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**PETROGRAPHY AND PHASE EQUILIBRIA MODELING OF MID-
P ALUMINOUS GNEISSES DERIVED FROM
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Abstract

A large portion of the hinterland of the central Grenville Province is characterized by mid-pressure granulite-facies metamorphic rocks. In the Manicouagan – Lac du Milieu region of Quebec, aluminous gneisses derived from hydrothermally altered felsic protoliths provide a record of anatectic processes and of the pressure-temperature (P – T) evolution during orogenesis. Samples collected in areas separated by several tens of kilometers consist of garnet + biotite + quartz + K-feldspar $\pm$ plagioclase $\pm$ sillimanite, with retrograde cordierite in some, and most display microstructural evidence of partial melting. However, they have a wide range of bulk compositions and textures, and may be grouped in two types: sillimanite-rich rocks, with a high alumina index in AFM space and sillimanite-poor rocks with low alumina index and in which sillimanite is not part of the peak assemblage.

Phase equilibria modeling with Thermocalc constrained the P – T field of the peak mineral assemblage(s) at 800–900 °C and 6–11 kbar, with melt solidification in the range

22 of 800–865 °C and 6–8 kbar. The presence of sillimanite inclusions in garnet, and the
23 scarcity of retrograde cordierite are consistent with moderate dP/dT gradient ‘hairpin’ P –
24 T paths. The data suggest that this part of the mid- P hinterland in the central Grenville
25 experienced a rather uniform high-temperature metamorphic evolution, with large
26 temperature variations relative to pressure. This is consistent with the thermal evolution
27 of middle crust beneath an orogenic plateau.

28 **Key words:** Phase Equilibria Modeling; Aluminous rocks; Grenville Province; Partial
29 melting; Pseudosection.

30
31 **Introduction**

32 High-temperature metamorphic processes play a major role in the tectonic
33 evolution of continental crust in mountain belts, and most particularly in “large hot
34 orogens” (Beaumont et al., 2006) where the middle to lower crust experiences partial
35 melting, melt transfer and ductile rheology (Brown, 2007). Formerly deep crust is widely
36 exposed in ancient orogens by late tectonism and erosion. A prime example of such an
37 orogen is the Grenville Province, which was active between 1.09 and 0.98 billion years
38 ago (Rivers et al, 2012; Rivers, 2008) along the eastern part of North America. Large
39 portions of the hinterland of the central Grenville Province reached mid-pressure
40 granulite-facies metamorphic conditions, and constitute prime material for the
41 investigation of high temperature metamorphic processes in deep orogenic crust.

42 Aluminous rocks are particularly useful for metamorphic studies involving
43 anatexis because in addition to their mineral assemblages being sensitive to changes in
44 pressure (P) and temperature (T), they are also the most fertile with respect to melting
45 due to the presence of micas (Spear et al., 1999). Most studies of aluminous systems in
46 the literature have focused on metasedimentary rocks (e.g. metapelites and
47 metagreywakes). However, aluminous bulk chemistries may also be derived from
48 hydrothermal alteration of a wide range of felsic to mafic protoliths (Bonnet and
49 Corriveau, 2007), in which case they may display significantly different features
50 relative to typical metasedimentary rocks (Lasalle and Indares, 2014).

51 The aim of this contribution is to investigate phase relations, melt productivity
52 and P – T paths recorded by gneisses that have acquired (to some extent) their aluminous
53 character by pre-metamorphic hydrothermal alteration of felsic protoliths and display a
54 wide range of chemical compositions and microstructures. This paper builds on
55 Hindemith et al. (2017) who focuses on the origin of these rocks, and discussed in detail
56 the evidence regarding the hydrothermally altered nature of samples from most of these
57 localities. The rocks come from three locations separated by several 10s of kilometers in
58 the hinterland of the central Grenville Province, and the new P – T data provide constraints
59 on the thermal evolution of this part of the orogen. The general approach used involves
60 documentation of microstructures, mineral chemistry and phase equilibria modeling.

61

62 **Geological context**

63

64 The Grenville Province mainly consists of accreted arc terranes developed along
65 the southeast margin of proto-Laurentia between 1700 and 1230 Ma (Gower and Krogh,
66 2002; Rivers and Corrigan, 2000) that were subjected to a continental collision between
67 1090 and 980 Ma (Grenvillian Orogeny; see review in Rivers, 2008). Large parts of the
68 orogenic hinterland were metamorphosed under mid-*P* (and locally high-*P*) granulite-
69 facies conditions during the Ottawan phase of the Grenvillian Orogeny (1090–1020 Ma),
70 resulting in partial melting and high crustal ductility, and constitute a key argument for
71 the development of an orogenic plateau (Rivers, 2008). These parts are juxtaposed with
72 domains that experienced lower grades of metamorphism (low-*P* segment and Orogenic
73 Lid; Rivers, 2012), in a configuration that is interpreted to be the result of orogenic
74 collapse during the late-Ottawan (Rivers, 2012). The aluminous rocks of this study come
75 from the Manicouagan and Lac du Milieu areas of the mid-*P* hinterland in the central
76 Grenville Province (Fig. 1).

77

78 **The Manicouagan and Lac du Milieu areas**

79 The mid-*P* hinterland south of Manicouagan (Fig. 1) has been described by
80 Dunning and Indares (2010) and Indares and Moukhsil (2013), whereas its eastern
81 extension in the Lac du Milieu area was documented by Moukhsil et al. (2012). The two
82 areas display a wide variety of pre-Grenvillian plutonic and supracrustal rocks ranging in
83 age from ~1.7 Ga to ~1.2 Ga, as well as Grenvillian age granite and anorthosite. The

84 main lithologic associations of interest for this study are the Complexe de la Plus-Value
85 (PLV), and the layered bimodal suite (LBS).

86 The PLV is a metasedimentary sequence that extends from the Lac du Milieu area
87 to the southern Canyon domain, south of the Manicouagan reservoir (Fig. 1). It is made
88 up of quartzofeldspathic gneiss, metapelite, quartzite with minor marble and calcsilicate
89 rocks, and is inferred to have been deposited at ~1.5 Ga on the margin of Laurentia
90 (Moukhsil et al, 2012; Lasalle et al, 2013). Locally, the PLV in the Lac du Milieu area
91 (PLV-LM) contains an association of rusty-weathering aluminous gneiss – fine-grained
92 quartz-rich rocks with aluminous minerals. This association is interpreted to be derived
93 from pre-metamorphic hydrothermal alteration of the paragneisses but the exact
94 protoliths are unknown (Indares and Letourneau, in press).

95 The LBS, exposed in the central part of the Canyon domain (southern
96 Manicouagan; Fig. 1), is made up of mafic and felsic layers, inferred to represent
97 remnants of a volcanic belt emplaced in an extensional crustal setting (Indares and
98 Moukhsil, 2013), at 1238 ± 13 Ma (Lasalle *et al.*, 2013). Some felsic layers are white
99 gneisses of andesitic composition that locally grade into aluminous varieties and
100 garnetite, inferred to represent hydrothermally altered felsic volcanic rocks, mainly by
101 addition of $\text{Fe} + \text{Mg} \pm \text{Mn}$ (Hindemith et al., 2017).

102 In addition, an association of aluminous rocks – fine-grained quartz-rich rocks
103 with aluminous minerals – white felsic gneiss is also exposed amongst tonalitic
104 migmatites of the Gabriel complex, on the Gabriel Islands (eastern Manicouagan
105 reservoir; Indares and Dunning, 2004; Strowbridge, 2006; Hindemith et al., 2017; Fig. 1).

106 However, the age of the Gabriel Complex is unknown, and the affiliation of this
107 association, which will be referred to here as the Gabriel aluminous gneisses (GAG), is
108 uncertain.

109 Mid-*P* granulite facies metamorphism in the Manicouagan and Lac du Milieu
110 areas is manifested by the presence of orthopyroxene $\pm$ leucosome (quartz and feldspars)
111 in mafic rocks, and garnet + sillimanite + biotite + K-feldspar + plagioclase $\pm$ leucosome
112 (quartz and feldspars) in aluminous rocks. At the outcrop scale, leucosomes are abundant
113 in the metapelites of the PLV, which have a marked migmatitic character, but they are
114 less obvious in the other aluminous gneisses. The age of metamorphism is constrained
115 between 1080 and 1040 Ma based on monazite and zircon data from the Manicouagan
116 area (Dunning and Indares, 2010; Lasalle *et al.*, 2014). The only modern published
117 metamorphic study of aluminous rocks in this area was done by Lasalle and Indares
118 (2014), who used phase equilibria modeling to investigate the *P–T* evolution of
119 metapelites of the PLV and some hydrothermally altered felsic rocks of the LBS.
120 However, it was not successful in retrieving *P–T* paths for the hydrothermally altered
121 rocks. This contribution focuses on aluminous gneisses from the PLV-LM (aluminous
122 and quartz-rich rock association), the LBS in the Canyon domain, and the GAG.

123

124 **Petrography**

125 The samples from all three locations have the mineral assemblage of garnet +
126 biotite + quartz + K-feldspar $\pm$ sillimanite $\pm$ plagioclase with a wide range of textures and
127 mineral proportions. In addition, cordierite is present in the PLV-LM and in some of the

128 GAG samples. The general distribution of phases was imaged by means of false color
129 mineral maps of polished thin sections (Fig. 2). These maps were acquired with a FEI
130 Quanta 600 scanning electron microscope (SEM) and JKTech Mineral Liberation
131 Analysis (MLA) software at the Bruneau Center, Memorial University of Newfoundland.
132 Analytical conditions included accelerating voltage of 25 kV, a 10 nA beam current, and
133 a dwell time of 10 ms. This method also produced mineral proportions, listed in Table 1.
134 Detailed key microstructures observed by optical microscopy are shown in Fig. 3.

135 In terms of the proportion and mode of occurrence of aluminosilicate, the samples
136 can be classified as: (a) sillimanite-rich, where sillimanite is part of the main assemblage,
137 and occurs in the matrix and in large proportions relative to the total aluminous minerals;
138 (Table 1; Figs. 2a–c, f, h) and (b) sillimanite-poor, where sillimanite is restricted within
139 aluminous nodules, inclusions in garnet, or absent altogether (Table 1; Figs. 2d, e, g). In
140 addition, two of the samples are quartz rich, and these are the most fine-grained (Figs. 2c,
141 h).

143 **Samples from the PLV-LM**

144 The samples from the PLV-LM belong to the sillimanite-rich group, but the ratio
145 of quartzofeldspathic to aluminous minerals (Table 1), and their microstructures are
146 varied. Sample 10-AI-20a consists of diffuse layers rich in sillimanite, biotite and
147 elongated clusters of framboidal garnet, in a quartzofeldspathic matrix (Fig. 2a). Sample
148 10-AI-58-3b1 is mostly made of interwoven sillimanite+biotite, contains elongate garnet
149 porphyroblasts and has scarce quartz and feldspars, mainly concentrated in a boudinaged

150 layer (vein?) (Fig. 2b). Finally, sample 10-AI-58-2A is quartz rich and fine-grained,
151 shows an even distribution of the aluminous minerals in the matrix and contains a quartz
152 vein with rafts of feldspars (Fig. 2c). We note that the later 2 samples represent distinct
153 aluminum-rich and quartz-rich layers of the same outcrop.

154 All samples exhibit strong foliation defined by the aluminous minerals. Garnet
155 porphyroblasts and clusters contain inclusions of acicular sillimanite, biotite, quartz and
156 feldspar, or polyminerallic inclusions of quartz, sillimanite and biotite rimmed by
157 feldspar (Fig. 3a). In addition, garnet is also corroded by biotite + sillimanite ±
158 plagioclase, or cordierite (Fig. 2a). Sillimanite in the matrix is prismatic, and biotite
159 commonly forms symplectites with quartz, and locally feldspars, in the vicinity of garnet
160 (Fig. 3b). Cordierite occurs as elongated interstitial patches (Figs. 2b, c, 3c) which
161 replace sillimanite and garnet (Fig. 3d). Plagioclase is minor, and in 10-AI-58-3B1 it is
162 restricted to sites of garnet replacement, and around K-feldspar.

163

164 **Samples from the LBS**

165 The selected two samples from the LBS belong to the sillimanite-poor group and
166 lack cordierite (Table 1). In 331-e2-11, the aluminous minerals are mostly concentrated
167 in elongate clusters, which define the foliation. Sillimanite is restricted to these clusters,
168 where it is overgrown by garnet. Garnet is porphyroblastic and contains both
169 monomineralic (quartz, sillimanite, biotite) and polyminerallic inclusions where biotite
170 and rounded quartz grains are surrounded by thin films of feldspars (Figs. 3e, f). Garnet
171 rims are partly replaced by biotite, which locally forms symplectites with quartz. The

172 matrix is rich in K- feldspar, with plagioclase mainly concentrated at the rims of the
173 aluminous clusters. Quartz occurs as scattered and resorbed domains, some of which have
174 angular shape and were interpreted by Hindemith et al. (2017) as relict xenocrysts.

175 Sample 216a consists of a garnet-bearing layered quartzofeldspathic matrix, and
176 has very small amounts of biotite and sillimanite. Garnet has a wide range of sizes and
177 shapes, and some porphyroblasts are elongate and overgrow sillimanite. Biotite occurs
178 scattered in the matrix, or clustered at garnet rims. Sillimanite in the matrix, is scarce, and
179 occurs in seams mantled by plagioclase. Quartz largely occurs as variably recrystallised
180 and resorbed ribbons.

181

182 **Samples from the GAG**

183 Two of the three samples selected from the GAG (HJ-58c and HJ-57-C2) belong
184 to the sillimanite-rich group, but sillimanite is restricted to a single garnet inclusion in 10-
185 AI-76. Sample HJ-58c consists of a quartzofeldspathic matrix with plagioclase-rich
186 layers, disseminated garnet, biotite and sillimanite, and a few garnet porphyroblasts (Fig.
187 2f). Biotite and sillimanite define a foliation. Garnet porphyroblasts contain inclusions of
188 biotite, sillimanite, quartz, and K-feldspar, and are locally replaced by biotite
189 ($\pm$ symplectitic with quartz), sillimanite $\pm$ plagioclase.

190 Sample 10-AI-76 is characterized by a K-feldspar-rich matrix with an aluminous
191 nodule, scattered garnet of varied sizes, and interstitial biotite (Fig. 2g). The aluminous
192 nodule consists of spinel, corundum, and sillimanite enclosed in garnet partly overgrown
193 by biotite (Fig. 3g) and mantled by plagioclase. Sample HJ-57-C2 is fine-grained,

194 homogeneous and rich in quartz and sillimanite (Fig. 2h). It also contains cordierite,
195 which is variably altered to pinite and occurs as wisps in the matrix, as well as associated
196 with garnet (Fig. 3h)

197

198 **Interpretation of the microstructures**

199 The dominant mineral assemblage and the textures observed in the rocks of the
200 three locations are consistent with the biotite dehydration reaction: biotite + sillimanite +
201 quartz $\pm$ plagioclase $\rightarrow$ melt + garnet + K-feldspar. Acicular sillimanite inclusions within
202 elongate garnet, and relict reactants surrounded by former melt films (to be discussed
203 below) are evidence of this reaction in the forward direction. Furthermore, corrosion of
204 garnet by biotite (locally with quartz symplectites) $\pm$ sillimanite $\pm$ plagioclase is
205 indicative of this reaction in reverse.

206 Cordierite is interpreted as retrograde in sample 10-AI-58-3B1 from the PLV-LM
207 because it replaces garnet and locally, sillimanite (Fig. 3d). However, in the quartz-rich,
208 fine-grained samples 10-AI-58-2A from the PLV-LM, and HJ-57-C2 from the GAG, it is
209 not conclusive whether cordierite is part of the peak assemblage or a retrograde phase
210 because they lack reaction microstructures.

211 The aluminous nodule from sample 10-AI-76 of the GAG is interpreted as a relict
212 hydrothermal alteration feature in Hindemith et al. (2017). The local enrichment of
213 alumina resulted in a chemical potential gradient of Al in the protolith, and was later
214 overprinted by granulite-facies metamorphism.

215 Samples from all locations display microstructures that are consistent with the
216 presence of former melt, including: pools of feldspar with resorbed quartz $\pm$ sillimanite $\pm$
217 biotite, variably enclosed in garnet porphyroblasts (Figs 3e, f) and inferred to represent
218 pseudomorphs after melt; biotite-quartz symplectites concentrated at garnet rims (Fig.
219 3b); and straight edges of garnet against former melt inclusions (Figs. 3e; Vernon and
220 Collins, 1988; Sawyer, 1999; and Holness et al., 2011). In addition, the interstitial shape
221 of cordierite in 10-AI-58-3B1 (Fig. 2b) suggests that this phase may have grown in sites
222 formerly occupied by melt, and corrosion of garnet in these sites (Fig. 3d) is consistent
223 with melt crystallization in the cordierite stability field. Finally, a boudinaged
224 quartzofeldspathic layer in the same sample, may represent a former melt pathway.

225 226 **Mineral Chemistry**

227 Some mineral composition data were compiled from previous work (GAG:
228 Strowbridge, 2006; LBS: Hindemith et al., 2017; and 216a: Lasalle and Indares, 2014).
229 Samples from the PLV were analysed at Carleton University on a Cameca Camebax
230 MBX electron microprobe in WDS mode with a 20kV accelerating potential, a 20nA
231 beam current, and beam size of 5 microns for all minerals except feldspars for which a 10
232 micron beam was used. Additional data from the GAG were acquired at Memorial
233 University of Newfoundland on a JEOL JXA-8230 electron microprobe in WDS mode,
234 using a 15kV accelerating potential, a 20 nA beam current and a 3 micron beam. The
235 chemical composition ranges of garnet, biotite and feldspars are listed in Tables 2 to 4
236 and key parameters (averages per sample) are graphically presented in Fig. 4.

237

238 ***Garnet***

239 In most cases, garnet is almandine-rich, with moderate pyrope, and minor (less
240 than 5% each) grossular and spessartine contents (Table 2). The composition ranges of
241 garnet from the 3 locations overlap (Figs. 4a, b) and garnets from the PLV-ML have the
242 highest x(Fe) ratios (0.64–0.76 vs 0.53–0.76 in the and GAG). One exception is garnet in
243 HJ-57-C2 (GAG) where spessartine reaches 26–34% of garnet, and x(Fe) is as low as
244 0.30–0.37. Garnet porphyroblasts in all locations show a 1–7% increase in almandine and
245 equivalent decrease in pyrope at the rims, whereas small grains are homogeneous.

246

247 ***Biotite and Cordierite***

248 Biotite is characterized by x(Fe) in the range of 0.30–0.46 in the PLV-LM and
249 LBS, and 0.32–0.50 in the GAG, with values as low as 0.06–0.20 in sample HJ-57-C2
250 (Table 3 and Fig. 4c). Ti and Al^{VI} contents in biotite cover wide ranges. In individual
251 samples, biotite in contact with garnet has slightly lower x(Fe) than in the matrix.
252 Cordierite is present in two samples from the PLV-LM, and one from the GAG. It is Mg-
253 rich, with an x(Fe) of 0.22–0.26 in the PLV-LM and 0.05 in the GAG.

254

255 ***Feldspars***

256 The anorthite contents of plagioclase are in the range of 8–30% (Table 4).
257 Plagioclase from the LBS is the most sodic and that from the PLV-LM is the most calcic
258 (Fig. 4d). Plagioclase grains are generally homogeneous, but there are some exceptions:

259 in samples from the LBS, and in the sample with aluminous nodules of the GAC (Figs.
260 2d, e, g) plagioclase becomes increasingly An-rich with proximity to garnet (e.g. 8–20%
261 in the matrix versus 17–20% near garnet in 331-E2-11; see also Hindemith et al., 2017);
262 and in sample 10-AI-20A from the PLV-LM, plagioclase cores are enriched in Ca
263 relative to the rims. K-feldspar has Or_{76–92}.

264

265 **Interpretation of mineral chemistry**

266 The variation in mineral chemistry between samples may reflect differences in the
267 bulk compositions and/or differences in the metamorphic evolution of the three sampled
268 localities. In contrast, variations within individual samples reflect incomplete re-
269 equilibration during cooling from the metamorphic peak.

270 The generally uniform composition of garnet cores is consistent with
271 homogenisation at high metamorphic *T*. The slight increase in x(Fe) at the rim of garnet
272 porphyroblasts suggests partial resetting during cooling since higher x(Fe) indicates
273 lower *T* (Spear, 1991). In combination with the slightly higher spessartine contents at the
274 rim, this zoning is indicative of a retrograde net transfer reaction (Spear, 1991). This is
275 consistent with textural evidence of biotite + sillimanite ± plagioclase growth at the
276 expense of garnet rims, inferred to have occurred during melt crystallization. In addition,
277 the observation that biotite has the lowest x(Fe) where it is adjacent to garnet suggests
278 that there was also retrograde Fe-Mg exchange between these two phases (Spear, 1991).
279 Within this framework, the composition of the garnet cores is inferred to represent the
280 highest metamorphic *T*. However, it is possible that at early stages of cooling the

281 composition of the entire garnet was reset, this being most likely for small garnets that do
282 not show retrograde zoning (Spear, 1991). Therefore, metamorphic conditions predicted
283 by using garnet cores should be considered as ‘minimum’ peak conditions. Concerning
284 biotite, because a large part of it may have been produced during melt crystallisation, and
285 because its composition resets more easily during cooling (Spear 1991), its composition
286 is not appropriate for the determination of the thermal peak.

287

288 **Phase equilibria modeling**

289 The P – T pseudosections in this study were calculated with THERMOCALC 3.33
290 (Powell and Holland, 1988) using the bulk composition of each sample in the Na_2O –
291 CaO – K_2O – FeO – MgO – Al_2O_3 – SiO_2 – H_2O – TiO_2 – O (NCKFMASHTO) system. Mn was
292 not considered in this system because of the low contents in the bulk composition and
293 this omission should not affect the results and interpretation significantly, except for HJ-
294 57-C2, which was not used for P – T determinations.

295

296 **Bulk Compositions**

297 The bulk compositions of the samples were calculated as molar % of oxides by
298 combining modal proportions and the average chemical composition for each mineral
299 (average of rim and core chemical analysis of many grains in all textural settings),
300 determined with the SEM–MLA and the EMP, respectively. This was the method of
301 choice because: (a) it gives bulk compositions representative of the investigated thin
302 sections; and (b) it allows the calculation of a bulk H_2O on the basis of the major hydrous

303 minerals, which is important for the location of the solidus in the pseudosections. It also
304 allows one to exclude secondary minerals and alterations, sulfides, and minor phases such
305 as apatite, monazite and zircon which cannot be accounted for in the NCKFMASHTO
306 system. The bulk H₂O was calculated as the sum of the H₂O in biotite and cordierite
307 using the methods outlined in White et al. (2007) and Harley et al. (2002), respectively.

308 The bulk compositions are listed in Table 5, and are plotted on an AFM diagram
309 (Fig. 5). Most have x(Fe) between 0.50 and 0.63 with the exception of HJ-57-C2.
310 Samples from the PLV-LM have higher alumina index than those from the LBS (~0.32–
311 0.50 vs ~0.15–22), whereas the alumina index in those from the GAG has a wide range of
312 values. Samples with an alumina index of >0.25 and <0.25 correspond to the sillimanite-
313 rich and sillimanite-poor groups, respectively. One sillimanite-rich sample, HJ-57-C2,
314 has an anomalous bulk composition that plots far from the main clusters, with much
315 higher alumina index, and x(Mg), consistent with advanced argillic alteration.

316

317 ***P–T* pseudosections**

318 The calculations used the internally consistent dataset tcds55 (Powell and
319 Holland, 1988), created November 22, 2003, and the phases liquid – quartz – garnet – K-
320 feldspar – plagioclase – sillimanite – biotite – cordierite – muscovite – rutile – ilmenite –
321 orthopyroxene were considered in the calculations. The following a–x models were used:
322 garnet (White et al., 2007), biotite (White et al., 2007), cordierite (Holland and Powell,
323 1998), muscovite (Coggon and Holland, 2002), plagioclase and K-feldspar (Holland and
324 Powell, 2003), orthopyroxene (White et al., 2002), ilmenite (White et al., 2000) and melt

(White et al., 2007). The NCKFMASHTO system also takes into account Fe^{3+} , which is calculated based on an input value of O (oxygen) in the bulk composition. For consistency, a low value of oxygen equal to 0.05 was used, which is supported by the scarcity of ilmenite, lack of magnetite and negligible proportion of andradite in garnet.

The pseudosections were calculated in the P – T range of 3–14 kbar and 700–1000 °C. The full topologies and mineral proportion isopleths (isomodes) for two pseudosections from each area are shown in Fig. 6, P – T paths and mineral composition isopleths are shown in Fig. 7 within a reduced range of 5–13 kbar and 800–950 °C, as well as melt reintegrated diagrams in Fig. 8. Finally, two additional samples were modeled in Fig. 9 to explore the stability of cordierite in these rocks.

General Topologies

Within the sillimanite stability field, all topologies of the P – T pseudosections are muscovite-free and can be divided into two main parts: (1) a mid- P area between the cordierite-in boundary and the sillimanite–kyanite transition line, and (2) a low- P area, with cordierite-bearing topologies. Cordierite becomes stable at P conditions below ~5 to 8 kbar.

Mid- P topologies of the sillimanite-rich samples (Figs. 6a, b, e) are characterized by a sub-vertical solidus within the T range of 810–840 °C, and a sub-vertical biotite-out line at higher T . Biotite is completely consumed at $T < 900$ °C in all but in the P – T pseudosection of 10-AI-58-3B1 (Fig. 6b), which is quartz poor (see Table 1). In this case quartz is eliminated first, and biotite is stable up to 1050 °C (outside of the range of the

347 diagram). Mid- P topologies of the sillimanite-poor samples show a sub-vertical solidus at
348 820–870 °C, but up- T orthopyroxene becomes stable before the elimination of biotite
349 (Figs. 6c, d, f), where both biotite and orthopyroxene boundaries have moderate positive
350 dP/dT gradients. In these pseudosections, the biotite-out boundary occurs at higher T than
351 in the sillimanite-rich samples (compare Figs 2c, d, f, with Figs 2a, e), and sillimanite is
352 either stable at near solidus conditions (Figs. 5c, d) or is not stable at all in the mid- P
353 topologies (Fig. 5f). The later is the case of samples with sillimanite restricted to garnet
354 inclusions. In addition, some mid- P topologies show a plagioclase-out boundary up- T at
355 near solidus conditions (10-AI-58-3B1 in Fig. 6b which has only retrograde plagioclase),
356 or at higher T . Mid- P topologies are generally extended into the kyanite stability field,
357 where muscovite becomes stable on the high- P /low- T side of some pseudosections, but is
358 out of range for others.

359 Low- P topologies are controlled by shallow dP/dT gradient field boundaries
360 involving cordierite, garnet and orthopyroxene in the case of the sillimanite-rich samples.
361 Minor phases include one or more of rutile, ilmenite and magnetite, where rutile is stable
362 throughout or on the high- P side when paired with ilmenite.

363 For the samples investigated, the P – T stability field of the inferred peak mineral
364 assemblage (Field I) falls within the mid- P part of the pseudosections. For the sillimanite-
365 rich samples, the peak assemblage liquid + biotite + garnet + quartz + sillimanite + K-
366 feldspar $\pm$ plagioclase is stable between the solidus and the biotite-out boundary, or
367 between the plagioclase-out and quartz-out boundaries (sample 10-AI-58-3B1; Fig. 6b).
368 For the sillimanite-poor samples, the peak assemblage liquid + biotite + garnet + quartz +

369 K-feldspar + plagioclase is stable between the solidus and the orthopyroxene-in
 370 boundary, in the case where sillimanite is restricted to garnet inclusions (Fig. 6f); or the
 371 sillimanite-out and orthopyroxene-in boundaries (Figs. 6c, d). In terms of T and P
 372 conditions, Field I falls within the range of 820–900 °C in the PLV-LM and LBS, and
 373 830–920 °C in the GAG, at the P interval between ~6–7 (defined by the cordierite-out
 374 boundary) to 10–12 kbar (defined by the sill–ky transition).

375

376 *Isomodes and isopleths*

377 The isomodes for plagioclase, sillimanite, biotite, garnet and liquid were
 378 calculated in Field I for each sample and are shown in Fig. 6. In this field, the proportions
 379 of plagioclase, biotite (and sillimanite in the sillimanite-rich samples) decrease, while
 380 liquid and garnet increase with increasing T . These trends are consistent with biotite
 381 dehydration melting according to a generalized reaction: biotite + quartz $\pm$ sillimanite $\pm$
 382 plagioclase $\rightarrow$ garnet + K-feldspar + melt. The maximum melt fraction within Field I
 383 varies between 3 and 10 modal % at the biotite-out boundary of Field I, and reaches 15%
 384 at the quartz-out boundary in the case of 10-AI-58-3B1 (Fig. 6b). These values seem to
 385 be correlated with the proportion of biotite in the solidus (and the T interval between the
 386 solidus and the biotite-out boundary) in accordance to the above mentioned reaction.

387 The isopleths $x(g)$, $z(g)$, $ca(pl)$ and $x(bi)$ in Field I for each sample are shown in
 388 Fig. 7. These are defined in THERMOCALC as $x(g) = Fe/(Fe+Mg)$ and $z(g) =$
 389 $Ca/(Fe+Mg+Ca)$ of garnet, $ca(pl)$ is the anorthitic content in plagioclase, and $x(bi)$ is
 390 $Fe/(Fe+Mg)$ of biotite. Within the fields of interest, the $x(g)$ and $x(bi)$ isopleths have a

steep, positive dP/dT gradient and their values decrease with increasing T except where the peak mineral assemblage is sillimanite-free, in which case they have a steep negative dP/dT gradient (Figs 7c, d, f). The $z(g)$ isopleths generally have a shallow positive dP/dT gradient and values that increase with P , however, in fields that lack plagioclase, $z(g)$ becomes sub-vertical, and in sillimanite-absent fields, these have a moderate negative slope. Isopleths of $ca(pl)$ typically have moderate positive dP/dT gradients.

P–T paths

The $P–T$ pseudosections discussed above can potentially constrain the $P–T$ evolution of the investigated rocks within the $P–T$ field(s) of the observed mineral assemblages. For instance, (a) the intersection of relevant garnet composition isopleths corresponding to measured garnet compositions, can be used to narrow down the peak $P–T$ conditions; and (b) because the retrograde phases produced depend on the angle at which the $P–T$ path cuts their isomodes, the distribution of these isomodes can be used to predict the shape of the $P–T$ paths during melt crystallization. In addition, the actual mineral proportions in the rock compared with those along the solidus should match at the $P–T$ conditions of melt crystallization.

Table 6 shows the $x(g)$, $x(bi)$ and $z(g)$ isopleths for each sample, and the predicted $P–T$ coordinates of the peak and of melt crystallization. Isopleths $z(g)$ and $x(g)$ from the cores of large garnet were used to constrain the peak $P–T$ and isopleths $x(bi)$ are shown for comparison. However, in several cases, the $x(g)$ isopleths, which are T -dependent, are close to the solidus, suggesting potential reequilibration of garnet cores

413 during cooling. Additionally, isopleths of $z(g)$ are widely spaced in the fields of interest
414 and provide little constrain on the P of the peak and of melt crystallization. Finally, the
415 mineral modes do not change significantly along the solidus between the cordierite
416 boundary and the sillimanite-kyanite transition line, so the P of melt crystallization
417 cannot be well constrained unless the sample has retrograde cordierite (see Table 7 for
418 comparison of mineral modes at the solidus with the modes of the thin section). For these
419 reasons, a range of possible P – T paths are shown in grey (Fig. 7).

420 The P – T paths for the samples from the PLV-LM are shown in Figs. 7a and b.
421 The minimum peak T and P predicted by the intersection of $z(g)$ and $x(g)$ isopleths is
422 840–865 °C and 9–11 kbar. The moderate replacement of porphyroblastic garnet by
423 biotite + sillimanite $\pm$ plagioclase $\pm$ quartz in these rocks is consistent with the retrograde
424 portion of the P – T path cutting the isomodes for these phases at a high angle, therefore
425 following a moderate dP/dT gradient. In addition, the presence of retrograde cordierite in
426 10-AI-58-3B1, constrains the P – T path in Fig. 7b to cross the solidus in the cordierite
427 stability field (at ~820–840 °C and 6–7 kbar).

428 For the two samples from the LBS, the $x(g)$ isopleths coincide with the solidus so
429 the peak T is only constrained by the entire T range of Field I, at 830–875 °C with a P of
430 8–11 kbar (Figs. 7c, d). The replacement of garnet rims by biotite in these samples is
431 consistent with a P – T path at an angle to the biotite isomodes and a peak T that is higher
432 than the T of the solidus. The conditions of melt solidification were constrained to 825–
433 875 °C and 6.5–8 kbar based on the lack of orthopyroxene and cordierite. The lack of

434 retrograde sillimanite in these samples is consistent with sillimanite becoming stable in
435 the pseudosections at near solidus conditions.

436 In the pseudosections of the two samples from the GAG, the $x(g)$ isopleths
437 coincide with the solidus, however, based on the T range of Field I in each sample and
438 the $z(g)$ isopleth of the garnet core, the peak T is constrained to 840–845 °C at P of 8–11
439 kbar (Figs. 7e, f). Partial replacement of garnet rims by sillimanite and biotite is
440 consistent with a shallow to moderate dP/dT gradient. Because of the lack of cordierite in
441 these samples, the P – T path may cross the solidus anywhere above the cordierite or
442 orthopyroxene boundaries, at 835–840 °C and 6–8 kbar.

443

444 **Melt reintegration**

445 All the P – T pseudosections discussed earlier are characterized by a steep solidus
446 and the lack of muscovite from the mid- P topologies. These features, together with the
447 good preservation of the peak mineral assemblages, are consistent with some melt loss
448 (White and Powell, 2002). This is common for anatectic rocks in orogenic environments
449 because the melt is squeezed out under deviatoric stress (Brown, 2007). The current bulk
450 compositions therefore represent those attained after the loss of melt. Considering this,
451 topologies calculated at T below the solidus (liq=0 line) in these diagrams are not valid,
452 and this hampers the prediction of prograde P – T paths. For this reason, additional
453 pseudosections were calculated using melt-reintegrated compositions for selected
454 samples to outline the topologies relevant to the prograde P – T paths.

455 To construct these diagrams, the composition of the first infinitesimal amount of
456 melt was calculated where the prograde portion of the path was projected to have crossed
457 the solidus. Although this P is not well constrained, it has been shown that the
458 suprasolidus topologies of the melt-reintegrated pseudosections are not significantly
459 affected by P (Indares et al., 2008; Groppo et al., 2012). Proportions of melt with this
460 composition were then added to the original bulk composition until an amount sufficient
461 to make the water saturated solidus stable was achieved. Therefore, in using this method,
462 we assume that the rock was water saturated and that the melt loss occurred in one step.

463 It is not possible to calculate the exact amount of lost melt, and in the case of
464 rocks with complex earlier histories, as the ones investigated here, the assumption of
465 water saturated solidus is questionable. However, it has been determined that the
466 suprasolidus topologies of the melt-reintegrated pseudosections do not change
467 significantly with varying quantities of added melt because they are largely controlled by
468 the solid phases (White and Powell, 2002). Therefore, these melt-reintegrated phase
469 equilibria diagrams can provide some insight on the prograde portion of the paths, and, in
470 addition, allow one to assess the potential melt productivity of the examined rock types.

471 Fig. 8 displays melt-reintegrated P – T pseudosections for 5 samples, a diagram for
472 sample 216a is not included because of difficulties in calculating it. In each diagram, the
473 thick dashed white line represents the “dry” solidus calculated with the actual bulk
474 compositions of the rock (before melt reintegration; see Fig. 6). The “wet” solidus in the
475 melt-reintegrated diagrams occurs at much lower T , outside of the range of the
476 pseudosection. The portion of the pseudosection at T above the “dry” solidus is quite

477 similar in the both the original and melt-reintegrated pseudosections, and therefore will
478 not be discussed again here.

479 The amount of melt added in order to produce a water-saturated solidus was
480 higher (20–25 modal %) in the sillimanite-rich samples HJ-58c and 10-AI-20-A, and
481 much lower (8–12 modal %) in the sillimanite poor samples (331-e2-11 and 10-AI-76)
482 and in the sillimanite-rich but quartz-poor 10-AI-58-3B1. The higher range compares
483 well with that calculated by Lasalle and Indares (2014) for sillimanite-rich rocks from the
484 LBS and a metapelite from the PLV.

485 In the melt-reintegrated diagrams, melt isomodes increase with T between the wet
486 solidus and biotite-out boundary (the dehydration melting region), and have moderate
487 dP/dT slopes (compared with steep dP/dT slopes in the original bulk pseudosections). For
488 the sillimanite-rich samples, the predicted amount of melt reaches ~36–40% at the
489 biotite-out boundary, and 30% at the quartz-out boundary in 10-AI-58-3B1 (Figs. 8a, b,
490 d). In contrast, the predicted melt production for the sillimanite-poor samples at the same
491 boundary reaches only ~20–30% (Fig. 8c, e). Melt production in the melt reintegrated
492 pseudosections is much greater between the conditions of the dry solidus and the biotite-
493 out boundary than in pseudosections calculated using the original bulk compositions.

494

495 *General topologies in melt reintegrated pseudosections and P – T paths*

496 For all samples, the stability field of muscovite has been shifted to lower P in the
497 melt-reintegrated diagrams. The appearance of sillimanite or K-feldspar coincides
498 approximately with the disappearance of muscovite for at least a portion of the muscovite

499 boundary. A sub-vertical garnet-in boundary occurs at the T range of 700 – 800 °C.
500 Because Mn, which expands the stability of garnet to lower T (White et al., 2014), was
501 not considered in the bulk composition, yet a small percentage of garnet is spessartine,
502 the T of the garnet-in boundary in the pseudosections is likely overestimated.

503 The prograde portions of the P – T paths were constrained by the sillimanite–
504 kyanite transition line since garnet overgrows sillimanite to a large extent. This is
505 consistent with a moderate dP/dT prograde P – T path. For sample 10-AI-76, there is field
506 where sillimanite becomes stable below the “dry” solidus within the T range of 700–800
507 °C. This explains the presence of sillimanite in this sample only as inclusions in garnet.

508

509 **The effect of bulk composition on cordierite stability**

510 Unlike sample 10-AI-58-3B1, where cordierite clearly replaces garnet, samples
511 10-AI-58-2a and HJ-57-C2 have significant cordierite distributed throughout the matrix
512 as wispy aggregates (Figs. 2c, 3c). These samples are fine-grained therefore prone to
513 extensive retrograde re-equilibration. Because of this, it is inconclusive as to whether or
514 not cordierite is part of the peak assemblage. Although these samples were not used to
515 infer P – T conditions and paths, their pseudosections are included in Fig. 9 to investigate
516 the role of bulk composition on the stability of cordierite. Both samples are part of the
517 sillimanite-rich group.

518 It is known that increasing $x(\text{Mg}) = \text{Mg}/(\text{Fe} + \text{Mg})$ expands the cordierite stability
519 field to higher P (Spear et al. 1999; Groppo et al., 2013) and this correlation can be seen
520 to some extent by comparing the $x(\text{Mg})$ of the bulk compositions (Table 5) with the

521 locations of the cordierite-in boundary in the calculated pseudosections. In particular by
522 comparing the P of the intersection between the solidus and the cordierite-in boundary
523 with the bulk $x(\text{Mg})$, samples can be divided into three groups. (1) Sample 10-AI-76 has
524 the lowest $x(\text{Mg})$ at <0.40 , and correspondingly, the lowest P at which cordierite
525 becomes stable, ~ 5 kbar (Fig. 6f). (2) The majority of the samples have an intermediate
526 $x(\text{Mg})$ in the range of $0.40\text{--}0.50$, and for these cordierite becomes stable between 6.01
527 and 7.11 kbar. (3) Finally, in HJ-57-C2, which has the highest $x(\text{Mg})$ of 0.74 , cordierite is
528 stable up to ~ 8 kbar (Fig. 9b). The high spessartine contents of garnet in this sample yet
529 lack of Mn used in the model may have affected the locations of boundaries in the phase
530 equilibrium diagram, and may explain the difficulty in calculating lower- P boundaries. In
531 samples with the lowest $x(\text{Mg})$ and P of the cordierite-in boundary, orthopyroxene
532 becomes stable before cordierite. Additionally, in fields that are sillimanite-absent,
533 cordierite first becomes stable at lower P compared with samples with similar $x(\text{Mg})$ in
534 sillimanite-present fields.

535 Based on the P of the cordierite-in boundary compared with the $P\text{--}T$ paths from
536 each area, it is likely that cordierite was not part of the peak assemblage of 10-AI-58-2a
537 (Fig. 9a) since the peak- P is much higher than that at which cordierite first becomes
538 stable ($\sim 6.5\text{--}7.5$ kbar for the cordierite boundary compared with $8\text{--}11$ kbar for the peak- P
539 of other samples in the area). However, because the cordierite-in boundary occurs at $\sim 8\text{--}9$
540 kbar between the solidus and biotite boundaries in HJ-57-C2, it is possible that cordierite
541 was a peak phase in this sample.

542

543 **Summary and conclusions**

544 Aluminous gneisses derived from hydrothermally altered protoliths have been
545 recognized in three distinct lithologic units and locations in the Manicouagan and Lac du
546 Milieu areas, in the hinterland of the Central Grenville Province. These rocks were
547 metamorphosed under mid-*P* granulite facies-conditions during the Grenvillian orogeny,
548 and contain the mineral assemblage garnet + biotite + quartz + K-feldspar $\pm$ plagioclase $\pm$
549 sillimanite $\pm$ cordierite with a wide range of mineral proportions and microstructures, as
550 well as evidence for anatexis.

551 In terms of the mode of occurrence (& proportion) of sillimanite, the rocks can be
552 classified into: (a) sillimanite-rich samples, which have sillimanite present in the peak
553 metamorphic assemblage and show a high Alumina index on AFM space; and (b)
554 sillimanite-poor samples, with sillimanite only as inclusions in garnet and with a low
555 Alumina index. In addition, some samples are quartz-rich and fine grained.

556 Microstructural evidence of former melt is preserved in all rocks except the
557 quartz-rich varieties, as former melt inclusions in garnet pseudomorphed by feldspars,
558 relict quartz grains surrounded by thin feldspar rims, inter-granular films in the matrix
559 and biotite-quartz symplectites. In addition, the interstitial form of cordierite in the matrix
560 of some samples suggest that this phase may have pseudomorphed former melt pockets.

561 *P–T* pseudosections (in the NCKFMASHTO system) calculated with the
562 measured bulk compositions show different topologies for the sillimanite-rich and poor
563 samples, and the predicted phase relations support microstructural evidence that in the
564 later samples, sillimanite is an earlier prograde phase, not stable at the thermal peak. In

565 addition, they provide the framework to assess the effect of the bulk $x(\text{Mg})$ on the
566 stability of cordierite, and support microstructural evidence that cordierite is of retrograde
567 origin. The P – T pseudosections are generally consistent with observed mineral
568 assemblages and compositions, with the exception of 10-AI-76. This sample has an
569 aluminous nodule containing spinel and sillimanite that was not predicted by the
570 pseudosection. This is because of the locally enriched Al bulk composition as a result of
571 relict hydrothermal alteration texture.

572 In all these P – T pseudosections, a lower- T limit for the conditions of the thermal
573 peak is constrained by the solidus whereas the upper- T limit is placed by: orthopyroxene-
574 out boundary in some sillimanite-poor rocks in which orthopyroxene appears before the
575 elimination of biotite, and the biotite-out boundary in all other rocks (except for a quartz-
576 poor variety). However, use of this limit implies that biotite was stable at the thermal
577 peak, which is not obvious based on the observed microstructures. In addition, the use of
578 mineral composition isopleths to constrain peak P – T conditions in these mid- P granulite
579 facies rocks was limited because of potential partial resetting of $x(\text{Mg})$ isopleths of
580 garnet, which is T dependent, during cooling, and the wide spacing of $z(\text{g})$ isopleths of
581 garnet in the mid- P range.

582 The P – T conditions of the metamorphic peak predicted by the P – T
583 pseudosections fall in the range 830–875 °C at 8–10 kbar, the lower T part of which is
584 probably underestimated owing to the limitations discussed above. In addition, the melt is
585 predicted to have solidified upon cooling down to at a T of 800–865 °C at 6–9 kbar
586 (Table 2.6). In this case, the T -range corresponds to the different locations of the dry

587 solidus in the different samples. The P of melt solidification is better constrained in
588 samples that contain retrograde cordierite, because in this case it corresponds to the
589 intersection between the solidus and the cordierite boundary. In addition, local retrograde
590 replacement of garnet by biotite $\pm$ sillimanite in most samples require retrograde P – T
591 paths with moderate dP/dT gradients. The prograde portion of the P – T paths was
592 investigated by means of melt-reintegrated P – T pseudosections, and based on
593 microstructural grounds they were constrained to fall below the sillimanite/kyanite
594 transition.

595 The two sets of P – T pseudosections (calculated with the measured vs melt
596 reintegrated bulk compositions) provide also some insight on the melt productivity of the
597 modeled bulk compositions. The melt proportions predicted at the biotite-out boundary in
598 the P – T pseudosections calculated with the measured bulk compositions are mainly
599 correlated with the proportion of biotite in the solidus. In contrast, the melt proportions
600 calculated at the same boundary, in the melt reintegrated P – T pseudosections are
601 correlated with the rock type, and are highest for the sillimanite-bearing samples.

602 Metamorphic P – T paths predicted by the P – T pseudosections, by comparing
603 textural and mineral composition data of the rocks with the calculated topologies, are
604 broadly similar in all locations and rock types. The generalized P – T path has an overall
605 “tilted hair pin” shape with a moderate dP/dT gradient. The data suggest that this part of
606 the mid- P hinterland in the central Grenville experienced a uniform metamorphic
607 evolution, with large T variations relative to P , consistent with the thermal evolution of
608 middle crust beneath an orogenic plateau.

609

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617

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619

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Table 1. Mineral modes.

Mineral	PLV			LBS			GAG	
	10-AI-20A	10-AI-58-3B1	10-AI-58-2A	331-e2-11	216a	HJ-58c	10-AI-76	HJ-57c2
Garnet	7.10	9.57	4.43	3.64	12.48	13.98	4.75	0.16
Biotite	12.20	34.84	7.47	19.01	4.09	6.75	5.20	5.38
Sillimanite	5.31	30.92	3.55	0.59	0.50	1.99	0.02	9.41
Cordierite	0.05	2.25	2.84	0.00	0.00	0.00	0.00	0.76
<i>Sum Al</i>	24.66	77.58	18.29	23.25	17.94	22.73	9.98	15.71
<i>Sill/Altot</i>	21.55	39.86	19.41	2.53	2.77	8.77	0.21	59.90
Quartz	36.57	12.53	68.86	14.88	32.12	35.08	23.26	73.65
K-Feldspar	27.36	6.51	4.39	39.35	30.27	12.70	48.67	10.49
Plagioclase	11.39	3.30	8.40	22.15	19.22	29.13	17.69	0.06
<i>Sum QF</i>	75.32	22.35	81.65	76.38	81.61	76.90	89.62	84.21
Rutile	0.00	0.02	0.03	0.00	0.36	0.28	0.00	0.08
Magnetite	0.02	0.00	0.00	0.00	0.08	0.09	0.00	0.00
Ilmenite	0.00	0.05	0.02	0.38	0.00	0.00	0.16	0.00
Spinel	0.00	0.00	0.00	0.00	0.00	0.00	0.25	0.00
Group	SR	SR	SR	SP	SP	SR	SP	SR

Sum Al: Sum of aluminous minerals; Sum QF: Sum of quartzofeldspathic minerals;
 Sill/Al_{tot}: Sillimanite proportion divided by Sum Al; SR: Sillimanite-rich;
 SP: Sillimanite-poor

Table 2. Compositional ranges of garnet.

Location	Sample	End Members of Grt (%)				x(Fe) (%)
		Alm	Prp	Grs(+And*)	Sps	
PLV-LM	10-AI-58-3B1	64.4-71.7	23.8-31.5	2.65-2.94	1.12-1.66	64.4-71.7
PLV-LM	10-AI-20A	65.2-75.7	19.4-29.6	2.25-3.23	1.98-2.67	65.1-75.7
PLV-LM	10-AI-58-2A	67.3-72.2	23.7-28.7	2.94-3.13	0.89-1.05	68.4-72.2
LBS ¹	331-e2-11	62.3-68.0	20.5-37.1	7.3-10.2	1.0-3.1	62.7-76.1
LBS ¹	216a	51-65	30-46	3-4	0-1	53-68
GAG ²	HJ-58C	61-65	32-36	2-3	1	63-67
GAG ¹	10-AI-76	64.3-65.9	22.3-25.2	6.3-6.6	4.5-4.9	72.0-74.5
GAG	HJ-57-C2	21.1-23.2	39.2-48.0	4.0-4.9	26.6-34.4	30.6-36.9

¹Hindemith et al. (2017) ²Strowbridge (2006) *Andradite is negligible

Alm: Almandine; Prp: Pyrope; Grs: Grossular; Sps: Spessartine, And: Andradite

x(Fe) = Fe/(Fe+Ca+Mg)

Table 3. Compositional ranges of biotite and cordierite.

Location	Sample	Biotite			Cordierite	
		Ti (p.f.u)	Al(VI) (p.f.u)	x(Fe) (%)	F	x(Fe) (%)
PLV-LM	10-AI-58-3B1	0.27-0.30	0.15-0.17	39-42	0.09-0.11	21.8-25.5
PLV-LM	10-AI-20A	0.06-0.22	0.24-0.36	30-42	0.14-0.23	X
PLV-LM	10-AI-58-2A	0.26-0.31	0.16-0.21	36-42	0.08-0.10	22.0-24.6
LBS ¹	331-e2-11	0.12-0.24	0.17-0.37	35.6-45.6	0-0.07	X
LBS ¹	216a	0.26-0.36	0.08-0.12	34-39	N/A	X
GAG ²	HJ-58C	0.28-0.29	0.10-0.11	32	N/A	X
GAG ¹	10-AI-76	0.49-0.68	0.21-0.40	38.6-49.6	0.09-0.31	X
GAG	HJ-57-C2	0.13-0.26	0.01-0.09	5.9-19.6	N/A	4.5-5.0

¹Hindemith et al. (2017) ²Strowbridge (2006)

X: none in sample; N/A: Not measured;

x(Fe) = Fe/(Mg+Fe); p.f.u: Per formula unit

Ti and Al^{VI} are on the basis of 11 oxygen

Table 4. Compositional ranges of feldspars.

Location	Sample	Plagioclase			K-feldspar		
		Ab	An	Or	Ab	An	Or
PLV-LM	10-AI-58-3B1	71.9-73.9	25.3-27.2	0.77-1.11	14.1-17.9	0.18-0.67	81.5-85.7
PLV-LM	10-AI-20A	75.8-83.2	16.0-22.7	0.72-1.57	10.8-15.9	0-2.5	83.1-86.3
PLV-LM	10-AI-58-2A	70.4-69.5	28.7-29.5	0.91-1.36	8.9-14.9	0.28-0.93	83.4-90.3
LBS ¹	331-e2-11	78.9-90.3	8.4-20.6	0.6-1.4	7.9-21.3	0.2-0.6	78.4-91.2
LBS ¹	216a	73-94	13-26	N/A	N/A	N/A	N/A
GAG ²	HJ-58C	78	20-21	1-2	8.7-18.8	0.01-1.8	79.4-91.1
GAG ¹	10-AI-76	73.3-76.3	22.7-26.9	1-1.7	8.4-19.0	0.2-0.7	76.3-92.1
GAG	HJ-57-C2	X	X	X	9.7-10.4	0.32-0.36	88.6-89.8

¹Hindemith et al. (2017) ²Strowbridge (2006)

X: none in sample; N/A: Not measured;

Ab: Albite; An: Anorthite; Or: Orthoclase

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Table 5. Bulk compositions.

	Oxides (mol %)										sum	x(Fe)	x(Na)	x(K)	x(Ca)
	SiO ₂	Al ₂ O ₃	MgO	FeO	MnO	CaO	Na ₂ O	K ₂ O	TiO ₂	H ₂ O					
PLV															
10-AI-20-A	73.75	10.25	3.8	4.56	0.09	0.67	1.49	3.52	0.3	1.58	100	0.55	0.26	0.62	0.12
10-AI-58-3B1	52.57	21.25	8.55	8.53	0.06	0.36	0.44	3.02	1.46	3.77	100	0.50	0.12	0.79	0.09
10-AI-58-2A	83.64	6.35	2.75	3.03	0.03	1.63	0.26	1.02	0.17	1.13	100	0.52	0.09	0.35	0.56
LBS															
331-e2-11	67.09	10.87	4.69	4.94	0.05	1.11	2.73	5.21	0.89	2.41	100	0.51	0.30	0.58	0.12
216a	73.97	9.38	3.94	4.91	0.00	1.36	1.75	3.60	0.63	0.46	100	0.55	0.26	0.54	0.20
GAG															
HJ-58C	72.17	9.93	4.25	6.49	0.08	1.62	2.81	1.59	0.29	0.76	100	0.60	0.47	0.26	0.27
10-AI-76	74.94	9.99	1.89	3.20	0.14	1.28	2.57	5.03	0.36	0.61	100	0.63	0.29	0.57	0.14
HJ-57c2	88.54	6.54	1.72	0.61	0.03	0.01	0.18	1.41	0.34	0.62	100	0.26	0.11	0.88	0.00
Melt reintegrated															
10-AI-20-A	71.87	9.82	3.08	3.77	0.00	0.66	1.85	3.45	0.24	5.25	100	0.55	0.31	0.58	0.11
10-AI-58-3B1	53.96	19.58	7.52	7.54	0.00	0.41	0.78	3.05	1.28	5.69	100	0.50	0.18	0.72	0.10
10-AI-76	73.94	9.78	1.73	2.96	0.00	1.22	2.63	4.85	0.32	2.58	100	0.63	0.30	0.56	0.14
331-e2-11	66.78	10.64	4.31	4.59	0.00	1.07	2.79	5.03	0.81	3.85	100	0.52	0.31	0.57	0.12
HJ-58C	68.79	9.46	4.04	6.19	0.00	1.54	2.68	1.52	0.28	5.51	100	0.60	0.47	0.26	0.27

$x(\text{Fe})=\text{Fe}/(\text{Fe}+\text{Mg})$; $x(\text{Na})=\text{Na}/(\text{Na}+\text{K}+\text{Ca})$; $x(\text{K})=\text{K}/(\text{Na}+\text{K}+\text{Ca})$; $x(\text{Ca})=\text{Ca}/(\text{Na}+\text{K}+\text{Ca})$

Table 6. Mineral isopleths of samples and P – T path coordinates.

Location	Sample	Isopleths			P – T path coordinates			
		x(g)	x(bi)	z(g)	Min Max T	Peak P	T at solidus	P at solidus
PLV	10-AI-20A	65	38	2.2- 3.2	840	9-11	820	6-7
PLV	10-AI-58-3B1	64	41	2.6- 2.9	865	8-11	839	6.1
PLV	10-AI-58-2A	67	42	2.9-3.1	X	X	841	6.5
LBS	331-e2-11	62	51	2.3- 4.2	830	9-11	825	6.5-8
LBS	216a	53	37	3-4	875	8-11	874	7.2
GAG	HJ-58C	65	40	2.4- 3.0	845	8-11	835	6.5-8.5
GAG	10-AI-76	73	43	4.5- 5.9	840	8-11	840	6-8.5

Isopleths used in determining the minimum peak – are highlighted in bold.

X: Could not be determined.

x(g)=Fe/(Fe+Ca+Mg) in garnet; z(g)=Ca/(Fe+Ca+Mg) in garnet; x(bi)=Fe/(Fe+Mg) in biotite

Table 7. Mineral modes compared with the modes at the solidus of the phase equilibria diagram.

	10-AI-20A		10-AI-58-3B1		331-e2-11		216a		HJ-58c		10-AI-76	
	TS mode	Solidus	TS mode	Solidus	TS mode	Solidus	TS mode	Solidus	TS mode	Solidus	TS mode	Solidus
Garnet	7.10	8.86	9.57	13.51	3.64	5.76	12.48	17.30	13.98	19.41	4.75	7.02
Biotite	12.20	13.91	34.84	31.95	19.01	21.76	4.09	4.31	6.75	7.77	5.20	5.50
Sillimanite	5.31	5.31	30.92	25.91	0.59	0.00	0.50	0.34	1.99	1.81	0.02	0.00
Cordierite	0.05	0.00	2.25	4.42	0.00	0.00	0.87	0.00	0.00	0.00	0.00	0.00
Quartz	36.57	36.86	12.53	12.04	14.88	15.94	32.12	32.15	35.08	34.00	23.26	24.26
Feldspars	38.75	34.99	9.81	11.47	61.50	56.16	49.49	45.19	41.83	37.00	66.35	62.63

TS mode: Mineral mode in thin section

Solidus: Mineral mode where the P – T path crosses the solidus

729 **FIGURE CAPTIONS**

730 **Fig. 1.** Generalized map of the Manicouagan and Lac du Milieu areas showing the
731 locations of the samples. Inset: location of the study area in the Grenville Province
732 (Modified after Indares and Moukhsil, 2013).

733 **Fig. 2.** SEM–MLA mineral maps for thin sections (46mm length) of aluminous gneisses
734 from the PLV, LBS and GAG. The maps were produced by Letourneau (2011; 2a – 2c);
735 Hindemith et al., 2017; 2d, 2g); Lasalle and Indares (2014; 2e) this study (2f, 2h). Sericite
736 is derived from alteration of feldspars.

737 **Fig. 3.** Photomicrographs showing characteristic microstructures: (a) polyminerallic
738 inclusion in garnet (10-AI-58-3B1, XP) (b) biotite-quartz symplectites (10-AI-20a, PPL);
739 (c) interstitial cordierite in the matrix (10-AI-58-2a, PPL); (d) cordierite corroding: (i)
740 garnet and (ii) sillimanite; cordierite pseudomorphing sillimanite prisms (10-AI-58-3B1,
741 PPL); (e) rounded quartz with thin rims of feldspar in the matrix (331-E2-11, XP); (f)
742 polyminerallic inclusions in garnet with thin rims of feldspar (331-E2-11, quartz plate);
743 (g) aluminous nodule (10-AI-76, PPL); (h) garnet and cordierite relationship (HJ-57-C2,
744 PPL). PPL: Plane polarized light, XP: Cross polars. Mineral abbreviations: garnet (g);
745 biotite (bi); sillimanite (sill); K-feldspar (ksp); quartz (q); spinel (sp); cordierite (crd).

746 **Fig. 4.** Average mineral composition parameters for: garnet on (a) Alm–Prp–Grs+And
747 and (b) Alm–Prp–Sps ternary diagrams, (c) biotite in terms of Ti (p.f.u = per formula
748 unit) vs. $x(\text{Fe}) = \text{Fe}/(\text{Fe}+\text{Mg})$ and (d) plagioclase on a Na–Ca–K ternary diagram.

749 **Fig. 5.** Bulk compositions plotted on an AFM diagram.

750 **Fig. 6.** P – T pseudosections showing the general topologies and relevant isomodes in
751 Field I (red box) for (a) 10-AI-20A, and (b) 10-AI-58-3B1 from the PLV-LM, (c) 331-
752 e2-11 and (d) 216a from the LBS, and (e) HJ-58c and (f) 10-AI-76 the GAG. Mineral
753 abbreviations: garnet (g); biotite (bi); sillimanite (sill); K-feldspar (ksp); plagioclase (pl);
754 quartz (q); silicate melt (liq); rutile (ru); ilmenite (ilm); cordierite (cd); orthopyroxene
755 (opx); muscovite (mu). Phases in the top right box are stable throughout.

756 **Fig. 7.** P – T pseudosections from Fig. 6 with a reduced range showing relevant isopleths
757 (as %) and inferred P – T paths. Phases in the top right box are stable throughout.

758 **Fig. 8.** Melt reintegrated P – T pseudosections for selected samples. Phases in the top right
759 box are stable throughout.

760 **Fig. 9.** P – T pseudosections for samples (a) 10-AI-58-2a and (b) HJ-57-C2, both of which
761 have interstitial cordierite at the thin section scale.

762

763

Figure 1

- Quartz
- Biotite
- Sillimanite
- Garnet
- Plagioclase
- K-feldspar
- Cordierite
- Spinel
- Corundum
- Sericite

Figure 4

Figure 5

Figure 26

Figure 7

Figure 8

Figure 9