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Conditions and Timing of LP-HT Metamorphism in the Montresor Belt, Rae Province, Nunavut

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ABSTRACT

P-T-t estimates for the Montresor Belt, obtained using phase equilibria and geospeedometry modelling integrated with in-situ U-Th-Pb monazite geochronology, shed new light on the tectono-metamorphic effects of the Snowbird phase of the Trans-Hudson orogeny. Typical metapelitic assemblages of the lower Montresor group consist of white mica, biotite, plagioclase, quartz, and andalusite, which in some rocks is partly or completely pseudomorphed by white mica. The observed assemblages reflect peak *P-T* conditions centring at ca. 575 °C and 3 kbar. Rocks with high bulk Fe/Mg contents contain compositionally zoned garnet, permitting the addition of further constraints on the conditions of metamorphism in the Montresor Belt: Core compositions of earliest-grown garnets indicate initial garnet crystallization at ca. 535 °C and 2.3 kbar suggesting a nearly isobaric *P-T* path of prograde metamorphism with a gradient of ca. 50 °C kbar⁻¹. Chemical age-dating of monazite inclusions in garnet yields ages of ca. 1870 ± 9 to 1837 ± 9 Ma. Retrograde, pseudomorphic andalusite replacement by white mica at ca. 540 °C is inferred to have been controlled by variations in bulk rock chemistry. Morphologically corroded and chemically hetero-

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geneous monazite adjacent to white mica pseudomorphs suggest that andalusite replacement took place at ca. 1792 ± 10 Ma, possibly associated with extension and movement along the detachment fault separating the upper and lower Montresor groups. Simulations of diffusion across chlorite- and biotite-filled cracks in garnet assumed to be coeval with andalusite replacement suggest that the rocks have experienced the retrograde event for at least 20 My.

Key words: diffusion, geospeedometry, phase equilibria, pseudomorphism

1 Introduction

The Trans-Hudson orogeny was a continent-scale, Paleoproterozoic event characterized by widespread metamorphism, deformation, and plutonism at ca. 1.91 - 1.75 Ga, part of which is preserved in the Canadian Shield (Corrigan et al., 2009). An understanding of the associated mineralogical, chemical, and textural changes of rocks involved in the Trans-Hudson orogeny is required to infer the pressure and temperature (P - T) conditions and timing of their formation (e.g., Borghi et al., 2002; Brown, 2014; Tirone and Ganguly, 2010; Tinkham and Ghent, 2005), which in turn are necessary to elucidate the tectono-metamorphic evolution of the crust during the Paleoproterozoic.

Garnet is one of the most commonly utilized minerals for the study of metamorphic P - T conditions. Growth zoning of garnet has been extensively studied and shown to be sensitive to the P - T conditions of metamorphic events (e.g., Borghi et al., 2002; Evans, 2004; Gaidies et al., 2006, 2008, 2015; Tirone and Ganguly, 2010; George and Gaidies, 2017). The usefulness

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14 of garnet is due in part to its thermodynamic stability over a wide range of P - T conditions
15 in a variety of chemical systems and its well-described thermodynamic properties (e.g.,
16 Berman, 1988; Holland and Powell, 1998). Growth zoning of major elements in garnet is
17 commonly retained during greenschist to amphibolite-facies metamorphism due to relatively
18 slow intracrystalline diffusion (e.g., Florence and Spear, 1991; Gaidies et al., 2008; Caddick
19 et al., 2010). Given the T and time (t) dependence of diffusion for different ionic species,
20 modification to growth zoning due to intracrystalline diffusion may be used to estimate the
21 duration of metamorphic events.

22 Determining the absolute timing of metamorphism is critical when reconstructing tectonic
23 processes (Brown, 2014; Foster et al., 2002). The combination of P - T and time information
24 allows for the determination of the duration and style of metamorphic events, increasing the
25 understanding of the tectonic and geological evolution of an area (Brown, 2014; Tirone and
26 Ganguly, 2010). In this study, metamorphic monazite ages are combined with metamorphic
27 P - T estimates obtained from phase equilibria modelling to determine the absolute P - T - t
28 conditions experienced by rocks from the Montresor Belt during the Trans-Hudson orogeny.
29 Relative time constraints are obtained from geospeedometry calculations for cation diffusion
30 across cracks in garnet filled with biotite and chlorite during retrograde metamorphism.

31 2 Regional Geology

32 The Canadian Shield forms the nucleus of North America, spanning the region from Baf-
33 fin Island to Hudson Bay and into the northern continental United States (Fig. 1). It is

comprised of several Archean cratons that are separated by Proterozoic orogenic belts. Proterozoic amalgamation of these cratons into the Canadian Shield caused regional deformation and metamorphism, including variable reworking of the bordering Archean cratons, as well as magmatism, and sedimentation (Hoffman, 1989). This resulted in a complex metamorphic history and some of the most extensively reworked continental crust on Earth (Berman et al., 2010).

The Rae cover sequence is a widespread but discontinuously preserved package of metasedimentary rocks (Rainbird et al., 2010), interpreted to unconformably overlie Archean basement rocks (ca. 2.6 Ga) of the Rae Craton (Davis et al., 2014; Percival et al., 2015b, 2017). It is exposed in isolated belts known as the Amer, Ketyet River, Montresor, Chantrey, and Penrhyn groups (Berman et al., 2015; Davis et al., 2014; Patterson, 1986), which have similar stratigraphic attributes (Rainbird et al., 2010). Detrital zircon geochronology constrains deposition of the Rae cover sequences to 2.19 - 1.92 Ga (Percival et al., 2015b, 2017).

2.1 The Montresor Belt

The Montresor Belt is located approximately 250 km north of the community of Baker Lake, Nunavut. It is one of the smaller northern outliers of preserved Rae cover sequence (Rainbird et al., 2010). The Montresor Belt was first discovered in 1961 during Operation Northern Keewatin, a reconnaissance program conducted by the Geological Survey of Canada (Heywood, 1961) and was first studied in detail by Frisch and Patterson (1983).

The belt is 73 km long and 8 km wide, extending from the Montresor River in the northeast

54 to the Back River in the southwest (Frisch, 2000). Structurally it forms an open synclinal
55 fold with the southeast limb dipping at 30° and the northwest limb dipping $40 - 70^{\circ}$, with
56 localized dip up to 80° . The base of the syncline is prominently visible as a magnetic low
57 in modern aeromagnetic maps (Tschirhart et al., 2015). The stratigraphically lower part of
58 the Montresor Belt is dominantly composed of meta-arenite and metacarbonate, with minor
59 metapelitic schist (Frisch and Patterson, 1983; Percival et al., 2015b). Archean basement
60 rocks to the Montresor Belt are dominantly of a tonalite-granodiorite-granite composition
61 (Frisch, 2000).

62 The upper part of the Montresor Belt is primarily comprised of low-grade, weakly meta-
63 morphosed siltstone and dolostone of the upper Montresor group (Fig. 2). These overlie
64 higher-grade amphibolite facies meta-arenites, calc-silicates, and metapelites of the lower
65 Montresor group that are discontinuously exposed on the edges of the syncline. The lower
66 Montresor group is commonly cut by granodiorite, granite, and pegmatitic dykes (undated;
67 Percival et al., 2015b). Previous studies suggested that these higher-grade amphibolite fa-
68 cies rocks belong to the basement (Frisch and Patterson, 1983; Frisch, 2000), but recent field
69 and geochronological work has demonstrated that these units are part of the Montresor Belt
70 (Percival et al., 2015b). These rock sequences can be regionally correlated with rocks seen
71 in other Rae cover sequence groups, such as the Amer Belt located 50 km to the south.

72 A mylonitic ductile fault zone separates the lower Montresor group from the Archean base-
73 ment. Observed imbrication of the Archean basement and lower Montresor group suggests
74 the presence of thrust faults at the base of the syncline. It has been suggested that this
75 faulting may relate to folding and thrusting described in the Amer Belt (Patterson, 1986;

Rainbird et al., 2010; Percival and Tschirhart, 2017). The upper and lower Montresor groups are separated by a low angle detachment fault. A difference in metamorphic grade between the greenschist facies upper Montresor group and amphibolite facies lower Montresor group suggests the detachment fault is late- to post-metamorphic with a vertical offset of several km. Later folding created the synclinal structure observed today.

Two major orogenic events occurred within the northwestern Rae Province after the deposition of the Rae cover sequence. These are referred to as the Thelon-Taltson orogeny (ca. 2.0 - 1.92 Ga) and the Trans-Hudson orogeny (ca. 1.91 - 1.75 Ga). However, a lack of ca. 2.0 Ga accessory mineral growth suggests that the extent of the Thelon-Taltson orogeny did not reach the Montresor Belt (Berman, 2010). This, along with a prominence of ca. 1.85 Ga metamorphic monazite ages (Davis et al., 2014), suggests that the Montresor Belt was regionally metamorphosed during the Trans-Hudson orogeny (Percival et al., 2015b). It has also been suggested that metamorphism may have been associated with localized pegmatitic intrusions (Frisch, 2000).

2.2 The Trans-Hudson orogeny

The Trans-Hudson orogen is the largest Proterozoic orogenic belt preserved in the Canadian Shield (Corrigan et al., 2009). The Trans-Hudson orogeny resulted from the assembly of the Rae, Hearne, Wyoming and Superior cratons to the Superior Province starting at ca. 1.91 Ga (Hoffman, 1988), during the closure of the Manikewan Ocean and subsequent formation of the supercontinent Nuna at ca. 2.0 - 1.8 Ga (Corrigan et al., 2009; Pehrsson et al., 2013).

96 The Trans-Hudson orogeny occurred in two distinct phases, referred to as the Snowbird
97 phase and the Main phase (Berman et al., 2007; Berman, 2010).

98 During the Snowbird phase at ca. 1.91 - 1.87 Ga, the Rae craton collided with the Hearne
99 craton (ca. 1.91 Ma) along the Snowbird Tectonic Zone, ca. 300 km south of the study
100 area (Fig. 1), and with the Meta-incognita microcontinent (ca. 1.87 Ma), ca. 1000 km to
101 the east (Berman et al., 2007; Berman, 2010; St-Onge et al., 2006). The Main phase of the
102 Trans-Hudson orogeny occurred between ca. 1.86 and 1.75 Ga (Berman, 2010), with the
103 terminal collision of the Superior craton to the southeast of Rae-Hearne block occurring at
104 ca. 1.82 Ga (Berman, 2010; Corrigan et al., 2009). The Trans-Hudson orogeny is spatially
105 and temporally associated with two distinct periods of syn- to post-tectonic magmatism,
106 the ca. 1845 - 1795 Ma Hudson suite and the ca. 1765 - 1750 Ma Nuelin suite (Davis et al.,
107 2014; Peterson et al., 2002; van Breemen et al., 2005). In the northwestern Rae Province,
108 the Trans-Hudson orogeny caused low- P metamorphism at amphibolite to granulite facies,
109 with an estimated peak P of 5 kbar (Berman, 2010).

110 2.3 Sample locations

111 The positions within the Montresor Belt of the samples investigated in this contribution are
112 given in Fig. 2 and Table 1. In a previous study, 12PBA-27A has been interpreted to have
113 experienced amphibolite facies conditions of 3.3 kbar and 575°C (Berman et al., 2014). Two
114 in-situ U-Th-Pb monazite analyses yielded metamorphic ages of 1861 ± 7 and 1844 ± 6 Ma
115 (Berman et al., 2015; Davis et al., 2014).

3 Methods

3.1 Whole-rock chemical composition

The bulk chemical compositions of the rocks studied were analyzed at Actlabs in Ancaster, ON (Canada). Samples were prepared from sections of rock that were macroscopically viewed to be chemically homogeneous and representative of the rock. Major element concentrations of 12PBA-27A were determined on a fused disc using a Panalytical Axios wavelength dispersive XRF spectrometer. All the remaining samples were analyzed using ICP-MS on a Perkin Sciex ELAN. Major element chemistries for the studied samples are provided in Table 2.

3.2 Electron probe micro-analysis (EPMA)

Quantitative EPMA traverses as well as qualitative elemental distribution maps of Fe, Mg, Ca, Mn, and Y on the largest garnet crystal of 12PBA-27A were conducted with a JEOL-JXA 8230 at University of Ottawa (Canada). The operating conditions for the mapping used an accelerating voltage of 20 kV and a beam current of 100 nA. A WDS detector was used with a dwell time of 200 ms on each analysis, and maps were produced with a pixel size of $7 \mu\text{m}^2$. An accelerating voltage of 20 kV and a beam current of 20 nA were used for the spot analyses with a counting time of 20 s for the major elements. All other EPM analyses on garnet and other phases were completed using a Cameca Camebax MBX electron microprobe at Carleton University (Ottawa, Canada). Operating conditions were an accelerating voltage of 20 kV and a beam current of 20 nA. Peak counting times ranging

from 15 - 45 s on each analysis were chosen to obtain around 40000 counts on each element. Background measurements on each side of the analyzed peak were done at 50 % of the total peak counting time. A Cameca PAP matrix correction procedure was used to convert counts into elemental wt% (see Pouchou and Pichoir, 1985). Several well-defined natural and synthetic standards were used for internal calibration of the microprobes at University of Ottawa and Carleton University.

3.3 *X-ray micro-computed tomography (XR- μ CT)*

XR- μ CT was done using a Bruker Skyscan 1173 tomography scanner at Carleton University. A 2.5 cm diameter by 2.8 cm long cylindrical core of 12PBA-27A was drilled for scanning. The sample was rotated 180° within the X-ray field with a step size of 0.4°. At each rotation step an image was taken with an exposure time of 1.35 s, and an image pixel size of 14.9 μ m. The scan used an operation setting of 130 kV and 61 μ A for the X-ray source. X-rays were filtered through a 0.25 mm brass plate before hitting the sample to pre-harden the beam and to reduce beam-hardening effects. Flat field corrections were performed every 60 min throughout each scan. Raw image data were converted into cross-sections using Skyscan NRecon software. Quantification of the distribution and size of garnet in the sample was done with BLOB 3D (Ketcham, 2005) and Avizo software. The XR- μ CT data was used to locate representative garnet porphyroblasts of the 5 largest size classes in the scanned core so that they could be isolated and centrally sectioned for EPMA.

3.4 Phase equilibria modelling

Rock-specific equilibrium assemblage diagrams and garnet end-member isopleths were calculated using THERIAK-DOMINO software (de Capitani and Petrakakis, 2010). This was done in the MnO-Na₂O-CaO-K₂O-FeO-MgO-Al₂O₃-H₂O-TiO₂ model system using the internally consistent database of Holland and Powell (1998). In addition, feldspar (FSP) was modelled as a ternary solid solution (Baldwin et al., 2005), garnet (GRT) as a quaternary solution according to Connolly (2005), biotite (BT) corresponds to the model of White et al. (2007), chloritoid (CTD) solid solution follows White et al. (2000) with the addition of the Mn-bearing end-member, and chlorite (CHL) follows that of Gaidies et al. (2008). White mica (WM) is modelled according to Coggon and Holland (2002), cordierite (CRD) corresponds to the model of Baldwin et al. (2005), and the thermodynamic model of Holland and Powell (1998) was extended to consider the Mn-bearing end-member of staurolite (ST) assuming ideal mixing. An ideal solution model was used for ilmenite (ILM). The metamorphic fluid was assumed to be pure H₂O due to negligible amounts of carbon present in the studied samples. SiO₂ and H₂O are considered to be in excess for all calculations, and all iron is considered as FeO since ferric iron-bearing phases are not present in the samples.

3.5 Monazite U-Th-Pb geochronology

Monazite chemical age dating was done using a Cameca SX ultrachron electron microprobe at the University of Massachusetts in Amherst, MA (USA). Monazite grains were analyzed in-situ in order to preserve textural relationships within the sample. Prior to analysis, thin

sections were double-coated first in 250 μm Al and then a layer of C. Microprobe operating conditions were an accelerating voltage of 15 kV and a beam current of 20 nA. Selected monazite grains were mapped for U, Th, Y, and Ca, prior to detailed spot analysis in order to visualize the extent of chemical variation across the grains. 2 - 6 spot analyses were done per grain, and the average of these analyses determines the ages reported here. The first of these spots is a background analysis that is used to correct for background effects on later spots in that grain. Each spot analysis is approximately 0.5 μm in diameter. Sets of spot analyses are constrained, as best possible, to domains of uniform chemical composition. Count times were 1400 s for Pb, 500 s for U, 400 s for Th, and 100 s for Y. A Cameca PAP matrix correction (Pouchou and Pichoir, 1985) was used to correct for matrix effects on the analyses, and a time-dependent intensity (TDI) correction was used for accurate estimation of initial intensities for U, Th, and Pb. Moacyr monazite (ca. 508 Ma) was used as a consistency standard, and was analyzed periodically during each session. For a complete description of the analytical method used see Jercinovic et al. (2008) and Williams et al. (2007).

4 Results

4.1 Petrography

Modal percentages of phases of the rocks studied (Tab. 1) are determined using volume estimations obtained from optical petrography and XR- μCT data. The notation of S_1 defines the main foliation of the rock, and S_0 defines a planar fabric seen as inclusion trails within

some porphyroblasts. White mica in all samples is predominantly muscovite as determined using EPMA.

4.1.1 14PBA-67A

This sample (Fig. 3A) contains an assemblage of quartz (40 - 50 vol%), biotite (10 - 20 vol%), white mica (5 - 10 vol%), plagioclase (10 - 20 vol%), and andalusite (2 - 5 vol%) which is commonly partly replaced by white mica. Accessory minerals include tourmaline, zircon, and ilmenite. Chlorite seen at the edges of biotite grains is inferred to be part of a retrograde assemblage. S_1 is defined by fine-grained biotite and white mica weakly wrapping andalusite porphyroblasts that measure up to 2 cm in diameter. Inclusions within andalusite are composed primarily of quartz and biotite, with rare plagioclase and white mica.

4.1.2 14PBA-85A

14PBA-85A (Fig. 3B) is composed of an assemblage of biotite (10 - 20 vol%), white mica (20 - 30 vol%), quartz (20 - 30 vol%), plagioclase (5 - 10 vol%), and andalusite (10 - 20 vol%) and contains trace amounts of zircon, monazite, and ilmenite. Similar to the other samples studied, minor amounts of retrograde chlorite replaces biotite. A well-developed S_1 is defined by biotite and white mica and wraps up to 2 cm sized andalusite porphyroblasts. Andalusite is irregularly shaped and randomly distributed containing abundant inclusions of quartz, plagioclase, biotite, and white mica. These inclusions are not aligned.

212 4.1.3 14PBA-97A

213 Quartz (20 - 30 vol%), biotite (10 - 20 vol%), white mica (10 - 20 vol%), plagioclase (10 - 20
214 vol%), and andalusite (10 - 20 vol%) form the dominant mineral assemblage of this sample
215 (Fig. 3C). Biotite grains are commonly rimmed by retrograde chlorite. Accessory minerals
216 include tourmaline, zircon, pyrite, and ilmenite. Porphyroblastic andalusite is up to 3 cm in
217 diameter and contains randomly oriented inclusions of quartz, plagioclase, biotite, and white
218 mica. S₁ formed by biotite and white mica wraps the porphyroblasts. Grain boundaries and
219 cracks within andalusite are replaced by fine-grained white mica. Andalusite porphyroblasts
220 show sericite alteration along the crystal faces and along fractures. Approximately 50 % of
221 the original andalusite volume has been altered to sericite.

222 4.1.4 14PBA-100

223 This sample (Fig. 3D) contains a mineral assemblage of quartz (30 - 40 vol%), white mica (20
224 - 30 vol%), biotite (10 - 20 vol%), plagioclase (5 - 10 vol%), and andalusite (10 - 20 vol%).
225 Rare retrograde chlorite replaces the edges of some of the matrix biotite grains. Accessory
226 minerals include tourmaline, zircon, pyrite, and ilmenite. White mica and biotite define S₁
227 which wraps andalusite porphyroblasts that are up to 3 cm in diameter. Andalusite porphy-
228 roblasts contain randomly oriented inclusions of quartz, plagioclase, biotite, and muscovite.

229 4.1.5 12PBA-27A

230 This sample contains an assemblage of garnet (2 - 5 vol%), biotite (20 - 30 vol%), quartz (20
231 - 30 vol%), plagioclase (10 - 20 vol%), coarse-grained white mica (2 - 5 vol%), and chlorite
232 (1 - 2 vol%) (Fig. 3E). Fine-grained, irregularly shaped, sericite-rich clusters up to 5 cm
233 in diameter comprise another 10 - 20 vol%. These clusters contain abundant inclusions of
234 quartz, plagioclase, and biotite and are interpreted to be white mica pseudomorphs after
235 andalusite based on their morphology and the andalusite-bearing phase assemblages seen
236 in the other lower Montresor group samples. Accessory minerals in the sample include il-
237 menite, pyrite, pyrrhotite, tourmaline, zircon, and monazite. Chlorite occurs both as part
238 of the peak assemblage in the rock matrix and along grain boundaries and cracks within
239 garnet and biotite where it is inferred to be of retrograde origin. S_1 is well-defined by biotite
240 and commonly wraps both garnet and the sericite-rich clusters. Coarse-grained white mica
241 in the rock matrix is seen crosscutting S_1 at random orientations and is also interpreted to
242 be of retrograde origin. Garnet crystals are less than 2 mm in diameter and are commonly
243 embayed with pitted surfaces observable in the XR- μ CT data (Fig. 4). Garnet contains abun-
244 dant inclusions of quartz and plagioclase, ilmenite, pyrite, pyrrhotite, zircon, and monazite.
245 These inclusions define a straight to weakly curved S_0 that runs at an oblique angle to S_1 .
246 Garnet typically contains roughly parallel microscopic to macroscopic fractures (Fig. 4) ap-
247 proximately perpendicular to S_1 , which are commonly filled with biotite and chlorite. Where
248 sericite (pseudomorphic) is in contact with garnet, it also occurs in the cracks. Straight grain
249 boundaries are observed between biotite, garnet, and the white mica pseudomorphs.

250 4.2 Garnet shape, size and spatial distribution in 12PBA-27A

251 Garnet in 12PBA-27A is close to spherical but highly corroded, and pitted surfaces are
252 typical (Fig. 4). Its crystal size distribution (CSD) as determined by XR- μ CT is given in
253 Fig. 5. Note that the number of crystals given is per volume of the scanned rock cylinder.
254 The CSD illustrates a general increase of crystal abundance with a decrease of crystal size
255 which is uncommon for garnet that formed in a single metamorphic event (e.g., Gaidies
256 et al., 2011). Polymetamorphism may be a possible explanation but resorption of garnet
257 may be a more likely cause given the petrographic evidence with chlorite present in cracks
258 of garnet and along its surfaces (Fig. 3).

259 Results of the statistical analysis of the 3D spatial and size distribution of garnet using
260 the software Reduce3D (Hirsch, 2011) are given in Fig. 5. Whereas the application of the
261 pair correlation function (PCF) indicates that garnet crystal centres are clustered at any
262 test distance, the mark covariance function (MCF) suggests that garnet crystal sizes are
263 randomly distributed. Hence, the MCF results point to garnet growth kinetics independent
264 of nutrient diffusion across the rock matrix but that garnet growth was rather controlled by
265 the rate of interface reactions (e.g., Kretz, 1974; Carlson, 1989, 1991; Hirsch et al., 2000;
266 Hirsch, 2011; Petley-Ragan et al., 2016; Gaidies et al., 2017). However, given the evidence
267 of significant garnet resorption, results of the MCF statistics may be of limited use. The
268 PCF, however, does not consider crystal sizes but crystal centres instead which should be
269 unaffected by resorption processes. Assuming that crystal centres correspond to nucleation
270 sites, the PCF results indicate the uneven, clustered distribution of energetically favourable

locations for garnet nucleation during metamorphism of this sample.

4.3 Whole-rock chemistry

The AFM diagram (Fig. 6) indicates that, with the exception of 12PBA-27A, all samples investigated in this study plot at F/M ratios similar to the average pelite of Symmes and Ferry (1991) but differ with regards to their A-values. 12PBA-27A is slightly enriched in F/M compared to the average pelite. Whereas the bulk chemistry of 14PBA-67A is almost identical to the reference sample of Symmes and Ferry (1991) in terms of its position in an AFM projection, 14PBA-85A, 14PBA-97A, and 14PBA-100 have higher A-values and 12PBA-27A has a lower A-value compared to the reference sample.

4.4 Chemical composition of garnet

The qualitative garnet chemical maps (Fig. 7) obtained for the central section of the largest garnet porphyroblast in 12PBA-27A show concentric chemical zoning of X_{Fe} , X_{Mg} , X_{Ca} , and X_{Mn} . Inspection of the XR- μ CT data indicates that this section is not more than ca. 30 - 90 μ m away from the true geometric centre of the grain. The X_{Mg} chemical map (Fig. 7C) clearly displays the lower X_{Mg} in the vicinity of the parallel cracks running through the garnet. Changes to the chemical zoning near cracks can also be seen in the X_{Fe} (Fig. 7D). A 600 μ m wide area of high Y concentration is seen at the garnet core, as well as a ring of elevated Y concentration within 50 - 100 μ m of the rim (Fig. 7F).

An EPMA traverse was also performed on the central section of this garnet (Fig. 8), with X_{Fe}

290 $= (\text{Fe}^{2+}/\text{Ca}+\text{Mg}+\text{Fe}^{2+}+\text{Mn})$, $X_{Ca} = (\text{Ca}/\text{Ca}+\text{Mg}+\text{Fe}^{2+}+\text{Mn})$, $X_{Mg} = (\text{Mg}/\text{Ca}+\text{Mg}+\text{Fe}^{2+}+\text{Mn})$,
 291 and $X_{Mn} = (\text{Mn}/\text{Ca}+\text{Mg}+\text{Fe}^{2+}+\text{Mn})$. The garnet, with a radius of 1.24 mm, shows an over-
 292 all trend of increasing X_{Fe} and X_{Mg} , and decreasing X_{Mn} and X_{Ca} from core to rim. X_{Fe}
 293 values range from 0.78 in the core to 0.82 in the rim. Repeating peaks or spikes in X_{Fe} values
 294 as high as 0.815 near the core correspond to zones adjacent to cracks in the garnet. X_{Mg}
 295 shows values of 0.05 in the core rising slightly to 0.08 in the rim. X_{Mg} values decrease to
 296 0.04 adjacent to fractures. A sharp decrease in X_{Mg} values is seen within 50 μm of the rim.
 297 The zoning of X_{Mn} reaches 0.10 in the core and smoothly drops to 0.05 in the rim. A sharp
 298 increase within 50 μm of the rim brings X_{Mn} up to 0.06.

299 Additional central sections of garnet of size classes 2 to 5 were analyzed by EPMA. These
 300 indicate that X_{Fe} and X_{Mg} core contents increase whereas X_{Mn} core content decreases with
 301 a decrease of crystal size (Table 3), a feature commonly observed in populations of garnet
 302 porphyroblasts in which garnet crystal sizes increase with the age of their cores (George and
 303 Gaidies, 2017).

304 4.5 *P-T conditions of metamorphism of the lower Montresor group*

305 Sample 12PBA-27A contains the peak mineral assemblage quartz + plagioclase + garnet
 306 + andalusite + biotite + chlorite + ilmenite (Fig. 3E). The sericite clusters observed in
 307 this sample are inferred to be white mica pseudomorphs after andalusite. Thus, whereas
 308 andalusite porphyroblasts are completely pseudomorphed by white mica, andalusite is still
 309 interpreted to be part of the stable peak mineral assemblage. White mica is not interpreted

310 to be part of the stable assemblage as it is only observed in these pseudomorphs and as
311 individual grains in the matrix that lie at an angle to S_1 , and hence are interpreted to
312 be retrograde. Phase equilibrium modelling yields the observed mineral assemblage at P - T
313 conditions of ca. 525 - 540 °C and 1.9 to 2.3 kbar (Fig. 9A).

314 Garnet core compositions used for isopleth thermobarometry (Gaidies et al., 2006) applied
315 to 12PBA-27A were carefully chosen to represent those parts of the crystal least affected
316 by the alterations adjacent to cracks (Fig. 8). However, the obtained area of garnet-core
317 end-member isopleth intersection does not focus on a point in P - T space but stretches from
318 ca. 528 to 542 °C and 2.2 to 2.4 kbar (Fig. 9F). Hence, the area of isopleth intersection
319 overlaps with the P - T field obtained for the observed mineral assemblage of this sample.
320 The centre of the intersection area is at a T ca. 10 °C higher than the garnet-in curve.

321 The peak assemblages in 14PBA-67A, 14PBA-85A, 14PBA-97A and 14PBA-100 are quartz
322 + plagioclase + andalusite + biotite + ilmenite + white mica (Fig. 3) and can be predicted
323 for similar P - T conditions (Fig. 10). The P - T fields overlap, and the centre of their inter-
324 section area plots at ca. 575 °C and ca. 3 kbar (star in Fig. 10E), assumed here to reflect
325 the peak metamorphic conditions experienced by these samples. These conditions are ca.
326 45 °C and ca. 0.75 kbar higher than the P - T conditions inferred from the peak assemblage
327 developed in 12PBA-27A. The arrow in Fig. 10E connects those P - T fields and may be part
328 of the prograde P - T path experienced by the lower Montresor group, with a gradient of ca.
329 50 °C kbar⁻¹.

330 4.6 *U-Th-Pb microprobe ages of monazite*

331 In-situ monazite geochronology was done for sample 12PBA-27A (see Appendix A1 for the
332 geochronological dataset). Monazite in this sample proved to be sparse, with generally less
333 than 20 grains per thin section large enough (more than 3 μm in diameter) to be considered
334 for analysis. Three monazite grains included in garnet were analyzed (2b, 19a, 19b) (Fig.
335 11B). These monazite grains are located towards the core of the garnets. Almost all monazite
336 inclusions are subhedral and are present in other mineral inclusions in garnet, such as apatite
337 or quartz, or along cracks in garnet. Note that these cracks are not filled with chlorite and
338 biotite and texturally unrelated to the approximately parallel sets of cracks running across
339 the sample. The calculated weighted average is 1829 ± 25 Ma for monazite 2b, 1905 ± 29
340 Ma for 19a, and 1845 ± 5 Ma for 19b (Fig. 11b).

341 In addition, a total of 10 matrix monazite grains were analyzed, with a variety of grain
342 morphology and composition seen across the monazite population (Fig. 11a). Seven of these
343 monazite grains are euhedral to subhedral (monazite 15-core, 17-core, 18-core, 22, 23a, 23b,
344 23c) and are inferred to represent the matrix monazite population. Grains 15, 17, and 18
345 have rims that differ in composition from the core. These were analyzed separately and
346 not included in the same age calculation (Fig. 11b). The matrix monazite grains commonly
347 display heterogeneous compositions with respect to Ca and Th (Fig. 11a). For example, the
348 core of monazite 15 shows an analyzed Th content that varies from 21190 to 54862 ppm.
349 The age calculations for single monazite grains show a wide spread in ages (Fig. 11b), from
350 1837 ± 9 Ma to 1870 ± 9 Ma. This spread in calculated ages does not appear to correlate

with variation in composition.

Chemical mapping of monazite 15 shows an anomalous low Y rim (15-rim, Fig. 11a). The low Y rim (ca. 5000 ppm Y) appears homogeneous in Y, Ca, Th, and U whereas the core (ca. 7000 ppm Y) is heterogeneous in Th and Ca. This is the only matrix monazite in the studied population to show this compositional feature. Analysis of the low Y rim yields a calculated age slightly younger than the core of the grain (Fig. 11b), with the core giving an age of 1870 ± 13 Ma and the rim an age of 1837 ± 9 Ma.

Chemical mapping of monazites 17 and 18 shows narrow (less than $5 \mu\text{m}$) high-Y rims (17-rim, Fig. 11a). Cores of these grains have Y contents of ca. 2000 ppm and the rims have Y contents of ca. 5000 ppm. The calculated ages for the high Y rims are slightly younger, but within error of the calculated ages for the monazite cores (Fig. 11b). Monazite 17 gives calculated ages of 1853 ± 4 Ma for the rim and 1852 ± 22 Ma for the core. Monazite 18 yielded slightly older calculated ages with 1870 ± 9 Ma for the core and 1866 ± 10 Ma for the rim.

Two monazite grains (f25a, f25b) in proximity to a white mica pseudomorph after andalusite differ from the other matrix monazite grains in that they display a highly corroded morphology (f25b, Fig. 11a). These grains show chemical heterogeneity with respect to Ca, Th and U. The largest chemical variation is with respect to Th, where totals range from ca. 5500 to 19000 ppm. In general, parts of the monazite grain with low Th also yield lower U content. As lower Th content is correlated with younger calculated ages, the spot analyses were divided in order to calculate a weighted average age for the low-Th and high-Th do-

372 mains. Another monazite grain, f25c, shows a subhedral grain morphology. Analysis of f25c
373 yields a Th content similar to the high-Th domains of f25a and f25b as well as a similar age.
374 Calculated ages for the high-Th domains of f25a and f25b are 1842 ± 29 Ma and 1836 ± 19
375 Ma, respectively (Fig. 11b). A weighted average of ages for the low-Th domains of monazites
376 f25a and f25b, and the analysis on f25c, gives an age of 1792 ± 10 Ma.

377 5 Interpretation and Discussion

378 5.1 *Influence of fluid infiltration on metamorphic phase equilibria, and the conditions of* 379 *andalusite replacement*

380 The reason for the considerable size of the garnet core isopleth intersection area and its dis-
381 tance to the garnet-in curve obtained for 12PBA-27A is a too high spessartine content and a
382 too low almandine content predicted for this sample compared to the observed concentrations
383 in the garnet core (Fig. 9). In general, off-centring of the chemical core and/or diffusional
384 modification of the garnet core composition may explain those discrepancies. However, sim-
385 ulations of garnet crystallization using the THERIA_G software (Gaidies et al., 2008) could
386 not predict the core or rim compositions of garnet in 12PBA-27A regardless of the P - T
387 scenario for reasonable geological time scales. This suggests that intracrystalline diffusion
388 alone cannot explain the obtained discrepancies.

389 Significant overstep during garnet nucleation (e.g., Gaidies et al., 2011; Pattison et al.,
390 2011; Spear, 2017; Gaidies, 2017) or inappropriate thermodynamic models for the rock-

forming phases, as well as diffusion-controlled crystallization (e.g., Carlson, 1989) might be alternative explanations. The latter seems unlikely given the clustered distribution of crystal centres and the random distribution of garnet crystal sizes in 12PBA-27A (Fig. 5). Instead, the petrographic evidence of rock fracturing and the retrograde replacement of garnet by chlorite and biotite along cracks and grain surfaces may point to fluid infiltration post garnet crystallization. Therefore, a bulk rock composition not fully representative of the thermodynamically effective chemical system during garnet growth may be the most likely cause of the differences between the observed and predicted garnet compositions. The relatively low P - T conditions obtained for the assemblage in 12PBA-27A compared with the other samples of this study may also be due to the uncertainties in the phase equilibria modelling of 12PBA-27A resulting from the modification of its bulk chemical composition post peak metamorphism.

Whereas partly or completely replaced andalusite is observed in samples 12PBA-27A, 14PBA-67A and 14PBA-97A (Fig. 3), andalusite in 14PBA-85A and 14PBA-100 is largely intact and not replaced by white mica (Fig. 3). The reason for these differences in mineral content and microstructure of the samples studied may be their rock-specific phase equilibria. For 14PBA-85A and 14PBA-100, and assuming that P did not change from its peak at ca. 3 kbar, the andalusite-out reaction (going down- T) is predicted to take place at ca. 500 °C and less than 450 °C, respectively (Fig. 10). The andalusite-out curves for 12PBA-27A, 14PBA-67A and 14PBA-97A are predicted at a significantly higher T of ca. 540 °C at ca. 3 kbar (Fig. 9, 10). We therefore infer that the reason for the replacement of andalusite only in certain samples may be due to primary differences in the bulk chemical compositions of

413 the rocks.

414 Since andalusite is completely replaced by white mica in 12PBA-27A but only partly in
415 14PBA-67A and 14PBA-97A, we hypothesize that the replacement took place close to
416 their andalusite-out curves. Assuming similar andalusite dissolution kinetics at the different
417 andalusite-out conditions of the samples studied, 14PBA-85A and 14PBA-100 may have
418 cooled too quickly during later stages in their P - T histories for the transformation to occur
419 whereas the higher T andalusite-out curve may have been crossed at a slower rate. Differ-
420 ent degrees of completeness of the andalusite-out reaction in 12PBA-27A, 14PBA-67A and
421 14PBA-97A may reflect local variations in fluid availability required for the transformation,
422 with more fluid present in 12PBA-27A than in the other samples. However, given that the
423 thermobarometric results obtained for the different samples are within error of each other
424 and in agreement with results obtained by Berman et al. (2015) for the lower Montresor
425 group, it may be assumed that post-peak fluid infiltration had a relatively minor impact on
426 the bulk rock composition of 12PBA-27A. Whereas the modification of the bulk rock com-
427 position of 12PBA-27A had a negligible influence on the estimated P - T conditions obtained
428 from the configuration of its stable peak assemblage, it resulted in a measurable effect on
429 the predicted garnet chemical composition.

430 5.2 *Timing of peak metamorphism, garnet growth, and fluid infiltration*

431 The monazite ages determined in this study are similar to results previously determined
432 for metamorphism in the Montresor Belt. Davis et al. (2014) report in-situ U-Th-Pb mon-

azite ages of 1861 ± 7 and 1844 ± 6 Ma. These ages are within the range of ca. 1870 - 1837 Ma matrix monazite ages from sample 12PBA-27A obtained in this study. This range of monazite ages is interpreted to represent continuous stages of monazite growth during prograde metamorphism, with peak metamorphism possibly occurring at ca. 1837 Ma. This would suggest that metamorphism in the Montresor Belt reached peak conditions ca. 73 Ma after the earliest estimate for the onset of the Trans-Hudson orogeny at 1.91 Ga. This time lag has previously been interpreted as the time needed for crustal thickening and thermal incubation after collision (England and Thompson, 1984).

Garnet in 12BPA-27A can only be as old or younger than the monazite grains it overgrew which within error have ages similar to the dominant matrix monazite population: 1829 ± 25 Ma and 1845 ± 5 Ma. Matrix monazite f25c, texturally associated with white mica pseudomorphs after andalusite, yields an age of ca. 1790 ± 12 Ma and is assumed to have crystallized during fluid infiltration. It is interpreted here that the low Th domains of monazite f25a and f25b which are of similar age compared to f25c and also spatially associated with the pseudomorphs, represent altered domains that became reset at the time of fluid infiltration (Tartése et al., 2011; Williams et al., 2011). Since monazite in garnet is significantly older than monazite f25c, f25a and f25b, it may be assumed that garnet growth had finished before the pseudomorphic replacement of andalusite occurred. Monazite 19a is also included in garnet and has an older calculated age of 1905 ± 29 Ma. While this age may represent an earlier onset of metamorphism than indicated by the other monazite grains, there is a large error of ± 29 Ma making it difficult to constrain exact timing. It is also possible that this is an inherited monazite from an older source.

455 5.3 *Modification of garnet growth zoning, and the duration of the alteration event*

456 The reason for the numerous cracks observed across the rock volume and in garnet of 12PBA-
 457 27A (Figs. 4, 7) may be mechanical fracturing caused by internal stress buildup due to
 458 fluid infiltration and subsequent hydration reactions (e.g., Jamtveit and Austrheim, 2010;
 459 Keleman and Hirth, 2012; Floess and Baumgartner, 2013). Alternatively, the ca. parallel
 460 cracks may primarily reflect tectonic extensional forces, and then acted as preferred conduits
 461 during fluid infiltration. The compositional zoning of garnet in the vicinity of these cracks
 462 (Fig. 7, 8, 12) is interpreted here to reflect the results of diffusional exchange of the respective
 463 components between garnet and biotite/chlorite in the cracks providing valuable insight into
 464 the duration of the alteration event.

465 Assuming 1-dimensional multicomponent diffusion perpendicular to the crack surface, the
 466 general diffusion equation can be written (Onsager, 1945)

$$\frac{\partial C_i}{\partial t} = \frac{\partial}{\partial x} \sum_{j=1}^{n-1} D_{ij} \left(\frac{\partial C_j}{\partial x} \right) \quad (5.1)$$

where C_i is the concentration of component i in the n - component system, t is time, x is distance, and D_{ij} are the interdiffusion coefficients. According to Lasaga (1979),

$$D_{ij} = D_i^* \delta_{ij} - \left(\frac{D_i^* z_i z_j X_i}{\sum_{k=1}^n z_k^2 X_k D_k^*} \right) (D_j^* - D_n^*) \quad (5.2)$$

where δ_{ij} is the Kronecker delta, z is charge and X is mole fraction of the diffusing compo-

nent. The tracer diffusion coefficient D_i^* is given by

$$D_i^* = D_0 \exp \left(\frac{-E_a - (P - 1) \Delta V^+}{RT} \right) \quad (5.3)$$

with the pre-exponential constant D_0 , the activation energy E_a at 1 bar, the activation volume ΔV^+ , and the universal gas constant R . In our simulations, diffusion of Mn, Fe, Mg, and Ca is calculated numerically using an implicit finite differences scheme, and Ca is considered the dependent component. D_0 , E_a and ΔV^+ are those determined by Chakraborty and Ganguly (1992). Dirichlet conditions are assumed for the inner and outer boundaries, and step functions are used to approximate the initial distributions of X_{Mg} and X_{Fe} in the vicinity of the cracks, as illustrated in Fig. 12. No initial compositional gradients are assumed for X_{Mn} and X_{Ca} .

The width of the diffusion fields observed in the secondary zoning of X_{Mg} in the vicinity of the cracks is smaller than the diffusion distance that can be observed in the X_{Fe} zoning (symbols in Fig. 12) reflecting the larger diffusivity of Fe compared to that of Mg at similar T . Dissolution and precipitation of garnet adjacent to the cracks may be an alternative explanation for the secondary zoning but this process would not result in the observed systematic differences in the widths of the zoning patterns for individual garnet components. In general, the width of the observed diffusion fields are an order of magnitude wider than would be expected for common diffusion zoning that develops at typical cooling rates of more than 50 °C/My.

Fig. 12 shows the predicted X_{Mg} and X_{Fe} zoning that develops in the vicinity of cracks for

different durations of diffusion. The data plotted are based on the assumption that T and the garnet chemical composition in direct contact with biotite and chlorite along the cracks did not change during the diffusion. An estimate for the maximum T during the diffusion may be obtained from the presence in the cracks of biotite, chlorite and white mica (where in contact with the pseudomorphs after andalusite) indicating that andalusite replacement happened during or after crack formation. Using the equilibrium phase relations illustrated in Fig. 9 and assuming that there was no significant change in P compared to that during peak metamorphism, a maximum T of 540°C is obtained and used in the diffusion simulations shown in Fig. 12.

The cracks may seem planar macroscopically (Fig. 4) but their intersections with the plane of the thin section indicate that they likely have irregular, uneven surfaces at the microscopic scale (Fig. 7). Hence, diffusion across them most likely resulted in asymmetric diffusion profiles visible at that same length scale. It may be expected that this asymmetry will be enhanced by diffusion across additional submicroscopic cracks too small to be resolved with the analytical equipment used here, and that this would result in a stronger asymmetric zoning pattern and more scatter in the data for faster diffusing components (such as Fe) compared to slower ones (such as Mg). Fig. 12 illustrates the asymmetric X_{Mg} and X_{Fe} zoning, where between 50 to 100 My of diffusion are required to reproduce the compositional zoning of X_{Mg} in sector I, and between 10 to 30 My to predict the X_{Mg} zoning in sector II. The asymmetry is even more significant for X_{Fe} where the modelling suggests more than 100 My for the zoning in sector I and between 10 to 40 My for X_{Fe} in sector II. We argue that the asymmetry in the chemical zoning observed is due to uneven crack surfaces at the μm -scale

and the presence of submicroscopic sets of cracks that possibly anastomose. Inspection of the BSE and X-ray images (Fig. 7) supports that assumption. This is why X_{Mg} zoning in sector II is used here for the estimation of the duration of diffusion across the cracks.

It is important to note that the angle between the plane of the thin section and the set of cracks across the sample (and garnet) is ca. 60° (Fig. 4). Therefore, the extent of diffusion as well as the distances between cracks in Figs. 7 and 8 are only apparent length scales exceeding the real dimensions. Fig. 12 illustrates the simulation results taking account of the non-orthogonal geometry of the modelled system. According to those calculations, a best fit between the modelled and the observed X_{Mg} zoning is obtained for 10 to 30 My of diffusion at 540°C with most data points clustering around a diffusion duration of 20 My (Fig. 12). These findings support the hypothesis of an extended period of time relatively early during cooling as outlined above based on the observation that andalusite is replaced only in rocks with a relatively high andalusite-out T . Diffusion simulations were also conducted assuming crack formation at 515°C but resulted in geologically unrealistic timescales exceeding the proposed durations by almost an order of magnitude.

5.4 Regional context for low pressure - high temperature (LP-HT) metamorphism

According to our findings, garnet growth and peak metamorphism occurred at or after ca. 1837 ± 9 Ma and before ca. 1792 ± 10 Ma at conditions of $535 - 575^\circ\text{C}$ and $2.3 - 3$ kbar. These conditions represent LP-HT metamorphism (De Yoreo et al., 1991; Lux et al., 1986; Scrimgeour et al., 2001). In collisional events such as the Trans-Hudson orogeny, LP-HT

metamorphic conditions cannot be solely a result of crustal thickening (White, 2005; Sandiford and Powell, 1991). This is based on models which show that the Moho T would exceed the liquidus at the base of the crust if the amount of heat in $LP-HT$ metamorphism is attributed to crustal thickening alone (Sandiford and Powell, 1991). Instead, $LP-HT$ metamorphism in orogenic settings has been attributed to a combination of high heat flow at the base of the crust and advection of heat within the crust via plutonism and fluid circulation. This is supported by the observation that orogenic $LP-HT$ terranes are commonly associated with syn- to post-tectonic plutonic activity (De Yoreo et al., 1991; Lux et al., 1986; Sandiford and Powell, 1991; Stuewe et al., 1993). Syn- to post-tectonic plutonism also occurred during the Trans-Hudson orogeny, with the emplacement of the ca. 1845 - 1795 Ma Hudson granite suite as an example (van Breemen et al., 2005). It is estimated that melting at the base of the crust begins 10 - 40 Ma after the onset of crustal thickening (De Yoreo et al., 1991). This may be demonstrated by the time difference between the Snowbird collision at ca. 1.87 Ga (Berman et al., 2007) and the earliest age estimate of the associated Hudson suite at ca. 1845 Ma (van Breemen et al., 2005).

There is some debate as to whether plutons associated with $LP-HT$ metamorphic events are a driving heat source for metamorphism or a reflection of high heat flow at the base of the crust (De Yoreo et al., 1991). Monazite geochronology in this study indicates that peak metamorphism in the Montresor Belt occurred at ca. 1837 Ma. This is synchronous with the emplacement ages of the Hudson granitoids at ca. 1845 - 1795 Ma (van Breemen et al., 2005). These ages demonstrate that magmatic activity occurred with peak metamorphic conditions, suggesting that in this case syn-metamorphic plutonism likely advected heat to

higher crustal levels during regional metamorphism. In light of the older monazite ages (1870 - 1845 Ma), heating began shortly after the Snowbird collision, culminating in deep crustal melting that produced Hudson granitoids. Undated dykes and sills of granite cutting lower Montresor units may represent this intrusive suite in the field area.

The metamorphic grade of meta-siltstones of the upper Montresor Belt is poorly defined by assemblages including white mica, biotite, chlorite, quartz and feldspar. A single U-Pb SHRIMP monazite age of $ca. 1847 \pm 15$ Ma is interpreted as the time of upper greenschist facies metamorphism. Pressure constraints are not available but the difference in metamorphic grade between the upper and lower Montresor groups suggests that extension and movement along the detachment fault separating the groups occurred some time after peak metamorphic conditions. Late stage (D3) folding and shearing within the central Rae Province is estimated to have occurred at $ca. 1780$ Ma based on zircon and monazite ages from the Walter Lake shear zone of the Committee Bay belt (Berman et al., 2010; Sanborn-Barrie et al., 2014), and this event may be responsible for the synclinal structure of the Montresor Belt. Since no macroscopic fabric elements are associated with the open fold, the metamorphic grade was significantly lower at this time in the Montresor Belt than in the Committee Bay region, where metapelitic migmatites were forming (Carson et al., 2004). This suggests that extension and movement along the detachment fault separating the upper and lower Montresor groups must have occurred between $ca. 1837$ Ma (peak metamorphism) and $ca. 1780$ Ma (late-stage folding). With these age constraints in mind, it remains probable that regional extension and displacement along the detachment fault is coincident with fluid infiltration at $ca. 1792 \pm 10$ Ma.

The P - T - t results from this study for the Montresor Belt can be compared to metamorphic conditions predicted for the Trans-Hudson orogeny across the Rae Province. The Committee Bay group (ca. 200 km east) shows greenschist to amphibolite facies metamorphism occurring at 1.86 - 1.82 Ga (Sanborn-Barrie et al., 2014). Similarly to the lower Montresor group, the Woodburn Lake group (ca. 200 km south) experienced peak metamorphic conditions of 540 - 570 °C and 2.8 - 5 kbar at ca. 1834 ± 5 Ma (Pehrsson et al., 2013). These P - T - t conditions have been attributed to the Snowbird phase of the Trans-Hudson orogeny at ca. 1870 Ma which drove crustal thickening and consequent metamorphism (Berman et al., 2015; Sanborn-Barrie et al., 2014). In contrast, the Main Phase of Trans-Hudson deformation and metamorphism is expressed to the northeast, in the collision zone between the eastern Superior promontory and Trans-Hudson internides (St-Onge et al., 2006). There, eclogite-facies metamorphism records active subduction constrained by monazite U-Pb ages of ca. 1831 ± 5 Ma (Weller and St-Onge, 2017). Regional metamorphic monazite has a complex growth history and suggests that peak T conditions occurred ca. 1805 Ma (Gagné et al., 2009).

6 Conclusions

The rocks of the lower Montresor group investigated in this study were metamorphosed as a result of crustal thickening during the Trans-Hudson orogeny at ca. 1870 - 1837 Ma. They experienced peak metamorphic conditions of ca. 575 °C and 3 kbar reflecting a geothermal gradient of ca. 55 °C/km typical of LP-HT metamorphism in continental crust of common

composition. Syn-tectonic plutonism and increased fluid circulation are assumed to have been a driving source of heat during the regional metamorphism. Diffusional modification of garnet growth zoning in the vicinity of cracks formed during retrogression at ca. 1792 Ma indicates an extended period of elevated T also after peak metamorphism. Metamorphic phase equilibria and geospeedometry calculations point to a T of ca. 540 °C that prevailed in the crust for ca. 20 Ma during the retrogression event. In addition to the modification of garnet growth zoning, the retrogression event resulted in the crystallization of secondary biotite, chlorite and monazite, and the pseudomorphic replacement of andalusite by white mica.

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Table Captions

Table 1. Main mineral contents of the samples (in order of decreasing abundance), and sample locations (NAD 1983 UTM Zone 14N).

Table 2. Bulk chemical composition of the studied samples (wt%).

Table 3. Core compositions of centrally sectioned garnet crystals of the 5 largest size classes (see garnet CSD in Fig. 5) of 12PBA-27A.

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824 **Tables**

Table 1

	Mineralogy ^a	Easting	Northing
14PBA-67A	Qtz, Bt, Pl, Wm, And ^b	578149	7325933
14PBA-85A	Wm, Qtz, Bt, And, Pl	577392	7325307
14PBA-97A	Qtz, Bt, Wm, And ^b , Pl	578370	7325495
14PBA-100	Qtz, Wm, Bt, And, Pl	578828	7325118
12PBA-27A	Qtz, Bt, And ^b , Pl, Grt, Wm	578147	7325917

^a Wm = white mica. Other abbreviations after Whitney and Evans (2004).

^b Partly or completely pseudomorphed by Wm.

Table 2

	SiO ₂	Al ₂ O ₃	MnO	MgO	Fe ₂ O ₃	CaO	Na ₂ O	K ₂ O	TiO ₂	LOI	Σ
14PBA-67A	62.11	20.50	0.05	2.30	5.63	0.73	0.80	4.87	0.63	3.01	100.70
14PBA-85A	55.73	24.47	0.05	2.63	6.84	0.28	0.32	4.68	0.83	3.02	98.99
14PBA-97A	47.58	26.83	0.08	3.00	8.36	1.53	1.77	4.95	0.95	3.97	99.16
14PBA-100	60.26	25.05	0.04	1.93	4.22	0.10	0.30	4.31	0.73	2.15	99.15
12PBA-27A	60.85	16.43	0.05	2.34	11.47	1.08	0.80	3.53	0.55	2.98	100.17

Table 3

Size class	1	2	3	4	5
SiO ₂	35.74	36.43	37.12	36.77	36.64
TiO ₂	0.01	0.01	0.04	0.03	0.05
Al ₂ O ₃	20.54	20.82	20.46	21.15	21.07
FeO	34.85	34.48	34.73	35.87	36.30
MnO	4.38	3.71	3.68	3.17	2.43
MgO	1.27	1.52	1.42	1.64	1.77
CaO	1.72	2.08	2.42	1.96	1.85
Total	98.55	99.05	99.87	100.59	100.11
Oxygen	24.00	24.00	24.00	24.00	24.00
Si	5.92	5.98	6.05	5.96	5.96
Ti	0.00	0.00	0.00	0.00	0.00
Al	4.01	4.03	3.93	4.04	4.04
Fe ³⁺	0.16	0.00	0.00	0.00	0.00
Fe ²⁺	4.67	4.74	4.73	4.86	4.94
Mn	0.61	0.52	0.51	0.44	0.34
Mg	0.31	0.37	0.35	0.40	0.43
Ca	0.31	0.37	0.42	0.34	0.32
X _{Mn}	0.10	0.09	0.08	0.07	0.06
X _{Ca}	0.05	0.06	0.07	0.06	0.05
X _{Fe}	0.79	0.79	0.79	0.80	0.82
X _{Mg}	0.05	0.06	0.06	0.07	0.07

Figure Captions

Fig. 1. Simplified geological map of the Western Churchill Province. Groups belonging to the Rae Cover Sequence (Montresor group, Amer group, Ketyet River group) are shown in yellow. The Montresor Belt and approximate location of map in Fig. 2 is shown by the black dashed box. Mo = Montresor Belt. STZ = Snowbird Tectonic Zone. CFZ = Chesterfield Fault Zone. CMZ = Chantrey Mylonite Zone (modified after Rainbird et al., 2010).

Fig. 2. (A) Geological map of the Montresor Belt, and (B) the locations of the samples studied (modified after Percival et al., 2015a).

Fig. 3. Microphotographs of (A) 14PBA-67A, (B) 14PBA-85A, (C) 14PBA-97A, (D) 14PBA-100, and (E) 12PBA-27A. Wm = white mica; all other mineral abbreviations are after Whitney and Evans (2004).

Fig. 4. XR- μ CT results obtained for a cylinder of rock of 12PBA-27A. Wm = white mica. Other mineral abbreviations are after Whitney and Evans (2004).

Fig. 5. (A) Garnet CSD and (B) results of spatial statistical correlation functions (Hirsch, 2011) applied to garnet of 12PBA-27A. The $2\text{-}\sigma$ envelope in (B) corresponds to theoretical results obtained from non-clustered interface-reaction controlled garnet crystallization simulations according to Hirsch et al. (2000). The *vertical bars* correspond to the mean nearest-neighbour distance centred on the mean, with a $2\text{-}\sigma$ width. PCF - pair correlation function; MCF - mark correlation function. The semi-major axes are used as mark value according to Petley-Ragan et al. (2016).

845 Fig. 6. AFM projection (Thompson, 1957) from muscovite, quartz and H₂O of the whole
846 rock chemistries of the samples investigated in this study (Table 2). Total iron is assumed
847 to be FeO.

848 Fig. 7. Backscattered electron image (A) and X-ray maps (B-F) of a central section through
849 the largest garnet porphyroblast (size class 1) of sample 12PBA-27A. The location of the
850 compositional profile as illustrated in Fig. 8 is shown in (A) as well as the higher resolution
851 (HR) transect of Fig. 12. Note the numerous subparallel cracks commonly filled with Chl
852 and Bt, and the chemical alterations adjacent to them.

853 Fig. 8. Compositional profile across the largest garnet of 12PBA-27A (size class 1). See Fig.
854 7 for the position of the A-A' traverse and the location of the higher-resolution profile (HR)
855 of Fig. 12. The arrows mark the chemical core used for isopleth intersection (Fig. 9), and
856 the vertical dashed lines mark the positions of cracks.

857 Fig. 9. (A) Equilibrium assemblage diagram and (B-E) garnet end-member isopleths for the
858 garnet-bearing sample 12PBA-27A. Qtz and H₂O are part of any assemblage. *P-T* conditions
859 of garnet-bearing assemblages are shown in grey, and the predicted formation conditions of
860 the observed peak assemblage is shown in dark blue. WM= white mica; all other phases
861 after Whitney and Evans (2004). See text for details.

862 Fig. 10. (A-D) Equilibrium phase relations for the garnet-free samples 14PBA-67A, 14PBA-
863 85A, 14PBA-97A, 14PBA-100, and (E) assumed *P-T* path of metamorphism for the rocks
864 of the lower Montesor group. Qtz and H₂O are part of any assemblage. *P-T* conditions of
865 garnet-bearing assemblages are shown in grey, and the predicted formation conditions of the

866 observed peak assemblages are shown in dark blue. WM= white mica; all other phases after
867 Whitney and Evans (2004).

868 Fig. 11. (A) Examples of the textural and chemical properties of some matrix monazites in
869 12PBA-27A, and (B) the calculated weighted average ages obtained in this study.

870 Fig. 12. Results of diffusion simulations in the vicinity of cracks in garnet. Zoning of X_{Ca}
871 and X_{Mn} is minor and not shown for legibility.

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872 **Figures**

Figure 1

Fig. 1.

Fig. 2.

Fig. 3.

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Fig. 9.

Fig. 10.

Fig. 11.

Fig. 12.

873 **Supplementary Material**

874 **Appendix A1.** Geochronology dataset.

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