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Cretaceous to Eocene evolution of the southeastern Canadian Cordillera: Continuity of Rocky Mountain thrust systems with zones of "in-sequence" mid-crustal flow

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The ∼400 km wide, east-verging, retrowedge side of the southeastern Canadian Cordillera was predominantly formed in a tectonic setting of oblique plate convergence during the Cretaceous to Eocene. This paper documents the internal geometrical development of the retrowedge. In the External zone, the Rocky Mountains and Foothills are characterized by three major east-verging, Late Cretaceous to Eocene, thin-skinned, piggyback thrust and fold systems. They root westward into a basal décollement and accommodated ∼180 km of shortening. The Western Internal zone is characterized by tracts of metamorphic rocks and metamorphic core complexes (e.g. Kettle, Okanagan, Priest River and Valhalla), some of which are basement-cored domes (e.g. Frenchman Cap, Thor-Odin, and Spokane). They have a downward-younging progression of Late Cretaceous to Eocene metamorphism and deformation in infrastructural flow zones characterized by transposition foliation, migmatites, flow folds and 1–7 km thick shear zones. In the Eastern Internal zone, a relict ∼100–200 km wide Early Cretaceous orogen, that predated emplacement of ca. 100 Ma plutons, is nested between the External and Western Internal zones. The geology and architecture of the Internal and External zones can be explained by progressive development of major Late Cretaceous to Eocene shear zone systems in the Internal zone that can be directly linked with coeval thrust and fold systems in the External zone. The linkage was via Late Cretaceous activation and Late Cretaceous to Early Eocene reactivation of the 150–200 km-wide central portion of the Rocky Mountain basal décollement that lies beneath and translated the intervening Early Cretaceous orogen. During the latest stages of shortening, in the Early Eocene, extensional shear zone systems in the Internal zone, localized on tectonothermal culminations, were concomitant with shortening in the External zone. Motion of deep-seated Early Eocene décollements beneath some of these culminations may have contributed to their doming. Crustal shortening ended at ca. 52 Ma due to a change in tectonic setting to that of a transtensional tectonic regime, coinciding with the end of thrusting in the External thrust belt and with crustal-scale extension in the Western Internal zone.

1. Introduction

In order to understand the mechanics of an orogen it is important to understand the progressive geometric and kinematic evolution of structures throughout the crust, in particular, interactions between structures in the ductile core of the orogen and those in the thrust and fold belt of the foreland. However, the relationships between the structures in the core and thrust belt may be difficult to ascertain. For example, if the infrastructure has overridden the foreland at the youngest stages of orogenesis, perhaps though thrusting or channel extrusion, then relationships between older coeval structures in different parts of the orogen may be, in part, obscured. In orogens such as the Alps (Escher et al., 1993), southern Appalachians (Hatcher et al., 2007) and Scandinavian Caledonides (Gee et al., 1985), crystalline megathrust sheets, or "allochthons," which include metamorphic complexes, overrode the foreland belt forming its upper boundary. In the Himalayan orogen, the Greater Himalayan migmatitic gneiss sheet, interpreted as a mid-crustal channel (Searle et al., 2006) with the regional-scale Main Central Thrust at its base (Stephenson et al., 2001), flowed over or was sheared over the evolving foreland thrust belt

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(Robinson et al., 2006). In the case of the Grenville orogen, mid- and lower-crustal rocks are exposed at the surface across most of the orogen; however, the suprastructure and foreland have been eroded away (Rivers, 2009; Hynes and Rivers, 2010).

From its inception as a small cold accretionary orogen (Beaumont et al., 2006), in the Middle Jurassic, the southern Canadian Cordilleran orogen was deformed, widened and thickened throughout the Cretaceous and Early Eocene; the foreland thrust and fold belt expanded by both incorporating and overriding foreland basin strata (Evenchick et al., 2007, and references therein). By the Late Cretaceous, the orogen had evolved into an ~800–900 km wide warm "Cordilleran" (Beaumont et al., 2006) bivergent orogen. Its development was arrested in the Early Eocene due to a change in plate convergence vectors, changing to a setting of transtension (Price, 1994), before the orogen could become a large hot orogen like the Grenville and Himalaya. In the Internal or core zone, the deepest levels are exposed, in part, due to Eocene extensional exhumation. From bottom to top, the structural section comprises undeformed basement that lay beneath the Cordilleran deformation event (Crowley, 1999; Crowley et al., 2008; Gervais et al., 2010), a stacked sequence of mid-crustal migmatitic infrastructural flow zones that young downward (Parrish, 1995; Reid et al., 2003; Glombick, 2005; Williams and Jiang, 2005; Hinchey et al., 2006; Gibson et al., 2008; Gordon et al., 2008; Carr and Simony, 2006; Gervais, 2009; Gervais et al., 2010), each with evidence of its infrastructure to suprastructure transition (Murphy, 1987; Reid et al., 2003; Glombick, 2005; Hinchey et al., 2006; Carr and Simony, 2006) and the erosional remnants of an upper crustal supracrustal lid that preserves remnants of both a Middle Jurassic orogen (Evenchick et al., 2007, and references therein) and an Early Cretaceous orogen (this study, and references herein). The subsurface geometry and thickness of the crust is well constrained by geophysical data (Cook et al., 1992; Cook, 1995).

The general concept that there was a dynamic connection between deformation in the core or Internal zone of the southeastern Canadian Cordillera and that of the foreland thrust and fold belt or External zone has been suggested for some time (Bally et al., 1966; Price and Mountjoy, 1970). Parrish (1995) and Brown (2004) noted the downward younging of ductile deformation in the Internal zone and suggested that progressive deformation of the core zone with transient shear zones and some form of underplating could link with progressive development of the foreland, although they did not suggest particular correlations. Brown et al. (1992) and Johnson and Brown (1996) proposed that the Monashee décollement in the Internal zone accommodated displacement that fed into the Rocky Mountain basal décollement of the foreland thrust belt, and Doughty et al. (1998) and Carr and Simony (2006) argued that particular shear zones in the Spokane and Valhalla complexes, respectively, were linked with the basal décollement. Tectonic models have focused on the concept of an orogenic wedge (Price, 1981, 1994; Brown, 2004; Stockmal et al., 2007; Evenchick et al., 2007) and tectonic wedging (Price, 1986; Colpron et al., 1998). Analyses of the Internal zone have focused on the structural style and role of infrastructural flow zones including models that incorporate channels within the core zone that are somehow linked to and coeval with contraction and thrusting in the foreland (Johnston et al., 2000; Glombick, 2005; Williams and Jiang, 2005; Brown and Gibson, 2006; Williams et al., 2006; Gervais and Brown, 2011); models that invoke detachment flow (c.f. Williams and Jiang, 2005), rather than channel flow, linked to shortening in the foreland thrust belt (Carr and Simony, 2006), or models that attribute infrastructural flow to extensional orogenic collapse, perhaps with a contribution of weakening and/or buoyancy forces related to in situ anatexis (Norlander et al., 2002; Whitney et al., 2004; Teyssier et al., 2005, and references therein; Glombick et al., 2006; Kruckenberg et al., 2008).

The analysis presented in this paper addresses questions as to the internal geometric evolution of the orogenic retrowedge in the southern Canadian Cordillera as it accommodated the westward underthrusting of the craton (Coney and Evenchick, 1994). This is the first study that attempts to track the detailed geometric and structural development, stage by stage, through the Cretaceous to Eocene. We distinguish: (i) Late Cretaceous to Paleocene, (ii) Paleocene and (iii) Late Paleocene to Eocene structures in the Internal core zone, and for structures of each of these ages, we propose linkages with thrust and fold systems of the same age in the External zone via Late Cretaceous activation and Late Cretaceous to Eocene reactivation of the central portion of the Rocky Mountain basal décollement. Our refined kinematic model for the southeastern Canadian Cordillera in the Late Cretaceous to Eocene yields insights into the systematics of shear zone – thrust and fold belt systems, shear zone localization and the lateral brittle–ductile transition, and the role of the basal décollement in the orogenic retrowedge.

2. Geological setting of the southeastern Canadian Cordillera

2.1. Mesozoic – Early Tertiary tectonic setting and lithotectonic elements

In this paper we focus on the east-verging foreland, or retrowedge, side of the orogen in the southeastern Canadian Cordillera, between latitudes 49°N and 54°N. The physiographic locations presented on Fig. 1 (e.g. Rocky Mountains, Selkirk Mountains, etc.) are used to locate geologic elements and will be used throughout the paper as a reference frame.

The geology of the southeastern Canadian Cordillera comprises the following lithologic and tectonic elements (Fig. 1; see Price, 1994, and references therein; and Evenchick et al., 2007, and references therein, for further overviews and in-depth citations to published literature). The *External zone* of the orogen, or Foreland thrust and fold belt, occurs in the Foothills, and in the Main and Front ranges of the Rocky Mountains. It consists of Paleozoic and Mesozoic strata of the North American platform and margin, and Mesozoic to Paleocene strata of the Foreland Basin which were deformed during successive pulses of Cretaceous to Eocene thin-skinned, east-verging thrusting and folding above a basal décollement. The *Eastern Internal zone* of the fold and thrust belt occurs in the western Rocky Mountains and the ranges on the west side of the Rocky Mountains (i.e. Purcell, Selkirk, and Cariboo mountains). It includes dominantly clastic Proterozoic and Paleozoic North American rocks. Locally sheets of Laurentian (Canadian Shield) basement were tectonically interleaved and interfolded with the cover (e.g. Malton complex). The *Western Internal zone*, the metamorphic and plutonic core of the orogen, is located in the southern Selkirk and western Purcell mountains and, in part, in the Cariboo, Selkirk and Monashee mountains, and in the uplands to the west of these ranges (Fig. 1). The Western Internal zone comprises clastic Proterozoic and Paleozoic North American strata as well as the inboard part of the hinterland, namely the inner accreted terranes, including the Slide Mountain marginal basin and Quesnel volcanic arc terranes (see Fig. 1 legend), that were accreted by eastward thrusting onto North American strata in the Late Triassic to Middle Jurassic due to westward underthrusting of the craton. In the Western Internal zone, Laurentian (Canadian Shield) basement rocks are exposed in the Thor-Odin and Frenchman Cap domes. Predominant igneous suites, in the Internal zone, that provide important timing constraints on tectonothermal events include: Middle Jurassic calc-alkaline plutons derived above a subduction

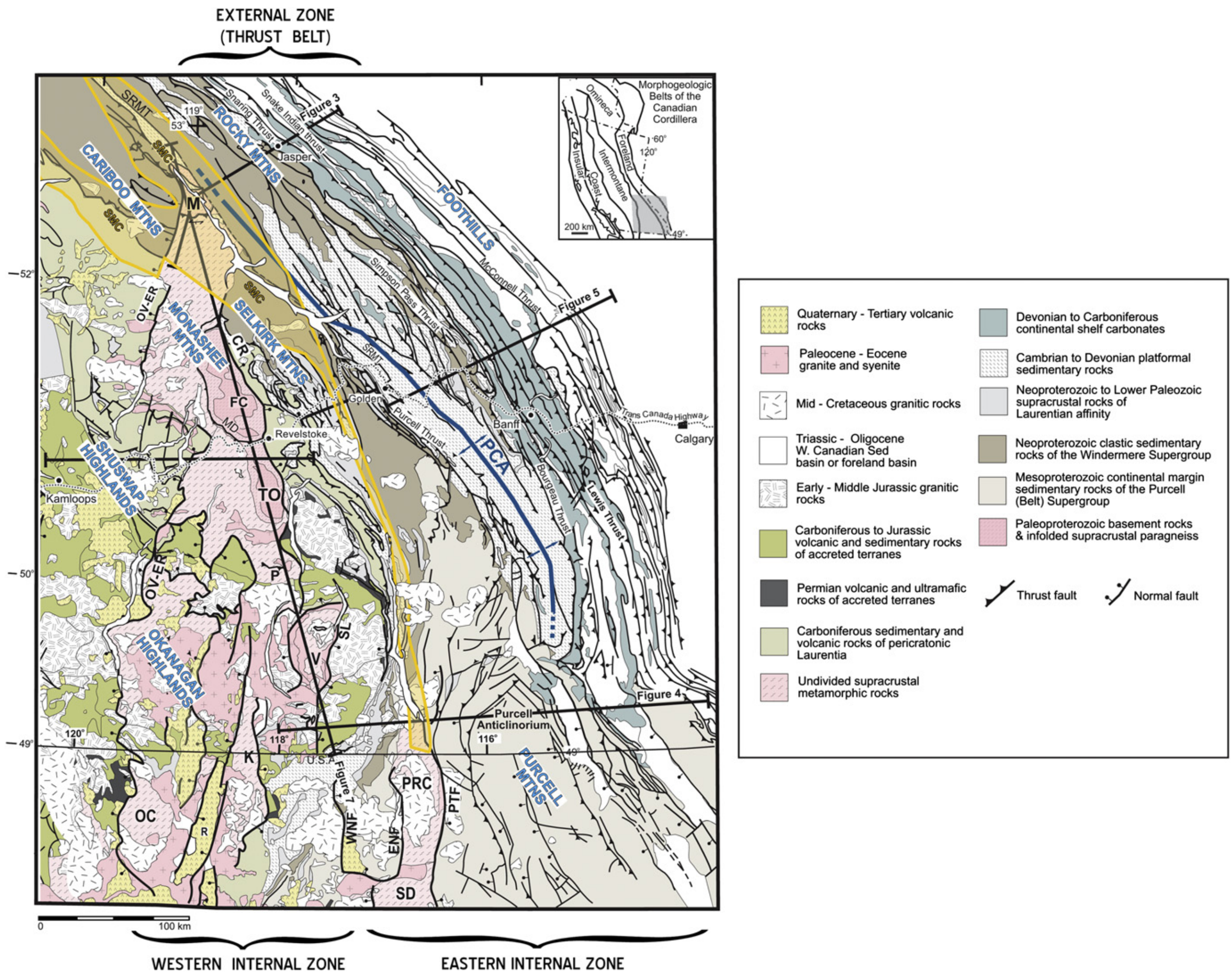

Fig. 1. Geological map of the southeastern Canadian Cordillera showing the External, Western Internal and Eastern Internal zones, cross-section locations for Figs. 3–5 and 7, and general locations of the mountain ranges referred to in the text (modified after Wheeler and McFeely, 1991; Carr, 1991, and Johnson and Brown, 1996). 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 and bounding Western (WNF) and Eastern Newport Faults (ENF). PTF = Purcell Trench Fault and SD = Spokane dome, In the Western Internal zone, K = Kettle – Grand Forks Complex, OC = Okanagan–Okanogan complex, R = Republic Graben, V = 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), and Slocan Lake-Champion Lake fault systems (SL). PCA = Porcupine Creek Anticlinorium; SMC = Selkirk–Monashee–Cariboo metamorphic complex; SRMT = southern Rocky Mountain Trench.

zone in a continental arc; and Early Cretaceous and Paleocene granites and leucogranites generated by crustal thickening and anatexis (Armstrong, 1988; Ghosh and Lambert, 1995).

2.2. Early Cenozoic extension

The tectonometamorphic culminations of complex high-grade metamorphic rocks in the Internal zone (Fig. 1) have Paleocene and Eocene cooling histories (Parrish et al., 1988; Harms and Price, 1992; Doughty and Price, 1999; Kruckenberg et al., 2008; Gervais et al., 2010). These culminations are generally bounded by outwardly dipping ductile and brittle normal faults. The extent and importance of extension in the genesis of individual complexes is debated (Journeay and Brown, 1987; Parrish et al., 1988; Teyssier et al., 2005; Glombick et al., 2006; Thompson et al., 2006; Brown, 2010) and the importance of extension in the doming of some of the culminations has been questioned (Carr and Simony, 2006; Gervais et al., 2010, and references therein). In our view, the normal faults are important and it is largely because of extension and related denudation that the rocks that were in the middle crust in the Cretaceous to Eocene are, in part, exposed for our study of their record of compressional tectonics. The timing, nature and origin of the culminations differ, as will be explained in this paper, and therefore the importance of extension varies significantly from culmination to culmination.

Price (1994) suggested that the major period of Paleocene to Eocene extension marks the end of compression, and therefore approximates the end of thrust propagation in the foreland belt. In the Scandinavian Caledonides, Fossen (2000) documented extension in two modes. At an early stage (Mode I) extensional faults merged into shallow detachments; the basal décollement was not cut by extensional faults, and extensional faulting was synchronous with thrusting in the foreland (Northrup, 1996). Later, Mode II extension involved the entire crust and marked the end of thrusting. A similar scenario may apply in the southern Canadian Cordillera in the Eocene. Prior to ca. 52 Ma, extensional detachment

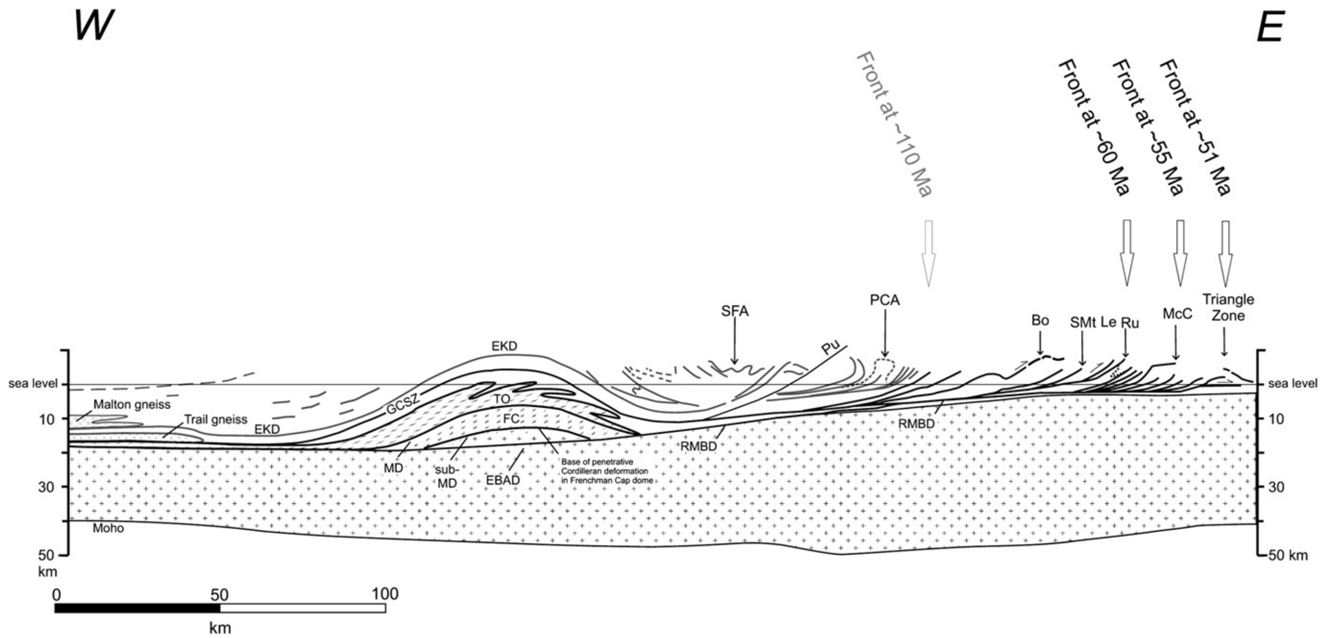

Fig. 2. Generalized schematic southwest–northeast cross section across the southeastern Canadian Cordillera showing linkages between flow zones in the Western Internal zone and thrust fault systems in the External zone, and the location of the Early Cretaceous and Late Cretaceous to Eocene orogenic fronts in the External zone. Structural culminations include the Frenchman Cap (FC) and Thor-Odin (TO) domes. The Selkirk Fan (SFA), Purcell thrust (Pu) and Porcupine Creek Anticlinorium (PCA) are located in the Eastern Internal zone. Shear zones and faults include the following: (1) Early Cretaceous basal décollement (EKD), (2) Cretaceous Gwillim Creek shear zone (GCSZ) and Bourgeau (Bo), Sulphur Mountain (SMt), Lewis (Le) and Rundle (Ru) thrusts; (3) Late Cretaceous to Eocene Monashee décollement (MD) and McConnell thrust system (McC); and (4) Paleocene to Eocene Basal décollement (EBAD) and Foothills thrusts.

faulting was localized in tectonothermal culminations in the Internal zone, and this extensional faulting was concomitant with thrusting in the foreland. After ca. 52 Ma, east–west extensional features penetrated the entire crust and thrusting ended. Early, ca. 59–54 Ma gently to moderately dipping extensional shear zones in the Valhalla complex (Carr et al., 1987), eastern margin of the Monashee complex (Parrish et al., 1988), early motion on the Okanagan Valley – Eagle River detachment system (Brown, 2010 and references therein) and perhaps in the Kettle dome and Priest River complex (Doughty and Price, 1999) could represent Mode I extension during foreland thrusting. Mode II, crustal-penetrating, east–west extension was beginning at ca. 52–51 Ma as evidenced by: motion on regional detachment fault systems (e.g. continued motion on the western margin of the Okanagan complex); N–S striking brittle faults, syenitic plutons with N–S striking vertical walls and steep N–S striking lamprophyre dyke swarms; extensive Eocene magmatism and volcanism with geochemistry and isotopic signatures indicating mantle contamination; and, rapid low temperature cooling of deep structural levels of tectonothermal culminations (Armstrong, 1982, 1988; Tempelman-Kluit and Parkinson, 1986; Parrish et al., 1988; Stinson and Simony, 1994; Teyssier et al., 2005; Glombick et al., 2006; Lemieux, 2006; Gordon et al., 2009; Brown, 2010, and references therein; Gervais et al., 2010). The timing of Mode II extension is synchronous with known dates of the last stages of ductile compressional structures exposed in the Internal zone of the orogen (i.e. deepest levels of Cordilleran deformation in Frenchman Cap dome, Gervais et al., 2010).

2.3. *Stratigraphic continuity between Internal and External zones of the orogen*

In our description of the kinematics of the orogen we accept the observations that Proterozoic and Paleozoic successions of the North American (Laurentian) platform can be traced from the External thrust belt westward into the Internal zone. The key successions include Mesoproterozoic (McMechan, 1981; Aitken and McMechan, 1991; Price and Sears, 2000), Neoproterozoic (Pell and Simony, 1987; Ross and Murphy, 1988; Grasby and Brown, 1993), Ediacaran and lower Cambrian (Kubli and Simony, 1992; Lickorish and Simony, 1995) and Upper Cambrian, Ordovician and Silurian (Aitken and Norford, 1967; Price and Mountjoy, 1970; Price et al., 1980; Root, 2001) successions. Recently, Johnston (2008) and Hildebrand (2009) proposed continental-scale tectonic-process models that include a ribbon continent "Saybia" or "Rubia" separated from Laurentia ("ancient North America") by an ocean basin until Early Cretaceous collision along a cryptic suture in the western Rocky mountains. These models, which attempt to resolve several tectonic problems, fail in the southeastern Canadian Cordillera because of the unequivocal former lateral continuity of tectonostratigraphic assemblages from the Alberta platform to the region deformed and metamorphosed in the Jurassic and Early Cretaceous.

3. The Late Triassic – Jurassic and Early Cretaceous orogens

3.1. *The Late Triassic – Middle Jurassic "small cold accretionary orogen"*

In the Internal zone, west of the Rocky Mountains, thrusting, folding, metamorphism and plutonism occurred in a 100–300 km wide belt prior to the Middle Jurassic due to obduction of the most inboard of the accreted terranes (see Evenchick et al., 2007 and references therein). Deformation began in the Late Triassic and/or Early Jurassic with the NE-vergent thrust emplacement of rocks of Slide Mountain and Quesnel terranes over Proterozoic and Paleozoic strata of the western margin of Laurentia (Struik, 1981). This resulted in a suprastructure of structurally imbricated low-grade rocks with upright structures, and an infrastructure of great southwest-verging recumbent nappes. Locally, Paleoproterozoic basement was structurally imbricated with supracrustal rocks, as exposed in the Malton complex (McDonough and Simony, 1988; Murphy et al., 1991; McDonough and Parrish, 1991). Structures were overprinted by a broad belt of low-grade metamorphism and were intruded by Middle Jurassic plutons with high-temperature low-pressure aureoles (c.f. Evenchick et al., 2007, and references therein). Part of this orogenic welt was exhumed and became the suprastructure through subsequent orogenesis, and part was progressively reactivated or overprinted by younger deformation.

3.2. The Early Cretaceous orogen in the Eastern Internal zone

An Early Cretaceous orogen, with a fold and thrust belt, a metamorphic core zone and a late- to post-kinematic plutonic arc, is embedded within the younger Late Cretaceous to Eocene orogenic edifice (Figs. 1–5). This orogen occupies much of the Eastern Internal zone in the Purcell, eastern Selkirk and eastern Cariboo mountains and the northeastern portion of the Monashee Mountains; its thrust belt and eastern margin lie in the Western Rocky Mountains (Fig. 1). The thrust belt is 50–150 km wide with the 350 km long Porcupine Creek anticlinorium at its leading edge (Figs. 1 and 2). The anticlinorium is a detachment fold above the Lower Cambrian orthoquartzite. The basal detachment of the fold is also the eastern edge of the Early Cretaceous basal décollement (Stretch, 1997; Larson and Price, 2006, and references therein). Granitic plutons with ages of ca. 109 Ma intruded the thrust faults on the steepened western flank of the anticlinorium near Latitude 49°30′ N (Larson et al., 2006). In the northern part of the anticlinorium, deeper décollements were activated (Ferri, 1984; Gal et al., 1989; Lickorish, 1993), and at Latitude 52°30′ N, east of the Malton complex, the basal décollement of the anticlinorium emerges in a structural culmination as the Ptarmigan décollement, a mylonitic thrust (Murphy, 2007; Dechesne, 1992). The Ptarmigan décollement is linked to the Monarch–Snaring thrust system, west of Jasper (Mountjoy, 1980), and with the décollement under the Malton complex (Figs. 1 and 3). The Malton complex consists of several km-thick sheets of basement (Morrison, 1982; McDonough and Simony, 1988). The ca. 2100–1860 Ma U–Pb ages (Murphy et al., 1991; McDonough and Parrish, 1991) match those of North American (Laurentian) basement of Alberta (Ross, 1991). The basement sheets were interleaved with cover and imbricated in the Middle Jurassic (Raeside and Simony, 1983; Murphy et al., 1991). The Early Cretaceous basal décollement dips southwestward under the Malton complex toward the Frenchman Cap dome (Fig. 3). It is relatively high structurally because the trailing edges of all of the Late Cretaceous to Eocene thrust faults and their common basal décollement project underneath it. Near Golden (Figs. 2 and 5), the Early Cretaceous basal décollement dips westward under the thrusts of the northern Purcell Mountains (Kubli and Simony, 1994) and under the Selkirk Fan (Colpron et al., 1998). Early Cretaceous thrusts of the southern Purcell Mountains merge into a basal décollement that may be imaged by reflection traces on geophysical reflection profiles (Cook, personal communication, 2010; Cook

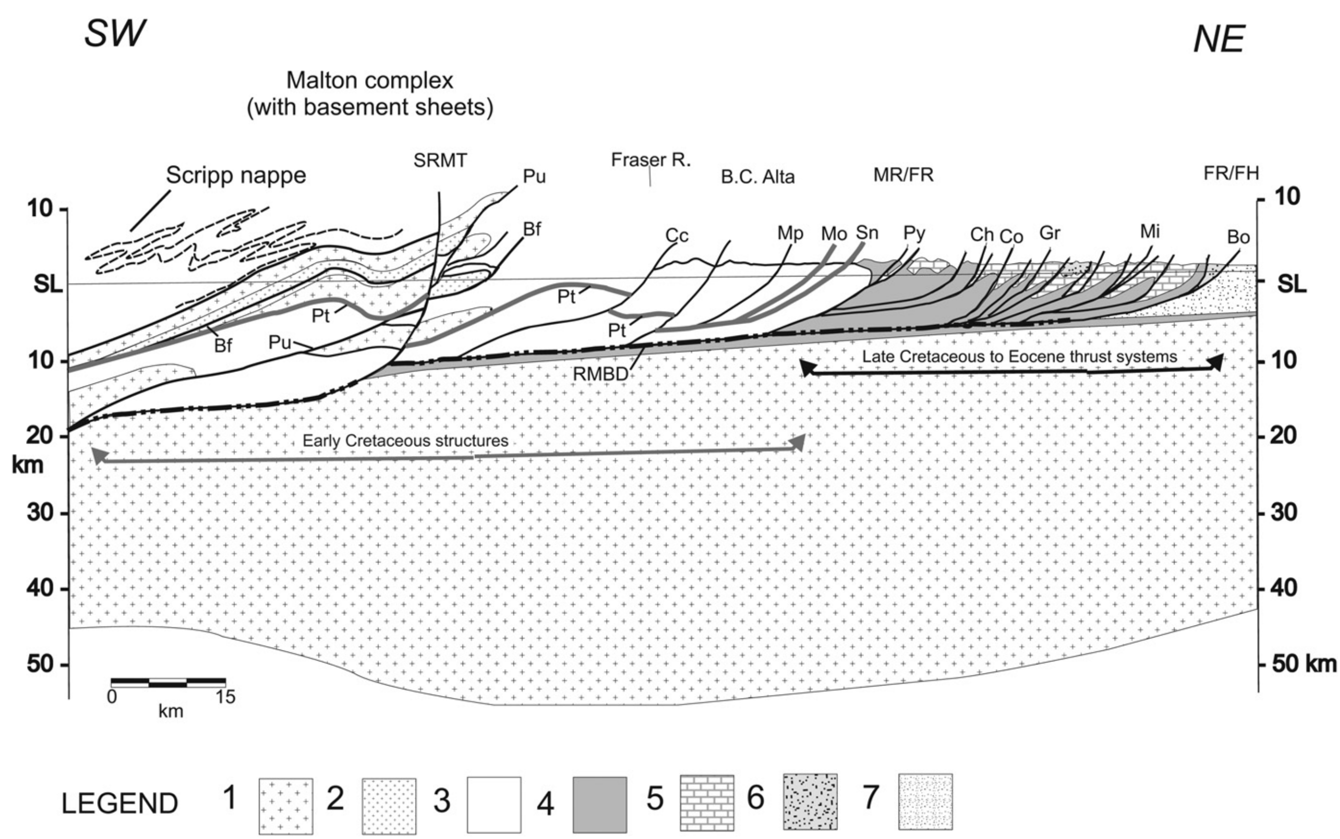

Fig. 3. Southwest–northeast cross section through the Malton complex (Fig. 1) showing the Early Cretaceous orogen located at higher structural levels and to the west of the Late Cretaceous to Eocene thrust and fold belt, that roots into the Rocky Mountain basal décollement (RMBD). The Early Cretaceous Ptarmigan (Pt)-Bear Foot (Bf) thrust system, in the internides of the Early Cretaceous orogen, was linked to the Snaring (Sn) and related thrust faults in its foreland (modified after McDonough and Simony, 1988). Note that the Purcell thrust (Pu) is out-of-sequence, and a normal faults daylights in the Southern Rocky Mountain Trench (SRMT). In the legend: 1 and 2 are basement rocks; 3 = metamorphosed Upper Proterozoic strata; 4 = Lower Paleozoic strata; 5 = upper Paleozoic strata; 6 = Mesozoic strata; and 7 = Mesozoic and Tertiary strata of the foreland basin. Thrust faults include: Bo = Boule; Mi = Miette; Gr = Greenock; Co = Colin; Ch = Chetamon; Py = Pyramid; Sn = Snaring; Mo = Monarch; Mp = Moose Pass; Cc = Chatter Creek; Pt = Ptarmigan; and Bf = Bear Foot. FH = Foothills; FR = Front Ranges; Mr = Main Ranges.

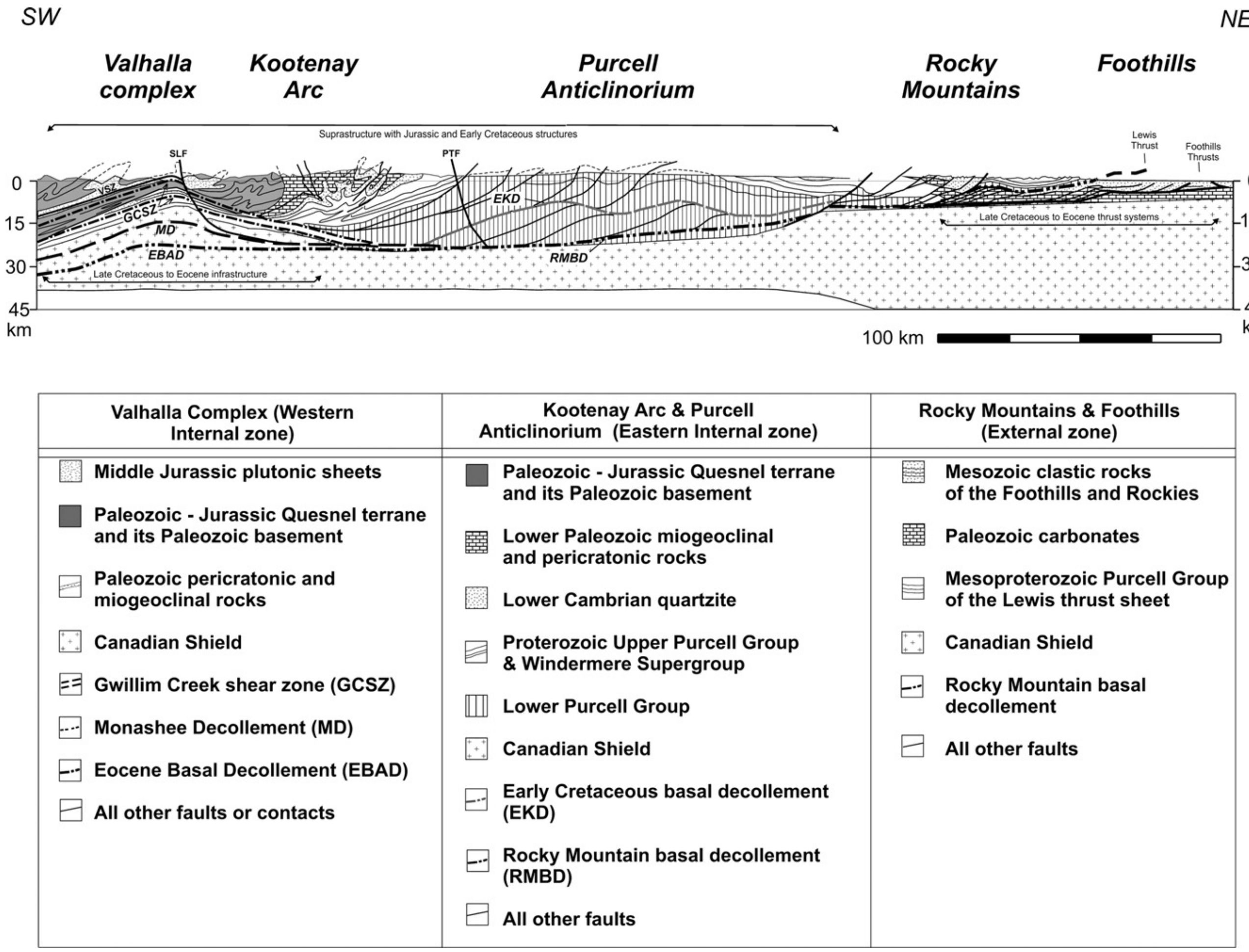

Fig. 4. West–east cross section through southern Valhalla complex (Fig. 1) showing the linkage between the Early Cretaceous to Eocene infrastructure in the west, located in and beneath the Valhalla complex, and the coeval thrust system in the External zone via the Rocky Mountain Basal décollement (RMBD) (after Carr and Simony, 2006). Shear zones in the Late Cretaceous to Eocene infrastructure are: the Eocene Rocky Mountain basal décollement (EBAD), Gwillim Creek Shear zone (GCSZ), and the Monashee décollement (MD). The suprastructure to the Late Cretaceous – Eocene orogen, lying west of the Valhalla complex and in the Kootenay Arc and Purcell anticlinorium, comprises rocks that were deformed in the Jurassic and Early Cretaceous carried by the Early Cretaceous décollement (EKD). Subsurface geometry from the Purcell Anticlinorium eastward is modified (Cook, personal communication, 2010) from Cook and Van der Velden (1995). Arching of Valhalla complex was caused by insertion of a proposed basement thrust sheet beneath the complex in the Early Eocene, carried by the Eocene basal décollement (EBAD).

and Van der Velden, 1995; Travis, 2007) along the top of what is shown in our section (Fig. 4) as a duplex structure.

The Selkirk–Monashee–Cariboo metamorphic complex (SMC, Fig. 1) is the exhumed portion of the metamorphic core of the Early Cretaceous orogen. It is ~500 km long and 2–100 km wide. The northeast margin of the complex is defined by regionally northeast-dipping garnet and biotite isogradic surfaces that cut across fold and thrust structures of the western Rocky Mountains (Gal et al., 1989). The metamorphism of the complex, particularly its eastern part, is of Early Cretaceous age with 140–125 Ma thermal peak metamorphism (Parrish, 1995, and references therein; Digel et al., 1998; Reid, 2003). Since the metamorphism was, in part, synchronous with deformation, the age of metamorphism also approximates, at in least part, the age of deformation. Biotite porphyroblasts that overgrew the metamorphic and mylonitic fabrics near the Ptarmigan décollement (Fig. 3) cooled at 110–100 Ma, based on K–Ar (Price and Mountjoy, 1970) and $^{40}Ar/^{39}Ar$ dates (Van den Driessche and Maluski, 1986). The décollement is therefore older. Regionally metamorphosed rocks and thrust faults in the Purcell Mountains were cross-cut by mid-Cretaceous (ca. 110–90 Ma) plutons (Archibald et al., 1983; Höy and van der Heyden, 1988; Wheeler and McFeely, 1991; Colpron et al., 1998). The western margin of the Early Cretaceous metamorphic complex, for example, north of the Frenchman Cap dome, is one of complex overprinting by Late Cretaceous to Paleocene metamorphism and deformation (Digel et al., 1998; Crowley et al., 2000; Gibson et al., 2008). Evidence for Early Cretaceous metamorphism or an Early Cretaceous basal shear zone has not been noted in or near the basement domes and core complexes with the exception of the Valhalla complex, where Early Cretaceous orthogneisses are present as strongly foliated screens within a weakly foliated mid-Cretaceous (~110 Ma) pluton (Fig. 6; Carr and Simony, 2006).

The Cadomin conglomerate occurs at the base of a thick Early Cretaceous foreland basin clastic unit in the Front Ranges and Foothills. It contains abundant pebbles from Lower Cambrian orthoquartzite and Carboniferous cherty carbonate formations typical of Early Cretaceous thrust sheets of the western Rocky Mountains to the west of the Monarch-Snaring thrust system (Schultheis and Mountjoy, 1978). This implies that there had been denudation of the Early Cretaceous thrust stack, to the level of the Lower Cambrian, in Early Cretaceous time. Ross et al. (2005) suggested that this stage of denudation may have reduced the orogenic

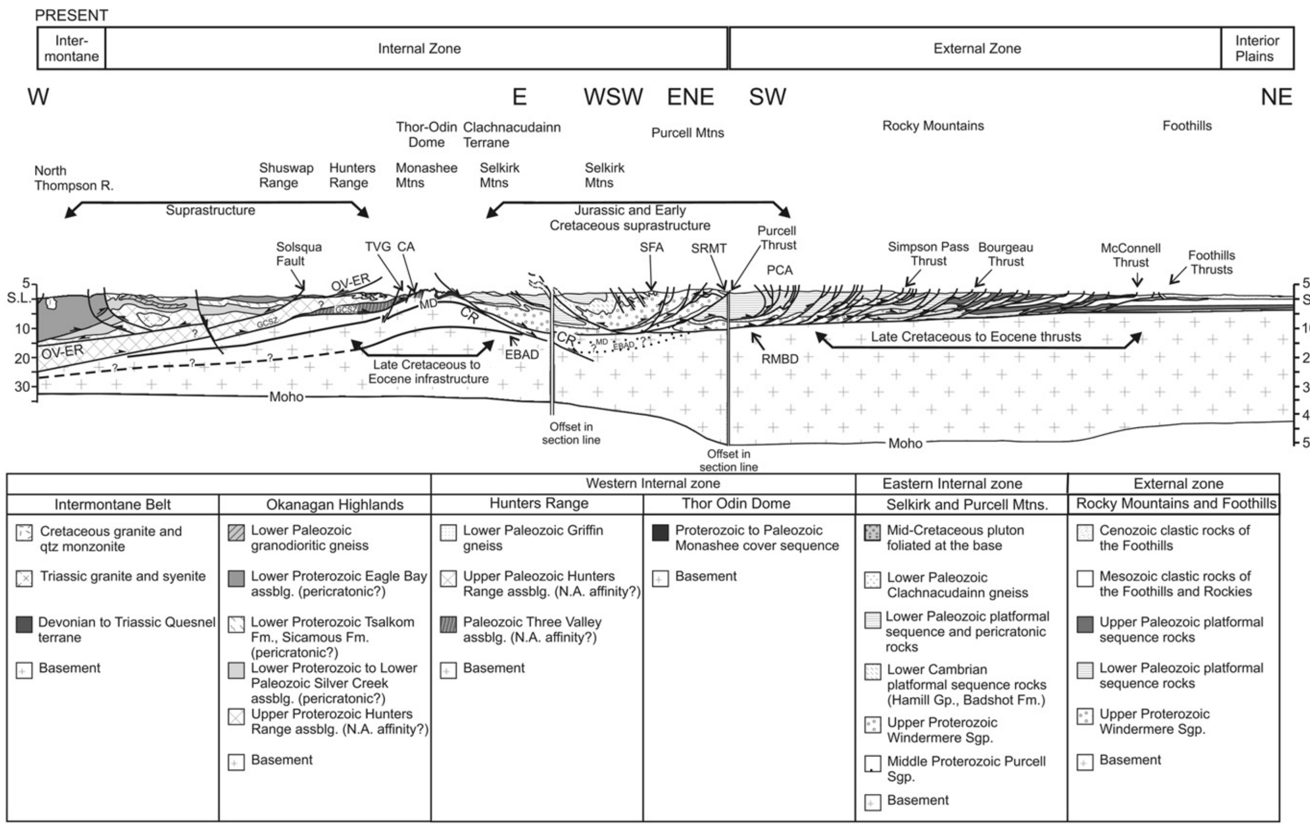

Fig. 5. Generally west–east oriented cross section through Thor-Odin dome (modified after Hinchey, 2005) showing the linkage between the Monashee décollement (MD) in the Internal zone with the McConnell thrust in the External zone via the Rocky Mountain Basal décollement (RMBD). CR = Columbia River normal fault; EBAD = Eocene basal décollement; OV-ER = Okanagan – Okanogan Valley – Eagle River normal fault system; PCA = Porcupine Creek Anticlinorium; SFA = Selkirk Fan Axis; and, SRMT = southern Rocky Mountain Trench.

wedge to below critical taper, and that thrusting did not resume until out-of-sequence thrusting and other mechanisms such as tectonic wedging (Price, 1986; Colpron et al., 1998) had increased the taper again.

The Early Cretaceous thrust belt and the eastern part of the Early Cretaceous metamorphic complex have apparently not been significantly deformed during the Late Cretaceous to Eocene, but rather were translated during younger orogenesis, and served to separate the Late Cretaceous to Eocene thrust belt from its metamorphic core zone, where the more westerly part of the Early Cretaceous metamorphic complex was largely overprinted.

4. Late Cretaceous to Eocene tectonic evolution of the southeastern Canadian Cordillera

In the following sections we describe timing constraints on the Late Cretaceous to Eocene structures of the Internal and External zones of the retrowedge, provide arguments in support of the existence of the Rocky Mountain basal décollement beneath the retrowedge and suggest direct linkages between specific structures in the core and thrust belt via the basal décollement.

4.1. External thrust belt and the Rocky Mountain basal décollement

The External zone, largely coinciding with the Late Cretaceous to Eocene Rocky Mountain thrust belt, can be regarded as an archetypical thrust and fold belt (Stockmal et al., 2007). An array of thrust faults that interleave and overlap along strike have the typical "thin-skinned" geometry, cutting across strata at a low to moderate angle, flattening with depth and repeating the same Cambrian to Triassic stratigraphy from thrust sheet to thrust sheet. These features indicate that the thrust faults merge into a common basal décollement, the Rocky Mountain basal décollement (Price, 1994). Most of the shortening was taken up by three or four major thrusts, and with their associated lesser faults they each constitute a system of interlinked faults. On a large scale, the relative ages of each of the constituent thrust systems have been broadly established, and they developed "in-sequence" from southwest to northeast.

4.1.1. The Late Cretaceous to Paleocene Lewis thrust system

The Lewis thrust is the most important thrust fault in the Rocky Mountains with >90 km of displacement in its central portion, and a strike length of >400 km. Near its tip lines, it is linked to thrusts that extend its strike length another 100 km in both directions. Thermochronometric analysis of stratigraphic, coalification, and fission track data (Osadetz et al., 2004; Feinstein et al., 2007) indicate that displacement of Mesoproterozoic strata of the Belt–Purcell Supergroup along the Lewis thrust fault was underway by ca. 75 Ma. In the southern part of the Lewis thrust sheet in Montana, faults at the leading edge cut through a 76 Ma (Campanian – Maastrichtian) volcanic marker and were cut by 59 Ma porphyry dikes (Sears, 2001; Schmidt, 1978). Motion on the Lewis thrust is thus bracketed between ca. 75 and 59 Ma. To the west, in the region north of the USA – Canada border, structures linked to the Hosmer–Wigwam–McDonald thrust system are indirectly linked to the Lewis thrust sheet (Price, 1994; Larson et al., 2006) and they deformed the Late Cretaceous (Turonian) strata within the Lewis thrust sheet. Motion on these faults could be as old as 85–80 Ma.

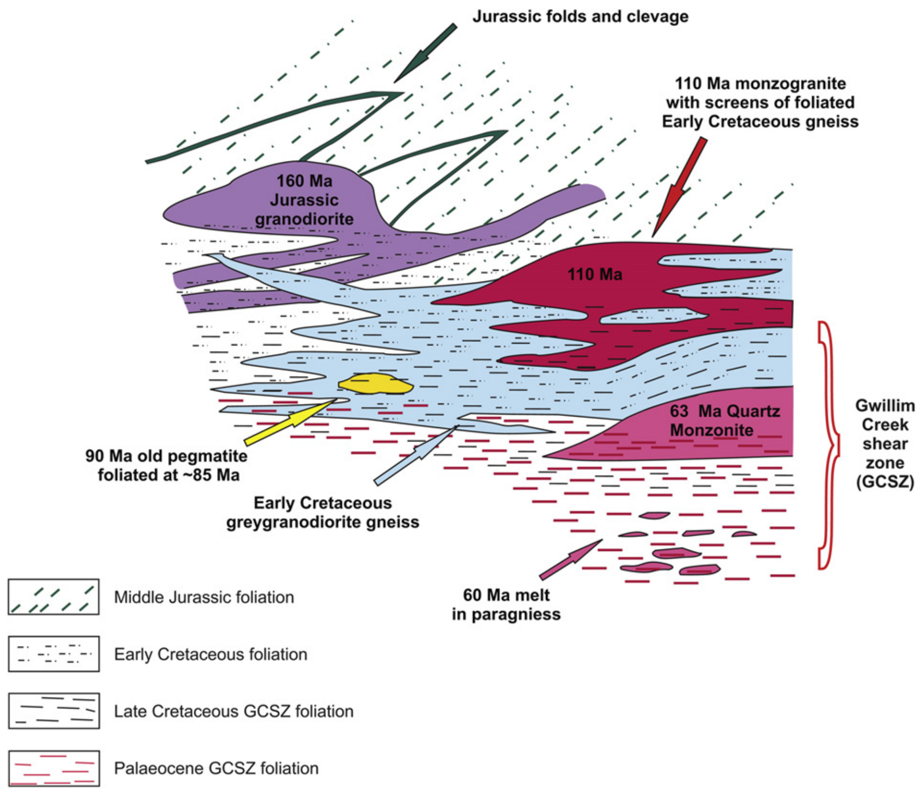

Fig. 6. Diagrammatic section depicts: (1) stitching relationships, based on cross-cutting relationships of variably deformed plutons, between Cretaceous to Paleocene infrastructure in the Valhalla complex and its overlying suprastructure; and (2) the "downward younging" overprinting relationships in Valhalla complex, described from oldest to youngest. Folds of Middle Jurassic age and associated cleavages were cut by Middle Jurassic (170–160 Ma) tonalite-granodiorite laccolithic plutons. The Early Cretaceous granodiorite-granite sheet complex, the "grey gneiss," cross-cut the feeder sill complex beneath the Middle Jurassic pluton. Both the Jurassic pluton and the "grey gneiss" have a pre-110 Ma foliation as they were cut by mid-Cretaceous (ca. 110 Ma) porphyritic monzogranite. The upper part of the mid-Cretaceous pluton is undeformed; however, the lower portion was deformed by the Gwillim Creek shear zone (GCSZ). A ca. 90 Ma pegmatite in the "grey gneiss" was recrystallized and strongly foliated at ∼85 Ma and places timing constraints on the upper, older portion of GCSZ. At the deepest structural levels exposed, leucocratic intrusions place timing constraints on the deepest and youngest part of the GCSZ. Paleocene (ca. 63 Ma) porphyritic quartz-monzonite, is undeformed in its top portion but foliated in its bottom portion. Ca. 60 Ma melt lenses in paragneiss cooled rapidly at ∼59–58 Ma in deepest exposed portion of GCSZ. (Carr and Simony, 2006).

The McDonald thrust continues northward as the Bourgeau thrust, the major western thrust of the Late Cretaceous – Eocene thrust belt (Price, 1994; Fig. 1). At the latitude of the Trans Canada Highway (Figs. 1 and 2), the Bourgeau, Sulphur Mountain and Rundle thrusts are "in-sequence" from west to east, although they are interlinked and their timing of motion overlapped (Price, 2001). They are, in turn, linked to the Lewis thrust (Bielenstein, 1969; Stockmal, 1979; McMechan, 1995). The entire interlinked Lewis–Bourgeau thrust system evolved during the interval from 85 to 59 Ma.

4.1.2. The Paleocene to Eocene McConnell thrust and Foothills structures

The McConnell thrust system, in the Front Ranges, is the next major thrust system east of the Lewis thrust, and as shown on the map, it is clearly separate from the Lewis thrust (Figs. 2 and 5). It extends ∼400 km along strike from the Jasper area (Figs. 1 and 3), where it is represented by the Miette thrust, to the region just north of Latitude 49°N. Shortening and thickening on the McConnell thrust system, amounting to ∼40–50 km in the Banff region, was accommodated by translation on the thrust as well as shortening on an imbricate system of splays of the McConnell thrust (Price and Mountjoy, 1970; Price and Fermor, 1985) and complicated duplexes that occur at several scales above the McConnell sole thrust (McMechan, 2001).

Timing on the McConnell thrust system is only approximately bracketed. The thrust fault cuts lithified clastic Campanian – Maastrichtian rocks in its footwall, and east of the fault trace, within the Foothills, lithified Paleocene clastic rocks are preserved in synclines (Ollerenshaw, 1978). In a cross section based on 1:50,000 scale geological maps, Price and Fermor (1985) show that a reasonable trajectory of the McConnell thrust, east of its map trace and above the present erosion level, projects above the Paleocene rocks, thus the fault is younger than Paleocene. There may be some overlap between late-stage motion on the Lewis thrust and thrusts of the McConnell system (Boyer, 1992; Fermor, 1999), but Fermor argues that motion on the McConnell thrust is largely younger than the bulk of the displacement on the Lewis thrust. A reasonable age bracket for the onset of motion for thrusts of the McConnell system is thus 63–60 Ma.

Precise age constraints are not available for the end of motion on the McConnell thrust. Attempts at direct dating are plagued with sampling, analytical and interpretation problems (c.f. van der Pluijm et al., 2006 and rebuttal by Price, 2007). Unfortunately, the stratigraphy of the foreland basin provides only broad age limits. We discuss three kinds of information from the Foothills belt that may provide constraints. The Foothills are the frontal belt of the orogen comprising recessive Mesozoic foreland basin clastic rocks that have been eroded to subdued ridge topography. Closely spaced thrusts repeat Cretaceous sandstone-shale sequences and merge down dip with a décollement at the top of the Paleozoic carbonates, which themselves are repeated by more widely spaced thrusts. There is no fundamental through-going structural boundary between the Front Ranges and the Foothills. The eastern limit of the Foothills belt is marked by the "triangle zone;" that is mainly a passive roof duplex where the Cretaceous clastic rocks that were carried by the most easterly thrusts were inserted as an intercutaneous wedge into Upper Cretaceous and Paleogene strata. East of the triangle zone at the latitude of Jasper, the lateral equivalents of strata that were deformed in the triangle zone include well-lithified clastic rocks of the Paskapoo Formation containing the Paleocene to Eocene boundary that has been dated at ca. 57.1 Ma (Lerbekmo et al., 2008). Vitrinite reflectance and coal rank from near that boundary indicates that there must have been an additional 2–2.5 km of overlying Eocene and younger strata (Kalkreuth and McMechan, 1996). This suggests that deformation in the eastern Foothills continued into the Eocene, to later than ca. 57 Ma.

We have argued earlier that, in the Eocene, local extension overlapped with compression in the Internal zone prior to ca. 52 Ma at which time extension of the full crustal section occurred, thus making younger thrusting unlikely. It is therefore reasonable to suggest that thrusting on the east edge of the Foothills was plausible until, perhaps, ca. 51 Ma, but that younger motion was unlikely. Stockmal et al. (1997) quoted apatite fission track data that suggest that thrusting could have continued to just before 49 Ma. A set of thrust faults in the eastern Foothills, merge down dip with the Rocky Mountain basal décollement east of where the McConnell thrust splays from it. Duplexes in the eastern Foothills fold the McConnell thrust and thrusts in the western Foothills that are linked to it (Ollerenshaw, 1978; Fermor, 1999; McMechan, 1995, 2001; Langenberg et al., 2002). Therefore, the main period of motion on the eastern Foothills thrusts is younger than the main period of motion on the McConnell thrust.

For the southern Lewis thrust system, Sears (2001) estimated displacement rates of between 3 and 9 km per million years, for the time interval between 75 and 59 Ma, using a range of magnitudes of displacement, which vary along strike. Assuming that that the Eastern Foothills thrusts may have moved at similar rates as Rocky Mountain thrusts (i.e. 5–10 km per million years), estimating the end of compression at ca. 51 Ma (discussion above) and using known magnitudes of displacement of ~30 km (Fermor, 1999), we roughly estimate that the onset of motion on the Eastern Foothills structures is bracketed between 57 (51 + 6) Ma and 54 (51 + 3) Ma. This simplistic procedure gives a crude estimate of the time of the end of motion on the McConnell thrust; however, this estimate is consistent with the reasonable relative ages of displacement and the available stratigraphic control.

In summary, timing controls on thrusting in the External zone are as follows. The Rocky Mountain basal décollement was first activated and propagated during motion of the Hosmer–Wigwam–McDonald–Lewis–Bourgeau–Rundle system in the Late Cretaceous to Early Paleocene (ca. 88–80 to 59 Ma). It continued to move or was reactivated during the Paleocene (ca. 65–60 to 57–54 Ma) during displacement on the McConnell thrust system, and finally, in the Eocene (ca. 57–51) during shortening in the triangle zone and Eastern Foothills.

4.1.3. *The westward continuation of the Rocky Mountain basal décollement*

Controls on stratigraphy and structure of the thin-skinned Rocky Mountain thrust belt permit palinspastic restoration such that the strata carried in each thrust sheet are restored to their pre-thrusting position above an incipient basal décollement. While there is some interpretation involved, it is a geometric exercise that yields approximate but reasonably reproducible results. Cross sections (Bally et al., 1966; Price and Mountjoy, 1970; Price, 2003; Price, 1981; Price and Fermor, 1985) yield internally consistent results that restore the present-day erosional leading edge of the Bourgeau thrust to the vicinity of the Frenchman Cap dome, Thor-Odin dome, Valhalla complex and Priest River complex (Price and Sears, 2000). Therefore, the Rocky Mountain basal décollement must continue westward below the Bourgeau thrust sheet, and below the nested Early Cretaceous orogen with its basal décollement, to a line ~100 km west of the restored position of the Bourgeau thrust estimated by Price and Sears (2000). Interpretations of seismic reflection data (Bally et al., 1966; Cook et al., 1992; Cook, 1995 and references therein; Cook and Van der Velden, 1995) support this geometric-stratigraphic interpretation. The deepest portions of the Frenchman Cap, Thor-Odin and Spokane domes contain Laurentian basement that is autochthonous or nearly so (see later discussion) and therefore the Rocky Mountain basal décollement must have extended westward well beyond the location of these complexes.

Previously, Brown et al. (1992); Johnson and Brown (1996); Doughty et al. (1998) and Carr and Simony (2006) proposed that major shear zones in the tectonothermal complexes had, at one time, been linked to the Rocky Mountain basal décollement. The geometry that we have described makes these proposed linkages plausible and we will discuss them below.

4.2. *Late Cretaceous to Paleocene deformation: the Gwillim Creek shear zone and linkages with the Lewis thrust system*

In our descriptions of the geology of the Internal zone, it is convenient to use the infrastructure – suprastructure terminology of Wegmann (1935) as used by Murphy (1987), Carr and Simony (2006) and Culshaw et al. (2006), recognizing that the timing of deformation of the infrastructure must be specified since this framework is relative to the deformation event under consideration.

4.2.1. *The Valhalla and Priest River complexes and the Gwillim Creek shear zone*

The Valhalla complex between latitude 49°10′ N and 50° N (Fig. 1) is a 20–30 km × 90 km fault bounded, doubly plunging, broad antiform with three distinct subculminations. Its Cretaceous to Paleocene infrastructure comprises migmatitic paragneiss intruded by sheeted and laccolithic granitoid orthogneiss plutons of Middle Jurassic, Early Cretaceous, Paleocene and Eocene age (Carr and Simony, 2006). The infrastructure of the complex was exhumed, at least in part, in the Eocene by an extensional shear zone and fault system (Carr et al., 1987; Parrish et al., 1988) that has two components. The east-verging, ductile Valkyr shear zone, which is arched over the complex forming its upper boundary, daylights in shear zones west of the complex (Parrish et al., 1988). The ductile to brittle, moderately east-dipping Slocan Lake – Champion Lake normal fault zone system bounds the eastern margin of the complex. This composite extensional fault system is hinged at its southern end near Latitude 49° N; the magnitude of displacement decreases from ~10–15 km in the central part of the culmination to 0.5–1.5 km at the south end (Simony and Carr, 1997). Thus, in the south, the pre-extensional relationships between the infrastructure in the Valhalla complex and the suprastructure above the extensional fault systems are preserved, exposing a >30 km thick continuous crustal section.

In the upper part of the crustal section, the suprastructure comprises Early Jurassic arc volcanic and volcaniclastic rocks, Late Paleozoic clastic sedimentary rocks and the unconformably underlying Trail gneiss, the Devonian basement of the Quesnel Terrane (Simony et al., 2006), and preserves pre-Middle Jurassic deformation and metamorphism. In the upper part of the section, the Jurassic and Late Paleozoic chlorite- to biotite-grade rocks were folded into north trending, km-scale Jurassic folds with axial planar slaty cleavage. Downward, through an interval of 1–2 km, the rocks were caught up in younger infrastructural deformation and metamorphism. The grade increases to garnet and andalusite, folds are tighter and minor folds are transposed into the WSW plunging mineral lineation. Similarly, the Middle Jurassic plutons, unfoliated in their upper level, become increasingly foliated toward their base with a strong WSW plunging mineral lineation. Metamorphic grade increases downwards: at the deepest levels of the infrastructure, in the southern part of the complex, the rocks are at sillimanite + muscovite grade, and in the rest of the complex they are at sillimanite + K feldspar + melt grade. In these high-grade rocks, cross-cutting relationships between Early Cretaceous, mid-Cretaceous (ca. 100 Ma), Late Cretaceous and Paleocene igneous sheets and foliations bracket stages of deformation in the infrastructure (Fig. 6), including Early Cretaceous deformation (described in Section 3.2) and Late Cretaceous to Paleocene stages of deformation interpreted to have formed above and within the Gwillim Creek shear zone, which lies at the base of the exposed section (Carr and Simony, 2006), and will be described below. As illustrated in Fig. 6, deformation was progressively younger downward through the structural section, and higher older structures and fabrics were deformed and overprinted as a consequence of younger deformation in the underlying levels. These structural relationships are, in essence, similar to structural geometries such as folded thrusts in thrust belts that are used to argue for "in-sequence" thrusting. The overprinting relationships cannot be explained merely by cooling and blocking of thermochronometers as a consequence of progressive exhumation, which would not produce the structural overprints. The dated plutonic rocks not only serve as time markers for progressive stages of deformation, but they also effectively stitch the suprastructure to the Late Cretaceous to Paleocene infrastructure. Therefore, all of these rocks constitute one coherent, though internally deformed, thrust sheet on the order of 30 km thick with the Late Cretaceous to Paleocene ductile Gwillim Creek shear zone at its base (Carr and Simony, 2006).

4.2.2. The Gwillim Creek shear zone in the Valhalla complex

The Late Cretaceous–Paleocene Gwillim Creek shear zone is a top-to-the-east ductile thrust-sense shear zone that was first recognized in the Valhalla complex where it occurs at the base of an ~3 km thick, downward increasing strain gradient in all of the three subsidiary culminations in the complex (Carr and Simony, 2006, and references therein). The shear zone is estimated to be 5–7 km thick on the basis of the exposed thickness of strained rocks and interpretation of seismic reflection data (Eaton and Cook, 1990) beneath the central culmination. Kinematic indicators on the mesoscopic and microscopic scale all indicate top-to-the-east shear (Schaubs et al., 2002, and references therein).

The Gwillim Creek shear zone was activated in the Late Cretaceous at ~85 Ma and was reactivated ~65–60 Ma (Carr and Simony, 2006) based on relationships between dated plutons and pegmatites, and sheared rocks. The timing of the youngest stage of displacement across the Gwillim Creek shear zone is constrained to the interval between ~65 Ma and ~60 Ma based on the ages of syn-tectonic leucosome (Gordon et al., 2008), deformed pegmatite, deformed yet cross-cutting ~63 Ma plutonic rocks (Carr and Simony, 2006) and syn-tectonic monazite (Heaman and Parrish, 1991; Parrish, 1995; Spear and Parrish, 1996; Schaubs et al., 2002; Spear, 2004). At the deepest exposed levels of the shear zone, flow took place at ~800 °C and 800 MPa and was accompanied by anatexis (Schaubs et al., 2002; Gordon et al., 2008). In the Paleocene, the ductile sheet was transported up and over a low-angle, 10–12 km high frontal ramp, and its base was cooled from below on the upper "flat" east of the ramp (Schaubs et al., 2002; Spear, 2004). The ramp interpretation is based on petrological evidence that the deepest exposed structural levels in the Gwillim Creek shear zone were quenched prior to quenching of overlying rocks in the hanging wall of the shear zone (Spear and Parrish, 1996; Schaubs et al., 2002). The ramp geometry of the Gwillim Creek shear zone is supported by geologic relationships showing that it cut upward in the eastward direction of transport, as well as southward, relative to layers in the plutonic and structural edifice (Carr and Simony, 2006). This ramp geometry is consistent with a minimum displacement of ~40 km (Carr and Simony, 2006) and is consistent with evaluation of thermal models by Spear and Parrish (1996) and Spear (2004).

4.2.3. The Spokane dome and the Gwillim Creek shear zone

The Priest River complex (Fig. 1) lies in the Internal zone, in the Selkirk mountains, in a tectonic setting more external than that of the Valhalla complex. It was exhumed in a complex manner between four ductile and brittle extensional fault systems in the Eocene (Doughty and Price, 1999). The Spokane dome mylonite zone is arched over the Spokane dome in the deep, southern core of the complex. The mylonite zone is ~4 km thick with a top-to-the-east shear sense. It has a 500–1000 m thick upper transition zone and a distinct ca. 100 m thick basal zone where the foliation of underlying rocks was transposed into the mylonitic foliation in a zone of intense folding and foliation. The outcrop of the basal zone outlines the Spirit Lake window just south of Latitude 48° N. The bulk of the rocks within the mylonite zone and below it are interpreted to be the high-grade metamorphic equivalent of the Pritchard Formation at the base of the Belt Supergroup (Rhodes and Hyndman, 1984; Doughty and Price, 1999; Doughty and Chamberlain, 2008). Archean gneiss, with its stratigraphic cover, is considered autochthonous basement by Doughty et al. (1998), and lies near the base of the mylonite zone in a small antiformal culmination north of the Spirit Lake window.

The Spokane dome mylonite zone mainly records Early Eocene extension (Doughty and Price, 1999) but in the deeper parts of the mylonite zone, monazite and xenotime within the metamorphic and structural fabrics indicate peak metamorphism during deformation at ca. 72 Ma (Doughty et al., 1998), and U–Pb data from zircon record Mesoproterzoic and Late Cretaceous to Paleocene crystallization (Doughty and Chamberlain, 2008). Given these ages, Doughty et al. (1998) link the deeper part of the Spokane dome mylonite to the Gwillim Creek shear zone. Given the presence of the Archean basement, Doughty et al. (1998) also link the deeper portion of the mylonite zone to the Rocky Mountain basal décollement. Our west–east cross section at the latitude of the Valhalla complex (Fig. 4) is consistent with these proposed linkages in the Spokane dome. The geological relationships in the Spokane dome, when projected northward along the structural trend, project into our cross section roughly where the Gwillim Creek shear zone is shown merging with the Rocky Mountain basal décollement within the base of the Belt-Purcell Supergroup; thus interpretations from the two different regions are consistent.

4.2.4. The Gwillim Creek shear zone to the west of the Valhalla complex

The Gwillim Creek shear zone can be traced to the north and south of the Valhalla complex on the basis of geological mapping and reflection seismic surveys (Cook et al., 1992; Cook, 1995; Carr and Simony, 2006; Travis, 2007). It projects into the subsurface to the west of the Valhalla complex, presumably lying at the base of the rocks of the Quesnel terrane or deeper. It could lie within gneisses of the Grand Forks – Kettle River complex (Fig. 1) where U–Pb ages of monazite suggest an episode of Late Cretaceous metamorphism and deformation at 84 ± 3 Ma (Laberge and Pattison, 2007). COCORP reflection seismic data from a transect located south of the Canada – USA border indicate a zone (Zone E) of coherent west-dipping reflectors at a depth of ~20 km under and east of the Okanogan dome (Fig. 1; Sandford et al., 1988). That zone of reflectors could represent a shear zone like the Gwillim Creek shear zone.

With respect to the continuation of the Gwillim Creek shear zone to the west and northwest of the Valhalla complex, we suggest that the trace reemerges on the southern flank of the Thor-Odin dome and projects northward west of the western flanks of the Thor-Odin and Frenchman Cap domes (Fig. 1). Late Cretaceous and/or Paleocene metamorphism (Carr, 1995; Parrish, 1995; Kuiper, 2004; Glombick et al., 2006; Lemieux, 2006) and deformation, including a strong transposition foliation and northeast verging folds have been documented on the southern and western flanks of the Thor-Odin dome (Carr, 1991, 1992, 1995; Parrish, 1995; McNicoll and Brown, 1995; Johnston et al., 2000; Kuiper, 2004), and Late Cretaceous ages of metamorphism and deformation (Scammell, 1993; Parrish, 1995; Gibson et al., 1999; Crowley et al., 2000) occur on the western flank of the Frenchman Cap dome. In this proposed model, the rocks exposed to the south of the Thor-Odin dome and west of the Thor-Odin and Frenchman Cap domes could be structurally equivalent to the rocks lying above the Gwillim Creek shear zone in Valhalla complex. The trace of the Gwillim Creek shear zone has not been separated out in the mapping of these regions; rocks in the proposed Gwillim Creek shear zone and its hanging wall have been, in part, strongly overprinted and transposed by younger deformation (Glombick, 2005; Williams and Jiang, 2005; Hinchey et al., 2006, and references therein).

4.2.5. *Linkages of the Gwillim Creek shear zone with foreland thrust faults via the Rocky Mountain basal décollement*

On the east side of the Valhalla complex, the Gwillim Creek shear zone dips moderately to the east. It is likely that the Valkyr shear zone merges with it down dip, and the east-dipping Slocan Lake normal fault cuts it and could, in part, have merged with it. East of the complex, the Gwillim Creek shear zone must lie at depth under rocks of the Quesnel terrane and Kootenay Arc (Fig. 4). This is where the Late Cretaceous to Eocene Rocky Mountain basal décollement lies, and the two presumably merge. The Late Cretaceous to Paleocene Gwillim Creek shear zone cannot emerge in the Selkirk and Purcell mountains where thrusting, folding and metamorphism are pre-mid-Cretaceous as established by the ca. 110–90 Ma K–Ar dates of cross-cutting plutons and metamorphic cooling (Archibald et al., 1983; Höy and van der Heyden, 1988). Thus Carr and Simony (2006) concluded that the Gwillim Creek shear zone extends beneath the Selkirk and Purcell mountains and lies within the Rocky Mountain basal décollement. To the east of the Purcell Mountains, in the Western Rocky Mountains, thrusting and folding were also pre-mid-Cretaceous, on the basis of ca. 109 Ma cross-cutting plutons (Larson et al., 2006), so the Gwillim Creek shear zone cannot emerge there. The most westerly thrusts that could be connected to the Late Cretaceous Gwillim Creek shear zone are those of the Hosmer–Wigwam–McDonald–Bourgeau thrust system that daylight west of and were indirectly linked with the Lewis thrust system. The early ~85–75 Ma motion on the Gwillim Creek shear zone may have been largely taken up by this composite thrust system. Subsequent motion on the Gwillim Creek shear zone, between ~75 and ~60 Ma, would have been largely taken up by the Lewis thrust.

In the Late Cretaceous, the Rocky Mountain basal décollement extended westward under the Purcell Mountains (see section 4.1), the most plausible link between Late Cretaceous motion on the Gwillim Creek shear zone and the Hosmer–Wigwam–McDonald–Bourgeau system was via the Rocky Mountain basal décollement. Therefore, in the Late Cretaceous, and until about 60 Ma, the Gwillim Creek shear zone can be considered as part of the Rocky Mountain basal décollement. Motion on both the Gwillim Creek shear zone and the Lewis thrust ended at ca. 59 Ma. Although linking the Gwillim Creek shear zone, the Rocky Mountain basal décollement and the Bourgeau-Lewis system may seem speculative, reasonable palinspastic restoration, the structure of the Purcell Mountains and timing constraints on deformation leave little choice.

In Section 4.1.1 we presented geometric arguments that, in the Banff region at the latitude of the Trans Canada Highway transect, the Bourgeau, Sulphur Mountain and Rundle thrust faults formed as an interlinked system (Price, 2001) that was linked to the Lewis thrust. These thrusts passed down dip into the Late Cretaceous to Paleocene Rocky Mountain basal décollement that lies beneath the Purcell and Selkirk mountains and beneath the Early Cretaceous orogen (Price and Fermor, 1985; Colpron et al., 1998). These thrusts

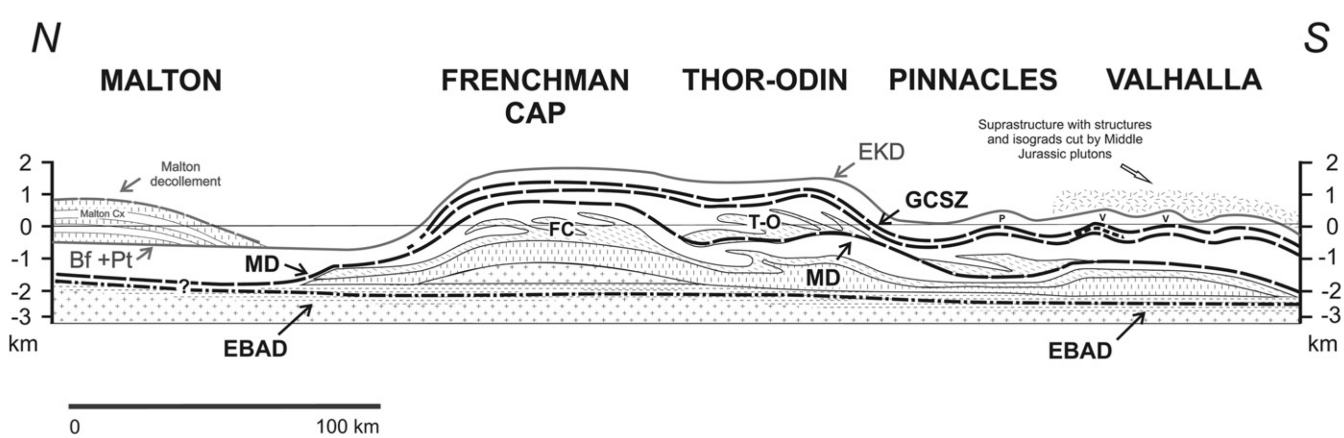

Fig. 7. Schematic north–south cross section through the Western Internal zone of the southeastern Canadian Cordillera illustrating the relative structural relationships between tectonothermal culminations and their bounding shear zones. The symbols within the basement are meant to depict relative strain. Upright plus signs indicate relatively undeformed basement, whereas vertical and diagonal patterns represent penetrative deformation and horizontal patterns and heavy lines indicate high strain. The structurally highest complex is the Malton complex. It comprises basement sheets that were internally imbricated and stacked in the Jurassic and roofed by the Jurassic Malton décollement. The Malton complex was translated by the Early Cretaceous basal décollement (EKD) comprising the Bear Foot thrust (Bf) and Ptarmigan décollement (Pt) (see Fig. 3). The Late Cretaceous to Paleocene Gwillim Creek shear zone (GCSZ) carried Valhalla complex (V) and the Pinnacles culmination (P). The Late Cretaceous to Eocene Monashee décollement (MD) bounds the upper margin of the basement-cored Frenchman Cap dome (FC) and, in this possible interpretation (see text for discussion) carried the basement-cored Thor-Odin dome (TO). EBAD = proposed subsurface Eocene basal décollement.

therefore were most probably linked, via the Rocky Mountain basal décollement, to the northern segment of Gwillim Creek shear zone.

4.3. Paleocene to Eocene deformation in the Internal zone and linkages with the thrust belt in the eastern Rocky Mountains and Foothills

4.3.1. The Frenchman Cap and Thor-Odin domes of the Monashee complex, and the Monashee décollement

Two basement-cored, domal culminations, the Frenchman Cap dome in the north and the Thor-Odin dome in the south form an $\sim 140 \times 30$ km, north to northwest trending composite culmination, termed the Monashee complex (Fig. 1). It is surrounded in map view and structurally overlain by deformed and metamorphosed Neoproterozoic and Paleozoic assemblages. It is bounded on the eastern margin by the moderately dipping, ductile–brittle, Eocene Columbia River normal fault system, and the northern and western upper margins of the Frenchman Cap dome are bounded by the Monashee décollement, which is arched over the dome and cut by the Columbia River fault. At its southern margin, the Frenchman Cap dome plunges SSE at 20°–25° beneath the northern flank of the Thor-Odin dome where the plunge is near horizontal (Hill, 1975; Read and Klepacki, 1981). Laurentian (western Canadian) basement rocks in the domes have ages in the ca. 2.3–1.8 Ga bracket (Armstrong et al., 1991; Parkinson, 1992; Crowley, 1999) and are structurally or unconformably overlain by supracrustal rocks, including marker quartzites and some rocks that may be as young as Cambrian to Devonian (Crowley, 1997; Scammell and Brown, 1990; Scammell and Parrish, 1993; Hinchey, 2005, and references therein).

A 6.5 km thick structural section is exposed within the Frenchman Cap dome beneath the Monashee décollement (Crowley and Parrish, 1999; Crowley et al., 2001, 2008; Gervais et al., 2010, and references therein). Paleoproterozoic basement in the basal 1000–2000 m has little or no Cordilleran deformation. At the top of this zone is a 600 m thick basal transition zone from undeformed basement into an overlying ~ 2 km thick "upper basement zone" which was strongly transposed, sheared and metamorphosed between ca. 54 and 49 Ma (see section 4.3.3). This upper basement zone is structurally overlain by an ~ 4.5 km thick zone of strongly deformed supracrustal cover with regional-scale fold nappes and top-to-the-NE sense of shear. Within this ~ 4.5 km thick zone, the lower 1.5 km contains structures and metamorphic assemblages that are as young as ca. 50 Ma (see section 4.3.3), whereas the upper ~ 3 km underwent deformation and metamorphism between ca. 60 and 55 Ma (Gervais et al., 2010, and references therein) related to the Monashee décollement (see below). In the upper zone, Gibson et al. (1999) report monazite as old as ca. 78 Ma in a high-pressure assemblage (Gervais et al., 2010) which may be related to Gwillim Creek shear zone strain as proposed in Section 4.2.4.

In contrast to the core rocks of the Frenchman Cap dome where undeformed basement is exposed, rocks throughout all exposed structural levels of the Thor-Odin dome apparently underwent Eocene penetrative deformation and anatexis (Hinchey et al., 2006, and references therein); this, and the axial plunge geometry of the domes (i.e. the Frenchman Cap dome apparently plunges beneath the Frenchman Cap dome), suggest that the Frenchman Cap dome represents a deeper level of the Cordilleran edifice than the Thor-Odin dome. Basement and supracrustal cover rocks in the Thor-Odin dome were infolded by northeast verging, tight to isoclinal folds with axial traces that extend for 60 km along strike (Reesor and Moore, 1971; Williams and Jiang, 2005); transposition, folding and anatexis were ongoing at ca. 56 Ma and had ended by ca. 54 Ma (Hinchey et al., 2006, and references therein). These Eocene structures were arched over the dome (Read and Klepacki, 1981; Williams and Jiang, 2005) and are interpreted to have formed during compression that began in the Late Cretaceous (Williams and Jiang, 2005; Hinchey et al., 2006; Paul McNeill, pers comm., 2005).

The upper boundary of the Frenchman Cap dome is interpreted as the Monashee décollement, an east-vergent, ductile 300–1000 m thick shear zone that is arched over the dome (Brown et al., 1986; Gervais et al., 2010, and references therein). It was defined on the northwest flank of the dome where the ca. 62–55 Ma timing of ENE -verging shear is well constrained by numerous zircon and monazite U–Pb dating studies carried out in conjunction with structural analysis (Parrish, 1995; Crowley and Parrish, 1999; Gibson et al., 1999; Crowley et al., 2001; Foster et al., 2004; Gervais et al., 2010, and references therein). Note that the timing of the Monashee décollement overlaps with the age of penetrative deformation in the Thor-Odin dome (Hinchey et al., 2006). In the Frenchman Cap dome, the Monashee décollement truncated folds in its footwall, juxtaposed older over younger rocks, and rocks in the hanging wall (see Section 4.2.4) and footwall have disparate P–T histories and structural style (Gibson et al., 1999; Gervais et al., 2010, and references therein). Brown and Gibson (2006) and Gervais and Brown (2011) interpret the Monashee décollement as the base of a channel.

The Monashee décollement was originally interpreted to extend southward and bound the upper boundary of the Thor-Odin dome as well as the Frenchman Cap dome (Brown et al., 1986, 1992; McNicoll and Brown, 1995; Gervais et al., 2010, and references therein), in part, based on the paradigm that rocks in the domes, representing undeformed basement of the craton, were overthrust by high-grade polydeformed metamorphic rocks with a Mesozoic Cordilleran tectonothermal history (Brown et al., 1986); however, this interpretation has not held up for the Thor-Odin dome on the basis of subsequent structural and geochronology studies (Williams and Jiang, 2005; Hinchey, 2005). Rather, Williams and Jiang (2005) concluded that the deformation history and geometry of the transposition foliation, intrafolial folds and superimposed polyphase folds was similar throughout both the Thor-Odin dome and overlying gneisses, and therefore did not support the interpretation of a lithological or structural dislocation or shear zone such as the Monashee décollement proposed by McNicoll and Brown (1995) for the western and southern flanks of the dome. The rocks in the Thor-Odin dome and those structurally overlying the dome are considered to be part of the same structural panel representing a zone of mid-crustal flow (Williams and Jiang, 2005; Williams et al., 2006). This is supported by dating studies showing that deformation at the deepest exposed structural level of the dome was ongoing at ca. 56 Ma and ended at ca. 54 Ma, based on the U–Pb SHRIMP ages of zircon from an isoclinaly folded and cross-cutting leucosome (Hinchey et al., 2006), and outlasted that of structurally overlying rocks which were cross-cut by 58 Ma pegmatites (Carr, 1992). It requires further mapping and research to resolve the differences between interpretations, as outlined above, of the presence of the Monashee décollement on the upper margin of the Frenchman Cap dome, and its apparent absence in the Thor-Odin dome. However, despite the discrepancies in interpretations, researchers agree about the structural style and timing of deformation in the domes. The important information necessary for our analysis is that ca. 62–54 Ma penetrative deformation and shearing in the Monashee décollement of the Frenchman Cap dome and throughout the core of the Thor-Odin dome are younger and structurally deeper in the orogen than that of the Valhalla complex and the Gwillim Creek shear zone, therefore we follow the regional plunge and project the Monashee décollement from the Frenchman Cap dome, through the lower structural levels of the Thor-Odin dome and southeastward into the subsurface beneath the Valhalla complex, and structurally below the Gwillim Creek shear zone (Figs. 4 and 7). Undeformed basement lies below the ductile and complexly deformed rocks of the Frenchman Cap dome. Continuation of such undeformed basement south and southeastward under the Thor-Odin, Valhalla and Spokane domes, as shown in Figs. 4 and 7, is consistent with the southward regional plunge.

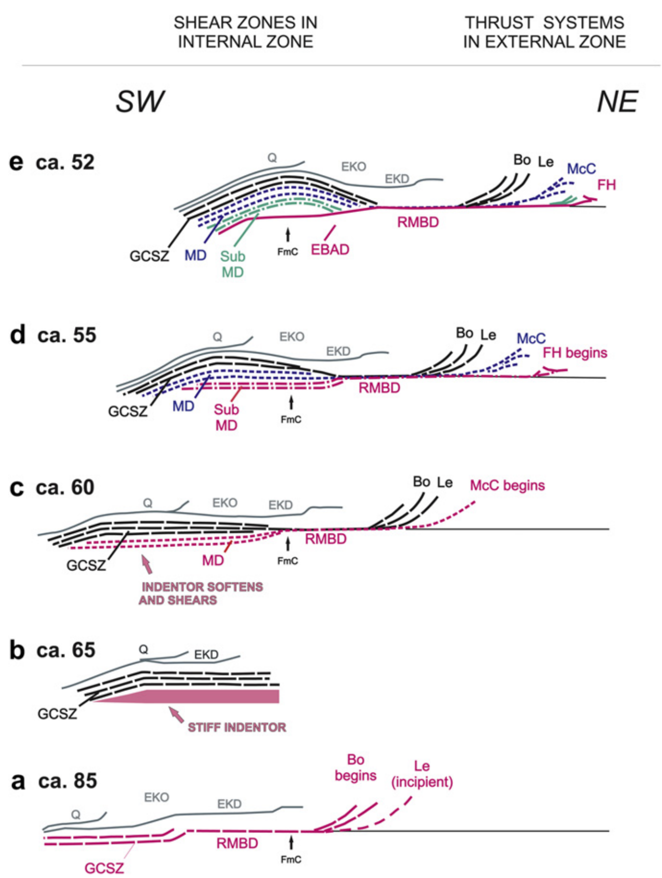

Fig. 8. Schematic cross sections, at successive stages, in the Late Cretaceous to Eocene development of the retrowedge illustrate a model of downward younging of shear zones in the Internal zone (EBAD = Eocene basal décollement; GCSZ = Gwillim Creek shear zone; MD = Monashee décollement) and continuity of these shear zones with the Rocky Mountain basal décollement (RMBD) and coeval thrust fault systems of the Front Ranges and Foothills in the External zone (Bo = Bourgeau; FH = Foothills; Le = Lewis; McC = McConnell). Red-colored shear zones and faults depict the active structures at the stage under consideration. In the Internal zone, the downward migration of zones of successively younger shear strain occurred in response to westward underthrusting of a stiff crustal indentor, stage by stage, with episodes of heating, weakening and localization of shear zones within the upper part of the indentor. At each stage, as the indentor softened and deformed, the older, higher shear zone strain hardened. The reference arrow FmC is a reference marker showing the incipient location (8a–d) or location (8e) of the Frenchman Cap dome. EKD = Early Cretaceous basal décollement; EKO = Early Cretaceous orogen and Q = Quesnel terrane, all translated on deeper younger shear zone systems. (8a) At ca. 85 Ma, rocks of the Gwillim Creek shear zone (GCSZ) were soft and weak due to metamorphism and anatexis. Transport flow had begun in the GCSZ, and the Rocky Mountain basal décollement (RMBD) and Bourgeau thrust (Bo) system were activated. (8b) By 65 Ma, a stiff indentor caused a deflection (ramp) in the GCSZ, and the GCSZ and Lewis thrust (Le) were in motion. (8c) At ca. 60 Ma, the GCSZ was strain hardening and stiffening, and motion on the Lewis thrust was ending. Subsequently, the indentor softened, the Monashee décollement formed and the McConnell system began to move. (8d) At ca. 55 Ma, the Monashee décollement had strain hardened above an indentor, strain migrated downward and a shear zone below the Monashee décollement was activated, as seen in the Frenchman Cap dome. Motion on the McConnell thrust system was ending and the thrusts of the Eastern Foothills began to move. (8e). At ca. 52 Ma, during the last stages of compression, a shear zone developed in the stiff basement (EBAD) and was linked via the RMBD to the youngest structures in the Foothills. Motion on the Eocene basal décollement (EBAD) contributed to doming of the Valhalla complex, Thor-Odin dome and Frenchman Cap dome.

4.3.2. Links between the Monashee décollement and the McConnell thrust

On the eastern margin of the Frenchman Cap dome, the Monashee décollement was down thrown ~20 km in the hanging wall of the Columbia River normal fault, it passes at depth under the Jurassic and Early Cretaceous structures of the Selkirk, Purcell and western Rocky mountains (Brown et al., 1986), where the Rocky Mountain basal décollement lies (Colpron et al., 1998) and the two are most reasonably linked (Brown et al., 1992; Johnson and Brown, 1996; Cook, 1995). If it is reasonable that motion on the Gwillim Creek shear zone was linked, via the Late Cretaceous – Early Paleocene Rocky Mountain Basal décollement, to motion on the Hosmer–Wigwam–McDonald–Bourgeau and Lewis–Rundle thrust systems of the Rocky Mountains (Carr and Simony, 2006; Section 4.2.5), then perhaps in a similar manner, motion on the Monashee décollement was linked, via the Paleocene Rocky Mountain Basal décollement, to motion on thrust systems that lie below and to the east of the Lewis thrust, that is the McConnell thrust system (Figs. 2 and 5), with thrust motion in the ca. 63–54 Ma time interval.

4.3.3. Linkages of Eocene Internal zone deformation and Foothills structures

In the Frenchman Cap dome, at a structural level beneath the Monashee décollement and immediately above the undeformed Proterozoic basement in the core, an ~3.5 km thick zone of highly deformed basement and cover rocks, with top-to-the-east shear zones, represents the deepest and youngest Cordilleran deformation exposed in the Cordillera (Crowley, 1999; Gibson et al., 1999; Crowley et al., 2001, 2008; Gervais et al., 2010) with deformation ending at ca. 54–49 Ma at the lowest level (Gervais et al., 2010). If during the interval between ca. 62–54 Ma, the Monashee décollement was the western continuation of the Paleocene Rocky Mountain Basal décollement and, in this manner, was linked via the Rocky Mountain Basal décollement to the McConnell thrust system, then this underlying deepest zone of deformation in the Frenchman Cap dome may all or in its deeper part, be linked, via the Eocene Rocky Mountain Basal décollement to structures of the same age in the Eastern Foothills (Sub-MD of Fig. 2).

In addition, we propose the existence of an Early Eocene Basal décollement in the subsurface, beneath the Internal zone of the orogen (EBAD, Figs. 4 and 5), that linked with the Eocene Rocky Mountain basal décollement. A consequence of such a décollement is that transport or flow of rocks up ramps carried by such a décollement provides a mechanism and explanation for late doming of tectonothermal culminations such as the Frenchman Cap dome, as proposed by Carr and Simony (2006) for the Valhalla complex. The west–east antiformal shape of the dome would have formed in response to movement up a frontal ramp, whereas the northern and southern plunges on the northern and southern flanks of the dome, respectively, would have formed in response to movement of a basement slice up and over the frontal ramp with consequent inversion of the originally inward-dipping lateral ramps bounding the sides. Note that shear zones under Frenchman Cap dome have been proposed by Brown et al. (1986, 1992), and that there are reflection horizons below the ones interpreted as the Gwillim Creek shear zone and the Monashee décollement on LITHOPROBE geophysical transects (Cook et al., 1992).

5. Conclusions and discussion

On the retrowedge side of the southeastern Canadian Cordillera, crystalline rocks of the Internal zone are situated at some distance behind the Late Cretaceous to Eocene east-verging foreland thrust belt in the External zone. No megathrust sheet or crystalline sheet with a basal shear zone overrode the External foreland thrust belt, thus the geometric and kinematic linkages between the Internal zone and External thrust belt can be reconstructed.

Three Mesozoic to Eocene orogenic events are recognized in the southeast Cordillera: Early to Middle Jurassic; Early Cretaceous; and, Late Cretaceous to Eocene. The Early to Middle Jurassic small cold accretionary orogen is mainly preserved in the suprastructure of the Internal zone of the orogen. The >200 km wide Early Cretaceous orogen is nested within the Late Cretaceous to Eocene orogen. In its core it contains the Malton complex which was formed by Jurassic imbrication and stacking of basement sheets and Early Cretaceous translation on a basal décollement. It serves as an example of a basement-cored culmination that is not an Eocene metamorphic core complex.

The wide warm Late Cretaceous to Eocene orogen has: (i) frontal east-verging thin-skinned thrust and fold systems (e.g. Lewis–Bourgeau, McConnell and Foothills) that root into a basal décollement that passes beneath the aforementioned Early Cretaceous orogen; and, (ii) tracts of metamorphic rocks, metamorphic core complexes and basement-cored domes in the Internal zone that have a structurally downward-younging progression of Late Cretaceous to Eocene deformation and metamorphism with structural overprinting that cannot be explained merely by cooling and blocking of thermochronometers as a consequence of progressive exhumation.

We propose a model of geometrical progression for the Late Cretaceous to Eocene development of the retrowedge. Structures and shear zones in the Internal zone are linked with coeval thrust systems in the External zone via the central portion of the Rocky Mountain basal décollement (Fig. 8), first activated in the Late Cretaceous and then reactivated in the Late Cretaceous to Paleocene and again in the Eocene. Between ca. 85 Ma and 59 Ma, early motion on the Gwillim Creek shear zone was linked to the Hosmer–Wigwam–McDonald–Bourgeau thrust systems (Fig. 8a and b), and younger Late Cretaceous to Paleocene motion is linked to the Lewis–Rundle thrust systems (Fig. 8b and c); the ca. 62–54 Ma Monashee décollement is linked to the McConnell thrust system (Fig. 8c and d); the ca. 53–49 Ma deformation below the Monashee décollement in the Frenchman Cap dome is linked to the thrusts in the eastern Foothills (Fig. 8e); and a possible subsurface Eocene basal décollement, proposed on the basis of tectonic arguments and reflective horizons in geophysical reflection data, is linked to the youngest thrust systems in the eastern edge of the Foothills (Fig. 8e).

The Early Cretaceous orogen and the rocks above the Late Cretaceous Gwillim Creek shear zone would have served to insulate the underlying rocks of the incipient Internal zone of the orogen, thus resulting in a mechanism of progressive heating, weakening and localization of basal shear zones. The craton was progressively underthrusting the developing retrowedge. At the base of the wedge, cooler stiffer rocks lay to the east of the Internal zone, at each stage, acting as an indentor. Thus, the downward migration of shear zones was coupled with flow of the hot mass of the internal zone up and over an indentor, strain softening of it, and incorporation of it into the wedge in successive stages. Although ramps are shown In Figs. 2 and 8 for illustrative purposes, the ramp geometry would have been progressively destroyed by progressive ductile strain; however, because of the conservation of mass, some of the excess volume from the foot of the ramp is preserved in the hanging wall.

The base of the exposed Gwillim Creek shear zone serves as an example. It was refrigerated from below prior to the cooling of the overlying rocks in its hanging wall as a result of overthrusting of hot rocks over cold footwall rocks on the upper "flat" of the ramp (Schaubs et al., 2002; Spear, 2004), and the rock volume from the foot of the ramp is now preserved above the Gwillim Creek shear zone in the Valhalla complex. Our general model is consistent with those of Beaumont et al. (2006, 2010) demonstrating the lateral transition from stiff cool crust to hotter weaker crust where the stiff cool crust acted as an indentor, with development of a ramp at the edge of the indentor, and flow of hot rocks up and over the ramp. Unless the front of the stiff crustal indentor was perfectly rectilinear, features that look like lateral ramps are to be expected. We propose that in the Internal zone, the footwall of the Gwillim Creek shear zone, insulated by the overriding sheet, heated, became weakened and strained, thus forming the younger and deeper Monashee décollement. In the Frenchman Cap dome, downward migration of strain related to the Monashee décollement is preserved in an arrested form (Crowley et al., 2001; Gervais et al., 2010). At the time of its initiation at ca. 60 Ma (Fig. 8c), the Monashee décollement was at the base of Cordilleran deformation. At that stage of development, the rocks beneath the Monashee décollement in the core of the Frenchman Cap dome had not yet been metamorphosed. The Monashee décollement was effectively the western continuation of the Rocky Mountain basal décollement, albeit thicker and weaker relative to the more localized basal décollement to the east, because it was well below the brittle–ductile transition. The interpretation of the Monashee décollement as the basal décollement is consistent with palinspastic reconstructions of the thrust belt. At each stage of shear zone development, the active shear zone in the Internal zone was effectively the basal décollement of its time. Subsequent cessation of motion of the active shear zone in the Internal zone, through strain hardening, correlates with cessation of motion of the thrust system linked to it. Similarly, activation of the next structurally deeper and younger shear zone in the Internal zone correlates with activation of the next thrust system in the External zone. If the shear zones are segments of the basal décollement abandoned as the retrowedge evolved, and if motion on them correlates with motion on the thrust systems with strike lengths of 400–600 km, it would seem reasonable that the shear zones also have strikes lengths of that order of magnitude. The Gwillim Creek shear zone may be traceable for ~400 km from the Spokane dome through the Valhalla dome and perhaps to the northwest margin of the Frenchman Cap dome. Perhaps the Monashee décollement is traceable for a similar distance, as we suggest in Fig. 7. The shear zones may well merge down dip and along strike and loose their identity. Recognition of the downward-younging compressional shear zones in the Internal zone and the notion that they represent abandoned strands of the basal décollement lead to the possibility of learning more about the nature and evolution of the basal décollement in the Cordillera and in other orogens.

There are a number of arguments supporting the contention that the basal décollement was effective and weak. The eastern portion of the Early Cretaceous orogen was not structurally overprinted during Late Cretaceous to Eocene orogenesis. It was effectively translated without significant deformation on the Late Cretaceous to Eocene basal décollement. The décollement lay below the brittle–ductile transition. This is illustrated by features in the hanging walls of the most westerly Late Cretaceous thrusts such as strong slaty cleavage with chlorite, white mica and local chloritoid, and the local deformation of quartz grains in Neoproterozoic siliciclastic rocks. These observations, together with evidence for repeated reactivation of the Rocky Mountain basal décollement between ca. 85 Ma and ca. 50 Ma, imply that the basal décollement of the orogenic retrowedge was weak. This may be an example of the findings of Stockmal et al. (2007) who show, through finite element modeling of thrust belts, that structures resembling the Rocky Mountain thrust belt can be achieved only if the décollement is modeled as very weak.

The Late Cretaceous to Paleocene Bourgeau–Rundle–Lewis thrust system and the Paleocene to Eocene McConnell thrust system have, in their central portions, some of the largest displacements, ~100 km and 50 km, respectively, of the southern Canadian Rocky Mountains. At their leading edge, in Paleozoic carbonates, displacement and deformation was facilitated by high fluid pressure as well as fluid assisted recrystallization of calcite (Kennedy and Logan, 1997). In the Internal zone in the Frenchman Cap dome, Thor-Odin dome, Valhalla complex and Priest River complex there are large displacements on shear zones, such as the Late Cretaceous to Paleocene Gwillim Creek shear zone and the Paleocene to Eocene Monashee décollement. High-grade metamorphic rocks with abundant leucosome lie within and above the shear zones suggesting that metamorphic and melt reactions weakened these zones (Hollister and Crawford, 1986). Models that propose channel flow, within the Internal zone, focus on rocks that lie within these migmatitic infrastructural zones, particularly those within the Thor-Odin dome and those overlying the Frenchman Cap dome (c.f. Williams and Jiang, 2005; Brown and Gibson, 2006; Gervais and Brown, 2011). At the latitude of the Frenchman Cap dome, localized channel flow within the Paleocene to Eocene infrastructure is not incompatible with geological relationships, as depicted in cross sections by Brown and Gibson (2006) and Gervais and Brown (2011), interpreted as a mid-crustal channel that extruded to higher structural levels above the present erosion surface of the Western Rocky Mountains. In more southerly transects, illustrated in Figs. 2, 4 and 5, structures east of the Frenchman Cap dome, Thor-Odin dome and Valhalla complex are all Jurassic and Early Cretaceous, and formed at low grade. Late Cretaceous to Eocene infrastructural flow zones dip eastward into the subsurface on the eastern margins of the domes. Any proposed Late Cretaceous – Eocene channel (Williams and Jiang, 2005) must lie below the older low-grade structures and above the Rocky Mountain basal décollement. There is insufficient room for a channel between the low-grade rocks and the underlying stiff basement east of the Selkirk Fan (SFA in Figs. 2 and 5) and east of the west limb of the Purcell Anticlinorium (Fig. 4). That geometry and the continuity of regional markers within the infrastructural zones do not preclude channel flow but place severe limits on the amount of flow. The strata in the proposed channels are ones expected there, not crust that has flowed into place from further west. Clearly channel flow is not the main mechanism for the retrowedge (c.f. Carr and Simony, 2006).

The question arises as to the interplay between between contractional and extensional structures in the Cordilleran metamorphic core complexes (Journeay and Brown, 1987; Parrish et al., 1988) and the formation of gneiss domes (Whitney et al., 2004, and references therein). We propose that there were two stages of Early Eocene extension in the Internal zone, similar to Modes I and II described by Fossen (2000) in the Caledonides. The younger stage, starting at ca. 52 Ma, coincided with the onset of crustal-scale extension accompanied by N–S faulting, dyke emplacement and voluminous magmatism, and marked the cessation of crustal shortening. The older ca. 59–52 Ma stage of extension is marked by ductile–brittle extensional shear zones localized on the margins of tectonothermal culminations (e.g. the Valhalla complex (Carr et al., 1987), the eastern margin of the Monashee complex (Parrish et al., 1988), early motion on the Okanagan Valley – Eagle River detachment system (Brown, 2010 and references therein) and perhaps in the Kettle dome and Priest River complex (Doughty and Price, 1999)) and coincided with continued shortening (see Section 4) on shear zones in the Internal zone and foreland thrusting. Notwithstanding the roles of buoyancy forces and tectonic exhumation, doming of metamorphic complexes such as the Valhalla complex and Frenchman Cap dome may, in part, have been facilitated by translation of incipient culminations on a subsurface basal décollement, and transport up frontal ramp structures in a crust (i.e. underthrusting of an indentor) that remained stiff.

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