Metamorphism and deformation of the Grand Forks complex: implications for the exhumation history of the Shuswap core complex, southern British Columbia

Joel F. Cubley and David R.M. Pattison

Abstract: The Grand Forks complex (GFC) is an elongate, north–south-trending metamorphic core complex in the Shuswap domain of southeastern British Columbia. It comprises predominantly upper-amphibolite- to granulite-facies paragneisses, schists, orthogneisses, amphibolites, and calc-silicates of the Paleoproterozoic to Paleozoic Grand Forks Group. The GFC is juxtaposed against low-grade rocks of the Quesnel terrane across two bounding Eocene normal faults: the Kettle River fault (KRF) on the east flank and the Granby fault (GF) on the west flank. Peak metamorphic Sil + Kfs ± Grt ± Crd (Sil, sillimanite; Kfs, potassium feldspar; Grt, garnet; Crd, cordierite) assemblages in paragneiss and Hbl ± Opx ± Cpx (Hbl, hornblende; Opx, orthopyroxene; Cpx, clinopyroxene) assemblages in amphibolite in the GFC formed at 750 ± 25 °C, 5.6 ± 0.5 kbar (1 kbar = 100 MPa; 20 ± 2 km depth). Stratigraphically overlying Sil + St-bearing pelitic schists (St, staurolite) within the complex record peak conditions of 600 ± 15 °C, 5.5 ± 0.25 kbar. Crd + Ilm + Spl (Crd, cordierite; Ilm, ilmenite; Spl, spinel) and Crd + Qtz (Qtz, quartz) coronal textures in paragneiss, and Cpx + Opx + Pl + Mt (Pl, plagioclase; Mt, magnetite) symplectites in amphibolite, formed at 735 ± 20 °C, 3.3 ± 0.5 kbar, indicating high-temperature, near-isothermal decompression of the GFC of ~2.3 ± 0.7 kbar (~8.2 ± 2.5 km) from peak conditions. Transitional greenschist–amphibolite metamorphic assemblages in the hanging wall of the KRF indicate conditions of ~425 ± 25 °C and 2.2 ± 0.6 kbar (~8 ± 2 km depth), with local contact metamorphism around Jurassic intrusions as high as 630–650 °C at ~2.5 ± 0.5 kbar. The pressure contrast across the Kettle River fault prior to greenschist facies displacement was ~0.8 ± 0.7 kbar, for a vertical offset of ~2.9 ± 2.5 km. This is similar to estimates for the Granby fault on the west flank of the GFC. The GFC therefore experienced a two-stage exhumation history: early high-temperature decompression at upper-amphibolite–to granulite-facies conditions, followed by low-temperature exhumation at greenschist-facies conditions owing to movement on the Eocene Granby and Kettle River faults.

Résumé: Le complexe de Grand Forks (GFC) est un complexe à noyau métamorphique; il a une forme allongée, de tendance nord-sud, et il est situé dans le domaine Shuswap au sud-est de la Colombie-Britannique. Il comprend principalement des paragneisses, des schistes, des orthogneisses, des amphibolites et des calc-silicates du Paléoproterozoïque au Paléozoïque des Grand Forks. Le GFC est juxtaposé contre des roches peu métamorphisées du territoire de Quesnel travers de deux failles normales datant de l’Éocène : la faille de Kettle River sur le flanc est et la faille de Granby sur le flanc ouest. Les assemblages du métamorphisme de crête, Sil + Kfs ± Grt ± Crd, dans les paragneisses, et les assemblages Hbl ± Opx ± Cpx dans les amphibolites du GFC se sont formés à des températures et à des pressions de 750 ± 25 °C et de 5.6 ± 0.5 kbar (1 kbar = 100 MPa; à une profondeur de 20 ± 2 km). Recouvrant stratigraphiquement ces assemblages, des schistes péliquités à Sil + St dans le complexe enregistrent des conditions de crête de 600 ± 15 °C et 5.5 ± 0.25 kbar. Des textures coronales Crd + Ilm + Spl et Crd + Qtz dans les paragneisses et des symplectites Cpx + Opx + Pl + Mt dans les amphibolites, sont formées à 735 ± 20 °C, 3.3 ± 0.5 kbar, indiquant une décompression à température élevée, quasi-isotherme, du GFC à ~2.3 ± 0.7 kbar des conditions de crête (à une profondeur ~8.2 ± 2.5 km). Des assemblages transitionnels métamorphiques au faciès des schistes verts-amphibolites dans l’épisode supérieur de la faille de Kettle River indiquent des conditions de ~425 ± 25 °C et 2.2 ± 0.6 kbar (à une profondeur ~8 ± 2 km), le métamorphisme de contact local autour des intrusions jurassiques atteignant 630–650 °C à une pression ~2.5 ± 0.5 kbar. La différence de pression en travers de la faille de Kettle River, avant le déploiement au faciès des schistes verts était ~0.8 ± 0.7 kbar, donnant un décalage vertical ~2.9 ± 2.5 km. Cela est semblable aux estimées pour la faille de Granby sur le flanc ouest du GFC. Le complexe de Grand Forks a donc subi un historique d’exhumation en deux étapes : une première décompression à température élevée sous des conditions de faciès des amphibolites supérieur au faciès des granulites puis une exhumation à basse température sous des conditions de faciès des schistes verts en raison de mouvement sur les failles de Granby et de Kettle River à l’Éocène.

[Traduit par la Rédaction]
Introduction

The Grand Forks complex (GFC; Brock 1903) is a fault-bounded metamorphic core complex in southeastern British Columbia (Fig. 1). It is known as the Kettle dome south of the United States – Canada border (Cheney 1980). The Grand Forks complex exposes a thick Paleoproterozoic to early Paleozoic, highly metamorphosed sedimentary package of North American affinity (Preto 1970b; Armstrong et al. 1991; Höy and Jackaman 2005). The complex is bounded on the east by the east-dipping Kettle River fault and on the west by the west-dipping Granby fault, both Eocene ductile-to-brittle normal faults (Figs. 1, 2). Juxtaposed across these faults are volcanic, volcanioclastic, sedimentary, and minor intrusive rocks of low metamorphic grade belonging to the pericratonic Queuel terrane (Fyles 1990; Acton et al. 2002; Unterschutz et al. 2002).

The GFC is one of a number of core complexes within the southern Omineca belt, the crystalline hinterland of the Cordilleran orogen. In southeastern British Columbia, Washington State, and Idaho these are known collectively as the Shuswap metamorphic core complex (SMC) (e.g., Monger et al. 1982). The Shuswap complex experienced late-Mesozoic to Paleocene compressional deformation during Cordilleran orogenesis, followed by regional extension in the Eocene (Parrish et al. 1988; Johnson and Brown 1996; Vanderhaeghe et al. 1999; Teyssier et al. 2005). Unlike nearby complexes such as the Monashee complex (e.g., Norlander et al. 2002; Teyssier et al. 2005), Valhalla complex (e.g., Spear et al. 2002), the Shuswap complex experienced late-Mesozoic to Paleocene compressional deformation during Cordilleran orogenesis, followed by regional extension in the Eocene (Parrish et al. 1988; Johnson and Brown 1996; Vanderhaeghe et al. 1999; Teyssier et al. 2005). Unlike nearby complexes such as the Monashee complex (e.g., Norlander et al. 2002; Teyssier et al. 2005), Valhalla complex (e.g., Spear et al. 2002), the Shuswap complex experienced late-Mesozoic to Paleocene compressional deformation during Cordilleran orogenesis, followed by regional extension in the Eocene (Parrish et al. 1988; Johnson and Brown 1996; Vanderhaeghe et al. 1999; Teyssier et al. 2005).

The purpose of the current study is to extend the work of Laberge and Pattison (2007) by investigating metamorphic and deformational relationships across the eastern bounding normal fault, the Kettle River fault (KRF), and characterizing the metamorphic history of the core complex across a broad region between Christina Lake and the Granby River, corresponding roughly to the NTS 082E/01 1:50 000 map sheet. Metamorphic investigations in both the GFC and the KRF hanging wall focus on pelitic and basic lithologies, as these provide the most robust P–T (pressure–temperature) constraints. Information regarding other lithologies, e.g., calc-silicates, can be found elsewhere (e.g., Laberge 2005; Laberge and Pattison 2007; Cubley and Pattison 2009). This study serves as a framework for two forthcoming geochronological studies, Cubley et al. (a, b, accepted for publication), on the timing of peak metamorphism, exhumation, and cooling of the Grand Forks complex.

Regional geology

The Grand Forks complex forms a broad domal culmination that plunges shallowly to the south, with the deepest stratigraphic levels exposed in British Columbia and stratigraphically higher rocks exposed south of the border in Washington State (Preto 1970b; Cheney 1980; Orr and Cheney 1987; Fig. 3). The GFC comprises a succession of highly deformed and metamorphosed Proterozoic to lower Paleozoic sedimentary units (paragneiss, schist, quartzite, marble, and calc-silicate) mapped by Preto (1970b) as the Grand Forks Group. Interlayered with the metasedimentary rocks are Proterozoic to lower Paleozoic amphibolites, pegmatites, and orthogneisses that have been coevally deformed and metamorphosed. Representative field photographs are shown in Fig. 4, and a map of associated outcrop locations is shown in Fig. 5. A major Proterozoic unconformity exists between the stratigraphically lowest Paleoproterozoic metasedimentary unit (Pr1) and the overlying Neoproterozoic Pr2 quartzite unit (Höy and Jackaman 2005). Stratigraphically overlying the quartzite marker unit is a Neoproterozoic to lower Paleozoic metasedimentary unit (Pr3) comprising paragneiss, pelitic schist, calc-silicate, marble, and amphibolite of remarkably similar character to that found below the unconformity (Höy and Jackaman 2005). In the southwest corner of the study area, these lower units (Pr1–Pr3) of the Grand Forks Group are overlain by a cover sequence of lower metamorphic grade comprising Neoproterozoic to lower Paleozoic pelitic schist and amphibolite (Pr4, Pr5) (Preto, 1970b; Orr and Cheney 1987; Höy and Jackaman 2005).

All of the units just discussed are cut by steeply dipping, granitic to monzonitic intrusives that postdate ductile deformation within the complex. These have been previously correlated with the regionally extensive, composite Jura-Cretaceous Okanagan Batholith (Tempelman-Kluit 1989; Höy and Jackaman 2005) but may be Eocene in age based on mineralogical, textural, and geochemical similarities to Eocene intrusives of the Colville Batholith in northeastern Washington (Holder and McCarley Holder 1988; Laberge 2005). Metamorphic grade within the core complex mostly ranges from upper-amphibolite-facies (Sil + Kfs ± Grt (Sil, sillimanite, Kfs, potassium feldspar, Grt, garnet) in metapelites; Hbl + Pl ± Cpx (Hbl, hornblende; Pl, plagioclase; Cpx, clinopyroxene) in metabasites (mineral abbreviations after Kretz 1983)) to granulite-facies (Grt + Kfs + Crd (Crd, cordierite) in metapelite; Cpx + Opx (Opx, orthopyroxene) in metabasites). An exception is a domain of middle-amphibolite-facies
(Sil + St zone; St, staurolite) conditions in the uppermost stratigraphic level (Pr5) exposed in British Columbia (Preto, 1970b; this study).

In the hanging wall of both the Granby and Kettle River fault normal faults are low-grade metamorphic rocks of the pericratonic Quesnel terrane (Quesnellia). These are interpreted to be rocks of island arc affinity that were obducted onto the western margin of North America in the Middle Jurassic (Høy and Dunne 1997; Erdmer et al. 2001; Unterschutz et al. 2002). In the hanging wall of the KRF in British Columbia, the rocks consist of a sedimentary, volcanic, and volcanoclastic succession deposited during Carboniferous to Early Jurassic time (Fig. 2). This stratigraphy is dominated by pelites, carbonates, and volcanics of the Carboniferous–Permian Knob Hill Group, sequentially overlain by Triassic volcanics, clastics, and limestones of the Brooklyn Formation (Nicola Group) and Jurassic volcanics of the Rossland Group (Fyles 1990; Höy and Jackaman 2005; Laberge and Pattison 2007). These are capped by Eocene volcanic, volcanoclastic, and syn-rift clastic rocks associated with extension and graben formation (Suydam and Gaylord 1997). Two main intrusive suites were emplaced into the Quesnel country rocks of both hanging walls: granites and granodiorites correlated with the ~173–159 Ma Middle Jurassic Nelson Suite (e.g., Parrish 1992; Ghosh 1995) and syenites and monzonites of the 51.1 ± 0.5 Ma Eocene Coryell Batholith (Carr and Parkinson 1989; Acton et al. 2002). Additional intrusive suites in the KRF hanging wall include diorites of the 215.9 ± 1.4 Ma Triassic Josh Creek diorite and leucogranites correlated with the 62–52 Ma Ladybird intrusive suite (Acton et al. 2002).

The Granby and Kettle River normal faults are low-temperature extensional features related to orogen-wide extension that began in the early Eocene (Parrish et al. 1988, Mulch et al. 2007). The Kettle River fault has been best characterized in northeastern Washington, where mylonitic fabrics in the footwall are locally cut by subparallel brittle faults (Rhodes and Cheney 1981, Mulch et al. 2007). On the western margin of the core complex, the Granby fault is a shallowly west-dipping listric normal fault that displays predominantly brittle fabrics, but thin isolated footwall mylonites have been documented (Preto, 1970b; Carr and Parkinson 1989; Fyles 1990; Laberge and Pattison 2007). The timing of latest movement on both the Granby fault and KRF is constrained to be after 51.1 ± 0.5 Ma, as both faults cut Coryell intrusives of that age (Carr and Parkinson 1989; Acton 1998).

**Structure**

**Grand Forks complex**

The main structural elements in the Grand Forks complex are summarized in Fig. 6. The earliest structural fabrics observed are straight to locally folded inclusion trails in garnet and staurolite porphyroblasts (S1), commonly oriented at a high angle to the enclosing gneissic fabric. The predominant structural fabric in the Grand Forks complex is a well-defined gneissosity (S2) that is commonly subparallel to sedimentary layering (S0) and is manifested as mineralogically segregated gneissic layering, preferred mineral orientation, and layer-parallel stromatic leucosomes. This gneissosity is seen in all units within the Grand Forks Group, as well as interlayered orthogneisses and amphibolites (Fig. 4). Minor intrafolial folds (F2) are developed within the gneissosity. In the upper units, Pr4 and Pr5, the gneissosity is replaced by a strong schistosity also deemed to be S2.

Locally developed high-strain shear zones subparallel to S2 fabrics are seen on the scale of metres to tens of metres across the core complex. Kinematic analysis of fabrics within these zones typically indicates top-to-the-east shear, but isolated zones of antithetic shear have been observed. There is no apparent increase in the abundance or degree of deformation within these shear zones with proximity to the bounding Kettle River and Granby normal faults.

Mineral lineations (L2) measured within the S2 gneissic foliation (sillimanite, hornblende, and quartz) trend west-northwest–east-southeast, similar to those within other core complexes of the Shuswap domain (e.g., Johnson and Brown 1996; Simony and Carr 1997; Kruckenberg et al. 2008; Gervais and Brown 2011). The orientation of undeformed, elongate minerals (Sil, Hbl) within this gneissosity suggests D2 developed at about the same time as peak metamorphism. The presence of abundant microfractures normal to the foliation and lineation suggest that extension approximately parallel to this D2 foliation postdated metamorphic recrystallization (Preto 1970b).

A number of folding episodes (F3a, F3b, F4, and F4d) postdate the development of the S2 gneissosity, L2 mineral lineations, and local shear zones. These folds are visible on thin-section, outcrop, and map scales. The major folding episode in the GFC is F3a, represented in map scale and in north–south cross-sections as large-scale features such as the Morrisey Creek Antiform (Figs. 2, 3C). This corresponds to a Phase 1 fold of (Preto 1970b). F3a folds are typically upright to weakly north- or south-vergent, display open to isoclinal fold morphologies, and are associated with variably developed crenulations. Shallowly plunging fold axes trend ~280°–100°.

A small number of folds characterized as F3a (Fig. 7A) have shallowly plunging fold axes trending 340°–160° and are visible in the map and cross-sections of the GFC (Figs. 3A, 3B). In the field they occur as metre-scale folds in amphibolites and orthogneisses and are predominantly east-vergent, with upright to recumbent fold morphologies that range from tight to open. The relative tightness of the F3a folds distinguishes them from broad to open F3 folds (discussed later in the text). Rare axial planar cleavage fabrics associated with F3a folds (S3a) are near vertical and may have helped localize the emplacement of late, post-kinematic granitoid and monzonite dikes that have a similar 340°–160° trend (unit Eg in Fig. 2). In thin section, L1c crenulations and S1c crenulation...
cleavages mechanically deform the peak metamorphic assemblage in pelites, and L₂ mineral lineations wrap around F₃b folds. The relative timing of F₃a and F₃b folds is obscure, as both fold types were not observed in a single outcrop. However, Preto (1970b) lumped a number of F₃b folds in the southwest part of the field area into an “Isolated Structures” category (e.g., CL905, Fig. 6B) and established that these postdated F₃a folds. Both F₃a and F₃b folds predate emplacement of post-kinematic Eocene granitoids (Figs. 3, 7B). Because these folds (most notably the east–west-trending F₃a folds) show no evidence of warping the trace of the Kettle River fault, they are interpreted to have formed prior to movement on this fault — an interpretation corroborated by the fact that the post-kinematic granitoids that truncate F₃a and F₃b folds are brecciated in the fault zone.

The final folding episode in the GFC, F₄, is characterized by upright, gentle folds that locally fold L₃a crenulation lineations in the southeastern part of the study area. These folds are correlated with the Phase 2 folds of Preto (1970b).

**KRF hanging wall**

A detailed structural characterization of the hanging wall to the KRF was presented by Acton (1998) and Acton et al. (2002). Additional structural mapping was conducted in the southeast corner of the study area by Höy and Jackaman (2005). The structures in the hanging-wall rocks bear no relationship to those in the footwall gneisses of the GFC. Compositional banding in metasiltstones is attributed to transposition of original bedding (S₀), and this transposed fabric has subsequently undergone at
Fig. 4. Field photographs of characteristic lithologies in the Grand Forks complex (A–D) and in the KRF hanging wall (E–H). (A) CL945: Garnet-rich Sil + Grt + Kfs (Sil, sillimanite; Grt, garnet; Kfs, potassium feldspar) paragneiss with stromatic leucosome. (B) CL455: Sil + Grt + Kfs paragneiss with top-to-the-east shear and boudinaged leucosome. (C) CL388: Gneissic fabrics in Cpx + Opx + Hbl (Cpx, clinopyroxene; Opx, orthopyroxene; Hbl, hornblende) Type B metabasite, with layer-parallel injections of leucogranitic melt. (D) CL345: Di + Pl + Qtz-bearing (Di, diopside; Pl, plagioclase; Qtz, quartz) calc-silicate gneiss at the south end of Christina Lake. (E) CL1013: Ovoid cordierite porphyroblasts in unfoliated argillaceous pelite. (F) CL294: Net-textured granitic leucosome in migmatitic Crd + Kfs (Crd, cordierite) metapelite in the contact aureole of the Nelson intrusive suite. (G) CL478: Rossland Group volcaniclastic with coarse andesitic clasts surrounded by epidote-rich matrix. (H) TGI20: Ductilely deformed, Di-bearing marble in the Burnt Basin area northeast of Christina Lake.
least two phases of deformation (Acton et al. 2002). The timing of
dominant D2 deformation is bracketed to the Late Triassic –
Early Jurassic, between U–Pb zircon ages for the deformed Josh
Creek diorite (216 Ma) and the undeformed Fife diorite (197–
181 Ma) (Acton et al. 2002).

Kettle River fault
Along the length of the GFC and Kettle dome, the eastern
margin of the core complex is defined by the east-dipping, brittle–
ductile Kettle River fault (KRF). It extends north and south of the
current field area for a total strike length of 95 km (Figs. 1, 2).
The KRF is poorly exposed in British Columbia, where it is
largely obscured under Christina Lake and Quaternary over-
burden (Preto 1970b; Rhodes and Cheney 1981; Acton 1998;
Höy and Jackaman 2005). Previous estimates for the vertical
separation across the KRF range from 5 km (Bowman 1950)t o
15 km (Parrish et al. 1988), with decreasing displacement to
the north (Orr and Cheney 1987; Parrish et al. 1988). A central
goal of the present study is to more accurately constrain the
vertical separation across the KRF.

Deformation in the immediate hanging wall of the KRF in
British Columbia comprises predominantly steeply dipping, brit-
tle fabrics that deform all observed units, including leucogranites
 correlated with the Paleocene to Eocene Ladybird suite (Fig. 7C)
and 51.1 Ma Coryell intrusives (Parrish et al. 1988; Carr
and Parkinson 1989; Acton 1998; this study). Brittle deforma-
tion fabrics occur up to 300–500 m from the fault trace and
include brecciation and local cataclastic textures. Chlorite,
epidote, and calcite alteration is common, and minor late
pyrite mineralization is associated with this alteration. Mineral
fibre slickenlines are common on steep Chl/Ep-coated (Chl,
chlorite; Ep, epidote) fracture surfaces, with near-vertical
plunges suggesting predominantly dip-slip movement.

Two previously unrecognized ductile shear zones have been
identified in the hanging wall of the KRF east of Christina
Lake (Fig. 2). These shear zones deform granitoids of the
Middle Jurassic Nelson suite and the Paleocene Ladybird
leucogranite suite, as well as the older Quesnel stratigraphy
(Fig. 2). The westernmost of these two shear zones, hereby

Fig. 5. Map of sample localities referred to in this paper. The prefix “CL” has been removed from all sample names, the one exception
being sample TGI20 in the KRF hanging wall. The location of the study of Laberge and Pattison (2007) (L&P 2007) is indicated by the
black rectangle. Refer to Fig. 2 for legend and abbreviations.
referred to as the Texas Point shear zone (TPSZ), is in structural contact with GFC footwall gneisses northeast of Christina Lake (Figs. 2, 3). It is characterized by slaty gneissic granitoids observed in discontinuous bands along the north-eastern shore of Christina Lake and along the eastern edge of the Sandner Creek drainage (Fig. 7D). The degree of strain within the TPSZ is variable, and undeformed leucogranite bodies within the shear zone suggest the emplacement of leucogranites may have outlasted TPSZ deformation. Preliminary kinematic analysis suggests that the TPSZ has a normal, top-to-the-east sense of motion. Acton (1998) reported high-strain, gneissic fabrics in hornblende granitoids at Texas Point (CL309, Fig. 5), and Parrish et al. (1988) grouped these gneissic rocks within the Grand Forks complex, implying that the Kettle River fault turns sharply away from Christina Lake at Texas Point. However, the occurrence of TPSZ fabrics northward along the lake and up the Sandner Creek drainage supports the mapped fault trace of Preto (1970b), Höy and Jackaman (2005), and others. Ductile deformation fabrics associated with the TPSZ are truncated by the Coryell Batholith along the east shore of Christina Lake, with subsequent KRF brittle deformation affecting all units (Acton 1998; this study).

The second hanging-wall ductile shear zone, hereby referred to as the Gladstone shear zone (GSZ) (Fig. 2), lies to the east and structurally above the TPSZ. Preliminary kinematic analysis in the field suggests a top-to-the-east, normal sense of movement for the GSZ, but owing to limited exposure and a lack of offset of marker units, the sense of movement is not well constrained. The TPSZ and GSZ strike parallel to each other, with shallow to moderate dips to the east. Mineral lineations (Cubley 2012) in both shear zones trend east-southeast, and planar fabrics are folded in upright, broad folds that have east-southeast-trending fold axes parallel to the mineral lineations (Cubley 2012).

In contrast to Washington State (Rhodes and Cheney 1981; Hurich et al. 1985; Mulch et al. 2007), footwall mylonitic fabrics clearly attributable to the KRF are not observed in British Columbia (Preto, 1970b; this study). Thus the transition from a footwall mylonite zone to a relatively undeformed gneissic core, as observed in Washington state (e.g., Hurich et al. 1985; Mulch et al. 2007), is not documented. Where footwall rocks close to the KRF trace are observed northeast and west of Christina Lake, fabrics are typically not mylonitic. In the Cascade area south of Christina Lake, however, local high-strain layers dip shallowly to the east with strong east–west lineations, distinct Qtz-ribboning, boudinage of calc-silicate layers, and feldspar porphyroclast rotation. The east–west mineral lineations within this zone have orientations parallel to those in other parts of the complex. Based on the absence of muscovite and (or) chlorite, these fabrics appear to be high-temperature features. These high-strain layers are folded by F3b folds and are truncated by massive leucogranites and pegmatites (Fig. 7B). All footwall units are subsequently cut by brittle deformation features characterized by minor brecciation and steeply dipping Chl ± Ep veining.

The lack of a well-developed KRF footwall mylonite zone in British Columbia has been noted by other authors (e.g., Rhodes and Cheney 1981). Explanations include a decreasing degree of mylonitization with downward structural level (Rhodes and Cheney 1981) and (or) decreasing northward displacement along the KRF (Parrish et al. 1988). In north-eastern Washington, ductile movement in footwall mylonitic quartzites of the KRF has been dated at ~49 Ma (Berger and Snee 1992; Mulch et al. 2007) under greenschist-facies conditions (~415 °C) (Mulch et al. 2007). Brittle deformation fabrics are typically subparallel to shallowly dipping footwall mylonites (Rhodes and Cheney 1981; Hurich et al. 1985), in contrast to the high-angle brittle slickensides and brecciation zones observed in British Columbia. This low-temperature extension at high structural levels is proposed to postdate the timing of regionally extensive, high-temperature extension at

---

### Structural Elements within the Grand Forks Complex

<table>
<thead>
<tr>
<th>Deformation (D)</th>
<th>Folding(F)</th>
<th>Schistosity (S)</th>
<th>Lineation (L)</th>
<th>Timing Relative to Metamorphism (M)</th>
<th>Timing Relative to Exhumation Stages</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>D1 - cryptic event(s)</strong></td>
<td>F1 - folded inclusion trails</td>
<td>S1 - Sil or Qtz inclusion trails in Grt or St</td>
<td>?</td>
<td>Pre-peak metamorphism</td>
<td>Pre-exhumation</td>
</tr>
<tr>
<td><strong>D2 - main gneissosity forming event</strong></td>
<td>F2 - intrafolial folds within gneissic layers</td>
<td>S2 - main gneissosity in Grand Forks Group</td>
<td>L2 - E-W-trending Hbl, Sil, Qtz lineations</td>
<td>Deformation pre-to-syn-peak metamorphism</td>
<td>Predates high-T horizontal movement of rock units (Preto, 1970; this study)</td>
</tr>
<tr>
<td><strong>D3a - E-W folding event</strong></td>
<td>F3a - Open to isoclinal upright folds with E-W fold axes (outcrop and map scales) Phase 1 of Preto (1970)</td>
<td>None observed</td>
<td>L3 - local crenulation lineations</td>
<td>Post-peak metamorphism</td>
<td>Postdates high-T horizontal movement of rock units, predates brittle KRF (Preto, 1970; this study)</td>
</tr>
<tr>
<td><strong>D3b - NW-SE folding event</strong></td>
<td>F3b - isolated outcrop scale folds, local crenulations of S2 “isolated structures” Preto (1970)</td>
<td>S3b - locally developed penetrative crenulation cleavage</td>
<td>L3 - rare crenulation lineations</td>
<td>Post-peak metamorphism (L3s lineations wrapped around F3b folds)</td>
<td>Postdates high-T horizontal movement of rock units, predates brittle KRF (Preto, 1970; this study)</td>
</tr>
<tr>
<td><strong>D4 - Broad doming of GFC</strong></td>
<td>F4 - Open to broad folds on outcrop and map scales (folds L3a lineations) Phase 2 of Preto (1970)</td>
<td>None observed</td>
<td>None observed</td>
<td>Post-peak metamorphism</td>
<td>Syn-to-post KRF? (Cheney, 1980; Rhodes and Cheney, 1981)</td>
</tr>
</tbody>
</table>

---

Fig. 6. Summary diagram outlining the defining characteristics of the five main “internal” deformation elements within the Grand Forks complex (GFC). Evidence from other studies is cited where appropriate. KRF, Kettle River fault.
mid-crustal levels by a few million years (Mulch et al. 2007).

Analytical methods

Electron microprobe analyses

Chemical analyses of major mineral phases were conducted on the JEOL JXA-8200 electron microprobe at the University of Calgary using wavelength-dispersive spectrometry (WDS). Operating conditions were held constant for all mineral phases, with an accelerating voltage of 15 kV, a beam current of 20 nA, and a beam diameter of 5 μm. All results were subjected to matrix corrections based on the ZAF (atomic number – absorption – fluorescence) method (Reed 1996). Compositional zoning of garnet was determined by X-ray element mapping (Mn, Ca, Fe, Mg), using beam currents between 50 and 100 nA and typical dwell times of 20–30 ms. Representative chemical analyses are presented in Tables 1–3 for GFC pelites, GFC amphibolites, and KRF hanging-wall lithologies, respectively.

Thermodynamic modelling

Thermodynamic modelling was performed on pelitic and basic samples from the GFC and KRF hanging wall to determine metamorphic $P$–$T$ conditions. Bulk chemical XRF analyses were obtained using a Philips PW2440 4 kW automated spectrometer system at McGill University, with carbon and sulfur measured on an Eltra CS-800 infrared analyzer. Complete XRF analysis results for modelled lithologies are given in Appendix A. From the analyzed XRF bulk compositions, a projection was made from pyrrhotite (footwall) or pyrite (hanging wall) to remove sulfur from the chemical system. The reduction in iron results in a negligible 1–2 °C shift in suprasolidus phase boundaries.

Ten-component MnNCKFMASHT ($\text{MnO–Na}_2\text{O–CaO–K}_2\text{O–FeO–MgO–Al}_2\text{O}_3–\text{SiO}_2–\text{H}_2\text{O–TiO}_2$) isochemical phase diagrams were constructed using the Gibbs free-energy minimization software Theria-Domino (de Capitani and Brown 1987; de Capitani and Petrakakis 2010). This program utilizes...
the Holland and Powell (1998) thermodynamic database (updated to database 5.5 in 2003) with the following solution model additions: amphibole per Diener et al. (2007); clinopyroxene per Green et al. (2007); garnet, biotite, ilmenite, and melt per White et al. (2007); and plagioclase per D.L. Tinkham (personal communication, 2010). The plagioclase model is an adaptation of the Holland and Powell (2003) plagioclase model but is formulated to automatically handle the C1–I1 transition.

For migmatitic GFC samples, modelling was conducted only for suprasolidus conditions. A number of assumptions were made: (1) all water available for melting is contained within hydrous phases, owing to negligible porosity in high-grade metapelites, and (2) above the solidus, the rock is closed to the loss of melt and therefore to the loss of H2O. Any water generated by suprasolidus dehydration reactions is assumed to be dissolved in the melt phase (dehydration melting). Fixed whole-rock water contents were calculated based on the total amount of H2O contained in hydrous phases immediately down-temperature of the wet solidus. This calculation was made at 5.8 kbar (1 kbar = 100 MPa), the pressure of peak metamorphism in the GFC calculated by Laberge and Pattison (2007) and Cubley and Pattison (2009). Fractional effects were ignored, consistent with relatively rapid elemental diffusion during high-grade metamorphism. Stable assemblage fields are shaded according to variance, with darker colours representing higher variance. Mineral composition isopleths were constructed using Theriak-Dominio, with the shaded areas representing the range of isopleths corresponding to the measured compositions of the natural minerals and the heavy dashed line representing the average composition.

Modelling of KRF hanging-wall metabasites was conducted in a smaller NCFMASH (Na₂O–CaO–FeO–MgO–Al₂O₃–SiO₂–H₂O) chemical system, following the precedent of recent studies by Dale et al. (2005) and Elmer et al. (2006). Projections were made from ilmenite and K-feldspar, the latter identified during microprobe analysis of each modelled sample. Modelling was conducted with clinopyroxene excluded from the thermodynamic database, as clinopyroxene was otherwise predicted to be stable at all P–T conditions of interest, at odds with the observed mineral assemblages.

Mineral equilibria thermobarometry

Mineral equilibria thermobarometry was used for additional P–T estimation as a complement to constraints from thermodynamic modelling. Unless otherwise indicated, all thermobarometry was conducted using THERMOCALC version 3.33 (Powell and Holland 1994 and more recent updates), with end-member mineral activities calculated using the AX program (Holland and Powell 2000; updated in 2008 (v2)). THERMOCALC 3.33 utilizes the tc-ds55s.txt data set of Holland and Powell (2003). To establish the position of individual mineral equilibria, pressures and (or) temperatures were specified by the user and the other variable calculated by THERMOCALC (avP or avT).

For amphibole–albite–quartz thermobarometry in KRF hanging-wall metabasites, the program TWQ (v. 1.02) of Berman (1991) was used in place of THERMOCALC because the Holland and Powell data set (tc-ds55s.txt) and newer version of TWQ (v. 2.02; Berman and Aranovich 1996) do not include thermodynamic data for ferro-tschermakite and ferro-pargasite. TWQ 1.02 utilizes the thermodynamic database of Berman (1988) with the amphibole solution model of Mäder et al. (1994) and the plagioclase solution model of Furman and Lindsley (1988). Paragonite spinel-cordierite thermometry in the FMASZn (FeO–MgO–Al₂O₃–SiO₂–ZnO) system required independent equilibria calculations based on the calibration of Nichols et al. (1992), as neither THERMOCALC nor TWQ accounts for the garnite end-member of spinel.

Petrography, mineral chemistry, and thermodynamic modelling of GFC rocks in the KRF footwall

Metapelites

Migmatitic K-feldspar + biotite ± garnet ± sillimanite ± cordierite paragenisses

The distribution of metamorphic mineral assemblages in pelites, amphibolites, and carbonates from the Grand Forks complex is shown in Fig. 8. In paragneisses within units Pr1 and Pr3, the predominant peak metamorphic assemblage in metapelites is Sill + Kfs + Bt ± Grt ± Crd (Bt, biotite). Kfs + Grt + Bt and Kfs + Bt assemblages occur in more psammitic horizons. These assemblages are indicative of upper-amphibolite- to granulate-facies conditions (Yardley 1989). Most paragneisses are migmatitic and contain Kfs + Qtz-rich stromatic leucosome layers parallel to the S₂ gneissosity, but thicker leucosome layers and coarse pegmatite injections locally crosscut the planar fabric (Fig. 4B). Leucosome and pegmatite in these metatexites compose up to 30% of the rock volume. Sillimanite-rich Sill + Bt + Pl ± Grt ± Crd mesoscale layers in aluminous paragneisses (terminology after Kriegsman 2001) and quartz-rich psammitic interlayers comprise the rest of the rock.

In the petrography sections that follow, mineral assemblages, textures, and chemistry will be discussed as follows: in lithologies that contain coronas or symplectites, the minerals incorporated within those textures constitute the “coronal assemblage” or “symplectite assemblage”. The “matrix” is the remainder of the rock not including the reaction textures.
Granby fault  
Kettle River fault  
Christina Lake 
Pr3  
Pr1  
Pr1m  
Eig  
mPg  
Progn3  
Eig  
Pr2  
Pr2  
Progn1  
Pr4  
mJg  
mEc  
mEc  
mEc  
mJg  
mJg  
mJg  
Trb  
CPk  
CPm  
CPm  
CPm  
eJv  
um  
mEc  
mJg  
mJg  
mEc  
mEc  
Trd  
CPm  
CPm  
CPm  
CPm  
eJv  
um  
mEc  
mJg  
mJg  
mEc  
mEc  
Kilometres  
2  
5  
10  
Grand Forks  
Basite - Calcsilicate Mineral Assemblages  
Metapelite Mineral Assemblages  
Metapelite (Hanging Wall)  
Metapelite (Footwall)  
Basite (Hanging Wall)  
Basite (Footwall)  
Calcsilicate (Hanging Wall)  
Calcsilicate (Footwall)
Where no reaction textures are observed, the matrix represents the entire rock.

**Paragneiss matrix assemblage**

Sillimanite + garnet-bearing aluminous paragneisses in units Pr1 and Pr3 yield the most detailed record of the metamorphic history of GFC. Along with sillimanite and garnet, these gneisses contain Kfs + Bt + Pl + Qtz + Ilm (Bt, biotite; Ilm, ilmenite), with varying abundances of matrix cordierite. Apatite, tourmaline, rutile, magnetite, zircon, and monazite are common accessory phases. Quartz, plagioclase, and K-feldspar make up approximately 65% of the modal mineralogy of the average paragneiss. Biotite (~20%) is concentrated in the mesosome but is also found in leucosomes and monomineralic selvages along the mesosome–leucosome boundary. Anhedral to subhedral garnet (5%–7%) is commonly found on the boundary between Pl + Qtz-rich psammitic layers and sillimanite + biotite-rich mesosomes. Garnet is wrapped by matrix sillimanite, but locally fibrolite occurs as inclusions within garnet cores, suggesting that garnet and sillimanite growth overlapped. Matrix sillimanite (~5%) comprises elongate prismatic grains typically rimmed by pinitized cordierite, ilmenite, and spinel (more discussion to follow). Cordierite is found in the leucosomes and psammitic interlayers of some samples as partially pinitized grains. The presence of matrix cordierite in a Sil + Grt + Kfs + Crd-bearing peak assemblage suggests granulite facies in metapelites (e.g., Yardley 1989; Pattison et al. 2003).

**Paragneiss coronal assemblage**

In nearly all paragneiss samples, Crd-bearing coronal structures are developed around garnet and sillimanite (Fig. 9). The Crd-bearing coronal texture around sillimanite is present regardless of whether or not garnet is present in the matrix assemblage (Figs. 9A, 9B). Where fully developed, this corona has the following structure, going from sillimanite to the innermost corona layer, commonly in a symplectite with quartz, surrounded by a quartz-rich outer layer (Figs. 9C, 9D). The spinel reported by Laberge and Pattison (2007) in cordierite-rich coronas enveloping garnet most likely formed from reaction of sillimanite that wrapped the garnet prior to corona formation (Fig. 9C).

**Mineral chemistry**

Garnet shows considerable compositional variation both among samples and among individual porphyroblasts in a single sample. Zoning patterns in garnet reflect retrograde processes and typically display either (1) a complete lack of zoning, apart from thin Mn + Fe-rich rims (see CL455 garnet in Cubley and Pattison 2009) or (2) a weak zoning pattern suggesting homogenization at peak temperature followed by re-equilibration at the rims (e.g., CL592 in Fig. 10A).

Biotite compositions in paragneisses vary systematically with textural location. The Mg# (Mg# = Mg/(Fe + Mg)) of biotite in mesosomes (range of 0.34–0.46, mean of ~0.41) is on average higher than in leucosomes (range of 0.33–0.44, mean of ~0.39), and titanium concentrations of biotite in leucosomes are typically higher than in mesosomes (Table 1). Cordierite in the matrix and in coronas around garnet has an average Mg# of ~0.46 (0.45–0.48), whereas cordierite in coronas around sillimanite is Mg-rich (~0.54–0.60). Spinel in cordierite coronas around sillimanite has an average Mg# of ~0.08 (0.07–0.11). Matrix plagioclase grains average ~An23 (An23–An25) (An, anorthite), whereas plagioclase grains in coronas around sillimanite show a progressive decrease in XAn (XAn = Ca/(Na + K + Ca)) with increasing distance from sillimanite. For example, in sample CL831, plagioclase nearest sillimanite is ~An13, whereas plagioclase in contact with K-feldspar is ~An27. Corona plagioclase consistently has higher XAn than leucosome plagioclase.

**Thermodynamic modelling**

A MnNCKFMASHT isochemical phase diagram for representative paragneiss sample CL831 is shown in Fig. 11A. The matrix mineral assemblage in the sample is Si1 + Grt + Kfs + Bt + Pl + Qtz + Ilm. The broad stability field for this assemblage is outlined in heavy black. Mineral composition isopleths were calculated for XAn in garnet, Mg# in garnet (Fig. 11B), Mg# in biotite (Fig. 11C), Mg# in cordierite (Fig. 11D), and anorthite content in plagioclase (Fig. 11E). The XAn (XAn = Mn/(Mn + Mg + Ca + Fe2+)) and Mg# values from homogenized garnet cores, XAn in plagioclase, and Mg# in biotite yield temperature estimates in a narrow zone between 730–765 °C but provide little constraint on pressure (Fig. 11F). Two individual mineral equilibria were employed to help constrain the pressure. A barometer based on the equilibrium 3east + py + gr + qtz = 3phl + 6an (east,
eastonite; py, pyrope; gr, grossular; qtz, quartz; phl, phlogopite; an, anorthite) provided an estimate of 5.4 kbar at 750 °C, whereas the equilibrium 3east/H₂O + 2gr/H₂O + 6qtz/H₂O + 2phl/H₂O + ann (alm, almandine; ann, annite) provided an estimate of 5.8 kbar. A peak P–T estimate of 750 °C and 5.6 kbar is thus proposed for CL831. This estimate is within error of an earlier P–T estimate from an additional GFC paragneiss sample, CL455 (Fig. 5), of 750 °C, 5.8 kbar (Cubley and Pattison 2009).

Temperature estimates for five other paragneiss samples using mineral composition isopleths fall in the same range as CL831 and CL455, with temperatures averaging 760 °C (750–775 °C) (Cubley 2012). Pressure estimation for these samples using either grossular isopleths in phase diagrams, or calcium-based thermobarometry (e.g., GASP) in THERMOCALC, yields results that are implausibly high (7–10 kbar). The stability field for the widespread Sil + Grt + Kfs + Bt + Crd mineral assemblage is relatively insensitive to bulk composition, ranging from 4.5 to 6.5 kbar at the temperatures of interest (750–800 °C) (e.g., CL455, Cubley and Pattison 2009; CL528, Cubley 2012). It is possible, however, that the Ca-based pressures represent a cryptic record of garnet growth at an earlier, higher pressure stage in the rock’s P–T history, the preservation of which may be due to the relatively slow diffusion of Ca in garnet compared with other elements like Fe and Mg that equilibrated at the lower pressures of the thermal peak.

Pressure–temperature conditions of the coronal assemblages in migmatitic metapelites are less well constrained. The main coronal features observed are (1) reaction of garnet to cordierite + quartz, whether or not sillimanite is present in the rock, and (2) formation of Crd + Ilm + Spl coronas around sillimanite, whether or not garnet is present in the rock. The first may be the result of continuous bulk-rock reaction that consumes garnet, such as the MnNCKFMASHT Reaction 1:

\[ \text{Grt + Kfs + Ilm + melt = Crd + Bt + Pl + Qtz} \]

Reaction 1 is predicted in response to decompression through the Grt + Kfs + Bt + Crd + Pl + Qtz + Ilm + Liq (Liq = melt) stability field (dashed box in Fig. 11A). Mineral isopleths of coexisting garnet and cordierite rims intersect at about 735 °C and 3.6 kbar in this field (Fig. 11F).
Table 1. Representative mineral compositions in paragneiss and pelitic schist from the Grand Forks complex (GFC).

<table>
<thead>
<tr>
<th>Sample</th>
<th>CL592</th>
<th>CL592</th>
<th>CL831</th>
<th>CL831</th>
<th>CL931</th>
<th>CL931</th>
<th>CL1003</th>
<th>CL1003A</th>
<th>CL931</th>
<th>CL1003A</th>
<th>CL1003A</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oxide</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SiO₂</td>
<td>37.62</td>
<td>37.70</td>
<td>37.39</td>
<td>37.16</td>
<td>36.88</td>
<td>37.13</td>
<td>37.47</td>
<td>35.56</td>
<td>34.33</td>
<td>34.85</td>
<td>34.10</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.00</td>
<td>0.00</td>
<td>0.02</td>
<td>0.01</td>
<td>0.00</td>
<td>0.01</td>
<td>0.00</td>
<td>2.92</td>
<td>4.47</td>
<td>2.73</td>
<td>4.53</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>FeO</td>
<td>34.98</td>
<td>34.97</td>
<td>35.05</td>
<td>35.36</td>
<td>35.31</td>
<td>35.26</td>
<td>35.26</td>
<td>35.94</td>
<td>36.87</td>
<td>36.87</td>
<td>37.16</td>
</tr>
<tr>
<td>MnO</td>
<td>0.39</td>
<td>0.46</td>
<td>1.33</td>
<td>1.74</td>
<td>2.14</td>
<td>9.58</td>
<td>5.84</td>
<td>0.10</td>
<td>0.08</td>
<td>0.05</td>
<td>0.15</td>
</tr>
<tr>
<td>MgO</td>
<td>4.34</td>
<td>4.21</td>
<td>3.83</td>
<td>2.60</td>
<td>4.21</td>
<td>2.54</td>
<td>3.35</td>
<td>7.56</td>
<td>5.91</td>
<td>9.83</td>
<td>8.18</td>
</tr>
<tr>
<td>CaO</td>
<td>2.00</td>
<td>2.08</td>
<td>0.74</td>
<td>0.93</td>
<td>0.78</td>
<td>1.60</td>
<td>0.65</td>
<td>2.62</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>Na₂O</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>K₂O</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>BaO</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>0.12</td>
<td>0.11</td>
</tr>
<tr>
<td>SrO</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>0.14</td>
<td>0.09</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>0.03</td>
<td>0.03</td>
<td>0.01</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.13</td>
<td>0.15</td>
<td>0.18</td>
<td>0.19</td>
</tr>
<tr>
<td>Y₂O₃</td>
<td>0.04</td>
<td>0.05</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.08</td>
<td>0.09</td>
<td>8.90</td>
<td>9.11</td>
<td>9.03</td>
<td>8.30</td>
</tr>
<tr>
<td>F</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>0.20</td>
<td>0.14</td>
</tr>
<tr>
<td>Cl</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>0.24</td>
<td>0.14</td>
</tr>
<tr>
<td>Total</td>
<td>100.51</td>
<td>100.76</td>
<td>99.86</td>
<td>99.74</td>
<td>99.45</td>
<td>99.89</td>
<td>95.53</td>
<td>95.61</td>
<td>95.52</td>
<td>96.53</td>
<td>96.65</td>
</tr>
</tbody>
</table>

Cations

| Si⁴⁺    | 2.99  | 2.99  | 3.02  | 3.01  | 2.97  | 3.00  | 3.00   | 2.71    | 2.65  | 2.67    | 2.63    |
| Ti⁴⁺    | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00   | 0.00    | 0.00  | 0.00    | 0.00    |
| Al³⁺    | 1.98  | 2.01  | 2.01  | 2.01  | 2.00  | 2.00  | 2.00   | 1.73    | 1.69  | 1.65    | 1.56    |
| Fe³⁺    | 2.33  | 2.32  | 2.37  | 2.50  | 2.26  | 1.72  | 1.97   | 1.34    | 1.48  | 1.29    | 1.47    |
| Mn²⁺    | 0.03  | 0.03  | 0.09  | 0.12  | 0.15  | 0.66  | 0.40   | 0.01    | 0.01  | 0.00    | 0.01    |
| Mg²⁺    | 0.51  | 0.50  | 0.41  | 0.31  | 0.50  | 0.31  | 0.40   | 0.86    | 0.68  | 1.12    | 0.94    |
| Ca²⁺    | 0.17  | 0.18  | 0.06  | 0.06  | 0.14  | 0.32  | 0.23   | 0.00    | 0.00  | 0.00    | 0.00    |
| Na⁺     | —     | —     | —     | —     | —     | —     | —      | —       | —     | 0.02    | 0.03    |
| K⁺      | —     | —     | —     | —     | —     | —     | —      | —       | —     | 0.87    | 0.90    |
| Ba⁺     | —     | —     | —     | —     | —     | —     | —      | —       | —     | 0.00    | 0.00    |
| Sr⁺     | —     | —     | —     | —     | —     | —     | —      | —       | —     | 0.00    | 0.00    |
| Cr³⁺    | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00   | 0.00    | 0.00  | 0.00    | 0.00    |
| Y³⁺     | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00   | 0.00    | 0.00  | 0.00    | 0.00    |
| F⁻      | —     | —     | —     | —     | —     | —     | —      | —       | —     | 0.01    | 0.01    |
| Cl⁻     | —     | —     | —     | —     | —     | —     | —      | —       | —     | 0.01    | 0.01    |
| Total   | 100.50| 100.53| 99.89 | 99.74 | 99.45 | 99.89 | 95.53  | 95.61   | 95.52 | 96.53   | 96.65   |

Note: rims in contact with Kfs.
For personal use only.
For the spinel-bearing coronal structure, there is no predicted stability field for spinel in any of the phase diagrams calculated for the whole rock composition. The simplified KFMASH bulk-rock reaction proposed for the decompression assemblage by Laberge and Pattison (2007):

\[ \text{Grt} + \text{Sil} + \text{melt} = \text{Crd} + \text{Spl} + \text{Qtz} + \text{Kfs} \]

does not satisfy the observation that spinel forms exclusively in coronas around sillimanite, whether garnet is present in the rock or not. The formation of the Crd + Ilm + Spl + Rt assemblage around sillimanite may instead be the result of a two-stage process. Garnet-absent KFMASHT Reaction 5:

\[ \text{Sil} + \text{Bt} + \text{Pl} + \text{Qtz} = \text{Crd} + \text{Kfs} + \text{Ilm} + \text{melt} \]

could account for the growth of cordierite and the second generation of ilmenite observed only in coronal structures. The source of the titanium for the ilmenite is most likely biotite. The growth of spinel and rutile both as rims around coronal ilmenite and as isolated grains within cordierite may then be the result of a second reaction, restricted to the corona, such as Reaction 4:

\[ \text{Crd} + \text{Sil} + \text{Ilm} = \text{Spl} + \text{Rt} \]

Alternatively, the growth of the Crd + Ilm + Spl coronas could be the result of a single KFMASHT reaction in a local Al + Fe-rich, Si-poor effective bulk composition (EBC) (Greenfield et al. 1998; Johnson et al. 2004), such as Reaction 5:

\[ \text{Sil} + \text{Bt} + \text{Qtz} = \text{Spl} + \text{Crd} + \text{Ilm} + \text{melt} + \text{Kfs} \]

To test this model, thermodynamic modelling of different EBCs was conducted incorporating three components: sillimanite, biotite, and a Sil + Bt-free matrix. The matrix composition was estimated by subtracting 5 vol\% sillimanite and 15 vol\% biotite from the bulk composition (the typical modal amounts of these minerals in the matrix of CL831), using the measured biotite composition (Table 1). Sillimanite was fixed as one end-member of the binary P–X system, whereas the second end-member was fixed as a mixture of biotite and Sil + Bt-free matrix in different ratios. This ratio was varied iteratively to produce the largest Crd + Ilm + Spl field. Figure 12A shows a P–X diagram constructed at 750 °C for a Bt:matrix ratio of 1:1.75. A stability field for Crd + Ilm + Spl occurs at a Sil:Bt + matrix of ~0.22:0.78. A P–T diagram using this bulk composition yields a stability field for the Crd + Spl assemblage between 4.25 and 2.5 kbar (Fig. 12B).

Whereas the EBC just mentioned predicts a stability field for the coronal assemblage, it fails to address the discrete coronal structure (layering) observed. The coronal texture is likely the result of chemical potential gradients that developed between sillimanite and the surrounding minerals during the high-temperature decompression episode (White and Powell 2011).

Mineral equilibria thermobarometry can be used to further constrain pressure conditions of the Crd + Ilm + Spl coronas. The equilibria Mg–Crd = 2Spl + Qtz in the FMASZn (FeO–MgO–Al2O3–SiO2–ZnO) system yields a pressure estimate of ~3.3 kbar (Fig. 11F), similar to the ~3.6 kbar estimate from coronal garnet and cordierite isopleths discussed earlier in the text. The combination of all constraints yields an estimate for the formation of the coronal assemblages of 735 ± 20 °C, 3.3 ± 0.5 kbar, similar to the 750 ± 30 °C, 3–4 kbar estimate of Laberge and Pattison (2007) on the west flank of the complex. When compared with the ~750 °C, ~5.6 kbar estimate of peak metamorphism, these results imply an episode of high-temperature decompression.

**Biotite ± staurolite ± garnet schists (unit Pr5)**

Immediately west of the village of Grand Forks (Fig. 2), unit Pr5 is composed of fine-grained, rusty amphibolites and pelitic schists that are non-migmatitic and preserve a schistose fabric. Two representative pelite samples were analyzed from unit Pr5, one of which is a Sil + St + Bt schist (sample CL1003) and the other a Grt + Bt schist (CL1001A).

**Petrography of staurolite ± sillimanite schists**

Staurolite-bearing schists in unit Pr5 have St + Bt + Pl + Qtz + Ilm ± Kfs peak metamorphic assemblages, with highly resorbed subhedral to anhedral staurolite porphyroblasts. Graphite and quartz inclusions form an internal planar fabric within the cores of staurolites (S1), at a high angle to the dominant foliation. Biotite grains define a foliation that wraps, and thus postdates, latest growth of cordierite, biotite, and a Sil + Bt matrix. The combination of all constraints yields an estimate for the formation of the coronal assemblages of 735 ± 20 °C, 3.3 ± 0.5 kbar, similar to the 750 ± 30 °C, 3–4 kbar estimate of Laberge and Pattison (2007) on the west flank of the complex. When compared with the ~750 °C, ~5.6 kbar estimate of peak metamorphism, these results imply an episode of high-temperature decompression.

**Mineral chemistry**

Staurolite porphyroblasts in sample CL1003 show chemically homogenous cores and thin, chemically distinct rims (Table 1). Core compositions have Mg# numbers averaging ~0.19 (0.17–0.21), whereas rim Mg# values average ~0.21.
Effective bulk composition (EBC) modelling of decompression reaction texture surrounding sillimanite. (A) P–X diagram illustrating relative contributions to an EBC of sillimanite and biotite + Bt-free matrix, the latter in a 1:1.75 ratio. The observed corona assemblage, Crd + Ilm + Spl + Pl + Kfs + Bt, is indicated by Field 8, at an overall sillimanite : Bt + matrix ratio of 0.22:0.78. (B) P–T diagram for the EBC determined in (A). Variation in Bt:Bt-free matrix composition and P–X fixed temperature affect EBC composition and thus P–T stability field, but resulting Spl-bearing P–T field (Field 7) generally lies between 4.5 and 2.5 kbar. This implies near-isothermal decompression from peak P–T conditions (Field 11). Abbreviations defined in text. 1 bar = 100 kPa.

(0.20–0.22), Cordierite in reaction coronas around staurolite porphyroblasts has an average Mg# of ~0.67 (0.65–0.68). Matrix plagioclase averages ~An15 (An14–An36.5), and matrix biotite has an average Mg# of ~0.48 (0.43–0.51) and XTi of ~0.05 (0.04–0.07).

Thermodynamic modelling

A MnNCKFMASHTO (MnO–Na2O–CaO–K2O–FeO–MgO–Al2O3–SiO2–H2O–TiO2–Fe2O3) isochemical phase diagram for representative Sil + St schist sample CL1003 is shown in Fig. 13A. The significant amount of Fe2O3 (3.2 wt%) in the sample (Appendix A) requires incorporation of both Fe2+ and Fe3+ in the chemical system, which reduces the stability range of garnet. The presence of coexisting sillimanite and staurolite in the absence of muscovite or chlorite constrains the peak assemblage to a small stability field (Field 1) at 600 ± 15 °C and 5.5 ± 0.25 kbar. Thermobarochemistry estimates from nearby garnet-bearing sample CL1001A (detailed in Fig. 13A) give a P–T intersection at ~580 °C, 5.75 kbar. Cordierite haloes around staurolite in CL1003 imply 1–2 kbar of decompression into the dashed Sil + Crd + Bt + Pl + Qtz + Ilm + Mt + H2O stability field (Fig. 13A). As in the gneisses, the local development of cordierite as reaction rims around staurolite, as opposed to more extensive recrystallization of the rock, suggests that decompression occurred with uniform or decreasing temperature. Petrology of garnet + biotite schists

Pelitic schist sample CL1001A, collected less than 10 m from CL1003, contains a staurolite-absent Grt + Bt + Pl + Qtz + Ilm + Kfs assemblage. Anhedral to subhedral garnets have inclusion-rich cores with thinner, inclusion-poor rims (Fig. 9F), and Bt + Qtz inclusion trails (S1) in garnet are oriented at a high angle to the external schistosity (S2). Biotite-absent, Pl + Qtz-bearing haloes around garnet are up to 1 mm in width. Potassium feldspar in the matrix of CL1001A, though still minor (<1%), is more abundant than in St-schists, although whether it is part of the stable mineral assemblage or represents unreacted detrital material is unclear.

Mineral chemistry

The X Mn (X Mn = Mn/(Mn + Ca + Fe + Mg)) in garnet cores averages ~0.22 (0.20–0.24), whereas rim values average ~0.12 (0.10–0.15) (Table 1; Fig. 10B). Core X Ca (X Ca = Ca/(Mn + Ca + Fe + Mg)) values average ~0.13 (0.12–0.14), and rim values average ~0.08 (0.07–0.09). The Mg# in garnet cores is lower (Mg# = ~0.15) than rim values (Mg# = 0.160–0.175). The textural and chemical relationships are consistent with two discrete phases of garnet growth (Fig. 10B). Garnet cores are rich in Mn + Ca, whereas garnet rims show depletions in Mn + Ca and a concurrent rise in Fe + Mg. Matrix plagioclase grains in CL1001A (An37–An40) are Ca-poorer than those in contact with garnet (An45–An49). Matrix biotite has Mg# compositions of 0.53–0.55 and an average X Ti of ~0.04 (0.03–0.04). Full thermodynamic modelling was not conducted because the limited number of phases in the sample resulted in poor P–T constraints in preliminary isochemical phase diagram sections.
Metabasites

Amphibolites in the southern Grand Forks complex were analyzed by Preto (1970a) and interpreted to have an igneous origin. The amphibolites are generally well-foliated schists and gneisses, often crosscut by granitic dikes and veinlets (Fig. 4C). Amphibolites throughout the complex contain Hbl + Pl + Qtz ± Bt ± Ttn (Ttn, titanite) with varying amounts of garnet, orthopyroxene, and clinopyroxene (Fig. 8). In units Pr1 and Pr3, two types of Opx + Hbl ± Cpx metabasites were observed: Type A amphibolites are garnet bearing, with only minor matrix Cpx + Opx, whereas Type B amphibolites (Fig. 9G) have coarse-grained orthopyroxene and clinopyroxene in the matrix but no garnet. Garnet within Type A amphibolites is typically large (up to 3–4 cm in size) and poikiloblastic. In rocks with matrix clinopyroxene, garnet is commonly partially or fully replaced (80%–100%) by symplectite overgrowths of Cpx + Opx + Pl + Mt (Fig. 9H), whereas in rocks with no matrix clinopyroxene, garnet shows little retrogression.

Most of the metamorphic mineral assemblages in units Pr1 and Pr3 indicate upper-amphibolite-facies conditions (Hbl + Pl ± Bt ± Cpx). However, some samples along the eastern margin of the complex contain orthopyroxene (Fig. 8), indicative of granulite-facies conditions. No orthopyroxene was reported in the western part of the GFC either in this study or by Laberge and Pattison (2007). This pattern may suggest higher P–T conditions along the eastern margin of the GFC or may be due to bulk compositional variation.

Amphibolites in the stratigraphically uppermost units of the GFC, Pr4 and Pr5, are finer grained than in units Pr1 and Pr3 and lack clinopyroxene or orthopyroxene. Garnet in amphibolites from units Pr4 and Pr5 is poikiloblastic and unzoned, and symplectites are absent. The modal amount of biotite is significantly higher in amphibolites from units Pr4 and Pr5 than from units Pr1 and Pr3.

Petrology of clinopyroxene + orthopyroxene + hornblende amphibolites

Type A Cpx + Opx + Hbl amphibolites yield the greatest insight into the P–T conditions of both peak metamorphic and low-pressure decomposition assemblages. Representative sample CL596 contains poikiloblastic garnet porphyroblasts 1 mm – 3 cm in diameter that are nearly completely replaced by a symplectite composed of orthopyroxene, clinopyroxene, plagioclase, and magnetite (Fig. 9H). The modal mineralogy of the symplectite is ~50% orthopyroxene, 5% clinopyroxene, 40% plagioclase, and 5% magnetite, with trace amounts of ilmenite, biotite, hornblende, and highly resorbed garnet (<<1%). The foliated rock matrix is made up of Hbl + Pl + Qtz + Ilm + Ap, with minor amounts of clinopyroxene and orthopyroxene in reaction rims surrounding hornblende and as small matrix grains (Cpx).
Mineral chemistry

Matrix hornblende in CL596 has Mg# values ranging from 0.45 to 0.47, aluminum contents ranging from 1.75 to 1.95 cations per formula unit (p.f.u.) and an XTi of 0.04 (0.03–0.05). Hornblende within the symplectites has slightly higher Mg# values, ranging from 0.46 to 0.53, with similar aluminum contents (Table 2). Clinopyroxene within the matrix and in symplectites has the same range of Mg# values (0.59–0.61) and aluminum contents between 0.05 and 0.07 cations p.f.u. Orthopyroxene in the matrix and in symplectites has Mg# values ranging from 0.46 to 0.47 and aluminum contents between 0.03 and 0.05 cations p.f.u. Plagioclase in the matrix of the rock has an average XAn of ~0.86 (0.83–0.87), whereas plagioclase within the symplectite structures has a higher XAn, averaging ~0.89 (0.88–0.90). Highly resorbed garnet fragments in the symplectite have Mg# values of 0.22–0.23, with an average XMn of ~0.02 and XCa of ~0.15.

Thermodynamic modelling

A MnNCKFMASHT isochemical phase diagram for representative amphibolite sample CL596 is shown in Fig. 13B. Whereas melt models exist for granitic melt relevant to metapelites (e.g., Holland and Powell 1998; White et al. 2001, 2007), at the present time no melt model exists for metab-
sites. Amphibolites were therefore modelled with melt suppressed in the thermodynamic database, recognizing that this introduces uncertainties to the following estimates. Water content was estimated from mineral stoichiometry of the /H11011 60 vol% hornblende in the sample (Table 2). The matrix assemblage, prior to development of the symplectite structures, is Cpx /H11001 Opx /H11001 Hbl /H11001 Grt /H11001 Pl /H11001 Ilm /H11001 Qtz. This assemblage falls within a thin stability field between 755–780 °C, with pressure unconstrained above 3.5 kbar (Field 6, outlined in solid black). The metabasite temperature constraint agrees well with that determined from migmatitic paragneisses (Fig. 11F).

The Cpx + Opx + Pl + Mt symplectites replacing garnet indicate decompression to below ~3.5 kbar, similar to the paragneiss coronal assemblage estimate of ~3.3 kbar. A simplified reaction that accounts for the occurrence of magnetite in Cpx /H11001 Opx /H11001 Pl /H11001 Mt symplectites is Reaction 6:

\[
\text{Cpx} + \text{Grt} + \text{O}_2 = \text{Cpx}_2 + \text{Opx} + \text{Mt} + \text{Pl} + \text{Qtz}
\]

as suggested by Choudhuri and Silva (2000). The presence of magnetite in the symplectites implies that oxidation was involved in symplectite formation. Unlike that found by

<table>
<thead>
<tr>
<th>Biotite</th>
<th>Plagioclase</th>
<th>Garnet</th>
</tr>
</thead>
<tbody>
<tr>
<td>CL38(I)</td>
<td>CL905(IV)</td>
<td>CL1204(V)</td>
</tr>
<tr>
<td>Matrix</td>
<td>Matrix</td>
<td>Matrix</td>
</tr>
<tr>
<td>37.26</td>
<td>36.02</td>
<td>35.29</td>
</tr>
<tr>
<td>4.51</td>
<td>4.04</td>
<td>2.54</td>
</tr>
<tr>
<td>13.67</td>
<td>21.00</td>
<td>21.73</td>
</tr>
<tr>
<td>0.08</td>
<td>0.08</td>
<td>0.19</td>
</tr>
<tr>
<td>15.10</td>
<td>11.52</td>
<td>11.96</td>
</tr>
<tr>
<td>0.04</td>
<td>0.03</td>
<td>0.20</td>
</tr>
<tr>
<td>0.17</td>
<td>0.14</td>
<td>0.10</td>
</tr>
<tr>
<td>8.95</td>
<td>8.79</td>
<td>7.77</td>
</tr>
<tr>
<td>0.06</td>
<td>0.42</td>
<td>0.28</td>
</tr>
<tr>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>0.02</td>
<td>0.20</td>
<td>0.15</td>
</tr>
<tr>
<td>—</td>
<td>0.02</td>
<td>0.02</td>
</tr>
<tr>
<td>94.48</td>
<td>96.76</td>
<td>94.80</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Plagioclase</th>
<th>Garnet</th>
</tr>
</thead>
<tbody>
<tr>
<td>Matrix</td>
<td>Matrix</td>
</tr>
<tr>
<td>2.79</td>
<td>2.74</td>
</tr>
<tr>
<td>0.25</td>
<td>0.23</td>
</tr>
<tr>
<td>1.29</td>
<td>1.26</td>
</tr>
<tr>
<td>0.86</td>
<td>1.34</td>
</tr>
<tr>
<td>0.01</td>
<td>0.01</td>
</tr>
<tr>
<td>1.68</td>
<td>1.31</td>
</tr>
<tr>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0.03</td>
<td>0.02</td>
</tr>
<tr>
<td>0.85</td>
<td>0.89</td>
</tr>
<tr>
<td>0.00</td>
<td>0.01</td>
</tr>
<tr>
<td>0.00</td>
<td>0.05</td>
</tr>
<tr>
<td>7.75</td>
<td>7.81</td>
</tr>
<tr>
<td>0.66</td>
<td>0.49</td>
</tr>
<tr>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>—</td>
<td>0.00</td>
</tr>
<tr>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>0.60</td>
<td>0.17</td>
</tr>
<tr>
<td>0.38</td>
<td>0.83</td>
</tr>
<tr>
<td>0.03</td>
<td>0.01</td>
</tr>
</tbody>
</table>

plagioclase, and garnet, respectively. Unit numbers are shown without the “Pr” prefix for to conserve space. M/FM = Mg/(Mg + Fe); XMn = Mn/(Mn + Mg)

Can. J. Earth Sci. Downloaded from www.nrcresearchpress.com by UNIV CALGARY on 12/20/12
For personal use only.
Choudhuri and Silva (2000), the clinopyroxene compositions do not differ significantly from matrix (Cpx1) to symplectite (Cpx2). Symplectic textures imply either or both of localized effective bulk compositions, or establishment of localized chemical potential gradients, between unstable minerals, complicating interpretations of $P$-$T$ conditions using Fig. 13B. Nevertheless, the agreement between metamabase and paragneisses with respect to conditions of peak metamorphism and corona or symplectic formation is notable.

Summary of Grand Forks complex metamorphism

Peak metamorphic conditions in the lower units (Pr1, Pr3) of the Grand Forks complex record upper-amphibolite- to granulite-facies metamorphic conditions in migmatitic paragneisses and amphibolites. Isochemical phase diagrams and mineral isopleth thermobarometry from a representative paragneiss, CL831, provide a good temperature estimate of $\sim 750 \pm 25 \, ^\circ C$, whereas additional barometry estimates constrain pressure to $\sim 5.6 \pm 0.5 \, \text{kbar}$. Low-pressure coronal amphibolite-facies in regional metamorphic and deformational events that peaked in the Middle Jurassic, prior to the intrusion of the Middle Jurassic Nelson intrusive suite (Acton 1998; Acton et al. 2002; Höy 2006). The distribution of metamorphic mineral assemblages in the hanging wall of the KRF is shown in Fig. 8. Separating regional metamorphism from contact metamorphism is difficult because the two overlap spatially and some mineral assemblages could belong to either.

Regional metamorphic assemblages are most clearly defined in metabasites and volcanioclastics of the Mollie Creek assemblage and Rossland Group. Greenschist-facies assemblages (e.g., Chl + Act + Ab + Ep + Cal; Act, actinolite; Ab, albite; Ep, epidote) dominate, but there are also local transition amphibolite-facies assemblages (e.g., Hbl + Act + Pl). Contact metamorphic assemblages are best characterized by Crd + And + Kfs (And, andalusite) schists and hornfelses of the Mollie Creek assemblage, which grade into migmatitic pelites approaching the Nelson suite intrusives. Semipelitic, marble, and calc-silicate lithologies lack mineral assemblages that yield good metamorphic constraints.

Petrography, mineral chemistry, and thermodynamic modelling of metabasites (regional metamorphism)

Minor volcanic greenstones and volcanic clasts within volcanioclastic rocks of the Rossland Group are characterized by Cpx + Bt + Pl + Qtz + Ttn + Act + Chl + Hbl + Ep + Cal assemblages interpreted to represent a combination of relict igneous and metamorphic minerals (Figs. 14A, 14B). Actinolite partially to wholly replaces clinopyroxene phenocrysts, and chlorite replaces biotite. Actinolite and chlorite, with or without associated epidote, additionally occur as fine-grained matrix minerals and collectively reflect greenschist-facies metamorphic conditions. Elemental mapping shows intergrowths of hornblende and actinolite in the matrix of some samples, suggestive of transitional greenschist–amphibolite-facies metamorphism like that described in the hanging wall of the Granby fault (LaBerge and Patterson 2007). Cpx + Bt andesitic volcanic clasts in volcanioclastic rocks show similar intergrowths (e.g., CL478; Figs. 4G, 14A, 14B). Close to the Kettle River fault, the matrix of Rossland Group volcanic and volcanioclastic rocks contains late, coarse Ep + Cal + Py + Qtz (Py, pyrite) mineralization.

Mineral chemistry

Mineral chemical analyses are listed in Table 3 for two samples in close proximity (Fig. 5): (1) CL478, a volcanioclastic sample with a Hbl + Act + Ep + Chl + Pl + Qtz + IIm metamorphic assemblage, and CL407, a homogenous volcanic sample characterized by an Act + Chl + Pl + Qtz + Kfs + IIm assemblage. In CL478, actinolite within Hbl + Act intergrowths (Figs. 14A, 14B) after clinopyroxene has an average aluminum content of $\sim 0.48$ cations p.f.u. (0.28–0.60), Na/Ca ratios of 0.04–0.05, and an average Mg# of $\sim 0.63$ (0.62–0.64). Hornblende within these intergrowths has an average aluminum content of $\sim 2.14$ cations p.f.u. (1.97–2.29), an average Na/Ca ratio of 0.19 (0.16–0.24) and an average Mg# of 0.50 (0.48–0.52). Matrix plagioclase has an average X$_{Ab}$ of 0.10 (An$_{37}$–An$_{55}$). CL407 contains more calcic plagioclase, averaging X$_{An}$ = 0.28 (0.26–0.28), but similar actinolite compositions, with an average Al content of 0.45 cations p.f.u. (0.30–0.53) and an average Na/Ca ratio of 0.04 and Mg# of $\sim 0.62$ (0.60–0.65).

Thermodynamic modelling

Phase equilibrium modelling of Rossland Group metavolcanics and volcanioclastics resulted in inconclusive results because of a number of factors, one of which is uncertainty over the effective (reactive) bulk composition. For example, chemical components in remnant igneous minerals and isolated volcanic clasts were unavailable for metamorphic growth yet formed part of the measured bulk-rock composition. For several samples, the predicted stability field for the mineral assemblage observed in the rocks covered such a wide range of pressure and temperature to be of limited use. For example, sample CL407, a relatively homogenous metavolcanic rock with only a few relict clinopyroxene phenocrysts, was modelled in the system NCFMASH.
(Na₂O–CaO–FeO–MgO–Al₂O₃–SiO₂–H₂O) and yielded a broad stability field for the observed Act/Hbl/Pl/Qtz/Kfs/IIm assemblage (Fig. 15). Plagioclase and actinolite mineral composition isopleths from CL407 fall at or just below the lower temperature limit of the assemblage stability field, ~400–500 °C.

Thermobarometry was conducted on Hbl + Act + Pl-bearing sample CL478 using mineral equilibria involving end-members of amphibole, plagioclase, and quartz (Fig. 15). The equilibria used (abbreviations from Berman 1991) and resultant pressures, for temperatures between 400 and 450 °C, are as follows: 3Tr + 5FeTs = 5Tsc + 3FeTr (~1.7–1.8 kbar); 8FeTs + 3Tr + 6Ab = 24aQz + 5Tsc + 6FePa (<2.5 kbar); 5Parg + 4FeTr = 4Tr + 5FePa (<3.2 kbar) (Tr, tremolite; FeTs, ferro-tschermakite; FeTr, ferro-tremolite; Ab, albite; aQz, alpha quartz; Tsc, tschermakite; FePa, ferro-pargasite; Parg, pargasite). These estimates are broadly consistent with the constraints of Bégin (1992), Dale et al. (2005), and Elmer et al. (2006) on the stability of Hbl + Act + Pl in typical metabasic bulk compositions (T ~ 400–480 °C, P < ~5 kbar). Combining all constraints, P–T conditions for the regional metamorphism of metabasic rocks in the KRF hanging wall are therefore poorly constrained to ~425 ± 25 °C and 2.2 ± 0.6 kbar, similar to the estimates of Laberge & Pattison (2007) (425 ± 40 °C and...
Table 3. Representative mineral compositions in pelitic, volcanic, and volcaniclastic lithologies from the hanging wall of the Kettle River fault (KRF).

<table>
<thead>
<tr>
<th>Sample:</th>
<th>Garnet</th>
<th>Biotite</th>
<th>Plagioclase</th>
<th>Amphibole</th>
<th>Cordierite</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Core</td>
<td>Rim</td>
<td>Matrix</td>
<td>Matrix</td>
<td>Matrix</td>
</tr>
<tr>
<td>CL070</td>
<td>37.30</td>
<td>37.23</td>
<td>35.52</td>
<td>57.48</td>
<td>44.52</td>
</tr>
<tr>
<td>CL070</td>
<td>37.23</td>
<td>34.56</td>
<td>35.02</td>
<td>58.37</td>
<td>42.33</td>
</tr>
<tr>
<td>CL066</td>
<td>57.48</td>
<td>56.50</td>
<td>61.31</td>
<td>58.37</td>
<td>54.24</td>
</tr>
<tr>
<td>CL070</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CL066</td>
<td>61.31</td>
<td>58.37</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CL070</td>
<td>56.50</td>
<td>58.37</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CL407</td>
<td>44.52</td>
<td>42.33</td>
<td></td>
<td>44.92</td>
<td>47.89</td>
</tr>
<tr>
<td>CL478</td>
<td>42.33</td>
<td>54.24</td>
<td></td>
<td>52.43</td>
<td>48.24</td>
</tr>
<tr>
<td>CL478</td>
<td>54.24</td>
<td>47.89</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CL478</td>
<td>47.89</td>
<td>48.24</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CL066</td>
<td>48.24</td>
<td>47.89</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CL070</td>
<td>48.24</td>
<td>47.89</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Note:</td>
<td>Core</td>
<td>Rim</td>
<td>Matrix</td>
<td>Matrix</td>
<td>Matrix</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Oxide</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SiO₂</td>
<td>37.30</td>
<td>37.23</td>
<td>35.52</td>
<td>57.48</td>
<td>44.52</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.04</td>
<td>0.05</td>
<td>3.52</td>
<td>1.92</td>
<td>0.25</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>20.94</td>
<td>20.91</td>
<td>20.27</td>
<td>17.13</td>
<td>12.76</td>
</tr>
<tr>
<td>FeO</td>
<td>33.41</td>
<td>33.47</td>
<td>20.93</td>
<td>20.56</td>
<td>28.46</td>
</tr>
<tr>
<td>MnO</td>
<td>2.89</td>
<td>3.09</td>
<td>0.21</td>
<td>0.06</td>
<td>0.73</td>
</tr>
<tr>
<td>MgO</td>
<td>3.41</td>
<td>3.30</td>
<td>7.02</td>
<td>13.07</td>
<td>10.49</td>
</tr>
<tr>
<td>CaO</td>
<td>2.04</td>
<td>1.94</td>
<td>9.01</td>
<td>7.53</td>
<td>9.70</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.12</td>
<td>0.15</td>
<td>0.23</td>
<td>0.09</td>
<td>0.03</td>
</tr>
<tr>
<td>K₂O</td>
<td>8.83</td>
<td>7.91</td>
<td>0.23</td>
<td>0.02</td>
<td>0.03</td>
</tr>
<tr>
<td>BaO</td>
<td>0.19</td>
<td>0.05</td>
<td>0.19</td>
<td>0.29</td>
<td>0.04</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>0.00</td>
<td>0.00</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Y₂O₃</td>
<td>0.32</td>
<td>0.02</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>F</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td>100.35</td>
<td>100.01</td>
<td>95.73</td>
<td>96.23</td>
<td>98.66</td>
</tr>
<tr>
<td>Cations</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Si⁴⁺</td>
<td>2.99</td>
<td>3.00</td>
<td>2.64</td>
<td>2.65</td>
<td>6.67</td>
</tr>
<tr>
<td>Ti⁴⁺</td>
<td>0.00</td>
<td>0.00</td>
<td>0.20</td>
<td>0.11</td>
<td>0.03</td>
</tr>
<tr>
<td>Al³⁺</td>
<td>1.98</td>
<td>1.98</td>
<td>1.82</td>
<td>1.53</td>
<td>2.25</td>
</tr>
<tr>
<td>Fe³⁺</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fe²⁺</td>
<td>2.24</td>
<td>2.25</td>
<td>1.34</td>
<td>1.30</td>
<td>3.57</td>
</tr>
<tr>
<td>Mn²⁺</td>
<td>0.20</td>
<td>0.21</td>
<td>0.01</td>
<td>0.00</td>
<td>0.09</td>
</tr>
<tr>
<td>Mg²⁺</td>
<td>0.41</td>
<td>0.40</td>
<td>0.80</td>
<td>1.47</td>
<td>2.34</td>
</tr>
<tr>
<td>Ca²⁺</td>
<td>0.18</td>
<td>0.17</td>
<td>0.00</td>
<td>0.01</td>
<td>0.06</td>
</tr>
<tr>
<td>Na⁺</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>K⁺</td>
<td>0.86</td>
<td>0.76</td>
<td>0.01</td>
<td>0.00</td>
<td>0.01</td>
</tr>
<tr>
<td>Ba²⁺</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cr³⁺</td>
<td>0.00</td>
<td>0.00</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Y³⁺</td>
<td>0.01</td>
<td>0.00</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>F⁻</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sum</td>
<td>8.01</td>
<td>8.01</td>
<td>7.69</td>
<td>7.88</td>
<td>15.33</td>
</tr>
<tr>
<td>M/FM</td>
<td>0.15</td>
<td>0.15</td>
<td>0.37</td>
<td>0.53</td>
<td>0.40</td>
</tr>
<tr>
<td>Xₘ₉</td>
<td>0.06</td>
<td>0.07</td>
<td></td>
<td></td>
<td>0.40</td>
</tr>
<tr>
<td>Xₙ₉</td>
<td>0.06</td>
<td>0.06</td>
<td></td>
<td></td>
<td>0.51</td>
</tr>
<tr>
<td>Xₙ₉</td>
<td>0.06</td>
<td>0.06</td>
<td></td>
<td></td>
<td>0.51</td>
</tr>
<tr>
<td>Xₙ₉</td>
<td>0.06</td>
<td>0.06</td>
<td></td>
<td></td>
<td>0.51</td>
</tr>
</tbody>
</table>

Note: Number of cations is based on structural formulae with 18, 12, 22, 23, and 8 oxygens for cordierite, garnet, biotite, amphibole, and plagioclase, respectively. M/FM = Mg/(Mg+Fe); Xₘ₉ = Mn/ (Mn+Mg+Ca+Fe²⁺); Xₙ₉ = Ca/(Mn+Mg+Ca+Fe²⁺); Xₙ₉ = Na/(Na+K+Ca); Xₙ₉ = Ca/(Na+K+Ca); Xₙ₉ = K/(Na+K+Ca).
Mineral chemistry

Dant throughout the rock matrix. Small, anhedral K-feldspar grains are abundant, wrapped by a late foliation (Fig. 14C). Andalusite is rare located Bt. Matrix plagioclase has anorthite contents averaging ~An 44. K-feldspar metapelites 2.3 ± 0.7 kbar) for rocks in the hanging wall of the Granby fault on the west flank of the GFC.

Petrography, mineral chemistry, and thermodynamic modelling of metapelites (contact metamorphism)

Non-migmatitic cordierite + biotite ± andalusite ± K-feldspar metapelites

Within 1.5 km of intrusive rocks of the Jurassic Nelson suite, pelitic lithologies within the Mollie Creek assemblage are characterized by spotted Crd + Bt + Pl + Qtz + Illm ± And ± Kfs hornfelses and schists (Fig. 4E). These assemblages contain large cordierite porphyroblasts in a fine-grained Bt + Pl + Qtz matrix. Representative sample CL066 (Fig. 5), located ~600 m from the intrusive contact, has a Crd + And + Bt + Kfs + Pl + Qtz + Illm assemblage. Cordierite forms large ovoid porphyroblasts up to 6 mm in length that are wrapped by a late foliation (Fig. 14C). Andalusite is rare (<1%) and restricted to thin zones between adjacent cordierite porphyroblasts. Small, anhedral K-feldspar grains are abundant throughout the rock matrix.

Mineral chemistry

Cordierite porphyroblasts in CL066 are compositionally homogeneous, with Mg# values averaging ~0.52 (0.51–0.53) and X_Mn = Mn/(Mn + Fe + Mg) ranging from 0.025 to 0.027 (Table 3). Biotite Mg# values average ~0.37 (0.36–0.38), with X_Ti averaging ~0.06 (0.05–0.07), independent of textural setting. Matrix plagioclase has anorthite contents averaging ~An_{44} (An_{43}–An_{45}).

Thermodynamic modelling

Figures 16A, 16B show a MnNCKFMAHST isochronal phase diagram and selected compositional isopleths, respectively, for sample CL066. The stability field for the peak assemblage, Crd + And + Bt + Kfs + Pl + Qtz + Ilm (Field 5) provides P–T constraints of 615–625 °C and 1.8–2.7 kbar (Fig. 16A). The choice of this stability field assumes that the matrix K-feldspar present in CL066 is of subsolidus origin, consistent with the lack of migmatitic textures. Cordierite Mg# isopleths, biotite Mg#, and plagioclase anorthite contents corresponding to measured values pass close to this P–T range (Fig. 16B). Combining the isopleth results with mineral assemblage constraints (Fig. 16B), a P–T estimate of 650 ± 25 °C, 2.5 ± 0.5 kbar is proposed for sample CL066.

Gedrite + garnet + cordierite metapelites

Rare gedrite-bearing metasediments are interlayered with Crd-bearing pelites in the Mollie Creek assemblage just east of Christina Lake. The modal mineralogy of sample CL070, a representative Ged + Grt + Crd + Bt + Pl + Qtz + Gr hornfels located ~500 m from the intrusive contact, is ~50% biotite, 20% gedrite, 5% garnet, 1% cordierite, and 25% plagioclase and quartz. Gedrite forms acicular splays intergrown with small garnets (Fig. 14D) and is typically unoriented relative to the foliation. Large, heavily altered cordierite porphyroblasts have inclusions of Grt, Ged, Pl, and Qtz.

Mineral chemistry

Garnet in sample CL070 is weakly zoned and has Mg# values averaging ~0.15 (0.13–0.15) (Table 3), X_Mn averaging ~0.07 (0.06–0.08), and X_Ca averaging ~0.06 (0.05–0.08). Plagioclase in CL070 averages ~An_{39} (An_{37}–An_{40}). Gedrite Mg# values average ~0.40 (0.39–0.41), cordierite Mg# values average ~0.59 (0.59–0.60), and biotite Mg# values range from 0.52 to 0.53, with an average X_Ti of ~0.04.

Thermodynamic modelling

A MnNCKFMAHST isochronal phase diagram and compositional isopleths are shown in Figs. 16C, 16D, respectively, for Ged + Grt + Crd-bearing sample CL070. The stability field for the peak assemblage is large and provides poor P–T constraints (Fig. 16C). Isopleths of Mg# of cordierite, gedrite, biotite, and garnet show poor convergence but suggest pressures below 4 kbar (Fig. 16D). Garnet–biotite Fe–Mg exchange thermometry using THERMOCALC (phl + alm = ann + py; phl, phlogopite; alm, almandine; ann, annite; py, pyrope) yields temperature estimates of 615–620 °C (Fig. 16D). The barometer 2an + 3east + 3py + 9qtz = 5phl + 3fcrd (ann, annite; east, eastonite; py, pyrope; qtz, quartz; phl, phlogopite; fcrd, ferro-cordierite) suggests pressures of ~2.4 kbar at 600 °C, whereas the barometer 12grun + 21mncrd = 28alm + 14psps + 75qtz + 12H_2O (grun, grunerite; mncrd, Mn-cordierite; alm, almandine; spss, spessartine; qtz, quartz) yields pressure estimates of 1.8 kbar at 600 °C. Combining all methods, P–T conditions of 630 ± 30 °C and 2.4 ± 0.6 kbar are inferred, similar to CL066.

Migmatitic cordierite-bearing metapelites

A zone of migmatization around the Middle Jurassic Nelson intrusive suite in the Sutherland Creek drainage east of Christina Lake extends for at least 500 m from the intrusion –
Fig. 16. MnNCKFMASHT isochemical phase diagrams and summary isopleth diagrams for CL066 (Crd + And + Kfs metapelite) and CL070 (Ged + Grt + Cordierite grains overgrow an earlier foliation defined by strong biotite inclusion trails, corroborating the observation of Acton et al. (2002) that emplacement of the Nelson intrusive suite and its associated contact metamorphism postdates the dominant D2 regional deformation in the hanging wall.

country rock contact (e.g., CL294; Fig. 5). Samples are characterized by Crd + Bt + Kfs + Pl + Qtz + IIm ± Sil ± Grt assemblages. In the field, thin stromatic anatectic leucosomes occur parallel to the dominant foliation, but wider crosscutting leucosomes in net-textured migmatites truncate preexisting fabrics, including the stromatic migmatites (Fig. 4F).
Fig. 17. Summary pressure–temperature (P–T) path for exhumation of the Grand Forks complex (GFC), highlighting the multiple exhumation stages. Peak metamorphism at ~59–50 Ma is followed by ~2.3 kbar of decompression at high temperatures (725–750 °C), leading to the formation of the decompression assemblages (Crd + Ilm + Spl, Crd + Qtz). The GFC then had to cool ~300 °C before low-T (~415 °C) movement along the Kettle River fault (KRF) at ~49 Ma, juxtaposing it against the currently exposed hanging wall. The maximum displacement on the KRF is ~0.8 ± 0.7 kbar, assuming isobaric cooling of the GFC following high-T decompression. This is only 25% of the apparent pressure contrast across the KRF. 1 kbar = 100 MPa.

Summary of hanging-wall regional and contact metamorphism

Acton (1998) speculated that some Crd + Bt-bearing, non-migmatitic assemblages in the KRF hanging wall might be attributed to regional metamorphism in the Middle Jurassic. The pressure–temperature estimates from this study for similar samples are 2.5 ± 0.5 kbar and 630–650 °C, implying a transitory geothermal gradient at the time of metamorphism of 70–80 °C·km–1. Consequently, we propose that these Crd-bearing rocks are due to contact metamorphism from the emplacement of the Middle Jurassic Nelson intrusive suite. The hanging-wall migmatitic samples in the Sutherland Creek drainage east of Christina Lake record even higher metamorphic temperatures.

Metabasic samples of transitional greenschist–amphibolite-facies suggest regional metamorphic temperature conditions away from the intrusions of ~425 ± 25 °C, nearly 200 °C below contact metamorphism temperature estimates (Fig. 17). The ~2.2 ± 0.6 kbar pressure estimate for the regional rocks overlaps with that for the contact metamorphic rocks (2.5 ± 0.5 kbar), implying limited burial or exhumation between regional and contact metamorphism.

Discussion

High-temperature metamorphism and decompression in the Grand Forks complex

Peak metamorphic conditions in metapelitic and metabasic gneisses of the GFC are estimated to be 750 ± 25 °C and 5.6 ± 0.5 kbar. The timing of this metamorphism is constrained to 59–50 Ma, based on U–Pb geochronology of metamorphic monazite (Cubley et al. 2011; Cubley et al. b, accepted for publication). Coronal assemblages in both rock types imply conditions of ~735 ± 20 °C and 3.3 ± 0.5 kbar. Combined, these P–T estimates suggest that the Grand Forks complex underwent high-temperature, upper-amphibolite-facies decompression on the order of 2.3 kbar, equivalent to ~8 km (assuming an average crustal density of 2.85 g·cm–3; Fig. 17). These estimates are comparable to the ~2 ± 1 kbar (7 ± 3.5 km) estimate from the western part of the GFC reported by Laberge and Pattison (2007), who documented peak metamorphism at 800 ± 35 °C and 5.8 ± 0.6 kbar and coronal development at 750 ± 30 °C and 3–4 kbar. The results of this study are also within error of preliminary estimates from the eastern edge of the GFC by Cubley and Pattison (2009). Lack of systematic variation of mineral assemblages, coronal textures, and related P–T results across the lower units (Pr1–Pr3) of the GFC suggests little variation in peak metamorphism and extent of high-temperature decompression across the complex.

Lower-temperature, middle-amphibolite-facies Sill + St schists from the stratigraphically overlying unit Pr5 record peak P–T conditions of 600 ± 15 °C and 5.5 ± 0.25 kbar, followed by high-T decompression of 1–2 kbar (3–6 km). The lower peak metamorphic temperatures and smaller magnitude of exhumation are consistent with the higher stratigraphic position of these schists. On the other hand, the similarity of calculated metamorphic pressure between the St–Sill schists and underlying migmatites is problematic.

The near-isothermal decompression recorded in the GFC is consistent with a rapid exhumation event (Fig. 17). Rapid exhumation is consistent with previous results from the west-
At about 10% of the time of footwall peak metamorphism, resulting in a greater fault–

high-temperature stage accommodates

The extent to which this depth contrast across the KRF

measures the vertical separation across the fault due to Eocene,

low-temperature movement depends on a number of assump-
tions. The estimate previously discussed would be applicable

If the hanging-wall sedimentary rocks were buried further prior to Paleocene–Eocene meta-
morphism in the footwall, the pressure contrast between hanging-wall and footwall assemblages would be reduced compared with the estimate discussed earlier. There is no petrographic evidence of such overprinting of contact meta-
morphic assemblages in the study area. Alternatively, if signif-
ificant erosion of the hanging wall occurred between the Middle Jurassic and the Paleocene–Eocene, the hanging wall could have been at a depth less than 9.0 km (2.5 kbar) at the time of footwall peak metamorphism, resulting in a greater fault-
related displacement compared with the estimate discussed ear-
lier. Understanding these limitations, we consider ~3 km to be the best estimate for vertical separation across the Kettle River fault. This vertical separation estimate is significantly less than the >15 km proposed by Parrish et al. (1988) but within error of an earlier 5 km estimate by Bowman (1950).

Two-stage exhumation of the Grand Forks complex and comparison with other core complexes of the Shuswap domain

The data just discussed suggest that exhumation of the GFC occurred in two stages: the first at high temperatures (upper-amphibolite- to granulite-facies) and the second at low tem-

peratures (greenschist-facies) in response to movement on the Eocene Kettle River and Granby normal faults (Fig. 17). The time of footwall peak metamorphism, resulting in a greater fault–

related displacement compared with the estimate discussed ear-

lier. Understanding these limitations, we consider ~3 km to be the best estimate for vertical separation across the Kettle River fault. This vertical separation estimate is significantly less than the >15 km proposed by Parrish et al. (1988) but within error of an earlier 5 km estimate by Bowman (1950).

Two-stage exhumation of the Grand Forks complex and comparison with other core complexes of the Shuswap domain

The data just discussed suggest that exhumation of the GFC occurred in two stages: the first at high temperatures (upper-amphibolite- to granulite-facies) and the second at low tem-

peratures (greenschist-facies) in response to movement on the Eocene Kettle River and Granby normal faults (Fig. 17). The high-temperature stage accommodates ~75% of overall exhu-
mation, whereas the subsequent low-temperature stage ac-
counts for the remaining 25%.

Offset across the Kettle River fault

The pressure difference between the low-pressure decom-
pression assemblage of the Grand Forks complex (3.3 ± 0.5 kbar) and the hanging-wall metapelites (2.5 ± 0.5 kbar) is 0.8 ± 0.7 kbar (statistically combining the two error estimates) or 2.9 ± 2.5 km (Fig. 17). This estimate is lower than, but within error of, estimates for the Granby fault by Laberge and Pattison (2007) (i.e., 1.3 ± 0.9 kbar or 5 ± 3.5 km). A vertical separation of ~1 kbar (3 km) on these faults is considerably less than the ~3.1 kbar (11 km) difference implied by the peak metamorphic conditions of the dramatically different meta-
morphic assemblages on either side of the fault.

The extent to which this depth contrast across the KRF

measures the vertical separation across the fault due to Eocene,

low-temperature movement depends on a number of assump-
tions. The estimate previously discussed would be applicable

If the hanging-wall sedimentary rocks were buried further prior to Paleocene–Eocene meta-
morphism in the footwall, the pressure contrast between hanging-wall and footwall assemblages would be reduced compared with the estimate discussed earlier. There is no petrographic evidence of such overprinting of contact meta-
morphic assemblages in the study area. Alternatively, if signif-
ificant erosion of the hanging wall occurred between the Middle Jurassic and the Paleocene–Eocene, the hanging wall could have been at a depth less than 9.0 km (2.5 kbar) at the time of footwall peak metamorphism, resulting in a greater fault-
related displacement compared with the estimate discussed ear-
lier. Understanding these limitations, we consider ~3 km to be the best estimate for vertical separation across the Kettle River fault. This vertical separation estimate is significantly less than the >15 km proposed by Parrish et al. (1988) but within error of an earlier 5 km estimate by Bowman (1950).

Two-stage exhumation of the Grand Forks complex and comparison with other core complexes of the Shuswap domain

The data just discussed suggest that exhumation of the GFC occurred in two stages: the first at high temperatures (upper-amphibolite- to granulite-facies) and the second at low tem-

peratures (greenschist-facies) in response to movement on the Eocene Kettle River and Granby normal faults (Fig. 17). The high-temperature stage accommodates ~75% of overall exhu-
mation, whereas the subsequent low-temperature stage ac-
counts for the remaining 25%.

Published by NRC Research Press
This two-stage history supports the contention of a number of workers (e.g., Mulch et al. 2007; Gordon et al. 2008; Simony and Carr 2011) that exhumation of the core complexes in the Shuswap domain, of which the GFC is a member, involved a combination of both high-temperature and low-temperature structural mechanisms. On the broad scale, these mechanisms were operative during the compressional and later extensional stages of Cordilleran orogenesis, with the earlier high-temperature exhumation implying extension in upper crustal levels simultaneous with compressional processes deeper in the crust (Gervais and Brown 2011; Simony and Carr 2011).

Recognition of these two stages has the effect of significantly reducing displacement estimates on the Eocene normal faults bounding other core complexes in the region. The Columbia River fault, bounding the eastern edge of the Monashee complex, is a brittle–ductile fault with vertical displacement estimates ranging from 5–9 km (Read and Brown 1981; Lane 1984; Johnson and Brown 1996) to >17 km (Parrish et al. 1988; Parkinson 1992), based on comparing peak metamorphic assemblages across the fault. However, Norlander et al. (2002) and Hinchey et al. (2007) estimated 3–6 kbar (11–21 km) of high-T decompression for the Thor-Odin dome, implying that subsequent low-T displacement on the Columbia River fault may only have been on the order of a few kilometres. The Slocan Lake fault, bounding the eastern edge of the Valhalla complex, has a minimum of 6–12 km vertical displacement along its central section (Carr et al. 1987), based on the contrast in peak metamorphic pressures across the fault. However, Hallett and Spear (2011) estimated 2 ± 0.5 kbar (5–9 km) of high-T decompression of the Valhalla complex (somewhat less than the >3 kbar (>10 km) estimate of Marshall and Simandl 2006), limiting low-T displacement on the Slocan Lake fault. Estimates of maximum vertical displacement on the Okanagan – Eagle River fault system, bounding the western edge of the southern Shuswap complex, vary considerably. Most estimates range from 8 km to ~20 km (Tempelman-Kluit and Parkinson 1986; Bordoux 1993; Johnson and Brown 1996; Brown 2010), although Erdmer et al. (1999) and Glombick et al. (2006) estimated much smaller (<3 km) displacements based on continuity of map units across the fault. More recent work on the Okanagan dome suggests 3–4 kbar of high-T, >650 °C decompression (Krukenberg and Whitney 2011), reducing the amount of exhumation due to low-temperature movement on the Okanagan – Eagle River fault system. In conclusion, the two-stage exhumation process for the Grand Forks complex and the 3–5 km of low-temperature displacement on its bounding Eocene normal faults seem to be typical for the Shuswap complex.

Acknowledgements

J.F. Cubley thanks D. Moynihan and B. Hamilton for their suggestions and input throughout this study. J. Baldwin and an anonymous reviewer are thanked for their constructive reviews of the manuscript. S. Wing, C. Bettigole, and G. Quade provided effective field assistance. Research was funded by a contract to J.F. Cubley and D.R.M. Pattison from Targeted Geoscience Initiative 3, Cordilleran Project TG6005, and NSERC Discovery Grant No. 037233 to D.R.M. Pattison.

References


Published by NRC Research Press


Published by NRC Research Press


Simony, P.S., and Carr, S.D. 1997. Large lateral ramps in the Eocene Valky school zone: extensional ductile faulting controlled by plu-


**Appendix A**

The appendix table appears on the following page.
Table A1. Bulk-rock X-ray fluorescence (XRF) chemical analyses for lithologies from the Grand Forks complex (GFC) and hanging wall of the Kettle River fault. Corresponding sample locations indicated in Fig. 5.

<table>
<thead>
<tr>
<th>Sample oxide</th>
<th>CL066</th>
<th>CL070</th>
<th>CL407</th>
<th>CL596</th>
<th>CL831</th>
<th>CL1003</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>61.62</td>
<td>60.25</td>
<td>53.94</td>
<td>49.02</td>
<td>62.28</td>
<td>62.00</td>
</tr>
<tr>
<td>TiO₂</td>
<td>1.12</td>
<td>1.11</td>
<td>1.03</td>
<td>2.17</td>
<td>0.98</td>
<td>0.78</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>16.66</td>
<td>15.74</td>
<td>15.95</td>
<td>13.31</td>
<td>18.96</td>
<td>17.00</td>
</tr>
<tr>
<td>FeO</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>5.95</td>
<td>3.72</td>
</tr>
<tr>
<td>Fe₂O₃*</td>
<td>9.41</td>
<td>10.06</td>
<td>8.16</td>
<td>16.34</td>
<td>1.58</td>
<td>3.21</td>
</tr>
<tr>
<td>MnO</td>
<td>0.14</td>
<td>0.20</td>
<td>0.14</td>
<td>0.25</td>
<td>0.24</td>
<td>0.06</td>
</tr>
<tr>
<td>MgO</td>
<td>2.61</td>
<td>3.28</td>
<td>4.61</td>
<td>6.05</td>
<td>2.45</td>
<td>2.68</td>
</tr>
<tr>
<td>CaO</td>
<td>1.83</td>
<td>3.21</td>
<td>6.44</td>
<td>10.72</td>
<td>0.19</td>
<td>1.99</td>
</tr>
<tr>
<td>Na₂O</td>
<td>1.45</td>
<td>2.50</td>
<td>3.20</td>
<td>1.03</td>
<td>0.59</td>
<td>1.99</td>
</tr>
<tr>
<td>K₂O</td>
<td>3.54</td>
<td>1.43</td>
<td>2.49</td>
<td>0.61</td>
<td>4.26</td>
<td>2.17</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.13</td>
<td>0.11</td>
<td>0.51</td>
<td>0.22</td>
<td>0.05</td>
<td>0.16</td>
</tr>
<tr>
<td>BaO</td>
<td>1151</td>
<td>612.6</td>
<td>2025</td>
<td>51.5</td>
<td>676</td>
<td>1945</td>
</tr>
<tr>
<td>Ce</td>
<td>31</td>
<td>21</td>
<td>63</td>
<td>43</td>
<td>110</td>
<td>24</td>
</tr>
<tr>
<td>Co</td>
<td>17</td>
<td>49</td>
<td>23</td>
<td>75</td>
<td>21</td>
<td>&lt;d/L</td>
</tr>
<tr>
<td>Cr</td>
<td>111</td>
<td>82</td>
<td>212</td>
<td>103.6</td>
<td>194</td>
<td>180</td>
</tr>
<tr>
<td>Cu</td>
<td>99</td>
<td>37</td>
<td>28</td>
<td>75</td>
<td>75</td>
<td>75</td>
</tr>
<tr>
<td>Ni</td>
<td>32</td>
<td>40</td>
<td>42</td>
<td>66</td>
<td>39</td>
<td>17</td>
</tr>
<tr>
<td>Sc</td>
<td>22</td>
<td>30</td>
<td>&lt;d/L</td>
<td>39</td>
<td>28</td>
<td>28</td>
</tr>
<tr>
<td>V</td>
<td>244</td>
<td>213.5</td>
<td>154</td>
<td>387</td>
<td>142</td>
<td>298</td>
</tr>
<tr>
<td>Zn</td>
<td>79</td>
<td>82</td>
<td>72</td>
<td>87</td>
<td>67</td>
<td>131</td>
</tr>
<tr>
<td>SO₃</td>
<td>0.99</td>
<td>—</td>
<td>0.55</td>
<td>—</td>
<td>0.10</td>
<td>0.42</td>
</tr>
<tr>
<td>CO₂</td>
<td>0.03</td>
<td>—</td>
<td>1.19</td>
<td>—</td>
<td>&lt;d/L</td>
<td>2.47</td>
</tr>
<tr>
<td>LOI</td>
<td>1.57</td>
<td>1.84</td>
<td>3.09</td>
<td>0.27</td>
<td>1.99</td>
<td>3.84</td>
</tr>
<tr>
<td>Total</td>
<td>100.26</td>
<td>99.84</td>
<td>99.82</td>
<td>100.10</td>
<td>100.31</td>
<td>100.29</td>
</tr>
</tbody>
</table>

Note: LOI, loss on ignition.

*Where FeO/Fe₂O₃ analyses were not done, all Fe reported as Fe₂O₃.