Middle Jurassic to earliest Cretaceous mid-crustal tectono-metamorphism in the northern Canadian Cordillera: Recording foreland-directed migration of an orogenic front

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ABSTRACT

In situ sensitive high-resolution ion microprobe monazite geochronology and garnet isopleth thermobarometry reveal a previously unrecognized Middle Jurassic to earliest Cretaceous mid-crustal tectono-metamorphic event in the eastern part of the Yukon-Tanana terrane (Finlayson Lake district, southeast Yukon) in the northern Canadian Cordillera. Intersection of garnet end-member compositional isopleths applied to single-stage, growth-zoned garnet records progressive garnet growth from 550 °C and 6.1–6.6 kbar to 600 °C and 7.5 kbar. Monazite textures, chemical zoning, and in situ U-Pb ages record a single protracted episode of monazite growth from ca. 169 to 142 Ma coeval with the development of transposition fabrics and the late stages of garnet growth. This event post-dates widespread Early Jurassic exhumation of Yukon-Tanana terrane rocks west of the Tintina fault in west-central Yukon, which were previously ductily deformed and metamorphosed in the Permo-Triassic. The lack of evidence for Permo-Triassic ductile deformation and high-grade metamorphism within the Finlayson Lake district, and its position east of the Permain arc center and west of Permain blueschists and eclogites, suggests this eastern part of the terrane occupied the cool forearc at this time. These data indicate younger, more protracted mid-crustal orogenesis in the northern Cordillera than was previously recognized, with deformation and metamorphism migrating toward the foreland and downward in the

INTRODUCTION

The Yukon-Tanana terrane is characterized by intense deformation and metamorphism that have been interpreted to be consequences of the accretion of the allochthonous peri-Laurentian terranes (e.g., Parrish, 1995; Gibson et al., 2008; Berman et al., 2007). Geochronologic investigations in the north-central part of the Yukon-Tanana terrane in Yukon indicate that the main episode of ductile deformation and amphibolite-facies metamorphism occurred in the Permo-Triassic (Berman et al., 2007; Beranek and Mortensen, 2011). K-Ar and 40Ar/39Ar data from green-schist- and amphibolite-facies rocks of the Yukon-Tanana terrane in the Yukon and Alaskan Cordillera (referred to herein as “northern Cordillera”) yield Early to Middle Jurassic cooling ages (Hansen et al., 1991; Hunt and Roddie, 1992; Stevens et al., 1993; Johnston et al., 1996; Gordey et al., 1998; Dusel-Bacon et al., 2002; Knight et al., 2013) that are interpreted to record widespread regional uplift and exhumation (Johnston et al., 1996; Berman et al., 2007; Beranek and Mortensen, 2011). The timing of such events contrasts with that documented in the southeastern Canadian Cordillera where deformation and metamorphism related to accretion of the peri-Laurentian terranes (Quesnel and Stikine terranes) and westward underthrusting of the North American plate apparently occurred in the Early Jurassic through early Paleogene (Evenchick et al., 2007, and references therein).

Furthermore, geochronologic data from the metamorphic hinterland of the orogen in the southern Canadian Cordillera have revealed that despite similarities in metamorphic grade and deformation patterns, deformation and metamorphism were diachronous with depth and across strike. Rocks presently in the upper structural levels were buried, heated, and exhumed in the Jurassic (Murphy et al., 1995; Parrish, 1995; Colpron et al., 1996; Crowley et al., 2000; Gibson et al., 2005, 2008), while structurally deeper rocks continued to be buried and heated from the Cretaceous to the earliest Eocene (Carr, 1991; Parrish, 1995; Gibson et al., 1999, 2005, 2008; Crowley and Parrish, 1999; Crowley et al., 2000). Additionally, numerous authors have suggested that the deformation migrated from the metamorphic and plutonic hinterland to the foreland thrust-and-fold belt to the northeast in the Mesozoic and Tertiary, and that there was a dynamic link between these two structural domains (e.g., Price, 1981; Brown et al., 1992; Johnson and Brown, 1996; Simony and Carr, 2011).

Herein, we present results of an integrated metamorphic and microstructural study coupled with in situ sensitive high-resolution ion microprobe (SHRIMP) analysis of monazite within amphibolite-facies rocks from the Finlayson Lake district (southeast Yukon) in the eastern part of the Yukon-Tanana terrane in the northern Cordillera. This study provides an example of a successful application of garnet isopleth thermobarometry that incorporates the effects of a changing bulk composition caused by the
Fractation of components during garnet growth, to determine pressure-temperature (P-T) conditions at successive increments along the prograde P-T path. The combined data reveal that deformation and metamorphism were diachronous within the Yukon-Tanana terrane, with younger events recorded at progressively deeper structural levels in a manner analogous to, and partly coincident with, the southeastern Canadian Cordillera.

GEOLOGIC SETTING

The Yukon-Tanana and Slide Mountain terranes occupy the innermost position of the accreted terranes in the northern Cordillera. The Yukon-Tanana terrane consists of a pre–Late Devonian metasedimentary basement (Snowcap assemblage) with lithological, geochemical, and isotopic compositions that suggest it represents a rifted portion of the western Laurentian continental margin (Piercey and Colpron, 2009; Colpron and Nelson, 2009). The Snowcap assemblage is overlain by three unconformity-bound Upper Devonian to Permian volcanic arc sequences (Finlayson, Klinkit, Klondike assemblages) that are coeval with oceanic chert, argillite, and mafic volcanic rocks of the Slide Mountain terrane (Colpron et al., 2006). The Yukon-Tanana and Slide Mountain terranes are thought to have originated as an arc and back-arc pair, respectively, during the Late Devonian to Permian opening of the Slide Mountain ocean (Nelson et al., 2006). Closure of the Slide Mountain ocean and the initial accretion of Yukon-Tanana to the western Laurentian margin is interpreted by Beranek and Mortensen (2011) to have occurred in the Late Permian.

In southeast Yukon, parts of the Yukon-Tanana and Slide Mountain terranes have been offset 490 km to the southeast, relative to the main body to the west, through a combined 430 km of dextral strike-slip displacement along the Tintina fault in the Paleogene, and 60 km of extension in the Cretaceous (Gabrielse et al., 2006). Figure 1 shows the restored position of this offset block of the Yukon-Tanana terrane (Finlayson Lake district) relative to its counterpart to the west. In the Finlayson Lake district, the Yukon-Tanana and Slide Mountain terranes are separated from marginal rocks of ancestral North America to the northeast by the Inconnu thrust, a major Jurassic–Cretaceous contractional fault that carried the Slide Mountain and Yukon-Tanana terranes on top of the North American continental margin sequence (Murphy et al., 2006).

In their pre-Paleogene configuration, rocks of the Finlayson Lake district lie immediately northeast of a tectonic window that exposes...
an Early to mid-Cretaceous high-grade metamorphic infrastructural domain of the Yukon-Tanana terrane southwest of the Tintina fault (Staples et al., 2013; Fig. 2). West of the Cretaceous domain, across the Australia Creek fault (Staples et al., 2013), Yukon-Tanana terrane rocks were transposed and metamorphosed in the Late Permian to Early Triassic and in the Early Jurassic (Berman et al., 2007; Beranek and Mortensen, 2011; Fig. 2) before they were exhumed to upper crustal levels in the Early to Middle Jurassic (Johnston et al., 1996; Dusel-Bacon, 2002; Berman et al., 2007). Southeast of the Cretaceous domain, across the Stewart River normal fault, are sub-greenschist-facies, weakly deformed to undeformed Devonian and Mississippian volcanic and plutonic rocks that record Paleozoic and Mesozoic 40Ar/39Ar cooling ages (Knight et al., 2013). Both the Permo-Jurassic metamorphic domain to the west and the Devono-Mississippian rocks to the southeast were presumably the suprastructural “lid” from beneath which the Cretaceous metamorphic domain (essentially a core complex) was exhumed along the Australia Creek and Stewart River faults, respectively, in the mid-Cretaceous (Staples et al., 2013).

Similar high-grade, ductily deformed rocks in east-central Alaska are interpreted as parautochthonous North American continental margin rocks (Lake George assemblage in Fig. 1) exhumed from beneath the Yukon-Tanana terrane along mylonitic extensional faults in the mid-Cretaceous (Hansen, 1990; Hansen and Dusel-Bacon, 1998; Dusel-Bacon et al., 1995, 2002, 2006). A parautochthonous North American origin for these rocks in east-central Alaska is based on Archean and Proterozoic U-Pb zircon inheritance and detrital ages, stratigraphic similarity with rocks in the Selwyn Basin, their structural position, and a lack of late Paleozoic arc assemblages characteristic of the Yukon-Tanana terrane (Dusel-Bacon, 2006; Nelson et al., 2006; Dusel-Bacon and Williams, 2009). Similar characteristics, including the exhumation of moderate- to high-pressure amphibolite-facies rocks from beneath the Yukon-Tanana terrane along mid-Cretaceous extensional shear zones, suggest that the Australia Mountain domain in west-central Yukon may also be a parautochthonous North American margin (Staples et al., 2013).

Permo-Triassic metamorphism in the central part of the Yukon-Tanana terrane was spatially and temporally associated with a belt of Permian arc magmatism (Klondike assemblage) (Mortensen, 1990; Gordey and Ryan, 2005; Berman et al., 2007). Permian magmatic rocks and a discontinuous belt of coeval blueschist and eclogite to the northeast crop out along the eastern side of the Yukon-Tanana terrane (Fig. 1; Creaser et al., 1997; Erdmer et al., 1998) and are interpreted to record a Permian northeast-facing magmatic arc and accretionary wedge, respectively (Mortensen, 1992; Nelson et al., 2006). By the Early to Middle Triassic (251–235 Ma), siliciclastic strata that contain detrital zircon inferred to have been sourced in the Paleoarc assemblages of the Yukon-Tanana terrane were deposited on the western Laurentian continental margin (Beranek and Mortensen, 2011). This depositional relationship implies that the Slide Mountain ocean had closed by the Early Triassic and the Yukon-Tanana terrane had been uplifted and accreted to the western Laurentian margin, forming a downward slope onto Laurentian strata. Beranek and Mortensen (2011) suggested that the Late Permian tectono-metamorphism within the Yukon-Tanana terrane is therefore the result of the accretion of the Yukon-Tanana terrane onto the western Laurentian margin.
The study area lies in the structurally lowest of three thrust sheets that imbricate Yukon-Tanana terrane in the Finlayson Lake district. This lowest sheet, the Big Campbell thrust sheet, is bound below by the Big Campbell thrust and above by the Money Creek thrust (Fig. 2; Murphy et al., 2006). The Big Campbell thrust sheet is a post-metamorphic thrust juxtaposing amphibolite-grade rocks of Yukon-Tanana terrane over Upper Triassic shale and spatially associated ultramafic rocks correlated with the Slide Mountain terrane (Fig. 2). The Big Campbell thrust has been inferred to be related to the post–Late Triassic, pre–mid-Cretaceous Inconnu thrust in a duplex structure (Murphy et al., 2006). The Big Campbell thrust sheet is exposed in the footwall of the North River fault, a regional-scale mid-Cretaceous normal fault that cuts across the thrust sheets of the Yukon-Tanana terrane in the Finlayson Lake district in a northeast-southwest trend (Figs. 3 and 4; Murphy, 2004). The Money Creek thrust sheet, and the structurally higher Cleaver Lake thrust sheet, are preserved to the southeast in the hanging wall of the North River fault. Rocks in the footwall of the North River fault yield mid-Cretaceous K-Ar cooling ages (Hunt and Roddick, 1992; Murphy et al., 2001), and have been ductily deformed at amphibolite facies, prior to the emplacement of the granites. The mid-Cretaceous granites are massive and discordant with respect to deformation in their wall rocks, except in their roof areas near the North River fault where a weak foliation is locally developed. Low-grade rocks in the hanging wall to the southeast compose the upper thrust sheets of the Yukon-Tanana terrane; these lack Cretaceous granites and record Mississippian K-Ar cooling ages (Hunt and Roddick, 1987; Murphy, 2004). The North River fault is inferred to have been the northern extension of the mid-Cretaceous Stewart River fault (Staples et al., 2013) in the western McQuesten and eastern Stewart River map areas prior to Paleogene offset on the Tintina fault (Fig. 2).

The metapelitic samples examined in this study lie in the footwall of the North River fault (Fig. 3), in the oldest exposed rock unit in the Finlayson Lake district, the pre–Upper Devonian North River formation. This formation consists of quartzose psammite, non-carbonaceous metapelite, marble, calcareous schist, and felsic metavolcanic members at or near its top (Murphy et al., 2006). The samples were collected from a horizon dominated by calcareous and pelitic schist immediately northeast of the Tintina fault (Fig. 3). The North River formation is considered part of the Snowcap assemblage (Colpron et al., 2006). In the Finlayson Lake district, this unit has been intruded and overlain by Late Devonian to Early Mississippian arc and back-arc facies igneous rocks (Murphy et al., 2006). Metamorphic grade and stratigraphic and structural depth broadly increase within the footwall of the North River fault from chlorite-bearing assemblages in the northeast to garnet-staurolite–bearing assemblages in the southwest, reflecting increased normal throw along the fault to the southwest. Samples were collected within the lowest exposed structural levels within the footwall, which were ductily deformed during or after amphibolite-facies metamorphism, as evidenced by wrapping of the penetrative foliation around garnet and staurolite porphyroblasts. Parallelism of compositional layering and lithologic contacts with a pervasive shallowly dipping foliation that contains rare intrafolial isoclinal folds suggests that these rocks experienced at least one generation of high-strain transposition.

Resolving the paleogeographic setting of the Yukon-Tanana terrane in the Late Jurassic–Early Cretaceous when metamorphism and deformation documented in this study occurred, particularly with respect to what lay outboard (to the west) of the Intermontane terranes, is critical to understanding the overall dynamic of the northern Cordilleran orogen at that time. In British Columbia and southeast Alaska, the Insular terranes (Peninsular-Alexander-Wrangellia) lie west of the Yukon-Tanana terrane, with the Alexander terrane lying structurally beneath rocks correlated with the Yukon-Tanana and Stikine terranes (Gehrels, 2001, 2002; Saleeby, 2000). Evidence for the initial accretion of the Alexander terrane to western Yukon-Tanana terrane in British Columbia and southeastern Alaska in the Middle to Late Jurassic include: (1) the presence of detrital zircons derived from both the Alexander and Yukon-Tanana terranes in Upper Jurassic–Lower Cretaceous strata of the Gravina belt (Kapp and Gehrels, 1998), (2) depositional overlap of Middle Jurassic (ca. 175 Ma) volcanic rocks and Upper Jurassic–Lower Cretaceous strata of the Gravina belt on both the Alexander and Yukon-Tanana terranes (Gehrels, 2001), (3) the presence of Late Jurassic (ca. 162–140 Ma) dikes that cross-cut a low-angle ductile tectonic boundary between the Yukon-Tanana terrane and the structurally underlying Alexander terrane in southeastern Alaska (Saleeby, 2000), and (4) an Early to Middle Jurassic shear zone within and along the eastern margin of the Alexander terrane further north in southeastern Alaska (McClelland and Gehrels, 1990).

In contrast, to the north, in mainland Alaska, earliest estimates of Insular terrane accretion are Late Jurassic, based on the cessation of magmatism, uplift and exhumation of the Talkeetna arc (Peninsular terrane), and deposition of conglomerates in the Late Jurassic (Clift et al., 2005; Trop and Ridgway, 2007). However, Hults et al. (2013) interpreted the presence of two Creta-Cretaceous flysch belts separated by a Late Cretaceous deformation zone, with each basin receiving detritus solely from Wrangellia (to the south) or the previously accreted Yukon-Tanana and Farewell terranes (to the north), to indicate that Wrangellia did not accrete to the Yukon-Tanana terrane until the Late Cretaceous.

These studies from different parts of the Insular-Intermontane terrane boundary imply that accretion of the Insular terranes outboard of the Yukon-Tanana terrane was diachronous, with oblique collision initiating in the south and younging to the northwest.

After restoring ~430 km of dextral offset along the Tintina fault in the Paleogene, and 60 km of extension in the Cretaceous (Gabrielse et al., 2006), as well as ~300–400 km of dextral displacement along the Denali fault (Eisbacher, 1976; Lowey, 1998, and references therein), the Finlayson Lake district of Yukon-Tanana terrane east of Wrangellia in mainland Alaska. However, Nelson et al. (2013) suggested that in the mid-Cretaceous (ca. 120–90 Ma) a crustal block comprising the Intermontane terranes (including the Yukon-Tanana terrane) was bound by a set of roughly coeval sinistral and dextral strike-slip fault systems and was extruded to the northwest. Total motion on the dextral system is estimated at ~250 km (Nelson et al., 2013), with as much as 400–800 km of sinistral displacement estimated along the western fault system (Monger et al., 1994; Nelson et al., 2013). Further restoration along these fault systems places the Finlayson Lake district to the east of the Insular terranes of coastal British Columbia and southeast Alaska, where their accretion to the Yukon-Tanana terrane occurred in Middle to Late Jurassic (McClelland et al., 1992; van der Heyden, 1992, Saleeby, 2000; Gehrels, 2001; Trop and Ridgway, 2007).

PETROLOGICAL ANALYTICAL METHODS

Microprobe and Conventional Garnet-Biotite Thermometry

Mineral compositions were quantitatively analyzed using a fully automated CAMECA SX50 instrument, operating in wavelength-dispersion mode with the following operating conditions: excitation voltage, 15 kV; beam current, 20 nA (10 nA for micas); peak count time, 20 s (40 s for F, Cl); background count time, 10 s (20 s for F, Cl); spot diameter, 5 μm. Quantitative data were obtained for garnet and...
Middle Jurassic to earliest Cretaceous tectono-metamorphism in the northern Canadian Cordillera

Figure 3. Geological map of a portion of the Finlayson Lake district showing the position of the sample locations within the footwall of the mid-Cretaceous North River fault and immediately east of the Tintina fault. Labels A and A’ locate the end points of the line of section for the cross-section in Figure 4. Map modified from Murphy et al. (2006).
Figure 4. Cross-section from points A to A’ on Figure 3, roughly perpendicular to the strike of the North River fault. This section shows that rocks of this study area are located in the footwall of the North River fault. This fault exposes ductily deformed, amphibolite-facies rocks with mid-Cretaceous K-Ar cooling ages within the footwall, against low-grade rocks with Mississippian K-Ar cooling ages in the hanging wall to the southeast. Fault abbreviations: NRF—North River fault; CLT—Cleaver Lake thrust. No vertical exaggeration.

biotite (Table 1). Garnet end-member compositions were calculated using an Fe³⁺ content estimated from charge balance and stoichiometry (e.g., Droop, 1987). Temperatures were calculated using the winTWQ program (version 2.32; Berman, 2007), which uses internally consistent thermodynamic data for end members and mixing properties to calculate an independent set of equilibria. The following garnet-biotite thermometer was used (mineral abbreviations after Kretz, 1983):

\[ \text{Alm + Phl} = \text{Pyr} + \text{Ann}. \]  

(1)

The winTWQ software incorporates solid-solution models for garnet and biotite (Berman, 2007). Absolute errors of thermometric data are considered to be approximately ±50 °C (Berman, 1991), with appreciably smaller errors associated with relative differences between samples. Pressures based on garnet-plagioclase equilibria were not calculated because of the low anorthite content (X_An = <10%), which reduces the reliability of the geobarometric estimates (Ashworth and Evirgen, 1985; Todd, 1998).

**Isochemical Phase Diagram Calculations**

Isochemical phase diagrams (equilibrium assemblage diagrams) and mineral-composition isopleths were calculated in the system MnNaKFeMgSiO₆ (MnO–Na₂O–CaO–K₂O–FeO∗–MgO–Al₂O₃–SiO₂–H₂O–TiO₂–Fe₂O₃) using the Gibbs free-energy minimization software TheriaK-Domino (de Capitani and Brown, 2007; de Capitani and Brown, 2007). This software incorporates internally consistent thermodynamic data to calculate an independent set of equilibria. The following garnet-biotite isopleths were calculated in the system MnNaKFeMgSiO₆ (MnO–Na₂O–CaO–K₂O–FeO∗–MgO–Al₂O₃–SiO₂–H₂O–TiO₂–Fe₂O₃) using the Gibbs free-energy minimization software TheriaK-Domino (de Capitani and Brown, 2007; de Capitani and Brown, 2007). This software incorporates internally consistent thermodynamic data to calculate an independent set of equilibria.

**TABLE 1. MICROPROBE ANALYSES USED IN THERMOBARIOMETRIC CALCULATIONS**

<table>
<thead>
<tr>
<th>Oxides</th>
<th>Sample 11RS033</th>
<th>Sample 11RS043</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Garnet core</td>
<td>Garnet rim</td>
</tr>
<tr>
<td>Si</td>
<td>2.96</td>
<td>2.97</td>
</tr>
<tr>
<td>Ti</td>
<td>0.01</td>
<td>0.01</td>
</tr>
<tr>
<td>Al</td>
<td>2.04</td>
<td>2.03</td>
</tr>
<tr>
<td>Fe²⁺</td>
<td>0.02</td>
<td>0.02</td>
</tr>
<tr>
<td>Fe³⁺</td>
<td>2.16</td>
<td>2.42</td>
</tr>
<tr>
<td>Mn</td>
<td>0.28</td>
<td>0.01</td>
</tr>
<tr>
<td>Mg</td>
<td>0.12</td>
<td>0.24</td>
</tr>
<tr>
<td>Ca</td>
<td>0.42</td>
<td>0.51</td>
</tr>
<tr>
<td>Na</td>
<td>N.D.</td>
<td>N.D.</td>
</tr>
<tr>
<td>K</td>
<td>N.D.</td>
<td>N.D.</td>
</tr>
<tr>
<td>Fe(Fe + Mg)</td>
<td>0.96</td>
<td>0.95</td>
</tr>
<tr>
<td>X_Aln</td>
<td>0.726</td>
<td>0.813</td>
</tr>
<tr>
<td>X_Pyr</td>
<td>0.040</td>
<td>0.082</td>
</tr>
<tr>
<td>X_Grt</td>
<td>0.140</td>
<td>0.101</td>
</tr>
</tbody>
</table>

Note: Garnet end-member abbreviations: Alm—almandine; Pyr—pyrope; Grs—grossular; Sp—atolls.

*All Fe assumed to be Fe²⁺. Number of cations normalized on the basis of 12 oxygen (garnet) or 11 oxygen (biotite).

**PETROGRAPHY AND MINERAL CHEMISTRY**

Four monazite-bearing samples (Fig. 3 inset map) that are lithologically and texturally similar were selected for in situ U-Pb monazite SHRIMP geochronology. Two of these samples (33 and 43) were selected for quantitative metamorphic (P-T) analysis. All investigated samples are quartz- and muscovite-rich, garnet-bearing schists that are intercalated with calcareous metapelite and meta-psammite of the “calcareous unit” of the North River formation described by Murphy (1998) and Murphy et al. (2006). The samples contain an assemblage of Garnet + St + Bt + Ms + Pl + Qtz + Ilm (after Rt). Muscovite and quartz compose the bulk of the mineral mode (70%). Garnet constitutes ~15% of the mode. Biotite (5%–10%) is much less abundant than muscovite, and staurolite and ulvospinel (White et al., 2000); and chlorite, cordierite, chloritooid, staurolite, and epidote (Holland and Powell, 1998). Garnet, biotite, staurolite, chloritooid, chlorite, and cordierite were extended to the Mn-bearing system as outlined in Tinkham et al. (2001). All other phases were treated as pure. A pure H₂O fluid was considered in excess in all calculations. Bulk rock compositions (Table 2) were determined by whole-rock x-ray fluorescence (XRF) analysis from thin-section offcuts. Ilmenite is the only Fe-bearing oxide, suggesting that these rocks are fairly reduced. Therefore, a nominal amount of Fe₂O₃ (0.01 mol% Fe₂O₃) was added to the model system to be consistent with the garnet and biotite solution models of White et al. (2007), which incorporate Fe₂O₃.
TABLE 2. NORMALIZED EFFECTIVE BULK ROCK COMPOSITIONS CORRESPONDING TO DIFFERENT STAGES OF GARNET GROWTH

<table>
<thead>
<tr>
<th>Element</th>
<th>Garnet core (point 1)</th>
<th>Garnet near-rim (point 2)</th>
<th>Garnet rim (point 3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>61.80</td>
<td>62.60</td>
<td>62.87</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.99</td>
<td>1.02</td>
<td>1.03</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>19.52</td>
<td>19.47</td>
<td>19.45</td>
</tr>
<tr>
<td>FeO</td>
<td>6.90</td>
<td>6.02</td>
<td>5.69</td>
</tr>
<tr>
<td>MgO</td>
<td>2.19</td>
<td>2.23</td>
<td>2.23</td>
</tr>
<tr>
<td>MnO</td>
<td>0.07</td>
<td>0.02</td>
<td>0.01</td>
</tr>
<tr>
<td>CaO</td>
<td>0.58</td>
<td>0.45</td>
<td>0.42</td>
</tr>
<tr>
<td>Na₂O</td>
<td>1.18</td>
<td>1.22</td>
<td>1.23</td>
</tr>
<tr>
<td>K₂O</td>
<td>3.53</td>
<td>3.65</td>
<td>3.69</td>
</tr>
<tr>
<td>H₂O</td>
<td>3.23</td>
<td>3.34</td>
<td>3.37</td>
</tr>
<tr>
<td>Total</td>
<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
</tr>
</tbody>
</table>

Note: Points 1, 2, and 3 refer to points shown in Figure 6.

A penetrative transposition foliation (Sₜ) is defined by the preferred alignment of muscovite, biotite, and ribbons of quartz. This foliation wraps tightly around garnet and staurolite porphyroblasts indicating that deformation outlasted growth of these porphyroblasts (Figs. 5A–5D). Garnet ranges from 2 to 4 mm in diameter and is typically rounded and subhedral to euhedral. Garnet is symmetrically chemically zoned from core to rim (Fig. 6), and characterized by a decrease in spessartine and Fe/(Fe + Mg) and an increase in almandine and pyrope components from core to rim, typical of prograde growth zoning (Hollister, 1966; Spear et al., 1990). All components in sample 43, as well as the spessartine component in sample 33, have a flat profile across the garnet core, suggesting rapid early growth over a short P-T interval. Zoning of the grossular component shows slightly more variation in the two samples, either decreasing from core to rim, or characterized by a slight rise outward from the core before decreasing toward the rim. The absence of discontinuities or sharp inflections in the chemical zoning profiles is most consistent with growth during a single metamorphic event. In some analyzed garnet, there is a slight rise in the Mn content at the garnet rim adjacent to quartz, suggestive of a small amount of garnet resorption during retrogression (Kohn and Spear, 2000). Despite the absence of increasing Fe/(Fe + Mg) values at the garnet rim, which is characteristic of retrograde diffusional re-equilibration (Spear, 1991), the possibility remains that the garnet rim Fe/(Fe + Mg) content may have increased by some unknown amount during cooling. A higher-than-peak Fe/(Fe + Mg) composition at the garnet rim may be further compounded if garnet experienced an intermitent period of garnet dissolution at the staurolite isograd, yielding an even larger underestimate of the thermal peak temperature (e.g., Florence and Spear, 1993). Chlorite occurs adjacent to garnet rims, and is interpreted to be retrograde. In a manner similar to the production of staurolite, the resorption of the garnet rim to form retrograde chlorite may result in the loss of the chemical composition reflective of peak metamorphism, resulting in an underestimate of peak metamorphism.

In samples 33 and 43, biotite within 0.4 mm of garnet has retrogressed to chlorite. Matrix biotite further than 0.4 mm from garnet has an unsystematic, slight compositional variation [Fe/(Fe + Mg) = 0.56 – 0.57]. Because biotite composes a small volume of the rock in comparison to garnet, even a small mass flux out of the rim of this Fe-rich garnet during resorption may increase the Fe/(Fe + Mg) content of biotite. Therefore, near-thermal peak temperatures were calculated using garnet-biotite Fe-Mg exchange thermometry (Figs. 7 and 8) with the composition of the least Fe-enriched biotite.

Staurolite occurs as rare 0.5 mm subhedral porphyroblasts with their long axis aligned roughly parallel to Sₜ. Inclusion trails (Sₜ) within staurolite porphyroblasts appear to curve into Sₜ (Fig. 5D) suggesting that staurolite growth was syn-kinematic with respect to Sₜ. The transposition foliation is deflected around staurolite porphyroblasts (Figs. 5C and 5D) indicating that this deformation outlasted staurolite growth. Subhedral to euhedral ilmenite and ilmenite-after-rutile laths occur both as randomly oriented inclusions in garnet and as matrix phases consistently oriented parallel to the foliation.

**P-T PHASE DIAGRAM SECTIONS AND DERIVED P-T PATHS**

Garnet Isohedral Thermobarometry

P-T conditions at successive increments along the prograde P-T path were determined from the intersection of garnet end-member compositional isopleths for core, near-rim, and rim compositions. This thermobarometric method requires that the evolving surface of the growing garnet is continuously in equilibrium with all matrix phases at a scale large enough to be representative of the bulk composition of the rock. However, due to chemical inhomogeneities and sluggish intergranular element diffusivities, the length scale for equilibrium may potentially be too small to satisfy the assumption of equilibrium at this scale (Chernoff and Carlson, 1997; Spear and Daniel, 2001; Hirsch et al., 2003). Most work that has found evidence for disequilibrium has done so by the identification of complex, patchy chemical zoning patterns that exhibit variations in the concentration of a single component that is anomalous, or discordant, with respect to the other end-member components (e.g., Chernoff and Carlson, 1997; Spear and Daniel, 2001; Hirsch et al., 2003). However, Mn profiles within garnet samples used in this study approach a bell-shape profile (Figs. 6A and 6B), and all end-member components vary sympathetically or antithetically in a symmetric manner with respect to one another and the garnet crystal faces. This pattern suggests that these elements were controlled by similar factors in the matrix, suggestive of a state of equilibrium.

Application of this technique to reconstruct a prograde P-T path from successive garnet growth increments is dependent on the evolving geological context.
To quantitatively link the modal proportion of Rayleigh fractionation model (Hollister, 1966) applied by Gaidies et al. (2006), which uses a (Marmo et al., 2002; Evans, 2004; Tinkham 2002), effective bulk compositions affected by the fractionation processes have been proposed to calculate effective core analysis and the increment of garnet growth that is of interest, and then subtracting these values from the bulk composition of the rock. The effective bulk composition calculated for different garnet compositions along a core–rim traverse were then utilized to estimate successive P–T points along the prograde P–T path using the intersection of garnet compositional isopleths. This technique, which relies solely on the garnet composition and the correlative effective bulk composition, provides a novel solution for determining P–T conditions in rocks such as these, for which plagioclase is either absent or low in anorthite content, thus significantly limiting the availability of conventional barometers.

Garnet compositions and the corresponding effective bulk rock composition used for each isopleth thermobarometric estimate are provided in Tables 1 and 2, respectively.

For garnet core compositions, the $X_{\text{Mn}}$ isopleths plot at the point of intersection with the other end members (Figs. 7A and 8A); however, the $X_{\text{Mn}}$ isopleths increasingly diverge from this point of intersection for successive rimward compositions (not shown on Figs. 7B, 7C, 8B, and 8C). This increasing discrepancy of the $X_{\text{Mn}}$ isopleth toward higher temperature and pressure with increasing distance from the garnet core is interpreted to reflect an error in the estimated extent of fractionation within garnet due to analyses of garnet that were not sectioned through the true center of the garnet. Because the spessartine component has the lowest concentration and is partitioned within garnet more strongly than the other end members, an error in the extent of garnet fractionation will have the largest impact on the calculated Mn content. For this reason, the isopleths of $X_{\text{Al}2\text{O}3}$, $X_{\text{Fe}2\text{O}3}$, and $X_{\text{Mn}}$ were used for intersections outside the garnet core.

**Derived P–T Path for Sample 33**

The isochemical P–T phase diagram section calculated for the unfractinated bulk composition of sample 33 (Fig. 7A; Table 2) was used to determine the P–T conditions during the incipient stages of garnet growth. Due to a lack of suitable inclusions within the garnet cores of this sample, the intersection of the compositional isopleths for the garnet core is the only indicator of the P–T conditions during the earliest stage of growth. Because Mn is strongly fractionated within garnet during growth (Hollister, 1966), the earliest stage of garnet growth in the samples of this study is assumed to be the point within the core of the garnet porphyroblast characterized by the highest $X_{\text{Mn}}$ content (point 1, Fig. 6A). The $X_{\text{Si}2\text{O}5}$, $X_{\text{Fe}2\text{O}3}$, and $X_{\text{Mn}}$ isopleths (dashed lines) from this point intersect in the Grt + Chl + Bt + Ms + Pl + Ilm + Qtz + H2O stability field at ~550 °C and 6.6 kbar (Figs. 7A and 7C).

Isoleth calculations for two additional points (points 2 and 3, Fig. 6A) take into account the effects of chemical fractionation during garnet growth. At point 2, the isopleths intersect at ~580 °C and 7.4 kbar (Fig. 7C). The equilibrium assemblage is predicted to have remained unchanged during growth from point 1 to point 2 (phase diagram calculated from the bulk composition of point 2 is not shown). The topology and mineral assemblages at the P–T conditions predicted along the prograde path (Fig. 7B) are similar to those in Figure 7A, with the most notable exception being the significantly decreased stability field for garnet. Previous work has shown that Mn stabilizes garnet to

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**Figure 5. Photomicrographs showing the deflection and tight wrapping of the transposition foliation around pre- to syn-kinematic porphyroblasts of garnet and staurolite in sample 33.**

Inclusion trails within staurolite porphyroblasts appear to curve into the transposition foliation around pre- to syn-kinematic porphyroblasts of garnet and staurolite in sample 33.
are given in Figures 6 and 7, respectively. Pressure-temperature estimates for points 1, 2, and 3 in garnet shown for samples 33 and 43 with photomicrographs showing the location of the microprobe traverse across garnet. Figure 6. Compositional profiles across garnet porphyroblasts of samples 33 (A) and 43 (B), with photomicrographs showing the location of the microprobe traverse across garnet. Pressure-temperature estimates for points 1, 2, and 3 in garnet shown for samples 33 and 43 are given in Figures 6 and 7, respectively.

Figure 6. Compositional profiles across garnet porphyroblasts of samples 33 (A) and 43 (B), with photomicrographs showing the location of the microprobe traverse across garnet. Pressure-temperature estimates for points 1, 2, and 3 in garnet shown for samples 33 and 43 are given in Figures 6 and 7, respectively.

substantially lower temperatures and pressures (Symmes and Ferry, 1992; Mahar et al., 1997; Tinkham et al., 2001). Therefore, the removal of Mn from the effective bulk composition as a result of being strongly fractionated into crystallizing garnet has the effect of progressively raising the lower garnet stability limit to higher temperatures and pressures. The staurolite stability field is slightly narrower at low pressures for this fractionated composition, but is expanded upward by ~0.2–0.3 kbar. For the composition at the garnet rim (point 3, Fig. 6A), the isopleths intersect within the Grt + Chl + Bt + Ms + Pl + Ilm (+ Qtz + H2O) stability field at ~595 °C and 6.1 kbar (Fig. 6A), in good agreement with the P-T estimate for the earliest increment of garnet growth in sample 33. However, for both samples 33 and 43, the isopleth intersection for the earliest increment of garnet growth lies ~15–20 °C above the garnet-in line calculated for the accompanying phase diagram. This discrepancy may simply reflect the uncertainty of the thermobarometric calculations. There are several other possible explanations for this observation. During the earliest stages of garnet growth, when garnet was very small, the length scale of diffusion may have been short enough to re-equilibrate the initial core composition to the continually changing equilibrium composition imposed on the rim. For a rock following a clockwise P-T path, this modification would move the garnet core isopleth intersection toward higher temperatures than experienced during initial garnet growth. Measurement of the garnet composition from an un-centered garnet section would have a similar effect. Additionally, the placement of the garnet core isopleth intersection after the low-temperature stability limit of garnet may be due to some amount of reaction overstepping caused by a kinetic impediment to garnet nucleation, such as limited volume diffusion of Mn. Assuming a larger amount of Fe2O3 in the model chemical system acted to shift the garnet-in line to higher temperature. However, the core isopleth intersection likewise shifted to higher temperatures, ruling out the uncertain Fe2O3 content as the source of discrepancy between the garnet-in line and the intersection of the core isopleths.

The compositions at points 2 and 3 (Fig. 6B) in sample 43 were calculated from the unfractionated bulk rock composition of this sample (Table 2). Garnet compositional isopleths intersect in the Grt + Chl + Bt + Ms + Pl + Ilm (+ Qtz + H2O) stability field at ~545 °C and 6.1 kbar (Fig. 6B), in good agreement with the P-T estimate for the earliest increment of garnet growth in sample 33. However, for both samples 33 and 43, the isopleth intersection for the earliest increment of garnet growth lies ~15–20 °C above the garnet-in line calculated for the accompanying phase diagram. This discrepancy may simply reflect the uncertainty of the thermobarometric calculations. There are several other possible explanations for this observation. During the earliest stages of garnet growth, when garnet was very small, the length scale of diffusion may have been short enough to re-equilibrate the initial core composition to the continually changing equilibrium composition imposed on the rim. For a rock following a clockwise P-T path, this modification would move the garnet core isopleth intersection toward higher temperatures than experienced during initial garnet growth. Measurement of the garnet composition from an un-centered garnet section would have a similar effect. Additionally, the placement of the garnet core isopleth intersection after the low-temperature stability limit of garnet may be due to some amount of reaction overstepping caused by a kinetic impediment to garnet nucleation, such as limited volume diffusion of Mn. Assuming a larger amount of Fe2O3 in the model chemical system acted to shift the garnet-in line to higher temperature. However, the core isopleth intersection likewise shifted to higher temperatures, ruling out the uncertain Fe2O3 content as the source of discrepancy between the garnet-in line and the intersection of the core isopleths.

The compositions at points 2 and 3 (Fig. 6B) are representative of intermediate and final stages of garnet growth, respectively; therefore, the isopleth calculations of these points take into account the effects of chemical fractionation during garnet growth. The isopleths for point 3 intersect at ~595 °C and 7.5 kbar (Fig. 8C) where biotite is added to the assemblage (phase diagram calculated from the bulk composition of point 2 is not shown). Near-peak conditions of ~600 °C and 7.5 kbar result from the calculated isopleths corresponding to the composition of point 3, although almandine and pyrope isopleths very narrowly (~2°) miss intersecting.
Figure 7. Isochemical phase diagram sections and compositional isopleths calculated for different stages of garnet growth in sample 33. (A) Calculated from the unfractionated bulk rock composition that corresponds to the incipient stage of garnet growth (garnet core, point 1 in Fig. 6A). (B) Calculated for the fractionated effective bulk rock composition that corresponds to the late stage of garnet growth (garnet rim, point 3). (C) Garnet compositional isopleths corresponding to the garnet core (point 1—dashed isopleths), near-rim (point 2—dotted isopleths), and rim (point 3—solid isopleths) compositions, calculated from the corresponding appropriate effective bulk compositions. The equilibrium mineral assemblages listed numerically from 1 to 25 correspond to the analogous number within stability fields too small to be directly labeled in A and B.
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**Figure 8. Isochemical phase diagram sections and compositional isopleths calculated for different stages of garnet growth in sample 43.**

(A) Calculated from the unfractionated bulk rock composition that corresponds to the incipient stage of garnet growth (garnet core, point 1 in Fig. 6B). (B) Calculated for the fractionated effective bulk rock composition that corresponds to the late stage of garnet growth (garnet rim, point 3). (C) Garnet compositional isopleths corresponding to the garnet core (point 1—dashed isopleths), near-rim (point 2—dotted isopleths), and rim (point 3—solid isopleths) compositions, calculated from the corresponding appropriate effective bulk compositions. The equilibrium mineral assemblages listed numerically from 1 to 25 correspond to the analogous number within stability fields too small to be directly labeled in A and B.
These conditions lie in the upper thermal stability limit of chlorite within the assemblage of Grt + Chl + Bt + St + Ms + Pl + Ilm (+ Qtz + H2O), ~15 °C lower than the garnet-biotite thermometer results. Similar to sample 33, the presence of rutile inclusions within garnet supports a low P-T origin for rutile prior to the incipient stages of garnet growth at 550 °C and 6.7 kbar.

The Rayleigh fractionation model used here to estimate the effective bulk composition at successive increments of garnet growth assumes a constant garnet-matrix coefficient (K_Mn) for Mn following the incipient stages of garnet growth. Because the equilibrium assemblage does not change until the latest increments of garnet growth, it follows that K_Mn is not affected by changes in phase relations. By contrast, the equilibrium constant between garnet and chlorite (the other most significant Mn-bearing phase) would be expected to vary, and thus change K_Mn to some unknown extent with increasing temperature during garnet growth. However, the tight isopleth intersections from core to rim, and agreement in P-T estimates between samples 33 and 43, suggest that any potential variation in K_Mn had a negligible effect, and the technique appears to yield a reasonable estimate of the physical conditions during progressive stages of garnet growth.

MONAZITE GEOCHRONOLOGY

Methods

Prior to U-Th-Pb analysis, petrographic and back-scattered electron (BSE) images of the in situ monazite grains were obtained to provide insight into their petrological context and internal zoning, identify cracks and mineral inclusions, and guide analytical spot placement (Figs. 9 and 10). In order to better characterize chemical zonation and potential age domains within individual monazite grains, chemical X-ray maps of Y, U, Th, and Ca in strategically selected monazite grains were produced using a Cameca SX50 electron microprobe at the University of Massachusetts operating at a high current (240–260 nA), with small step sizes (0.25–0.62 µm) and rastering of the electron beam. In situ U-Th-Pb analyses using the SHRIMP II at the Geological Survey of Canada in Ottawa were performed on monazite cored from polished thin sections and mounted in epoxy, together with pre-polished monazite standards according to the methods of Rayner and Stern (2002). Targeted areas of monazite were analyzed using a mass-filtered O2+ primary beam focused with a Kohler aperture to a spot measuring 9 × 12 µm. The methods employed followed the SHRIMP analytical protocols described in detail by Stern (1997), Stern and Sanborn (1998), and Stern and Berman (2000). Tera-Wasserburg concordia plots, data regression, and weighted mean calculations were made using the program Isoplot (Ludwig, 2008). Errors assigned to SHRIMP U-Th-Pb ages were determined using numerical propagation of all known sources of error as outlined by Stern (1997), Stern and Sanborn (1998), and Stern and Berman (2000). Uncertainties for individual analyses (ratios, ages, and error ellipses) shown in Figures 10, 11, and 12, Table 3, Table A1 in the GSA Data Repository1, and in the text are presented at the 1σ level.

Figure 9. Representative photomicrographs of samples 33 and 43 showing the alignment of monazite parallel to the transposition foliation (S_T), deflection of S_T around monazite, and the presence of strain caps and shadows around monazite. These textures suggest that monazite grew prior to and/or during the development of S_T. Note that monazite 043-M30 was not analyzed for U-Th-Pb geochronology.

The analyses were corrected for common Pb based on the 206Pb (Table A1) and 207Pb (Table 3) methods following the procedures of Stern and Berman (2000) and Ireland and Gibson (1998), respectively. The two correction methods yield ages indistinguishable within error. However, considering the potential for overcorrection using the 204Pb method due to errors arising from low 208Pb counts, background interference, and a 204 isobar (different ionic species having the same nominal mass), the 206Pb/238U chronometer corrected using the 204Pb method is thought to provide the most meaningful ages for this study. Accordingly, all ages quoted and displayed on the Tera-Wasserburg concordia diagram (Fig. 12) are based on the 206Pb/238U chronometer corrected using the 204Pb method.

Although the 206Pb/238U chronometer is considered ideal for monazite because it is not known to be affected by isotopic disequi-

GSA Data Repository item 2014229, Table A1: sensitive high resolution ion microprobe (SHRIMP) data for monazite corrected using the 204Pb method, is available at http://www.geosociety.org/pubs/ft2014.htm or by request to editing@geosociety.org.

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**Results and Interpretation**

Monazite in the four analyzed samples is typically elongate matrix grains with an average aspect ratio of 2.5:1, and measuring between 70 and 350 µm in their longest dimension. With the exception of monazite M27 and M35, the other nine analyzed monazite grains occur within, and are aligned parallel to, the dominant penetrative transposition foliation (S₁) (Fig. 9; Table 3), which is characteristic of nearly all observed monazite grains. Only one grain is partly enclosed within the rim of a garnet porphyroblast in sample 43.

Most monazite grains consistently display a radially symmetric zoning pattern that parallels the crystal faces and is characterized by high Th in the core that decreases continuously toward a low-Th rim (Fig. 10). This Th zoning is interpreted to reflect Rayleigh fractionation during a single episode of monazite growth (e.g., Kohn and Malloy, 2004; Pyle et al., 2005; Spear et al., 2008). Ca also decreases toward the rim, mimicking very closely the Th zoning pattern (Fig. 10). This similarity between Ca and Th zoning likely reflects a brabantite substitution in which Ca²⁺ charge balances Th⁴⁺ in the substitution for two rare earth element (REE) ions (REE⁵⁺).

Analyses of both core and rim of all monazite grains from all four samples yield ages that range between ca. 169 and 142 Ma (Fig. 11, Table 3), with individual samples also yielding a large spread in ages (up to 20 m.y.). Ages from the high-Th core either agree within error, or are consistently older than those from the low-Th rim within any individual grain (Table 3). However, in some cases the high-Th core of one grain is younger than the low-Th rim of another grain, within any individual sample. This pattern is interpreted to be due to slow intergranular transport of Th, creating a localized equilibration volume that is smaller than the distance between monazite grains within a single thin section.

The Y content is low and homogeneous throughout every monazite grain imaged in this study (Fig. 10). Garnet and xenotime are significant reservoirs of Y in metapelitic rocks, and the prograde growth of garnet is accompanied by the consumption of xenotime (Bea and Montero, 1999; Pyle and Spear, 1999). Given the abundance of garnet porphyroblasts in these rocks, it follows that monazite grown after significant garnet crystallization should be depleted in Y, precisely as observed. The near absence of monazite inclusions in garnet further suggests that monazite grew during the late stages of garnet growth, and hence near the thermal peak of metamorphism.

The consistent alignment of monazite parallel to S₁, and the deflection of S₁ with strain caps and shadows around monazite (Fig. 9) suggest that monazite grew pre- to syn-development of S₁. The Th-rich core, thought to reflect Rayleigh-type fractionation of Th during early stages of monazite growth, further constrains the timing because the zoning consistently parallels the crystallographic long axis. These observations suggest that the early stages of monazite growth occurred within the flattening plane of S₁ deformation, and were therefore syn-kinematic. Synkinematic growth of monazite coeval with the late stages of garnet growth is consistent with the observation that S₁ wraps tightly around garnet (Figs. 5A and 5B), indicating that deformation was also ongoing during the latest stage of garnet growth.
The nearly continuous array of monazite ages in this study (Fig. 11), as well as the homogenous Y zoning and outwardly decreasing Th zoning characteristic of Rayleigh fractionation, suggest a single, protracted episode of monazite growth in at least this part of the Finlayson Lake district. This interpretation is consistent with thermodynamic modeling of monazite growth in a low-Ca metapelite by Spear and Pyle (2010). They suggested that monazite can grow continuously in the subsolidus region for most reasonable metamorphic P-T paths, representing a significant part of the metamorphic episode. In our study, the low-Y monazite and its near absence as inclusions within garnet suggest that monazite growth occurred between an initial period of garnet growth at ~550 °C and 6 kbar and the peak of metamorphism at ~600 °C and 7.5 kbar. Nearly half (45%) of the U-Pb monazite analyses yield ages that cluster between ca. 155 and 152 Ma. This cluster may reflect a period of rapidly changing P-T conditions. Alternatively, this concentration of ages may simply be due to a small sample size, and may disappear with a larger number of analyses. The K-Ar cooling ages on Hbl, Bt, and Ms from rocks at this same structural level indicate that cooling through ~300–500 °C did not occur until the Early to middle Cretaceous (ca. 140–103 Ma; Hunt and Roddick, 1992).

**DISCUSSION**

**Absence of Older Permo-Triassic and Early Jurassic Metamorphism**

The samples from the Finlayson Lake district record no evidence of the Permo-Triassic tectono-metamorphism that is so prominent in Yukon within the Yukon-Tanana terrane southwest of the Tintina fault (Berman et al., 2007; Beranek and Mortensen, 2011). It is possible that Permian monazite was present, but was later consumed during retrogression or via a dissolution-reprecipitation process to form new monazite in the Middle to Late Jurassic. Alternatively, monazite paragenesis would have depended upon the location within the thermal structure of the orogen. For instance, the Permian amphibolite-facies metamorphism in west-central Yukon was coeval with, and is cospatial with, the arc that produced the Middle to Late Permian (~220 Ma) Klondike assemblage and the inferred distribution of Permo-Triassic rocks (ca. 260–240 Ma; Berman et al., 2007) amphibolite-facies metamorphism. As stated above, prior to offset along the Tintina fault in the Paleogene, and Cretaceous extension (Gabrielse et al., 2006), a belt of mid-Permian blueschist- and eclogite-facies rocks lay northeast of the Permian Klondike arc along the eastern margin of the Yukon-Tanana terrane (Erdmer et al., 1998; Creaser et al., 1997). Together, the coeval Klondike assemblage and belt of high P-T rocks reflects a Middle to Late Permian northeast-facing convergent margin (Nelson et al., 2006; Mortensen, 1992). The absence of a Permian metamorphic signature and the position of the Finlayson Lake district east of the Klondike arc center and west of high P-T subduction-complex eclogites suggest they occupied a position in the cool, brittle forearc region (Fig. 13A).

Additionally, the samples of this study show no record of an Early Jurassic metamorphic event. There are a few localized Early Jurassic plutons ~30 km to the north (Fig. 3; Murphy et al., 2001); however, these plutons are situated in a higher structural panel, positioned in the hanging wall of two normal faults that separate these plutons from the deeper structural levels of the study area. Likewise, a 201 Ma K-Ar muscovite age obtained by Stevens et al. (1982) comes from the hanging wall of the North River fault, and was thus in a different (higher) structural panel than the study area (footwall) prior to the mid-Cretaceous. Unlike the study area in the footwall, the hanging wall preserves Mississippian K-Ar cooling ages and was thus never significantly buried and heated after that time.

Contrary to the Finlayson Lake district, the present lack of evidence for Permian metamorphism within the Yukon-Tanana terrane in southwest Yukon (Aishihik region) may simply be due to the fact that the correct technique (e.g., dating of a high-T metamorphic chronometer such as monazite) to elucidate this information has not yet been applied to the region. There is a strong Jurassic signal in the Aishihik region (Johnston et al., 1996), but nothing precludes an earlier history in the Permian. For example, the Stewart River area of the Yukon-Tanana terrane is blanketed by Jurassic K-Ar cooling ages (Hunt and Roddick, 1992), but it was not until Berman et al. (2007) dated monazite and metamorphic zircon rims that Permian metamorphism was known.

**Revealing a Younger, More Protracted Metamorphism**

The identification of Middle Jurassic to Early Cretaceous (ca. 169–142 Ma) ductile deformation and high-P, moderate-T (~7.5 kbar, 600 °C) metamorphism within the Finlayson Lake district reveals a protracted and diachronous defor-
Middle Jurassic to earliest Cretaceous tectono-metamorphism in the northern Canadian Cordillera

In the Early to Middle Triassic, detrital zircons with provenance ties to the Paleozoic arc assemblages of the Yukon-Tanana terrane were deposited unconformably on the ancestral North American continental margin (Beranek and Mortensen, 2011), suggesting that the Yukon-Tanana terrane was on the North American continental margin and its detritus was deposited on it across the intervening Permo-Triassic suture zone. A 9 kbar and 600 °C metamorphic event is recorded within the Yukon-Tanana terrane west of the Finlayson Lake district between ca. 260 and 239 Ma (Beranek et al., 2007). This metamorphic event, and ca. 260 and 253 Ma U-Pb crystallization ages of syn- and post-kinematic Permian intrusive rocks, respectively, are interpreted by Beranek and Mortensen (2011) to record the accretion of the Yukon-Tanana terrane onto the western ancestral margin of North America. In this same domain west of the Finlayson Lake district, Beranek et al. (2007) documented an Early Jurassic (ca. 195 Ma) metamorphic overprint characterized by a path of increasing pressure from 5.4 to 7.6 kbar (~25 km depth). This Early Jurassic burial and metamorphism followed, or was partly contemporaneous with, the emplacement of Early Jurassic (ca. 205–197 Ma; Tafti, 2005; Hood, 2012; Knight et al., 2013) granitic intru-

![Figure 12. Tera-Wasserburg plot of samples 32, 33, 43 and 90 showing isotopic data uncorrected for common Pb. Error ellipses represent 1σ level of uncertainty, with the ellipses of each sample shown with a different line symbol.](image)

### TABLE 3. SENSITIVE HIGH-RESOLUTION ION MICROPROBE U-Th-Pb ANALYTICAL DATA FOR MONAZITE

<table>
<thead>
<tr>
<th>Spot*</th>
<th>Grain location</th>
<th>Texture †</th>
<th>Th§</th>
<th>206Pb/238U</th>
<th>f(206)207 (%)</th>
<th>Total 206Pb/238Pb</th>
<th>Total 238U/206Pb</th>
<th>Age**</th>
</tr>
</thead>
</table>
| Sample 32
| M20.1 | I to Sₚ | Core High | 10.0 | 1.189E-03 ± 1.533E-04 | 2.25 | 0.0640 ± 0.0008 | 38.710 ± 0.427 | 161.4 ± 1.8 |
|       | M20.2 | I to Sₚ | Rim Low | 8.6 | 1.242E-03 ± 2.262E-04 | 2.66 | 0.0673 ± 0.0017 | 40.684 ± 0.437 | 153.0 ± 1.7 |
| M21.1 | I to Sₚ | N.A. Hom | 2.4 | 4.578E-04 ± 3.656E-05 | 1.04 | 0.0544 ± 0.0003 | 41.566 ± 0.461 | 152.3 ± 1.7 |
| M21.2 | I to Sₚ | N.A. Hom | 7.4 | 2.515E-03 ± 6.728E-04 | 3.99 | 0.0780 ± 0.0050 | 40.230 ± 0.467 | 152.6 ± 1.8 |
| Sample 33
| M23.3 | I to Sₚ | Rim Low | 8.7 | 2.634E-03 ± 3.315E-04 | 5.27 | 0.0882 ± 0.0087 | 39.320 ± 0.640 | 154.0 ± 3.0 |
| M23.4 | I to Sₚ | Core High | 14.9 | 1.841E-03 ± 2.302E-04 | 3.92 | 0.0774 ± 0.0028 | 38.877 ± 0.459 | 158.9 ± 1.9 |
| M24.3 | I to Sₚ | Core High | 15.3 | 4.657E-03 ± 6.102E-04 | 6.80 | 0.1004 ± 0.0050 | 36.559 ± 0.574 | 162.9 ± 2.8 |
| M24.4 | I to Sₚ | Rim Low | 5.5 | 6.598E-03 ± 1.143E-03 | 10.78 | 0.1321 ± 0.0131 | 40.197 ± 0.406 | 142.0 ± 3.0 |
| M25.1 | I to Sₚ | Rim Low | 15.0 | 4.207E-03 ± 3.363E-04 | 6.08 | 0.0947 ± 0.0050 | 40.146 ± 0.583 | 149.6 ± 2.4 |
| M25.2 | I to Sₚ | Core High | 14.7 | 2.655E-03 ± 3.659E-05 | 6.09 | 0.0948 ± 0.0046 | 38.970 ± 0.619 | 154.1 ± 2.6 |
| M27.3 | I to Sₙ | Rim Low | 8.1 | 7.880E-03 ± 6.435E-04 | 12.78 | 0.1481 ± 0.0194 | 38.211 ± 0.396 | 146.0 ± 3.2 |
| Sample 43
| M30.3 | I to Sₚ | Rim Low | 7.6 | 2.906E-04 ± 1.136E-04 | 1.31 | 0.0566 ± 0.0006 | 38.469 ± 0.396 | 164.0 ± 1.7 |
| M31.3 | Garnet | Rim Low | 7.5 | 5.233E-04 ± 1.122E-04 | 1.43 | 0.0575 ± 0.0011 | 38.392 ± 0.515 | 164.1 ± 2.2 |
| M31.4 | Garnet | Core High | 13.0 | 1.226E-03 ± 2.077E-04 | 2.36 | 0.0650 ± 0.0013 | 36.820 ± 0.462 | 169.4 ± 2.1 |
| M35.3 | I to Sₚ | Rim Low | 11.6 | 1.026E-03 ± 1.219E-04 | 1.73 | 0.0599 ± 0.0007 | 38.493 ± 0.385 | 163.2 ± 1.6 |
| M36.3 | I to Sₚ | Rim Low | 8.1 | 3.207E-03 ± 2.923E-04 | 6.27 | 0.1328 ± 0.0171 | 39.852 ± 0.562 | 154.3 ± 2.3 |
| M36.4 | I to Sₚ | Core High | 14.2 | 7.970E-03 ± 1.247E-03 | 12.95 | 0.1496 ± 0.0171 | 36.142 ± 0.781 | 154.0 ± 5.0 |
| Sample 90
| M38.3 | I to Sₚ | Core High | 9.7 | 7.814E-04 ± 1.188E-04 | 1.52 | 0.0582 ± 0.0007 | 40.649 ± 0.418 | 154.9 ± 1.6 |
| M38.4 | I to Sₚ | Rim Low | 8.0 | 8.343E-04 ± 1.317E-04 | 1.62 | 0.0630 ± 0.0008 | 39.481 ± 0.650 | 159.3 ± 2.6 |
| M41.2 | I to Sₚ | N.A. Hom | 9.8 | 1.205E-03 ± 1.362E-04 | 2.12 | 0.0630 ± 0.0007 | 40.458 ± 0.435 | 154.7 ± 1.7 |

Note: Uncertainties reported at 1σ (absolute) and are calculated by numerical propagation of all known sources of error (Stern and Berman, 2000).

*Spot: example: M20.2 = second spot on monazite grain 20.

†Texture: location of monazite as inclusion in garnet, or as an elongate matrix grain (I to Sₚ = parallel to foliation, I to Sₙ = at a high angle to foliation).

§Th: relative concentration from X-ray map at the spot of the analysis. The "Grain location" of the analyses is not applicable (N.A.) for monazite with homogenous (hom) Th-zoning.

f(206)207 refers to the fraction of total 206Pb that is common Pb, calculated using the 207Pb method.

**Ages have been corrected for common Pb using the 206Pb method.

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sions at moderate depths (14–23 km; McCausland et al., 2002; Tafti, 2005). Widespread Early Jurassic K-Ar and $^{40}$Ar/$^{39}$Ar ages in this same domain west of the Finlayson Lake district are interpreted to record either cooling and exhumation associated with this contraction (Dusel-Bacon et al., 2002), or exhumation accommodated by a shift to extensional tectonics (Berman et al., 2007; Knight et al., 2013).

Prior to this study, there had been little evidence (Berman et al., 2007; Staples et al., 2013) for continued deep-seated tectono-metamorphism within the hinterland of the northern Cordilleran orogen following this Early Jurassic cooling event. Further south, in southwest Yukon and northern British Columbia, Middle Jurassic K-Ar and $^{40}$Ar/$^{39}$Ar mica ages have been attributed to cooling and exhumation following the southwestward obduction of the Cache Creek terrane onto the Stikine terrane (Johnston et al., 1996; Mihalynuk et al., 2004). However, it is uncertain to what extent these events affected rocks in the Finlayson Lake district 350–500 km along strike, and to the north, of the most northerly exposure of the Cache Creek terrane prior to movement along the Tintina fault (Nelson and Colpron, 2007).

Figure 13. Schematic crustal sections (not drawn to scale) of the northern Cordilleran margin showing the interpreted crustal position of the Finlayson Lake and other metamorphic domains before and during Mesozoic deformation and metamorphism of the Yukon-Tanana terrane. (A) Late Permian. The Finlayson Lake domain lay in the forearc region with respect to the Klondike domain. The Australia Mountain domain is inferred to have originated in distal extension of the ancestral North American (NAm) margin prior to underthrusting beneath the Yukon-Tanana terrane. This time frame post-dates extreme thinning of the Yukon-Tanana terrane in the latest Middle Permian (Johnston et al., 2007). (B) Late Jurassic–Early Cretaceous (ca. 169–142 Ma), prior to Paleogene offset along the Tintina fault and mid-Cretaceous extension within the Yukon-Tanana terrane, and after the westward underthrusting of the Yukon-Tanana terrane that produced the deformation and metamorphism recorded in the Finlayson Lake district, and initial shortening in the Selwyn fold belt. At that time, the Australia Mountain domain is inferred to have begun westward underthrusting beneath the Yukon-Tanana terrane en route to peak metamorphism of 9 kbar at ca. 146–118 Ma (Staples et al., 2013).
Data from the Finlayson Lake district reveal renewed tectonic burial, ductile deformation, and high-grade metamorphism in the northern Cordillera from the Middle Jurassic to Early Cretaceous. Therefore, while Yukon-Tanana rocks metamorphosed in the Permo-Triassic and again in the Early Jurassic occupied a high structural level from Early Jurassic onward, rocks to the northeast in the Finlayson Lake district occupied a lower structural level, being buried, heated, and ductily deformed at mid-crustal levels (~25 km depth) from the Middle Jurassic to earliest Cretaceous (ca. 169–142 Ma). Metamorphism continued at an even deeper crustal level (~30 km depth, as recorded in the Australia Mountain domain), propagating downward into the parautochthonous North American crust in the Early to mid-Cretaceous (ca. 146–118 Ma; Berman et al., 2007; Staples et al., 2013). This Early to mid-Cretaceous domain is exposed in a tectonic window juxtaposed against the Permo-Triassic/Early Jurassic metamorphic domain to the southwest across the Australia Creek normal fault and bounded on its northeast by the Tintina fault (Fig. 2; Staples et al., 2013). This domain is comparable to parautochthonous North American continental margin rocks (Lake George assemblage in Fig. 1) in east-central Alaska, which were exhumed from beneath the Yukon-Tanana terrane along mylonitic extensional faults in the mid-Cretaceous (Hansen, 1990; Hansen and Dusel-Bacon, 1998; Dusel-Bacon et al., 1995, 2002, 2006). Together, these data reveal a spatial and temporal pattern of structurally downward-younging deformation and metamorphism.

Additionally, the apparent absence of Permo-Triassic and Early Jurassic metamorphism in the younger and deeper structural domains (Finlayson Lake district and Australia Mountain) suggests that deformation and metamorphism migrated progressively into cooler rocks. This suggests that the downward propagation of deformation and metamorphism corresponded with the foreland-directed growth of the orogenic wedge. As the orogenic wedge propagated toward the foreland, cooler rocks in front of the wedge were progressively buried and heated as they were underthrust beneath the warmer overriding wedge to the west.

Despite this apparent foreland-directed Younging of deformation and metamorphism, the ca. 146–118 Ma Australia Mountain metamorphic domain lies southwest (outboard) of the older (ca. 169–142 Ma) Finlayson Lake metamorphic domain (Fig. 2). However, the Australia Mountain domain is interpreted to have originally lain east of the Finlayson Lake domain (Yukon-Tanana terrane) prior to accretion, based on the interpretation that it represents parautochthonous North American margin rocks. The present exposure of the Australia Mountain domain west of the Finlayson Lake district is attributed to underthrusting down to the west beneath the Finlayson Lake district in the Early Cretaceous, following burial and metamorphism of rocks of the Finlayson Lake district. K-Ar cooling ages (Hunt and Roddick, 1992) and retrograde monazite ages (Staples et al., 2013) reveal that both the Australia Mountain and Finlayson Lake metamorphic domains did not cool appreciably until they were exhumed along the mid-Cretaceous Australia Creek, Stewart River, and North River faults (Fig. 2; Murphy, 2004; Staples et al., 2013). Focused exhumation along the Australia Creek and Stewart River faults relative to the North River fault can explain the exposure of the youngest and deepest crustal level found in the Australia Mountain domain (9 kbar metamorphism) west (outboard) of the Finlayson Lake district (7.5 kbar metamorphism). Comparable exhumation of parautochthonous rocks of the Lake George assemblage occurred <200 km to the east in east-central Alaska (Dusel-Bacon et al., 1995, 2002; Hansen and Dusel-Bacon, 1998).

The Middle Jurassic to Early Cretaceous metamorphic event recorded in the Finlayson Lake district in the northern Cordillera was contemporaneous with amphibolite-facies metamorphic events recorded in the southern Cariboo and northern Monashee Mountains in the southeastern Canadian Cordillera (ca. 135 Ma: Currie, 1988; ca. 140 Ma: Digel et al., 1998; ca. 163 Ma: Crowley et al., 2000; ca. 148 Ma: Reid, 2003; ca. 153 Ma: Gervais and Hynes, 2012). Furthermore, this pattern of structurally downward-younging deformation and metamorphism described above is strikingly similar to, and in part contemporaneous with, that of the southeastern Canadian Cordillera. In the southeastern Canadian Cordillera, deformation and metamorphism progressed from the Early Jurassic to Eocene (Evenchick et al., 2007, and references therein), with younger events recorded at progressively deeper crustal levels (Parrish, 1995; Simony and Carr, 2011). There, rocks presently in the upper structural levels were buried, heated, and exhumed in the Jurassic (Murphy et al., 1995; Colpron et al., 1996; Crowley et al., 2000; Gibson et al., 2005, 2008), while structurally deeper levels continued to be buried and heated from the Cretaceous to earliest Eocene (Carr, 1991; Parrish, 1995; Gibson et al., 1999, 2005, 2008; Crowley and Parrish, 1999; Crowley et al., 2000; Simony and Carr, 2011). This pattern has been attributed to progressive structural burial and underplating of cooler rocks to the east as deformation migrates toward the craton (Parrish, 1995; Brown, 2004; Simony and Carr, 2011).

We propose that a similar process was active from Middle Jurassic to mid-Cretaceous time in the northern Cordillera. The foreland-directed and downward migration of deformation and metamorphism is interpreted to have been driven by continued migration and underthrusting of the North American continent from the east, together with the accretion and underthrusting of the Insular terranes (Peninsular-Alexander-Wrangellia) beneath the western side of the Yukon-Tanana terrane in the Middle to Late Jurassic (Fig. 13B; McClelland et al., 1992; van der Heyden, 1992; Saleeb, 2000; Gehrels, 2001; Trop and Ridgway, 2007). These results suggest that analogous orogenic processes may have been operating contemporaneously over 1000 km along strike during the development of the Canadian Cordillera, and is a model that may have implications for unravelling the history of other orogenic belts around the world.

CONCLUSIONS

In situ SHRIMP monazite geochronology and mineral equilibria modeling reveal a previously unrecognized Middle Jurassic to earliest Cretaceous tectono-metamorphic event recorded in the Finlayson Lake district of the eastern Yukon-Tanana terrane. The results of garnet isopleth thermobarometry applied to successive increments of single-stage, growth-zoned garnet record progressive growth from 550 °C and 6.1–6.6 kbar to 600 °C and 7.5 kbar. Monazite textures, chemical zoning, and in situ U-Pb ages record a protracted episode of monazite growth from ca. 169 to 142 Ma coeval with the development of transposition fabrics and the late stages of garnet growth. The absence of evidence for Permo-Triassic ductile deformation and metamorphism that is characteristic of the Yukon-Tanana terrane in Yukon further to the west, and its position between the Permian arc axis to the west and the belt of Permian blueschists and eclogites to the east, suggest that the Finlayson Lake district occupied the cold, brittle forearc region in the Permian. This eastern part of the Yukon-Tanana terrane was not buried and heated sufficiently to grow metamorphic monazite until the Middle Jurassic to earliest Cretaceous, revealing a younger, more protracted and diachronous mid-crustal tectono-metamorphic history within the hinterland of the northern Cordillera than was previously recognized. These data, together with the identification of an Early Cretaceous tectono-metamorphic domain in adjacent and structurally deep (9 kbar) parautochthonous North American rocks to the west (Australia Mountain domain; Staples et al., 2013), reveal that deformation and metamorphism migrated toward the foreland and
structurally downwards in the Middle Jurassic to Early Cretaceous. This pattern and timing are similar to and coeval with events documented in the southeastern Canadian Cordillera, which have been attributed to the progressive burial of cooler rocks to the east as deformation migrated toward the craton (Parrish, 1995; Brown, 2004; Gibson et al., 2008; Simony and Carr, 2011). Deformation and metamorphism is likewise considered to have migrated toward the craton and downwards in the Middle Jurassic to Early Cretaceous in the northern Cordillera as the oxygen propagated toward the foreland above the westward-underthrusting North American continent.

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