Anatetic record and contrasting $P-T$ paths of aluminous gneisses from the central Grenville Province

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**ABSTRACT**

Anatetic aluminous gneisses, some derived from sedimentary rocks of broadly pelitic composition and others from hydrothermally altered felsic volcanic rocks, are exposed in the mid-$P$ and high-$P$ segments of the hinterland in the central Grenville Province. These gneisses consist dominantly of garnet, biotite, K-feldspar, plagioclase and quartz, with sillimanite or kyanite, and display microstructural evidence of anatexis by fluid-absent reactions consuming muscovite and/or biotite. Melt-related microstructures, such as inter-granular films and/or interstitial quartz or feldspars enclosing relict phases, are most abundant in the metasedimentary samples. Despite anatexis at granulite-facies conditions, the hydrothermally altered rocks preserve earlier features attributed to the circulation of hydrothermal fluids, such as sillimanite seams, dismembered quartz veins, and garnet-rich aluminous nodules in a K-feldspar dominated matrix. Microstructural and mineral chemical data, integrated with $P$–$T$ pseudosections calculated with THERMOCALC for the metasedimentary rocks permit qualitative constraints on the $P$–$T$ paths. Data from a high-$P$ kyanite-bearing sample are consistent with a steep prograde $P$–$T$ path up to $\sim$14.5 kbar and 860–900 °C, followed by decompression with minor cooling down to the solidus at $\sim$11 kbar and 870 °C. This pressure-dominated $P$–$T$ path is similar to those inferred in other parts of the high-$P$ segment in the central Grenville Province. In contrast, the $P$–$T$ path predicted from a mid-$P$ sillimanite-bearing paragneiss has a strong $T$ gradient with $P$–$T$ of $\sim$9.5 kbar and 850 °C at the thermal peak, and a retrograde portion down to $\sim$8 kbar and 820 °C. In a broad sense, these two contrasting $P$–$T$ patterns are consistent with predictions of thermo-mechanical modelling of large hot orogens in which $P$–$T$ paths with strong $P$ gradients exhume deeper rocks in the orogenic flanks, whereas $P$–$T$ paths with strong $T$ gradients in the orogenic core reflect protracted lateral transport of ductile crust beneath a plateau.
24 Keywords

25 anatectic Al-gneisses; granulites; Grenville Province; $P-T$ pseudosections; THERMOCALC
INTRODUCTION

Anatectic rocks commonly occur in the exposed internal zones (hinterland) of large hot orogens and their microstructures have the potential to preserve a record of high-temperature processes in orogenic environments at a range of scales, from localized partial melting to transport mechanisms in ductile crust (Brown, 2010). Identification criteria of melt related microstructures are sufficiently well established so that links can be made between the partial melting history and microstructural evolution of anatectic rocks (cf., Vernon & Collins, 1988; Harte et al., 1991; Sawyer, 1998; Holness et al., 2011). Although in tectonically active metamorphic environments such microstructures are prone to obliteration by deformation and recrystallization, they can be preserved if protected by rigid porphyroblasts such as garnet, or if the melt crystallized late. In the case of aluminous anatectic rocks, the availability of a felsic melt model (White et al., 2002; 2007), allows for integrated interpretations of microstructures and mineral compositions in terms of $P$–$T$ paths within the framework of $P$–$T$ pseudosections (isochemical $P$–$T$ phase diagrams). Studies of aluminous systems commonly refer to metasedimentary rocks such as metapelites and metagreywakes (Johnson et al., 2008; White et al., 2007). However, aluminous rocks may also be derived from hydrothermally altered felsic volcanic precursors (Bonnet & Corriveau, 2007), in which case the overall aluminous bulk chemistry and the general microstructure at the onset of metamorphism may vary significantly from that of their metasedimentary counterparts.

Often compared to the Himalaya-Tibet system, the Mesoproterozoic Grenville Province is inferred to represent remnants of the oldest large hot orogen on Earth (cf., in Rivers et al., 2012). The Grenvillian orogeny resulted in granulite-facies metamorphism and crustal anatexis in large portions of the exposed hinterland of the Province, which are further divided into high-$P$ and
mid-$P$ segments, or belts. These are locally juxtaposed with crustal segments that experienced lower grade Grenvillian age metamorphism (low-$P$ and orogenic lid; Rivers, 2008, 2012) and this overall configuration is attributed to the collapse of an orogenic plateau (Rivers, 2012).

Heterogeneous flow of middle to lower orogenic crust during high-grade metamorphism is inferred to have been instrumental in the overall tectonic evolution of the hinterland (Jamieson et al. 2007, 2010; Jamieson & Beaumont, 2011). Therefore, understanding the metamorphic record of the Grenville is critical for the assessment of the current tectonic models. However, rigorous investigation of the $P$–$T$ conditions and paths as well as of the role of partial melting on the microstructural evolution of the anatectic rocks in the granulite-facies belts remains scarce.

Among the exceptions, a recent study has highlighted the metamorphic evolution of a high-$P$ segment in the Manicouagan area of the central Grenville Province (Fig. 1; Indares et al., 2008), where the recorded $P$–$T$ paths are in the kyanite stability field (Cox & Indares, 1999a, b; Yang & Indares, 2005; Indares & Dunning, 2001). The high-$P$ segment in this area is tectonically overlain by mid-$P$ granulite-facies rocks (Dunning & Indares, 2010) including sillimanite-bearing anatectic aluminous gneisses of diverse origins, such as metasediments of broadly pelitic composition and hydrothermally altered felsic volcanic rocks (Indares & Moukhsil, 2013; Lasalle et al., 2013). These aluminous gneisses belong to lithologic associations of the Canyon domain (Hynes et al., 2000; Dunning & Indares, 2010), and kyanite-bearing equivalents of some of them are also reported from the southern tip of the high-$P$ segment. Exposure of both high-$P$ and mid-$P$ portions of the hinterland in the Manicouagan area (Fig. 1) provides a unique opportunity to gain insights on high-$T$ metamorphic processes at a range of crustal depths.

The aim of this contribution is to investigate the effect of anatexis on the microstructural evolution of aluminous gneisses from the Canyon domain of the Manicouagan area and interpret
this evolution in terms of $P$–$T$ paths within the framework of $P$–$T$ pseudosections. This study provides the first (to our knowledge) comprehensive assessment of the metamorphic record of mid-$P$ rocks in the Grenvillian hinterland and a comparison between the $P$–$T$ evolution of two juxtaposed high-$P$ and mid-$P$ segments. In addition, it discusses the results in the light of recent geodynamic models for the evolution of the Grenvillian orogen, and explores the effect of granulite-facies metamorphism and partial melting in a range of aluminous rock types.

REGIONAL GEOLOGY

The Manicouagan area straddles the boundary between the Parautochthonous belt and the structurally higher hinterland, in the central part of the Grenville Province (Fig. 1). In this area, the Parautochthonous belt is represented by Archean and Paleoproterozoic units of the Gagnon terrane (van Gool et al., 2008), which were metamorphosed under granulite-facies conditions during the waning stages of the Grenvillian orogeny (c. 990–980 Ma; Jordan et al., 2006). In contrast, the hinterland consists of lithotectonic packages of Mesoproterozoic rocks (Dunning & Indares, 2010) with diverse metamorphic signatures acquired during the culmination of the Grenvillian orogeny (c. 1080–1040 Ma), and which represent a stack of formerly deep (high-$P$), intermediate (mid-$P$) and relatively shallow Grenvillian crustal levels.

The structurally lowest component of the hinterland is the Manicouagan Imbricate zone (MIZ), exposed along the northern shores of the Manicouagan reservoir and consisting mainly of imbricated 1.65 Ga anorthosite-gabbro suite(s) to the east, 1.46 Ga augen gneiss, supracrustal rocks and 1.17 Ga gabbro sills to the west (Fig. 1; Cox et al., 1998; Indares et al., 2000). The MIZ was metamorphosed under high-$P$ granulite- to eclogite-facies conditions (1500–1800 MPa and 800–900 °C) at 1.05–1.02 Ga, followed by a retrograde $P$–$T$ path dominated by
decompression (Cox & Indares, 1999a, b; Indares & Dunning, 2001; Indares, 2003; Yang &
Indares, 2005; Indares et al., 2008).

In contrast, to the south, c. 1.69 Ga mafic rocks of the Island domain and 1.5–1.2 Ga units of
the Canyon domain were metamorphosed under mid-\(P\) granulite-facies conditions at 1080–1040
Ma (Dunning & Indares, 2010; see below). Finally, the Hart Jaune terrane to the east (Fig. 1) is
made of c. 1.5 Ga rocks (Gobeil et al., 1997a, b) that escaped any perceptible Grenvillian age
deformation and metamorphism (Indares & Dunning, 2004). The boundary between the high-\(P\)
and the mid-\(P\) portions of the hinterland is masked by the Triassic Manicouagan impact crater,
but its eastern extension juxtaposes the MIZ with the Hart Jaune terrane (Fig. 1).

The Canyon domain

The Canyon domain, first defined by Hynes et al. (2000) in the southern part of the Manicouagan
area, mainly consists of layered units, including a c. 1.5 Ga supracrustal sequence which is part of
the complex de la Plus Value (PLV; Moukhsil et al., 2012), a c. 1.4 Ga mafic suite, and a 1.24 Ga
bimodal felsic–mafic sequence (LBS) with hydrothermally altered components (Hindemith &
Indares, 2013; Indares & Moukhsil, 2013; Lasalle et al., 2013; Dunning & Indares, 2010). In the
Canyon domain, mineral assemblages of garnet, clinopyroxene, orthopyroxene, plagioclase and
hornblende in mafic rocks, and garnet, sillimanite, K-feldspar±biotite in aluminous rocks, are
reflective of mid-\(P\) granulite-facies metamorphism (Dunning & Indares, 2010). However, rocks
typical of the LBS are also recognized north of the Manicouagan impact crater, in the southern tip
of the high-\(P\) segment (Fig. 1), in which case, aluminous rocks have kyanite instead of
sillimanite. Monazite ages from aluminous gneisses of the Canyon domain constrained the
granulite-facies metamorphism at c. 1080–1040 Ma (Dunning & Indares, 2010).
Sample localities

In the Canyon domain, aluminous gneisses occur in the metasedimentary sequence of the PLV and in the bimodal LBS (Fig. 1). In the latter, they are found as discrete layers, some of which are inferred to be of metasedimentary origin on the basis of zircon data (Lasalle et al., 2013), and as diffuse layers, some of which have aluminous nodules, within felsic gneisses. On the basis of field relations, petrography and zircon data, the diffuse layers are inferred to be derived from hydrothermally altered felsic volcanic rocks (Indares & Moukhsil, 2013; Lasalle et al., 2013; Hindemith & Indares, 2013). These will be referred to as HAF (hydrothermally altered felsic) rocks. The aluminous rocks investigated in this study come from various localities (Fig. 1) and include a metasedimentary rock from the PLV (location HJ60) and three gneisses from the LBS. Two of the latter, from the south, are HAF rocks, one grading into a felsic gneiss (location 216) and one with aluminous nodules (location 333). The other from the north (location 244), is an aluminous layer, several cm thick, inferred to be of sedimentary origin (Lasalle et al., 2013) in sharp contact with massive felsic layers. Location 244 is the only one in the high-P segment. In the following sections the symbol # is used to replace the word “sample” in front of sample numbers.

PETROGRAPHY

Figure 2 highlights the macroscopic characteristics of representative rock samples from each location. Samples from HJ60 and 244 consist of light-coloured quartzofeldspathic layers alternating with darker layers rich in garnet, Al-silicate, and biotite (Fig. 2a,b). In samples from location 216 the gneissosity is defined by thin layers rich in aluminous minerals and
quartzofeldspathic lenses (Fig. 2c). In contrast, the sample from location 333 consists of garnet
nodules (with biotite and sillimanite) unevenly distributed in a matrix rich in K-feldspar (Fig. 2d).

Microstructures

This study focuses on five samples with representative mineral assemblages and microstructures. The distribution of the minerals at the thin-section scale was imaged by false color maps (Fig. 3) generated using a Mineral Liberation Analysis (MLA) software (developed by JK Tech, University of Queensland, Australia; Gu, 2003) linked to a FEI Quanta 400 environmental scanning electron microscope (SEM) at the Bruneau Center of Memorial University. The SEM was setup with an accelerating voltage of 25 kV, a beam current of 10 nA, a 2mm frame size (or horizontal field width), a dwell time of 10 ms, and a step size of 50 μm. The MLA software was used to generate a point-counted estimation of mineral proportions and a composition-sensitive false-color map of the mineral associations and textural relationships (Shaffer et al., 2008; Shaffer, 2009). Smaller-scale microstructures were documented by optical microscopy and cathode-luminescence (CL) imaging (Figs. 4 & 5). SEM–MLA and CL imaging were of particular use in establishing the overall distribution of quartzofeldspathic phases and their mutual relations.

The main mineral assemblage of garnet, Al-silicate, biotite, K-feldspar, plagioclase and quartz is common to all samples (Fig. 3) with the Al-silicate being kyanite in #244 and sillimanite in every other case. Minor phases include rutile, ilmenite (in #HJ60b, #333x), apatite (in #244 and #216a, c), graphite (in #HJ60b), sulphides (all samples but #244), monazite and zircon. Mineral modal proportions estimated by SEM–MLA are shown in Table 1.
The metasedimentary samples (#244, #HJ60b) mainly consist of large garnet porphyroblasts in a Al-silicate and biotite-bearing quartzofeldspathic matrix (Fig. 3a,b; Table 1). In contrast, the HAF samples (#216a, c; #333x) have markedly lower proportions of garnet+biotite+Al-silicate (Fig. 3c–e; Table 1). A common feature in these samples is the concentration of sillimanite in thin seams that are locally mantled by plagioclase (#216a) or overgrown by variably elongated garnet. Sample 333x is distinct in that it has a K-feldspar dominated matrix with variably elongated domains of pure quartz and garnet that commonly forms grain aggregates, but that in one instance defines a large discontinuous lens (Fig. 3e).

Garnet

Garnet porphyroblasts are ~1.5 mm to 7 mm in size and are smallest and most evenly distributed in #216c. Generally they are sub-rounded (cf., #HJ60b; Fig. 4a) to elongate along the foliation. The elongated type commonly overgrows Al-silicate and is mantled by plagioclase (cf., #244, #216a and #333x; Figs. 3b,c,e & 4b). Garnet porphyroblasts generally contain inclusions of lobate quartz (cf., #HJ60b and #216a; Fig. 3a,c), Al-silicate (Fig. 4b,c) and, in all samples but #333x, polymineralic inclusions and/or embayments that are especially abundant in #HJ60b.

These inclusions consist of corroded quartz and/or biotite and/or sillimanite within cuspatate pools or films of optically continuous quartz/feldspar, and delimited by straight to smoothened garnet faces (Fig. 4d–f). In contrast, garnet aggregates in #333x consist of small rounded grains that coalesce in patches and contain abundant ‘drop-like’ inclusions (Fig. 4g). In all samples garnet rims are locally replaced by biotite (Fig. 4a) and, in #HJ60b, by biotite–sillimanite clusters or feldspar pools enclosing small sillimanite prisms (Fig. 4h).
Al-silicates and biotite

Al-silicates occur both in the matrix and as small inclusions in garnet. They consist of kyanite in #244 and sillimanite in all other samples (with one exception, see below). In #244, matrix kyanite forms large blades with inclusions of apatite, quartz, and quartz rimmed by K-feldspar (Fig. 5a). These blades are corroded and locally mantled by plagioclase, from which they are variably separated by a thin film of quartz (Fig. 5b). In addition, in #216c, one small prism of kyanite, replaced by quartz or feldspar along cleavage planes, is included in garnet (Fig. 4c).

Sillimanite (in #HJ60b; #216a, c; #333x) is prismatic and generally associated with biotite. Matrix sillimanite in #216a, c, is mantled by plagioclase (Fig. 3d) from which it is locally separated by a thin rim of quartz, and contains scarce inclusions of corroded quartz rimmed by feldspar (Fig. 5c,d). Sillimanite in sites of garnet resorption occurs as smaller prisms, which have in one instance coalesced in a large patch separated from garnet by a pool of K-feldspar (Fig. 5e).

Biotite forms large single laths or clusters, commonly with Al-silicate and variably replacing garnet (cf., #HJ60b and #216c; Fig. 3e,c), and as smaller interstitial flakes. In the proximity of garnet, biotite rims form symplectites with quartz (cf., #HJ60b; Fig. 5f,g).

Quartzofeldspathic matrix

In the metasedimentary samples #244 and #HJ60b, quartz and feldspars show a relatively homogeneous distribution and grain size (Fig. 3a,b), except for a few larger grains. In contrast, the matrix in the HAF samples is heterogeneous (Fig. 3c–e). In #333x, quartz is restricted to large ribbons and the matrix mainly consists of coarse grained K-feldspar with plagioclase exsolution. In #216a and c, the matrix consists of alternating quartz±plagioclase-rich and K-feldspar-rich layers, with quartz commonly forming large, partly recrystallized ribbons while plagioclase is
mostly concentrated in mantles around elongate garnet and sillimanite. Sample 216c contains one large plagioclase (~5 mm) with corroded rims and lobate inclusions of quartz. In addition, smaller-scale microstructures are observed in the matrix of #HJ60b and #216a including pools of feldspar engulfing corroded biotite and partly rimmed by garnet (#HJ60b only; Fig. 5h), and feldspar or quartz as interstitial pools between quartzofeldspathic minerals (Fig. 5i) or as thin films at grain boundaries (Fig. 5j–l).

Mineral chemistry

The mineral chemistry of garnet, biotite and plagioclase were determined using a 4 spectrometer Cameca Camebax MBX electron probe at Carleton University, and a collection of well characterized natural and synthetic standards. Analytical conditions were 15 kV accelerating potential and 20 nA beam current. Raw X-ray data were converted to elemental weight percent using the PAP model. Ranges of chemical composition parameters for analyzed garnet, plagioclase and biotite are shown in Table 2.

Garnet

Overall, garnet is almandine-rich with subordinate pyrope, minor grossular and negligible spessartine ($X_{sp}$≤0.02) components. Figure 6 presents representative zoning profiles of the largest garnet grains in each sample. Generally, $X_{alm}$ and $X_{Fe}$ are lowest in the cores ($X_{alm}$ 0.50–0.52 in #244, #216a, c, and 0.63 in #HJ60b, #331x) and variably increase at the rims, mostly adjacent to biotite. The zoning pattern of $X_{prp}$ is antithetic to that of $X_{alm}$ and both are consistent with diffusional homogenization of Fe and Mg in garnet at high-$T$ conditions, followed by resetting of
the rim composition by retrograde Fe–Mg exchange with biotite (Spear, 1991; Pattison & Begin, 1994a, b).

$X_{grs}$ is lowest in garnet from the sillimanite-bearing samples (Fig. 3d; $X_{grs} = 0.02–0.05$; #331x; #HJ60; #216a,c), in which zoning patterns range from flat (garnet in #331x) to ‘bell shaped’, mostly in the small garnets of #216c (Fig. 6b). In the kyanite-bearing #244, $X_{grs}$ in garnet varies between 0.03 and 0.15 (Fig. 6a) and shows a complex $X_{grs}$ profile with intermediate values ($X_{grs}$ 0.09) in the core gradually increasing towards the rims ($X_{grs}$ 0.14) then dropping sharply to minimum values at the outer rim ($X_{grs}$ 0.03–0.04). $X_{grs}$ variations in garnet core and inner rim are likely relics of growth zoning due to lower diffusion rates of Ca, even at high $T$, relative to Fe and Mg (Florence & Spear, 1991). Profiles in $X_{sps}$ are flat, to weakly increasing at the rims (#216c; Fig. 6b), consistent with partial replacement of garnet by biotite (Spear, 1991).

Biotite

Biotite laths adjacent to garnet and/or isolated in the matrix, as well as inclusion in garnet (in #216c and #331x only) were analyzed. $X_{Fe}$ of biotite ranges from 0.30 to 0.45, Ti from 0.15 to 0.32 cations p.f.u. (based on 11 O), and Al$^{VI}$ from 0.15 to 0.30 cations p.f.u., the latter being lowest in #216a,c (Fig. 7; Table 2). The correlation between the microstructural setting of biotite and the compositional parameters cited above (see also Spear & Parrish, 1996) is highlighted in Fig. 7. In #216c, #HJ60b, and #244, there is a trend of increasing $X_{Fe}$ and Ti in biotite with increasing distance from garnet. In #216a, Ti follows this trend but $X_{Fe}$ is uniform. In #333x there is a wide range of $X_{Fe}$ without any clear trend, while Ti contents are relatively uniform. There is an inverse correlation between Al$^{VI}$ and Ti, with the Al$^{VI}$ varying most relative to Ti in #333x.
Plagioclase

$X_{An}$ of analyzed plagioclase generally ranges between 0.02 (rarely in rims) and 0.26 and the least calcic plagioclase ($X_{An} 0.03–0.10$) is found in #333x (Table 2). In most cases, plagioclase adjacent to garnet has the same composition as that isolated in the matrix. However in #216a, plagioclase mantling sillimanite and/or garnet has the highest $X_{An} (0.24–0.25)$. The large plagioclase grain in #216c has a uniform composition with $X_{An} 0.19$, within the range of $X_{An}$ of plagioclase in other microstructural domains of the same section.

Interpretation of microstructures

Peak assemblage and evidence for partial melting

The dominant mineral assemblage consists of garnet, Al-silicate, biotite, K-feldspar, plagioclase and quartz (with Al-silicate being kyanite in #244 and sillimanite in the rest) and in all samples but #333x there is also microstructural evidence for the presence of former melt. This evidence includes films and pools of optically continuous interstitial and/or xenomorphic feldspar and/or quartz enclosing corroded phases. Such microstructures, found both in the matrix and within garnet (Figs. 4d–f & 5i–l), are typically interpreted to represent melt pseudomorphs (Vernon & Collins, 1988; Harte et al., 1991; Sawyer, 1998; Holness & Sawyer, 2008; Holness et al., 2011, Johnson et al., 2013).

This inferred peak mineral assemblage in the presence of former melt is consistent with the idealized biotite-consuming fluid-absent melting reaction,

$$\text{biotite} + \text{Al-silicate} + \text{quartz} \pm \text{plagioclase} \rightarrow \text{garnet} + \text{K-feldspar} + \text{liquid}, \ (R1)$$
which occurs in aluminous rocks at $T$ above the suprasolidus muscovite breakdown via reactions such as,

$$\text{muscovite} + \text{quartz} \pm \text{plagioclase} \rightarrow \text{Al-silicate} + \text{K-feldspar} + \text{liquid}, \quad \text{(R2)}$$

(Spear et al., 1999).

Microstructures consistent with fluid-absent melting consuming muscovite and biotite

Evidence for melting involving muscovite in #216a, c and #244 is provided by the lobate inclusions of corroded quartz rimmed by feldspar within sillimanite and kyanite, respectively, (Fig. 5a,c) suggesting that these Al-silicates formed in the presence of melt.

The biotite-consuming melting reaction R1 produces garnet, K-feldspar and liquid at the expense of biotite, Al-silicate, and quartz ($\pm$plagioclase). Microstructural evidence consistent with this reaction includes: (a) the presence of elongated garnet overgrowing Al-silicate in all samples (Figs. 3b,c,e & 4b); (b) the corroded aspect of most kyanite prisms in #244 (Fig. 5a,b); (c) the presence of feldspar pools (inferred to represent former melt) engulfing corroded biotite, sillimanite and/or quartz, as polymineralic inclusions or embayments in garnet (#HJ60b; #216a; Fig. 4d–f) and implying that garnet grew in the presence of melt; and (d) quartz or feldspar interstitial pools and films in the matrix (Fig. 5i–l), that are most common in #HJ60b. The preservation of those pools and films in the matrix is very rare in regionally metamorphosed terranes, but this is not the first time it has been reported (cf., Hartel & Pattison, 1996; Sawyer 2001; Jordan et al., 2006; Guilmette et al., 2010).

Microstructures consistent with retrograde reactions during melt crystallization
In all samples (Figs. 4a & 5f, g) garnet rims replaced by biotite (± symplectitic with quartz), biotite+sillimanite and/or sillimanite+melt pseudomorphs, are consistent with crystallization of melt during cooling within the $P$–$T$ field of reaction R1. These microstructures are best developed in the metasedimentary sample #HJ60b (Figs. 4h & 5e), perhaps suggesting that the other samples contained less melt during the metamorphic peak. The overall preservation of the peak assemblage in all samples and the lack of retrograde muscovite (despite evidence that partial melting initially occurred via muscovite breakdown in #216a, c and #244) implies that a large portion of melt was lost (White & Powell, 2002), a feature that is common when anatexis occurs in a tectonically active environment (Brown, 2002, 2004).

The microstructures described above provide a record of the anatectic history of the investigated samples that can be summarized as: (a) fluid-absent muscovite melting in the stability field of kyanite in the case of #244 (from the high-$P$ segment; Fig. 1) and sillimanite in the case of the other samples; (b) fluid-absent melting consuming biotite, leading to garnet growth in the presence of melt at the expense of biotite and Al-silicate (all samples); (c) partial melt escape during prograde metamorphism; and (d) melt solidification at $T$ above the stability field of muscovite (all samples). We note that the presence of one resorbed kyanite prism in garnet in #216a may be a prograde relic of an earlier metamorphic event.

**Inherited microstructures**

Despite anatexis at granulite-facies conditions, earlier features are recognized in the HAF samples and most particularly in #333x. Key microstructural elements in this rock are: (a) the atypical mineralogy of the matrix, which is made almost exclusively by perthitic K-feldspar; (b)
the elongate large quartz which may represent relict phenocrysts or folded and attenuated veins; (c) the distribution of sillimanite into thin seams that are reminiscent of fluid pathways; and (d) the general shape of the discontinuous garnet aggregate (highlighted by the dotted line in Fig. 3d) which is suggestive of a precursor Al and Fe-Mg-rich nodule that has been cracked, and infiltrated by the matrix. These characteristics are consistent with Al and K enrichment by hydrothermal alteration of a felsic precursor (Bonnet & Corriveau, 2007; Hindemith & Indares, 2013). In addition, an original volcanic protolith (ryolite?) is supported by the zircon data published in Lasalle et al. (2013). A distribution of sillimanite in seams is also clear in #216a, c (Fig. 3c,d). In addition, both #216a and c are characterized by coarse and flattened quartz aggregates that attest to the former presence of large quartz crystals and/or dismembered quartz veins within a finer grained rock.

MODELLING OF PHASE EQUILIBRIA

In this section, the microstructural evolution of the investigated rocks is evaluated within the framework of $P-T$ pseudosections that are used in conjunction with mineral chemistry data to infer $P-T$ paths. The calculations were performed in the Na$_2$O–CaO–K$_2$O–FeO–MgO–Al$_2$O$_3$–SiO$_2$–H$_2$O–TiO$_2$–Fe$_2$O$_3$ (NCKFMASHTO) system using the software THERMOCALC version 3.33 (updated from Powell & Holland, 1988) and the internally consistent dataset (file tc-ds55.txt) of Holland & Powell (1998) updated in November 2003. The a-x mixing models used were: garnet; biotite and silicate melt (White et al., 2007); cordierite (Holland & Powell, 1998); orthopyroxene (White et al. 2002); muscovite (Coggon & Holland, 2002); plagioclase and K-feldspar (Holland & Powell, 2003); ilmenite (White et al. 2000).
PT pseudosections were calculated for #HJ60b, #244, #216c and #333x. The bulk composition of each sample was obtained by combining the mineral proportions estimated by SEM–MLA with the mineral chemistry. The bulk H2O was calculated based on the proportion of biotite using the H–Ti substitution scheme of White et al. (2007; see also Indares et al., 2008; Guilmette et al., 2011) and varies between 0.55 mol% (#333x) and 1.67 mol% (#HJ60b). The amount of oxygen (O), which is used to calculate the bulk Fe2O3 was set at a minimum: 0.01 mol% for #HJ60b and #333x, and 0.03 mol% for #244 and #216c to allow for the calculation otherwise impossible, of certain phase boundaries. Finally, in the case of the apatite-bearing #244 and #216c, the CaO of apatite was excluded from the bulk composition to prevent from overestimating the Ca contents of Ca-bearing NCKFMASHTO model system phases.

The PT pseudosections are shown in the PT range of 5–15kbar (and 5–18 kbar for the kyanite-bearing #244) and 700–950°C, and the stable phases were labeled using the following abbreviation scheme: garnet (g); biotite (bi); sillimanite (sill); kyanite (ky); rutile (ru); ilmenite (ilm); cordierite (cd); orthopyroxene (opx); muscovite (mu); quartz (q); silicate melt (liq). Isopleths of phase proportions were calculated and superimposed onto the pseudosections to illustrate the production/consumption of relevant phases (dotted lines; Figs. 8 & 10). In addition, the distribution of selected mineral composition isopleths \( z(g) = \frac{Ca}{(Ca+Fe+Mg)} \) and \( x(g) = \frac{Fe}{(Fe+Mg)} \) in garnet; \( x(bi) = \frac{Fe}{(Fe+Mg)} \) in biotite is shown in Figs. 9 and 10.

PT pseudosections using the estimated bulk compositions

A first set of PT pseudosections (Figs. 8 & 9) was calculated using the estimated bulk composition of the samples. As the microstructures suggest that some melt was removed during
the prograde \( P-T \) evolution of the rocks, these bulk compositions are residual. Therefore these \( P-T \) pseudosections are only valid for assessing the peak and retrograde evolution of the rocks.

**General topologies**

In Figs. 8a–d and 9a–d, quartz and K-feldspar are predicted in every field over the \( P-T \) range modelled. The topology of the observed mineral assemblage \( \text{sill/ky+g+bi+pl+ksp+q+liq} \) \((\pm \text{ru} \pm \text{ilm}) \) (M1) is highlighted in Fig. 8. The M1 field is limited by the solidus at low \( T \) and the prograde disappearance of biotite at high \( T \), both of which have a large positive \( \frac{dP}{dT} \). The predicted overall \( T \)-range of the peak field is between 820 and 890 °C; its width is largest for the bulk composition with the highest \( \text{H}_2\text{O} \) (#HJ60b; Fig. 8b) and the bi-out line is at the highest \( T \) in bulk compositions with relatively high \( X_{\text{Fe}} \) (#244 and #216c; Fig. 8c,d).

In terms of pressure, the M1-ky field, which represents the peak for the kyanite-bearing sample #244, is located between the sill/ky transition near 11kbar and the line marking the prograde appearance of muscovite near 18 kbar (Fig. 8a). In contrast, in the case of the sillimanite-bearing samples, the M1-sill field is bounded at low \( P \) by the appearance of cordierite near 7 kbar and the sill/ky transition, within the \( T \) range of interest (Fig. 8b–d). The M1-sill field is transected by the ru-in and ilm-out lines with increasing \( P \). However, these boundaries cannot be reliably used to further constrain the \( P \) of the peak assemblage because their locations are highly dependent upon the amount of oxygen used in the bulk rock composition, which is unconstrained.

Within the general M1-field (Fig. 8a–d) isopleths of phase proportions have a steep positive \( \frac{dP}{dT} \), and with increasing \( T \), garnet and melt proportions increase, while those of Al-silicate and biotite decrease. Most marked variations in phase proportions are predicted for the M1 field of
#HJ60b (Fig. 8b), which is the widest; in this case the garnet proportion increases from ~18 to 30% and the liquid from 0 to 10%. In contrast only ~4% of garnet (from 19 to 23%) and ~3% of liquid are produced across the narrow M1-field of #333x (Fig. 8d). At $T > T_{M1}$, isopleths of phase proportions have a moderate positive $dP/dT$ and are more widely spaced.

The distribution of $z(g)$, $x(g)$ and $x(bi)$ isopleths in the M1 field are shown in Fig. 9a–d. The $z(g)$ isopleths have a moderate to low $dP/dT$, and their spacing decreases in the kyanite-stability field. In contrast, the $x(g)$ and $x(bi)$ have steep $dP/dT$ with values decreasing with increasing $T$.

**P–T constraints on the metamorphic peak and the retrograde evolution**

The $P–T$ conditions of the metamorphic peak can be further constrained by comparing the modelled $z(g)$ and $x(g)$ isopleths of garnet with the measured values of these parameters. However, this approach should be used with caution. In terms of grossular, the analyzed garnets show relic growth zoning in the cores, variably smoothened by diffusion, and a steep decrease in some outer rims, which is consistent with retrograde zoning. Because the true grossular content of garnet at the metamorphic peak cannot be evaluated, a range of values was considered (excluding those of the outer rims). The analyzed garnets show homogeneous cores in terms of $x(g)$, implying diffusional homogenization at high $T$. However, this process may have continued during early stages of cooling, therefore $x(g)$ can only be used as a minimum peak-$T$ estimate.

The $P–T$ conditions where the last melt crystallized, and which might be expected to record effective closure of the samples to diffusion can be constrained by the intersection of the isopleth $z(g)_{rim}$ with the solidus, and the phase proportions predicted by THERMOCALC at this intersection should be similar to those observed in the thin-sections.
Garnet from #244 shows the most prominent and irregular growth zoning (Fig. 6a). The range of $X_{\text{grs}}$ for the inner rims is between 0.12 and 0.10 (with an exception of an anomalous peak at $X_{\text{grs}} = 0.14$) which in terms of the modelled $z(g)$ values defines a $P$-range for the metamorphic peak between ~14 and 16 kbar (Fig 9a). In terms of $T$, this field is bounded by the $x(g)$ of garnet cores, which falls close to the solidus, at ~870 °C, and the bi-out line at ~895 °C. The intersection of the $z(g)$ of the outer rims of garnet coincides with the kyanite–sillimanite transition line and intersects the solidus at ~11 kbar and 870 °C. This implies a decompression of at least 3 kbar and a maximum cooling of only 25 °C during melt crystallization.

The upper $P$-limit of the metamorphic peak for the sillimanite-bearing samples is set by the kyanite–sillimanite transition line. The tightest $P–T$ constraints are provided by #HJ60b. Garnet $X_{\text{grs}}$ profiles for this sample are the smoothest (except for the outer rims; Fig. 6d) defining a narrow range of $z(g)$ and the $x(g)$ of the garnet cores fall close to the bi-out line (Fig. 9b). The $P–T$ range defined by the $z(g)$, $x(g)$ and the bi-out line is ~10 kbar and 840–850 °C. The intersection of the $z(g)$ of the outer rim of garnet with the solidus is at 8.2 kbar and 820 °C implying a decompression of ~2 kbar and cooling of 20–30 °C during melt crystallization.

The range of $z(g)$ in garnet cores in #216c falls within the kyanite field down to the sill-ky transition (Fig. 9c), which is at odds with microstructural evidence for sillimanite being present during the prograde evolution. In this sample, the intersection of $z(g)$ rims with the solidus is at 9.2 kbar and 870 °C. Finally, the predicted $P–T$ conditions are lowest for #333x (Fig. 9d; 7.8 kbar and 840 °C) and close to those of melt solidification (6.8 kbar and 840 °C; in the vicinity of the cd-in line). However, these $P–T$ conditions are poorly constrained due to the low grossular content of this sample (~1%) and the wide spacing of $z(g)$ isopleths at that range.
Generally the proportions of phases predicted by THERMOCALC at the low-\(T\) end of the recorded \(P-T\) path (Fig. 8a–d) are similar to those observed (Table 1), except for the proportions of plagioclase and K-feldspar in #333x, the predicted amounts of which are largely lower and higher respectively. The matrix in this rock is mostly K-feldspar containing exsolution lamellae of plagioclase, which may have formed at \(T\) below the solidus.

Additional constrains on the retrograde \(P-T\) path may be placed by interpreting the observed retrograde microstructures in the framework of the calculated \(P-T\) pseudosections. For instance, the lack of Al-silicate overgrowing garnet rims in #244 is consistent with a steep \(P-T\) path, at low angle to the als isopleths (Fig. 8a). By contrast, partial or complete replacement of garnet by biotite+plagioclase±Al-silicate in #HJ60, for which the \(P-T\) pseudosections predict the widest M1 field, is consistent with a \(P-T\) path with moderate \(dP/dT\) cutting at high angle the bi, pl and sill isopleths (Fig. 8b). The absence of cordierite from the sillimanite-bearing samples is consistent with the intersection of the inferred \(P-T\) path with the solidus above the cd-in line (Fig. 8b–d). Finally, the isopleths reflecting the measured \(x(g)\) at the garnet rims and \(x(bi)\) of the biotite in matrix should reflect closure to diffusion of Fe and Mg between these two minerals at relatively low-\(T\). In most cases these isopleths either overlap with the solidus or are located within the subsolidus fields.

**Evidence for melt loss, and melt reintegration**

The overall good preservation of the peak assemblages, as well as the microstructural evidence for early melting involving muscovite, but for melt solidification outside the muscovite stability field, implies that some melt was lost during the prograde evolution. Indeed, the low-\(T\) subsolidus fields within the calculated \(P-T\) pseudosections are muscovite-free (except at high-\(P\)). Thus, in
order to account for the possibility of melt loss and suggest a trajectory for the prograde evolution of the rocks (cf., White & Powell, 2002), a second set of $P$–$T$ pseudosections was calculated after reintegrating melt to the residual bulk compositions (Fig. 10).

Generally, the melt loss history of anatectic rocks cannot be known. However, as shown by White & Powell (2002) and Powell et al. (2005) the main suprasolidus stability fields do not change significantly in different scenarios of melt loss. Therefore, by reintegrating a calculated melt proportion and composition into the measured rock composition (White et al., 2004) useful insights can be gained on the prograde evolution. Here we considered the simplest scenario of melt loss in a single step, at the intersection of the solidus for the residual composition with a presumed prograde $P$–$T$ path (Fig. 10). The composition of the melt stable at this $P$–$T$ point was calculated by THERMOCALC, and an amount of that melt, sufficient to just saturate the solidus in H$_2$O at a reference $P$ of 8 kbar, was added to the bulk composition (cf., details of the approach in Indares et al., 2008). This was done to maximize the amount of mica present before melting, and therefore the extent of the liquid-present fields. The amount of reintegrated melt ranges between 19.9 mol% in HJ60b and 26.5 mol% in 333x.

Melt-reintegrated $P$–$T$ pseudosections

General topologies

The melt-reintegrated pseudosections are shown in Fig. 10a–d, along with isopleths of g, sill/ky, liq and z(g), and with the position of the solidus as modelled with the residual bulk compositions (from Fig. 8). For reference, this residual solidus divides the $P$–$T$ pseudosections in two parts. At higher $T$ within the kyanite field, the topologies of the two sets of pseudosections are almost
identical (compare Figs. 8 & 10), whereas within the sillimanite field, the topologies with the
lowest variance fields in the melt-reintegrated pseudosections are truncated either by the pl-out
line (in #HJ60b and #333x; Fig. 10b,d) or the sill-out line (#244; Fig. 10a). At lower T, across the
full P-range, topologies of the two sets of pseudosections are different.

In all the melt-reintegrated P–T pseudosections, the M1 field(s) extend over a much larger T-
interval and are terminated down-T by the disappearance of garnet between about 770 and 810°C
(Fig. 10a–d). In addition, muscovite is stable over a larger P-range. However, there are
significant differences between the topologies in #244, #216c, #HJ60b and those in #333x. In the
former case the mu-out line almost coincides with the ksp-in line and melt is produced mainly at
T above the mu-out–ksp-in lines at P below 12 kbar (Fig. 10a–c). In contrast, in the case of
#333x, K-feldspar is present throughout, the als-in line overlaps with the mu-out line across the
whole P–T range and garnet is absent at low T at any P (Fig. 10d). In this case, the g-free mu-
bearing field shows the lowest melt production with increasing T.

Constraints on the prograde P–T evolution

Constraints on the prograde P–T evolution may be placed by interpreting the observed prograde
microstructures and the measured zoning in the grossular content of garnet cores within the
framework of the extended suprasolidus topologies. Garnet from #244 contains kyanite
inclusions and shows a systematic increase in grossular towards a mantle zone, from ~0.08 to
0.12–0.15 (Fig. 6a). In terms of the modelled z(g) values this translates into increasing P–T
conditions from the g-in line at ~10.5 kbar and 775 °C to up to 15–16 kbar and 850–860 °C (for
z(g)=0.12 and 0.15, respectively) at the residual solidus (Fig. 10a). However, maximum grossular
contents in the garnet rims are mostly around 0.12, with the peak value of 0.15 being exceptional and probably representing a site of apatite dissolution in the melt (cf., Indares et al., 2008).

In all other samples garnet overgrows sillimanite (except for one kyanite inclusion in #216a). For these samples the melt-reintegrated $P$–$T$ pseudosections predict that garnet started growing at $T$ conditions above the mu-out line, and that Al-silicate is not produced with increasing $P$–$T$ conditions above the g-in line (Fig. 10b–d). It is therefore implied that all prograde Al-silicate predates garnet. The prograde $P$–$T$ path most consistent with both the petrography and the melt reintegrated pseudosections is restricted at $P$ below the sill-ky transition and has a strong $T$ gradient. Such an evolution is also consistent with the measured and modelled grossular contents in garnet cores in #HJ60b (Fig. 10b), but not for those of #216c, in which the $z(g)$ of the garnet cores fall in the kyanite field (Fig. 10c).

Sample #216c contains relict apatite and shows garnet distributed as numerous small grains in which Ca contents decrease markedly from core to rim (more than in larger garnets from other samples; e.g., #216a, #HJ60c; Fig. 6). A potential interpretation is that nucleation of the small Ca-zoned garnet in #216c was favored in sites of apatite dissolution in the melt, within micro-domains enriched in Ca relative to the average bulk (see also Indares et al., 2008). Since apatite cannot be modelled in THERMOCALC, it is not possible to interpret this high-Ca garnet in the context of $P$–$T$ pseudosections. Finally, in #333x the grossular content of garnet is too low and homogeneous to place any additional constrain on the prograde $P$–$T$ path.

**DISCUSSION AND CONCLUSIONS**
Anatetic record of aluminous rocks from the Grenvillian hinterland in the Manicouagan area

Aluminous gneisses of diverse origins, pelitic metasediments and metamorphosed hydrothermally altered felsic (HAF) volcanic rocks, from the mid-\(P\) segment of the hinterland in the central Grenville Province show garnet–sillimanite–biotite–quartz–K-feldspar–plagioclase assemblages and microstructural evidence of anatexis by fluid-absent melting reactions involving first muscovite, and then biotite, with increasing \(T\). The same assemblage (with kyanite instead of sillimanite) and microstructural evidence of melting is also observed in aluminous metasedimentary layers on the southern boundary of the high-\(P\) segment, to the north.

Evidence of melting involving muscovite is best shown in HAF rocks of a broadly pelitic composition (#216a, c), whereas the most abundant evidence of former melt linked with biotite breakdown comes from the metasedimentary rocks (#HJ60b, #244). The only rock type that lacks evidence of former melt is a HAF quartz-poor and K-feldspar rich nodular gneiss (#333x).

However, in all rocks, the presence of garnet overgrowing Al-silicate is consistent with the fluid absent melting reaction (R2). The peak assemblage is best preserved in HAF rocks, whereas the mid-\(P\) metasedimentary rock #HJ60b, which shows the most ample evidence of former melt, displays the most pronounced retrograde microstructures associated with melt crystallization.

\(P–T\) record

Microstructures and mineral chemistry data from the metasedimentary rocks (#244 and #HJ60b), were most suitable for \(P–T\) path predictions by phase equilibria modelling. Two types of \(P–T\) paths are predicted, depending on the Al-silicate present (Fig. 11). The data from the kyanite-bearing rock #244 are consistent with a steep prograde \(P–T\) path, reaching the metamorphic peak
at ~14.5 kbar and 860–900 °C, followed by decompression with minor cooling down to the solidus at ~11 kbar and 870 °C, just above the sillimanite–kyanite transition line (Fig. 11, paths A-A’). In contrast, the $P-T$ path predicted from the sillimanite-bearing rock #HJ60b (Fig. 11, path B) is characterized by a prograde portion with a strong $T$ gradient below the sillimanite–kyanite transition line, up to ~9.5 kbar and 850 °C, and a retrograde portion down to the solidus at ~8 kbar and 820 °C, involving moderate decompression and cooling. Although $P-T$ paths were not established for the HAF rocks, and it is possible that different locations in the mid-$P$ portion of the Canyon domain may have followed distinct $P-T$ trajectories, microstructures constrain the $P-T$ paths of all the sillimanite-bearing rocks below the sillimanite–kyanite transition and above the cordierite-in line. The two predicted $P-T$ paths of the respective samples end at the solidus (point at which the microstructural evolution of the rocks ended) and they do not show any overlap, although it may be possible that the high-$P$ path A-A’ merges with the retrograde portion of path B at lower $P$.

Regional patterns of metamorphism

The $P-T$ path predicted for the kyanite-bearing rock #244 is broadly similar to those inferred in other parts of the high-$P$ belt in the Manicouagan area (cf., Indares et al., 2008). These steep $P-T$ loops have been attributed to rapid exhumation of the MIZ over a crustal-scale ramp on the underlying Parautochthonous belt (Indares et al., 2000; Dunning & Indares, 2004) shortly after the peak of metamorphism, which was dated at c. 1040–1030 Ma by Indares & Dunning (2001).

In contrast, the $P-T$ path predicted for the sillimanite-bearing aluminous paragneiss #HJ60b is the first example ever inferred for the mid-$P$ portion of the hinterland of the central Grenville Province, and maybe the first one determined using modern methodology in the Grenvillian
hinterland in general. This path is characterized by larger $T$ and lesser $P$ variations relative to those of the high-$P$ rocks. In addition, monazite from individual sillimanite-bearing aluminous samples shows a larger spread in Grenvillian metamorphic ages (e.g., c. 1080–1050 Ma in #HJ60b; Dunning & Indares, 2010), consistent with a longer residence under high-$T$ conditions, and a tectonic transport dominated by lateral crustal flow during this time interval.

**Comparison with predictions of tectonic models**

In a broad sense, these two contrasting types of $P$–$T$ paths discussed above can be compared with those predicted by recent numerical modelling of the Grenville orogeny (e.g., Jamieson et al., 2010; Jamieson & Beaumont, 2011). These models involve a laterally heterogeneous crust that develops ductile fold nappes during high-$T$ metamorphism and have been most successful in explaining first order geological features in the western part of the Province, while in the case of the Manicouagan area, they were in part limited by the scarcity of geological constraints from the mid-$P$ hinterland at the time (Jamieson et al., 2010). However, despite the uncertainties involved (cf., Jamieson et al., 2010; Jamieson & Beaumont, 2011) and the fact that a given numerically modelled $P$–$T$ path can be reproduced by a range of combinations of model parameters, there are some general trends that emerge. In regions near the orogenic flanks $P$–$T$ paths are tight loops involving strong gradients in pressure whereas in the orogenic core they show large near isobaric heating segments reflecting protracted lateral transport of ductile crust beneath a plateau. These two contrasting patterns are well illustrated by our data from the Manicouagan area, which show steep $P$–$T$ paths exhuming deeper rocks in the edge of the hinterland (MIZ) versus moderate-gradient $P$–$T$ paths followed by intermediate-depth crustal levels farther in the orogenic core, within the Canyon domain.
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Figure captions

Fig. 1. Simplified geological map of the Manicouagan area with location of the 4 samples of interest and inset map showing the general framework of the Grenville Province (updated from Dunning & Indares, 2010).

Fig. 2. Photographs of the rock samples: (a) HJ60b: metasedimentary gneiss from the PLV, (b) 244: metasedimentary gneiss from the LBS, (c) 216 and (d) 333x: aluminous gneisses derived from hydrothermally altered felsic rocks of the LBS.

Fig. 3. False-color maps of thin-sections generated by SEM–MLA: (a) #HJ60b, (b) #244, (c) #216a, (d) #216c, (e) #333x.

Fig. 4. Photomicrographs of garnet microstructures viewed in: PPL: plane-polarized light; XPL: crossed-polarized light; XPLq: crossed-polarized light with accessory quartz plate; CL: cathodoluminescence. (a) Large garnet porphyroblast with rims corroded by biotite±sillimanite (#HJ60b; PPL). (b) Elongate garnet with inclusions and tails of sillimanite needles/prisms, mantled by plagioclase (#216a; PPL and XPL). (c) Garnet porphyroblast with rounded inclusions of quartz and one larger inclusion of kyanite; the later shows is resorbed, and partly replaced by quartz or feldspar along cleavages (#216a; XPLq). (d) Embayment in garnet filled by a large squarish quartz grain thinly rimmed by K-feldspar (#HJ60b; CL). (e) Embayment in garnet filled with plagioclase (central part with a faint outline) and resorbed quartz, rimmed by a thin film of feldspar (#244; CL). (f) Embayment in garnet filled by resorbed biotite, sillimanite and quartz in a pool of K-feldspar; biotite rims are locally overgrown by biotite+ quartz symplectite and a
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**Fig. 5.** Photomicrographs of other microstructures viewed in; PPL: plane-polarized light; XPL: crossed-polarized light; XPL: crossed-polarized light with accessory quartz plate; CL: cathodoluminescence. (a) Kyanite with a composite inclusion of rounded quartz partly rimmed by K-feldspar, (#244; CL). (b) Resorbed kyanite, thinly coated by quartz, between two garnet grains (#244; CL). (c) Composite inclusion of resorbed quartz rimmed by feldspar, in sillimanite (#216; CL). (d) Thin coating of quartz around sillimanite seams mantled by plagioclase (#HJ60b; CL). (e) Largely coalesced aggregate of sillimanite needles in a pool of feldspar corroding garnet; (#HJ60b; XPL); the location of this microtexture is shown as a box with a dotted outline in Fig. 4a. (f) Biotite lath rimmed by biotite+quartz symplectite against a corroded garnet rim (#HJ60b; PL). (g) Biotite+quartz symplectite rimming a biotite+sillimanite cluster (#HJ60b; PL). (h) Feldspar pool with resorbed biotite and garnet in the matrix of #HJ60b (XPLq). (i) Interstitial quartz pool in the matrix of #216c; XPLq). (j) Thin film of quartz around sillimanite prisms at the boundary of a coarse grain of K-feldspar (#HJ60b; XPLq). (k) “String of beads” made of quartz between two K-feldspar grains (#216c; XPL). (l) Finger of K-feldspar between two plagioclase grains (#216c; XPL).

**Fig. 6.** Zoning profiles of the largest garnet porphyroblasts analyzed in each sample (except for #333x in which garnet is unzoned).
**Fig. 7.** Biotite composition plots: Ti vs. $X_{Fe}$ and Ti vs. Al$^{IV}$.

**Fig. 8.** $P$–$T$ pseudosections calculated with the estimated bulk rock composition specific to each sample #244, #HJ60b, #216c, #333x, showing the general topologies and the distribution of relevant phase proportion isopleths; the highlighted topology (field M), corresponds to the mineral assemblage observed in the rocks. Mineral abbreviations: garnet (g); biotite (bi); sillimanite (sill); kyanite (ky); rutile (ru); ilmenite (ilm); cordierite (cd); orthopyroxene (opx); muscovite (mu); quartz (q); silicate melt (liq).

**Fig. 9.** Same set of $P$–$T$ pseudosections as in Fig. 8, showing the distribution of relevant mineral composition isopleths and inferred $P$–$T$ paths. For mineral abbreviations see caption of Fig. 8.

**Fig. 10.** Melt-reintegrated $P$–$T$ pseudosections showing the distribution of relevant isopleths, and the inferred overall $P$–$T$ evolution (for #244 and #60b only; see explanation in text). The solidus calculated with the measured bulk compositions (from Figs. 8 & 9) is also shown for reference. Blank fields were not calculated. Notes on the melt reintegration: (i) In #244, a proportion of 22% of liquid was added at the intersection of $z(g) = 11\%$ with the solidus (at $P=14.55$ kbar – $T=859.9^\circ C$); (ii) In HJ60b, a proportion of 20% of liquid was added at the intersection of $z(g) = 4\%$ with the solidus (at $P=8.83$ kbar – $T=808.5^\circ C$); (iii) In 216c, a proportion of 24% of liquid was added at the intersection of $z(g) = 4\%$ with the solidus (at $P=12.81$ kbar – $T=868.5^\circ C$); (iv) In 333x, a proportion of 26% of liquid was added at the intersection of $z(g) = 2\%$ with the solidus (at $P=10.20$ kbar – $T=832.2^\circ C$). For mineral abbreviations see caption of Fig. 8.
Fig. 11. Schematic $P-T$ paths from the high and mid-$P$ portions of the hinterland in the Manicouagan area.
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<th>bi&lt;sup&gt;a&lt;/sup&gt;</th>
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<th>ksp</th>
<th>q</th>
<th>ru</th>
<th>ilm</th>
<th>ap</th>
<th>zr</th>
<th>mnz</th>
<th>gr</th>
<th>sulph</th>
<th>ser&lt;sup&gt;b&lt;/sup&gt;</th>
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<td>14.3</td>
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<td>0.03</td>
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<td>19.3</td>
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<sup>a</sup> including chlorite present as replacement phase after biotite

<sup>b</sup> sericite present as replacement after feldspar
Table 2: Mineral chemistry

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<tr>
<th></th>
<th>Garnet</th>
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<th>Plagioclase</th>
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<td>Xprp</td>
<td>Xgrs</td>
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<td>0.24–0.33</td>
<td>0.02–0.04</td>
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<tr>
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<td>0.34–0.41</td>
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<td>0.03–0.04</td>
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<td>0.22–0.35</td>
<td>0.01</td>
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</table>

*Ti, Al⁶⁺: cations on the basis of 11 oxygens

a extreme values recorded by the rim of a plagioclase grain in the matrix (in #216c)

b extreme values recorded by the rim of a biotite grain in the matrix (in #216c)