed). The strike of the fitted plane is at azimuth 158° (or 338°), the dip angle is 72° , and the dipping direction is at azimuth 68° . (a) The relocated hypocenters on a vertical plane along the simulated strike direction (azimuth 338°); (b) the relocated hypocenters on a vertical plane along the dipping direction (azimuth 68°). The colors of the solid circles show the time progression of the events.

Figure 15. (a) The lower hemispherical projection of the moment tensor. The plus sign + next to a station name shows that the polarity at that station is positive. The plus sign with letter *P* shows the location of compressional axis. *p1* and *p2* show the two nodal planes. Plane *p2* is identified as the approximate fault on which the 1988 Saguenay mainshock occurred. (b) Waveform comparison between the observed long-period Rayleigh wave (10 s ~ 40 s) records at station HRV and the synthetic seismograms generated using the moment tensor, shown at panel (a). In each waveform pair, the upper trace is the observed waveform, while the lower trace is the synthetic waveform. The characters and numbers on the left side of each pair are, from top to bottom: the station code, component (TAN, RAD, and VER), the epicentral distance and the azimuth in degrees, and the maximum amplitude ratio between the observed and the synthetic.

747 **Table 1.** The GSC catalog of earthquakes with magnitude $m_N \geq 2.5$

No.	date	time	lat./°	long./°	H	GSC	δH	m_N
1	1988/11/23	09:11:27	48.132	-71.200	26.5	29.0	-2.5	4.6
2	1988/11/25	23:46:04	48.117	-71.183	28.5	28.9	-0.4	5.9 M_W
3	1988/11/26	03:38:08	48.142	-71.299	22.0	26.4	-4.4	4.1
4	1988/11/26	04:06:58	47.947	-71.053	29.0	23.3	5.7	2.5
5	1988/11/26	04:47:26	47.984	-71.154	29.5	27.8	1.7	2.7
6	1988/11/26	04:54:20	48.142	-71.264	22.0	22.3	-0.3	2.9
7	1988/11/26	06:36:31	48.100	-71.091	33.0	32.2	0.8	2.6
8	1988/11/26	08:13:21	48.160	-71.021	26.5	23.6	2.9	2.5
9	1988/12/11	14:30:14	48.001	-71.164	29.0	26.8	2.2	2.8
10	1989/01/19	21:36:38	48.063	-71.008	25.0	24.8	0.2	3.6
11	1989/04/18	20:49:08	48.115	-71.160	29.0	28.8	0.2	2.5

12	1989/10/31	04:50:03	48.666	-72.328	17.0	18.0		3.4
13	1990/07/06	03:42:16	48.433	-72.136	10.0	18.0		2.8
14	1990/12/31	03:53:58	47.617	-72.587	24.0	18.0		4.5
15	1991/10/12	23:03:45	48.302	-71.308	11.5	15.0	-3.5	3.3
16	1995/09/23	03:06:41	48.303	-71.489	12.0	18.0		2.7
17	1998/01/10	01:51:04	48.194	-70.908	6.0	18.0		3.6
18	1999/02/23	01:10:36	48.018	-70.674	21.5	18.0		2.7
19	2000/07/12	15:01:49	47.562	-71.064	24.5	18.0		4.2
20	2000/07/12	15:06:35	47.551	-71.078	24.5	18.0		3.5
21	2004/05/04	15:41:15	47.938	-70.667	23.5	22.0	1.5	3.5
22	2004/05/08	05:46:45	47.935	-70.666	23.0	22.0	1.0	3.3
23	2006/03/25	18:23:59	48.304	-71.483	11.0	18.0		2.9
24	2008/04/15	09:51:34	48.585	-72.263	26.0	18.0		2.8
25	2008/11/02	07:26:27	47.431	-71.440	16.5	18.0		3.6
26	2010/04/24	07:08:54	47.055	-71.085	24.5	22.5	2.0	3.3
27	2011/06/13	17:47:14	48.194	-71.175	24.5	18.0		2.6
28	2011/12/20	04:12:12	47.522	-72.215	10.0	18.0		3.0
29	2012/04/23	12:56:01	48.025	-71.553	14.0	8.5	5.5	3.0
30	2012/09/17	04:57:16	48.215	-71.282	8.5	18.0		2.6
31	2013/12/26	04:37:51	48.011	-71.144	18.0	17.5		3.3
32	2013/12/26	04:52:36	48.014	-71.145	18.0	18.0		2.7
33	2016/06/29	19:03:11	48.130	-71.235	13.5	18.0		2.6

748 **Note:** These earthquakes occurred between latitudes 47.0°N and 49.0°N and longitudes
749 69.5°W and 73.0°W for the time period from 23 November 1988 to 29 June 2016.
750 Column H (km) lists the focal depths determined with RDPM. Column GSC (km) lists
751 values from the GSC catalog (the focal depth of 18.0 km is the default focal depth
752 value). Column δH (km) lists the depth difference between columns H and GSC. The
753 events indicated with bold text were mentioned in this article.

754

755 **Table 2.** Catalog of the events selected for relocation.

No.	date	time	lat./°	long./°	H	m_N
1	19881123	09:11:27.3	48.132	-71.200	26.50	4.6
2	19881125	23:46:04.5	48.117	-71.183	28.91	5.9 M_W
3	19881126	03:38:08.2	48.142	-71.299	22.00	4.1
4	19881126	04:47:26.2	47.984	-71.154	29.50	2.7
5	19881126	04:54:20.8	48.142	-71.264	22.00	2.9
6	19881126	06:36:31.1	48.100	-71.091	33.00	2.6
7	19881126	08:13:21.0	48.160	-71.021	26.50	2.5
8	19881211	14:30:14.3	48.001	-71.164	29.00	2.8
9	19890119	21:36:38.4	48.063	-71.008	25.00	3.6
10	19890418	20:49:08.4	48.115	-71.160	29.00	2.5

756 **Note:** The depth value, 28.91 km, for the main shock comes from the GSC catalog.

757 Other depth values come from the RDPM procedure (see Table 1).

758

759 **Table 3.** Catalog of the relocated events.

No.	date	time	lat./°	long./°	H	e_EW	e_NS	e_Z	m_N
1	1988-1123	09:11:27.58	48.1234	-71.1805	26.50	13.4	15.8	0.0	4.6
2	1988-1125	23:46:04.71	48.1121	-71.1785	28.91	10.9	14.5	0.0	5.9 M_W
3	1988-1126	03:38:08.74	48.1485	-71.2266	22.00	14.5	16.2	0.0	4.1
4	1988-1126	04:47:25.97	48.0402	-71.1185	29.50	16.6	23.2	0.0	2.7
5	1988-1126	04:54:20.97	48.1368	-71.2130	22.00	14.9	16.6	0.0	2.9
6	1988-1126	06:36:31.26	48.0834	-71.1336	33.00	12.0	17.1	0.0	2.6
7	1988-1126	08:13:20.70	48.0950	-71.1174	26.50	19.6	21.1	0.0	2.5
8	1988-1211	14:30:13.94	48.0619	-71.1353	29.00	21.8	27.4	0.0	2.8
9	1989-0119	21:36:37.78	48.0523	-71.0806	25.00	30.2	35.2	0.0	3.6
10	1989-0418	20:49:08.47	48.1036	-71.1614	29.00	13.6	15.5	0.0	2.5

760 **Note:** Events 7 and 9 did not display the same spatial trend formed by the other eight
 761 events, and therefore, they were removed from the simulation of the rupture plane for
 762 the mainshock. Columns e_EW, e_NS, and e_Z are output errors in meter at E-W, N-S,
 763 and vertical directions, respectively. Focal depths are fixed during the inversion, so
 764 the values in column e_Z are zero.

765

766 **Table 4.** Comparison of the parameters of nodal plane 2 for the 1988 Saguenay
767 mainshock
768

Strike (°)	Dip (°)	Slip (°)	Max. amplitude ratio				σ	Authors
			TAN	RAD	VER			
317	64	60	1.25	0.74	0.59	0.3153		Harvard's CMT catalog
328	51	70	0.81	0.95	0.81	0.1578		Haddon (1995)
320	65	78	0.86	1.17	0.95	0.1304		Somerville et al. (1990)
325	67	54	1.13	1.03	0.82	0.1294		North et al. (1989)
299	57	64	1.16	1.02	0.82	0.1395		This study

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Figure 1. Map of the region surrounding the Charlevoix Seismic Zone and the epicenter of the 1988 M_W 5.9 Saguenay earthquake (1988). Colours indicate the elevation of the ground. The earthquake hypocentres (circles) are events with $m_N \geq 2.8$ that occurred between 1985 and 2017 inclusively, with symbol sizes related to the magnitude and the colour to the focal depth. The dashed lines show the major faults that define the Saguenay Graben (Du Berger et al., 1991). The eastern extent of the southern boundary is approximate. CNSN stations are marked by squares; while the locations of the Quebec City and Saguenay community are marked by diamonds.

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784 **Figure 2.** Vertical component records at Charlevoix seismic stations A54, A61, A64,
785 A16, A11, and A21, generated by aftershock 31 in Table 1. The time at the left side of
786 each trace is the record start time. There is only a small difference between the lengths
787 of the line segments indicated with δt . Each record's epicentral distance is indicated
788 along the respective trace. The difference between the closest epicentral distance and
789 the furthest is ~ 32 km, but there is only a subtle difference in the time durations
790 between the sPg and Pg phases when the top and bottom traces are compared. This is
791 a unique feature of depth phase and can be used for depth phase identification.

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795 **Figure 3. (a)** Regional depth phase *sPg* modeling at station A64 (epicentral distance ~
 796 96 km). Traces 1, 2, 3, 4, 5, 6, and 7 were generated with depths 17.2, 17.3, 17.4, 17.5,
 797 17.6, 17.7, and 17.8 km, respectively. The trace indicated with A64/hhz is the *P*-wave
 798 segment recorded at A64, generated by event 31 in Table 1. The used crust model is in
 799 Figure 4. The time difference between *sPg* and *Pg* along trace 4 best matches that
 800 recorded. The modeled focal depth for this event is about 17.5 km. (b) Synthetic traces
 801 generated using a crustal model 4% slower than the model in Figure 4 at depths 16.5,
 802 16.6, 16.7, 16.8, 16.9, and 17.0 km, and at the same distance. The time difference
 803 between *sPg* and *Pg* along trace 3 best matches that recorded. The modeled focal depth
 804 for the same event is 16.7 km.

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Figure 4. Crustal velocity models used to generate synthetic seismograms.

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 813 **Figure 5.** Synthetic seismograms generated at depths 22, 23, 24, 25, 26, and 27 km,
 814 and distance 100 km, using the same crustal model (Figure 4), but a different P wave
 815 velocity beneath the Moho. The S wave velocity and the density beneath the Moho
 816 were also adjusted, accordingly. The P wave velocity in the layer above the Moho is
 817 7.1 km/s; the velocities beneath the Moho are 7.5 km/s, 7.7 km/s, and 8.0 km/s,
 818 respectively. The PmP amplitudes along the top six traces are weak; while the
 819 amplitudes along the bottom six traces are strong.
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823 **Figure 6.** Synthetic seismic traces generated at 14 azimuths; using depth 22 km,
824 distance 100 km; and crustal model in Figure 4, and focal mechanism in Figure 14, in
825 which the strike of the nodal plane 1 is 160°, the dip direction is 250°; the strike of the
826 nodal plane 2 is 299°, the dip direction is 29°. The amplitudes of the PmP phase
827 change with azimuth. Weak amplitudes occurred at directions somewhere between the
828 strike and the dip directions.

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833 **Figure 7.** Synthetic regional phases. The top trace 0.91° was generated using a distance
 834 0.91°, depth 22.1 km, and the crustal velocity model in Figure 4. Other traces were
 835 generated with the same depth and velocity model, and a distance increment of 0.1°.
 836 At the distance 0.91°, sPmP is not well developed, while sPg is. The sPmP phase is clear
 837 after 2.0°. The time difference between PmP and Pg changes with distance. Between
 838 1.5 ~ 2.0°, Pg and PmP, sPg and sPmP are hard to separate. The time difference between
 839 sPg and Pg slightly changes with distance. The time difference between sPmP and PmP
 840 only very slightly changes with distance. The Pn phase appears at 1.31° and is weak.
 841 The time differences between Pn and Pg, as well as Pn and PmP increase with distance.

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Figure 8. (a) Epicenter distribution in the 1988 Saguenay earthquake area and its vicinity. The focal depth of each earthquake is indicated by the color of the solid circle that represents the earthquake. The color palette at the right shows the depth value associated with each color. The number near a circle is the year in which the earthquake occurred. (b) The focal depth distribution with time. The length of each vertical solid line segment shows the focal depth for an event of group 1 in Table 1, while the length of each dashed vertical line segment shows the focal depth in group 2. The position of a vertical line segment along the horizontal axis shows the time of the event occurrence. The date and the time of each event of Table 1 were converted to days counted from the origin of the foreshock.

Figure 9. Full waveform records generated by aftershocks 3 and 6 in Table 1. The time differences between phase pairs sPg and Pg, Sg and Pg are almost identical along these two traces. The time at the left side of each trace is the start time of the original record.

Figure 10. Regional depth phase sPg modeling at station A61 (distance ~ 83 km). Each synthetic trace was generated using the depth indicated at the left. The bottom trace is the *P*-wave segment recorded at A61, generated by event 3 in Table 1. The modeled focal depth for this event is about 22.0 km.

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872 **Figure 11.** Projections of the 10 relocated events obtained using the crustal model in
873 Figure 4. (a) The distribution of the stations used; (b) the distribution of the 10
874 epicenters; (c) the projection of the 10 hypocenters on vertical profile A-A' at azimuth
875 68°, shown in (b); and (d) the projection of the 10 hypocenters on vertical profile B-B'
876 at azimuth 158°. The arrows point to the aftershocks 7 and 9 in Table 3.

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881 **Figure 12.** Epicentre distributions. The solid circles show those relocated 10 events,
 882 while the diamonds show the same 10 events in the GSC catalogue.

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Figure 13. Spatial distribution of the 10 relocated events. The frame was manually adjusted to show the linear trend of the hypocenters, looked at a north-west direction. The colors of the solid circles show the time progression of the events. The numbers 1 to 10 are the event index in Table 2 and Table 3. The size of each solid circle is proportional to the magnitude of the event depicted by the circle. The m_N 2.5 aftershock on 26 November 1988 and the m_N 3.6 aftershock on 19 January 1989 (No. 7 and 9, respectively) did not fit the trend. The symbol Lat means latitude; while Lon means longitude.

Figure 14. Projections of the fitted rupture plane for the mainshock using the hypocenters of the eight relocated earthquakes in Table 3 (No. 7 and 9 were removed). The strike of the fitted plane is at azimuth 158° (or 338°), the dip angle is 72°, and the dipping direction is at azimuth 68°. (a) The relocated hypocenters on a vertical plane along the simulated strike direction (azimuth 338°); (b) the relocated hypocenters on a vertical plane along the dipping direction (azimuth 68°). The colors of the solid circles show the time progression of the events.

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Figure 15. (a) The lower hemispherical projection of the moment tensor. The plus sign + next to a station name shows that the polarity at that station is positive. The plus sign with letter P shows the location of compressional axis. $p1$ and $p2$ show the two nodal planes. Plane $p2$ is identified as the approximate fault on which the 1988 Saguenay mainshock occurred. (b) Waveform comparison between the observed long-period Rayleigh wave (10 s ~ 40 s) records at station HRV and the synthetic seismograms generated using the moment tensor, shown at panel (a). In each waveform pair, the upper trace is the observed waveform, while the lower trace is the synthetic waveform. The characters and numbers on the left side of each pair are, from top to bottom: the station code, component (TAN, RAD, and VER), the epicentral distance and the azimuth in degrees, and the maximum amplitude ratio between the observed and the synthetic.

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