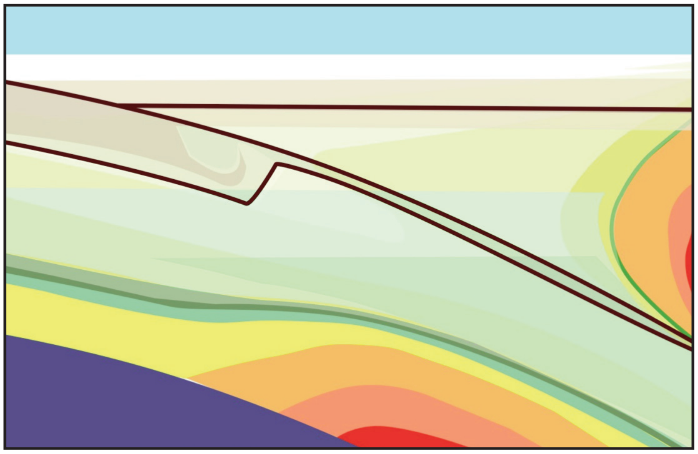
ANDREW HYNES SERIES: TECTONIC PROCESSES



Deconstructing the Infrastructure: A Complex History of Diachronous Metamorphism and Progressive Deformation during the Late Cretaceous to Eocene in the Thor-Odin–Pinnacles Area of Southeastern British Columbia

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SUMMARY

The Thor-Odin dome is a basement-cored tectonothermal culmination in southern British Columbia containing high-grade metamorphic rocks that were polydeformed in the Late Cretaceous to Eocene. The rocks south of the Thor-Odin dome that extend ca. 20 km to the Pinnacles culmination and Whatshan batholith comprise a heterogeneous tract of polydeformed medium- to high-grade metamorphic rocks and host the South Fosthall pluton near the base of the structural section. They lie in the footwall of the Columbia River fault (CRF) zone, a moderately east-dipping, ductile-brittle, normal fault that was active after ca. 55 Ma and reactivated periodically up to 30 Ma. This tract of rocks has been interpreted as a mid-crustal zone that was exhumed and cooled during Eocene extension or, alternatively, a mid-crustal channel that was bounded at the top by the CRF and was active during the Late Cretaceous to Eocene. However, the timing of metamorphism, deformation, anatexis in basement rocks, and intrusion of leucogranite plutons reveals that there are four tectonothermal domains within the tract that each experienced metamorphism, deformation and cooling at different times. These rocks record Cretaceous metamorphism and cooling in the upper structural levels and three stages of progressive metamorphism and penetrative deformation that migrated into deeper crustal levels in the Paleocene and Eocene producing a complex structural section that was exhumed in part due to motion on the Columbia River fault zone, and in part due to NE-directed transport over a basement ramp.

RÉSUMÉ

Le dôme de Thor-Odin correspond à une culmination tectonothermique d'un noyau de socle dans le sud de la Colombie-Britannique renfermant des roches métamorphiques de haute intensité polydéformées entre le Crétacé supérieur et l'Éocène. Les roches au sud du dôme de Thor-Odin qui s'étendent sur environ 20 km jusqu'à la culmination des Pinnacles et du batholite de Whatshan sont constituées d'une bande hétérogène de roches polydéformées à faciès métamorphique d'intensité moyenne à élevée qui constitue l'encaissant du pluton de South Fosthall près de la base de la colonne structurale. Elles se trouvent dans l'éponte inférieure de la zone de faille de la rivière Columbia (CRF), une faille normale à pendage modéré vers l'est, ductile-fragile, qui a été active après 55 Ma environ et a été réactivée périodiquement jusqu'à 30 Ma. Cette bande de roches a été interprétée comme une zone de mi-croûte qui a été exhumée et a refroidi durant l'extension éocène ou alors comme un canal mi-crustal qui a été limité au sommet par la CRF, et qui a été actif de la fin du Crétacé jusqu'à l'Éocène. Toutefois, la chronologie du métamorphisme, de la déformation, de l'anatexie dans les roches du socle, et de l'intrusion de plutons de leucogranite, montre qu'il existe quatre domaines tectonothermiques pour chaque bande qui ont subit du métamorphisme, de la déformation et du refroidissement à différents moments. Ces roches exhibent un métamorphisme et un refroidissement crétacé dans les niveaux structuraux supérieurs et trois stades de métamorphisme pro-

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gressif et de déformation pénétrative qui ont migré dans les niveaux crustaux profonds au Paléocène et à l'Eocène constituant ainsi une colonne structurale complexe qui a été exhumée en partie en raison du mouvement de la zone de faille de Columbia River, et en partie en raison du transport vers le N.-E. sur une rampe de socle.

Traduit par le Traducteur

INTRODUCTION: SUPRASTRUCTURE–INFRASTRUCTURE FRAMEWORK AND THE SOUTHEASTERN CANADIAN CORDILLERA

The rocks of the southeastern Canadian Cordillera have a protracted history of deformation and metamorphism. At the present latitude of the southern Canadian Cordillera, the orogeny developed into a doubly vergent, medium-sized, warm orogenic belt (Evenchick et al. 2007; Simony and Carr 2011) in a transpressional setting during the Cretaceous to Eocene (Monger and Price 2000). Crustal shortening on the western edge of the Laurentian craton occurred in the Jurassic and continued until the Late Cretaceous to Paleocene and juxtaposed pericratonic and oceanic terranes with the Laurentian margin successions (Monger et al. 1982; Colpron et al. 1996). By the Middle Jurassic some of the accreted terranes, e.g. the Slide Mountain ocean basin which closed before the Late Permian (Klepacki 1985) and rocks of the Quesnel terrane which include Laurentian-derived clastic sediments and pericratonic juvenile oceanic arc-derived rocks (Unterschutz et al. 2002), had been obducted over the pericratonic Kootenay terrane (Brown et al. 1986; Ross et al. 2005; Evenchick et al. 2007). The main periods of crustal thickening and shortening of supracrustal rocks in the eastern retrowedge side of the orogen occurred between ca. 100 Ma and 52 Ma (Simony and Carr 2011 and references therein). The collision of the Alexander terrane and Wrangellia with the Cordilleran margin occurred during the mid-Cretaceous to Tertiary (Monger et al. 1982; Monger and Journeay 1994).

Late Cretaceous to Eocene tectonothermal events overprint older structures formed during Paleozoic basin inversion, Late Triassic to Jurassic accretion of inboard terranes (Monger et al. 1982; Colpron et al. 1996), and/or structures developed during Jurassic to Cretaceous shortening and deformation, all formed in a transpressional setting (Evenchick et al. 2007; Simony and Carr 2011 and references therein). At ca. 52 Ma, the plate tectonic setting changed from a fully transpressional setting to one that included a component of local extension due to more oblique convergence of the Kula and Laurentian plates (Lonsdale 1988; Andronicos et al. 2003). This led to a structural regime dominated by strike-slip faults in the western and northern Cordillera and, in rocks of present day southeastern British Columbia, regional extension. This extension was characterized by magmatism, north-south striking, shallow to moderately dipping ductile–brittle extensional detachments with fault traces between 100 and 200 km long (Fig. 1); steep brittle north-south striking fault systems; and tectonic exhumation of rocks that were in the mid-crustal infrastructure in the Eocene (Parrish et al. 1988; Carr 1991a; Johnson and Brown 1996; Adams et al. 2005; Johnson 2006; Kruse and Williams 2007). At the latitude of the Thor-Odin dome (Fig. 1), the east-dipping Columbia River fault zone and the west-dipping Okanagan Valley–Eagle River fault zones are important Eocene, ductile–brittle extensional fault systems (Parrish et al. 1988; Johnson and Brown 1996; Teyssier et al. 2005; Johnson 2006 and references therein; Thompson et al. 2006; Brown et al. 2012).

The Thor-Odin–Pinnacles area (Fig. 2a) is located in the Western Internal zone of the southeastern Canadian Cordillera (Fig. 1) which contains predominantly metamorphic and igneous rocks, and includes several tectonothermal culminations which expose high grade metamorphic and plutonic rocks. Some of these culminations contain exposures of Laurentian basement rocks, for example the Priest River complex (Doughty et al. 1998), the Malton gneiss complex (Murphy 1987), the Frenchman Cap dome (Armstrong et al. 1991; Parkinson 1991) and the Thor-Odin dome (Reesor and Moore 1971). In others, for example, the Kettle dome/Grand Forks complex, the Okanogan dome (Parrish et al. 1988; Kruckenberg et al. 2008), and the Valhalla complex (Carr et al. 1987; Hallett and Spear 2011), the basement rocks are not exposed. All these domes have been described as metamorphic core complexes, or part of metamorphic core complexes (Ewing 1980; Armstrong 1982; Coney and Harms 1984), characterized by domal culminations of deeply exhumed, high-grade metamorphic rocks bounded by shallowly outward-dipping normal faults with low-grade metamorphic rocks in the hanging walls (Whitney et al. 2013); in detail, they may have different tectonometamorphic histories (Gervais and Brown 2011; Simony and Carr 2011).

Mechanisms that have been investigated to explain the evolution and geometry of tectonothermal culminations in the Cordillera can be discussed in terms of three main categories. These are i) channel flow models (Glombick 2005; Teyssier et al. 2005; Brown and Gibson 2006; Lemieux 2006; Kuiper et al. 2006; Williams et al. 2006; Gervais and Brown 2011; Rey et al. 2011); ii) thrust sheet models (Brown et al. 1986; Price 1994; McNicoll and Brown 1995), some with a component of interdeformation within sheets (Carr and Simony 2006; Hallett and Spear 2011); and iii) diapirism (Vanderhaeghe et al. 1999, 2003; Norlander et al. 2002; Teyssier et al. 2005; Gordon et al. 2008; Rey et al. 2009, 2011). The relative importance of exhumation along Eocene extensional fault systems (Parrish et al. 1988; Thompson et al. 2004, 2006; Brown 2010) is also under debate. Each culmination must be evaluated individually and this study aims to clarify the Paleocene–Eocene evolution of the Thor-Odin dome with reference to some of the main mechanisms that have been proposed for dome evolution.

In this study, we discuss the Thor-Odin–Pinnacles area in terms of an orogenic infrastructure–suprastructure framework. The terminology is that of Murphy (1987), Carr and Simony (2006), Culshaw et al. (2006) and Williams et al. (2006), which describes the infrastructure at lower structural levels as composed of high-grade, penetratively deformed metamorphic rocks with transposed structures and isoclinal regional folds. The upper structural levels, or suprastructure, contain rocks of lower metamorphic grades with variably oriented upright open folds, and thrust faults. From north to south, the Monashee complex consisting of the Frenchman Cap and Thor-Odin domes (Fig. 1), and the Thor-Odin to Pinnacles area represents a structural section including Laurentian basement rocks deep in the Frenchman Cap dome (and possibly the Thor-Odin dome) beneath the Cordilleran orogenic base

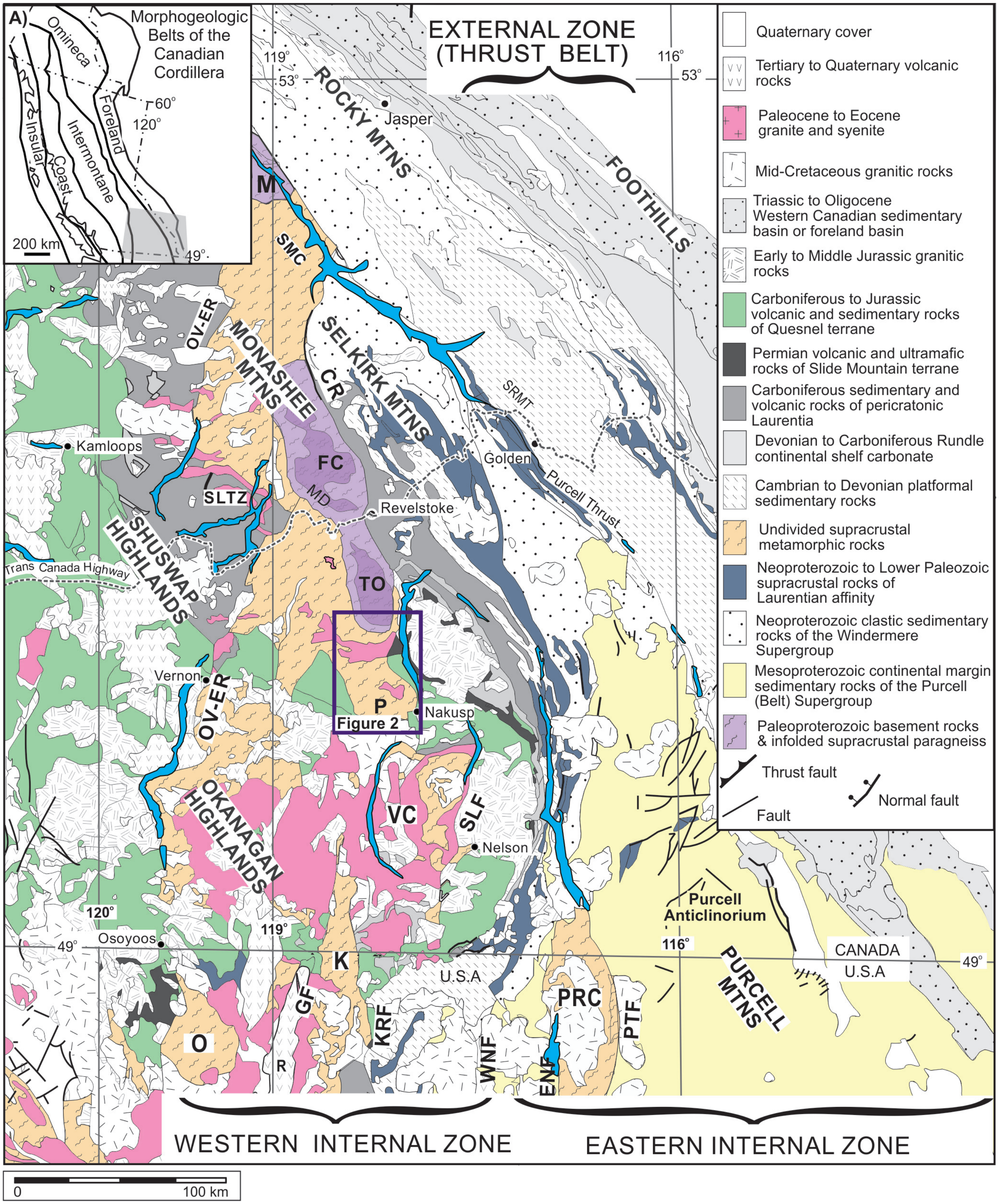

Figure 1. Geological map of the southeastern Canadian Cordillera showing the External, Western Internal, and Eastern Internal zones, (modified after Wheeler and McFeely 1991; Carr 1991b; Johnson and Brown 1996; Simony and Carr 2011). The shaded area of the inset locates the map within the morphogeological belts of the Canadian Cordillera. In the Eastern Internal zone: PRC = Priest River complex, the bounding Western (WNF) and Eastern Newport faults (ENF), and the PTF = Purcell Trench fault. In the Western Internal zone: K – Kettle dome–Grand Forks complex, O – Okanogan dome, R – Republic Graben, VC – Valhalla complex; and complexes with basement rocks include the Frenchman Cap (FC) dome; Malton complex (M); and Thor-Odin dome (TO). Eocene normal fault systems that bound high-grade rocks in the Western Internal zone include the Okanagan Valley–Eagle River fault system (OV–ER); Columbia River fault (CR), SLTZ – Shuswap Lake Transfer Zone and Slocan Lake-Champion Lake fault systems (SLF), the Granby fault (GF) and the Kettle River fault (KRF). SMC – Selkirk-Monashee-Cariboo metamorphic complex; SRMT – southern Rocky Mountain Trench.

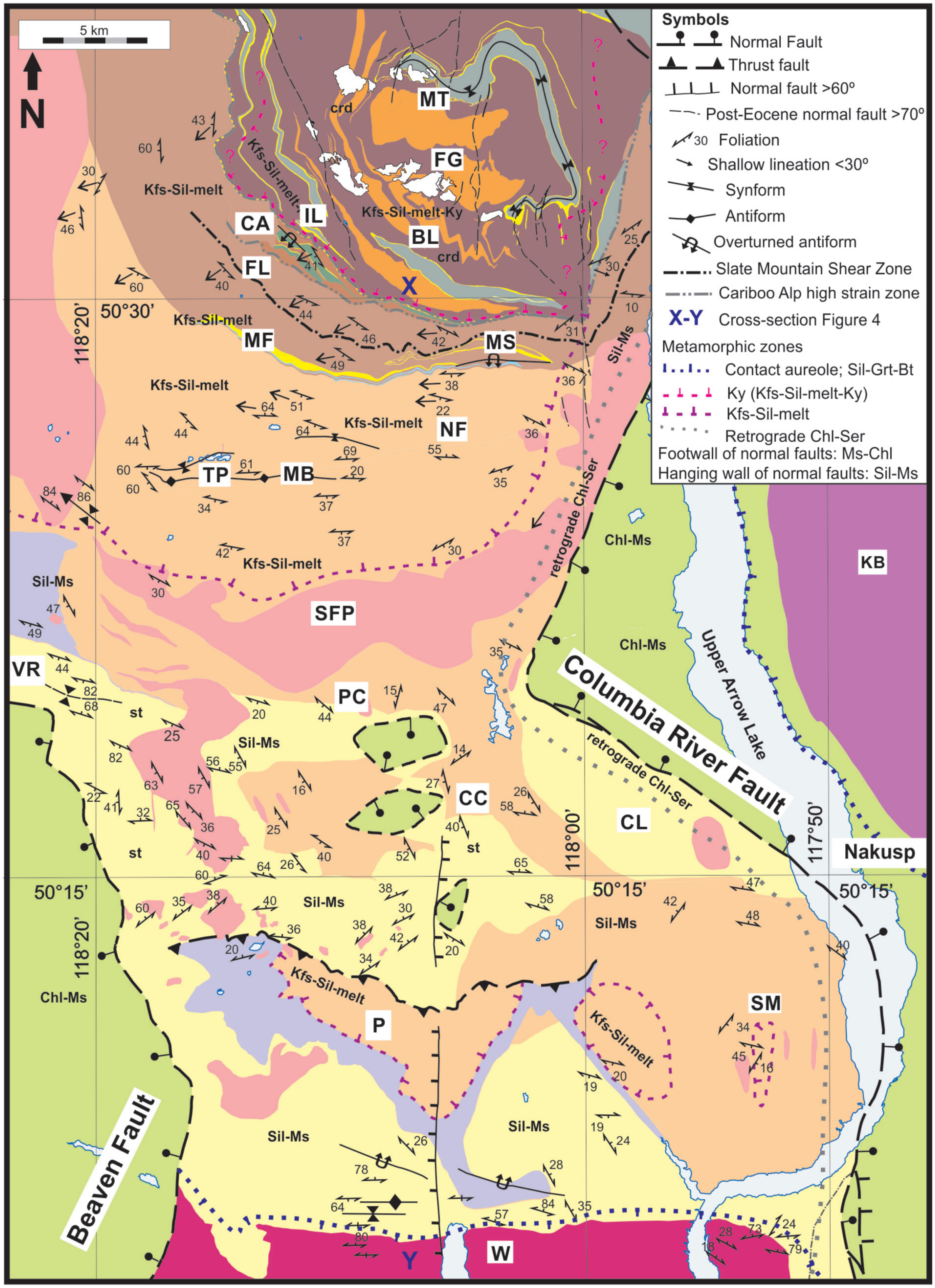

Figure 2a. Simplified geological map of the Thor-Odin–Pinnacles study area. This map shows metamorphic zones, major study locations and lithology. Structural information focuses solely on the youngest generations of foliation in each area, major folds, and fault zones; bedding and overprinted foliations are omitted for clarity. It is important to note that the structures and metamorphic data shown here do not represent coeval events, for further discussion see text and Figure 3 for details. The legend for this map outlining the lithology of major tectonostratigraphic units is included in Figure 2b, along with illustrations of the dominant fold styles in different areas. X and Y indicates the position of the cross-section in Figure 4. (Modified after Reesor and Moore 1971; Coleman 1990; Carr 1990, 1991b; Kruse et al. 2004; Thompson et al. 2004; Hinchey 2005.) Mineral abbreviations after Kretz (1983).

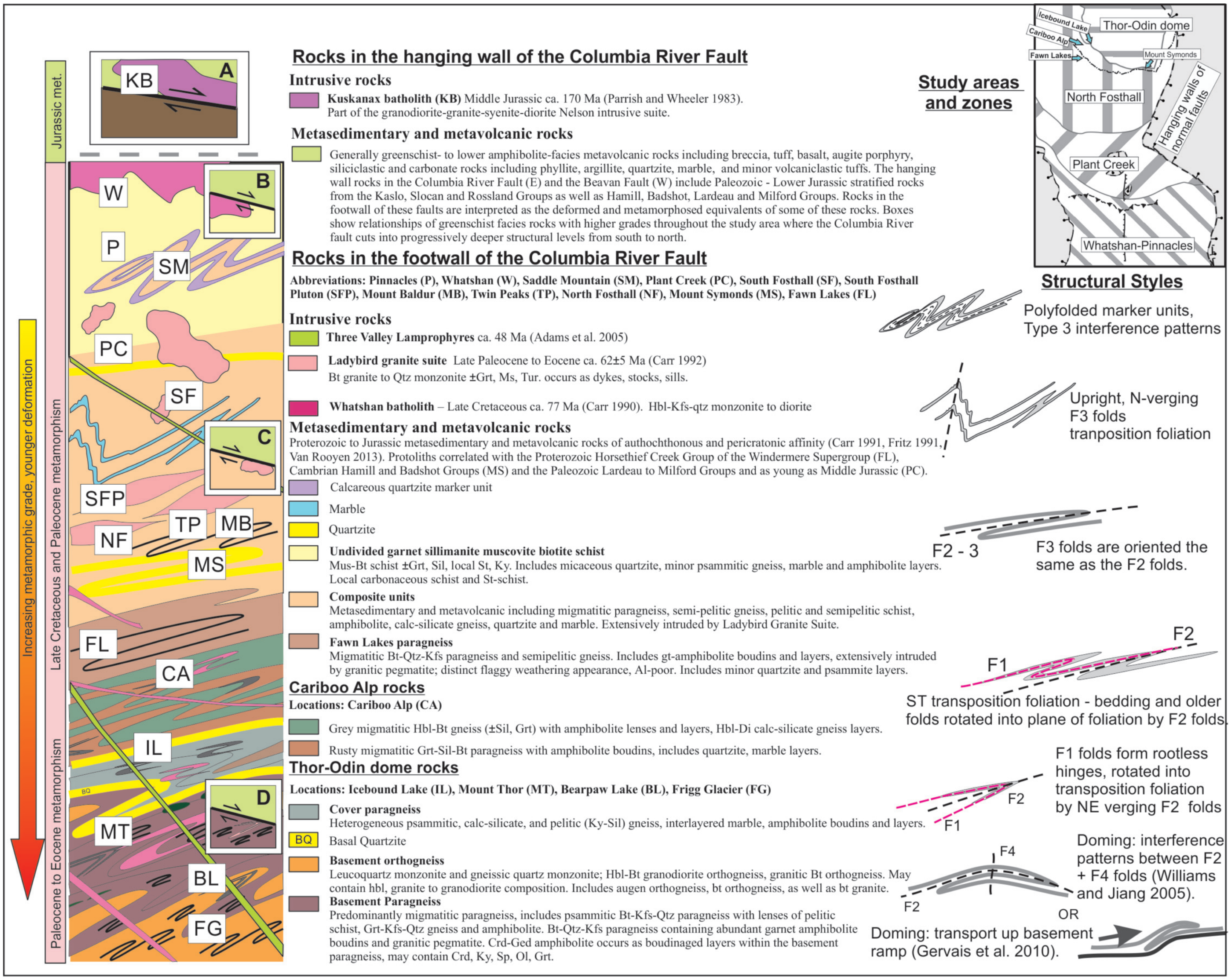

Figure 2b. Simplified lithostratigraphic column summarizing the tectonic elements, structural styles and major lithologic units, and projected locations of study areas. The figure inset contains a sketch map of the study area, where the extent and locations of the different zones discussed in the text are given for easy geographical references. It also serves as the legend for Figures 2a, 3, and 4. Specific areas discussed in the text are Bearpaw Lake (BL), Cariboo Alp (CA), Fawn Lakes (FL), and Frigg Glacier (FG), Icebound Lake (IL), Mount Baldur (MB), Mount Symonds (MS), Mount Thor MT), North Fosthall (NF), Plant Creek (PC), Pinnacles (P), Saddle Mountain (SM), South Fosthall (SF), South Fosthall pluton (SFP), Twin Peaks (TP), and Whatshan batholith (W). Data from Read and Wheeler 1976; Archibald et al. 1983; Parrish and Wheeler 1983; Parrish and Armstrong 1987; Parrish et al. 1988; Coleman 1990; Carr 1990, 1991a, b, 1992, 1995; Smith and Gehrels 1992; Colpron and Price 1995; Vanderhaeghe et al. 1999; Johnston et al. 2000; Norlander et al. 2002; Kuiper 2003; Gibson et al. 2004, 2005; Adams et al. 2005; Glombick 2005; Lemieux 2006; Hinchey et al. 2006, 2007; this study.

(Crowley 1999; Crowley et al. 2008; Gervais et al. 2010). The Laurentian basement rocks are tectonically overlain by a heterogeneous package of mid-crustal rocks in which the age of deformation youngs downward towards the basement (Parrish 1995; Glombick 2005; Williams and Jiang 2005; Carr and Simony 2006; Hinchey et al. 2006; Gibson et al. 2008; Gordon et al. 2008; Gervais et al. 2010). Internal infrastructural transitions of different ages have been documented within this mid-crustal package of rocks (Murphy 1987; Glombick 2005; Hinchey et al. 2006; Carr and Simony 2006). The hanging walls of the Eocene extensional faults contain Paleozoic to Early Jurassic stratigraphic units, and these rocks preserve the Middle Jurassic to Early Cretaceous deformational and metamorphic histories acquired when they were in the suprastructure of the orogen (Evenchick et al. 2007 and references therein; Simony and Carr 2011 and references therein).

This study integrates published and new geochronological, metamorphic and thermochronological datasets to describe and construct a history of the region from the southern flank of the Thor-Odin dome to Pinnacles. The data show that there are tectonothermal belts that preserve different Cretaceous to Eocene deformation and metamorphic histories. An important contribution of this study is to extend the definition and description of infrastructure-suprastructure associations and penetrative deformation active in the infrastructure in the study area to include a temporal dimension, recognizing that these divisions are dynamic and the boundary between them migrates with time.

REGIONAL SETTING OF THE THOR-ODIN–PINNACLES AREA OF SOUTHEASTERN BRITISH COLUMBIA

Located southwest of Revelstoke and west of the Columbia valley, the study area is approximately 50 x 20 km in size and includes the southern part of the Thor-Odin dome and rocks that lie to the south of it in the Thor-Odin–Pinnacles area (Fig. 2). The rocks are generally southwest to south dipping in a panel that has a roughly 12–15 km structural thickness, including the top 2–3 km of basement in the Thor-Odin dome. The deepest structural levels are exposed in the northern part of the study area, in the core of the high-grade metamorphic, migmatitic, Paleoproterozoic basement-cored Thor-Odin dome (Armstrong et al. 1991; Parkinson 1991). They are structurally overlain by Laurentian-derived metasedimentary rocks which are in turn overlain by metamorphosed rocks of the pericratonic Kootenay and Quesnel terranes (Monger and Price 2000). A number of volumetrically significant igneous units of Mesozoic to Eocene age occur in the study area. Some of the important igneous rocks include the Jurassic Nelson suite (Parrish and Armstrong 1987), the Cretaceous Whatshan batholith (Carr 1992), the Paleocene to Eocene Ladybird granite (Carr 1992; Hinchey et al. 2006) and the Eocene Three Valley lamprophyre suite (Adams et al. 2005). The Jurassic to Eocene intrusive rocks mark major periods of igneous activity in the Canadian Cordillera (Parrish et al. 1988; Gabrielse et al. 1991; Parrish 1995) and, taken with field relationships, they are important as strain markers and in providing geochronological constraints on timing of deformation.

One very important tectonic element in the study area is the Ladybird granite, a peraluminous anatectic granite suite derived from melting of basement rocks with Laurentian affinities (Carr 1990; Carr 1991a; Hinchey and Carr 2006). It makes up the largest igneous body in the study area, the South Fosthall pluton (Fig. 2) and is present throughout the study area (Carr 1991a, b). The Ladybird granite is a useful strain marker because it consists of pre-, syn-, and post-tectonic intrusions with well-constrained U–Pb zircon crystallization ages ranging between ca. 64 and 52 Ma (Parrish et al. 1988; Carr 1991a, 1992), discussed in more detail in subsequent sections.

The ca. 200 km long, generally north-south striking Columbia River fault zone (CRFZ) (Fig. 1) is a shallow to moderately east-dipping ductile–brittle extensional fault system bounding the eastern margin of the Frenchman Cap, Thor-Odin and Pinnacles culminations and intervening metamorphic rocks (Parrish et al. 1988; Carr 1990: Lemieux et al. 2003, 2004). We restrict our discussion to the southern part of the fault zone adjacent to the area between the southern flank of the Thor-Odin dome and northern margin of Whatshan batholith, termed the Thor-Odin–Pinnacles area (Fig. 2). The CRFZ shows significant displacements of 10–30 km and is interpreted as a crustal scale detachment (Parrish et al. 1988). This interpretation is supported by data from seismic reflection profiles (Cook et al. 1992; Varsek and Cook 1994; Cook 1995;), geological mapping (Read and Brown 1981; Brown and Read 1983; Johnson and Brown 1996; Johnson 2006) and thermochronology studies (Archibald et al. 1983; Vanderhaeghe et al. 2003; Van Rooyen 2013; Van Rooyen and Carr in press)

The Columbia River fault zone was active between ca. 58 and 55 Ma on the basis of strain recorded within mylonitic Paleocene to Eocene Ladybird granite and pegmatites in the lower plate of the fault zone where deformed Ladybird granite represents syntectonic intrusions (Parrish et al. 1988; Carr 1991a, 1992). After ca. 55 Ma the Columbia River fault zone experienced a major episode of motion, resulting in widespread cooling in the footwall rocks (Van Rooyen 2013; Van Rooyen and Carr in press). On the basis of field and thermochronological data, the fault zone is interpreted to have been reactivated in the brittle field and overprinted by steep brittle faults periodically until ca. 30 Ma (Lorencak et al. 2001). There is a sharp contrast in metamorphic ages and cooling histories between hanging wall rocks which generally preserve greenschist- to amphibolite-facies Jurassic metamorphism (discussed in the following paragraph) and the footwall rocks in the study area which show Cretaceous to Eocene amphibolite- to granulite-facies metamorphism (discussed under individual study areas below).

Jurassic to Early Cretaceous metamorphism is recorded in the Selkirk Mountains, Kootenay Arc, Purcell anticlinorium and in the Cariboo and Monashee mountains (Gibson et al. 2004, 2005). Specifically, the hanging wall of the CRF to the east of the Thor-Odin–Pinnacles area contains greenschist- to lower amphibolite-facies metasedimentary and metavolcanic rocks that experienced peak metamorphic pressures and temperatures of 5–8 kbar and 500–600°C in the Jurassic at 187–160 Ma (Read and Wheeler 1976; Archibald et al. 1983; Parrish and Wheeler 1983; Parrish and Armstrong 1987; Smith and Gehrels 1992; Colpron et al. 1996). These rocks are Paleozoic–Lower Jurassic stratified rocks from the Kaslo, Slocan and Rossland Groups as well as Hamill, Badshot, Lardeau and Milford groups (Read and Wheeler 1976; Thompson et al. 2006). These rocks were deformed and metamorphosed and were intruded by the extensive Middle Jurassic Nelson granodiorite suite (depths of 3–4 kbar), where sillimanite is present in the contact aureoles of plutons in the Thor-Odin–Pinnacles map area (Parrish and Armstrong 1987; Carr 1991b). The intrusions are syn- to post-tectonic, and crosscut Middle Jurassic (and older) folds and foliations (Parrish and Armstrong 1987; Carr 1991b, 1992). These rocks were exhumed and cooled in the Late Jurassic – Early Cretaceous (cf. Evenchick et al. 2007 and references therein) and during the Late Cretaceous–Eocene, this upper crustal domain formed the suprastructure relative to structurally deeper rocks of the Late Cretaceous–Eocene infrastructure. Additional juxtapositions of Jurassic-age greenschist-facies rocks with Late Cretaceous amphibolite-facies rocks occur as klippen of the CRF in the central part of the study area (Brown and Read 1983; Parrish et al. 1988; Carr 1991b, 1992; Lemieux et al. 2003, 2004; Van Rooyen 2013) and in the southwestern part of the area across the Beaven fault, a moderately (average 45 degrees) west-dipping, predominantly brittle (at the present exposure level), normal fault which is part of the Eocene, west-dipping Okanagan Valley–Eagle River fault system (Tempelman-Kluit and Parkinson 1986; Parrish et al. 1988; Carr 1990; Johnson and Brown 1996; Johnson 2006 and references therein; Thompson et al. 2006; Brown 2010).

MAP SCALE GEOLOGY OF THE STUDY AREA FROM THE SOUTHERN THOR-ODIN DOME AND THE WHATSHAN BATHOLITH

The study area exposes an approximately 12–15 km thick, tilted structural section in the footwall of the CRF with the deepest exposed rocks to the north. The map in Figure 2a includes metamorphic mineral zones but it is important to note that these zones do not represent metamorphic events that are the same age. For example, the Sil-Kfs-melt zones (likely representing the reaction Ms + Qtz = Sil + Kfs + melt) include Sil-Kfs-melt metamorphism that is Early Cretaceous in the Whatshan–Pinnacles area (Carr 1992; Glombick 2005; Lemieux 2006), as well as Late Cretaceous to Paleocene in the South Fosthall and Twin Peaks areas (Carr 1991b, 1992; Glombick 2005; Lemieux 2006). Likewise, structures like the map-scale folds, for example, at Pinnacles and Mount Symonds, represent Early Cretaceous and Paleocene structures, respectively (Carr 1991b, 1992; Glombick 2005). The study areas are discussed in order of increasing structural depth, starting at the top of the tectonostratigraphic column (Fig. 2b) as follows: i) Whatshan–Pinnacles, ii) Plant Creek–South Fosthall, iii) North Fosthall–Mount Symonds, iv) Fawn Lakes–Cariboo Alp, and v) the Thor-Odin dome. For each area we present the important structures, lithologies, metamorphic data and geochronology data. Figure 2b illustrates the main structural styles of the rocks in the study area in a stylized tectonostratigraphic column, Figure 3 summarizes the geochronological and metamorphic data discussed in subsequent sections, and Figure 4 is a N–S cross-section from the southern flank of the Thor-Odin dome to the Whatshan batholith that includes interpretations of structures and the location of zones that were being penetratively deformed at different times.

Whatshan–Pinnacles

The Whatshan–Pinnacles area, in the southern part of the study area, makes up a roughly 3–4 km thick package of generally medium-grade Grt-Bt-Ms-bearing and Ms-Sil-bearing schist and paragneiss with minor local migmatitic Grt-Sil-Bt-Kfs-bearing rocks (Carr 1990; Thompson et al. 2004; Glombick 2005). These rocks are not generally characterized by transposition foliation because original stratigraphy is preserved and can be recognized, unlike at deeper levels where all lithologic and structural elements are transposed into a composite foliation (cf. Tobisch and Paterson 1988). The rocks of the Saddle Mountain area extend across Upper Arrow Lake in the southeast corner of the map on the basis of correlative marker units (e.g. calcareous quartzite), and are interpreted here to be at the same structural level as Pinnacles (Fig. 2). Upright, doubly vergent structures defining the main map pattern in this area with Type 3 interference patterns (Figs. 2b, 4) are older than Late Cretaceous, as well as refolded, flat-lying, nappe-style folds (Carr 1991a). The area around Pinnacles (Fig. 2a) has a mapped north-verging thrust fault which places higher grade Kfs-Sil-melt-bearing rocks over the dominant lower grade Sil-Ms-bearing rocks and is interpreted as a ductile, Late Cretaceous or Paleocene structure (Fig. 4). The Whatshan batholith has a U–Pb titanite crystallization age of 75.5 ± 1.5 Ma from a sample locale south of the Pinnacles area (Carr 1991a). This constrains the age of deformation in the host rocks to be older than ca. 75 Ma. This structural level is considered to be equivalent to the Joss Mountain area on the western flank of the Thor-Odin dome (N of the study area, outside of Fig. 2a), where deformed granitic pegmatites have been dated at ca. 73 Ma and undeformed pegmatites at ca. 70 Ma, indicating that the end of deformation is bracketed within that time period (Johnston et al. 2000).

The age of regional metamorphism in the Whatshan–Pinnacles area is constrained by U–Pb dating of monazite (Glombick 2005; Lemieux 2006) and spans the Early to Late Cretaceous, between ca. 130 and 90 Ma (the younger ages generally in structurally deeper areas), with some evidence for earlier metamorphism as old as ca. 200 Ma (Fig. 3). The foliation and pre- to syn-deformational metamorphic minerals aligned within it are cut by the Late Cretaceous Whatshan batholith, suggesting a pre-Campanian metamorphic history. In contrast, andalusite, garnet, and biotite are present in the contact aureole of the Whatshan batholith, overprinting the previously discussed regional metamorphic rocks (Glombick 2005). Kyanite and staurolite are present in the footwall rocks of the Beaven fault, but these cross-cut foliations and overprint older metamorphic assemblages (Reesor and Moore 1971; Carr 1991b). These minerals are restricted to minor aluminous lithologies in the footwall of the Beaven fault and are interpreted as mineral growth at conditions below peak metamorphism, possibly related to conditions during exhumation (Carr 1991b). Metamorphic titanite in the contact aureole of the Whatshan batholith dated at ca. 64 Ma is interpreted as a cooling age after the intrusion of the batholith (Carr 1992). The Whatshan–Pinnacles area represents rocks that are in the suprastructure of the orogen by ca. 75 Ma because the regional penetrative foliation and associated metamorphic mineral assemblages predated the low-pressure andalusite, garnet and biotite in the contact aureole of the Whatshan batholith (Carr 1992), and there is no evidence of overprinting of Paleocene or Eocene events on the regional metamorphism and deformation (Glombick 2005).

Plant Creek–South Fosthall

The central part of the study area, the Plant Creek–South Fosthall area, south of the South Fosthall pluton, comprises a structural thickness of approximately 2–3 km. The metamorphic grades of the rocks in the Plant Creek area (similar to the lower grade rocks between Whatshan and Pinnacles discussed above) are the lowest in the study area, and are mainly medium-grade Bt-Grt-Sil-Ms±St-bearing schist with mixed metasedimentary and metavolcanic protoliths (Plate 1a, 1b). The protolith ages of these rocks range from Cambrian to Devonian, with some as young as Jurassic (Van Rooyen 2013).

In general, the map pattern is controlled by tight asymmetric F3 folds with moderately to steeply dipping axial planes that fold the dominant transposition foliation, and refold coaxial F2 folds and rootless isoclinal folds (Carr 1991a). The axial planes of F3 folds north of Plant Creek dip to the south (Fig. 2b, Plate 1c), and axial planes of F3 folds in the southern part of the Plant Creek area dip to the north, exposing the lowest grade metamorphic rocks in the study area, in the structural depression (Fig. 4). This structural interpretation is supported by the geometry of crustal reflectors imaged by the Lithoprobe Project (Cook et al. 1992; Cook 1995). The cross section in Figure 4 illustrates the structural context of this area as part of a regional syncline with the lowest grade rocks in the core. Within this part of the study area are klippen of the hanging wall of the Columbia River fault, where Jurassic rocks metamorphosed at greenschist-facies conditions are in fault contact with the high-grade footwall rocks with Cretaceous to Eocene thermal and deformation histories being discussed here (Van Rooyen 2013; Van Rooyen and Carr in press).

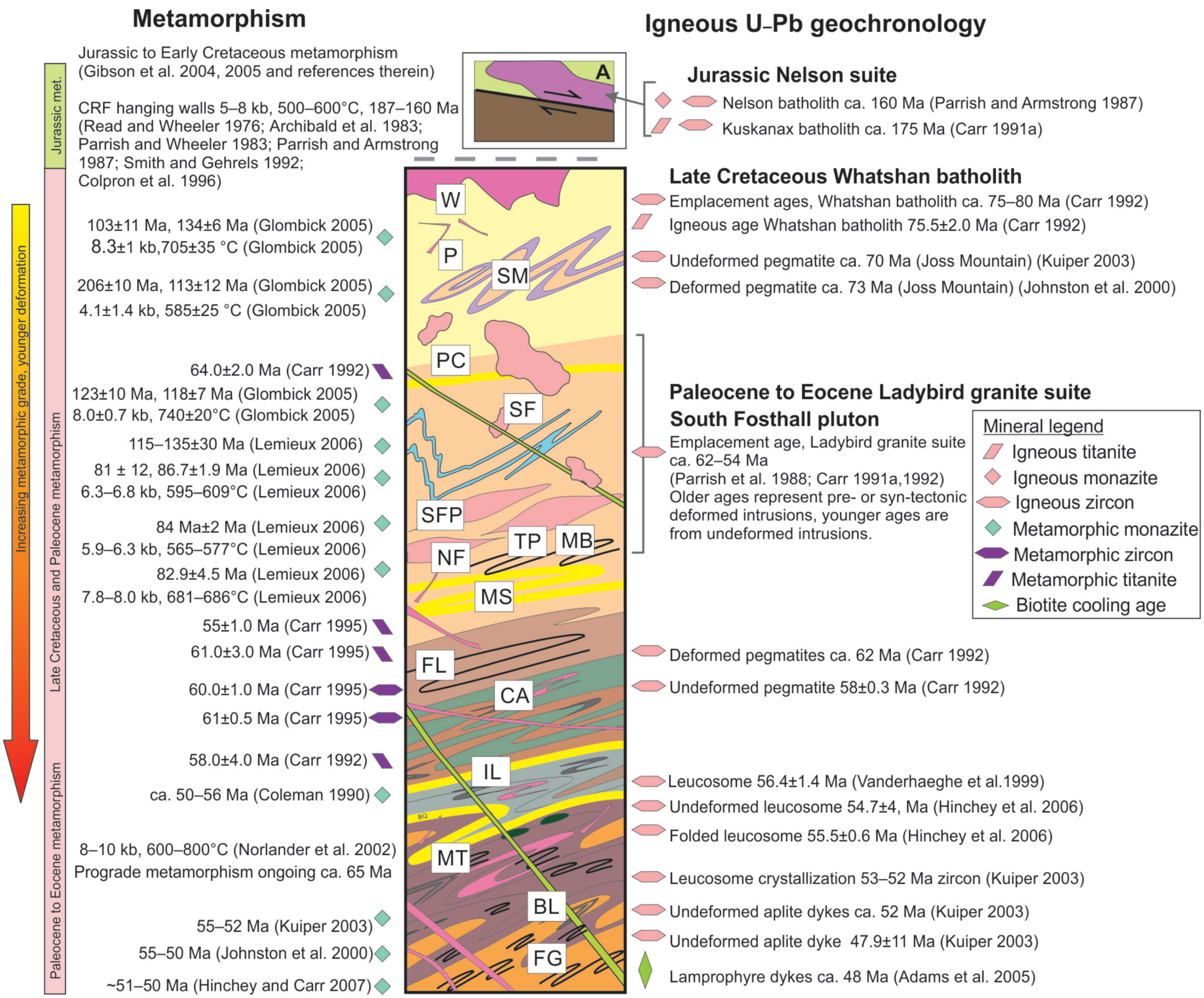

Figure 3. Tectonostratigraphic column of the study area summarizing the timing of metamorphism, pressure and temperature conditions, and U–Pb zircon ages of deformed and undeformed igneous rocks. The *P–T* determinations and dates are projected into the structural section as presented in column form, and represent the structural level at which they are listed. The *P–T* and geochronology data do not necessarily represent data from exactly the same rocks, but they are representative of the same general areas. Data from Read and Wheeler 1976; Archibald et al. 1983; Parrish and Wheeler 1983; Parrish and Armstrong 1987; Parrish et al. 1988; Coleman 1990; Carr 1990, 1991a, 1992, 1995; Smith and Gehrels 1992; Colpron et al. 1996; Vanderhaeghe et al. 1999; Johnston et al. 2000; Norlander et al. 2002; Kuiper 2003; Gibson et al. 2004; Adams et al. 2005; Glombick 2005; Lemieux 2006; Hinchey et al. 2006, 2007; this study.

Metamorphic minerals occur aligned with S2 foliations that developed together with F2 folds, which are generally refolded by F3 folds (Carr 1991b). In some cases F3 folds contain axial planar foliations. The timing of the onset of prograde metamorphism is not well constrained, but it is likely that prograde metamorphism in the Plant Creek area was ongoing during the Late Cretaceous (Fig. 3), spanning the time between ca. 100 and 80 Ma based on U–Pb monazite dating (Lemieux 2006). Folded rocks are cut by the syn- to post-deformational ca. 62–54 Ma Ladybird suite (Carr 1992). Hornblende $^{40}Ar/^{39}Ar$ cooling ages in the Plant Creek–South Fosthall area vary between ca. 62 and 58 Ma and biotite cooling ages vary between ca. 52.5 and 50.5 Ma, while muscovite and biotite cooling ages from Ladybird granite pegmatites are between 51 and 50.5 Ma (Van Rooyen 2013; Van Rooyen and Carr in press). At ca. 73–64 Ma the rocks of the Plant Creek-South Fosthall area, structurally above the South Fosthall pluton, were being penetratively deformed in the infrastructure (Fig. 4).

North Fosthall to Fawn Lakes

The North Fosthall area, from the South Fosthall pluton to the southern margin of the Thor-Odin dome, includes study areas at Twin Peaks, Mount Symonds, and Fawn Lakes (Figs. 2, 3). It makes up a structural thickness of roughly 4–5 km, 2–3 km of which is within the boundaries of the South Fosthall pluton. The age of the South Fosthall pluton is bracketed by U–Pb zircon ages between ca. 62 and 54 Ma (Carr 1992). The rocks between the South Fosthall pluton and the southern margin of the Thor-Odin dome comprise a south-southwest-dipping panel of heterogeneous, polydeformed, medium- to high-

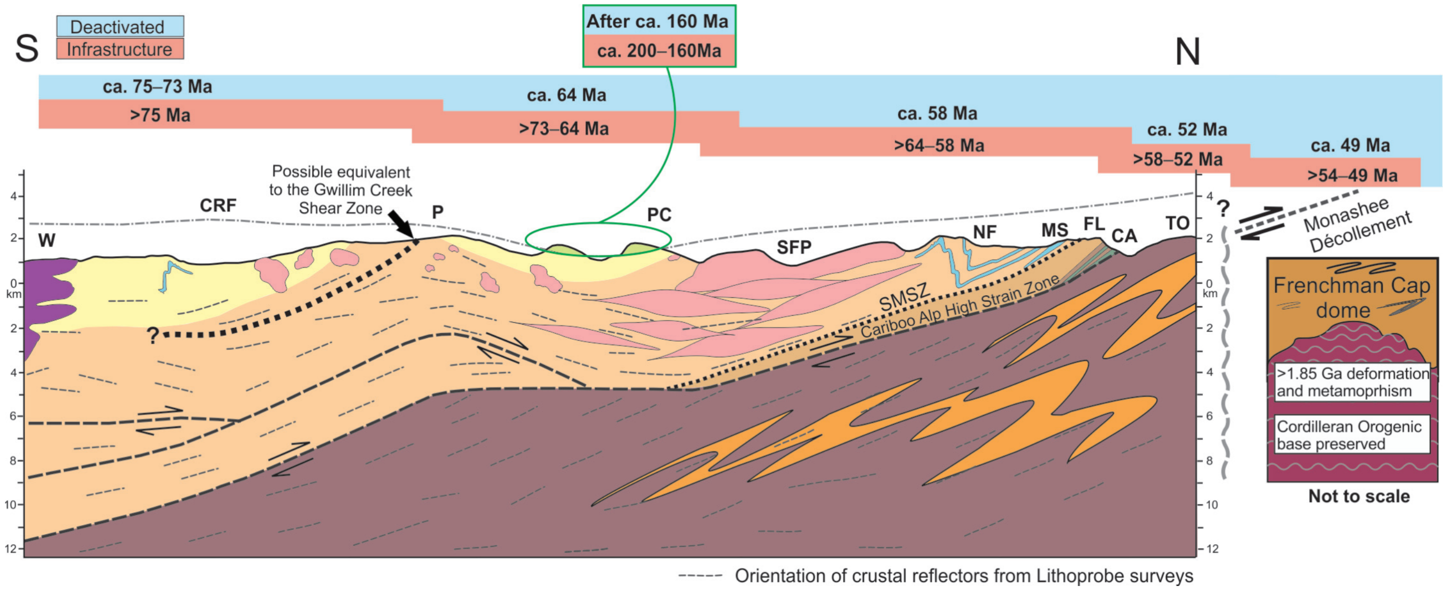

Figure 4. N–S cross-section of the study area on line X–Y in Figure 2a. This cross-section illustrates the general structural styles and orientations of units in the study area. Note that minor lithological units are omitted for clarity, and marker units (e.g. the blue calcareous units) have been given an exaggerated thickness to illustrate fold styles. The structural interpretation is supported by data from Lithoprobe seismic reflection studies (Carr 1995; Cook 1995). The shapes of the South Fosthall pluton and Whatshan batholith in the subsurface are speculative, but illustrate the interpretation discussed in the text that the intrusions are mostly sheet-like bodies within the country rocks. The proposed analogue for the Gwillim Creek shear zone in the Pinnacles area (as mapped on Figure 2) follows crustal seismic reflectors parallel to the deeper thrust slice interpreted to be located under Pinnacles. (SMSZ – Slate Mountain shear zone; CRF – Columbia River fault zone).

This cross-section also illustrates the interpretations we make based on U–Pb zircon, monazite and titanite geochronological data presented in Figure 3, in combination with $^{40}Ar/^{39}Ar$ cooling data. We suggest that there are at least four tracts of rocks with different structural, metamorphic and (in part) cooling histories in the Thor-Odin–Pinnacles area and that they represent progressively younger phases of metamorphism and deformation migrating downwards into the structural section. Also illustrated is the projected relationship with the Frenchman Cap dome which represents an even younger phase of deformation, and contains the base of Cordilleran deformation, below which only Paleoproterozoic deformation and metamorphism is evident (Gervais et al. 2010).

grade metasedimentary and metavolcanic rocks (Carr 1991a). The map pattern is created by northeast-verging km-scale tight asymmetric F2–F3 folds with moderately to steeply dipping axial planes (Reesor and Moore 1971; Carr 1991b), which fold a pervasive older transposition foliation present in all the rocks in this area. The southeastern dip of the panel is controlled by the long limbs of the F2–F3 folds (Figs. 2b, 4). Pegmatites occur as pre-, syn-, and post-deformational intrusions. Plate 1d shows an example of a syn-deformational pegmatite boudinaged along the foliation plane in the North Fosthall area.

The Twin Peaks area is located northwest of the South Fosthall pluton and consists mostly of migmatitic Sil-Grt-Bt-Kfs-bearing paragneiss with a well-developed composite transposition foliation, primarily moderately south- to southwest-dipping (Fig. 4). Ladybird granite intrusive rocks are abundant and are concordant with layering in the host rocks in some cases, and crosscut the layering in others. The regional transposition foliations in this area generally dip steeply (40–60°) towards the south and southwest, with some on the northern side of the valley dipping towards the north and northeast (Fig. 2a). The area is located on a regional scale northeast-verging F2 fold, similar in orientation and style to folds in other study areas (e.g. Mt. Symonds), but contains a fold closure with fold axis plunging to the west-southwest and axial plane dipping steeply to the south. Plate 1e shows parasitic Z folds in migmatitic Sil-Grt-Bt-Kfs-bearing paragneiss that reflect the regional fold style. These cm-scale folds have axial planes that dip steeply to the south and fold hinges that plunge shallowly to the west. Plate 1f shows migmatitic Hbl-Bt-bearing amphibolite gneiss with a well-developed transposition foliation in which the foliation is crosscut by coarse-grained granitic leucosome, suggesting that these rocks underwent progressive episodes of deformation and anatexis with syn-deformational leucosome intrusion. The timing of deformation in this area is constrained by geochronology in an adjacent area at the same structural level at Mount Baldur (Fig. 3). Mount Baldur is situated east of the Twin Peaks area, within the South Fosthall pluton (Carr 1990; Hinchey 2005) and is dominated by Ladybird granite with up to 30% screens of paragneiss, psammite, quartzite and amphibolite country rock within the granite (Hinchey and Carr 2006). Around the South Fosthall pluton the age of the transposition foliation is Late Cretaceous to Paleocene, bracketed by U–Pb zircon ages from deformed and undeformed Ladybird granite indicating that deformation was ongoing at ca. 62 Ma, and largely over by ca. 55 Ma (Carr 1992). The amphibolite-facies metamorphic assemblages are folded by the transposition foliation and the granite cuts the foliation, indicating that the peak metamorphism is pre- or syn-deformational (Carr 1991a; this study).

The Mount Symonds area (Fig. 2) comprises a mixed assemblage of migmatitic Sil-Kfs-Grt-Bt-bearing paragneiss, quartzite, amphibolite, marble and psammite, with a well-developed, moderately south-southwest-dipping composite transposition foliation. Several marker units have Proterozoic (i.e. Fawn Lakes Assemblage) to Cambrian protolith ages (i.e. Empress Marble south of Mount Symonds) which can be traced along strike for up to 30 km. The area is located on the long limb of a km-scale isoclinal fold, interpreted as F2, which formed during northeast-directed transport and compression. The axial plane of this fold dips steeply to the south, similar to

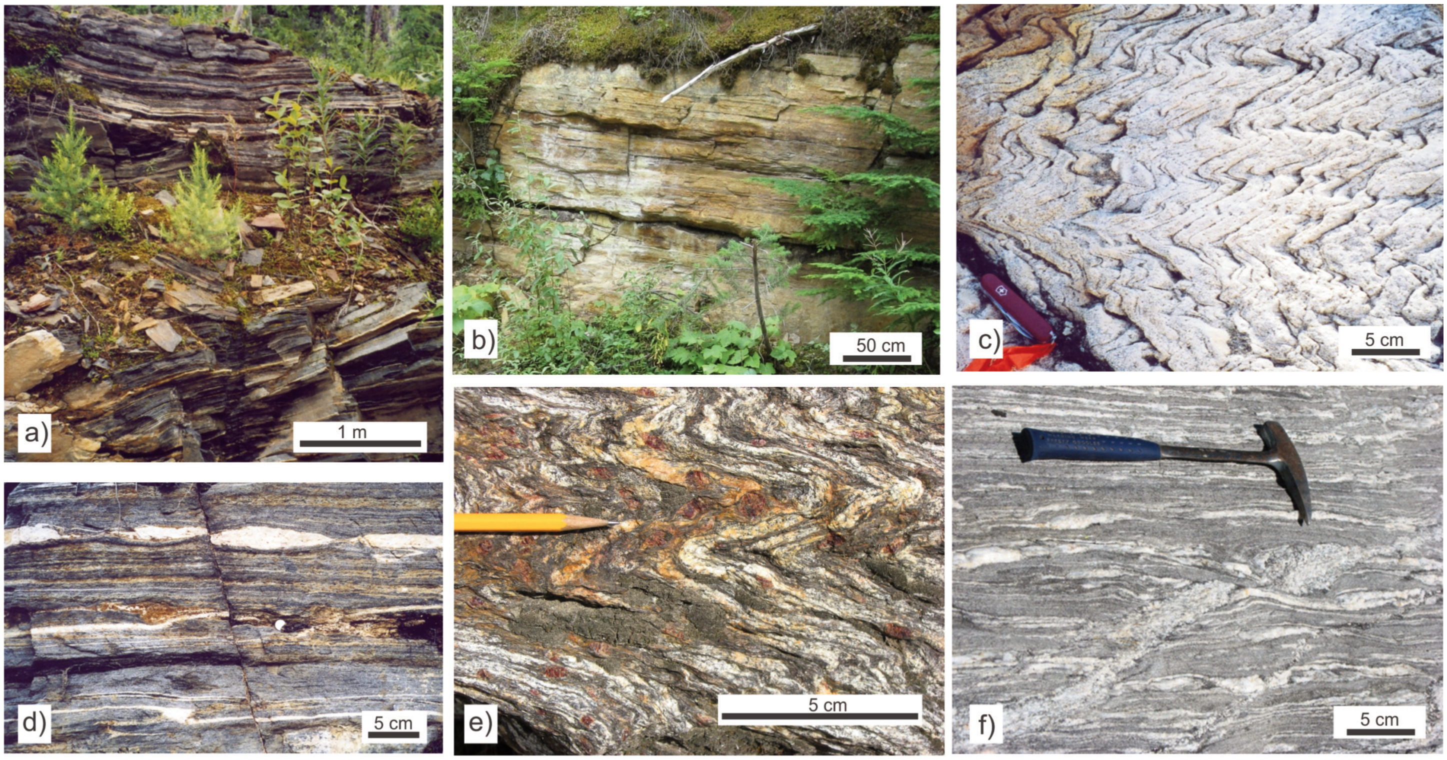

Plate 1. Illustrations of major structural styles and important geological units or lithologies throughout the structural section shown in Figures 2, 3, and 4. **(a)** Shallow south-dipping metasedimentary and calc-silicate rocks in the Plant Creek–South Fosthall area. These metamorphic rocks are typical of the mixed assemblage of upper amphibolite-facies metasedimentary and metavolcanic rocks found in the center of the study area. **(b)** Moderately south-dipping metasedimentary rocks south of the Plant Creek area showing stratigraphy parallel to foliation. **(c)** Parasitic folds in a mixed calc-silicate–marble unit in the Plant Creek area illustrating the dominant fold style in the central part of the map area. This sample is not in situ, but the outcrop folds in the same unit have axial planes dipping moderately to the south. **(d)** Boudinaged pegmatite layer in highly strained calc-silicate amphibolite in the North Fosthall area. The orientation of the boudinaged leucosome within the south-dipping foliation plane suggests elongation in an east-west direction. **(e)** Small scale folds in migmatitic Sil-Grt-Bt-bearing paragneiss at Twin Peaks illustrative of the regional fold style. These cm-scale folds have axial planes that dip steeply to the south and fold hinges that plunge shallowly to the west, similar to the regional fold closure mapped at Twin Peaks. **(f)** Migmatitic amphibolite orthogneiss at Twin Peaks with a well-developed transposition foliation. Below the hammer is a crosscutting vein of coarse-grained granitic leucosome. The leucosomes in the foliation are crosscut by this coarse-grained pegmatite vein which in turn has been deformed, illustrating the progressive nature of leucosome generation, pegmatite intrusion and deformation.

the regional km-scale folds (Fig. 2b) at Fawn Lakes and Mount Symonds. Pegmatites crosscut the transposition foliation in some instances, and are concordant with it in others indicating a combination of syn- and post-deformational intrusion of pegmatites.

The apparent age of at least some stages of tight F2 and F3 folding (Fig. 2b) and metamorphism is older than ca. 58 Ma and formed, in part, at ca. 62 Ma based on the age of metamorphic titanite (Carr 1992). The Ladybird granite west of Mount Symonds is concordant with folds in the host paragneiss and has a U–Pb zircon crystallization age for magmatic zircon of ca. 60 Ma (Vanderhaeghe et al. 1999), indicating that folding was still ongoing after that date. Metamorphism was ongoing in the Mount Symonds and North Fosthall rocks at ca. 61–60 Ma (Fig. 3) based on metamorphic ages in titanite and zircon (Carr 1995) (Fig. 3). In the Twin Peaks–Fosthall pluton area and at Mount Symonds hornblende cooling ages vary between ca. 57 and 53 Ma, and biotite cooling ages are in the 52.5–50.5 Ma range, while biotite cooling ages from Ladybird granite pegmatites are between 51 and 50.5 Ma (Van Rooyen 2013; Van Rooyen and Carr in press).

The Fawn Lakes area contains primarily migmatitic Grt-Bt-Kfs±Sil-bearing paragneiss with subordinate quartzite, amphibolite and psammite collectively mapped as the Fawn Lakes Assemblage (Carr 1991a). It is situated on the long limb of a regional scale northeast-verging F2 fold (Fig. 2b) and has a well-developed south-southwest-dipping composite transposition foliation in all units. The geometry of the panel of rocks is controlled by the enveloping surface of two phases of coaxial northeast-verging coaxial folds (Fig. 2b, Plate 1g). Metamorphism was ongoing in the Fawn Lakes rocks at ca. 61–60 Ma (Fig. 3) based on metamorphic ages in titanite and zircon (Carr 1995). At Fawn Lakes the hornblende cooling ages are similar to Plant Creek at ca. 62–58 Ma and biotite cooling ages are between 52.5 and 51 Ma (Van Rooyen 2013; Van Rooyen and Carr in press).

Cariboo Alp

The southwestern margin of the Thor-Odin dome at Cariboo Alp area is an approximately 1 km-thick structural package of highly strained rocks with a pervasive southwest-dipping transposition foliation (Plate 1h) and northeast-verging isoclinal folds (Reesor and Moore 1971; McNicoll and Brown 1995). This southwest-dipping panel of rocks is characterized by 1 x 5 km lozenges of transposed migmatitic Kfs-Sil-Grt-Bt-bearing gneiss folded and refolded by F2–F3 northeast-verging tight rootless antiforms and synforms (Fig. 2b, Plate 1i). The lozenges thin and pinch out into a high strain zone (Fig. 4) on the western, southern and southeastern flanks of the dome (McNicoll and Brown 1995). The rocks are predominantly

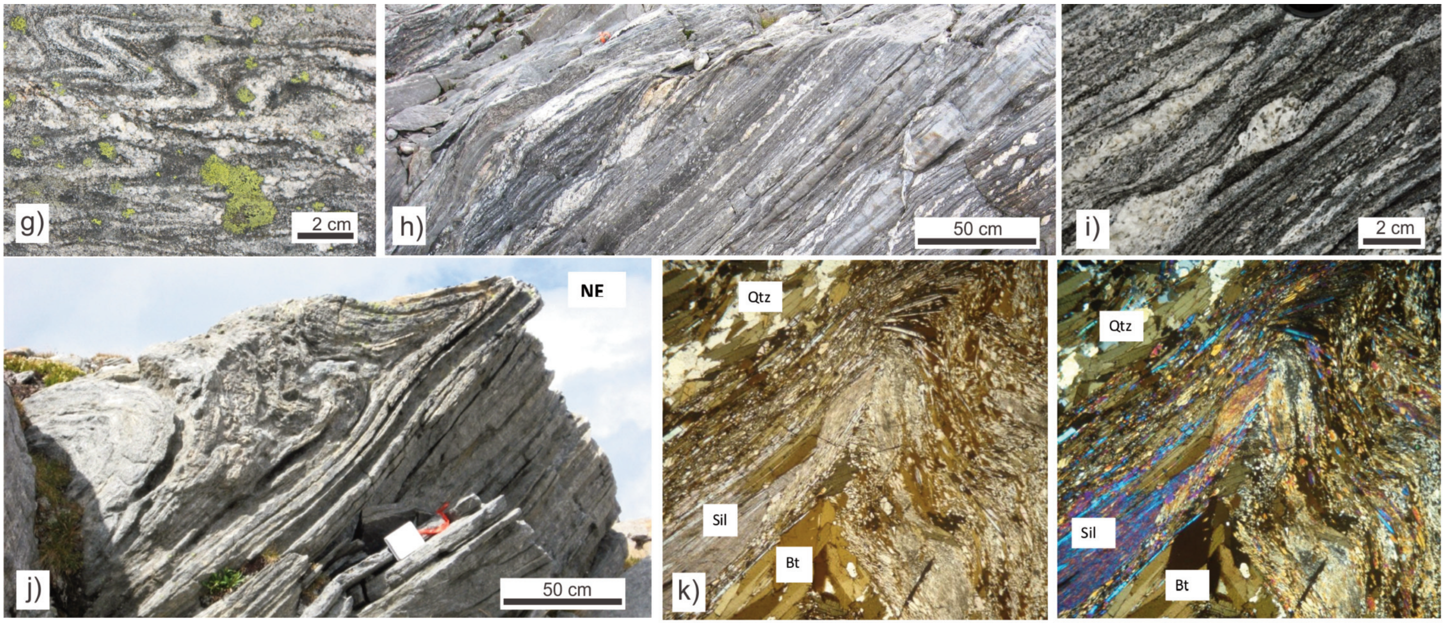

Plate 1. ***Cont'd*** **(g)** Grt-Kfs-Bt-bearing paragneiss at Fawn Lakes. The cm-scale folds show sheared out fold limbs and some coarse-grained leucosomes that are late in the transposition history. The axial plane of this fold dips steeply to the south, similar to the regional km-scale folds at Fawn Lakes and Mount Symonds. **(h)** Migmatitic Hbl-Bt-bearing orthogneiss at Cariboo Alp. Coarse-grained granitic leucosomes and boudinaged pegmatites, elongate in a northeast-southwest direction along the southwest dipping foliation planes, crosscut the transposition foliation. **(i)** Details of the grey Di-Hbl-Bt-bearing orthogneiss at Cariboo Alp. These photos show the composite transposition foliation, rootless isoclinal fold hinges, and cm-scale gneissic layering defined by biotite, diopside and hornblende alternating with quartz and plagioclase. **(j)** Fold in grey Sil-Bt-bearing quartzofeldspathic paragneiss at Cariboo Alp with axial plane parallel to the main SW-dipping transposition foliation. The fold vergence indicates a top-to-the-northeast directed sense of movement at Cariboo Alp. **(k)** Photomicrographs of Grt-Sil-Kfs-Bt-bearing paragneiss (DC184) at Cariboo Alp in plane-polarized light (left) and under crossed polars (right). The fold in this section shows a northeast-verging sense of motion, similar to the direction of motion determined from outcrop-scale folds. The section is oriented perpendicular to the northeast-southwest 'motion plane.' The folded sillimanite needles are interpreted as evidence for pre- or syn-deformational growth of the high temperature metamorphic assemblage.

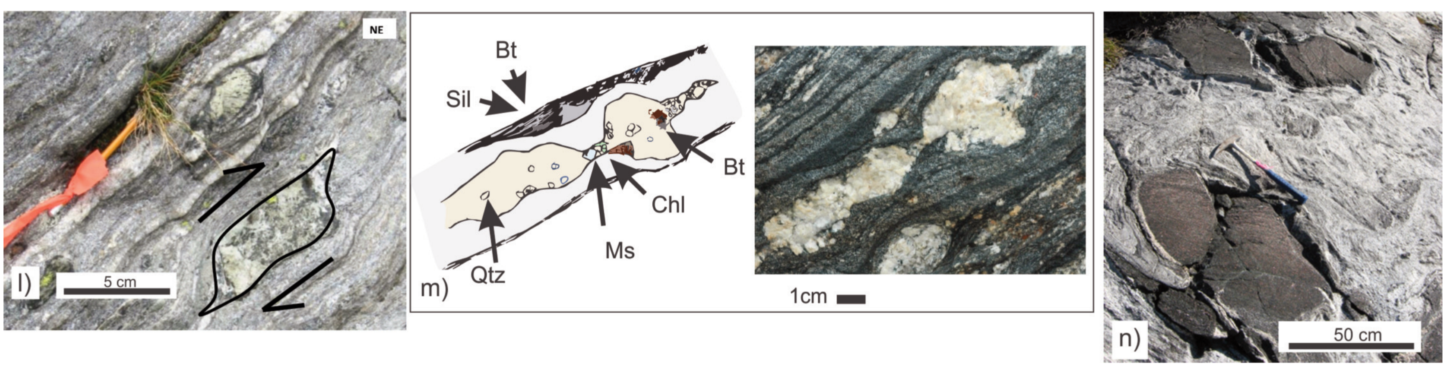

Plate 1. ***Cont'd*** **(l)** Asymmetric pressure shadows on a large diopside crystal aggregate in grey migmatitic Di-Hbl-Bt-bearing calc-silicate gneiss at Cariboo Alp. The shear sense is top-to-the-northeast, consistent with the northeast-verging folds observed at Cariboo Alp. **(m)** Muscovite crystals in leucosomes at Cariboo Alp. These muscovite (1 cm) crystals occur in pressure shadows of peritectic K-feldspar or plagioclase crystals in leucosome pods within the host migmatitic Hbl-Bt-bearing gneiss at Cariboo Alp. The muscovite is intergrown with chlorite, and interpreted as evidence for retrograde metamorphism. Both minerals are present as euhedral crystals 0.1 mm to 10 mm in diameter with sharp grain boundaries. As an alternative interpretation it is possible that the muscovite crystallized from the leucosome melt and that the chlorite overgrew it during retrograde metamorphism. This does not change the interpretation that the cooling age of the muscovite is regionally significant. **(n)** Garnet amphibolite rafts and lenses in migmatitic Grt-Bt-bearing quartzofeldspathic paragneiss at Icebound Lake.

alternating lenses and layers of grey migmatitic Sil-Grt-Kfs-Bt-bearing paragneiss and rusty migmatitic Kfs-Grt-Sil-Bt-bearing paragneiss with southwest-plunging sillimanite and hornblende mineral lineations and stretching lineations. The structural interpretation of the margin of the dome is controversial and has been interpreted as a ductile thrust-sense shear zone termed the Monashee décollement (Brown et al. 1992; McNicoll and Brown 1995), the border of a diapir (Vanderhaeghe et al. 1999, 2003; Norlander et al. 2002; Fayon et al. 2004), or as a high strain zone that formed as a result of ductile extension during normal faulting (Kruse and Williams 2007). It has also been proposed that there is no structural break at the Monashee décollement and that the highly strained rocks are part of a zone of crustal channel flow that includes both the dome and overlying rocks (Johnston et al. 2000; Williams and Jiang 2005; Kuiper et al. 2006; Williams et al. 2006).

The sense of shear interpreted from kinematic indicators and fold vergence (Fig. 2b) indicates top-to-the-northeast (Plate 1i–l) as the dominant transport direction, generally interpreted as forming in a compressional setting (McNicoll and Brown 1995). Plate 1h and 1i show details of the composite transposition foliation, such as rootless isoclinal fold hinges and cm-scale gneissic layering (Plate 1i) in the grey migmatitic Hbl-Bt-bearing orthogneiss. Primary layering, as seen in the different lithologies contained in the main paragneiss units, is concordant with the foliation. The most prominent folds at Cariboo Alp are tight to isoclinal northeast-verging F2 folds (Plate 1j and 1k). The axial planes and limbs of the F2 folds are approximately concordant with the predominant gneissic transposition foliation (Fig. 2b). The calc-silicate unit of the grey gneiss contains pressure shadows containing leucosome and hornblende around diopside (Plate 1l), indicating a top-to-the-northeast sense of movement. Shear bands are found between boudinaged pegmatite lenses and also indicate top-to-the-northeast sense of movement.

High strain deformation and the formation of the transposition fabric were coeval with metamorphism (Fig. 3) between ca. 62 and 58 Ma based on U–Pb dates in metamorphic titanite (58 ± 4 Ma) from paragneiss at Cariboo Alp (Carr 1992) and metamorphic zircon U–Pb dates (61 ± 0.5 Ma) in the Fawn Lakes Assemblage (Carr 1992). Deformation ongoing at Cariboo Alp stopped by ca. 58 Ma (Carr 1992), based on the U–Pb zircon crystallization age of an undeformed pegmatite (Fig. 3) cross-cutting the dominant transposition foliation (Carr 1992). U–Pb monazite ages from a quartzite at Cariboo Alp yielded prograde metamorphic ages of ca. 56–54 Ma with errors ranging between 1 and 4 Ma and were interpreted to represent post-deformational growth of monazite (Coleman 1990). Hornblende cooling ages at Cariboo Alp are between 55 and 53 Ma, biotite cooling ages are in the 52.5 to 50.5 Ma range, and muscovite cooling ages from leucosome within the host gneiss (illustrated on Plate 1m) and from Ladybird granite pegmatites are between 51 and 50.5 Ma (Van Rooyen 2013; Van Rooyen and Carr in press). The Cariboo Alp area represents infrastructure to the structurally higher and deactivated South Fosthall–Plant Creek zone by ca. 64 Ma. The Cariboo Alp area transitioned to being deactivated, relative to the deeper rocks within the Thor-Odin dome by ca. 58 Ma.

The Thor-Odin Dome

The Thor-Odin dome (Fig. 2) comprises a roughly 4–5 km structural thick block of polydeformed, migmatitic rocks that includes rocks with Laurentian basement affinity and supracrustal cover rocks complexly infolded with the basement rocks (Reesor and Moore 1971; Parkinson 1991, 1992; Hinchey et al. 2006 and references therein). The southern part of the Thor-Odin dome consists of migmatitic ortho- and paragneiss with a pervasive transposition foliation (Reesor and Moore 1971; Parkinson 1992; Spark 2001; Hinchey et al. 2006). The oldest structures are isoclinal folds, with limb lengths of 60–80 km involving Proterozoic Laurentian basement rocks and a cover sequence of metamorphic supracrustal rocks of uncertain age (Reesor and Moore 1971; Parkinson 1991). The large-scale domal structure is dominated by interference patterns between these early folds and three subsequent generations of folds (Fig. 2b), designated F2, F3, and F4 by Kruse et al. (2004) and Williams and Jiang (2005). The main prograde metamorphic assemblages in the migmatitic paragneiss are Grt-Sil-Kfs±Ky, with Crd±Sp forming during decompression from 8–10 kbar to 4–5 kbar after ca. 65 Ma (Norlander et al. 2002; Hinchey et al. 2006).

The following descriptions include results from Icebound Lake (this study), and Bearpaw Lake, Mount Thor, and Frigg Glacier as mapped and described by Hinchey (2005, her maps 3, 4, and 5). Icebound Lake, 1 km north of Cariboo Alp, is located on the southwestern flank of the Thor-Odin dome and represents the structurally highest rocks in the dome. The dominant diatexite migmatitic Grt-Kfs-Bt-bearing paragneiss (Plate 1n) is unconformably overlain by a southwest-dipping metasedimentary package dominated by Kfs-Sil-Grt-Bt-bearing paragneiss (Figs. 2b, 4), with a distinct layer of quartzite near the base of the package. There are well developed moderately southwest-dipping composite transposition foliations in metasedimentary rocks. These rocks are interpreted to be part of a southwest-dipping panel of transposed rocks on the lower limb of a regional southwest-dipping overturned synform, following the interpretation of Reesor and Moore (1971) and subsequently Williams and Jiang (2005) (Fig. 2b).

At Bearpaw Lake the dominant lithologies are migmatitic Hbl-Bt-bearing orthogneiss and migmatitic Grt-Sil-Bt-Kfs-bearing paragneiss containing abundant leucosome (Hinchey and Carr 2007). At Mount Thor, the dominant rock type is migmatitic Hbl-Bt-bearing orthogneiss, which is interpreted to be unconformably overlain by a heterogeneous package of moderately northeast-dipping metasedimentary rocks with a quartzite layer at its base (Reesor and Moore 1971; Duncan 1984; Spark 2001; Hinchey 2005). The Mount Thor area is located on the lower limb of a regional scale F2 northwest-verging fold (Fig. 2a) (Reesor and Moore 1971; Spark 2001; Hinchey 2005) (Fig. 2a). At Frigg Glacier, of the structurally deepest rocks exposed in Thor-Odin, the dominant lithologies are migmatitic Hbl-Bt-bearing orthogneiss and migmatitic Kfs-Hbl-Bt-bearing granodiorite; both containing abundant leucosome, with minor amphibolite boudins present throughout the area and the dominant structures are transposed foliations (Hinchey 2005).

Northeast-verging isoclinal F1 folds and tight, asymmetric, coaxial F2 and F3 folds generally dip to the southwest on the southwestern flank of the dome. All areas of the Thor-Odin dome and overlying rocks were affected by F4 folds (Fig. 2b); broad, upright folds which arch older structures. In the rocks overlying the Thor-Odin dome this folding did not significantly alter the map pattern created by F3 folds, but in the Thor-Odin dome itself, F4 folding was in part responsible for the outward-dipping domal geometry of the culmination (Williams and Jiang 2005) (Fig. 2b). In other words, the shape of the Thor-Odin dome is interpreted as a result of the interference pattern (Fig. 2b) between the late F4 and the dominant Eocene age F2–F3 structures within the dome (Hinchey 2005; Williams and Jiang 2005).

While there is no firm date for the onset of prograde metamorphism, it is likely that metamorphism was ongoing by the Paleocene to Eocene (Fig. 3), between ca. 65 and 56 Ma (Norlander et al. 2002; Hinchey et al. 2006). Monazite growth over a period of 62.3 ± 3 Ma to 50.1 ± 2 Ma (Coleman 1990; Hinchey and Carr 2007) has been interpreted as evidence for a protracted period of metamorphism. Monazite growth and recrystallization can occur over a wide range of temperatures in different lithologies and these ages do not necessarily represent peak metamorphic conditions (Parrish 1990; Spear 1993; Crowley and Parrish 1999; Gibson et al. 2004; Cubley et al. 2013). Monazite is predicted to be unstable above the Ms + Qtz = Sil + Kfs + melt reaction boundary (Cubley et al. 2013) and monazite growth in the migmatitic assemblages of the

Thor-Odin dome is therefore likely to represent post-peak metamorphism.

Deformation in the Thor-Odin dome interior was ongoing ca. 56–54 Ma as determined by U–Pb igneous zircon crystallization ages in deformed leucosomes (Hinchey et al. 2006). The age of at least some stages of isoclinal folding (Fig. 2b) and prograde metamorphism is as young as ca. 56–54 Ma (Norlander et al. 2002; Hinchey et al. 2006). Deformation ended between ca. 54 and 52.5 Ma (Hinchey et al. 2006), based on zircon U–Pb crystallization ages in undeformed leucosomes. Biotite cooling ages from the Icebound Lake area on the flank of the dome are between 52 and 51 Ma, and from the interior of the dome between 53 and 52 Ma, while muscovite cooling ages from Ladybird granite pegmatites at Icebound Lake are between 51 to 50.5 Ma (Van Rooyen 2013; Van Rooyen and Carr in press). The rocks of the Thor-Odin dome are estimated to have been exhumed from a depth of 26–33 km based on estimates of peak metamorphic pressures between 8 and 10 kbar (Norlander et al. 2002). When the rocks in the overlying Cariboo Alp area were structurally deactivated ca. 58 Ma, the rocks of the dome were still being penetratively deformed in the infrastructure. The Thor-Odin rocks were structurally deactivated during the ca. 54–52.5 Ma period when penetrative deformation ended. It is interesting to note that penetrative deformation in the Frenchman Cap dome north of Thor-Odin continued after ca. 52 Ma until ca. 49 Ma (Crowley 1999). Therefore, it is permissible to suggest that the Thor-Odin dome represents a higher structural level of infrastructure that was deactivated relative to the deeper rocks in the Frenchman Cap dome after ca. 52.5 Ma (Figs. 4, 5).

LATE CRETACEOUS TO EOCENE INTRUSIVE ROCKS – THE LADYBIRD GRANITE SUITE AS A STRAIN MARKER AND CONSTRAINT ON AGE OF DEFORMATION

The peraluminous, anatectic Ladybird granite suite is the dominant igneous suite in the Thor-Odin–Pinnacles area (Carr 1990, 1991a; Hinchey and Carr 2006). The largest intrusion of Ladybird granite in the area is the South Fosthall pluton, a roughly 4–5 km thick composite laccolithic complex where sheets of granite or pegmatite are concordant with the metasedimentary layers comprising the country rock (Carr 1990). The U–Pb zircon crystallization ages for the Ladybird granite range between ca. 64 and 52 Ma and are generally older towards the south in the Valhalla complex (Parrish et al. 1988) and the Grand Forks complex (Parrish 1992; Cubley et al. 2012, 2013). The youngest pegmatites related to the Ladybird granite are tourmaline-bearing and may locally be as young as ca. 50 Ma on the western margin of the Thor-Odin dome (Johnston et al. 2000).

Carr (1990) and Vanderhaeghe (1999) presented evidence of strain variations and strain partitioning within the South Fosthall pluton and surrounding Ladybird granite, indicating that the Ladybird granite was intruded, in part, while penetrative deformation was ongoing. This is consistent with observations in this study that some of the pegmatites in the Icebound Lake, Cariboo Alp, Fawn Lakes, and North Fosthall areas, including Twin Peaks and Mount Symonds, were deformed concordantly with the host rock, indicating that they experienced at least the last increment of deformation together. Other pegmatites of the area crosscut the fabrics in the host

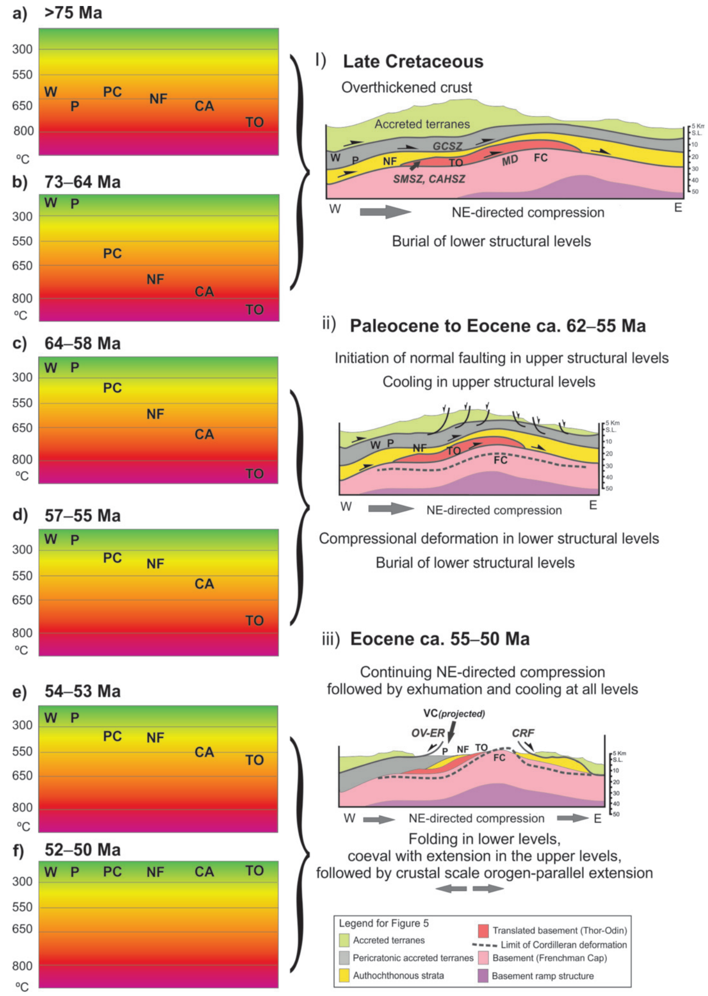

Figure 5. Illustration of the thermal evolution of the Thor-Odin–Pinnacles area as discussed in the text. The thermal cartoons on the left show the times at which different structural levels (as represented by specific study areas) passed through particular temperature ranges of 900°C for zircon crystallization, 650°C for wet granite solidus (Spear 1993), 550–570°C for argon closure temperature in hornblende, and 320–330°C for argon closure temperature in biotite (details about closure temperatures in Van Rooyen and Carr in press). They illustrate the simultaneous exhumation of some areas and burial of structurally deeper ones, based on metamorphic temperatures determined for various areas, and on $^{40}Ar/^{39}Ar$ cooling ages summarized in the text. They also illustrate the increased cooling rates in the lower structural levels, interpreted as resulting at least in part from the transport of these rocks over a basement ramp during NE-directed compression. The sketch cross-sections on the right show the general physical configuration of the area during different times and note important events and changes in deformation styles. (SMSZ – Slate Mountain Shear Zone; CAHSZ – Cariboo Alp High Strain Zone; CRF – Columbia River fault zone; OV–ER – Okanagan Valley–Eagle River fault zone; GCSZ – Gwillim Creek shear zone).

rocks and appear undeformed. The pegmatites in the Plant Creek area, in the upper part of the structural section, crosscut all the fabrics in the host rocks, indicating that the host rocks were not actively deforming during pegmatite intrusion. As mentioned in the preceding sections, muscovite and biotite cooling ages from Ladybird pegmatites are consistently between 51 and 50.5 Ma indicating rapid cooling in pegmatites throughout the area (Van Rooyen 2013; Van Rooyen and Carr in press).

The South Fosthall pluton is located between the shallowly east-dipping Columbia River fault and the moderately west-dipping Beaven fault. The structures in the South Fosthall pluton occur primarily as shallowly east-dipping mylonitic fabrics on the eastern side of the pluton, linked to the east-dipping Columbia River fault and, on the western side of the pluton, west-dipping directed mylonitic fabrics and shear zones are linked to the west-dipping Beaven fault (Parrish et al. 1988; Carr 1991b; Vanderhaeghe et al. 2003). Field relations and U–Pb zircon crystallization ages (by TIMS) in the east-dipping mylonitic fabric overprinting NE-directed compressional structures indicate that the Columbia River fault was actively deforming the Ladybird granite at 55.1 ± 1.5 Ma (Parrish et al. 1988), interpreted at the time as the earliest evidence of extensional activity on the Columbia River fault.

The South Fosthall pluton is one of several examples of laccolithic composite bodies of intrusive granite in the Eastern Internal zone of the Canadian Cordillera. The Pukeashun granite, an anatectic leucogranite suite of Eocene age, ca. 56 Ma, located to the northwest of the Thor-Odin dome in the Shuswap Lake transfer zone (SLTZ, Fig. 1) between the North Thompson–Adams and Okanagan Valley–Eagle River fault systems, occupies a similar structural setting as the Ladybird granite (Johnson 2006). The Ladybird granite is also present as a laccolithic complex of sheet-like bodies in the Valhalla complex south of Thor-Odin–Pinnacles, intruded between ca. 59 and 56 Ma (Carr et al. 1987), where the upper margin of the granite bodies coincide, in part, with the Valkyr Shear Zone, a zone of ductile deformation that is cut by the younger east-dipping Columbia River and the Slocan Lake normal faults (Carr 1995). Johnson (2006) suggested that the South Fosthall pluton, the Pukeashun granite and the Ladybird granite in the Valhalla complex, all occupy step-over zones between sets of extensional normal faults (Fig. 1), and that the distribution of these granite bodies may be linked to the presence of releasing zones in basement ramps. The emplacement of granite was localized in low-angle extensional shear zones, providing a weak rheology in which deformation was concentrated during extension (Johnson 2006).

DIACHRONOUS METAMORPHISM AND DEFORMATION IN THE THOR-ODIN–PINNACLES AREA; INFRASTRUCTURAL ZONES MIGRATING THROUGH TIME

The rocks of the Thor-Odin–Pinnacles area discussed in this study experienced protracted (but not necessarily continuous) deformation and metamorphism throughout the Cretaceous to Paleocene and into the Eocene, possibly overprinting Early Cretaceous or older events. We identify the time at which belts of rocks were penetratively deforming in the infrastructure, and when the formation of transposition foliation ended, and when the rocks cooled (Van Rooyen 2013; Van Rooyen and Carr in press). In the area between the Thor-Odin dome and the Whatshan batholith, the timing of tectonothermal events is younger structurally downwards, reflecting the downward migration of the active infrastructure boundary through time, in combination with upwards migration of rocks and NE-directed transport towards the foreland. Metamorphic grade generally increases downwards throughout the structural section into the dome. Based on data compilation which includes timing of metamorphism, deformation, anatexis in basement rocks, and intrusion of leucogranite, at least four belts of rocks, or tectonothermal domains, are recognized that experienced regional metamorphism and deformation at different times, with cooling rates that increase downward into the section (summarized below and on Figures 3 and 4).

Metamorphism in each structural level was broadly coeval with the penetrative ductile deformation, with thermal peak metamorphism predating the end of deformation. However, the timing of the onset of prograde metamorphism is uncertain. Periods of prograde metamorphism are recorded from ca. 130–90 Ma in rocks of the Whatshan area (Glombick 2005) continuing to sometime before ca. 75 Ma (Carr 1992); ca. 100–80 Ma in the Plant Creek–South Fosthall area (Lemieux 2006) and continuing to ca. 64 Ma (Carr 1992); ca. 73–62 Ma in the North Fosthall–Mount Symonds area (Carr 1991b, 1992); ca. 65–58 Ma at Cariboo Alp (Coleman 1990; Carr 1991b, 1992); and ca. 65–56 Ma in the Thor-Odin dome (Norlander et al. 2002; Hinchey et al. 2006, 2007).

Penetrative deformation progressively young down-section; from >75 Ma at the top in rocks of the Whatshan area (Carr 1991b, 1992), to ca. 56–54 Ma within the Thor-Odin dome (Hinchey et al. 2006) at the bottom, and represents progressive heating and deformation migrating downwards due to burial, radiogenic heating, incubation and structural weakening of structurally lower rocks. The rocks of the Whatshan–Pinnacles area were polyfolded in the Late Cretaceous (pre-ca. 75 Ma) and represent the infrastructure before ca. 75 Ma. They preserve Cretaceous metamorphism and deformation and were not remobilized during Late Paleocene to Eocene northeast-directed transport. At ca. 75–64 Ma, the Whatshan–Pinnacles area rocks were deactivated and acted as suprastructure to the underlying Plant Creek–South Fosthall area rocks, from the central part of the tract structurally above the South Fosthall pluton. This central part of the section was being penetratively deformed within the infrastructure at this time. After ca. 64 Ma, the Plant Creek–South Fosthall area rocks were deactivated, as demonstrated by the timing of the end of penetrative deformation. Rocks in the lowest part of the tract, below the South Fosthall pluton (i.e. Mount Symonds–Cariboo Alp area) continued to be deformed after ca. 64 Ma. After ca. 58 Ma the lowest part of the tract, the Mount Symonds–Cariboo Alp area, was no longer active relative to the infrastructure now comprised of the deforming rocks in the Thor-Odin dome. The Thor-Odin dome rocks remained in the infrastructure until ca. 52 Ma.

Hornblende cooling ages (for a closure temperature of ca. 550–570°C for the grain sizes and cooling rates in the study area – see Van Rooyen and Carr in press) span about 12 m.y.; ca. 62–58 Ma in the Plant Creek–South Fosthall area; ca. 57–55 Ma in the North Fosthall area; and 55–53 Ma on the upper margin of the dome at Cariboo Alp. Biotite cooling ages (for a closure temperature of 320–330°C) are all between 52.5 and 50.5 Ma throughout the entire structural section including the Thor-Odin dome, regardless of lithology, (including Ladybird pegmatites), indicating that the entire panel of rocks cooled through the closure temperature for biotite as a unit (ca. 320–330°C; see calculations in Van Rooyen and Carr in press). Muscovite cooling ages from Ladybird pegmatites are between 51.5 and 50.5 Ma from all parts of the structural section (for a closure temperature of 450–465°C – see calculations in Van Rooyen and Carr in press), indicating that the pegmatites all cooled at the same time regardless of structural position, and all cooled slightly later than their host rocks. This cooling history shows differential cooling rates throughout the study area, with rates increasing towards the deeper part of the section into the Thor-Odin dome. Cooling in the upper structural levels (as is the case with deformation and metamorphism already discussed) was coeval with deformation and anatexis in the lower structural levels, and the early cooling front (represented by hornblende cooling ages) migrated progressively down into the section. However, since the entire section has the same biotite and muscovite cooling ages, the area can be characterized as having experienced a major period of cooling and exhumation after 56–55 Ma, after which the tilted crustal section represented by this study area passed through the closure temperature for argon in biotite and muscovite close together in the period around 52–50 Ma. This interpretation is illustrated in Figure 5 as a thermal time series which shows the different times at which different study areas passed through particular temperatures. Figure 4 summarizes the times at which the different structural levels in the study area were actively deforming in the infrastructure and when they were deactivated.

HYBRID MODEL FOR THE EVOLUTION OF THE FRENCHMAN CAP – THOR-ODIN–PINNACLES TECTONOTHERMAL CULMINATIONS

As summarized in the introduction to this study, the formation of tectonothermal culminations in the southern Canadian Cordillera is primarily discussed in terms of diapirism, horizontal channel flow, or thrust-related deformation (cf. Carr and Simony 2006). In this section, we discuss the possible contributions of these mechanisms to the tectonic evolution of the Thor-Odin–Pinnacles area in the Late Cretaceous to Eocene.

It has been proposed that the driving force for the exhumation of Thor-Odin dome was gravitational instability created by the presence of anatectic melt and the subsequent ascent of partially molten crust in the core of the dome as a diapir (Vanderhaeghe et al. 1999, 2003; Norlander et al. 2002; Fayon et al. 2004; Teyssier et al. 2005). Evidence for decompression melting in the Thor-Odin dome (specifically the Bearpaw Lake area, Fig. 2) was reported by Norlander et al. (2002). Rocks underwent decompression from 8–10 kbar in the kyanite zone to 4–5 kbar, which equates to exhumation from ca. 25–33 km to ca. 15 km depth. These authors have suggested that this occurred as isothermal or near-isothermal decompression with coeval anatexis, based on reaction textures, mineral assemblages, and the extensive presence of anatectic melt in the rocks.

Models for the Thor-Odin dome invoking diapirism, delineated Cariboo Alp (Vanderhaeghe et al. 1999; Teyssier et al. 2005) as a possible boundary along which the dome is exhumed. In thermochronological studies of core complexes, as well as in numerical modelling, cooling ages always young towards the extensional shear zone (Tirel et al. 2006, 2008; Sullivan and Law 2007), which is the case here only for hornblende, not biotite or muscovite. Based on the hornblende and biotite cooling ages (Van Rooyen 2013; Van Rooyen and Carr in press), in combination with the timing of metamorphism and deformation, the geometry of the southwest-dipping package of rocks with their southwest-dipping transposition foliation in Cariboo Alp was likely formed before the Eocene cooling and exhumation of the dome. Movement along the Cariboo Alp area must have taken place at temperatures higher than 550–570°C (upper estimate of hornblende Tc), and before ca. 58 Ma (end of formation of transposition foliation), after which rocks on both sides equilibrated and cooled together to ca. 320–330°C by ca. 52 Ma. The Cariboo Alp area therefore did not accommodate km-scale extension or significant reactivation related to exhumation during the Eocene. Not only is the Cariboo Alp area not a major post-metamorphic extensional fault, it was also not involved in the exhumation of the dome, as discussed above.

Rey et al. (2009) proposed that one of the driving forces behind the formation of migmatite-cored tectonothermal culminations is the interplay between extension rates and the amount of melt present in the crust during extension. In two-dimensional thermal-mechanical modelling Rey et al. (2009) demonstrated that the fraction of melt present in the deeper structural levels of the crust during extension coupled with differences in extension rates produces domal structures with different characteristics. Their models showed that when extension rates are fast ($2*10^{-16}$/s, equivalent to 25.5 mm/y) domes are produced with weakly deformed migmatitic cores where migmatite crystallizes at low pressures (e.g. the Thor-Odin dome), and in settings where extension is slower ($2*10^{-15}$/s, equivalent to 0.85 mm/y), the cores of domes preserve pre-exhumation structures and crystallization of migmatite at depth (e.g. the Ruby Mountain–East Humboldt Range in Nevada). Increasing the melt fraction present produces faster exhumation rates and more extensive isothermal decompression (Rey et al. 2009).

As discussed earlier, the core of the Thor-Odin dome is not composed purely of diatexite migmatite and actually preserves significant NE-verging structures that developed during NE-directed transport toward the foreland. Some aspects of the Thor-Odin dome certainly reflect rapid melt ascent and crystallization at low pressure with fast extension rates as described in the model by Rey et al. (2009), such as a period of near-isothermal decompression between ca. 56 and 52 Ma (Norlander et al. 2002; Hinchey et al. 2006). In addition, the fast cooling rates predicted for domes like this (ca. 35–65°C/km) are actually lower than the rates suggested for the Thor-Odin dome (>100°C/km; Van Rooyen 2013; Van Rooyen and Carr in press). However, the cooling rates for the areas that structurally overlie the core of the dome also match or exceed the proposed rates but the pre-exhumation history and NE-verging structures are preserved in coherent stratigraphic successions. We propose that the Thor-Odin dome, therefore, preserves at least in part a history of pre-exhumation structural development that is not completely overprinted by Eocene anatexis and exhumation, as would be expected in a dome formed by diapiric ascent as the dominant process. In particular, the documented polyphase folding of leucosomes in the Thor-Odin dome (Hinchey et al. 2006) precludes this interpretation and the complexity of the infrastructure transitions and the cooling history of the rocks in the footwall of the Columbia River fault zone in the Thor-Odin–Pinnacles area are not consistent with the interpretation of the Thor-Odin dome as a diapir. The dome is best interpreted as a fold interference structure (Williams and Jiang 2005; Hinchey et al. 2006), perhaps amplified by some buoyancy effect, but the preservation of NE-directed compressional structures is not consistent with that of a diapir.

A variant of gravity-driven models to explain domal geometries presented by Rey et al. (2011) relates to so-called 'double domes,' where the main domal culmination contains two sub-domes, separated by a steeply dipping high strain zone, illustrated specifically in the Naxos dome in Greece, and the Montagne Noire in France. In two-dimensional thermal-mechanical modelling these double domes, of which the Thor-Odin dome could be an example, develop during initial convergence where crustal flow is primarily horizontal followed by the rotation of this flow in an upward direction during extension (Rey et al. 2011). Based on currently available data and the interpretations presented in this study, the Thor-Odin dome does not seem to be a good candidate for developing through the double dome model of Rey et al. (2011) because there is currently no good estimate for where the steep dividing shear zone would be. The dominant steep features in the Thor-Odin area are N–S striking, steep to near-vertical brittle faults (some dextral strike-slip, some normal) that post-date and dismember metamorphic sequences (Kruse and Williams 2007; Kuiper et al. 2015) and largely reflect extension after peak metamorphism and subsequent exhumation.

Another category of tectonic mechanisms proposed in the Cordillera to form tectonothermal culminations involves ductile extrusion and channel flow (Johnston et al. 2000; Williams and Jiang 2005; Brown and Gibson 2006; Glombick et al. 2006; Kuiper et al. 2006; Gervais and Brown 2011), in which mid-crustal rocks flow in a channel and are bounded by two opposite-verging shear zones at the top and base, active at the same time. The Frenchman Cap dome, immediately to the north of the Thor-Odin dome (Fig. 1) has been extensively studied with respect to exhumation mechanisms and deformation history. Crowley et al. (2008) and Gervais et al. (2010) documented the preservation of ca. 1.85 Ga Proterozoic deformation, metamorphism, and magmatism in the deepest exposed basement rocks of the Frenchman Cap dome, approximately 1.5 km structurally below the basal quartzite of the cover rocks. The uppermost 1.5 km of these Proterozoic rocks preserve a Cordilleran thermal overprint documented by monazite growth, but no documented Cordilleran deformation. Below the thermal overprint all recorded events are Proterozoic in age, and include ca. 1.9 Ga metamorphism and deformation, and 1.85 Ga igneous intrusions by the Bourne Granite suite. The preservation of Proterozoic deformation in the core of the Frenchman Cap dome indicates that it could not have formed as a result of vertical diapiric rise of partially molten crust (Crowley et al. 2008; Gervais et al. 2010), as proposed for the Thor-Odin dome (Vanderhaeghe et al. 1999, 2003; Norlander et al. 2002; Fayon et al. 2004; Teyssier et al. 2005) and the Okanogan dome (Kruckenberg et al. 2008). As is the case in the Frenchman Cap dome, the preservation of extensive folded and refolded migmatites and the presence of generally NE-directed deformation fabrics in the Thor-Odin dome are inconsistent with diapirism as the sole mechanism of exhumation.

Gervais and Brown (2011) proposed a model of formation for the Frenchman Cap dome that is based in the sequential extrusion of mid-crustal material up a basement ramp, in which the Monashee décollement represents the lower boundary of a mid-crustal channel and the Okanagan Valley–Eagle River Fault represents the upper boundary, or lid, of the channel. In this model, the rocks of the Lower Selkirk Allochthon (taken to be equivalent to the rocks between Thor-Odin and Pinnacles) were exhumed while the underlying Monashee complex rocks were being buried. Movement on the east-verging Monashee décollement in the Frenchman Cap dome was coeval with amphibolite-facies metamorphism (forming kyanite and K-feldspar) and penetrative ductile deformation between ca. 60 and 55 Ma (Parrish 1995; Crowley and Parrish 1999; Crowley et al. 2001; Gibson et al. 2004). The main period of formation for north-northeast-directed folds and transposition foliations in a top-to-the-east shear zone in the Frenchman Cap dome was between ca. 53 and 49 Ma (Crowley et al. 2001; Gervais et al. 2010) (Fig. 4).

Despite documented differences in the pressures, temperatures, and deformation histories of the Frenchman Cap dome and the Thor-Odin dome discussed elsewhere (Hinchey 2005; Gervais 2009) some aspects of the evolution of the Thor-Odin dome are broadly similar in the Frenchman Cap dome. Specifically, the thermochronology data showing that the upper structural levels (e.g. Plant Creek area) were being cooled through 550°C while the lower levels were being penetratively deformed, followed by crustal scale extension over the whole area, is kinematically consistent with the channel flow model for Frenchman Cap discussed by Gervais and Brown (2011). Carr and Simony (2006) argued against a full model of channel flow in the southeastern Canadian Cordillera and have suggested instead that channel flow was active for a short duration and was arrested before extrusion occurred; however, the duration of channel flow or the extent to which the channel extruded in either dome does not affect the major conclusions of this work. In subsequent work, Simony and Carr (2011) presented a model in which sequential ductile thrusts in the hinterland of the Cordillera can be linked to major thrust faults in the foreland fold and thrust belt that suggested that channel flow is not a necessary mechanism in the SE Cordillera and that the deformation histories can be explained by internal deformation in crystalline thrust sheets.

We propose in this study that the Thor-Odin dome–Pinnacles area is best viewed as a coherent crystalline thrust sheet that was transported and shortened during NE-directed compression (Fig. 5). During this process it was transported up a basement ramp (analogous to the Frenchman Cap dome), during which the ages of metamorphism and penetrative deformation young downwards into the section as progressively lower levels get heated and deformed (Fig. 5). This type of transport of ductile rocks over basement ramps has also been documented for the Valhalla dome south of Pinnacles, where the Gwillim Creek shear zone represents a major ductile shear zone and represents the top of a series of Late Cretaceous to Eocene belts of younging-downward infrastructure (Carr and Simony 2006; Hallett and Spear 2011; Simony and Carr 2011), consistent with the transitions documented in this study. The Gwillim Creek shear zone is known to have transported rocks over a basement ramp (Hallett and Spear 2011) and it was suggested by Simony and Carr (2011) that this type of transport is a mechanism that can be extended to the Thor-Odin dome. The Gwillim Creek shear zone was active between ca. 90 Ma and 60 Ma and is interpreted to have transported a coherent crystalline thrust sheet during NE-directed compression in the Cretaceous, with displacements of 80–100 km (Carr and Simony 2006). It was active during a period of thickening of the orogen in the Late Cretaceous, which allowed for metamorphism and deformation in the rocks above the nascent shear zone, which were then transported up a basement ramp during Paleocene northeast-directed compression and cooled against a cold basement footwall (Hallet and Spear 2011). The hanging wall of the Gwillim Creek shear zone records a downward-increasing strain gradient through an approximately 30 km-thick thrust sheet, which contains (among others) Jurassic to Cretaceous intrusive rocks (Simony and Carr 2006). The downward-increasing strain gradient reflects the progression of deformation through the structural section; penetrative deformation youngs downward throughout the section, preserving the youngest deformation in the lowermost structural levels. This is similar to the general progression of deformation and metamorphism in the Thor-Odin–Pinnacles area described in this study. The Gwillim Creek shear zone is cut by the 59–55 Ma, extensional Valkyr-Slocan Lake fault system (Carr and Simony 2006), which is analogous to the Columbia River fault zone in the Thor-Odin–Pinnacles area. We suggest that the Thor-Odin–Pinnacles area may host an analogue for the Gwillim Creek shear zone in the Pinnacles culmination, where a thrust-sense ductile fault marks the approximate transition between deformation in the period between 75 and 64 Ma in the Whatshan–Pinnacles area, and after ca. 64 Ma in the Plant Creek area. In this area, the Slate Mountain Shear Zone and Cariboo Alp High Strain Zone (Figs. 2, 4, and 5) may be additional shear zones accommodating NE-directed compression on which overlying thrust sheets were carried over and incubated at structurally deeper levels. Extending this model to the Frenchman Cap dome, we suggest that the Monashee décollement carries rocks over the Frenchman Cap dome in the same manner, consistent with Gervais et al. (2010) and Gervais and Brown's (2011) model of sequential extrusion of mid-crustal rocks above a basement ramp.

CONCLUSIONS

In our view, it is not appropriate to look only at the core of the Thor-Odin dome to evaluate possible mechanisms for its formation; the structurally overlying panel (here illustrated as the area between the dome and the Whatshan batholith) has to be taken into account as well. When this area is viewed as a whole, the preferred model for the evolution of the Thor-Odin–Pinnacles area suggested by this study is, therefore, a hybrid model in which the primary tectonic process driving deformation and metamorphism was crustal thickening in a convergent setting (Fig. 5). Diapirism and gravity-driven buoyant exhumation of deep crustal rocks were definitely important factors in the exhumation of some of the rocks of this area (Norlander et al. 2002); however, the structural styles in the rocks dictate that they were limited in duration and spatial extent to areas in the core of the Thor-Odin dome. The rocks of the Thor-Odin–Pinnacles area were progressively heated and deformed as a result of crustal thickening as they were transported to the east to northeast over a basement ramp in the Late Cretaceous to Paleocene (cf. Gervais et al. 2010; Hallett and Spear 2011).

The top of the section (Whatshan to Pinnacles to South Fosthall) was the first to be penetratively deformed, and also the first to be deactivated. The structural style in the Whatshan–Pinnacles area is sufficiently different from that in the Plant Creek to Mount Symonds areas that the Whatshan–Pinnacles area is interpreted here as analogous to the upper plate of the Gwillim Creek shear zone as exposed in the Valhalla complex to the south, here represented by the South Fosthall to Mount Symonds areas. The structurally lowest part of the section in the core of the Thor-Odin dome hosts the youngest penetrative deformation, and was the last to be deactivated. This downward-younging progression continues into the Frenchman Cap dome, where the last increments of NE-directed deformation and transposition were coeval with cooling and deactivation in Thor-Odin. Compression-driven deformation in Frenchman Cap dome continued to ca. 49 Ma, after which it was also deactivated. The initial cooling to below ca. 550–570°C between 62 and 58 Ma in the upper part of the section (Pinnacles to South Fosthall) occurred as a result of syn-convergent exhumation of the suprastructure. The switch to cooling across the whole area, linked to crustal-scale extension, occurred only when the Columbia River fault cut across the entire crustal section, when all the rocks in the Thor-Odin–Pinnacles area cooled through ca. 320–330°C as a unit between ca. 52.5 and 51 Ma. This crustal scale extension marked the beginning of a period of intrusion by mantle-derived igneous rocks throughout the southeastern Canadian Cordillera (Adams et al. 2005) in an extensional tectonic regime.

In conclusion, the rocks between the Whatshan–Pinnacles area and the interior of the Thor-Odin dome record Cretaceous metamorphism and cooling in the upper structural levels, and three stages of infrastructural flow at progressively deeper crustal levels in the Late Cretaceous, Paleocene and Eocene. The generally downward-younging progression of metamorphism and deformation, as well as the structural style and transposition foliation in the lower part of the section, is consistent with a model in which rocks were progressively heated and deformed during compressional transport, possibly up a basement ramp, which facilitated doming and exhumation. This is the first study to discuss this area in terms of infrastructure transitions and to explain contrasting histories in different parts of the Thor-Odin–Pinnacles area. This framework makes it possible to explain seemingly conflicting geological histories in adjacent areas while linking the different histories into an internally consistent geological model, and can potentially be applied to complex orogenic belts in multiple tectonic settings.

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