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Journal:	<i>Canadian Journal of Earth Sciences</i>
Manuscript ID	cjes-2018-0039.R1
Manuscript Type:	Article
Date Submitted by the Author:	04-May-2018
Complete List of Authors:	Markley, Michelle; Mount Holyoke College, Geology and Geography Dunn, S.R.; Dept of Geography and Geology Jercinovic, Michael; U. of Mass., Geosciences Peck, William; Department of Geology, Williams, Michael; Dept of Geosciences,
Keyword:	Grenville, Monazite, Central Metasedimentary Belt, U-Th-Pb, shear zone
Is the invited manuscript for consideration in a Special Issue? :	N/A

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**Monazite U-Th-Pb Geochronology of the Central Metasedimentary Belt Boundary Zone
(CMBbz), Grenville Province, Ontario Canada**

Markley, Michelle J. * (1), Steven R. Dunn (1), Michael J. Jercinovic (2), William H. Peck (3),
and Michael L. Williams (2)

(1) Department of Geology and Geography, Mount Holyoke College, South Hadley,
Massachusetts 01075

(2) Department of Geosciences, University of Massachusetts, Amherst, Massachusetts 01003
USA

(3) Department of Geology, Colgate University, Hamilton, New York 13346 USA

*corresponding author: Dr. Michelle J. Markley, Associate Professor of Geology at Mount
Holyoke College, full street address: Department of Geology and Geography, Mount
Holyoke College, 50 College Street, South Hadley, MA 01075 USA, work phone (413) 538-
2814, work fax (413) 538-2239, email: mmarkley@mtholyoke.edu

Abstract

The Central Metasedimentary Belt boundary zone (CMBbz) is a crustal-scale shear zone that juxtaposes the Central Gneiss Belt and the Central Metasedimentary Belt of the Grenville Province. Geochronological work on the timing of deformation and metamorphism in the CMBbz is ambiguous, and the questions that motivate our study are: How many episodes of shear zone activity did the CMBbz experience, and what is the tectonic significance of each episode? We present electron microprobe data from monazite (the U-Th-Pb chemical method) in order to directly date deformation and metamorphism recorded in five garnet-biotite gneiss samples collected from three localities of the CMBbz of Ontario (West Guilford, Fishtail Lake, and Killaloe). All three localities yield youngest monazite dates ca. 1045 Ma; most of the monazite domains that yield these dates are high Y rims. In comparison to this common late Ottawa history, the earlier history of the three CMBbz localities is less clearly shared. The West Guilford samples have monazite grain cores that show older high Y domains and younger low Y domains; these cores yield a prograde early Ottawa (1100-1075 Ma) history. The Killaloe samples yield a well-defined prograde, pre- to early Shawinigan history (i.e. 1220-1160 Ma) in addition to some evidence for a second early Ottawa event. In other words, the answers to our research questions are: three events; a Shawinigan event possibly associated with crustal thickening, an Ottawa event possibly associated with another round of crustal thickening, and a late Ottawa event that resists simple interpretation in terms of metamorphic history but that coincides chronologically with crustal thinning at the base of an orogenic lid.

KEYWORDS: Grenville, Monazite, Central Metasedimentary Belt, U-Th-Pb, shear zone.

Introduction

The Grenville Province of New York, southern Ontario, and Quebec provides a window into the middle and lower crust of a major orogenic belt. From northwest to southeast are the granulite-facies Central Gneiss Belt; the lower-grade Central Metasedimentary Belt; and the Central Granulite Terrain, including the Adirondack Highlands of upstate New York (Figure 1; terminology of Wynne-Edwards 1972). Geologists have struggled to place these Mesoproterozoic metamorphic and igneous rocks into tectonic context (Wynne-Edwards 1972, Rivers et al. 1989, Easton 1992, Carr et al. 2000, Hanmer et al. 2000, Culshaw et al. 2006, Rivers 2008, Jamieson et al. 2010; McLelland et al. 2013). In a recent summary of Grenvillian tectonics, Rivers (2015) argued that the Grenville Province was a long-lived, accretionary margin at the edge of Laurentia during the Mesoproterozoic prior to continental collision during the Grenville Orogeny. Two late, relatively well-defined accretionary events were the Elzeviran Orogeny (1245-1225 Ma), associated with opening and closure of a back-arc basin and the amalgamation of terranes in the Composite Arc Belt (Figure 1a), and the Shawinigan Orogeny (1190-1140 Ma) in the Frontenac-Adirondack and Parry Sound regions (Figure 1b). The active margin culminated in the collisional Grenvillian Orogeny (1090-980 Ma) within the supercontinent Rodinia. The Grenvillian Orogeny can be subdivided into two phases (Rivers et al. 2012, Rivers 2015). The Ottawa Phase (1090-1020 Ma) was a long hot orogenic event during which the Grenville Province accommodated thousands of kilometers of shortening in the hanging wall of the Allochthon Boundary Thrust (ABT in Figure 1b), doubling the thickness of the continental crust (Jamieson et al. 2010). Eventual late-Ottawan orogenic collapse was probably associated with channel flow in the mid-crust at the base of an orogenic lid (Rivers 2012). During the Rigolet Phase (1005-980 Ma), shortening migrated to the foreland at the Grenville Front (Figure 1a).

Our study area is a major Grenvillian shear zone: the Central Metasedimentary Belt boundary zone (CMBbz; Figure 1a), a boundary recognized by Wynne-Edwards (1972) and later workers as a regionally significant shear zone. The CMBbz is characterized by a southeast-dipping foliation (Hanmer and McEachern 1992), and geophysical data show that the zone is a crustal-scale structure that dips moderately to the southeast at depth (White et al. 2000). At least four geochronological studies have presented data from this shear zone, and each study suggests that deformation and metamorphism occurred at a different time. Compounding this uncertainty, shear zone kinematics are ambiguous. As a result, the different estimates of timing imply different interpretations of the tectonic significance of the CMBbz: from pre-Shawinigan ductile thrusting at the active margin of Laurentia to late-Ottawan extension at the base of the orogenic lid built during continental collision. The existing data suggest that the CMBbz experienced a polyphase tectonic and metamorphic history. Here we present U-Th-Pb monazite dates from garnet-biotite gneisses from three localities in the CMBbz. We directly date deformation and metamorphism using the approach of Williams et al. (2007 and 2017), an approach that is well suited to assessing complex histories in metamorphic rocks that have experienced multiple deformation events. Our two research questions are: How many episodes of shear zone activity did the CMBbz experience, and what is the tectonic significance of each episode within the context of the overall Grenvillian tectonics?

Tectonic Significance of the Central Metasedimentary Belt boundary zone (CMBbz)

The CMBbz (Figure 2) is a relatively wide (25-50 km) and long (200 km) amphibolite-facies shear zone that includes map-scale, lenticular, meta-plutonic thrust sheets interlayered with well-foliated (i.e. straight) gneisses, meta-volcanic rocks, and marble breccia, all intruded

by a variety of igneous rocks of vastly different scales and compositions (Hanmer & McEachern 1992). The CMBbz dips to the southeast and contains a ubiquitous concordant foliation that varies in grade and intensity. Stretching lineations are much less common and vary in orientation; kinematic indicators are equally uncommon and ambiguous. Hanmer and McEachern (1992) distinguished a package of aluminous gneisses (containing orthoamphibole-bearing rocks) at the bases of many of the thrust sheets. They hypothesized that the rheology of these rocks may have contributed to the localization of shear within the CMBbz during ductile thrusting towards the northwest. Unlike the Allochthon Boundary Thrust (Figure 1b) and the Grenville Front to the northwest (Figure 1a), the CMBbz lacks an inversion of metamorphic grade (Rivers et al. 1989, Anovitz and Essene 1990; Easton 1992; Davidson 1998). And unlike the Robertson Lake Shear Zone and Carthage-Colton Shear Zone to the southeast (Figure 1b), the CMBbz also lacks sharp discontinuities in metamorphic grade. From the footwall (Muskoka terrane) to the hanging-wall (Hastings Low of the Elzevir terrane), the metamorphic grade decreases from 750°C and >8kbar to <500°C and <6 kbar (Anovitz and Essene 1990; Streepey et al. 1997; Dunn et al. 2016). Because there is no consensus in the literature on the CMBbz's name or exact boundaries, we follow Hanmer and McEachern's (1992) delineation, which includes some of the Central Gneiss Belt and Bancroft Terrane of Easton (1992).

The footwall of the CMBbz is the Gneiss Belt (for brevity, we drop the word Central from the traditional terminology), composed largely of the meta-granitoids associated with the pre-Ottawan active continental margin of Laurentia. Protolith ages >1450 Ma are common throughout the Gneiss Belt in Ontario (Carr et al. 2000). The hanging wall of the CMBbz is the Composite Arc Belt of the Metasedimentary Belt, made up of meta-igneous and meta-sedimentary rocks associated with pre-Shawinigan island arcs and basins formed on oceanic

crust. These meta-igneous rocks show younger protolith ages <1300 Ma (Lumbers et al. 1990), and deposition of the protoliths to the meta-sedimentary rocks probably occurred 1300-1245 Ma in shallow marine environments (Easton, 1992; Carr et al. 2000). The base of the CMBbz hanging wall is defined by a string of Elzeviran gabbro bodies at the edge of the Elzevir Terrane (Fig. 2; Hanmer and McEachern 1992). Many workers have considered the timing of intrusion of these gabbro bodies (U-Pb zircon ages of 1250-1230 Ma reported by Pehrsson et al. 1996) as an upper age limit for metamorphism and deformation associated with ductile thrusting in the CMBbz (Hanmer et al., 2000; McLelland et al., 2010). Therefore, the first phases of deformation and metamorphism in the CMBbz probably related broadly to accretion at the edge of Laurentia and likely included the late stages of the Elzeviran Orogeny. In addition, the CMBbz probably accommodated both Shawinigan thrusting of the Composite Arc Belt (i.e. Metasedimentary Belt) onto the ancient Laurentian margin (i.e. Gneiss Belt) and extensive reactivation during Ottawa convergence and later collapse (Rivers et al. 2012, McLelland et al. 2013).

Detailed geochronological work on deformation and metamorphism within the CMBbz supports this long, polyphase tectonic history. From U-Pb dates of zircon in syntectonic pegmatites along the length of the Ontario CMBbz, McEachern and van Breemen (1993) argued that ductile thrusting began by 1190 Ma (early Shawinigan), with a major pulse of deformation in a relatively short interval of the Ottawa orogeny from 1080-1060 Ma, followed by waning deformation in the shear zone that persisted until 1000 Ma (Rigolet). Hanmer and McEachern (1992, referencing the data published in 1993) speculated that the earlier event may have been the tectonically more significant one. But Timmermann et al. (1997) argued against the earlier event based on the early Ottawa timing of syntectonic igneous rocks in the direct footwall of the CMBbz (U-Pb zircon dates of 1080-1060 Ma from the Muskoka terrane; Figure 1b). Mezger et

al. (1993) reported U-Pb titanite dates from the CMBbz: igneous pegmatites, skarns, and syenite dikes gave U-Pb titanite ages in the range 1075-1045 Ma, and “sphenes of unambiguously metamorphic origin” yielded late Ottawaan U-Pb dates of 1045-1030 Ma. Interpreting these U-Pb titanite dates as records of mineral growth, Mezger et al. (1993) argued that the CMBbz, and the rest of the western Metasedimentary Belt, underwent only one significant late Ottawaan metamorphic/deformation event 1045-1030 Ma, with one small region in the northern CMBbz having experienced an earlier event. Although Mezger et al.’s (1993) assessment of the tectonic history of the CMBbz differs significantly from McEachern and van Breemen’s (1993), both studies reported dates older than 1080 Ma in the same northern region of the CMBbz. A recent compilation of geochronological data from this northern region of the CMBbz (Duguet and Easton 2017) suggests that Shawinigan metamorphic and igneous dates there are widespread. Agustsson et al. (2013) reported metamorphic U-Pb zircon dates of 1150-1100 Ma from the ca. 1340 Ma Dysart complex (Figure 2), one of Hanmer’s (1988, 1989) original thrust sheets. They interpreted these dates to represent a significant episode of intrusion and metamorphism from 1150-1100 Ma in the heart of the central CMBbz. They also reported one early Ottawaan metamorphic date from this part of the central CMBbz. In summary, these geochronological studies disagree about the timing of specific metamorphic events in the CMBbz, but taken as a whole, they suggest that the CMBbz may have been active throughout the Shawinigan Orogeny and the Ottawaan phase of the Grenvillian Orogeny. A lower age limit for deformation in the CMBbz is likely late Ottawaan, from the youngest rocks of the 1090-1030 Ma late syenite suite intruding the CMBbz along its length, from Minden to the Ottawa River (Easton, 2013).

A different approach to understanding metamorphic gradients and dates in the Ottawaan phase of orogeny is the Ottawaan Orogenic Lid model (Rivers 2012). This model is based in part

on hornblende-cooling-age data reported by many workers, in particular Cosca et al. (1991 and 1992). Rivers (2012) argued that relatively old (i.e. >1020 Ma) $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende cooling ages mark the remnants of a late Ottawa orogenic plateau (thermo-mechanically speaking). These old $^{40}\text{Ar}/^{39}\text{Ar}$ ages occur in the hanging wall of the CMBbz (i.e. the Hastings Low of the Elzevir terrane, near Madoc) and further east in the Frontenac terrane and Adirondack Lowlands (Figure 1b). In the Rivers (2012) model, the CMBbz was near or at the boundary of the ductile mid-crust beneath a rigid orogenic plateau that developed during the peak Ottawa and later phases of the orogeny. Rivers and Schwerdtner (2015) and Schwerdtner et al. (2016) recently extended this model to suggest that the Gneiss Belt in the CMBbz footwall was essentially a late Ottawa metamorphic core-complex-type shear zone with transtensional kinematics.

In summary, the CMBbz is made up of meta-sedimentary and meta-igneous rocks characterized by a southeast-dipping foliation and amphibolite facies metamorphic assemblages. Protolith ages for meta-sedimentary and some meta-igneous rocks in the CMBbz are 1350-1250 Ma. Most of the deformation and metamorphism in the CMBbz is bracketed between 1250-1030 Ma (i.e. including the Elzevirian and Shawinigan Orogenies and the Ottawa phase of the Grenvillian Orogeny), but a finer scale of agreement on the tectonic significance and chronological continuity or discontinuity of orogenic events in the CMBbz is currently elusive in the literature. Near the end of this time span, during the final stages of the Ottawa orogeny, the CMBbz likely occupied a position near the bottom of a rigid orogenic plateau.

Samples

We present monazite data from five samples collected from three localities of the CMBbz: West Guilford, Fishtail Lake, and Killaloe (black squares in Figure 2; the two Killaloe

samples are from the same outcrop and appear as a single black square). Table 1 summarizes sample locations, all of which are at the base of thrust sheets within the CMBbz (Hanmer and McEachern 1992). All of the samples are tectonites that show shallowly to moderately east-dipping foliation that strikes parallel to the local mapped boundary of the CMBbz (Figure 2). All five samples are garnet-biotite gneiss with foliation defined by compositional layering and oriented biotite grains (Figure 3). Major minerals are quartz, K-feldspar, plagioclase feldspar, biotite, and garnet. Garnet grains are mm- to cm-scale, equant, and subhedral to anhedral; they show inclusions but do not appear to preserve an older fabric. Garnets in two samples (HA13 and 11AM-1) show rims that are relatively lower in calcium than their cores, but garnets in the other samples are unzoned. Quartz and feldspar grains are also mm- to cm-scale and equant, with straight to curved grain boundaries, and quartz shows undulose extinction. Sample HA13 contains sillimanite and is the only sample that displays a lineation. Some samples are from lithological units that Hanmer and McEachern (1992) referred to as “cordierite-gedrite and sillimanite-garnet gneisses” at the base of map-scale crystalline thrust sheets within the CMBbz. In a Supplemental Document, we include more detailed descriptions of each sample and give more information about how other workers have mapped each locality.

Monazite U-Th-Pb Geochronology

The overall goal of this work is to constrain the age of deformation and metamorphism using in-situ electron microprobe compositional mapping and chemical dating. Like zircon, monazite has a high closure temperature for lead and actinide series elements. Therefore U-Th-Pb dates from monazite that grew during amphibolite facies metamorphism are likely mineral growth ages rather than cooling ages (Cherniak et al. 2004, Cherniak and Pyle 2008, Cherniak

2010), although fluid-mediated dissolution-reprecipitation in the presence of aqueous fluids at even low temperatures can reset monazite dates (Harlov et al. 2011; Williams et al. 2011). Our approach follows that of Williams et al. (2007 and 2017); we use in situ analysis to link U-Th-Pb monazite ages to specific deformation and metamorphic events inferred from petrographic analysis.

Methods Analyses were carried out at the University of Massachusetts (Amherst) Department of Geosciences Electron Microprobe/SEM Facility following the specific strategies outlined by Jercinovic and Williams (2005), Williams et al. (2006; 2017), and Jercinovic et al. (2008). We cut samples perpendicular to foliation. Polished thin sections were commercially prepared from cut samples and then coated with 20 nm of aluminum followed by 8 nm of carbon (Jercinovic et al. 2008). We identified monazite and conducted basic textural and compositional mapping of thin sections using a CAMECA SX-50 microprobe for wavelength-dispersive spectroscopy mapping of full thin sections for major elements (Mg and Ca) and trace elements associated with monazite and other actinide-rich minerals (Ce or La, Y, and Zr) (Williams et al. 2017). After identifying grains of monazite suitable for dating, we used the SX-50 to obtain higher-resolution compositional maps of the individual grains, focusing on elements likely to distinguish compositional domains within monazite grains (Y, Th, U, Ca, and Nd or Si). Most of the samples we discuss here contain additional grains of monazite that we did not map or attempt to date because our approach favors detailed analysis of selected grains with well characterized textural and compositional context (Williams et al. 2007, 2017).

The CAMECA “Ultrachron” microprobe was used for quantitative monazite analyses. The typical analytical protocol (see Williams et al. 2007) includes U, K, Th, Ca, S, P, Y, Sr, Si, Pb, La, Ce, Nd, Pr, Eu, Sm, Tb, Dy, Er, Gd, Tm, Yb, Ho, and As. We collected data using Probe

for EPMA (Probe Software Inc.), in tandem with CAMECA's Peaksite automation. Count times for individual elements followed Williams et al. (2017). We ran Brazilian Moacyr monazite (ca. 506 Ma) as a consistency standard before and after each day-long working session, and these results suggest that our data are extremely consistent (± 2 Ma). Background intensities for U, Th, and Pb were calculated using the multi-point method of Allaz et al. (2013). We used one background analysis for each compositional domain. Then, multiple peak measurements (typically 5 -8) were made near the background point in order to achieve the minimum uncertainty (Williams et al. 2017). We present a single weighted mean "date" for each compositional domain. All errors on dates are reported at the 2σ level; we calculated these errors by propagating the counting error and the background error through the age equation for each peak measurement and calculating the weighted mean of the peak estimates using a student's *t* distribution to account for small sample populations. One disadvantage of the EMPA chemical dating method is that we are not able to evaluate inherited (common) lead, and dates assume that all lead is the daughter of uranium and thorium decay. We present these dates along with complete compositional data for 62 monazite domains (Table 2). We gathered all the data here during three different working sessions - February 2015 (the BY13 pair), July 2016 (11AM-1), and June 2017 (11AM-20 and HA13).

West Guilford Samples Monazite grains in both West Guilford samples show multiple core domains. In sample HA13, some extremely high Y domains are in the inner cores of grains, and these high Y internal domains yield dates that are the oldest in the sample, ca. 1095 Ma (Table 2). Many grains also show outer cores that are extremely low in Y, and these domains yield younger ages, ca. 1080 Ma (Table 2). Although the geometric relationships between various core domains in sample 11AM-20 (Figure 3A) are more ambiguous with respect

to their relative ages (Figure 4), dates suggest that the higher Y core domains are older than the lower Y core domains (Figure 5; Table 2). For both samples, grains that are enclosed within garnet (“gt” in Table 2) also yield old dates. Monazite grains in both samples also show relatively Y-rich rims that yield young ages, ca. 1045 Ma (Figures 4 and 5; Table 2). In sample HA13, Y-rich rims are quite thin and show no geometric variation with orientation or location. But in sample 11AM-20, Y-rich rims are generally thicker along the grain edges oriented as pressure shadows with respect to the foliation (Figure 4).

Fishtail Lake Sample

The most distinctive feature of monazite in the Fishtail Lake sample (11AM-1; Figure 3B) is inner core domains that are unusually rich in Nd, poor in Y, and yield ca. 1080 Ma dates (Figure 6). Outer core domains are relatively high in Th and low in Nd, and the dates from outer core domains do not clearly define a single population. Monazite grains lack consistent rims. Two grains are partly enclosed in garnet (Figure 7), and two young dates come from relatively high Y domains in these grains (Figure 6). The three other grains are oriented with their longest axes parallel to foliation. Tentatively, we group the five older domain dates (ca. 1080 Ma) together based on their very low Y content, even though this grouping ignores the stark difference in Nd content, and we group the three younger domains (ca. 1050 Ma) together based on higher Y content, even though this grouping ignores the stark difference in settings (Table 2).

Killaloe Samples

In sample BY13-2 (Figure 3C), the geometries of different core domains are geometrically complex and not resolvable into inner and outer cores, but seven high Y core domains yield older ages ca. 1220-1180 Ma, and three low Y core domains yield younger ages ca. 1060 Ma (Figure 8; Table 2). Sample BY13-1 yields similar results, but also shows some low Y core domains that yield older dates 1180-1160 Ma. For sample BY13-1, grains that

are enclosed in garnet yield dates that fall in the oldest high Y group (Table 2). Taken together, these old dates show a surprisingly steady decrease in Y during the time span 1220-1160 Ma (Figure 9). The youngest dates for this sample pair are from the distinctive high Y rims on most grains in sample BY13-2 (Figures 8 and 9; Table 2). Slightly older than these high Y rims are some lower Y rims and unzoned grains in BY13-1 and a few low Y outer cores in BY13-2, all ca. 1060 Ma (Figure 9). For sample BY13-2, grains that are enclosed in garnet yield surprising dates. One grain that is partly enclosed in garnet and another grain that appears fully enclosed in garnet yield dates from high Y rims that fall in the youngest group (Figure 10; Table 2). Both of these grains occur near the same fracture in the same garnet grain (Figures 3C and 10).

Discussion: Monazite Petrochronology

Monazite composition is influenced by other minerals that are growing or dissolving during monazite growth. Garnet is particularly important in regulating the yttrium content of monazite. Garnet's affinity for yttrium and its significant volume in many rocks make it one of the most important yttrium reservoirs in garnet-bearing rocks. Monazite that grows along with garnet will be relatively low in yttrium, whereas monazite that grows during garnet dissolution is relatively yttrium rich (Pyle and Spear, 1999; Spear and Pyle, 2002; Kohn et al., 2005; Williams et al. 2007; 2017). Garnet is not the only mineral that can influence monazite yttrium content. Xenotime, apatite, and allanite are common minerals that may also play a role (Pyle et al., 2001; Spear and Pyle, 2002), but garnet is so clearly linked to monazite yttrium content that the history of garnet growth and consumption is considered to be the key control on yttrium in monazite.

In garnet-bearing gneisses like our samples, a simple (i.e. fictional) model metamorphic history might involve an early generation of Y-rich monazite associated with the protolith or

prograde metamorphic history, followed by the deformation-assisted growth of garnet, associated with a second generation of Y-poor monazite, followed by partial dissolution of garnet during retrogression or decompression associated with a late generation of Y-rich monazite. In this simple model, a garnet-bearing rock might have monazite grains with pre-garnet Y-rich inner cores, post- or syn-garnet Y-poor outer cores, and Y-rich rims. Monazite grains enclosed by garnet might be expected to record only the early events (Spear and Pyle, 2002; Kohn et al., 2005; Mahan et al., 2006; Williams et al., 2017).

Each of our garnet-biotite gneiss samples shows a relatively simple fabric that appears, at first glance, to record a single deformation event associated with the development of a strong foliation during amphibolite-grade metamorphism. The lack of compositional zoning in most garnet (the exceptions are HA13 and 11AM-1), in combination with a lack of older fabric preserved as inclusions, supports this apparent simplicity. Weak undulose extinction in quartz grains is slight evidence for late-stage strain, but this texture is typical of gneiss and even granite. In other words, the metamorphic mineral assemblages in these samples show little evidence for polymetamorphism or preservation of prograde (or retrograde) textures. The polymetamorphic histories of these rocks are largely cryptic but are preserved in monazite grain populations.

Monazite in the West Guilford samples (11AM-20 and HA13) records a relatively simple progression similar to that presented above. Cores do not show clear inner and outer domains in a geometric sense, but Y-rich core domains are generally older than Y-poor core domains (ca. 1095 Ma and ca. 1080 Ma, respectively), and grains enclosed in garnet yield only these early dates. The transition from older Y-rich core domains to younger Y-poor core domains may record prograde metamorphism, and we associate peak Ottawa metamorphism and garnet growth with low-Y monazite core domains that yield ca. 1080 Ma dates. Monazite grains in both

West Guilford samples show relatively Y-rich rims that yield young ages, ca. 1045 Ma. In sample 11AM-20, Y-rich rims are generally thicker along the grain edges oriented as pressure shadows with respect to the foliation. In other words, the dates, compositions, and geometries of monazite rims suggest that they record late Ottawaan garnet breakdown during retrogression and/or decompression associated with reactivation of foliation.

Monazite inclusions in garnet deserve particular discussion. Intracrystalline diffusion in garnet is so limited that it is very unlikely that monazite could grow or dissolve while enclosed within garnet (Spear and Pyle, 2002). Thus, one would expect that the youngest age obtained on a monazite inclusion would represent the maximum age of garnet growth. The monazite grain in West Guilford sample 11AM-20 that is wholly enclosed within garnet (Figures 3A and 4) is straightforward to interpret. This monazite grain yields two ages that overlap within error, 1082 ± 11 and 1096 ± 7 Ma, and is low in Y relative to monazite in the matrix. Two monazite inclusions in garnet have been analyzed from the other West Guilford sample, HA13, and both of these grains yield domain ages similar to garnet inclusions in 11AM-20. Low Y cores of matrix monazite are similar in age to these inclusions and suggest that the maximum age of garnet growth for these samples is ca. 1080 Ma, during the peak of Ottawaan orogenesis. Higher Y rims of matrix monazite are younger (ca. 1045 Ma). The anhedral nature of garnet in this sample suggests some garnet consumption took place after the period of maximum garnet growth. Relatively high Y in the rims of matrix monazite could be the result of Y release during garnet consumption. In summary, monazite dates from the West Guilford samples record three distinct episodes of metamorphism: a very early Ottawaan stage ca. 1095 Ma, a peak Ottawaan stage ca. 1080 Ma likely associated with garnet growth, and a late Ottawaan retrogression ca. 1045 Ma.

In the Fishtail Lake sample (11AM-1), monazite core domains that are unusually rich in Nd and poor in Y yield ca. 1080 Ma dates, apparently peak Ottawa dates that agree within error with dates from some other low Nd domains in this sample. One of these dates comes from a grain that is partly enclosed in garnet (Figure 7). The youngest dates in this sample (ca. 1050 Ma) come from higher Y domains: the outer core of a matrix grain and two grains that are partly enclosed in garnet (Figures 6 and 7). These dates are mid to late Ottawa. Unlike the samples from West Guilford, this sample from Fishtail lake bears monazite without consistent rims and yields young dates from monazite grains that are partly enclosed in garnet.

Despite the relation between dates and Y content in monazite in the Fishtail Lake sample (i.e. early Ottawa dates from low Y domains and later Ottawa dates from high Y domains), a definitive interpretation of these dates in the context of garnet growth is elusive. Although the lower-Ca rims of garnets are likely evidence for late retrogression (or decompression), the timing of both garnet growth and retrogression probably occurred after ca. 1050 Ma and was not marked by growth of high-Y rims on monazite grains. An alternative interpretation is that garnet growth occurred ca. 1080 Ma and coincided with the growth of low Y monazite cores; late retrogression occurred ca. 1050 Ma and coincided with the growth of high Y domains with unusual geometries and locations. This alternative interpretation is not fully plausible because two monazite grains that yield the young ages are partly enclosed in garnet. We note however, that both these garnet-enclosed monazite grains occur in locations that would have been favorably oriented as pressure shadows (at the edges of the garnet grains) with respect to the foliation. In addition, the orientation of the long axes of matrix grains further suggests that some deformation of this sample may have occurred after the first period of monazite growth. In summary, monazite dates

from the Fishtail Lake sample record two distinct Ottawa episodes of monazite growth, but a definitive petrochronological interpretation of these dates remains uncertain.

Monazite in the two Killaloe samples yields a history that begins well before Ottawa time. The oldest dates from core domains are 1220 -1160 Ma and show a surprisingly steady decrease in Y with age, perhaps related to a protracted prograde metamorphic event. This time span is long and starts early, but it essentially corresponds with the Shawinigan orogeny. Our tentative arguments about the prograde paths of these samples rest on the age relations between core domains that show stark contrasts in Y content. In other words, none of the monazite in these samples shows clear inner and outer cores. In both West Guilford samples, however, and also in the Killaloe sample that yields the oldest dates in this study (BY13-2), high Y core domains are consistently older than low Y core domains, even though the geometric relations between the domains are ambiguous. We interpret old dates from high Y domains as a record of early steps along the prograde paths of samples, when monazite may have been forming during the dissolution of other high Y phases (such as xenotime) in the absence of metamorphic garnet. Low Y domains in monazite grain cores likely relate to the timing of peak metamorphism after garnet or other high Y phases had begun to grow and incorporate yttrium.

Like the Fishtail Lake sample, the Killaloe samples yield an ambiguous Ottawa history. Some lower-Y rims, unzoned grains, and a few low-Y outer core domains may represent an Ottawa event ca. 1060 Ma. The youngest dates are from high-Y rims that appear to represent a well-defined Ottawa event ca. 1045 Ma. The result that two of these young dates come from monazites enclosed in garnet (one partly, and one fully) poses a significant challenge to interpreting these late Ottawa dates in terms of retrogression (or decompression) associated with dissolution of garnet. Some garnet growth appears to post-date the growth of all the

monazite domains and grains in these samples. The lack of prograde zoning in garnet in the Killaloe samples further suggests that the temperatures attained were sufficient to homogenize garnet interiors by intracrystalline diffusion. This event occurred after ca. 1045 Ma, and the complex suite of monazite dates from these Killaloe samples is untethered to a metamorphic history involving the garnet currently present in these samples, perhaps involving metamorphic reactions with an older generation of garnet or with other Y-bearing phases such as xenotime.

An alternative explanation is that the youngest dates from the Killaloe samples are associated with a late fluid-mediated dissolution-reprecipitation event rather than a metamorphic event involving mineral reactions. Although experimental and empirical studies have shown that volume diffusion is not an effective mechanism for modification of monazite composition at amphibolite facies conditions (e.g., Spear and Pyle 2002; Cherniak et al. 2004), dissolution-reprecipitation in the presence of aqueous fluids at even low temperatures can result in the development of rims and overgrowths as well as complex, irregular domains of varying composition in monazite (Harlov et al. 2011; Williams et al. 2011). Patchy, multi-domain monazite grains were produced in experimental studies by Williams et al. (2011). Their results showed that fluid-mediated dissolution-reprecipitation can reset monazite dates to zero in altered domains while leaving original dates intact in unaltered regions (see also Harlov et al. 2011). Irregularly zoned monazite in rocks from many terranes has yielded apparently accurate age determinations by SIMS U-Pb and U-Th-Pb dating methods (Spear and Pyle 2002; Kohn et al. 2005; Mahan et al. 2006; Wawrzenitz et al. 2012; Williams et al. 2017). Because monazite reset by this mechanism commonly shows very low concentrations of Th (Harlov et al. 2011; Williams et al. 2011), we are hesitant to invoke this mechanism to explain the youngest dates from the Killaloe samples because they come from monazite domains that show relatively high

Th concentrations, similar to the other older domains (Table 2). But a fluid-mediated dissolution-reprecipitation event could be responsible for altering monazite chemistry and resetting monazite dates in grains that are partly enclosed or within fractures in garnet. If this alternative interpretation is correct, then garnet in the Killaloe samples may have grown during either peak Shawinigan (ca. 1180 Ma low Y monazite core domains) or peak Ottawa (ca. 1060 Ma low Y domains) metamorphism, and the late Ottawa event ca. 1045 Ma may record fluid infiltration.

Discussion: Tectonic Implications

Both the Killaloe and West Guilford samples yield dates that appear to record prograde metamorphic paths. For the Killaloe region, we argued (above) that the dates 1220-1160 Ma, and the associated decrease in Y concentration of monazite domains with time (Figures 9 and 11), record the timing of prograde metamorphism. We infer that prograde metamorphism peaked at about 1180 Ma, consistent with the timing of peak Shawinigan tectonism (McLelland et al. 2010). This estimate agrees with the U/Pb dates of syntectonic pegmatites (McEachern and van Breemen 1993) and the interpretation that prograde metamorphism and crustal thickening in the CMBbz was well under way at this time (Hanmer and McEachern 1992). Our evidence for a record of Shawinigan events in this part of the CMBbz is also consistent with observations that the northern Bancroft Domain (i.e. the CMBbz near Killaloe) preserves a pre-Ottawan history (Mezger et al. 1993; Duguet and Easton 2017). These oldest monazite dates in our study, ca. 1220 Ma high Y core domains from Killaloe sample BY13-2, are younger than likely protoliths to the Killaloe samples, and we do not interpret them as inherited ages. Instead, we favor an interpretation of these ages as recording crustal thickening at the paleo-margin of Laurentia and

maybe even local terrane assembly because the dates we present here are consistent with the well-established Elzeviran and Shawinigan timing of these events (Carr et al. 2000; Rivers 2015).

In the West Guilford region, we argued that dates 1095-1080 Ma and the associated decrease in Y concentration of monazite domains with time also represent prograde metamorphism (Figures 5 and 11). We interpret these dates as representing a second prograde event in the CMBbz, the ca. 1080 Ma peak metamorphism and crustal thickening associated with Ottawa orogenesis (McLelland et al. 2010). Both of the other localities (i.e. Fishtail Lake and Killaloe) yield low Y monazite domains with Ottawa ages (Figure 11). We infer that Ottawa ages from low Y monazite domains in samples from all three localities suggest that the younger Ottawa metamorphic record is preserved throughout the CMBbz.

Why is the older Shawinigan metamorphic record preserved only in the Killaloe region? One possible explanation is that the Fishtail Lake and West Guilford regions did not experience the older Shawinigan event. It is interesting to note that the Elzevir domain in the hanging wall of the CMBbz contains Flinton Group sediments that were apparently deposited during the time interval between the Shawinigan and Ottawa orogenies, i.e. after 1150 Ma but likely before 1080 Ma (Sager-Kinsman and Parrish 1993). This apparently late timing of sedimentation nearby suggests that perhaps some parts of the CMBbz were not juxtaposed until Ottawa time. If so, the rocks sampled from the Killaloe locale may have experienced prograde metamorphism and burial during the Shawinigan orogeny and then remained deep in the orogen during Ottawa events, but the rocks sampled from the West Guilford and Fishtail Lake locales may have experienced prograde metamorphism only during the Ottawa Orogeny.

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Alternatively, we think it is more likely that the record of the older Shawinigan event is simply not preserved in samples with extensive records of the younger Ottawa event. Certainly, all of the samples in this study likely have protoliths that predate the Shawinigan orogeny, yet none of the samples yields pre-Shawinigan dates. Specifically, the Killaloe samples are adjacent to meta-igneous rocks of the Gneiss Belt (Lumbers, 1982) whose protoliths generally yield ages >1.4 Ga, yet the oldest U-Th-Pb monazite dates we obtained from these Killaloe gneisses are ca. 1.2 Ga. The West Guilford samples are mapped directly in the footwall of the Dysart thrust sheet, and the Dysart Gneiss Complex has recently yielded igneous ages of 1.35-1.30 Ga from U/Pb zircon dates from granodiorites and metamorphic ages of 1.15-1.10 Ga from tonalites (Agustsson et al. 2013). Yet the oldest U-Th-Pb monazite dates we present here from the West Guilford gneisses overlap only with the youngest ages reported by Agustsson et al. (2013). This disconnect between our data and U/Pb zircon protolith ages even holds true for the Fishtail Lake sample, which is related to a suite studied in much more detail by Peck and Smith (2005). U/Pb zircon dates from this suite suggest that volcanic protolith ages for these gneisses are ca. 1.3 Ga (W.H. Peck, unpublished data, 2011), but the oldest U-Th-Pb monazite dates reported here are ca. 1.1 Ga, consistent with the youngest U/Pb zircon dates (from texturally metamorphic zircon) in the unpublished U/Pb zircon data. These considerations lead us again to the conclusion that all of our samples have earlier tectonic histories than are recorded by monazite.

All three CMBbz sample regions share youngest ages of ca. 1045 Ma. For the West Guilford and Killaloe regions, these youngest ages are best represented in Y-rich rims on monazite grains. In the West Guilford samples, a petrochronological interpretation is straightforward, and the high-Y composition of the youngest monazite population is likely associated with monazite growth during partial dissolution of garnet associated with

retrogression. But in the Killaloe region, a petrochronological interpretation suggests that garnet growth occurred after this youngest generation of monazite rims, or that dates from monazite rims are a record of late fluid infiltration, or both. In either case, the ca. 1045 Ma timing, in association with the inferences of fluid-mediated resetting and retrogression, supports a link to the late extension and thinning in Rivers' (2012) Ottawa Orogenic Lid model. Sample 11AM-20 (from West Guilford) provides a subtle window into the nature of this youngest event; young high-Y rims on monazite grains in this sample are thickest in directions parallel to foliation, consistent with a late, retrogressive (or decompression-related) reactivation of foliation during crustal thinning. Whether or not the Fishtail Lake sample records this late event is less clear. The highest Y monazite domains yield similar dates ca. 1050 Ma, but the relations between domain geometry and petrography are ambiguous.

If our interpretation of the ca. 1045 Ma dates is correct, monazite grains in these garnet-biotite gneisses record late Ottawa extension. It is interesting to note that many workers have associated late Ottawa extension in this region only with finer-grained mylonitic shear zones and straight gneisses, including CMBbz straight gneiss (Hanmer and McEachern 1992; Schwerdtner and Rivers 2014), marble from the extensional Bancroft Shear Zone (Carlson et al. 1990), and orthogneiss and amphibolite from the Salerno Creek Deformation Zone (Easton and Kamo 2011). Because there are no precise geochronological constraints on the timing of mylonitization in these late and relatively narrow cross-cutting shear zones, it is impossible to compare their timing directly with the youngest monazite dates that we present here from relatively coarser-grained rocks in the CMBbz, which is a much wider shear zone.

Conclusions

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Our two research questions were: How many episodes of shear zone activity did the CMBbz experience, and what is the tectonic significance of each episode? We have presented U-Th-Pb monazite dates from garnet-biotite gneiss samples collected from three localities of the CMBbz: West Guilford, Fishtail Lake, and Killaloe. All three localities yield monazite dates ca. 1050-1045 Ma (Figure 11). In the West Guilford region, the monazite domains that yield these dates show compositional and geometric evidence that they are associated with late retrogressive (or decompression-related) metamorphism. In the Killaloe region, these dates are possibly associated with late fluid infiltration. Chronologically these dates are associated with crustal thinning and extensional shear zone activity at the base of the late Ottawa Orogenic Lid (Rivers 2012), and this timing agrees with Mezger et al.'s (1993) estimate of peak metamorphism based on U/Pb dates on titanite from metamorphic rocks in the CMBbz. Some of Mezger et al.'s (1993) dates from the region, especially those from the adjacent Bancroft Shear Zone, are even younger than ours and may represent cooling ages, similar to even younger $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende cooling ages reported by Cosca et al. (1991, 1992).

In comparison to the common late Ottawa history that we document, the earlier history of the three CMBbz regions is more complex. Monazite domains in the Fishtail Lake sample are ambiguous in terms of compositions and geometries, and we are hesitant to offer a decisive interpretation of the earlier Ottawa (i.e. 1085 Ma) dates that this sample yields. The West Guilford samples yield clearer evidence that Ottawa dates in the same range record a prograde history corresponding with garnet growth. The cores of monazite in these samples show older high Y domains and younger low Y domains, and dates from a monazite grain included in garnet agree with this interpretation. The Killaloe samples yield some Ottawa ages, but a much more likely Shawinigan prograde history ca. 1220-1160 Ma, based on monazite core domain

compositions. For the Killaloe samples, the relation of monazite domain dates, compositions, and geometries to the growth history of garnet are difficult to infer. The inclusion of young monazite in garnet suggests that at least some garnet growth occurred after ca. 1045 Ma. If this explanation is correct, then garnet growth probably occurred during a significant episode of deformation and metamorphism that post-dates both the Shawinigan event and the Ottawa event that we infer from monazite dates. Late resetting of monazite rims associated with fluid infiltration in both matrix monazite and monazite in garnet fractures may provide an alternative scenario in which garnet growth occurred during prograde metamorphism associated with either the Shawinigan or Ottawa events we document.

The dates we present here are consistent with previous geochronology in the CMBbz. Two U/Pb zircon studies that were specifically conducted to assess the timing of deformation in the CMBbz (McEachern and van Breemen 1993; Timmerman et al. 1998) emphasized the importance of crustal thickening and shear zone activity during peak Ottawa and Shawinigan time. But these studies disagreed in their interpretations of the relative significance of the two events in the CMBbz. Our results do not settle this disagreement but are consistent with a conciliatory interpretation that the CMBbz in Ontario experienced metamorphism during both orogenies. Our results do clearly indicate that the apparently simple foliations in the CMBbz are actually composite structures that formed initially in older rocks and were then reworked during several periods of deformation and syntectonic mineral growth, giving rise to multiple generations of monazite growth.

Finally, we note that we do not report evidence for preservation of detrital populations of monazite. Nor do we report evidence for significant shear zone activity or metamorphism during the interval 1160-1100 Ma, between the relatively well established chronological limits of

Shawinigan and Ottawa events that are recorded everywhere from the Adirondack Mountains to the Grenville Front (Rivers 2015, McLelland et al. 2013). As a result, we do not find any evidence to support the interpretations of U/Pb zircon and titanite dates in this range that suggest that metamorphism may have occurred throughout the CMBbz during this interval (Mezger et al. 1993, Agustsson et al. 2013). In summary, monazite U-Th-Pb geochronology records three events: a prograde Shawinigan event, a prograde Ottawa event, and a late Ottawa event associated with crustal thinning at the base of an orogenic lid.

Acknowledgements

Thanks to the undergraduate geology students who worked with us in the field. From the 2013 Mount Holyoke College group: Michaela Kim, Maneh Kotikian, Dana Reuter, Katherine Rudzki, and Camilla Yohn-Barr. From the 2011-2012 Keck Geology Consortium project: Kenjo Agustsson, Naomi Barshi, Neva Fowler-Gerace, Calvin Mako, Edward Marshall, Abigail Monreal, Bo Montanye, Jacquelyne Nesbit, and Callie Sendek. The Keck Project was made possible by the support of the Keck Geology Consortium and NSF-REU grant 1005122. Also, thanks to Sean Regan for his patience and wisdom in the Ultrachron lab and to Mike Easton for his patience and wisdom in the field. Comments by Toby Rivers, as editor, Félix Gervais, and an anonymous reviewer improved this manuscript.

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732

733 **Table**

734 see attached .xlsx files:

735 “Table 1: Sample Locations”

736 “Table 2. Monazite domain compositions”

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738 **Figure Captions**

739 Figure 1. Terrane maps of the Grenville Province of Ontario and New York (in light gray, with
740 major lakes and rivers in dark gray). **A)** The three tectonic subdivisions of Wynne-Edwards
741 (1972) are: the Central Gneiss Belt (CGB), the Central Metasedimentary Belt (CMB), and the
742 Central Granulite Terrain (CGT). The Central Metasedimentary Belt boundary zone (CMBbz) is
743 the shear zone at the boundary between the CGB and CMB. The Composite Arc Belt (CAB)
744 follows the designation of Easton (1992) and Carr et al. (2000). **B)** The terrane/domain
745 subdivisions of Davidson (1984) and Carr et al. (2000) are the Parry Sound (PS), the Muskoka,
746 many other terranes in the CGB that we do not label, the Bancroft (*sensu lato*, we equate the
747 Bancroft Domain with the CMBbz, an equivalence that is in first-order agreement with the
748 diverse delineations in the literature), the Elzevir, the Mazinaw (Ma), the Sharbot Lake (SL), the
749 Frontenac (Fr), the Adirondack Lowlands (AL), and the Adirondack Highlands. Shear zones are
750 the Allochthon Boundary Thrust (ABT; Rivers et al. 1989), the Bancroft (BSZ), the Mooroton
751 (MoSZ), the Robertson Lake (RLSZ), the Maberly (MaSZ), and the Carthage-Colton (CCSZ). In
752 addition to Ottawa, cities and towns marked with an open circle are Sudbury (S), Haliburton (H),
753 Killaloe (K), Bancroft (B), and Madoc (M).

754
755 Figure 2. Geological map of the study area from Hanmer and McEachern (1992). The Central
756 Metasedimentary Belt boundary zone (CMBbz) is a mid-crustal amphibolite facies shear zone at
757 the boundary between the Gneiss Belt and Metasedimentary Belt. Sample locations for which we
758 report U-Th-Pb dates on monazite are black squares. Other sample locations for which we have
759 some monazite data are grey squares, and sample locations without monazite are open squares.
760 Data presented here come from three localities in the CMBbz: West Guilford - WG, Fishtail

Lake - FL, and Killaloe – K. Some towns are marked with an x: Minden, Bancroft, Killaloe, and Pembroke. The Bancroft Shear Zone (BSZ) is a late extensional shear zone that is distinct from the CMBbz at this scale. Labeled meta-plutonic thrust sheets are the Redstone (Rd), Dysart (Dy), Glamorgan (Gl), and Foymount (Foy).

Figure 3. Composite X-ray maps of three garnet-biotite gneiss samples: Ca in green, Mg in red. For major minerals: biotite grains are red; plagioclase feldspar grains are green; garnet grains are brown; quartz and potassium feldspar grains are black. White circles with black outlines are monazite grain locations, and grains that we mapped are labeled (m10, for example). White ellipses are parallel to foliation (S). A) Sample 11AM-20 from the West Guilford region. The edges of this slide are epoxy with grains of sand, and the field of view is 3.5 cm wide. B. Sample 11AM-1 from the Fishtail Lake region. Zoned grains are garnet (mostly lower Ca in rims). The location of monazite grain m3 is in the circle with a white rim, in order to show the small garnet grain that partly encloses it. The right and left edges are epoxy with grains of sand, and field of view is 3.2 cm wide. C) Sample BY13-2 from the Killaloe region. Field of view is 3.5 cm wide (the white scale bar in the lower left corner is 5 mm).

Figure 4. Monazite grain maps for sample 11AM-20 (West Guilford). Monazite grain maps in this image (and the following two grain map images) are simultaneously processed in order to best compare compositions between grains in the sample. Orientation of monazite grains in this image (and the following two grain map images) is identical to the full section scan (Figure 3). In other words, the orientation of foliation relative to each grain map in this figure is from the upper right to the lower left. Some monazite grains have long axes parallel to foliation (e.g. m8); others

are at a high angle (e.g. m9). One grain (m6) is enclosed in garnet (gt). In this image (and the following two grain map images), dashed lines in the U maps indicate analysis locations.

Figure 5. Monazite compositions and dates for West Guilford samples. The two samples are plotted on different graphs with the same axes and scales. The y-axis units are yttrium cations, calculated from data in Table 2 assuming a two-cation formula for monazite: $(\text{Ce},\text{La},\text{Th})\text{PO}_4$. The grey arrows (in this image and in Figure 9) represent transitions (with respect to age) between lower- and higher-Y populations of monazite grains and domains. We distinguish between lower- and higher-Y populations using relative concentration with respect to each sample.

Figure 6. Monazite grain maps for sample 11AM-1 (Fishtail Lake). Two larger grains show distinctive inner core domains that are unusually rich in Nd. Two grains (m3 and m5) are partly enclosed in garnet. The other three grains are oriented with their long axes parallel to foliation (vertical in this frame of reference).

Figure 7. Photomicrographs of two monazite grains (m3 and m5) that are partly enclosed in garnet in sample 11AM-1 from Fishtail Lake.

Figure 8. Monazite grain maps for sample BY13-2 (Killaloe, same outcrop as BY13-1). Monazite core domains are geometrically complex and not resolvable into inner and outer cores. Many of these grains also show partial thin rims that are high in Y. Grains and grain domains show no preferred orientation with respect to foliation (horizontal in this frame of reference).

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807

808 Figure 9. Monazite compositions and dates for Killaloe samples.

809

810 Figure 10. Photomicrographs of two monazite grains that are enclosed in garnet in sample BY13-
811 2 from Killaloe. One grain (m3) is partly enclosed in garnet; another (m4) is fully enclosed. Both
812 occur adjacent to the same crack in the garnet grain.

813

814 Figure 11. Summary of U-Th-Pb monazite dates. The y-axis shows a gaussian probability for
815 each date, based on the weighted mean of the multiple analyses in an individual compositional
816 domain in the form of a 2σ error. Dates from relatively high Y domains and grains are orange,
817 and dates from relatively low Y domains and grains are purple. We argue that: early high Y
818 monazite domains indicate prograde monazite growth before the appearance of garnet; low Y
819 monazite domains indicate monazite growth in the presence of garnet; and late high-Y monazite
820 domains indicate monazite growth during garnet breakdown associated with retrogression and/or
821 decompression. All samples appear to share a late Ottawa history of garnet breakdown. Both
822 the Killaloe and West Guilford samples appear to record earlier prograde histories, but only the
823 Killaloe samples preserve a pre- and early Shawinigan history.

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825 **Supplemental Document: Detailed Sample Descriptions**

826 The **West Guilford samples (11AM-20 and HA13)** are from outcrops about 4 km apart.
827 In hand sample, 11AM-20 is a relatively fine-grained garnet-biotite gneiss with a weak foliation
828 defined by oriented biotite grains (Figure 3A). Major minerals are quartz, K-feldspar, plagioclase
829 feldspar, biotite, and garnet; muscovite, apatite, monazite, tourmaline, graphite, an oxide, and
830 pyrite are accessory minerals. Garnet is sparse and irregularly shaped. Quartz and feldspar grains
831 are relatively equant, tens to hundreds of microns in size, with straight to rounded grain
832 boundaries often meeting in triple-grain junctions. Quartz shows weak undulose extinction.

833 In hand sample, HA13 is a sillimanite-bearing garnet-biotite gneiss with a strong foliation
834 and lineation defined by elongate quartz aggregates and oriented biotite and sillimanite. The
835 lineation dips shallowly to the southeast, and the strong foliation and elongate quartz grains
836 make HA13 the most strongly foliated sample in the study. Gneiss at this outcrop shows cm-
837 scale felsic layers that may be deformed leucosome. Major minerals are quartz, plagioclase
838 (albite), garnet, K-feldspar, and biotite. Minor minerals are magnetite and sillimanite. Grain sizes
839 are hundreds of microns. Garnet grains are slightly zoned, equant and subhedral.

840 Both West Guilford samples are tectonites of the well-foliated, map-scale shear zone
841 between the Redstone and Dysart gneiss bodies (Figure 2) of Hanmer (1988, 1989) and Hanmer
842 and McEachern (1992), who understood these gneiss bodies as map-scale crystalline thrust
843 sheets (i.e. nappes) within the CMBbz. In contrast, the simplified map of the OGS (2011) in this
844 region defined the CMBbz more narrowly, as the tectonite units surrounding the lowermost
845 Redstone nappe. As a result, West Guilford is within the CMBbz of the Hanmer group, but it is
846 to the southeast (i.e. in the hanging wall) of the simplified CMBbz of the OGS (2011). Both
847 sample locations appear on Hanmer's (1989) detailed geologic map along strike of each other

within a continuous unit at the base of the Dysart nappe labeled “Poc: undifferentiated sillimanite-kyanite-garnet-ortho-amphibole-cordierite gneiss.” In contrast, the OGS (2011) map shows the location of sample HA13 as “45 clastic metasedimentary rocks,” but the location of sample 11AM-20 as “46 carbonate metasedimentary rocks.” The OGS mapping implies that the 11AM-20 outcrop is a large silicic xenolith in the tectonic marble breccia of this region. Outcrop here is not continuous enough to distinguish between this scenario and the scenario mapped by Hanmer (1989) who showed this outcrop as part of a long continuous unit of silicic gneisses. The West Guilford samples are a few km to the northeast of the West Guilford sample in the U/Pb zircon study of McEachern and van Breemen (1993).

The **Fishtail Lake sample (11AM-1)** is a garnet-biotite gneiss with foliation defined by compositional banding and oriented biotite grains (Figure 3B). Major minerals are quartz, K-feldspar, plagioclase feldspar, biotite, and garnet. Garnet grains are relatively small (mm-scale), equant, and subhedral; they show mild zoning and inclusions but do not appear to preserve an older fabric. Quartz and feldspar grains are relatively large (mm- to cm- scale) and equant, with straight to curved grain boundaries, and quartz shows undulose extinction.

Sample 11AM-1 is from an outcrop that is unambiguously within the CMBbz. Hanmer and McEachern (1992) show this locality as just below the base of the Grace thrust sheet and clearly as CMBbz, albeit in the Bancroft section for which they did not present detailed maps. The OGS includes this outcrop in its CMBbz “51 tectonite unit” (OGS, 2011). The older OGS map (Lumbers and Vertolli 2003) shows this unit as a “siliceous, micaceous, clastic metasedimentary rock,” equivalent to Hanmer’s (1989) Poc unit. This location is a few km east-northeast of the classic study area for cordierite-gedrite rocks by Lal and Moorhouse (1969).

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The **Killaloe samples (BY13-1 and BY13-2)** are from one outcrop. In hand sample, **BY13-1** is a garnet-biotite gneiss with a weak foliation defined by oriented biotite grains and compositional banding (light bands appear to be dismembered pegmatite). **BY13-2** (Figure 3C) is the same lithology but shows a stronger foliation, again defined by oriented biotite grains and compositional banding. For both samples, major minerals are quartz, plagioclase feldspar, K-feldspar, biotite, garnet, and apatite; accessory minerals are muscovite and magnetite. Grains are on the scale of millimeters, garnets are almost 1 cm wide, equant, and unzoned, although some garnet cores contain extensive inclusions of the other major minerals. Quartz and feldspar grains are equant with straight to curved grain boundaries and triple grain junctions, and quartz extinction is weakly undulose.

The BY13 samples are within the CMBbz of Hanmer and McEachern (1992) in what they map as tectonites just below the base of the Stafford thrust sheet. In contrast, the OGS (2011) puts this outcrop 3-4 km west of the CMBbz tectonites (i.e. in the footwall) in “41 migmatitic rocks and gneisses of undetermined protolith” within the Gneiss Belt. Lumbers (1982) shows this outcrop on a 1:100,000 map as older meta-sediments variably intruded by granitic rocks of the Algonquin Batholith. These samples are not necessarily well served by grouping within Hanmer’s (1989) Poc unit. The major mineral assemblages are nearly identical to the other three samples, but the BY13 samples contain significant apatite, and gneiss at this outcrop contains pods and layers of felsic leucosome. Nonetheless, the BY13 location does appear on the map of Hanmer and McEachern (1992) adjacent to the cordierite-gedrite gneisses that they associated with the sillimanite-garnet gneisses. The location of this sample locality is about 2 km west of the Killaloe samples in McEachern and van Breemen (1993).

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The occurrence of garnet and biotite (and in one case, sillimanite) in these quartzo-feldspathic gneisses tempts us to infer clastic sedimentary protoliths for them, and Hanmer and McEachern (1992) documented a strong spatial association between these garnet gneisses and the unusual cordierite-gedrite gneisses described by Lal and Moorhouse (1969). The protoliths to the cordierite-gedrite gneisses may well be altered volcanic rocks (Peck and Smith, 2005) that predate the Shawinigan orogeny. If so, a sedimentary and/or volcanic protolith to some or all of our garnet-biotite gneiss samples seems like a reasonable inference.

These samples were collected over the course of two summer field seasons. Abby Monreal collected samples 11AM-1 and 11AM-20 during a 2011-2012 Keck Geology Consortium project (Monreal, 2012); Dunn and Markley collected HA13 and the pair of BY13 samples during a summer 2013 field campaign aimed at a systematic sampling of CMBbz outcrops for this study. We had initially cast a much wider net for our study, including dozens of samples from the Central Gneiss Belt and Elzevir terrane of the Central Metasedimentary Belt (Figure 1). We specifically targeted late shear zones, including CMBbz straight gneiss (Hanmer and McEachern 1992), marble from the extensional Bancroft Shear Zone (Carlson et al. 1990), and orthogneiss and amphibolite from the Salerno Creek Deformation Zone (Easton and Kamo 2011). But most of the samples we collected do not contain monazite (open squares in Figure 2). In a future publication, we hope to present data from five other samples from the CMBbz (including localities in South Minden and Hoare Lake) and two samples from the Central Gneiss Belt (near Whitney) that contain monazite with complex U-Th-Pb systematics that appear reset by fluid-related dissolution and reprecipitation (grey squares in Figure 2).

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Table 1. Sample Locations (WGS 84)

Region	Sample	Lat.°N	Long.°W
West Guilford	11AM-20	45.0949	78.6386
West Guilford	HA13	45.1037	78.5983
Fishtail Lake	11AM-1	45.1575	78.1803
Killaloe	BY13-1	45.5401	77.4462
Killaloe	BY13-2	45.5401	77.4462

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[illegible]

BY13-2 m4 (gt)	1222	5	5	99.05	0.99	0.00	0.08	0.00	29.46	0.09	3.31	0.43	3.51	12.44	26.32	13.75	3.21	1.56	0.15	2.06	0.92	0.04	0.22	0.09	0.00	0.06	0.13	0.25
BY13-2 m11	1218	12	5	101.46	1.03	0.00	0.08	0.02	30.22	0.05	2.20	0.52	2.44	14.02	28.23	13.80	3.29	1.51	0.12	1.96	0.70	0.03	0.18	0.25	0.07	0.04	0.50	0.21
BY13-2 m10	1214	12	5	101.48	1.28	0.00	0.08	0.01	30.28	0.05	2.62	0.61	2.61	13.43	27.65	13.87	3.26	1.57	0.12	1.89	0.72	0.07	0.16	0.22	0.05	0.05	0.63	0.25
BY13-2 m8	1208	6	6	100.36	1.18	0.00	0.10	0.02	29.79	0.00	2.97	0.79	2.96	12.81	26.90	13.78	3.21	1.63	0.14	2.00	0.80	0.01	0.22	0.24	0.07	0.04	0.41	0.30
BY13-2 m12	1204	37	3	101.10	1.02	0.00	0.09	0.01	29.98	0.07	2.54	0.56	2.63	13.45	27.88	14.01	3.32	1.56	0.13	1.92	0.72	0.06	0.16	0.21	0.11	0.03	0.41	0.23
BY13-1 m6 (gt)	1198	12	6	101.25	0.70	0.00	0.20	0.00	30.71	0.03	2.82	0.58	2.59	14.35	28.53	13.09	3.27	1.35	0.12	1.70	0.74	0.04	0.16	0.00	0.06	0.00	0.02	0.25
BY13-1 m6 (gt)	1193	7	6	101.08	0.95	0.00	0.21	0.01	30.51	0.03	3.89	0.76	1.66	13.30	27.77	13.97	3.29	1.57	0.12	2.04	0.63	0.00	0.06	0.00	0.01	0.00	0.01	0.34
BY13-2 m4 (gt)	1182	13	6	100.08	0.89	0.00	0.09	0.02	29.32	0.16	3.69	0.38	1.31	13.31	28.05	14.55	3.41	1.70	0.10	2.14	0.54	0.04	0.07	0.00	0.10	0.00	0.02	0.26
BY13-1 m7	1182	13	5	99.37	1.12	0.00	0.11	0.00	29.87	0.02	4.72	0.51	1.20	12.75	27.14	13.92	3.33	1.68	0.12	1.97	0.51	0.01	0.04	0.00	0.10	0.00	0.02	0.33
BY13-1 m8	1178	10	6	101.56	0.93	0.00	0.14	0.00	30.53	0.00	4.12	0.42	1.65	13.30	28.44	14.05	3.41	1.62	0.12	1.86	0.62	0.00	0.07	0.00	0.09	0.00	0.02	0.29
Killaloe - low Y Shawinigan cores 1180-1160 Ma																												
BY13-2 m3 (gt)	1180	7	4	101.21	0.86	0.00	0.10	0.00	30.13	0.07	3.77	0.29	0.65	13.81	28.91	15.02	3.50	1.62	0.05	1.82	0.30	0.00	0.05	0.00	0.09	0.00	0.02	0.24
BY13-1 m8	1173	9	5	101.00	1.10	0.00	0.16	0.00	30.08	0.02	5.01	0.29	0.59	13.30	28.56	14.51	3.42	1.58	0.06	1.66	0.28	0.02	0.02	0.00	0.10	0.00	0.02	0.31
BY13-1 m7	1170	7	5	100.18	1.08	0.00	0.21	0.01	29.90	0.03	5.13	0.15	0.45	13.06	28.24	14.60	3.43	1.60	0.05	1.71	0.22	0.00	0.01	0.00	0.11	0.00	0.02	0.29
BY13-1 m8	1161	10	5	101.22	1.12	0.00	0.21	0.00	30.01	0.03	5.46	0.22	0.23	13.41	28.92	14.71	3.47	1.56	0.03	1.46	0.13	0.00	0.00	0.00	0.09	0.00	0.02	0.31
Killaloe - low Y Ottawa cores, rims, and domains ca. 1060 Ma																												
BY13-1 m14	1072	20	3	100.80	0.63	0.08	0.20	0.00	30.02	0.00	2.10	1.07	0.55	17.05	31.15	12.04	3.15	1.04	0.02	1.26	0.22	0.00	0.02	0.00	0.00	0.00	0.03	0.27
BY13-2 m10	1068	10	5	101.91	0.87	0.00	0.13	0.00	30.19	0.08	3.72	0.36	1.01	13.55	29.14	14.91	3.51	1.70	0.07	1.88	0.41	0.00	0.06	0.00	0.11	0.01	0.02	0.23
BY13-1 m10	1066	5	6	101.85	1.17	0.03	0.34	0.02	30.32	0.10	4.98	0.77	0.50	14.29	29.21	13.55	3.30	1.32	0.03	1.32	0.21	0.00	0.01	0.00	0.08	0.00	0.02	0.35
BY13-2 m6	1056	22	3	100.22	0.84	0.02	0.15	0.00	29.50	0.05	3.84	0.32	0.82	13.74	29.11	14.52	3.45	1.46	0.07	1.66	0.32	0.05	0.05	0.00	0.06	0.00	0.02	0.23
BY13-2 m2	1054	10	6	99.20	0.91	0.00	0.14	0.00	29.26	0.08	3.89	0.35	0.94	13.37	28.08	14.52	3.46	1.58	0.06	1.80	0.38	0.03	0.04	0.00	0.12	0.00	0.02	0.23
BY13-1 m8	1052	10	5	101.64	0.99	0.01	0.29	0.01	30.16	0.05	4.42	0.84	0.57	15.80	29.83	12.46	3.16	1.11	0.05	1.28	0.23	0.00	0.01	0.00	0.06	0.00	0.02	0.33
Killaloe - high Y Ottawa rims ca. 1045 Ma																												
BY13-2 m11	1055	12	5	100.88	0.69	0.01	0.28	0.00	29.65	0.18	3.39	0.31	3.20	12.95	27.45	14.13	3.32	1.59	0.13	1.96	0.87	0.05	0.23	0.01	0.16	0.09	0.04	0.20
BY13-2 m3 (gt)	1046	20	4	101.21	0.71	0.00	0.19	0.01	30.29	0.00	3.20	0.37	3.02	13.00	27.50	14.27	3.34	1.67	0.14	2.01	0.86	0.08	0.19	0.03	0.10	0.04	0.03	0.20
BY13-2 m10	1045	8	4	101.59	0.70	0.00	0.30	0.00	29.91	0.22	3.55	0.32	3.17	13.06	27.54	14.17	3.29	1.63	0.17	1.94	0.86	0.09	0.23	0.00	0.12	0.09	0.03	0.21
BY13-2 m4 (gt)	1039	21	3	99.30	0.70	0.00	0.30	0.00	28.83	0.09	3.76	0.23	3.17	12.83	27.00	13.82	3.23	1.54	0.16	2.06	0.90	0.11	0.23	0.02	0.03	0.04	0.05	0.20
BY13-2 m5	1035	14	5	99.25	0.69	0.04	0.24	0.01	29.14	0.06	3.29	0.35	3.13	12.87	26.98	13.97	3.27	1.59	0.13	1.89	0.83	0.08	0.24	0.01	0.09	0.11	0.04	0.20
BY13-2 m9	1033	13	4	99.58	0.76	0.05	0.23	0.01	29.30	0.08	3.31	0.30	2.18	13.24	27.62	14.28	3.34	1.66	0.11	1.96	0.66	0.04	0.14	0.04	0.06	0.01	0.03	0.19
BY13-2 m13	1032	20	4	100.40	0.75	0.00	0.18	0.01	29.80	0.04	3.20	0.37	3.06	13.02	27.44	14.05	3.30	1.61	0.14	1.86	0.83	0.06	0.20	0.03	0.14	0.07	0.04	0.20
Killaloe - oddballs																												
BY13-1 m10	1113	11	6	101.39	0.83	0.00	0.22	0.01	30.16	0.04	3.75	0.52	0.53	14.56	29.61	13.98	3.39	1.54	0.04	1.61	0.23	0.00	0.01	0.01	0.09	0.00	0.01	0.27
BY13-1 m13	1091	9	5	101.78	1.00	0.00	0.23	0.03	30.56	0.05	3.97	0.78	3.02	18.92	29.33	8.45	2.54	0.50	0.08	0.84	0.62	0.07	0.22	0.05	0.08	0.08	0.08	0.31
BY13-1 m12	1030	18	5	101.95	0.83	0.02	0.24	0.01	30.50	0.00	3.72	0.64	1.06	13.99	29.05	13.99	3.36	1.71	0.10	1.93	0.43	0.00	0.04	0.04	0.05	0.00	0.02	0.26

Figure 1. Terrane maps of the Grenville Province of Ontario and New York (in light gray, with major lakes and rivers in dark gray). A) The three tectonic subdivisions of Wynne-Edwards (1972) are: the Central Gneiss Belt (CGB), the Central Metasedimentary Belt (CMB), and the Central Granulite Terrain (CGT). The Central Metasedimentary Belt boundary zone (CMBbz) is the shear zone at the boundary between the CGB and CMB. The Composite Arc Belt (CAB) follows the designation of Easton (1992) and Carr et al. (2000). B) The terrane/domain subdivisions of Davidson (1984) and Carr et al. (2000) are the Parry Sound (PS), the Muskoka, many other terranes in the CGB that we do not label, the Bancroft (*sensu lato*, we equate the Bancroft Domain with the CMBbz, an equivalence that is in first-order agreement with the diverse delineations in the literature), the Elzevir, the Mazinaw (Ma), the Sharbot Lake (SL), the Frontenac (Fr), the Adirondack Lowlands (AL), and the Adirondack Highlands. Shear zones are the Allochthon Boundary Thrust (ABT; Rivers et al. 1989), the Bancroft (BSZ), the Mooroton (MoSZ), the Robertson Lake (RLSZ), the Maberly (MaSZ), and the Carthage-Colton (CCSZ). In addition to Ottawa, cities and towns marked with an open circle are Sudbury (S), Haliburton (H), Killaloe (K), Bancroft (B), and Madoc (M).

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Figure 2. Geological map of the study area from Hanmer and McEachern (1992). The Central Metasedimentary Belt boundary zone (CMBbz) is a mid-crustal amphibolite facies shear zone at the boundary between the Gneiss Belt and Metasedimentary Belt. Sample locations for which we report U-Th-Pb dates on monazite are black squares. Other sample locations for which we have some monazite data are grey squares, and sample locations without monazite are open squares. Data presented here come from three localities in the CMBbz: West Guilford - WG, Fishtail Lake - FL, and Killaloe - K. Some towns are marked with an x: Minden, Bancroft, Killaloe, and Pembroke. The Bancroft Shear Zone (BSZ) is a late extensional shear zone that is distinct from the CMBbz at this scale. Labeled meta-plutonic thrust sheets are the Redstone (Rd), Dysart (Dy), Glamorgan (Gl), and Foy (Foy).

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Figure 3. Composite X-ray maps of three garnet-biotite gneiss samples: Ca in green, Mg in red. For major minerals: biotite grains are red; plagioclase feldspar grains are green; garnet grains are brown; quartz and potassium feldspar grains are black. White circles with black outlines are monazite grain locations, and grains that we mapped are labeled (m10, for example). White ellipses are parallel to foliation (S). A) Sample 11AM-20 from the West Guilford region. The edges of this slide are epoxy with grains of sand, and the field of view is 3.5 cm wide. B. Sample 11AM-1 from the Fishtail Lake region. Zoned grains are garnet (mostly lower Ca in rims). The location of monazite grain m3 is in the circle with a white rim, in order to show the small garnet grain that partly encloses it. The right and left edges are epoxy with grains of sand, and field of view is 3.2 cm wide. C) Sample BY13-2 from the Killaloe region. Field of view is 3.5 cm wide (the white scale bar in the lower left corner is 5 mm).

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Figure 4. Monazite grain maps for sample 11AM-20 (West Guilford). Monazite grain maps in this image (and the following two grain map images) are simultaneously processed in order to best compare compositions between grains in the sample. Orientation of monazite grains in this image (and the following two grain map images) is identical to the full section scan (Figure 3). In other words, the orientation of foliation relative to each grain map in this figure is from the upper right to the lower left. Some monazite grains have long axes parallel to foliation (e.g. m8); others are at a high angle (e.g. m9). One grain (m6) is enclosed in garnet (gt). In this image (and the following two grain map images), dashed lines in the U maps indicate analysis locations.

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Figure 5. Monazite compositions and dates for West Guilford samples. The two samples are plotted on different graphs with the same axes and scales. The y-axis units are yttrium cations, calculated from data in Table 2 assuming a two-cation formula for monazite: (Ce,La,Th)PO₄. The grey arrows (in this image and in Figure 9) represent transitions (with respect to age) between lower- and higher-Y populations of monazite grains and domains. We distinguish between lower- and higher-Y populations using relative concentration with respect to each sample.

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Figure 6. Monazite grain maps for sample 11AM-1 (Fishtail Lake). Two larger grains show distinctive inner core domains that are unusually rich in Nd. Two grains (m3 and m5) are partly enclosed in garnet. The other three grains are oriented with their long axes parallel to foliation (vertical in this frame of reference).

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Figure 7. Photomicrographs of two monazite grains (m3 and m5) that are partly enclosed in garnet in sample 11AM-1 from Fishtail Lake.

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Figure 8. Monazite grain maps for sample BY13-2 (Killaloe, same outcrop as BY13-1). Monazite core domains are geometrically complex and not resolvable into inner and outer cores. Many of these grains also show partial thin rims that are high in Y. Grains and grain domains show no preferred orientation with respect to foliation (horizontal in this frame of reference).

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Figure 9. Monazite compositions and dates for Killaloe samples.

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Figure 10. Photomicrographs of two monazite grains that are enclosed in garnet in sample BY13-2 from Killaloe. One grain (m3) is partly enclosed in garnet; another (m4) is fully enclosed. Both occur adjacent to the same crack in the garnet grain.

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Figure 11. Summary of U-Th-Pb monazite dates. The y-axis shows a gaussian probability for each date, based on the weighted mean of the multiple analyses in an individual compositional domain in the form of a 2σ error. Dates from relatively high Y domains and grains are orange, and dates from relatively low Y domains and grains are purple. We argue that: early high Y monazite domains indicate prograde monazite growth before the appearance of garnet; low Y monazite domains indicate monazite growth in the presence of garnet; and late high-Y monazite domains indicate monazite growth during garnet breakdown associated with retrogression and/or decompression. All samples appear to share a late Ottawa history of garnet breakdown. Both the Killaloe and West Guilford samples appear to record earlier prograde histories, but only the Killaloe samples preserve a pre- and early Shawinigan history.

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