**PRESSURE-TEMPERATURE-TIME-DEFORMATION PATHS OF
FORMER MID-CRUSTAL ROCKS, NORTHERN MONASHEE COMPLEX OF
THE SOUTHEASTERN CANADIAN CORDILLERA: A MODEL OF
SYNCONVERGENT EXHUMATION BY SEQUENTIAL DUCTILE
EXTRUSION**

**by
Félix Gervais, B.Sc., M.Sc.**

**A thesis submitted to the Faculty of Graduate Studies and Research
in partial fulfillment of the requirements for the degree of
Doctor of Philosophy, Department of Earth Sciences**

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À mon père Paul. ...

"The young specialist in English Lit, ...lectured me severely on the fact that in every century people have thought they understood the Universe at last, and in every century they were proved to be wrong. It follows that the one thing we can say about our modern "knowledge" is that it is wrong.

... My answer to him was, "... when people thought the Earth was flat, they were wrong. When people thought the Earth was spherical they were wrong. But if you think that thinking the Earth is spherical is just as wrong as thinking the Earth is flat, then your view is wronger than both of them put together."

Isaac Asimov, *The Relativity of Wrong*, Kensington Books, New York, 1996, p 226.

FRONTISPIECE

Me, pondering the wrongs in the middle of Frisby Glacier.

ABSTRACT

It is generally accepted that the high-grade rocks exposed in the southeastern Canadian Cordillera were exhumed from the middle crust by large-magnitude extension in the Paleocene-Eocene following gravitational collapse of a thickened crustal welt and/or change in far-field stress. Some models further proposed that gneiss domes of the Monashee Complex, the deepest structural level of the southeastern Canadian Cordillera, were produced by extension-assisted diapirism. However, a small number of studies also proposed that part of the exhumation occurred during convergence. This thesis aims at gaining a better understanding of the processes that led to the exhumation of former mid-crustal rocks to the surface of the Earth.

The first part of the thesis focuses on the core zone of the northern dome of the Monashee Complex, Frenchman Cap. After confirming previous studies in that all gray granite dykes in basement gneiss of the dome are part of a Paleoproterozoic suite, these dykes are used as time markers to demonstrate that only the upper ~1.5 km of the exposed 5-6 km basement section has a penetrative foliation developed during the Cordilleran orogeny. Below, strain patterns identified as Cordilleran are limited to open folds and meter-scale shear zones. Furthermore, phase equilibria forward modeling points to a maximum burial depth of ~20 km for the deepest exposed structural level of the dome. Data presented in this first part suggest that a downward-migrating base of convergence was frozen-in in the Eocene, and preclude diapirism as a viable doming mechanism for Frenchman Cap dome (FCD).

The focus of the second part shifts toward upper structural levels, and various exhumation processes are tested for the Lower Selkirk allochthon (LSa). LSa is a 10-15 km thick panel of migmatitic rocks that underlies the older lower-grade rocks of the Intermontane Belt and overlies the younger rocks of the Monashee Complex. Structural, metamorphic and geochronologic data are combined to construct a detailed Pressure-Temperature-time-deformation path (P-T-t-d) for the thrust shear zone at the base of the LSa. This path indicates that the last increments of thrust shear strain took place between ~62 and 57 Ma, and were associated with near-isothermal decompression. A compilation of previously published data further suggests that normal shearing along the shear zone that defines the roof of the LSa also occurred in the Paleocene. Combined with several depth-time paths that point to synchronous burial of the underlying Monashee Complex, this coeval motion of two oppositely-verging shear zones bounding a migmatitic panel suggests that the LSa was exhumed to upper crustal level primarily by Paleocene syn-convergent ductile extrusion, rather than large-magnitude extension.

The four P-T-t paths constructed in the last part indicate that rocks of the Monashee cover sequence, which are now sandwiched in between the LSa above and the Monashee Complex basement below, were exhumed from a depth of ~40 km to less than ~14 km depth by ductile extrusion after ductile extrusion of the LSa. Fold nappes developed during extrusion, most likely induced by internal density contrast within the extruding channel. A first-order force balance analysis indicates that underthrusting of a felsic basement beneath a denser orogenic wedge should produce a lateral pressure gradient large enough to drive low-viscosity rocks, such as migmatites, up a 30° ramp.

Consequently, results of this thesis strongly argue against large-magnitude extension for exhumation of high-grade rocks of the southeastern Canadian Cordillera at the latitude of the Monashee Complex. Extension probably played a role but only for the last stages of exhumation, once the rocks were already exhumed to upper crustal levels.

ACKNOWLEDGEMENT

It is an honour for me to (most likely) be the last graduate student supervised by Richard L. Brown. Not only has this great geoscientist significantly contributed to the advancement of Cordilleran tectonics and structural geology in general, but he also has formed and inspired a generation of "tectonicians" that are now leaders in governmental research centres, universities and industries. Like many others before me, Dick has allowed my potential to flourish to its near maximum and has nourished my ambition to become one of tomorrow's leaders. Merci Dick!

The next person I am grateful to is my friend Jim Crowley, without whom this thesis would be a mere diminutive of itself. Jim was there for my first experience in the mountains in 2003 and again for two weeks of my last field season in 2005; he provided invaluable tips during the planning of my two field campaigns; took seven full days of his busy (no kidding!) time to teach me the intricacies of geochronological processing, from clean rock crushing to U-Pb analysis; but most importantly lay the geochronologic foundation for my metamorphic and structural analysis. In addition, the endless arguments we had about the basement story allowed every possible hypothesis to be turned upside down and led to writing of the Crowley et al. (2008) paper and to the final version of chapter 2.

Sharon Carr and John Percival, my two internal examiners, also deserve a special acknowledgment. Not only did their critical, meticulous and rigorous corrections of the two previous drafts of this thesis induce a permanent adaptation to stress hormones :), but more importantly they taught me the art of rigorous scientific writing and self-criticism. I still have a great deal to learn about these two crucial skills, but at least I now have a solid basis and know what is expected. I also greatly acknowledge the informal but meticulous review of chapter 3 by Kyle Larson.

Technical assistance by Peter Jones (electron microprobe), Mike Jackson and George Mrazek (thin-sections preparation), Brian Cousens and John Blenkinsop (Rb-Sr and Sm-Nd chemical analysis), Antonio Simonetti (U-Pb La-MC-ICP-MS geochronology) are gratefully acknowledged. Sam Bowring is also acknowledged for letting me use the mineral separation lab at MIT.

The ideas presented in this thesis have greatly benefited from discussions with Sharon Carr, Deanne van Rooyen, George Skippen, Brian Cousens, Giorgio Ranalli and Juan Carlos Afonso from Carleton University; John Percival, and Rob Berman from the GSC, Paul Williams, Dan Gibson, Andy Parmenter, Guido Gosso and Maria Iole Spalla in the mountains and around a boot-mug of cold Tall Timber at the Ol'Frontier in Revelstoke; Dan Gibson, Phil Simony, Ray Price, Davide Zanoni, Brad Johnson, Don Murphy, Ed Ghent and David Pattison at various Cordilleran Tectonic Workshops; Laurent Godin, Ray Price, Alfredo Camacho, Kyle Larson, Dawn Kellett and Rob Harrap from the highlighting tectonic discussion groups at Queen's University.

I am also grateful for the excellent helicopter support provided by Matt and his Kokanee Helicopters crew as well as for the excellent field assistance provided by Denis Vachon and Nancy Williams.

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Un énorme merci à mon père Paul, à qui cette thèse est dédiée. En plus de l'apaisant

soutien financier que tu m'as procuré, je te serai éternellement reconnaissant pour cette incroyable aptitude à dire les bons mots aux moments les plus opportuns, de même que pour la confiance inébranlable que tu as toujours manifestée envers moi. Un remerciement non-conventionnel à ma mère Louise qui a su éveillé, dès le berceau, un émerveillement candide mêlé d'une avide curiosité devant les beautés de la nature. Je considère que cette avide curiosité combinée à un sens créatif sont des aptitudes aussi importante que l'esprit analytique et la rigueur scientifique pour la recherche en science pure. Merci à la gang de St-Sac et compagnie, et à Kratt, Pat et Karine pour avoir grandement contribué à évacuer le "trop-plein" de temps à autre! Vive le suit rose!

Enfin, un merci tout spécial à ma Muse Nancy. Sans ces cinq années d'amour, de rires, de bouffes & vins, bref de bonheur constant, je ne sais pas si j'y serais arrivé...

ORIGINAL CONTRIBUTION

This research aimed at gaining a better understanding of the process(es) that led to the exposure of high-grade rocks within and surrounding the Monashee Complex of the southeastern Canadian Cordillera. This required integrating previously published data into newly acquired structural, metamorphic and geochronologic data. The choice of the study areas and study objectives were entirely mine. Fieldwork was conducted in parts of the following 1: 20 000 BC TRIM sheets (North American Datum 1983; UTM Zone 11): 082M.066, 082M.056, 082M.046, 082M.047, 082M.048, 082M.037, 082M.038, 082M.027, 082M.028, 082M.017, and 082M.018. I originally planned mapping the area located at the northern toe of Bourne Glacier (sheet 082M.038) at the 1: 10 000 scale, but bad weather and an injury to my field assistant resulted in a map that is more representative of the 1: 25 000 scale.

Phase equilibria modeling with the program *Perple_X* was learned and conducted entirely by myself. Pseudosections for specimen DG-167 had been calculated beforehand by Foster et al. (2004), but were calculated without a melt model and without Ti as a system component. Incorporating these modifications into the calculations had significant effects that are discussed in the text.

Prior to modeling, I examined thin-sections under an optical microscope; I acquired chemical compositions of metamorphic minerals with an electron microprobe at Carleton University under the supervision of Peter Jones; I obtained major and trace element analyses by X-Ray fluorescence (XRF) under the supervision of Ron Hartree at University of Ottawa.

For the geochronologic part of this study, I carried out U-Pb laser ablation – multiple collector – inductively coupled plasma – mass spectrometry (LA-MC-ICP-MS) analyses on a total of four specimens, three leucosomes and one gray granite, from which 85 analyses on 70 zircon grains and 31 analyses on 15 titanite grains were conducted. I conducted all stages of specimen preparation, including crushing, grinding, mineral separation, and grain mounting at the Massachusetts Institute of Technology under the supervision of Dr. James L. Crowley. Accompanied by Jim Crowley, I conducted LA-MC-ICP-MS U-Pb analyses at the Radiogenic Isotope Facility of University of Alberta under the supervision of Dr. Antonio Simonetti.

Prior to U-Pb geochronology, I imaged the internal morphology of zircon and titanite by backscattered scanning electron (BSE) and cathodoluminescence (CL) at Carleton University.

I conducted Rb-Sr and Sm-Nd isotopic analyses on intrusive basement rocks from Frenchman Cap dome under the supervision of Brian Cousens and John Blenkinsop. Specimen preparation, including rock crushing, grinding, isotope dilution, and mass spectrometry were conducted at Carleton University under the supervision of Brian Cousens and John Blenkinsop.

I drew all figures and prepared all tables presented in this thesis, unless otherwise cited. This thesis presents novel structural, metamorphic and tectonic models that are the fruit of my interpretation of the data. Finally, I am entirely responsible for any inaccuracies that may have slipped through.

METHOD OF PRESENTATION

This thesis is presented as five chapters, including an introduction and a conclusion. My intent is to submit chapters 2, 3, and 4 to scientific journals. Due to its length, it is possible that chapter 4 will be subdivided into two separate papers. I apologize for the redundancies introduced by the journal format, especially in chapter 3 that incorporates parts of chapter 1.

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CHAPTER 1

GENERAL INTRODUCTION

1.1 Tectonic setting

The ~900 km wide southern Canadian Cordillera can be subdivided in five morphogeological belts based on different physiographic and geological characteristics (Fig. 1.1a; Gabrielse et al., 1991). The Intermontane Belt is a relatively flat region of lower elevation (~1200 m from Tribe, 2005 and Flück et al., 2003) than the Coast Belt to the west and Omineca Belt to the east (~1600 m for the southern Omineca Belt; Flück et al., 2003). The topography of the two outer belts — Insular and Foreland — are rather rugged and of high elevation (~1800 m for the Foreland Belt; Flück et al., 2003). The Insular, Intermontane and Foreland belts mainly comprise rocks of low metamorphic grade, whereas the Coast and Omineca belts are zones of amalgamation and accretion comprising variously deformed metamorphic and plutonic rocks (Monger et al., 1982). The study area surrounds Frenchman Cap dome of the Omineca Belt and fieldwork was conducted in the alpine landscape and on logging roads northwest of the town of Revelstoke (Fig. 1.1a).

The subsurface expression of the southern Canadian Cordillera has been characterized by the Lithoprobe Geoscience Program conducted in the 80's and 90's (e.g. Brown et al., 1992; Varsek and Cook, 1994; Buriannyk and Kaneseewich, 1995; Carr, 1995; Hyndman and Lewis, 1999). Several important features are illustrated on the cross-section passing through the Intermontane and Omineca belts presented on figure 1.1b. This section is the interpreted model of Buriannyk and Kaneseewich (1997) for Line 7 of the Lithoprobe's 1990 Southern Cordillera Refraction Experiment, combined with

interpreted reflectivity of Cook et al. (1992). This section shows a flat Moho that rises rapidly from a depth of ~40 km beneath the Omineca-Foreland belts boundary to a depth of ~32 km beneath the western Omineca and Intermontane belts. Hyndman and Lewis (1999) demonstrated that this gradient in Moho depth coincides with a major boundary between a cold and thick lithosphere to the east and a hot and thin lithosphere to the west. Hence, the temperature at the Moho is ~600°C beneath the Foreland Belt compared to ~900°C beneath the Omineca and Intermontane belts. The thickness of the lithosphere is ~150 km beneath the Foreland Belt compared to ~50 km beneath the Omineca and Intermontane belts (Hyndman and Lewis, 1999).

The cross-section also shows a good fit between seismic reflectors and P-wave velocity structures. The lower crust is formed by a 10-15 km thick panel with high P-wave velocity that indicates a mafic composition (Zelt et al., 1995; Burianyk and Kanasewich, 1997). The presence of two lobate structures defined by medium V_p (green and yellow centred on at km ~210 and ~275 on Fig. 1.1b) bounded by reflectors interpreted as the subsurface expression of major faults/shear zones bounding structural domains of the Omineca Belt is noteworthy. Cropping out at ~ km 225 of the section, the Okanagan Valley fault system (OVfs) is the west-dipping, top-to-the-west brittle-ductile shear zone that separates the low-grade rocks of the Intermontane Belt to the west from the sillimanite-migmatites of the Lower Selkirk allochthon to the east (see below for terminology). Cropping out at ~km 275 of the section, the Monashee décollement (MD) is the west-dipping top-to-the-east shear zone that separates the LSa from the underlying Monashee Complex (MC). Although not shown on the figure of Burianyk and Kanasewich (1997), the trace of the east-dipping, top-to-the-east Columbia River Fault

(CRF) is also revealed by the P-wave velocity structure to the east of the easternmost “lobe”.. Subsurface structures revealed by crustal velocity thus suggest that the southern Omineca Belt is characterized by two shear zone-bounded lobes that root at the top of a mafic lower crust. Finally, the section suggests that the Intermontane Belt is ~20 km thick above the mafic lower crust. These geometric constraints will be used to construct the tectonic model of chapter 4.

A comprehensive synthesis of the Jurassic-Cretaceous tectonic evolution of the central and southeastern Canadian Cordillera was presented by Evenchick et al. (2007) who linked a wide range of data across the orogen. An important assumption made by Evenchick et al. was that the Bowser and Sustut basins, part of which are deformed in the Skeena Fold Belt, were located at the latitude of the study area in the Jurassic-Cretaceous period. This assumption is justified by the ~800 km of dextral strike-slip displacement on the Fraser and associated faults estimated by Gabrielse et al. (2006) based on independent evidence. Of this displacement, ~500 km would have occurred in the Late Cretaceous to Eocene time (Gabrielse et al., 2006). The two basins and the Skeena Fold Belt would thus have been located immediately west of the southern Omineca Belt in the Jurassic-Late Cretaceous period. This reconstruction allowed Evenchick et al. to link: 1) patterns of sedimentation in the two basins flanking the Omineca Belt: the Bowser/Sustut basins to the west and the Foreland basin to the east; 2) deformation in the two fold-and-thrust belts flanking the Omineca Belt: the Skeena Fold Belt to the west and the Foreland Fold belt to the east; 3) deformation, metamorphism and plutonic activity in the Omineca Belt. The consistency between the timing of major events across the orogen suggests that their reconstruction is not only permissible, but indicated. The following paragraph

summarizes the tectonic evolution presented by Evenchick et al. (2007), complemented by other relevant studies.

The Jurassic was a period of final amalgamation of terranes that were thrust and obducted over North American pericratonic terranes and basement. This imbrication resulted in the formation of a crustal edifice that would become the Omineca Belt. Southwest-verging fold nappe formation at depth was accompanied by the development of a highland that started to shed material from the carapace of accreted terranes in both basins flanking its margin. After a period of northeast convergence, continued thickening and marine deposition, the period between 145-135 Ma marked an abrupt change across the Cordillera. This change was characterized by regional uplift and deposition of more than 1.5 km of nonmarine sediments in both the Bowser and Foreland basins, which was accompanied by the formation of zones of penetrative northeast verging deformation at mid-crustal levels of the Omineca Belt. Thrusting is inferred to have started soon after this period in both the Skeena and Foreland Fold belts, flanking respectively the western and eastern margins of the Omineca Belt. The period between ~115-90 Ma was characterized by: 1) ductile deformation, high-grade metamorphism and pluton emplacement in both the Omineca and Coast Belt; 2) by intense deformation in the Skeena Fold Belt; 3) by the first deposition of muscovite-bearing clasts in the Sustut basin; 4) by subsidence and deposition of >1.7 km of sediments in the Foreland Basin.

The next important period — spanning from ~74-59 Ma— is the milestone for this study because it is during this timeframe that ~200 out of the ~250 km of net shortening took place in the Foreland Belt at this latitude (Price and Sears, 2000). In addition, the late Campanian (~72 Ma) marked a major depositional change in both basins flanking the

Omineca Belt with a change from low to high-energy sediment input in the Foreland and Sustut basins. It is very interesting to note that this period also coincides with a rapid northward migration of the stratigraphic hinge line in the Foreland Basin as determined by reciprocal stratigraphy (Catuneanu et al., 2000). The lobate shape of this hinge line centred on southern Canadian Cordillera from late Campanian to early Paleocene (~72-61 Ma) was further interpreted as evidence of localized tectonic loading by Catuneanu et al. (2000). This period was also the locus of intense deformation in the two fold belts and of top-to-the-east shearing at mid-crustal level of the Omineca Belt (Carr, 1992; Scammell, 1993; Parrish, 1995; Spear and Parrish, 1996). Significant deposition took place during the Paleocene in the Foreland Basin with an estimate of ~2.5 km of sediments eroded above the 3 km thick ~58 Ma Paskapoo Formation (Kalkreuth and McMechan, 1996). Thrusting also continued for some time after ~58 Ma because the Paskapoo Formation is deformed (Hamblin, 1997). Paleocene to Early Eocene high-grade metamorphism and deformation is also recorded at the deepest structural levels of the Omineca Belt in the Monashee Complex (e.g. Crowley and Parrish, 1999; Vanderhaeghe et al., 1999; Crowley et al., 2001; Hinchey et al., 2006). This was partly synchronous with shearing along crustal-scale normal brittle-ductile shear zones flanking the Omineca Belt (e.g. Parrish et al., 1988; Carr, 1992; Johnson, 2006). Finally, it is instructive to note that Stock and Molnar (1988) reconstructed past plate motion in the Pacific for the period 70-35 Ma. These authors documented a progressive reduction in northeast-southwest convergence rate between the northern Farallon plate and North America from an average of 126 mm/yr between 68 and 56 Ma, to 117 mm/yr between 56 and 47 Ma, to 99 mm/yr between 47-40 Ma (uncertainty varying between 37 and 53 mm/yr). From

these reconstructions, it thus appears that a high convergence rate in excess of 100 mm/yr was maintained until ~47 Ma.

1.2 A quantitative subdivision for the southeastern Canadian Cordillera

Before focussing on a discussion about tectonic models that have been proposed for the southeastern Canadian Cordillera, it is important to clarify and justify the terminology used in this thesis. The new terminology proposed is an attempt at clarifying the confusion that is present in the literature. A quick review of papers published between 2000 and 2008 reveals that there is no consensus on the way to name the different tectonometamorphic units of the southeastern Canadian Cordillera. For example, the westernmost high-grade rocks of the Omineca Belt located west of Revelstoke (Fig. 1.1a for location) have been referred to as “Kootenay Terrane” (e.g. Johnston et al., 2000); “undivided high grade rocks” (Carr and Simony, 2006); “Middle Unit of the Shuswap Complex” (Teyssier et al., 2005); “Late Proterozoic-Mesozoic metamorphic assemblage of the Middle-crustal zone” (Kruse and Williams, 2007); “Neoproterozoic Windermere Supergroup of the Selkirk allochthon” (Gibson et al., 2008); “Monashee Cover sequence” (Thompson et al., 2006) — to name a few. A unique terminology based on testable criteria would greatly facilitate the comprehension of tectonic issues for geologists not familiar with the Cordillera and students starting Cordilleran studies, and could lead to less divergence of opinion in the Cordilleran community. This section starts by partially reviewing the evolution of ideas that led to the different subdivision schemes. It highlights the need to use thermochronologic, metamorphic and structural data to subdivide high-grade rocks. A new terminology is then proposed based on testable and objective criteria. This should allow refining or modifying the proposed boundaries of the

different tectonometamorphic units with the acquisition of new data.

Geological maps are commonly subdivided on the basis of lithostratigraphy. The geological map of Fig. 1.3 (modified from Gibson, 2003) presents a regional map subdivided into lithostratigraphic units. As can be observed on this map, lithostratigraphic units are grouped into the Selkirk allochthon and structurally underlying Monashee Complex. These two terms were introduced by Read and Brown (1981) based on the mapping of a zone of mylonitic rocks recording thrust-sense shearing, the Monashee décollement, which separates underlying autochthonous basement rocks and unconformable cover rocks of the Monashee Complex from overlying allochthonous rocks of the Selkirk allochthon. A problem with such a lithostratigraphic-based subdivision is that it could be quite difficult to assign a specific rock assemblage to a lithostratigraphic unit in rocks that have been transposed and metamorphosed up to upper amphibolite facies. For example, the light brown unit of the Windermere Supergroup on Fig. 1.3 is supposed to have been deposited in the Neoproterozoic, but Scammell (1993) presented evidence suggesting that rocks from the light brown unit north of the Monashee Complex could in fact be a mix of metasedimentary rocks ranging in age from Neoproterozoic to Early Paleozoic. Thompson et al. (2006) also challenged the lithostratigraphic subdivision shown on Fig. 1.3 on the basis of a reinterpretation of the paleogeographic origin of different lithostratigraphic packages. The purpose here is not to judge these disparate arguments, but rather to highlight that a subdivision based on lithostratigraphy in highly metamorphosed and highly strained rocks remains qualitative and open to discussion unless supported by detailed geochronologic and detritus geochronologic data. This latter

approach has been used with success in the Cariboo Mountains (NW of Fig. 1.3) by Reid (2003) who distinguished lithostratigraphic units based mainly on Nd isotopic analyses. Unfortunately, such studies are rare and another subdivision scheme based on available data would be preferable.

Another approach is to use metamorphic grade as a subdivision scheme. For example, Okulitch (1984), building on Wheeler (1965) and Read and Brown (1981) defined the “Shuswap Complex” as all rocks at the sillimanite grade. However, this approach can potentially link rocks that were metamorphosed at very different times as was demonstrated by Gibson et al. (2005, 2008) for sillimanite-grade rocks of the Selkirk Mountains (Fig. 1.3).

A more quantitative subdivision scheme is to use geochronologic data. Accordingly, Parrish et al. (1988) used K-Ar mica dates in the 45-60 Ma range to subdivide the southern Omineca Belt. However, cooling ages only put a timing constraint on the low-temperature cooling of rocks that record much higher peak temperature and could group together rocks that were only juxtaposed late in their tectonic evolution. An obvious problem with the cooling ages subdivision scheme is that it does not single-out the Monashee Complex. Furthermore, Parrish et al. (1988) presented data for their subdivision up to the latitude of the northern Monashee *Complex*, but their map was later extrapolated up to the northern Monashee *Mountains* in most studies afterward (Fig. 1.3). This extrapolation resulted in some inconsistencies. For example, most maps show the eastern boundary of the Shuswap Complex to follow the Columbia River up to the northern end of the Columbia River Fault (see Fig. 1.3), whereas lithologic units and isograds are continuous across the river (see Fig. 1.3 and Gibson et al., 2005 & 2008).

Since Parrish et al. (1988), 20 years of active research have been conducted by at least five research groups in the area surrounding the Monashee Complex. Because geochronologic data were acquired in most of these studies, I suggest building on the pioneer work of Parrish et al. (1988) and I propose a new subdivision scheme based on geochronologic data.

The map of Fig. 1.4 along with three simplified cross-sections shown on Fig. 1.5 present the new subdivision adopted in this thesis along with thermochronologic data from which it is based. References for this compilation are presented in Table 1.1. The northernmost section A-B (Fig. 1.5a) that runs SW-NE is modified from Gibson (2003), the C-D (Fig. 1.5b) cross-section that runs W-E is modified from Johnson and Brown (1996); whereas cross-section E-F (Fig. 1.5c) that runs W-E combines the interpretation of the Lithoprobe seismic reflection lines 7-9 by Glombick et al. (2006a) and the interpretation of the Lithoprobe seismic refraction line 7 by Burianyk and Kanesevich (1997). The main criteria that distinguish the three units are the timing of ductile deformation and thermal peak. On Fig. 1.3 and 1.4, U-Pb ages of deformed intrusion/leucosome are shown by triangles, U-Pb ages of monazite in schists by pentagons, U-Pb ages of posttectonic dykes by circles, and cooling ages by squares. On the map of Fig. 1.4, most cooling ages are for hornblende except when an adjacent letter indicates otherwise (*t* for U-Pb titanite, *b* for $^{40}\text{Ar}/^{39}\text{Ar}$ of biotite, and *ap* for apatite fission track of apatite). Numerical values of these ages are shown on the simplified cross-sections of Fig. 1.5, and referenced on Table 1.1. For $^{40}\text{Ar}/^{39}\text{Ar}$ analyses, only plateau ages defined by spectra of more than 80% of ^{39}Ar are shown. Black symbols represent ages older than 120 Ma, gray symbols represent ages between 120-60 Ma, and

white symbols are ages between 59-47 Ma. From all these data, three main *tectonometamorphic* units are singled-out. From upper to lower structural levels: the light gray unit is the Upper Selkirk allochthon (USa) that was deformed and metamorphosed before 120 Ma (black symbols); the medium gray unit is the Lower Selkirk allochthon (LSa) that was mainly deformed and metamorphosed between 120 and 60 Ma (gray symbols); and the dark gray unit is the Monashee Complex (MC) that was deformed and metamorphosed between 59 and 47 Ma (white symbols). These names were chosen rather than the upper-middle-lower crustal zones of Carr (1991) because they assure continuity with the traditional subdivision presented in Read and Brown (1981).

1.2.1 Cross-section C-D

Cross-section C-D (Fig. 1.5b) is the easiest to interpret. Along this section, all the illustrated boundaries are shear zones. The westerly-dipping, top-to-the-west brittle-ductile Okanagan Valley fault system (OVfs) separates the USa from the underlying LSa; the westerly-dipping, top-to-the-east Monashee décollement (MD) separates the LSa from the underlying Monashee Complex; and the easterly-dipping Columbia River Fault separates the USa from the underlying Monashee Complex. Notice that I follow Johnson (2006) and will refer to the Okanagan Valley fault system (OVfs) for the western boundary between the USa and underlying LSa. The OVfs thus groups the Adams-North Thompson fault, the Eagle River fault, the Kalamalka shear zone, and the Silver Star shear zone (Glombick et al., 2006b; Johnson, 2006). Along this section, the hanging wall of the OVfs consists of rocks at the greenschist facies or lower, whereas the footwall of the fault consists of migmatitic rocks at the sillimanite-K-feldspar grade that are mylonitic and contain numerous top-to-the-west shear sense indicators (Johnson, 1994 &

2006). Johnson (2006) documented that top-to-the-west structures are formed by the peak metamorphic minerals garnet-sillimanite-K-feldspar, by the retrograde minerals muscovite and chlorite, as well as by brittle structures. Johnson's interpretation was that top-to-the-west shearing would have started at high-temperature and continued down temperature and pressure until the rocks had been cooled and exhumed above the ductile-brittle transition. Hangingwall rocks cooled through the apatite fission track temperature ($110 \pm 10^\circ\text{C}$, Green et al., 1989) by 67 Ma (Lorencak et al., 2001), whereas footwall rocks record complex and protracted zircon growth in sheared leucosome. An interesting study that has not been published, nor cited, is Gilley (1999) who conducted SHRIMP U-Pb dating of zircon grains from three leucosome specimens collected in the immediate footwall of the OVfs near the town of Sicamous (Fig. 1.4). Gilley (1999) documented a first episode of zircon growth occurring between ~ 160 - 155 Ma, and almost continuous zircon growth from ~ 120 to ~ 57 Ma.

A well-developed L-S transposition fabric can be observed in the Hunters Range (HR on Fig. 1.4) between the OVfs and the Monashee décollement (Johnson, 2006). The planar fabric is formed by the original bedding, various generations of veins and flattened leucosome, and aligned planar minerals that have all been transposed parallel to a transposition foliation. The linear fabric is formed by elongated mineral aggregates, elongated quartzo-feldspathic pressure shadows around garnet, aligned metamorphic minerals, and rod-shape leucosome that together define a stretching lineation. These structures are common throughout the LSa and will thus not be described for the other cross-sections (Journey, 1986; Carr, 1991; Scammell, 1993; Johnston, 1998; Glombick et al., 2006a). In the immediate footwall of the OVfs, lineations plunge shallowly to the

west, whereas lineations trend NE-SW throughout the Hunters Range (Johnson, 2006). Johnson (1994) reported that kinematic indicators and mylonite are rare in this area coring the LSa, whereas kinematic indicators abound within mylonites of the two oppositely verging shear zones bounding the LSa. Johnson (1994) interpreted these features as reflecting an increase of the kinematic vorticity number (Hanmer and Passchier, 1991) from the core of the LSa to its margins.

The lower boundary of the LSa with the underlying Monashee Complex is the Monashee décollement, for which the type locality crops out along cross-section A-B described below. The presence of the Monashee décollement at the latitude of Thor-Odin dome has been questioned by Johnston (1998), Vanderhaeghe et al. (1999) and Williams and Jiang (2005). These authors did not identify a strain gradient increasing toward the décollement. However, these authors also recognized that the trace of the Monashee décollement around Thor-Odin dome does coincide with a lithologic boundary that nearly coincides (<1 km horizontal divergence) with the kyanite-out isograd (Harrap, 1990; Carr, 1991; McNicoll and Brown, 1995; Johnston et al., 2000). In addition, this boundary nearly coincides with a geochronologic boundary that separates rocks deformed and metamorphosed between ~120 and 60 Ma (gray symbols on Fig. 1.4 and 1.5) from rocks deformed and metamorphosed between ~59 and 47 Ma (white symbols). Near Three Valley Gap (TVG on Fig. 1.4), zircon grew in leucosome between ~100 and 70 Ma with a cluster of dates between ~75 and 72 Ma (Gilley, 1999; Kuiper, 2003). The transposition foliation started to develop prior to ~73 Ma and continued to after ~65-67 Ma (Johnson, 1994; Johnston et al., 2000). In contrast, there is no evidence of deformation and high temperature metamorphism prior to ~60 Ma in the underlying Monashee

Complex (Johnston et al., 2000; Kuiper, 2003).

The eastern boundary between the Monashee Complex and the USa is marked by the brittle-ductile Columbia River Fault (CRF) that juxtaposes the staurolite-kyanite and lower grade rocks of the Clachnacudainn salient in its hanging wall (Crowley et al., 1996) to the migmatitic-kyanite-sillimanite-K-feldspar grade rocks of the Monashee Complex in its footwall (Brown and Brown and Psutka, 1980; Read and Brown, 1981; Brown and Murphy, 1982; Lane et al., 1989; Kruse, 2007; Mulch et al., 2007). Johnson and Brown (1996) restored the unmodified version of this cross-section to its preextension state. They showed that 45 km of net extension accommodated by the Okanagan Valley and Columbia River faults links together stratigraphic units in the hanging wall of both faults that are similar “...in terms of stratigraphy, metamorphic grade, structural style, and their unique association with widespread Devonian-Mississippian orthogneisses.” It is important for this thesis to note that this restoration puts rocks of the Monashee Complex and LSa at a depth of less than ~14 km; an exhumation mechanism other than extension is therefore required to bring these rocks from mid-crustal levels at >25 km depth to less than 14 km.

1.2.2 Cross-section E-F

The next cross-section examined is the southern section E-F modified from Glombick et al. (2006a), which is largely based on the interpretation of seismic reflection and refraction Lithoprobe profiles that were conducted at this latitude (Fig. 1.5c). As for section C-D, most rocks of the LSa along this transect are at the sillimanite-K-feldspar grade (Glombick, 2005 partly modifying Carr, 1991) except in the Pinnacles area (Pnc on Fig. 1.4) where the muscovite-out, and sillimanite-in isograds have been mapped towards

the L_{Sa}-U_{Sa} boundaries surrounding the high-grade core.

Glombick et al. (2006b) conducted detailed U-Pb geochronologic studies in the area surrounding the town of Vernon (SW corner of the map of Fig. 1.4). They demonstrated that the hanging wall of the Okanagan Valley fault system (OVfs) was deformed between ~172 and ~162 Ma; whereas migmatitic rocks of the Aberdeen Complex in the footwall of the fault record a protracted history of zircon growth between ~155 and 150 Ma, at ~90 Ma, and between ~66 and ~51 Ma. The maximal age of deformation in the complex is constrained by the ~52 Ma emplacement age of a monzonite pluton that postdates the foliation in enclosing gneiss (Glombick et al., 2006b). Noteworthy is the 61.6 ± 4.2 Ma age obtained from two concordant titanite fractions in a migmatitic diorite gneiss that suggests cooling below the titanite closure temperature at this time (~600-700°C depending on grain size and cooling rate, Cherniak, 1993). The last increments of ductile top-to-the west shearing along the OVfs are constrained to have occurred between ~50 and ~47 Ma, the U-Pb age of a deformed felsic dyke and of an undeformed mafic dyke respectively (Glombick et al., 2006b citing Heaman et al., 1999). These results are consistent with the $^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages obtained by Bardoux (1993) just south of the southern limit of Fig. 1.4. Biotite in the hanging wall of the OVfs yielded a plateau $^{40}\text{Ar}/^{39}\text{Ar}$ age of ~160 Ma, whereas hornblende and biotite in the footwall rocks yielded $^{40}\text{Ar}/^{39}\text{Ar}$ ages between ~52 and ~49 Ma, respectively. The U_{Sa}-L_{Sa} boundary is thus a top-to-the-west detachment that juxtaposes rocks that were deformed and cooled to <300°C in the Jurassic to rocks that were at temperature higher than 550-500°C (closure temperature of hornblende) until ~50 Ma.

Between Vernon and Nakusp (Fig. 1.4), two tectonic windows expose infrastructure

within USa exposure, whereas three outliers of USa are present within LSa exposure. Glombick et al. (2006a) showed that these observations are due to the subhorizontal attitude of the USa-LSa boundary in this area.

Carr (1992) used the age of variously deformed Ladybird leucogranite to constrain the timing of transposition between ~61 and ~55 Ma in the Pinnacles area (Pnc on Fig. 1.4), and at ~59-58 Ma for the Monashee décollement at the base of the LSa. A ~57 Ma $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende plateau age yielded by a specimen collected in the hanging wall of the Monashee décollement (Spark, 2001) is also consistent with the U-Pb dates and suggest rapid cooling of this part of the LSa soon after cessation of ductile shearing. These dates from the LSa contrast with the timing of deformation and metamorphism in the core of Thor-Odin dome of the underlying Monashee Complex that is bracketed between ~60 and 50 Ma (Vanderhaeghe et al., 1999; Hinchey et al., 2006, 2007).

Finally, along the E-F section, the easterly-dipping Columbia River Fault separates the LSa that records Late-Cretaceous-Paleocene high-temperature deformation in its footwall from the USa intruded by the posttectonic ~173 Ma Kuskanax Batholith (Parrish and Wheeler, 1983) in its hanging wall. From studies conducted in the southernmost part of Fig. 1.4, Lemieux (2006) concluded that the net displacement on the Columbia River Fault was minimal because markers could be mapped across the fault. The displacement thus progressively decreases from the core of the Monashee Complex southward.

1.2.3 Cross-section A-B

The northern cross-section A-B is discussed last because its interpretation is less straightforward than the other two, in part because the pitch of the average regional stretching lineation in the average foliation plane changes from almost down-dip to

almost strike-parallel from south to north across the map (Fig. 1.4). Hence, by contrast with the other two sections that are approximately parallel to the average regional stretching lineation and perpendicular to the average regional foliation, drawing this section perpendicular to the foliation strike results in an obliquity larger than 45° to the lineation trend. An assumption of the cross-section is that rocks of the L_{Sa} located north of the Monashee Complex and west of the Columbia River (between Scrip Creek, SC, and Monashee décollement, MD on Fig. 1.4) project along strike *beneath* the U_{Sa} that is exposed east of the river. This assumption finds its justification in the continuous exposure of L_{Sa} rocks from the Selkirk Mountains, to the northern Monashee Mountains, and then toward the SW around the Monashee Complex (Fig. 1.4). Hence, rocks of the L_{Sa} illustrated in the subsurface beneath the U_{Sa} on cross-section A-B (Fig. 1.5a) were most likely adjacent to rocks from Scrip Creek (SC) to the Monashee décollement (MD) exposed north of the Monashee Complex (Fig. 1.4) before being down-faulted by the Columbia River Fault.

The OVfs is not as well defined as in the other two sections because there are less data available. However, a major discontinuity is suggested by $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende plateau dates ranging between ~111 and 74 Ma in the immediate hangingwall of the fault and a crystallization age of ~56 Ma for a deformed leucogranite in the footwall (Johnson, 1994). It is from studies conducted at the latitude of this cross-section that the Monashee décollement was best documented (Journeay, 1986; Scammell, 1986; Brown et al., 1992). There it marks: 1) a lithologic break with rocks characteristic of the Monashee Complex stratigraphic assemblage (quartzite, marble, pelitic schists) in the footwall, and rocks characteristic of the semi-pelite amphibolite (SPA) division of the Neoproterozoic

Horsethief Group in the hangingwall; 2) an abrupt increase in leucogranite volume from <15 % in the footwall to more than 30% in the hangingwall; 3) a metamorphic discontinuity with kyanite migmatite in the footwall and sillimanite migmatite in the hangingwall; 4) a geochronologic break. Accordingly, rocks from the LSa in the hangingwall of the décollement record protracted zircon growth in sheared leucosome pods and leucogranite sheets at ~118-115, ~97, ~93 and ~85 Ma (Parrish, 1995; Scammell, 1993) and top-to-the-east shearing ending prior to the ~57 Ma crystallization age of an undeformed leucogranite dyke cutting the foliation at right angles (Scammell, 1993). By contrast, extensive geochronologic studies conducted in rocks of the Monashee Complex indicated that high temperature metamorphism and deformation occurred from ~60-50 Ma (Crowley and Parrish, 1999; Gibson et al., 1999; Crowley et al., 2001; Foster et al., 2004).

Scammell (1993) integrated structural, metamorphic and geochronologic data from the LSa located between SC and MD on Fig. 1.4 into a model of Cretaceous synconvergent “dynamic spreading”, or ductile extrusion. From U-Pb dating of variously transposed leucogranite, Scammell documented that high temperature top-to-the-east shear strain lasted several ten’s of my in the Cretaceous. There is a trend of cooling ages decreasing from northeast to southwest across strike of his study area. This trend is illustrated on cross-section A-B of Fig. 1.5a where titanite U-Pb ages decrease from ~98-90 Ma at Scrip Creek to ~74 Ma at Ruddock Creek, while $^{40}\text{Ar}/^{39}\text{Ar}$ ages decrease from ~75-66 Ma to ~63-61 Ma across the same transect. It is thus possible that cooling and exhumation were diachronous across the area.

Along cross-section A-B, the Columbia River Fault separates footwall kyanite

migmatites of the Monashee Complex, which were deformed and metamorphosed in the Paleocene-Eocene period, from hangingwall lower-amphibolite to greenschist facies rocks of the USA, which were deformed and metamorphosed in the Jurassic (Colpron et al., 1996; Gibson et al. 2005 & 2008). Sillimanite migmatitic schists that record penetrative east-verging deformation resurface in the footwall of the Selkirk detachment (Fig. 1.4) in the Selkirk Mountains, and are in lithologic, structural and metamorphic continuity with rocks located west of the Columbia River, near SC on Fig. 1.4. Numerous transposed leucosome and synkinematic leucogranite were formed/emplaced between 105-90 Ma, and this period also coincides with the timing of the thermal peak inferred from U-(Th)-Pb analyses of monazite from many migmatitic sillimanite-bearing schists (Sevigny et al., 1989; Scammell, 1993; Crowley et al., 2003; Gibson, 2003; Gibson et al., 2005). This suggests that peak metamorphic conditions were reached, and that a large amount of east-verging shear strain accumulated during this time period. The Selkirk detachment thus constitutes a major tectonic boundary with rock in the hangingwall having been deformed and metamorphosed in the Jurassic, whereas rocks in the footwall were deformed and metamorphosed in the Cretaceous. The minimal age of deformation is poorly constrained by the ~71-57 Ma crystallization ages of numerous undeformed leucogranite dykes that crosscut the transposition foliation and post-transposition folds (Sevigny et al., 1989; Scammell, 1993; Crowley et al., 2003; Gibson et al., 2005). Mica $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages range from ~55-52 Ma in the northern part of this transect (Scammell, 1993; Ghent and Villeneuve, 2006), which is only 2-6 my older than mica $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages obtained in the LSA cropping out to the west of the Monashee Complex (Fig. 1.5).

It is important to stress that isograds and map units are continuous *along strike* across the Columbia River from the Selkirk to the Monashee mountains, and that no tectonic boundary has been identified *across strike* toward the southwest in the Monashee Mountains (Fig. 1.4). Therefore, high-grade rocks located in the Selkirk and northern Monashee Mountains are physically linked to rocks located west of the Monashee Complex and they all share similar metamorphic, structural and geochronologic characteristics. The continuity of the LSa around the Monashee Complex and into the Selkirk Mountains is emphasized because it has been ignored in most conclusions of studies conducted in the southern part and south of the map of Fig. 1.4 (e.g. Parrish et al., 1988; Bardoux and Mareschal, 1994; Vanderhaeghe and Teyssier, 1997; Vanderhaeghe et al., 1999; Teyssier and Whitney, 2002; Teyssier et al., 2005). A key observation is that the northeasternmost exposure of LSa is not bounded by brittle-ductile normal faults, as is the case along section E-F (Fig. 1.4 & 1.4c). Another key implication is that the Selkirk detachment occupies the same structural position as the Okanagan Valley fault system, with USa in its hangingwall and LSa in its footwall. The two structures thus likely formed one structure that was eroded away above the Monashee Complex and down-faulted by the Columbia River Fault to the east of the complex, as suggested by Brown and Gibson (2006) and illustrated on cross-section A-B of Fig. 1.5a. Furthermore, from studies conducted near the Purcell Thrust (Fig. 1.4), Crowley et al. (2003) and Gibson (2003) documented that same migmatitic rocks of the LSa deformed and metamorphosed from Mid-Cretaceous up to Paleocene described above structurally overlying lower-amphibolite and greenschist facies rocks that record a main history of deformation and metamorphism between ~171 and 130 Ma. Boggs (2004) also obtained $^{40}\text{Ar}/^{39}\text{Ar}$

hornblende plateau ages of ~89 and ~83 Ma in the lower plate of the same area suggesting that the main period of cooling below 550-500°C took place while overlying rocks of the LSa were at their thermal peak. There would thus be another USa-LSa boundary at the bottom of the northeasternmost exposure of LSa, but this boundary would have the opposite polarity to that of the OVfs, with the USa in the footwall and LSa in the hangingwall. Determining the nature of this boundary will be a matter of future research because it passes within rocks inferred to have remained stitched together throughout their tectonic history that involved three generations of folding outlasted by the development of metamorphic isograds (Simony et al., 1980; Ghent et al., 1980; Raeside and Simony, 1983; Digel et al., 1998; Ghent and Simony, 2005). Finally, the western boundary of the LSa in the northern part of Fig. 1.4 is not well constrained. It is tentatively traced parallel to the North Thompson segment (NTS on Fig. 1.4) of the OVfs.

1.2.4 Summary

In summary, data presented on the map of Fig. 1.4 and on the cross-sections of Fig. 1.4 clearly demonstrate that the southeastern Canadian Cordillera can be subdivided in three distinct tectonometamorphic units based on distinct timing of metamorphism, ductile deformation and cooling. Indeed, rocks of the USa were metamorphosed, deformed and exhumed above the brittle-ductile transition prior to ~120 Ma, as indicated by the prevalence of black symbols on Fig. 1.4 and 1.5. Rocks of the LSa record protracted high-grade metamorphism and high-temperature deformation between ~120 and 60 Ma, as indicated by the prevalence of gray symbols on Fig. 1.4 and 1.5, and were exhumed above the brittle-ductile transition between ~75 and 54 Ma. And rocks of the Monashee Complex were metamorphosed, deformed and exhumed above the brittle-

ductile transition between ~60 and 49 Ma, as indicated by the prevalence of white symbols on Fig. 1.4 and 1.5. Although not as diagnostic as the above timing constraints, these three tectonometamorphic units are generally characterized by distinct metamorphic assemblages. The USa mainly comprises staurolite- and lower-grade rocks with a few localities at the sillimanite-grade. The LSa mainly comprises sillimanite-K-feldspar-grade rocks, but in the southern part of the map of Fig. 1.4 the LSa is at the muscovite-sillimanite-grade (although migmatitic), whereas in the northeastern part of Fig. 1.4, the LSa is at the muscovite-kyanite-grade and migmatitic. Rocks of the Monashee Complex are commonly migmatitic and contain kyanite, which is variously transformed into sillimanite. Strain patterns are not diagnostic because most rocks of the LSa and Monashee Complex and some areas of the USa are characterized by L-S tectonites with a stretching lineation that constantly plunges to the west-northwest to west-southwest. However, southwest-verging asymmetric minor folds and kinematic indicators are common in the USa, whereas east- to northeast-verging asymmetric minor folds and kinematic indicators prevail throughout the LSa and the Monashee Complex.

The boundary between these three units are generally well-defined shear zones, such as the brittle-ductile, top-to-the-west Okanagan Valley fault system (OVfs) between the USa and the structurally underlying LSa, the ductile top-to-the-east Monashee décollement between the LSa and the structurally underlying Monashee Complex, the brittle-ductile top-to-the-east Columbia River Fault between the USa and structurally underlying LSa or Monashee Complex. However, the boundaries between the LSa and the USa in the northern part of the map of Fig. 1.4 remain to be characterized. Consequently, this chapter demonstrates that it is possible to subdivide the southeastern

Canadian Cordillera into three main tectonometamorphic units based on quantitative and testable criteria. The location and nature of the boundaries between these units, and the criteria used for the subdivision could be refined and/or modified in the future with the acquisition of new data. In my opinion, it would be highly beneficial for the Cordilleran tectonic community to adopt this new subdivision as it would greatly facilitate the understanding of tectonic issues.

1.3 Tectonic models

The tectonic evolution of southeastern Canadian Cordillera has always been vigorously debated and is now more disputed than ever with proposed models as diverse as: 1) ductile thrusting and duplexing (Journeay, 1986; Spear, 2004; Carr and Simony, 2006); 2) progressive underthrusting in an orogenic wedge under critical taper (Parrish, 1995; Brown, 2004); 3) channel flow and/or ductile extrusion (Scammell, 1993; Johnston et al., 2000; Williams and Jiang, 2005; Brown and Gibson, 2006; Glombick et al., 2006a; Kuiper et al., 2006); 4) channel flow and diapirism toward a extensional rolling-hinge (e.g. Vanderhaeghe et al., 1999; Teyssier et al., 2005); and 5) exhumation by extension and crustal boudinage (Price and Monger, 2003; Kruse and Williams, 2007). The extensive and multidisciplinary database available for the SE Canadian Cordillera makes it an ideal natural laboratory to test these contrasting tectonic models, which are all a subject of active studies worldwide. The focus of this thesis is devoted to gaining a better understanding of the processes that led to the exhumation of high-grade rocks exposed in the southeastern Canadian Cordillera. In this section, each of the models is briefly described. In accordance with England and Molnar (1990) *exhumation* will refer to the upward vertical motion of rocks with respect to the surface. This is not to be confused

with *uplift* that is the upward vertical motion of rocks with respect to the geoid. Thus the term *exhumation* does not necessarily imply exposure of a rock at the surface.

Although large-magnitude extensional models are widely accepted (e.g. Coney and Harms, 1984; Parrish et al., 1988; Bardoux and Mareschal, 1994; Teyssier and Vanderhaeghe, 1997; Vanderhaeghe et al., 1999; Teyssier et al., 2005), other authors argued that at least part of the exhumation occurred during convergence (Journeay, 1986; Brown and Journeay, 1987; Scammell, 1993; Johnson and Brown, 1996; Johnston et al., 2000; Glombick et al., 2006a).

Some extensional models suggested that high-grade rocks were exhumed along crustal-scale ductile-brittle normal faults dipping outward from the metamorphic core (Fig. 1.6a). According to these models, a change in far-field stress after the orogen had been thickened to >50 km induced gravitational collapse and extension of the crust that was accommodated along crustal-scale normal faults (Parrish et al., 1988), or listric normal faults in the upper crust and ductile thinning in the lower crust (Bardoux and Mareschal, 1994). Restoration of balanced cross-sections indicated net extension of ~80-100 km (Parrish et al., 1988; Bardoux and Mareschal, 1994).

Other extensional models placed the emphasis on ductile flow of mid-crustal rocks during gravitational collapse of a thickened orogen (e.g. Block and Royden, 1990; Wernicke, 1992), and domes coring the metamorphic complex of southeastern Canadian Cordillera were interpreted as the product of extension-assisted diapirism (Vanderhaeghe et al., 1999; Norlander et al., 2002; Teyssier and Whitney, 2002). These diapiric models evolved into one of channel flow toward a rolling-hinge presented by Teyssier et al. (2005) who described a model that is very similar to the analogue models of Brun and

Van den Driessche (1994) and to the thermomechanical numerical models of core complex formation by Tirel et al. (2004 & 2008). According to the model of Teyssier et al. (2005), illustrated on Fig. 1.6b, extension initially localized at the thermal boundary between the cold craton and hot orogen (Hyndman and Lewis, 1999) and led to the formation of a master detachment. As the crust extended from the west, domino-style normal faults thinned the upper crust above the master detachment. Thinning of the upper crust created a pressure gradient that drove the flow of a hot and ductile middle crust. Coupling of mid-crustal flow and dissection of the upper crust by normal faulting resulted in the flattening and rotation of the master fault, effectively creating a rolling-hinge. West of the hinge, the combined effect of the upper crust moving westward and underlying mid-crustal rocks flowing eastward toward the thinned upper crust above the hinge created a top-to-the-west shear couple; whereas a top-to-the-east shear couple developed by the same process to the east of the hinge. Teyssier et al. (2005) argued that a diapir developed and flowed vertically en-masse toward the kinematic hinge at the climax of extension in the Eocene. Isotherms were bowed upward during the formation of the rolling hinge such that a transient brittle-ductile transition developed, above which mid-crustal rocks were progressively cooled. As the upper crust was pulled westward, progressively younger mid-crustal rocks were exhumed at the hinge.

In stark contrast with gravitational collapse models, other authors considered that exhumation was a two-stage process initially involving a period of synconvergent exhumation followed by extension. One of these models involves underthrusting of a stiff indenter and/or basement duplexing beneath the thrust sheet (Fig. 1.6c) that would have induced erosion and normal faulting at the surface to stabilize the gravitational instability

created by thickening (Journeay, 1986; Brown and Journeay, 1987; Spear, 2004; Carr and Simony, 2006). Exhumation by this model resulted from tectonic denudation and erosion. Furthermore, the entire crustal edifice would have moved as a block along a basal shear zone, although internal ductile deformation was also taking place.

Brown (2004) built on Parrish (1995) by integrating results of more than 25 years of active research in the southeastern Canadian Cordillera and proposed a model — inspired by Platt (1986) — involving progressive westward underplating beneath an orogenic wedge in critical taper. In this model, the location of the basal shear zone and isotherms are fixed with respect to an external reference frame and rock particles cycle through the wedge (Fig. 1.6d). Hence, underplated rocks get deformed and metamorphosed as they are accreted to the basal shear zone and are then exhumed by the combined effect of underplating from below and denudation — by extensional fault and erosion— at the surface to maintain the wedge in a critical taper state.

Finally, Scammell (1993) was the first to suggest a model of synconvergent ductile extrusion in the southeastern Canadian Cordillera (Fig. 1.6e); an idea that was revived by Johnston et al (2000), Brown and Gibson (2006), Glombick et al. (2006a) and Kuiper et al. (2006). Several papers of the volume “Channel flow, ductile extrusion and exhumation in continental collision zones” (2006) discussed the problem of semantics surrounding channel flow and ductile extrusion processes and highlighted the fact that there is no consensus on which term should be used (Godin et al., 2006b; Grasemann et al., 2006; Jones et al., 2006; Williams et al., 2006). In the absence of consensus, I will refer to *channel flow* as the process by which a kilometric layer of low-viscosity material leaves its lid and base behind as it flows *horizontally* down a pressure gradient in the

lower to middle crust. The term *ductile extrusion* will refer to the process by which a slice of deformable rock material is detached from its base and lid and flows towards the surface of the Earth, without necessarily reaching it. The upward translation of the slice is accommodated by ductile shear zones of opposite shear sense and by internal ductile deformation. This definition has no genetic implications with regards to the driving force; which could be: 1) a lateral pressure gradient induced by a topographic gradient or density contrast between two rock columns; 2) active erosion at the orogenic front; 3) convergence of two blocks that expels the rock mass between them; or 4) any combination of the above.

1.4 Study objectives

The aim of this research is to compile available data and acquire new key data that would allow for a reduction of the number of possible tectonic models in the southeastern Canadian Cordillera. The exhumation process of the LSa and the northern Monashee Complex is the main subject of this study. Because the rolling-hinge model of Teyssier et al. (2005; Fig. 1.6b) implies extensive mid-crustal ductile flow, whereas the progressive underthrusting model of Brown (2004; Fig. 1.6d) implies a base of Cordilleran penetrative strain, it is important to determine the extent of Cordilleran flow at the deepest exposed structural level. The key issue of the second chapter of this thesis is thus to determine whether there is penetrative ductile Cordilleran strain at the deepest exposed structural level of the southeastern Canadian Cordillera that crops out in the core of the Frenchman Cap dome. The first part of chapter 2 focuses on assessing whether all gray granite dykes in the core of the dome are part of a plutonic suite that could be used as a relative time marker of deformation. The second part presents new field relationships,

structural, metamorphic and geochronologic data from the lower basement level of Frenchman Cap dome to characterize the state and nature of Cordilleran strain.

The third chapter examines the tectonic evolution of the LSa, with special attention devoted to exhumation processes. Key observations are first extracted from an extensive multidisciplinary dataset and then a Pressure-Temperature-time-deformation (P-T-t-d) path is constructed for the Monashee décollement at the base of the LSa. This path provides information that was missing and allows testing the different tectonic models of exhumation described above (Fig. 1.6).

Finally, the fourth chapter investigates the exhumation process (es) of the Monashee Complex and also examines the structural evolution of two fold nappes. The approach taken in this chapter is to construct P-T-t paths across the two fold nappes by integrating available geochronologic data and results of phase equilibria modeling of migmatitic aluminous metapelitic schist, which locally contain the three Al_2SiO_5 polymorphs. Detailed particle paths are extracted from these P-T-t paths and included in a geodynamic model that shows the formation and exhumation of fold nappes by sequential ductile extrusion of partially molten rocks from middle to upper crustal conditions during underthrusting of a rigid basement.

1.5 Methodology

1.5.1 Field investigations

Frenchman Cap dome has been the subject of extensive mapping at various scales (Wheeler, 1965; McMillan, 1970; Fyles, 1970; Höy and Brown, 1980; Brown and Psutka, 1980; Journeay, 1986; Lane et al., 1989; Scammell, 1993; Gibson et al., 1999; Björnson, 2003); and of many geochronologic studies (Parrish, 1995; Crowley, 1999; Crowley and

Parrish, 1999; Gibson et al., 1999; Crowley et al., 2001; Foster et al., 2004). After reviewing literature data, I decided to diverge from the conventional approach that generally consists in mapping a delimited area and relating the structural, metamorphic and geochronologic information from this area to regional observations. Rather, field investigations consisted in filling holes on previous maps; mapping along new logging roads; collecting specimens for metamorphic investigations from areas where timing constraints were available; and revisiting key areas for collection of metamorphic specimens and verification of previous observations. Areas investigated during fieldwork are located on the simplified geological map of Fig. 1.7. Notably, the map of Frenchman Cap basement drawn by Journeay (1986) was extended further south to encompass the areas north of Bourne Glacier where Crowley et al. (2001 & 2008) collected their specimens for U-Pb geochronology (Fig. 1.7). Also, geochronologic data presented in Crowley and Parrish (1999) were placed into a metamorphic and structural framework by mapping a well-exposed logging road that runs perpendicular to the structural grain north of Ratchford Creek (Fig. 1.7). During ~5 months of fieldwork, detailed structural and metamorphic observations were collected at 490 stations. Data collected during fieldwork are reported in Appendix 1 and 2. I also had access to the collections of rock chips collected by Journeay (1986), many of which are stained for feldspar identification.

1.5.2 Petrography

Metamorphic paragenesis and microstructures were documented in 100 thin-sections from rocks collected during this study and by Crowley (1997). These observations combined with detailed observations made in the field lay the foundations for subsequent quantitative analyses.

1.5.3 *Electron microprobe*

Quantitative mineral composition analyses were determined on an automated 4 spectrometer Camebax MBX electron microprobe at Carleton University by the wavelength dispersive X-ray analysis method (WDX). Operating conditions were: 15kv accelerating potential and a beam current of 20 nano-amperes (nA). Peak counting times for each analyzed element were 15-30 seconds or 40,000 accumulated counts. Peak counting time for fluorine was 60 seconds. Background measurements were made at 50% peak counting time on each side of the analyzed peak. Raw X-ray data were converted to elemental weight % by the Cameca PAP matrix correction program. A suite of well characterized natural and synthetic minerals and compounds were used as calibration standards and secondary standards.

1.5.4 *Conventional geothermobarometry and strategy for quantitative data acquisition*

Conventional geothermobarometry is used in this study as a complement to thermodynamic forward modeling (see next section). Geobarometers and geothermometers used in this study consist of the garnet-plagioclase- Al_2SiO_5 geobarometer (Holdaway, 2001), the clinopyroxene-plagioclase-quartz geobarometer (SCan, McCarthy and Patiño-Douce, 1998), the Al-in hornblende geobarometer (Anderson and Smith, 1995), the garnet-biotite geothermometer (Holdaway, 2000), and the hornblende-plagioclase geothermometer (Holland and Blundy, 1994). The average P-T mode of Thermocalc, which uses convergence of many phase equilibria calculated with average mineral composition (Powell et al., 1998), was also used in cases of chemical equilibrium at the thin-section scale. In order to test whether or not the mineral compositions represented an equilibrium composition, the first step of electron

microprobe work was to acquire core and rim compositions of three to four grains of each of the minerals investigated from different domains of the thin-section. It was judged that chemical equilibrium at the thin-section scale had been reached if there was no significant variation of composition ($\pm 5\%$ variation) from rim to rim of each mineral phase analyzed or even better if there was complete homogenization of composition. Of all specimens analyzed, only specimen FG462d proved to be at equilibrium (cf. chapter 3). In examples of disequilibrium, it was important to pick analyses within microdomains that presented a possible link between a microstructure and a metamorphic reaction. A microstructure common to metapelitic schists of the study area is resorbed garnet porphyroblasts replaced by plagioclase, sillimanite and biotite in the asymmetric pressure shadows. Where such microstructure was observed, multiple analyses of garnet and plagioclase or garnet and biotite in mutual contact were conducted in order to get an estimate of pressure and temperature during the development of that microstructure. Where larger grains (a few 100s of μm) are present, compositions were obtained for spots located further from margins to check for zoning.

1.5.5 Thermodynamic modeling

Rock slabs, 20 cm x 10 cm x 0.5 cm, of unaltered and homogenous rock specimens were crushed to fine powder. The bulk composition of each specimen was obtained by a well-calibrated in-house XRF protocol at University of Ottawa.

The program *Perple_X'06* (Connolly, 1990; Connolly and Pettrini, 2002) was used to calculate isochemical P-T diagrams by thermodynamic forward modeling. The bases of this forward modeling approach are the laws of thermodynamics (see Spear, 2003 for a comprehensive treatment of the subject applied to metamorphic petrology). The Gibbs

phase rule states that at equilibrium, there is no exchange of chemical elements between phases (p_1 - p_n) and each phase is at the same pressure (P) and temperature (T). Hence, a system composed of X independent system components $c_1, c_2, c_3 \dots c_X$ can be defined at equilibrium by the following equations, where $G(c_1)_1$ is the Gibbs free energy of component 1 in phase 1, P_1 and T_1 are the pressure and temperature of phase 1, respectively:

$$P_1 = P_2 = P_3 \dots P_n$$

$$T_1 = T_2 = T_3 \dots T_n$$

$$G(c_1)_1 = G(c_1)_2 = G(c_1)_3 \dots G(c_1)_n$$

$$G(c_2)_1 = G(c_2)_2 = G(c_2)_3 \dots G(c_2)_n$$

$$G(c_3)_1 = G(c_3)_2 = G(c_3)_3 \dots G(c_3)_n$$

$$G(c_X)_1 = G(c_X)_2 = G(c_X)_3 \dots G(c_X)_n$$

In the above table, where p is the number of phases and c is the number of independent system components, there are $p(c+2)$ variables and $(c+2)(p-1)$ equations, where $c+2$ is the number of rows and $p-1$ is the number of equations in each row. An additional p equation arises because the P , T and G of each component cannot be independently varied in each phase. The variance, or degree of freedom f , of a system equals the number of variables minus the number of equations.

$$f = \text{n. variables} - \text{n. equations}$$

$$= (p(c+2)) - ((c+2)(p-1)+p)$$

$$f = c + 2 - p$$

Now, if the chemical composition of the system is fixed, the composition of each phase adds p variables and there is one mass-balance equation for each component that adds c equations. The variance then becomes:

$$f = c + 2 - p + p - c = 2$$

Therefore, when the bulk composition is known, definition of any two independent variables determines the state of the system, including the composition of all minerals and their modal abundances. When calculating a P-T diagram by thermodynamic forward modeling, the bulk chemical composition of the examined specimen is known and the computer program *Perple_X'06* calculates the metamorphic assemblage that minimize the Gibbs free energy of the system at each P-T point. However, such calculations are extremely complex and computationally expensive because the Gibbs free energy of solid-solutions is a logarithmic function. Thus, the basis of the algorithm is to first subdivide each solid-solution into “pseudocompounds” that are treated as homogenous phases during the calculations. Then the program estimates the composition of each solid-solution by linking a set of given pseudocompounds together. Finally, it divides the P-T field into polygons that contain distinct mineral assemblages. The level of details of the diagram depends in part on the spacing of solid-solution subdivision and on the spacing of the P-T grid. Given the computational power of modern computers, it is possible to calculate a P-T section in a reasonable amount of time (~2-10hr) with a level of detail that far surpasses the uncertainties associated with the definition of solid-solution models. The interested reader is directed to the work of Powell and Holland (1998) for the computer program *Thermocalc*, and Connolly, (1990), Connolly and

Petrini (2002) and Connolly (2005) for the program Perple_X. It is of note that various names have been suggested for the type of P-T diagram presented in this contribution. I choose to use the name *pseudosection* because it shortens writing space and it is the most common name used in the literature today.

Pseudosections were calculated in the extended ten components model system $\text{TiO}_2\text{-MnO-FeO-MgO-CaO-Na}_2\text{O-K}_2\text{O-Al}_2\text{O}_3\text{-SiO}_2\text{-H}_2\text{O}$ (TiMnCaNaKFMASH). TiO_2 is included in order to calculate phase equilibria involving ilmenite and rutile, from which important petrological information is obtained. Thermodynamic calculations were conducted with the software Perple_X'06 (Connolly, 1990; Connolly and Petrini, 2002) using a gridded Gibbs free-energy minimization approach (Connolly, 2005) and the internally consistent and updated thermodynamic database of Holland and Powell (1998, revised 2002). Most mineral solid-solutions are from the version of the Holland and Powell dataset that was available for Perple_X at the end of 2006. It includes the Ti-Fe-Mg-Mn-biotite solution model of Powell and Holland (1999), the Mica (CH2) white mica solution model modified from Coggon and Holland (2002), and the melt model of White et al. (2001). The Newton et al. (1981) feldspar solid-solution model was used instead of the Holland and Powell feldspar model upon recommendation given in the Perple_X solution model files, which can be found at:

www.perplex.ethz.ch/perplex_documentation.html#SOLUTION_MODEL_GLOSSARY.

Most pseudosections are composite, including the portion below 650°C calculated without a melt phase and saturated in a C-O-H fluid as defined by Connolly and Cesare (1993). The amount of H_2O calculated at 650°C and 700 MPa was then used to calculate the higher temperature portion of the pseudosection. The composite nature of the

pseudosections required minor adjustments that were accounted for by visually adjusting the fields to agree with Schreinemaker's rules. All pseudosections were calculated with saturated SiO_2 . Notice that Perple_X does not calculate uncertainty.

1.5.6 Garnet isopleth thermobarometry

It is possible to derive the composition of each mineral phase at any given P-T point of a pseudosection calculated with Perple_X'06 and to use these *isopleths* for thermobarometry. Because garnet is a solid-solution consisting of the four end-members almandine (alm for Fe), pyrope (pyr for Mg), grossular (grs for Ca), and spessartine (sps for Mn), any combination of three isopleths of these end-members are thermodynamically independent. Garnet isopleths that intersect within a small P-T space can thus be used as a geothermobarometer. This geothermobarometer is robust because it does not make any assumption related to the equilibrium between minerals, as is the case for conventional thermobarometry (e.g. Spear, 1993). However, diffusion at high temperature has to be taken into consideration for the interpretation of the results. For example, the composition of a garnet grain can be homogenized at high temperature and its composition can equilibrate with the matrix phases on the retrograde path. It could thus be possible to obtain an intersection of garnet isopleths that would indicate the P-T conditions at which the garnet stopped equilibrating with the matrix. These P-T conditions would thus yield a minimum estimate of the metamorphic peak.

1.5.7 Whole rock major and trace element geochemical analyses

Specimens, which had a size of at least 10 x 10 x 10 cm and were free of any altered surfaces, were crushed using a Chipmunk Tm jaw crusher and then ground in an agate ring mill to a fine powder. Pieces of both machines were removed and thoroughly

cleaned using acetone between specimens. The fine powder was mixed and divided in two fractions: one for XRF analysis, one for Sm/Nd and Rb/Sr isotopic analyses. Major elements and trace elements (Zn, Ba, Ce, Co, Cr, Ga, La, Nb, Nd, Ni, Pb, Rb, Sr, Th, U, V, Y Zr) were analyzed by X-Ray Fluorescence (XRF) on fused disks using a Philips PW 2400X-ray fluorescence spectrometer at University of Ottawa. XRF precision was calculated using an in-house protocol that yields accuracy better than 1% and 10% for major and trace elements, respectively.

1.5.8 Sm-Nd and Rb-Sr isotopic analyses

Procedural steps for Sr and Nd isotope analyses included spiking, acid washing, dissolution, column chromatography, filament loading, and mass spectrometry. Chemical and mass spectrometry analyses were performed at the Isotope Geochemistry and Geochronology Research Center (IGGRC) at Carleton University, Ottawa, Canada. The initial step consisted in adding a 96% pure enrichment of a mixed ^{148}Nd - ^{149}Sm spike to the specimen in order to measure the concentration of Nd and Sm, the latter of which was not measured by X-Ray Fluorescence. The mass spectrometer used was a Thermo Finnigan Triton T1, operated in static mode. The chemical procedures of Cousens (1996) are outlined below. Rock powders were dissolved in three steps: (1) 50% HF + 16N HNO_3 , (2) 7N HNO_3 , and (3) 6N HCl. Following complete dissolution, the dried residue from the Sr split was taken up in 2.5N HCl and the dried residue from the Nd split was taken up in 1N HBr. Ion-exchange column chromatography was used to isolate the elements of interest. The dissolved Sr split was centrifuged and loaded into 14-mL Bio-Rad borosilicate glass chromatography Sr columns with Dowex 50-X8 cation resin. Sr was eluted with 2.5N HCl. The rare earth elements (REE) were separated with the Sr

columns (as above), but using 6N HCl. The REE fraction, dried and dissolved in 0.26N HCl, was loaded into 14 mL Bio-Rad borosilicate glass chromatographic columns containing a 2 cm-high bed of Teflon powder coated with HDEHP [di(2-ethylhexyl) orthophosphoric acid]. Nd was then eluted using 0.26N HCl. The final batch of specimens was processed in similar fashion on a set of Eichrom 2ml pre-packed LN resin columns. Sr was loaded onto single Ta filaments with H_3PO_4 . Isotope ratios were normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.11940$ to correct for fractionation. Two Sr standards are run at the IGGRC, NBS987 ($^{87}\text{Sr}/^{86}\text{Sr} = 0.710258 \pm 22$, $n=16$) and the Eimer and Amend (E&A) SrCO_3 ($^{87}\text{Sr}/^{86}\text{Sr} = 0.708027$). Nd specimens were loaded on one side of a Re double filament with H_3PO_4 . Isotope ratios were normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.72190$ to correct for fractionation. Analyses of the Nd Std standard averaged $^{143}\text{Nd}/^{144}\text{Nd} = 0.511817 \pm 8$ ($n=28$). The uncertainties quoted for the standards are at the 2σ level. Results of Rb-Sr and Sm-Nd isotopic analyses were equivocal and are not discussed in the text. Rather, analytic data are reported in Appendix 3.

1.5.9 U-Pb spot analyses of zircon and titanite

1.5.9.1 U-Pb laser ablation – multiple collector – inductively coupled plasma – mass spectrometry (LA-MC-ICP-MS)

Zircon and titanite grains from six rock specimens were isolated using standard mineral separation techniques. The clearest, most euhedral and faceted zircon and titanite grains were hand picked under the binocular microscope. Grains were mounted in epoxy and polished to reveal their centres. Cathodoluminescence (CL) imaging was conducted using an Electron Optics Services system interfaced to the Camebax microprobe, at Carleton University, consisting of a high sensitivity photo-multiplier (PM)

tube connected to a Baush and Lomb Cathodoluminescence Amplifier. Images were collected at 512 x 512 pixels resolution with 100-200 microsecond/pixel dwell times with the 4Pi Analysis digital imaging system. CL images were collected sequentially at 15 kV accelerating potential and 15 nA beam current. Backscattered electron (BSE) images of the polished titanite were acquired with an electron microprobe (EMP), a JEOL 733 Superprobe at the Massachusetts Institute of Technology, with an accelerating voltage of 15 KeV, a beam current of 30 nA, and the secondary electron detector mounted in place of the optical microscope ocular. Isotopic data were acquired by laser ablation – multiple collector – inductively coupled plasma – mass spectrometry (LA-MC-ICP-MS) at the radiogenic isotope facility (RIF) of the University of Alberta, in Edmonton.

Instrumentation includes a Nu Plasma MC-ICP-MS, coupled to a Nd:YAG UP213 laser ablation system. The MC-ICP-MS instrument is equipped with a modified collector block containing 12 Faraday collectors and three ion counters (Simonetti et al., 2005).

Analytical protocol and instrumentation is described in Simonetti et al. (2006).

The first step of each U-Pb session was to analyze rims of many grains with unequivocal clear zoning, in order to avoid overlapping distinct zones. A rough approximation of the date of each analysis was calculated between two analyses. After the first round, other analyses were conducted on the most interesting grains in terms of dates and zoning. Most commonly, there was space for only one or two spots of the squared 40 μm wide laser beam on a given grain. A few analyses, labelled with a * in the tables of this thesis, were conducted with a 20 μm wide laser beam for small zones that were judged important.

1.5.9.2 Standards

Three zircon and one titanite standard have been routinely analyzed during the U-Pb analytic sessions, which lasted three days, and the results are presented on Table 1.2. JCA-50 is an in-house zircon standard from the Massachusetts Institute of Technology dated at ~ 3717 Ma (J. Crowley pers. communication 2008). The thirteen U-Pb analyses of this specimen were all less than 6% discordant between $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ ages, individual $^{207}\text{Pb}/^{206}\text{Pb}$ dates were all within 1.3% of the true age, and the weighted mean average $^{207}\text{Pb}/^{206}\text{Pb}$ age yielded 3713.4 ± 7.4 Ma (MSWD = 2.2; prob. 0.01), which is $\sim 0.1\%$ from the true age.

BJWP is another in-house zircon standard from the Massachusetts Institute of Technology dated at ~ 720 Ma (J. Crowley pers. communication). Three of the twelve U-Pb analyses of this specimen were more than 10% discordant, but the remaining nine analyses were less than 2.0% discordant. Individual $^{207}\text{Pb}/^{206}\text{Pb}$ dates were all within 2.0% of the true age (discarding the three discordant analyses). Individual $^{206}\text{Pb}/^{238}\text{U}$ dates were all within 2.6% of the true age (including the three discordant analyses). The weighted mean average $^{207}\text{Pb}/^{206}\text{Pb}$ age yielded 719 ± 19 Ma (MSWD = 3.4; prob. 0.00), which is $\sim 0.14\%$ from the true age. The weighted mean average $^{206}\text{Pb}/^{238}\text{U}$ age yielded 717.0 ± 6.6 Ma (MSWD = 0.61; prob. 0.82), which is $\sim 0.42\%$ from the true age.

Temora 1 is the Phanerozoic zircon standard dated by Black et al. (2003) at 416.8 ± 1.1 Ma. Eight of the 29 U-Pb analyses of this specimen were more than 10% discordant. Individual $^{206}\text{Pb}/^{238}\text{U}$ dates were all within 5 % of the true age (including the discordant analyses), and the weighted mean average $^{206}\text{Pb}/^{238}\text{U}$ age yielded 415.9 ± 4.0 Ma (MSWD = 2.4; prob. 0.00), which is $\sim 0.22\%$ from the true age.

The titanite standard BLR-1 dated by Aleinikoff et al., (2007) at 1047.1 ± 0.4 Ma

was routinely analyzed. Titanite can incorporate a large amount of common Pb in its structure and an appropriate correction is important, especially for analytic methods that does not measure ^{204}Pb as the method used herein. Two methods have been proposed to correct for common lead in titanite (see Aleinikoff et al., 2004 and Storey et al., 2006). The first is to regress the uncorrected data on a Terra-Wasserburg semi-total Pb plot. If the data form an isochron, it can be assumed that the intercept on the y-axis yields the initial common Pb ratio, whereas the lower intercept on the concordia curve yields the age of the titanite. The second method is to calculate the weighted mean average of individual $^{206}\text{Pb}/^{238}\text{U}$ dates corrected with the *Age7corr* algorithm in Isoplot 3.3 (Age7corr, Ludwig, 2003), while using the y-intercept on the Terra Wasserburg plot and its uncertainty for the common Pb correction. The second method has the disadvantage of assuming concordancy of the analyses, but yields better precision. Discarding one of the 11 analyses of BLR-1 standard, which was clearly off the regression line on the Terra-Wasserburg plot, yielded an age of 1041 ± 110 Ma (MSWD = 0.79; prob. = 0.62) for the first method; and a weighted mean average of 1043 ± 35 Ma (MSWD = 0.09; prob. = 1.00) for the second method. Individual dates for the second method were all within 2.5% of the true value, whereas the weighted mean is ~0.4% from the true value. In summary, routine analyses of three zircon and one titanite standards demonstrate that the accuracy of individual dates were well within 5% of the true age whereas the weighted mean average were all within 1% of the accepted value. Hence, it is not surprising that the ages of the unknown analyzed in this study yielded uncertainties of a few my for Paleocene-Eocene rocks.

Table 1.1. References for Fig. 1.4 and 1.5

Cross-section A-B			Cross-section C-D-E			Cross-section F-G		
Location*	U-Pb	Ar/Ar AFT	Location*	U-Pb	Ar/Ar AFT	Location*	U-Pb	Ar/Ar
Pukeashun Mt. (PM)	11,	11	Salmon Arm (SA)		10	Vernon	2	3
MD north FCD (MD)	13, 21		OVfs Sicamous	11, 12	9, 11	Vernon antiform (VA)	2, 7	1
FCD-Sybley (SS)	13, 22, 23, 24, 26	25	Hunters Range (HR)	11	11	Trinity Hills (TH)	9	
FCD-Mt. Grace (MG)	22, 26		MD Three Valley Gap (TVG)	12, 13, 14, 15	16	Pinnacles (Pnc)	4, 5	9
Ruddock crk. (RC)	21	21	TOD NW	14, 15		MD Cariboo Alps (CA)	4, 35	16
Horne crk. (HC)	21	21	CRF Mt. Hall		17	TOD core (CR)	6, 3, 8	
Scrip crk. (SC)	21	21	Clachna-cudainn (Cldn)	18	19	CRF Pinnacles (Pnc)	7	
SFA	30, 31		Illecillewaet synclinorium (Ilcw)	20	20	Kuskana-x Batholith (KB)	27	
Selkirk-Monashee Mountains	28, 29 30, 31, 34	32						
Kinbasket Lk. (KL)	34, 30, 31, 34	33						

* Abbreviations used on the map of Fig. 1.4 are in bold and in parentheses.

Abbreviations: MD: Monashee décollement; FCD: Frenchman Cap dome; SFA: Selkirk Fan Axis; OVfs: Okanagan Valley fault system; TOD: Thor-Odin dome; CRF: Columbia River Fault. References: 1: Bardoux, 1993; 2: Glombick et al., 2006; 3: Vanderhaeghe et al., 1999; 4: Carr, 1992; 5: Lemieux, 2006; 6: Hinchey et al., 2006; 7: Parrish et al., 1988; 8: Teyssier et al., 2005; 9: Vanderhaeghe et al., 2003; 10: Lorencak et al., 2001; 11: Johnson, 1994; 12: Gilley, 1999; 13: Parrish, 1995; 14: Kuiper, 2003; 15: Johnston et al., 2000; 16: Sparks, 2001; 17: Mulch et al., 2006; 18: Crowley and Brown, 1994; 19: Colpron et al., 1999; 20: Colpron et al., 1996; 21: Scammell, 1993; 22: Crowley and Parrish, 1999; 23: Foster et al., 2004; 24: Gibson et al., 1999; 25: Sandborn, 1996; 26: Crowley et al., 2001; 27: Parrish and Wheeler, 1983; 28: Sevigny et al., 1990; 29: Digelet et al., 1998; 30: Gibson et al., 2005; 31: Gibson et al., 2008; 32: Ghent and Villeneuve, 2006; 33: Boggs, 2004; 34: Crowley et al., 2003; 35: Coleman, 1990.

Table 1.2. U-Pb analyses of zircon and titanite standards

	True age (Ma)	Mean $^{207}\text{Pb}/^{206}\text{Pb}$ age (Ma)	Max. % from true age individual dates	Mean $^{206}\text{Pb}/^{238}\text{U}$ age (Ma)	Max. % from true age individual dates
JCA-50 zircon	~3717	3713 ± 7 (95% conf.) 1 of 13 rejected MSWD = 2.2 prob. = 0.01 0.1% from true	1.3	3603 ± 50 (95% conf.) 1 of 13 rejected MSWD = 1.6 prob. = 0.10 2.9% from true	5.9
BJWP-1 zircon	~720	719.2 ± 19 (95% conf.) 2 of 12 rejected MSWD = 3.4 prob. = 0.000 0.1% from true	< 2.0 (1 date at 10.7)	717.0 ± 6.6 (95% conf.) 0 of 12 rejected MSWD = 0.61 prob. = 0.82 0.4% from true	2.6
Temora-1 zircon	~417	402 ± 17 (95% conf.) 3 of 29 rejected MSWD = 6.6 prob. = 0.00 3.6% from true	33.7	415.9 ± 4 (95% conf.) 0 of 29 rejected MSWD = 2.4 prob. = 0.00 0.2% from true	4.8
BLR-1 Titanite	~1047	n/a	n/a	1043 ± 35 (95% conf.) 1 of 11 rejected MSWD = 0.09 prob. = 1.00 0.4% from true	2.5

Figure 1.1a

Figure 1.1b

Figure 1.1. Physiographic map and geophysical cross-section of the southern Canadian Cordillera.

a) Physiographic map of the southern Canadian Cordillera subdivided into the five main morphogeological belts of the Canadian Cordillera. The location of the study area is indicated by a square. b) Cross-section from Burianyk and Kanasewisch (1995) showing the P-wave velocity subsurface structures (color pattern) obtained from seismic refraction surveys, and the main seismic reflectors. Notice the good fit between the outcropping structures, the trace of seismic reflectors and P-wave velocity domains. See text for further explanation.

Figure 1.2. The five morphogeological belts of the Canadian Cordillera.

The dark gray area labeled BB is the current location of the Bowser/Sustut basin and associated Skeena Fold Belt. The pale gray, stippled area, labeled BB, is the location of the Bowser/Sustut basin and associated Skeena Fold Belt in the Cretaceous estimated by Gabrielse et al. (2006) after restoration of ~800 km of dextral strike-slip motion.

Figure 1.3. Lithostratigraphic map of the southeastern Canadian Cordillera surrounding the Monashee Complex.

Modified from Gibson (2003).

LEGEND (Map & cross-sections)

▨ Devonian, Jurassic and Cretaceous Plutons

⊕ Paleocene Ladybird leucogranite

Tectonometamorphic units

▨ Upper Selkirk allochthon (Usa)

▨ Lower Selkirk allochthon (Lsa)

Monashee Complex

■ Cover sequence: metasediments infolded with basement slices

▨ Basement: polymetamorphosed para-orthogneiss

Geochronologic data**Type of dates**

▲ Deformed intrusions (U-Pb zrc, mnz)

◆ Thermal peak schist (U-Pb mnz)

● Crosscutting dykes (U-Pb zrc, mnz)

■ Hbl cooling ages or
b:biotite, m: muscovite, t: titanite, ap: apatite
Ar/ Ar for hbl, b, m; U-Pb for t;
fission track for ap

Colour chart

□ < 59-47 Ma

▨ 120-60 Ma

■ > 120 Ma

Structural symbols

⌋ Normal brittle-ductile fault

▼ Thrust-sense ductile shear zone

Figure 1.4. Revised tectonometamorphic map of the southeastern Canadian Cordillera surrounding the Monashee Complex. Inset (above) shows the five morphogeological belts of the Canadian Cordillera. Symbols on the map correspond to the collection site of specimens analyzed for geochronology, see legend for signification of symbols, Fig. 1.5 for value of ages and Table 1.1 (p. 38) for references. A-B, C-D and E-F are the traces of the cross-sections shown on Fig. 1.5. See Table 1.1 p.38 for abbreviations.

Figure 1.4

Figure 1.5. Cross-sections on which all available ages have been projected. See text for explanation. Legend is the same as for Fig. 1.4 except for: 1) the black fill that represents the mafic lower crust inferred from seismic refraction profiles (Burianyk and Kanesevich, 1995); and 2) for the basement of the Monashee Complex (X pattern) that only includes rocks inferred to have escaped Cordilleran tectonism, whereas on the map of Fig. 1.4 basement refers to lithologic units irrespective of their timing of deformation.

Figure 1.5

Figure 1.6. Exhumation models investigated in this study. See text for explanation. In d) MD3 represents the active shear zone, whereas MD1 & 2 are shear zones that are being passively exhumed by accretion of material from below and denudation at the surface.

a) Large-magnitude extension (as in Parrish et al., 1988; Bardoux and Mareschal, 1994)

b) Flow toward a rolling-hinge (as in Teyssier et al., 2005)

c) Ductile thrusting above an indenter (as in Spear, 2004; Carr and Simony, 2006)

d) Progressive underthrusting in critical taper (as in Brown, 2004)

e) Channel flow and ductile extrusion (as in Brown and Gibson, 2006; Glombick et al., 2006)

Legend

- Crustal contraction
- Crustal extension
- Sense of shear (top direction)
- Erosion
- Isotherms

- Low-grade rocks
- Medium-grade rock
- Migmatites cooled below brittle-ductile transition
- Flowing migmatites

Figure 1.6

Figure 1.7. Location of the field camps visited during this study. The map shows the basal quartzite that delimits basement rocks from cover sequence rocks of Frenchman Cap dome.

Figure 1.7

Figure 1.7. Location of field camps visited during this study

CHAPTER 2
CORDILLERAN OROGENIC BASE EXPOSED IN THE CORE OF
FRENCHMAN CAP DOME OF THE MONASHEE COMPLEX,
SOUTHEASTERN CANADIAN CORDILLERA

ABSTRACT

The aim of this study is to determine whether basement gneisses in the core of Frenchman Cap dome of the Monashee Complex record evidence of flow during the Cordilleran orogeny and to derive ensuing tectonic implications. It is first demonstrated that gray granite dykes previously dated at ~1850 Ma do form a distinctive suite throughout basement gneisses of the dome that can be used to distinguish Cordilleran from Precambrian strain. Accordingly, only a ~1.5 km structurally thick carapace of basement gneiss records penetrative Cordilleran strain, whereas the lowermost ~5 km thick basement section preserves a Paleoproterozoic foliation. Cordilleran deformation at the lower basement level identified during this study is limited to a meter-scale, top-to-the-east shear zone, and to N-S trending, upright folds developed at depths of less than ~15-20 km, respectively. Furthermore, new and previously published U-Pb dating of zircon, monazite and titanite suggest that contractional Cordilleran deformation at the lower structural levels of Frenchman Cap dome took place between ~52 and ~49 Ma. These characteristics suggest that the core of Frenchman Cap dome represents a downward-migrating base of Cordilleran strain frozen at the end of the orogeny. Preservation of a Paleoproterozoic foliation in its core implies that Frenchman Cap gneiss dome was not formed by extension-assisted diapirism.

2.1 Introduction

The core of the southern Canadian Cordillera is one of the best areas worldwide to understand the tectonic evolution of a large (>900 km wide) orogen because of relatively easy access and amazing outcrop quality provided by rapidly retreating glaciers. The tectonic history of the two domes of the Monashee Complex is of utmost importance for the understanding of the tectonic evolution of the southern Canadian Cordillera because the domes occupy the deepest exposed structural level of the orogen. Unfortunately, studies conducted in the northern dome (Frenchman Cap) point to a drastically different history than studies conducted in the southern dome (Thor-Odin). Crowley et al. (2001) documented a downward-decreasing strain gradient in the core of the northern dome from the deformation state of Paleoproterozoic gray granite and pegmatite dykes. By contrast, Vanderhaeghe et al. (1999) documented an Eocene metatexite/diatexite transition toward the core of the southern dome, and Norlander et al. (2002) revealed the isothermal decompression path followed by these Eocene migmatites. Given these contradictions, it is not surprising that tectonic models from the northern dome converged towards progressive downward migration of the base of convergent ductile shearing with time (Parrish, 1995; Gibson et al., 1999; Crowley et al., 2001; Brown, 2004); whereas models from the southern dome converged towards gravitational collapse of a thickened orogen accompanied by mid-crustal ductile flow toward a rolling-hinge developed under large-magnitude extension (Vanderhaeghe and Teyssier, 1997; Vanderhaeghe et al., 1999; Teyssier and Whitney, 2002; Teyssier et al., 2005). Faced with conflicting data, Vanderhaeghe et al. (1999) raised the question of whether the Paleoproterozoic crystallization age of the gray granite and pegmatite dykes dated by Crowley et al. (2001)

by conventional U-Pb geochronologic methods (ID-TIMS) could in fact reflect inheritance. This hypothesis has recently been repudiated by Crowley et al. (2008) who conducted additional U-Pb dating of similar dykes from the same locality as Crowley et al. (2001). Alternatively, Kuiper et al. (2006) suggested that rocks from this locality could in fact be a mega-scale boudin that floated in flowing rocks during Cordilleran deformation.

The first aim of this contribution is to determine whether the gray granite dykes dated by Crowley et al. (2001 & 2008) can be used as time markers to determine the relative age of deformation throughout the basement of the northern dome of the Monashee Complex. The second aim is to better characterize the strain gradient reported by Crowley et al. (2001). Finally, the nature and age of deformation at the deepest structural level of the dome will be assessed. Results of this study demonstrate that the strain gradient reported by Crowley et al. (2001) is in fact a relatively sharp boundary between a completely transposed orthogneissic basement and a ~4-5 km thick structural section of basement that preserves a Paleoproterozoic foliation. At these structurally deep exposures, strain identified as Cordilleran is limited to open folding and meter-scale shear zones developed in schistose slivers at a relatively shallow depth of less than ~15-20 km. These features are not compatible with models of gravitational collapse, large magnitude extension and extensive ductile flow of the middle crust.

2.2 Tectonic setting

See Chapter 1.

2.3 Geological setting of Frenchman Cap dome

The Monashee Complex (Fig. 1.4, p. 47) exposes the deepest structural level of the

Canadian Cordillera and consists of two gneiss domes: Frenchman Cap (FCD) in the north and Thor-Odin in the south. The complex is fault-bounded, with a westerly-dipping, top-to-the-east ductile shear zone called the Monashee décollement on its west flank; and an easterly-dipping, top-to-the-east brittle-ductile shear zone called the Columbia River Fault on its east flank (Fig. 2.1). FCD is cored by Paleoproterozoic gneiss and mantled by a metasedimentary cover sequence inferred to have been deposited on the passive margin of North America in the Precambrian (see Scammell and Brown, 1990 and Crowley, 1999). The base of the sequence consists of a quartzite layer that delineates the shape of both domes, and thus constitutes an important structural marker (Fig. 2.1). The cover sequence is infolded with transposed basement slices, but only basement rocks are present below the basal quartzite in the core of FCD (Fig. 2.1). In this thesis, FCD is divided into four structural levels that are illustrated on the cross-section of Fig. 2.2. The upper level of the cover sequence occurs immediately below the Monashee décollement, has a structural thickness of ~3 km, and comprises two fold nappes. This structural level records east-vergent, penetrative ductile deformation and high-grade metamorphism between ~60 and ~55 Ma (Parrish, 1995; Crowley and Parrish, 1999; Gibson et al., 1999; Crowley et al., 2001; Foster et al., 2004). The lower level of the cover sequence has a structural thickness of ~1.5 km and comprises the upright, lower limb of the lower fold nappe. This structural level records high-temperature deformation until ~50 Ma, which is 5 ± 2 my later than the upper level (Crowley and Parrish, 1999; Crowley et al., 2001). The upper basement level consists of Paleoproterozoic para- and orthogneisses that record penetrative Eocene deformation (see below), whereas the lower basement level consists of the same lithologic units that do not record penetrative

Cordilleran deformation (Crowley et al., 2001; Crowley et al., 2008). It will be demonstrated in this paper that the upper basement level has a structural thickness of ~1.5 km, whereas the exposed lower basement level reaches a structural thickness of ~4-5 km. All four structural levels are exposed on the moderately steep west flank of FCD, whereas the upper level of the cover sequence is not exposed on the gently dipping, east flank of the dome, which suggests a termination of the upper fold nappe above the surface exposure of the complex.

2.3.1 Structural geology at the lower level of the cover sequence and upper basement level

Journey (1986) mapped and conducted a structural analysis of the lower level of the cover sequence and of the upper basement level. As illustrated on Fig. 2.3c & f, a striking structural feature of the area is the cluster of SW-plunging stretching lineations and minor fold axes that Journey (1986) interpreted as resulting from rotation of fold axes during top-to-the-northeast shearing. As illustrated on Fig. 2.2, it is noteworthy that such parallelism is a characteristic structural feature throughout the ~10-15 km thick LSa (Journey, 1986; Scammell, 1993; chapter 3 of this thesis), which overlies the Monashee Complex, and throughout the upper and lower cover sequence of the complex (Journey, 1986). Other important structural features of the lower level of the cover sequence and of the upper basement level are northeast-verging, inclined folds that deform the transposition foliation and lineation. For simplicity, these folds will be referred to as F_3 folds. These folds were only reported where the dominant transposition foliation dips toward the NE to SE, but are absent where the dominant transposition foliation dips toward the W to SW. Journey (1986) interpreted the prevalence of F_3 folds on the east

flank of the dome as evidence for doming during east-directed shearing that would have placed the easterly-dipping flank of the dome in the shortening field of the strain ellipsoid — developing late folds— while the westerly-dipping flank would have remained in the extensional field — continuing to be flattened and boudinaged.

2.3.2 Timing of metamorphism and deformation at lower level of the cover sequence and upper basement level

The timing of deformation at the lower level of the cover sequence and upper basement level has been well constrained by Crowley et al. (2001) and Crowley and Parrish (1999). Thirty-two U-Pb analyses consisting of single or group of monazite grains from nine pelitic schists, collected at the lower level of the cover sequence and at the upper basement level, plot on or above concordia and yielded dates between ~61 and ~49 Ma (Crowley and Parrish, 1999), with 23 of these analyses clustering between ~53 and ~49 Ma. These results suggest that monazite started to grow during prograde metamorphism between ~61 and ~53 Ma and that thermal peak was reached between ~53 and ~49 Ma. Timing of deformation is also tightly bracketed at these two structural levels. Foliated and boudinaged pegmatite layers that are folded by F_3 northeast-verging folds yielded concordant to slightly discordant zircon U-Pb dates of 51 ± 1 Ma interpreted as the crystallization age (specimen 12, Crowley et al., 2001). A sheared and boudinaged sillimanite-bearing leucosome yielded a concordant U-Pb age of 50.5 ± 1.0 Ma, also interpreted as the crystallization age (specimen 258, Crowley and Parrish, 1999). An undeformed pegmatite, highly discordant to the gneissosity of its host basement gneiss, yielded a poorly defined linear array of zircon U-Pb analyses with a lower intercept coinciding with a concordant analysis at 49.5 ± 0.5 Ma, which was also

interpreted as the crystallization age. As argued by Crowley et al. (2001), this dataset indicates that at least part of the transposition fabric (foliation, gneissosity, lineation, isoclinal folds) and F_3 folds developed between ~51 and 49 Ma, which coincides with the latest period of monazite growth in pelitic schists.

2.3.3 *Timing of deformation at lower basement level*

Conventional (ID-TIMS) U-Pb geochronology conducted on specimens collected less than 2 km structurally below the basement-cover contact pointed to a drastically different tectonic history (Crowley et al., 2001). Three pegmatite layers and one gray granite sheet that each crosscut the migmatitic gneissosity of their host orthogneisses at a high angle yielded highly discordant monazite and zircon U-Pb analyses that each define a linear array with upper intercepts of: 1848 ± 8 Ma (gray granite PCA-305B-83), 1834 ± 9 Ma (pegmatite JC-91-93), 1845 ± 13 Ma (pegmatite JC-84-93), and 1845 ± 13 Ma (pegmatite JC-308-93; uncertainties modified from Crowley, 1999 by multiplying the 1σ analytical uncertainty by the square root of the MSWD and by the Student's-t factor, following Ludwig, 2003). The array formed by monazite and zircon analyses from gray granite specimen PCA-305B-83 also defines a linear array with discordant titanite analyses from another gray granite dyke (JC-324-93) collected at the upper basement level. Crowley (1997) argued that the gray granite sheets form a distinctive suite, which he called Bourne suite, emplaced at ~1850 Ma. Crowley et al. (2001) used Bourne granite as a time marker to distinguish Precambrian from Cordilleran deformation at basement levels of FCD. However, because of the high discordance of most U-Pb analyses, which were conducted on whole grain or group of grains, it was possible that the upper concordia intercepts dated the inherited source rather than the crystallization age. This

alternative became more likely with the high-resolution U-Pb dating (with SHRIMP) conducted in the Thor-Odin dome of southern Monashee Complex that demonstrated the composite nature of zircon grains from anatectic leucosome that generally consist of Paleoproterozoic cores surrounded by Tertiary rims (Vanderhaeghe et al., 1999; Kuiper, 2003; Hinchey et al., 2005).

In order to test the inheritance hypothesis for pegmatite and gray granite dykes from FCD, Crowley et al. (2008) investigated three pegmatite and two gray granite dykes from the northern toe of Bourne Glacier: the same area where earlier studies were conducted (area K on Fig. 2.3). Results of that study repudiated the inheritance hypothesis from three independent tests comprising field relationships, composition and zoning of monazite and zircon, and U- (Th)-Pb age of monazite and zircon. Interestingly, pegmatite JC-308-93 was recollected for high-resolution U-Pb analyses by laser ablation (LA-ICP-MS). This pegmatite contains blobs of a gray granitic phase similar in composition and texture to other gray granites. Sixteen zircon analyses yielded the same age within uncertainty as that of the earlier study (1856 ± 10 Ma by Crowley et al. 2008 vs. 1845 ± 13 Ma by Crowley 1997). In contrast, the 10 monazite analyses from that pegmatite yielded Eocene dates with a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 50.7 ± 1.8 Ma (MSWD = 0.7), which is the same as two monazite U-Pb analyses (composed of 2 and 4 grains) from an adjacent pelitic schist that yielded concordant dates of ~ 51.5 – 49.5 Ma (Crowley, 1999). The Crowley et al. (2008) study demonstrated the absence of Cordilleran penetrative ductile deformation at one locality of the lower basement level of FCD, and supported the contention that gray granite dykes at this locality are part of the Bourne suite (Crowley, 1999). However, it did not imply that these two statements apply

to the entire lower basement of FCD.

2.4 The Bourne granite suite

Structural analysis of polydeformed terrains is not a simple task. A useful approach is to examine the strain state of different suites of dykes, provided that reliable criteria exist to link a given dyke to a specific suite. The Crowley (1997), Crowley et al. (2001) and Crowley et al. (2008) studies suggested that gray granite sheets intruding basement gneisses of FCD are part of a suite that can be used as a time marker of deformation. However, this conclusion was based on similar ages of only three granitic sheets from the basement area covered on the map of Fig. 2.3, which represents only $\sim 1/10$ of the FCD area covered by basement gneiss (Fig. 2.1). In order to investigate whether all gray granite intruding basement gneiss of FCD are part of a ~ 1850 Ma suite, several investigations were undertaken at different locations of FCD basement (see Fig. 2.1 for locations of fieldwork). First, I documented the mineral assemblage, macroscopic textures, and field relationships with host gneisses of all gray dykes encountered during the ~ 12 weeks of fieldwork conducted in basement exposures of FCD. Then, 15 thin-sections were examined with an optical microscope and from these, 10 thin-sections were closely inspected with a scanning electron microscope (SEM). Four specimens, including one pegmatitic leucogranite commonly associated with the gray granite, as specimen JC-308, were analyzed by XRF for their bulk chemical composition. One of these specimens was analyzed for U-Pb geochronology with the high-resolution laser ablation technique.

2.4.1 *Field relationships, petrography and trace element profiles*

Bourne granite is distinguished in the field by a light gray color, a medium-grained and equigranular texture, and (at lower basement levels) by a weak to absent alignment of

minerals interpreted as a lack of solid-state fabric (Fig. 2.4a & c; criteria from Vernon, 2004 pp.456-463). They generally form sheets 0.5-1 m thick, but veinlets 2 cm thick and stocks occupying surfaces of ten's of square meters have also been observed. No systematic orientation of the sheets was noted. Another characteristic feature is the association between the gray granite and a coarse-grained to pegmatitic leucogranite (Fig. 2.4b). The pegmatitic leucogranite is commonly present at the margins of gray dykes but heterogeneous mixing of the two phases has also been observed. The common lobate contacts between these two phases and their local heterogeneous mixing (Fig. 2.4b) are interpreted as evidence for magma mingling. At Perry River locality (Fig. 2.1 for location), these two granitic phases are commonly associated with lenses of ultramafic composition. In fact, the largest volume of gray granite observed in FCD was in the aureole of an ultramafic complex at Perry River. More work would be needed to determine whether there is a cogenetic link between the granite and the ultramafic unit.

Petrographically, Bourne granite is characterized by: 1) a mineral assemblage formed by approximately 20% quartz, 25% plagioclase, 35% K-feldspar, 15% biotite and 5% allanite and in some specimens trace amount of monazite or titanite; 2) biotite having a very dark color that masks birefringence; 3) distinctive and abundant large dark-brown allanite (3-5 mm in diameter). Close inspection of 10 thin-sections of Bourne granite with the SEM revealed that most samples contain subhedral allanite with rare-earth elements (REE) zoning. They have REE-rich cores and REE-poor rims (Fig. 2.4d), features that have been interpreted as evidence for a magmatic origin (e.g. Beard et al., 2006). In contrast, some dykes contain oscillatory-zoned monazite as a magmatic phase instead of allanite (cf. Crowley et al., 2008). Monazite in these dykes is generally rimmed by a

double corona formed by an inner shell of apatite + thorite and an outer shell of allanite $\pm$ phengite (cf. Fig. 2.4c for an example of a single allanite coronae). The presence of magmatic monazite in some gray granites and magmatic allanite in others could indicate that they belong to two distinct suites or reflect a slightly more calcic composition for the allanite-bearing dykes (Cuney and Friedrich, 1987; Dini et al., 2004). It is thus important to determine if both types of gray granites belong to the same suite or not.

Three gray granite sheets, one from each investigated locality of the lower basement level, and one pegmatitic leucogranite mingled with a gray granite phase (from Perry River locality) were analyzed by XRF to derive their trace element profiles (Table 2.1). Inconclusive results from Sm-Nd and Rb-Sr isotopic analyses conducted on felsic and mafic basement rocks are presented in Appendix 2. Fig. 2.5 shows that the three gray granites have the exact same trace element profiles and that the white leucogranite has peaks and troughs for the same elements as that of the gray granite. Although this dataset is limited, it supports the hypothesis that gray granites from three different localities of the lower basement level of FCD are from the same suite.

2.4.2 *U-Pb geochronology*

Gray granite specimen JC-50-05 collected at the Perry River locality (see Fig. 2.1 for location), which is ~ 2 km structurally below the Bourne Glacier locality and ~ 1 -2 km structurally above the Frisby Creek locality, was dated using the high-resolution laser ablation MC-ICP-MS technique (see methodology in chapter 1 and Simonetti et al., 2005). At the collection site, this gray granite is heterogeneously mixed with the pegmatitic leucogranite (Fig. 2.4b) and with lenses of ultramafic igneous rocks, which together form a stock a few ten's of meters wide. In the field, the gray granite appears to

have a lineation defined by aligned K-feldspar, but not under the microscope where the granite has an equigranular texture with equant grains of quartz, very similar to the inverse-colored SEM picture of Fig. 2.4c. Importantly, no monazite was extracted from this sample. Sample JC-50-05 rather contains ~5% allanite crystals that are euhedral to subhedral, present a resorbed texture at their margins, and are zoned in REE (Fig. 2.4d). The shape of the grains and the zoning is interpreted as evidence for a magmatic origin. Specimen JC-50-05 thus constitutes an ideal candidate to determine whether allanite-bearing gray granites are related to the monazite-bearing gray granites dated by Crowley (1997) and Crowley et al. (2008). Titanite is a rare phase from this specimen that is present only in mm-scale isolated leucocratic pockets.

Twenty-one zircon grains extracted from the gray granite JC-50-05 were analyzed for U-Pb geochronology by the laser ablation MC-ICP-MS technique (see introduction for methodology and Simonetti et al., 2005). Cathodoluminescence (CL) images reveal that zircon is generally sector or oscillatory-zoned (Fig. 2.6b), but that this zoning is commonly blurred. CL images also show thin black rims around most of the grains (Fig. 2.6b, c). Eight zircon analyses yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates between ~1943 and 2098 Ma that are interpreted as inherited (Table 2.2). Twelve zircon analyses yield discordant (7-19%) U-Pb dates that define a linear array (MSWD = 2.3; prob. = 0.01) with an upper intercept of 1861 ± 58 Ma and a lower intercept of 344 ± 400 Ma (Fig. 2.6a; Table 2.2). Alternatively, forcing the lower intercept to 49 ± 1.1 Ma, which is the titanite age in specimen JC-50-05 (see below), yields an upper intercept of 1825 ± 16 Ma with a higher MSWD (4.7) and a null probability of fit. The former age of ~1861 Ma is interpreted as the crystallization age of the gray granite because of better statistical results and because

it is closer in age to dykes dated by Crowley et al. (2008). Interestingly, it was possible to use a 20µm laser beam to analyse the thin black rims (CL image) common to many zircon grains from the gray granite specimens examined by Crowley et al. (2008) and herein. The two analyses of black rims yield dates indistinguishable from dates obtained from oscillatory-zoned parts (Fig. 2.6c, Table 2.2), confirming that no part of zircon grains from the Bourne suite grew in the Tertiary.

Five titanite grains were also analyzed for U-Pb geochronology with the laser ablation MC-ICP-MS technique (Fig. 2.7; Table 2.4). Titanite grains commonly have a diameter of ~500µm, are subhedral and present patchy and planar zoning (Fig. 2.7a, b). Titanite naturally contains a significant amount of common Pb. Thus, great care must be taken in the method used for the common Pb correction (see Storey et al., 2006). Three methods can be used to correct for initial common Pb in isotopic techniques that cannot precisely measure ^{204}Pb such as the LA-MC-ICP-MS technique used herein (Aleinikoff et al., 2004). The first is to regress the uncorrected data on a Terra-Wasserburg semi-total $^{207}\text{Pb}/^{206}\text{Pb}$ vs. $^{238}\text{U}/^{206}\text{Pb}$ plot and use the lower concordia intercept as the age of titanite growth. The second method is to use the y- $^{207}\text{Pb}/^{206}\text{Pb}$ intercept on the Terra-Wasserburg plot and calculate the weighted average of $^{206}\text{Pb}/^{238}\text{U}$ dates corrected by the ^{207}Pb correction method (c.f. Stern, 1997; Williams, 1998). The third method is to use the model terrestrial Pb curve of Stacey and Kramers (1975) as the initial $^{207}\text{Pb}/^{206}\text{Pb}$ for the ^{207}Pb correction method. The last two methods have the disadvantage of assuming concordancy of U-Pb analyses, but yield more precise results. Grain M04 yields distinct data as that of grains M12, M13, M14 and M18 (Table 2.3). Because only three analyses were collected for grain M04, it is not possible to estimate the amount of

common Pb other than using the Stacey and Kramers (1975) method. By using a value of common Pb corresponding to 1850 Ma, three analyses from one titanite grain yield dates of ~1707, 1715 and 1399 Ma. This grain must have formed during crystallization of the granite and preserved its Paleoproterozoic isotopic information. The second method for common Pb correction is used for the remaining analyses (Aleinikoff et al., 2004; Storey et al., 2006). Regression of six uncorrected analyses from four grains on a Terra-Wasserburg plot yields a lower intercept of 49.2 ± 2.6 Ma (MSWD= 5.3; prob. fit = 0.0) and yields a y-axis ($^{207}\text{Pb}/^{206}\text{Pb}$) intercept of 0.82 ± 0.05 . Using the y-axis value of $^{207}\text{Pb}/^{206}\text{Pb}$ for common Pb correction yields a corrected weighted mean average $^{206}\text{Pb}/^{238}\text{U}$ age of 49.3 ± 1.1 Ma (MSWD =0.90; prob. fit = 0.48). The agreement between the two methods suggests that the analyses are concordant and the ~49 Ma age is interpreted as the time at which these titanite grains closed for Pb diffusion. It is not clear whether these grains crystallized in the Paleoproterozoic and were completely reset in the Eocene or whether they crystallized in the Eocene.

2.4.3 Interpretation

Combined with previous studies, it is possible to derive reliable criteria for identifying gray granite dykes of the Bourne suite and use them as time markers of deformation. Including Crowley (1999), Crowley et al. (2008) and this study, five gray granite specimens (also including pegmatitic leucogranite specimen JC-308 that contains blobs of Bourne granite) have been analyzed for U-Pb geochronology and yielded crystallization ages clustering between ~1848 and 1861 Ma (Fig. 2.8). Combining these five ages yields a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1852 ± 5 Ma (MSWD =0.73; prob. fit = 0.57). These dated granites cover the entire spectrum of gray granite types including:

a monazite-bearing gray granite (JC-19-03), a monazite rimmed by allanite-bearing gray granite (JC-13-03), an allanite-bearing gray granite mingled with a pegmatitic leucogranite and ultramafic lenses (JC-50-05-05), a pegmatitic white granite including blobs of gray granite (JC-308-03), and a stock of titanite-bearing gray granite (JC-324-93). In addition, these granites were collected at various locations of the upper and lower basement levels of FCD. Trace element profiles for three gray granite specimens from three locations of the lower basement level of FCD, including the ~1861 Ma specimen JC-50-03 dated herein and one pegmatitic leucogranite mingled with a gray granitic phase, are very similar and suggest that all gray granites share a similar chemical composition. In the field, gray granites are easily distinguished by their gray color, equigranular and medium-grained texture, their discordant relationship with the gneissosity in the host gneiss, and a common association with a pegmatitic leucogranite. Microscopically, gray granites are characterized by the presence of zoned allanite crystals (commonly obvious under polarized light) or by coronas of allanite, surrounding resorbed monazite, and by a very dark color of biotite, that masks birefringence. Consequently, macroscopic, microscopic, geochemical, and most importantly geochronologic data demonstrate that gray granite dykes intruding basement gneiss of FCD are part of a plutonic dyke suite emplaced at ~1850 Ma. Furthermore, preservation of Paleoproterozoic isotopic information in some titanite grains, but new crystallization and/or complete resetting of other grains at ~49 Ma, strongly suggest that the Bourne suite was not heated significantly until a few millions of years before ~49 Ma. This conclusion is the same as that reached by Armstrong et al. (1991), Parrish (1995) and Crowley and Parrish (1999). Consequently, structures older than the Bourne granite are

considered Paleoproterozoic, whereas structures younger than the suite are considered Eocene in age.

2.5 Field investigation

Most of Frenchman Cap dome has been mapped at least at the 1:50 000 scale and its structural geometry has been well characterized (McMillan, 1970; Fyles, 1970; Höy and Brown, 1980; Brown and Psutka, 1980; Journeay, 1986; Lane et al., 1989; Scammell, 1993; Gibson et al., 1999; Björnson, 2003). An exception is the core zone that has only been mapped at the 1: 253 440 scale by Wheeler (1965). Four superbly exposed areas from Frenchman Cap dome (FCD) were investigated in the course of this study (Fig. 1.7, p. 52 and 2.1). The upper basement level was investigated in an area located south of Pettipiece Lake where the contact between basement and overlying metasedimentary rocks of the Monashee Complex is well exposed. Three localities from the lower basement level were investigated. The Bourne Glacier, Perry River and Frisby Creek localities are located approximately 1.5-3, 4-5 and 6-7 kilometers, respectively, structurally below the cover-basement contact.

The upper basement level was mapped at the 1: 50 000 scale by Journeay (1986). Hence, fieldwork at this structural level consisted in comparing strain patterns and deformation history between cover and basement rocks along the cover-basement contact south of Pettipiece Lake (Fig. 2.1). The map of Journeay (1986) was extended further south to encompass the areas north of Bourne Glacier where Crowley et al. (2001 & 2008) collected their specimens for U-Pb geochronology (area K on Fig. 2.3). The complex geology of the Perry River and Frisby Creek localities were not mapped in detail. Rather, fieldwork in these two areas focussed on documenting intrusive

relationships between Bourne granite dykes and their host gneisses.

2.6 Lithologic units at upper and lower basement levels of Frenchman Cap dome

South of Pettipiece Lake (Fig. 2.3), the cover sequence consists of a 500-1000m thick layer of basal quartzite containing a varying amount of muscovite (5-20%). Below the cover sequence, basement is dominated by orthogneiss. The oldest lithologic units are a ~2.2 Ga quartz diorite and a <2.1 Ga paragneiss (Crowley, 1999). These two units crop out as rafts, lenses and slivers within a ~2.1 Ga (Crowley, 1999) biotite-hornblende ± garnet granodioritic augen gneiss (unit 2d of Journeay, 1986) that constitutes the volumetrically dominant unit throughout the basement. The volume of leucosome and the strain level in the gneiss are variable. Accordingly, mapping in outcrops identified as unit 2c in Journeay (1986) revealed that this unit is more likely a zone of more deformed augen gneiss rather than a different unit and both units 2c and 2d of Journeay were regrouped into one “granodioritic augen gneiss” unit on Fig. 2.3. The alaskitic gneiss units 2a and 2b of Journeay consisting of leucocratic gneiss containing more than 20% and less than 20% amphibolite, respectively, were regrouped into a single unit on the map of Fig. 2.3. It was possible to map these units at Bourne Glacier, but not at Frisby and Perry River localities because lithologic units there are interlayered at the meter scale. Amphibolites are intrusive in the leucogranitic gneiss and generally do not contain garnet or clinopyroxene porphyroblasts. Clinopyroxene- or clinopyroxene-garnet-leucosome are locally present but because they invariably occur at the margin of dykes in contact with migmatitic orthogneiss, they are interpreted as the product of a reaction between the partially molten host gneiss and the amphibolite, a process that has been experimentally documented by Patiño-Douce (1995).

As shown on the map of Fig. 2.3, interlayered augen granodioritic orthogneisses and paragneiss layers strike in a N-S direction north of Bourne Creek. Mapping of the glacially polished and continuous outcrops south of Bourne Creek failed to match corresponding units. There is thus a structural break approximately coinciding with Bourne Creek. Importantly, Bourne Creek runs parallel to the projection of the transition zone of Cordilleran strain described below.

2.7 Structural Geology

2.7.1 Upper basement level: south of Pettipiece lake

The numerous geochronologic studies conducted in Frenchman Cap dome (FCD) have demonstrated that rocks of the cover sequence have only been deformed and metamorphosed during the Cordilleran orogeny (Parrish, 1995; Crowley and Parrish, 1999; Gibson et al., 1999; Crowley et al, 2001; Foster et al., 2004). Deformation of the basal quartzite can thus be compared with deformation of basement rocks to determine qualitatively the level of Cordilleran strain experienced by basement gneiss. In the relatively small area examined here, I considered it appropriate to use a sequential notation of deformation F_1 , F_2 ... to compare the deformation history of basement and cover rocks. The use of this notation scheme does not imply that folds belong to distinct deformation events, as it is acknowledged that deformation was most likely progressive and involved rotation of fold axes toward the direction of minimum shortening. South of Pettipiece Lake (Fig. 2.3), laminations and bedding in the basal quartzite are isoclinally folded by F_1 folds. F_1 imparted a penetrative foliation dipping generally to the NE, and a well-developed NNE-plunging lineation. S_1 foliation is refolded by NE-verging F_2 folds characterized by a gently dipping long limb and a steeply dipping short limb, an axial

plane with a steeper dip than the foliation, and a fold axis trending NW-SE (Fig. 2.3b). This type of fold has often been described as “cascading” and reported elsewhere as F_3 (e.g. Brown and Psutka, 1980; Journeay, 1986; Höy, 1987). Basement rocks, including Bourne granite dykes, share the same deformation history as the basal quartzite except that an earlier foliation is folded by the first set of isoclinal folds. F_1 in the cover is thus F_2 in the basement. Bourne granite dykes are commonly oblique to an early foliation, but are thoroughly folded, foliated and lineated as can be observed on Fig. 2.3a and Fig. 2.9e. Journeay (1986) conducted detailed mapping and structural data collection from this area as well as in the adjacent area labelled “J” on Fig. 2.3e where the westerly dip of the regional foliation allows a direct comparison with area K of the lower basement level. Lower hemisphere, equal area projections from area J highlight the colinearity between SW-plunging L_2 stretching lineations (Fig. 2.3c) and F_2 fold axes (Fig. 2.3f). This colinearity is a pattern indicative of a high level of strain that is common to all overlying high-grade rocks of the Monashee Complex and LSa at the latitude of Frenchman Cap dome (Fig. 2.2; Journeay, 1986; Scammell, 1986; Scammell, 1993; Gibson et al., 1999; Johnson, 2006).

2.7.2 *Transition zone*

At the head of Bourne creek (Fig. 2.3e) the transition from the upper basement level that records penetrative Cordilleran strain to the lower basement level that does not, is approximately 600 m thick (between stippled lines on Fig. 2.3d). Unfortunately, this zone crops out on a precipitous cliff that precluded a detailed structural analysis. In contrast with the transposition foliation and the well-developed lineation characterizing rocks of the upper level, rocks of the transition zone are not strongly lineated but are

intensely folded. All folds observed had an easterly vergence despite variation in the attitude of their axial planes and axes. These variations are interpreted as reflecting different degrees of fold rotation during easterly-directed simple shearing. Fig. 2.9b illustrates a folded granodioritic augen gneiss typical of this zone that contrasts with typical gneiss from the upper (Fig. 2.9a) and lower (Fig. 2.9c) basement levels. As mentioned above, the transition zone coincides with the abrupt truncation of interlayered units striking ~N-S north of Bourne Creek, but not present south of Bourne Creek.

2.7.3 *Lower basement level: Bourne Glacier locality*

Basement rocks located structurally below the transition zone at the toe of the Bourne Glacier show a strikingly different structural style than rocks above. In the granodioritic augen gneiss units, cores of primary K-feldspar megacrysts are commonly preserved at lower basement level, whereas they are entirely recrystallized at upper basement levels (compare Fig. 2.9a vs. 2.9c and 2.9g). Rims of these primary megacrysts have been recrystallized into augen that are slightly stretched, but not as much as the long cigar-shaped augen commonly observed at the upper basement level. These differences strongly suggest a lower amount of finite strain at lower basement levels. The rare folds observed at Bourne Glacier are generally upright and symmetric with axes plunging gently toward the northwest. Throughout overlying rocks of the Lower Selkirk allochthon and of the Monashee Complex, folds varying between immature, northeast-verging folds with axial surfaces having a steep westerly dip, and axial-planar mature folds having axes rotated into sub-parallelism with the lineation are common. Importantly, such folds, which can be qualified as "Cordilleran", were not observed at lower basement levels. The stark contrast between upper and lower basement levels is best illustrated on lower

hemisphere equal-area projections of fold axes and lineations (encompassing stretching, mineral and intersection lineations) from both structural levels presented on Fig. 2.3. Lineations from Bourne Glacier spread on a $>180^\circ$ girdle (Fig. 2.3d), whereas lineations from the upper basement level cluster in the southwest quadrant (Fig. 2.3c). Fold axes from Bourne Glacier form a girdle within the northwest quadrant (Fig. 2.3g), whereas fold axes from the upper basement level cluster in the southwest quadrant, parallel to lineations (Fig. 2.3f).

The lower finite Cordilleran strain at the Bourne Glacier locality is clearly indicated by the deformation state of the Bourne granite dykes. Bourne granite dykes are generally highly discordant to the gneissosity in their host. (Fig. 2.9g) and have been observed to crosscut limbs of folds at a high angle (Fig. 2.9f). Bourne granite is weakly deformed and exhibits only a weak mineral alignment. Fig. 2.4c shows an inverse-color SEM image of a Bourne granite specimen JC-13-03 cut parallel to the weak mineral alignment observed in the dyke. It is clear from this image that quartz grains are, at most, slightly deformed, and that the igneous texture is largely preserved.

Finally, results of Crowley et al. (2008) demonstrated that the migmatitic foliation at Bourne Glacier formed between ~ 1925 Ma, the age of an isoclinally folded leucocratic layer axial-planar to the layering (Fig. 2.10) and ~ 1850 Ma, the age of the Bourne dyke suite that crosscuts this transposition fabric at a high angle within the same outcrop.

2.7.4 Lower basement level: Perry River and Frisby Creek localities

Kuiper et al. (2006) suggested that rocks from the Bourne Glacier locality could in fact be a mega-scale boudin, suspended in flowing rocks during Cordilleran deformation. In order to test this hypothesis, fieldwork was conducted at the two deepest structural

levels exposed in FCD. The Perry River locality is ~4-5 kilometers structurally below the basement-cover contact, whereas the Frisby Creek locality is ~6-7 kilometers structurally below the contact. At both localities, the pre-Bourne granite history is extremely complex and its documentation is beyond the scope of this study. Bourne granite dykes are ubiquitous at both localities and crosscut the migmatitic gneissosity and associated axial-planar folds at a high angle (Fig. 2.11a, c). Furthermore, no lineation was observed at Frisby Creek while the E-W lineation observed at some outcrops of Perry River is cut by Bourne granite dykes as shown on Fig. 2.11d. Consequently, regardless of local complexities described below, there is no foliation or gneissosity of Cordilleran age within a 6-7 km thick section of basement cropping out below the transition zone described above.

2.8 Localized Cordilleran strain at the lower basement level

2.8.1 Top-to-the-east shear zone at the Bourne Glacier locality

2.8.1.1 Field relationships

Despite evidence presented above for the absence of penetrative Cordilleran ductile strain at Bourne Glacier, a thin, 10-20 m thick, shear zone in a paragneiss sliver can be confidently attributed to Cordilleran deformation. The zone is characterized by abundant kinematic indicators, such as C-S fabric, shear bands and oblique boudin trains (Hanmer and Passchier, 1991) all indicating a top-to-the-east sense of shear (Fig. 2.12a). Shearing was apparently limited to softer mica-rich slivers because the surrounding orthogneiss was not affected. Pegmatite dyke specimen JC-308-03 dated by Crowley et al. (2008) at ~1856 Ma is trapped and dismembered into oblique boudins in the zone, demonstrating that shearing was younger than ~1856 Ma. The youngest monazite from a pelitic schist

located within meters of the shear zone yielded dates of ~51.5-49.5 Ma that most likely reflect timing of shearing because no growth of monazite or zircon occurred between ~1850 Ma and ~52 Ma at this locality (Crowley and Parrish, 1999; Crowley and Ghent, 1999; Crowley et al., 2008).

2.8.1.2 Metamorphic conditions

Metamorphic conditions during shearing can be derived by combining results of phase equilibria modeling with results from conventional thermobarometry. Fg-204 is a (muscovite) garnet-sillimanite-biotite schist collected within the shear zone. Garnet occurs as trace amounts of small, <2 mm long, crystals having an ellipsoidal shape aligned parallel to the lineation. Because it is localized and commonly crosscuts the foliation, muscovite is interpreted as a late phase that was not stable at the thermal peak. Sillimanite in the specimen is mainly localized in finer-grained mylonitic layers defining C planes in a top-to-the-east C-S fabric. Garnet is unzoned, has a composition of $X_{alm} = 0.79$, $X_{pyr} = 0.08$, $X_{sps} = 0.09$, $X_{grs} = 0.04$. Biotite has a uniform composition of $Mg/Mg+Fe = 0.37$. Matrix plagioclase composition is richer in anorthite ($X_{an} = 0.27$) than plagioclase in contact with garnet ($An = 0.24$). Applying the garnet-biotite thermometer and garnet-plagioclase-sillimanite-quartz barometer, recently calibrated by Holdaway (2000 and 2001, respectively), yields a pressure of 369 ± 80 MPa at 609 ± 25 °C (Table 4.3). The uniform composition of biotite and garnet suggests that equilibrium was attained, and the absence of zoning in garnet suggests that neither retrograde net-transfer reactions, nor Fe-Mg exchange with biotite have taken place.

Fig. 2.13 is a phase diagram section calculated with the computer program *Perple_X'06* (Connolly and Petrini, 2002; Connolly, 2005) for the bulk composition of

this specimen (Table 2.1) in the system TiMnNaCaKFMASH including the melt model of White et al. (2001) and the Ti-Mn biotite solution model of Powell and Holland (1999). Importantly, measured garnet isopleths ($X_{\text{pyr}} = 0.08$, $X_{\text{sps}} = 0.09$, $X_{\text{grs}} = 0.04$) converge toward a P-T field centred at ~ 350 MPa/ 625°C , near the field composed of garnet-biotite-sillimanite that is interpreted as the metamorphic assemblage stable at during shearing. This agreement between reverse and forward modeling strongly suggests that top-to-the-east shearing took place at relatively low pressure (below ~ 400 MPa) indicating less than ~ 14 kilometers depth (assuming a crustal density of 2900 kg/m^3).

2.8.2 *Upright folding and partial melting at lower basement levels*

2.8.2.1 Field relationships and structural geology

At the deepest exposed structural level of FCD, localized upright folds postdate the intrusion of Bourne granite dykes. At the Frisby Creek locality upright, open folds with shallow axes and steep, NNW-SSE axial planes fold the migmatitic gneissosity and Bourne granite dykes (Fig. 2.11b, c). Granitic leucosome containing coarse euhedral hornblende >5 mm in diameter and small garnet <2 mm in diameter are common and parallel to the NNW-SSE axial planes of the folds (indicated by arrows on Fig. 2.11). Fig. 2.14 is a lower hemisphere equal area projection of structural data from the Frisby Creek locality. It shows that the gneissosity is regionally folded into upright, SSE-plunging folds with axes that lie within the plane of steeply dipping, NNW-SSE striking hornblende-bearing leucosome. A common, apparently paradoxical, observation is presented on Fig. 2.11a and Fig. 2.11b. Fig. 2.11a shows an outcrop with a Bourne granite dyke oriented at a high angle to the migmatitic gneissosity that cuts across both limbs of an axial-planar fold. There is no upright fold and no hornblende-bearing

leucosome. By contrast, ~10 m further east, within the same perfectly exposed outcrop, upright folds and axial-planar hornblende-bearing leucosome veins abound, and another Bourne granite dyke (sample Fg-425 analyzed for its chemical composition) is oriented parallel to the axial surface of the upright folds (Fig. 2.11b). This kind of heterogeneous distribution of upright folds and associated hornblende-bearing leucosome interspaced with areas where Bourne granite dykes cut the gneissosity at a high angle is a striking characteristic from both Frisby Creek and Perry River localities. These observations imply that the migmatitic gneissosity is older than the emplacement age of the Bourne granite suite at ~1850 Ma, but that the upright folds and associated hornblende-bearing leucosome are younger than ~1850 Ma.

2.8.2.2 U-Pb geochronology of hornblende-bearing leucosome veins

Zircon grains from two hornblende-bearing leucosome veins from Perry River and Frisby Creek localities were analyzed for U-Pb geochronology by the laser ablation MC-ICP-MS technique (see chapter 1 for methodology and Simonetti et al., 2005). All but one zircon grain is very similar. They are euhedral, have an elongated prismatic shape, present clear oscillatory zoning from core to rim, a bright CL image, and numerous inclusions (Fig. 2.15a). Thirty-five analyses from thirty grains define a discordia with upper and lower intercepts of 2114 ± 27 Ma and 564 ± 140 Ma, respectively (MSWD = 0.74; prob. = 0.86; Fig. 2.15c). One zircon grain is different than the others in that it has an inherited core, a dark color on CL image, and faint oscillatory zoning (Fig. 2.15b). One analysis from that grain is concordant and two other analyses slightly discordant. The three analyses yield a weighted mean average $^{206}\text{Pb}/^{238}\text{U}$ age of 52.7 ± 1.1 Ma (MSWD = 0.47; prob. = 0.63).

2.8.2.3 Interpretation

Upright folding and leucosome formation at the deepest exposed structural level of FCD is interpreted to result from Eocene fluid-assisted melting. The age of the hornblende-bearing leucosome analyzed for U-Pb geochronology cannot be provided by the upper intercept of the discordia defined by thirty-five analyses because the ~2114 Ma age is older by >300 my than the emplacement age of the Bourne granite dykes that predate formation of the leucosome and associated upright folds. All zircon grains but one would thus be inherited from the host granodioritic orthogneiss that crystallized at ~2114 Ma. Hence, the ~53 Ma age of the lone Eocene zircon has to be considered as the crystallization age of the leucosome and associated folding. Contamination during mineral separation is judged extremely unlikely because the process was conducted under the rigorous protocol of the Massachusetts Institute of Technology and that no rock of this age was processed in this lab beforehand (J.L. Crowley, pers. comm. 2005). In addition, this age is consistent with the timing of deformation and metamorphic peak at upper basement levels (Crowley and Parrish, 1999; Crowley et al., 2001), as well as with a minor episode of monazite and titanite crystallization within the lower basement level (Crowley, 1999; Crowley et al., 2008; Fig. 2.7)

The age of leucosome in FCD would thus coincide with the crystallization age of leucosome in Thor-Odin dome (Vanderhaeghe et al., 1999; Hinchey et al., 2006). It is tempting to conclude that core rocks from both Thor-Odin and Frenchman Cap domes were extensively melted in Early Eocene and were thus extremely ductile. This hypothesis is unlikely, however, for the following reasons. First, N-S trending folds and leucosome in FCD are not ubiquitous, but rather localized. In between partially molten

and folded zones, Bourne granite dykes are undeformed and cut the migmatitic gneissosity at a high angle (Fig. 2.11a, c, d). Even in reworked zones, gneissic layering is commonly only weakly disrupted (Fig. 2.11b). Such strain localization is not expected in migmatitic terrains affected by dehydration melting. In these types of migmatites, melting starts at grain boundaries and the melt phase migrates to dilatant sites with deformation (e.g. Sawyer, 1999; Rosenberg and Berger, 2001, Brown, 2007). Rosenberg and Handy (2005) showed that a significant reduction of strength occurs when melt volume reaches ~7%, at which point the migmatite should flow and develop penetrative ductile fabrics. This is not what is observed in FCD where deformation is limited to open folds and discrete leucosomes. I argue that melting was rather induced by, and limited to zones of fluid infiltration in a dry orthogneiss.

Independent evidence for genesis of leucosome via fluid-assisted melting is the abundance of coarse hornblende in leucosome. It is possible to estimate metamorphic conditions at Frisby Creek from conditions of ~625°C/350 MPa calculated for the Bourne Glacier locality. Assuming a density of ~2800 kg/m³ for the felsic orthogneiss forming the ~4-5 km structurally thick section of crust between Bourne Glacier and Frisby Creek, and a geothermal gradient of ~15°C/km, peak metamorphic conditions at Frisby Creek would have been ~685-700°C at 500-600 MPa. The 800 MPa partial melting experiment of Naney and Swanson (1980) indicates that hornblende is a peritectic product of partial melting of granodiorite at ~700°C, provided that at least 5 wt% H₂O is added to the system. Naney and Swanson used a synthetic Fe-Mg granodioritic composition that is most likely comparable to that of the granodioritic augen gneiss exposed in FCD. Gardien et al. (2000) also concluded that a significant amount of water is required to

produce hornblende from partial melting of a quartzo-feldspathic gneiss. Thus, these studies demonstrated that at temperature of $\sim 700^{\circ}\text{C}$, a significant amount of water must have been introduced to the dry granodioritic augen gneiss to produce the coarse-grained hornblende phenocrysts observed in the leucosome of the lower basement level of FCD. In a context of open folding, it is likely that fluid would migrate toward sheared limbs and/or hinges of incipient folds. This should lead to a feedback relationship in that fluid-assisted melting would lead to strain partitioning and further migration of fluid toward areas of higher strain. Therefore, fluid-assisted melting could result in the heterogeneous distribution of partially molten and folded zones observed at Frisby Creek. The source of hydrous fluid in a dry orthogneissic basement is problematic and would require further investigations.

2.9 Discussion

2.9.1 The base of northeast-directed shearing

Data presented herein complement data presented in Crowley et al. (2008) and Crowley et al. (2001) and together demonstrate that gray granite dykes crosscutting the gneissosity of basement gneiss in Frenchman Cap dome (FCD) are part of the Bourne granite suite that was emplaced at ~ 1850 Ma. The Bourne granite suite was then used as time markers to show that only the upper ~ 1.5 km of FCD basement has a foliation/gneissosity of Cordilleran age. Below this carapace of high Cordilleran strain, Bourne granite dykes crosscut the gneissosity throughout the exposed 4-5 km thick lower basement level. The boundary between these two structural levels is transitional and characterized by abundant easterly-verging folds with few lineations. Waning of Cordilleran strain with structural depth is not related to strain partitioning into softer

lithologic units because it entirely occurs within the same granodioritic augen gneiss unit. This boundary also coincides with a major change of strain patterns. Indeed, above the transition zone, a 17 to 22 km thick structural section comprising the LSa, the upper and lower levels of the cover sequence and the upper basement level of the Monashee Complex is characterized by a well-developed, W- to SW plunging stretching lineation toward which minor fold axes have been almost completely rotated (Fig. 2.2) during easterly-directed shearing (Journeay, 1986; Scammell, 1986 & 1993; Johnston et al., 2000; Williams and Jiang, 2005; Kruse, 2007). In stark contrast, gneisses of the lower basement level of FCD have weak lineations that do not cluster on lower-hemisphere equal area projections and no kinematic indicators were observed.

Furthermore, the ~50 Ma northeast-verging F_3 folds on a northeasterly-dipping gneissosity at upper basement level, the ~52-49 Ma top-to-the-east westerly-dipping shear zone at Bourne glacier, and the ~53-49 Ma NNW-SSE trending upright folds on a southerly-dipping gneissosity at Frisby Creek (Fig. 2.14) are strain patterns compatible with E-W contraction occurring between ~53 and 49 Ma. This age range coincides or shortly predates: 1) the ~53-48 Ma $^{40}\text{Ar}/^{39}\text{Ar}$ mica cooling ages recorded throughout the LSa and the Monashee Complex (Scammell, 1993; Bardoux, 1993; Johnson, 1994; Sanborn, 1996; Vanderhaeghe et al., 2003; Ghent and Villeneuve, 2006); 2) the ~51-47 Ma bracket for the last increments of ductile shearing along the two outward-dipping brittle-ductile normal shear zones that bound these two tectonometamorphic units (Glombick et al., 2006a and references therein; Mulch et al., 2006); and 3) the ~48 Ma emplacement age of undeformed lamprophyre dykes with chilled margins that are ubiquitous throughout the Monashee Complex (Adams et al., 2005). Given that the

timing of E-W contraction at upper and lower basement levels of FCD immediately preceded rapid regional cooling that marked the end of orogeny in the southeastern Canadian Cordillera, it is most likely that the deformation patterns observed in FCD today are a snapshot of a downward-migrating base of east-vergent shearing frozen at the end of orogeny. Accordingly, the upper basement level would represent the lowermost transposed structural level, the transition zone would represent a less advanced stage of transposition, and the lower basement level would represent initial stage of deformation with localized deformation zones characterized by upright folds and thrust shear zones. If deformation had continued, gneisses of the lower basement level would most likely have been completely transposed like gneisses of the upper basement level. This downward-decreasing gradient of Cordilleran strain can thus be considered as the orogenic front at the base of the Canadian Cordillera, analogous to the Grenville Tectonic Front in eastern Canada (e.g. Bethune, 1997).

2.9.2 *Implication for dome forming mechanism*

In recent years, a lot of attention has been put on the domes of the Monashee Complex as being typical extensional gneiss domes resulting from extension-assisted diapirism (Vanderhaeghe et al., 1999; Norlander et al., 2002; Vanderhaeghe and Teyssier, 1997; Teyssier and Whitney, 2002; Teyssier et al., 2005). However, results of this contribution strongly argue against such a model for the formation of Frenchman Cap dome (FCD). The main argument against diapirism is the absence of a Cordilleran penetrative foliation in the core of FCD that precludes extensive ductile flow of rocks occupying the core of a large dome. It could be argued that a diapir exists below the surface of exposure, but if this was the case the deepest structural level of FCD would be

located above the apex of the diapir where experimental (Dixon, 1975) and natural (e.g. Gervais et al., 2004) studies of diapirs documented that the strain type should be oblate and the maximum shortening direction should be vertical. Such a strain regime is clearly incompatible with the E-W contraction indicated by the NNW-SSE upright folds and axial-planar subvertical leucosomes observed in FCD. Furthermore, a diapir rise would lead to a radial, normal-sense shear strain on its flanks or within the bended rocks above. This expected kinematic framework is not compatible with the thrust shear zone localized in the schistose sliver at Bourne Glacier. Extension assisted diapiric models are thus not viable dome-forming mechanisms for FCD.

Cordilleran strain that I identified at the lower basement level of FCD is limited to NNW-SSE trending upright folds and a thrust shear zone a few meters thick that most likely developed between ~52-49 Ma and were synchronous with northeast-verging shearing at the upper basement level and lower level of the cover sequence. Journeay (1986) argued that doming was synchronous with the formation of cascading, northeast-vergent folds (F_3) because these folds are limited to the easterly-dipping flanks of the dome that would have been in the shortening field of the strain ellipsoid during top-to-the-northeast simple shear. Testing this hypothesis requires a compilation of all structural data acquired in FCD and is beyond the scope of this thesis. However, if Journeay's hypothesis proves to be correct, possible synconvergent doming mechanisms include cross-folding (proposed by Duncan, 1984 for Thor-Odin dome), heterogeneous buckling during shearing (Journeay, 1986), basement duplexing (Journeay, 1986), and underthrusting of a strong basement indenter beneath an extruding channel (Beaumont et al., 2004). Although less likely, it is also possible that the dome entirely resulted from, or

was amplified by extensional doming mechanisms such as warping in the footwall of the brittle-ductile Columbia River Fault, or crustal-scale heterogeneous boudinage (Price and Monger, 2001). Therefore, this study rules out the extension-assisted diapiric model for the formation of FCD, but more work is still required to identify a specific model.

Finally, it is not conceivable that Thor-Odin dome formed by extension-assisted diapirism while FCD formed by one of the other models mentioned above because the formation of domes by diapirism with diameters >30 km would mobilize a large part of the total crustal thickness, whereas it was clearly not the case in FCD. Consequently, unless a crustal-scale shear zone is identified between the two domes, another doming mechanism has to be considered for Thor-Odin as well. In this respect, the reader is referred to the study of Read (1980) that presented a N-S cross-section across the two domes and showed the continuity of the basal quartzite.

2.10 Conclusion

1. Field relationships (Fig. 2.4a, b), macro- and microscopic textures (Fig. 2.4c), trace element profiles (Fig. 2.5) and U-Pb age of seven dykes (Fig. 2.6 & 2.8) demonstrate that gray granite dykes of the Bourne suite were emplaced in the basement of Frenchman Cap dome in the Paleoproterozoic. The dykes thus constitute excellent time markers to distinguish Cordilleran from Precambrian strain.
2. A transition zone of decreasing, northeast-directed shear strain, marked by abundant folds (Fig. 2.9b), occurs ~ 1.5 km below the cover-basement contact in Frenchman Cap dome (Fig. 2.9). Above this zone, and throughout the Monashee Complex and overlying LSa, strain is high and indicates easterly-directed shearing (Fig. 2.2); whereas below this transition zone, and within a 4-5 km thick structural section, a

penetrative Cordilleran foliation is absent (Fig. 2.9. f, g).

3. Cordilleran strain at the lower basement level of Frenchman Cap dome identified in this study is limited to a meter-scale, top-to-the-east shear zone located within schistose slivers interlayered with basement orthogneiss at the Bourne Glacier locality (Fig. 2.12) and to upright NNW-SSE trending folds associated with hornblende-bearing leucosome at the Frisby Creek and Perry River localities (Fig. 2.11b, c). Metamorphic and geochronologic results indicate that these structures developed at less than ~20 km depth between ~52-49 Ma, synchronous with top-to-the-east shearing at the upper basement level and at the base of the cover sequence.
4. All the above characteristics are interpreted to represent a frozen downward-migrating base of convergent shearing. The orogenic base of the southern Canadian Cordillera is thus exposed in the core of Frenchman Cap dome.
5. Preservation of a Paleoproterozoic foliation within a 4-5 km thick structural section is not compatible with doming by extension-assisted diapirism. Unless a major shear zone between the northern (Frenchman Cap) and southern (Thor-Odin) domes of the Monashee Complex is identified, diapric models are also not applicable to Thor-Odin.

Table 2.1. Major and trace element compositions.

Specimen	FG-204	JC-33	JC-50	FG-425b	FG-187
Rock type	schist	Leuco-granite	Bourne granite	Bourne granite	Bourne granite
Locality	Bourne Glacier	Perry River	Perry River	Frisby Creek	Bourne Glacier
UTM NAD83 Zone 11	390166E 5687987N	390288E 5686655N	389458E 5675424N	394005E 5677681N	390387E 586506N
Major (wt%)					
SiO ₂	62.06	77.26	72.20	72.94	70.44
Al ₂ O ₃	14.70	12.65	13.44	13.27	13.77
CaO	0.86	2.44	1.39	1.30	1.73
Fe ₂ O ₃ (T)	12.78	2.05	3.26	2.51	3.68
K ₂ O	3.76	0.67	5.80	6.16	5.67
MgO	2.84	0.13	0.32	0.33	0.43
MnO	0.07	0.03	0.05	0.03	0.04
Na ₂ O	0.88	4.33	2.66	2.54	2.60
P ₂ O ₅	0.07	0.02	0.06	0.08	0.07
TiO ₂	0.68	0.08	0.26	0.32	0.33
L.O.I.	1.50	0.70	0.80	0.80	0.30
Sum	100.20	99.80	99.65	99.77	99.04
Trace (ppm)					
Ba	92	77	830	1582	948
Ce	22	84	234	278	333
Co	118	5	4	3	6
Cr	20	14	24	12	12
Ga	68	21	18	16	20
La	15	49	107	115	188
Nb	52	13	21	12	25
Nd	91	39	81	92	120
Ni	26	27	58	19	17
Pb	124	20	58	37	47
Rb	73	33	155	216	225
Sr	26	354	212	165	153
Th	9	53	61	37	65
U	95	16	6	3	7
V	19	7	7	16	19
Y	106	43	32	16	25
Zn	239	32	50	44	57
Zr	62	592	242	304	355

Table 2.2. Results of U-Pb laser ablation MC-ICP-MS spot analyses of zircon grains from Bourne granite sample JC-50-03 (Perry River).

	$^{207}\text{Pb}/^{206}\text{Pb}$	$\pm 2\sigma$	$^{207}\text{Pb}/^{235}\text{U}$	$\pm 2\sigma$	$^{206}\text{Pb}/^{238}\text{U}$	$\pm 2\sigma$	rho	$^{207}\text{Pb}/^{206}\text{Pb}$ age	$\pm 2\sigma$
Igneous								Ma	Ma
L01-1	0.11283	0.00120	4.2059	0.1264	0.2704	0.0108	0.99	1845	19
L03-1	0.11335	0.00119	4.7414	0.1423	0.3034	0.0101	0.95	1854	19
L04-1	0.11075	0.00114	4.3862	0.1317	0.2872	0.0099	0.96	1812	19
L05-1	0.11089	0.00111	4.2827	0.1287	0.2801	0.0115	0.73	1814	18
L06-1	0.11287	0.00116	4.3997	0.1321	0.2827	0.0105	0.98	1846	19
L09-1	0.11047	0.00113	4.0772	0.1226	0.2677	0.0110	0.73	1807	19
L10-1	0.10900	0.00110	4.0762	0.1223	0.2712	0.0087	0.95	1783	18
L12-1	0.10974	0.00111	4.0086	0.1203	0.2649	0.0088	0.95	1795	18
S 10	0.11013	0.00122	4.4591	0.1338	0.2937	0.0094	0.94	1802	20
S-27	0.11195	0.00118	3.9921	0.1203	0.2586	0.0139	0.56	1831	19
S-30*	0.11316	0.00125	4.6192	0.1391	0.2961	0.0154	0.58	1851	20
S-34	0.11136	0.00114	4.1080	0.1233	0.2675	0.0082	0.94	1822	19
Inherited									
L15-1	0.11915	0.00155	4.9633	0.1489	0.3021	0.0094	0.91	1943	23
S 01	0.12997	0.00177	6.4141	0.1925	0.3579	0.0111	0.90	2098	24
S 04	0.12194	0.00143	5.7060	0.1712	0.3394	0.0108	0.93	1985	21
S 15	0.12104	0.00144	5.9049	0.1772	0.3538	0.0109	0.92	1972	21
S-11	0.12496	0.00134	5.4948	0.1649	0.3189	0.0102	0.94	2028	19
S-18	0.12598	0.00138	5.8682	0.1761	0.3378	0.0108	0.94	2043	19
S-22*	0.12333	0.00140	5.1223	0.1543	0.3012	0.0164	0.55	2005	20
S-26	0.12423	0.00137	5.3096	0.1594	0.3100	0.0105	0.95	2018	20

40 x 40 μm laser beam size except for * that were 20 x 20 μm . All uncertainties are at the 2 sigma level.

Table 2.3 Results of U-Pb laser ablation MC-ICP-MS spot analyses on titanite grains from sample JC-50-03 (Perry River).

	Uncorrected isotopic ratios and uncertainties				rho	Corrected Age ^a	
	²⁰⁷ Pb/ ²⁰⁶ Pb	± 2σ	²³⁸ U/ ²⁰⁶ Pb	± 2σ		²³⁸ U/ ²⁰⁶ Pb	± 2σ
						(Ma)	
M04	0.1235	0.0013	3.2269	0.1142	0.98	1707	59
M04_2	0.1393	0.0016	3.8906	0.1293	0.98	1399	45
M04_3	0.1232	0.0012	3.2133	0.1016	0.98	1715	53
M12_1	0.7391	0.0126	11.9103	0.5287	0.82	58	32
M13_1	0.3677	0.0056	74.7463	2.7583	0.82	50	3
M13_2	0.3807	0.0061	74.9314	2.4994	0.82	49	3
M14_1	0.1708	0.0023	109.3354	3.7372	0.82	49	2
M14_2	0.1917	0.0056	105.2591	4.3419	0.82	50	2
M18_1	0.5033	0.0087	58.7494	2.0252	0.82	45	5

^a ²³⁸U/²⁰⁶Pb age corrected by the ²⁰⁷Pb method (Stern, 1997) by taking the y-intercept on the Terra-Wasserberg plot (Fig. 2.7) as the value of common Pb.

Table 2.4. Composition of metamorphic minerals for sample FG204.

	Biotite	Garnet	Plagioclase	Plagioclase
Analysis#	9	1	14-core	15-rim
SiO ₂	34.56	36.45	61.55	62.05
TiO ₂	1.13			
Al ₂ O ₃	21.00	21.07	24.34	24.04
Fe ₂ O ₃	21.50	35.81		
MnO	0.10	3.44		
MgO	7.18	2.24		
CaO	0.01	1.38	5.71	5.24
Na ₂ O	0.28		8.36	8.87
K ₂ O	9.29		0.17	0.13
F	0.62			
Cl	0.09			
H ₂ O	3.58			
Total	99.34	100.39	100.13	100.33
Oxygens	12	12	8	8
Si	2.66	2.95	2.73	2.74
Al tot	1.90	2.02	1.27	1.25
Al iv	1.34	0.04		
Al vi	0.56	1.97		
Ti	0.06			
Fe	1.38	2.43		
Mn	0.01	0.24		
Mg	0.82	0.27		
Ca	0.00	0.12	0.27	0.25
Na	0.04		0.72	0.76
K	0.91		0.01	0.01
X _{grs}		0.04		
X _{pyr}		0.09		
X _{sps}		0.08		
X _{an}			0.27	0.24

Table 2.5. Results of U-Pb laser ablation MC-ICP-MS spot analyses of zircon grains from leucosome sample JC-42-03 (Perry River).

Anal.#	$^{207}\text{Pb}/^{206}\text{Pb}$	$\pm 2\sigma$	$^{207}\text{Pb}/^{235}\text{U}$	$\pm 2\sigma$	$^{206}\text{Pb}/^{238}\text{U}$	$\pm 2\sigma$	rho	$^{207}\text{Pb}/^{206}\text{Pb}$	$\pm 2\sigma$
								Age (Ma)	Ma
L02-1	0.12619	0.00130	5.5690	0.1672	0.3201	0.0113	0.96	2046	18
L04-1	0.12579	0.00128	5.5340	0.1662	0.3191	0.0119	0.98	2040	18
L05-1	0.12508	0.00129	5.4020	0.1621	0.3132	0.0105	0.95	2030	18
L09-1	0.12487	0.00126	5.4612	0.1639	0.3172	0.0109	0.96	2027	18
L11-1	0.12586	0.00133	5.5366	0.1662	0.3190	0.0119	0.97	2041	19
M-1	0.12580	0.00132	5.3383	0.1602	0.3078	0.0106	0.96	2040	19
M-11	0.12663	0.00132	5.5267	0.1659	0.3165	0.0115	0.97	2052	18
M-12	0.12563	0.00130	5.3646	0.1610	0.3097	0.0109	0.96	2038	18
M-13	0.12625	0.00128	5.4987	0.1650	0.3159	0.0105	0.95	2046	18
M-14	0.12344	0.00128	5.1350	0.1542	0.3017	0.0109	0.97	2007	18
M-16	0.12602	0.00128	5.1422	0.1543	0.2959	0.0095	0.95	2043	18
M-17	0.12737	0.00131	5.7227	0.1717	0.3259	0.0109	0.95	2062	18
M-18	0.12674	0.00128	5.6496	0.1696	0.3233	0.0108	0.95	2053	18
M-2	0.12560	0.00129	5.4596	0.1640	0.3153	0.0123	0.99	2037	18
M-20	0.12665	0.00129	5.6484	0.1696	0.3235	0.0123	0.98	2052	18
M-21	0.12420	0.00126	5.2965	0.1590	0.3093	0.0113	0.97	2017	18
M-22	0.12665	0.00130	5.5262	0.1659	0.3164	0.0110	0.96	2052	18
M-2-3	0.12676	0.00129	5.6911	0.1708	0.3256	0.0115	0.96	2054	18
M-3	0.12795	0.00130	5.8741	0.1763	0.3330	0.0117	0.96	2070	18
M-5	0.12558	0.00129	5.3982	0.1621	0.3118	0.0112	0.97	2037	18
M-8	0.12685	0.00129	5.5931	0.1679	0.3198	0.0116	0.97	2055	18
M-9	0.12646	0.00128	5.4717	0.1643	0.3138	0.0117	0.98	2049	18

Table 2.6. Results of U-Pb laser ablation MC-ICP-MS spot analyses of zircon grains from leucosome sample FG425c (Frisby Creek).

Anal.#	$^{207}\text{Pb}/^{206}\text{Pb}$	$\pm 2\sigma$	$^{207}\text{Pb}/^{235}\text{U}$	$\pm 2\sigma$	$^{206}\text{Pb}/^{238}\text{U}$	$\pm 2\sigma$	rho	age ^a	$\pm 2\sigma$
								Ma	Ma
M01-1	0.12333	0.0012	5.3082	0.1594	0.3122	0.0115	0.98	2005	18
M02-1	0.12616	0.0013	5.3490	0.1605	0.3075	0.0104	0.95	2045	19
M03-1	0.12517	0.0013	5.3551	0.1608	0.3103	0.0114	0.97	2031	18
M04-1	0.12671	0.0021	5.8128	0.1745	0.3327	0.0111	0.86	2053	30
M04-2	0.12246	0.0013	4.6956	0.1410	0.2781	0.0097	0.96	1992	19
M05-1	0.12733	0.0013	5.6746	0.1704	0.3232	0.0115	0.96	2061	18
M05-2	0.12578	0.0013	5.5441	0.1664	0.3197	0.0111	0.96	2040	18
M06-1	0.12550	0.0013	5.0017	0.1502	0.2891	0.0105	0.97	2036	18
M07-1	0.12736	0.0013	5.8605	0.1759	0.3337	0.0111	0.95	2062	18
M07-2	0.12393	0.0014	4.8923	0.1468	0.2863	0.0089	0.93	2014	21
M08-1	0.12530	0.0014	5.4179	0.1626	0.3136	0.0099	0.93	2033	20
M10-1	0.12666	0.0013	5.4389	0.1632	0.3114	0.0106	0.96	2052	18
M12-1	0.12628	0.0013	5.6572	0.1698	0.3249	0.0110	0.96	2047	18
S10-1	0.04714	0.00449	0.0539	0.0049	0.0083	0.0003	0.01	53	2
S10-2*	0.03522	0.00129	0.0399	0.0020	0.0082	0.0003	0.70	53	2
S10-3*	0.03825	0.00094	0.0426	0.0019	0.0081	0.0004	0.84	52	2

40 x 40 μm laser beam size except for * that were 20 x 20 μm . ^a Ages are $^{207}\text{Pb}/^{206}\text{Pb}$ for Precambrian date and $^{206}\text{Pb}/^{238}\text{U}$ for Tertiary dates.

Figure 2.1. (a) Inset. Morphogeological map of the Canadian Cordillera (modified after Wheeler and McFeely, 1991). (b) Simplified geological map of Frenchman Cap dome (FCD) of the northern Monashee Complex (modified from Björnson, 2003). The three localities discussed in the text are shown as squares.

Figure 2.2. SW-NE geological cross-section of Frenchman Cap dome. Notice that northeast-verging folding of lithologic contacts is limited to the easterly-dipping flank of the dome, whereas two fold nappes crop out only on the western flank. Lower hemisphere equal area projections from the Lower Selkirk allochthon and upper level of the Monashee Complex cover sequence highlight the parallelism between stretching lineations (L) and fold axes (F) that is indicative of a high level of strain (data from Journeay, 1986). See Fig. 2.1 for location of cross-section A-B.

Figure 2.3. Geological map, field relationships and structural data from the upper and middle basement levels of FCD. (a) Bourne granite dyke folded by NE-verging F_2 fold within the upper basement level of FCD. Lens cap for scale. (b) NE-verging, “cascading” F_2 folds (F_3 at basement level) developed in the basal quartzite of the cover sequence. (c & f) Cordilleran deformation at upper basement level (area J on the map) is characterized by the parallelism between L_2 stretching lineations and F_2 fold axes (data from Journeay, 1986). (d & g) The spread in orientation of L_2 lineations and F_2 fold axes from Bourne Glacier locality (area K on the map) contrasts sharply with the clustering observed in area J. (e) Geological map of upper and middle basement levels of FCD. Notice the truncation of lithologic units at the transition zone (stripped pattern).

Figure 2.3

Figure 2.4. Bourne granite suite. (a, b) Field relationships of Bourne granite specimen JC-50-03 from Perry River locality. Notice the high angle between the gneissosity of the host granodioritic augen gneiss and the dyke margin. Also, notice the very weak mineral alignment in Bourne granite. Notebook for scale. In (b), Bourne granite is mingled with a pegmatitic leucogranite phase. These two granite phases commonly occur together. Notebook for scale. (c) Inverted color scanning electron microscope (SEM) image of the ~1846 Ma Bourne granite sample JC-13-03 dated by Crowley et al. (2008). This specimen is one of the rare Bourne granite dykes that run parallel to the foliation of the host augen gneiss. This thin-section is cut parallel to the lineation and perpendicular to the foliation of the host gneiss. Notice that quartz (white) is clearly not aligned and has a euhedral to subhedral shape demonstrating the absence of solid-state fabric in the dyke. Also shown is the typical allanite corona surrounding a resorbed magmatic monazite. (d) Backscattered image of an allanite crystal in sample FG-425b from Frisby Creek locality. Allanite in this sample is interpreted as a magmatic phase based on its prismatic shape and concentric zoning in REE (lighter gray shades on the image correspond to higher REE content).

Figure 2.4

Figure 2.5. Trace element profiles for Bourne granite.

The profiles highlight the similarities between three Bourne granite specimens (labeled B.granite) from each of the three studied localities (B= Bourne Glacier; P= Perry River; F= Frisby Creek); and leucogranite specimen JC-33-03 (labeled “lc granite”) from Perry River locality.

Figure 2.6. Results of laser ablation MC-ICP-MS U-Pb spot analyses from zircon grains in Bourne granite sample JC-50-03.

Specimen collected at Perry River locality (see Fig. 2.1 for location). (a) U-Pb analyses of 20-40 μm spots on zircon grains yield a regression line (MSWD = 2.3) with an upper intercept of 1861 ± 58 Ma and a lower intercept of 344 ± 400 Ma. The upper intercept is interpreted as the crystallization age. (b) Cathodoluminescence (CL) image of a typical zircon from specimen JC-50-03 showing oscillatory zoning and dark rim. (c) CL image of a zircon from specimen JC-50-03 indicating that the dark rims common to many zircon grains of the suite are also Paleoproterozoic.

Figure 2.7. Results of laser ablation MC-ICP-MS U-Pb spot analyses from titanite grains in Bourne granite sample JC-50-03.

Specimen collected at Perry River locality (see Fig. 2.1 for location). (a & b). BSE images of titanite grains with subhedral shape (a) and zoning that could be interpreted as relict oscillatory zoning (indicated by arrows). (c) Terra-Wasserburg plot for the youngest U-Pb analyses. The upper intercept on the y-axis is interpreted as the initial common Pb composition, and the lower intercept on the concordia curve yields an age of 49.2 ± 2.6 Ma. (d) Weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 49.3 ± 1.1 Ma corrected for common Pb by using the y-intercept in (c) as the initial common Pb (cf. Aleinikoff et al., 2004; Storey et al., 2006). The ~ 49 Ma age is interpreted as the crystallization age of the youngest titanite grains or as the time at which the grains closed for Pb diffusion.

Figure 2.8. Weighted mean average $^{207}\text{Pb}/^{206}\text{Pb}$ age for pegmatite and Bourne granite analyzed for U-Pb geochronology in Frenchman Cap dome.

Figure 2.9. WNW-ESE geological cross-section and field relationships from the area surrounding the northern toe of Bourne Glacier (location on Fig. 2.3). (a-c) Strain gradient developed in the ~2.1 Ga granodioritic augen gneiss. At Bourne Glacier, primary cores of K-feldspar are still present in augens (c). The augen gneiss is highly folded within the transition zone (b) and completely transposed at upper structural level with a strong LS fabric (a). (d) WNW-ESE cross-section. Same fill patterns as on the map of Fig. 2.3e except for thick black lines illustrating the deformation state of Bourne granite sheets. (e) Isoclinal and recumbent fold developed in a Bourne granite dyke at upper basement level. (f) Thin Bourne granite dyke crosscutting a fold train at lower basement level. (g) Weakly deformed Bourne granite dykes crosscutting the foliation of the granodioritic augen gneiss at a high angle within lower basement level (Bourne Glacier). Vertical scale = horizontal scale. Scales: a) pen; b) hammer handle; c) compass pointing north; f) hammer; g) lens cap.

Figure 2.9

Figure 2.10.

Figure 2.10. Field photograph of pegmatitic layer (leucosome?) specimen JC-09-03 from Bourne Glacier locality.

This pegmatitic layer crystallized at ~1925 Ma (Crowley et al., 2008). It is folded into an isoclinal F_2 fold axial-planar to the gneissosity of the host gneiss and refolded by upright N-S trending F_3 folds. Pegmatite dykes specimens JC-08-05 and JC-308-05, dated respectively at ~1846 and ~1856 Ma by Crowley et al. (2008), were collected less than 100 m away within the same continuous outcrop. The age of the migmatitic gneissosity is thus bracketed between ~1925 and ~1856 Ma. Lines are traces of axial planes of the two generation of folds. Compass pointing north for scale (circle).

Figure 2.11. Field relationships between Bourne granite dykes and host gneisses at Frisby Creek and Perry River localities (see Fig. 2.1 for location). (a) A straight and undeformed Bourne granite dyke crosscuts both limbs of an axial-planar folded mafic layer (Frisby Creek locality). Hammer for scale. (b) ~10 meters east of (a) the migmatitic gneissosity is refolded by open NNW-SSE trending F_3 folds associated with steeply dipping, axial-planar, hornblende+garnet-bearing leucosome (black arrows). A Bourne granite dyke (contoured on the far left of the photo) is also transposed parallel to steep F_3 axial plane. F_3 is thus localized and younger than ~1850 Ma. Compass pointing to the N for scale. (c) A Bourne granite dyke crosscuts the gneissic layering at right angle, but is also folded by open, NNW-SSE trending F_3 fold (Frisby Creek locality). Compass pointing to the N for scale. (d) A Bourne granite dyke crosscuts an E-W stretching lineation (black arrow) developed in the host granodioritic augen gneiss (Perry River locality). Despite its typical "Cordilleran" orientation, the lineation must be older than the ~1850 crystallization age of the Bourne granite dyke. Notebook for scale.

Figure 2.11

Figure 2.12. Meter-scale ductile shear zone with a top-to-the-east sense of shear at Bourne Glacier locality.

(a) C-S fabric developed within a garnet-sillimanite-biotite schist containing leucosome probably derived from the same material as the ~1856 Ma pegmatite JC-308-03 (Crowley et al., 2008) that is dismembered into an oblique boudin train elsewhere in the same outcrop. Pencil for scale.

Figure 2.14. Equal area lower hemisphere projection of structural data from Frisby Creek locality.

Figure 2.15. Results of laser ablation MC-ICP-MS U-Pb spot analyses from zircon grains in leucosome samples Fg425c and JC42-05.

Specimen Fg425c was collected at Frisby Creek and specimen JC42-05 was collected at Perry River. (a) CL image of a typical oscillatory-zoned zircon inherited from the host granodioritic augen gneiss. (b) CL image of the only Eocene zircon extracted from the hornblende-bearing leucosome specimen Fg425c. (c) concordia diagram for 35 analyses from 30 inherited zircon grains extracted from the two leucosomes. The upper intercept age of 2114 ± 27 Ma is interpreted as the crystallization age of the host granodioritic augen gneiss. (d) The weighted mean average $^{206}\text{Pb}/^{238}\text{U}$ age of 52.7 ± 1.1 Ma from the lone Eocene zircon extracted from the two leucosomes is interpreted as the crystallization age of hornblende-bearing leucosome that is axial-planar to upright F_3 fold.

CHAPTER 3

TESTING EXHUMATION MODELS IN THE SOUTHEASTERN CANADIAN CORDILLERA

ABSTRACT

The aim of this study is to test five tectonic models for the exhumation of the Lower Selkirk allochthon (LSa), which is part of the Shuswap Complex of southeastern Canadian Cordillera. Three of these models involve exhumation during convergence: ductile thrusting above a ramp, progressive underthrusting under critical taper and ductile extrusion. The two other models involve exhumation during large-magnitude extension: exhumation along bivergent crustal-scale brittle-ductile normal shear zones and ductile flow toward a rolling hinge. A compilation of the extensive structural, metamorphic and geochronologic database available for the LSa is presented. The LSa is a 10-15 km thick panel of sillimanite-grade migmatitic supracrustals that records protracted, east-directed high-temperature ductile flow in the Cretaceous/Paleocene. A top-to-the-west ductile-brittle shear zone (the Okanagan Valley fault system) separates the LSa from the overlying staurolite- and lower-grade rocks of the Upper Selkirk allochthon deformed and metamorphosed in the Jurassic/Early Cretaceous. A top-to-the-east shear zone (the Monashee décollement) separates the LSa from the underlying kyanite-grade migmatites of the Monashee Complex deformed and metamorphosed between ~60-49 Ma. This contribution presents a new Pressure-Temperature-time-deformation path from the basal shear zone that indicates a steep decompression path at temperature >700°C from a depth of ~29 km to less than ~14 km between ~83 and ~57 Ma during top-to-the-east shearing. Although not as well constrained, a coeval

motion from the top-to-the-west ductile-brittle shear zone at the roof of the LSa is inferred from available data. Seven depth-time paths were derived from the LSa and underlying Monashee Complex and demonstrate that the LSa was exhumed to <14 km depth sequentially from NE to SW between ~90 Ma to 50 Ma, contemporaneous with burial of the Monashee Complex to a depth in excess of ~35 km. Those findings are incompatible with the popular models of exhumation by gravitational collapse and large-magnitude extension of a crustal welt in the Eocene. The results are rather compatible with ductile extrusion of the LSa between an underthrusting basement and an already exhumed USa.

3.1 Introduction

The southeastern Canadian Cordillera, with relatively easy access and beautiful alpine exposure, constitutes an excellent natural laboratory to study tectonic processes operative in the core zones of orogenic belts. Despite being one of the most studied mountain belts worldwide, its tectonic evolution is more debated than ever. A key issue concerns the exhumation process (es) that led to the exposure of a ~50 x 150 km area of migmatitic rocks. These migmatitic rocks are generally juxtaposed against rocks metamorphosed at lower-amphibolite/greenschist facies ~40-90 my before the main migmatitic event. The migmatitic exposures have often been ascribed the name *Shuswap Complex* (*cf. Okulitch, 1984*). Although large-magnitude extensional models are widely accepted (e.g. Coney and Harms, 1984; Parrish et al., 1988; Bardoux and Mareschal, 1994; Teyssier and Vanderhaeghe, 1997; Vanderhaeghe et al., 1999; Teyssier et al., 2005), other authors have argued that part of the exhumation occurred during convergence (Journeay, 1986; Brown and Journeay, 1987; Scammell, 1993; Johnson and

Brown, 1996; Johnston et al., 2000; Glombick et al., 2006a).

Some extensional models suggested that high-grade rocks were exhumed along crustal-scale ductile-brittle normal faults dipping outward from the metamorphic core (Fig. 3.1a). According to these models, a change in far-field stress after the orogen had been thickened to >50 km induced gravitational collapse and extension of the crust that was accommodated along crustal-scale normal faults (Parrish et al., 1988), or listric normal faults in the upper crust and ductile thinning in the lower crust (Bardoux and Mareschal, 1994). Restoration of balanced cross-sections indicated net extension of ~80-100 km (Parrish et al., 1988; Bardoux and Mareschal, 1994).

Other models put the emphasis on ductile flow during gravitational collapse of a thickened orogen (e.g. Block and Royden, 1990; Wernicke, 1992) and domes coring the Monashee Complex of southeastern Canadian Cordillera were interpreted as the product of extension-assisted diapirism (Vanderhaeghe et al., 1999; Norlander et al., 2002; Teyssier and Whitney, 2002). These diapiric models evolved into one of channel flow toward a rolling-hinge presented by Teyssier et al. (2005) who described a model that is very similar to the analogue models of Brun and Van den Driessche (1994) and to the thermomechanical numerical models of core complex formation by Tírel et al. (2006 & 2008). According to the model of Teyssier et al. (2005), illustrated on Fig. 3.1b, extension initially localized at the thermal boundary between the cold craton and the hot orogen (Hyndman and Lewis, 1999) and led to the formation of a master detachment. As the crust extended from the west, domino-style normal faults thinned the upper crust above the master detachment. Thinning of the upper crust created a pressure gradient that drove the flow of a hot and ductile middle crust. Coupling of mid-crustal flow and

dissection of the upper crust by normal faulting resulted in the flattening and rotation of the master detachment, effectively creating a rolling-hinge. The combined effect of the upper crust moving westward and underlying mid-crustal rocks, located west of the hinge, flowing eastward toward the thinned upper crust above the hinge created a top-to-the-west shear couple. A top-to-the-east shear couple developed by the same process to the east of the hinge. Teyssier et al. (2005) argued that a diapir developed and flowed vertically en-masse toward the kinematic hinge at the climax of extension in the Eocene. Furthermore, isotherms were bowed upward during the formation of the rolling hinge such that a transient brittle-ductile transition developed, above which mid-crustal rocks were progressively cooled. As the upper crust was pulled westward, progressively younger mid-crustal rocks were exhumed at the hinge.

In stark contrast with gravitational collapse models, other authors considered that exhumation was a two-stage process, initially involving a period of synconvergent exhumation followed by extension. One of these models involves underthrusting of a stiff indenter and/or basement duplexing beneath the thrust sheet (Fig. 3.1c) that would induce erosion and normal faulting at the surface to stabilize the gravitational instability created by thickening (Journeay, 1986; Brown and Journeay, 1987; Spear, 2004; Carr and Simony, 2006). Exhumation in this model comes from tectonic denudation and erosion. In this model, the entire crustal edifice moves as a block along a basal shear zone, although internal ductile deformation also takes place.

Brown (2004) built on Parrish (1995) by integrating results of more than 25 years of research in the southeastern Canadian Cordillera and proposed a model — inspired by Platt (1986) — involving progressive westward underplating beneath an orogenic wedge

in critical taper. In this model, the location of the basal shear zone and isotherms are fixed with respect to an external reference frame and rock particles cycle through the wedge (Fig. 3.1d). Hence, underplated rocks get deformed and metamorphosed as they enter the basal shear zone and are then exhumed by the combined effect of underplating from below and denudation — by extensional faulting and erosion— at the surface that maintains the wedge in a critical taper state.

Finally, Scammell (1993) was the first to suggest a model of synconvergent ductile extrusion in the southeastern Canadian Cordillera (Fig. 3.1e), an idea that was revived by Johnston et al (2000), and Glombick et al. (2006a). Several papers of the volume “Channel flow, ductile extrusion and exhumation in continental collision zones” (2006) discussed the problem of semantics surrounding channel flow and ductile extrusion processes and highlighted the fact that there is no consensus on which term should be used (Godin et al., 2006b; Grasemann et al., 2006; Jones et al., 2006; Williams et al., 2006). In this contribution, I will refer to *channel flow* as the process by which a kilometric layer of low-viscosity material leaves its lid and base behind as it flows *horizontally* down a pressure gradient in the lower to middle crust. In contrast, the term *ductile extrusion* will refer to the process by which a slice of deformable rock material is detached from its base and lid and flows towards the surface of the Earth, without necessarily reaching it. The upward translation of the slice is accommodated by ductile shear zones of opposite shear sense and by internal ductile deformation. This definition has no genetic implications with regards to the driving force; which could be: 1) a lateral pressure gradient induced by a topographic gradient or density contrast between two rock columns; 2) active erosion at the orogenic front; 3) convergence of two blocks that expels

the rock mass between them; or 4) any combination of the above.

This chapter first presents a compilation of the extensive database of structural, metamorphic and geochronologic data from the LSa, presents new pressure-temperature-deformation-time paths from the foreland-verging shear zone bounding its base (the Monashee décollement), and reconstructs five depth-time paths for different areas of the infrastructure. These paths are then compared to two depth-time paths reconstructed for the underlying Monashee Complex presented in the fourth chapter of this thesis. From a critical analysis of the available data, I conclude that the only model compatible with the current multidisciplinary dataset is a synconvergent ductile extrusion model followed by limited Eocene extension.

3.2 Tectonic setting

See section 1.1.

3.3 Tectonometamorphic setting of the Lower Selkirk allochthon (LSa)

There is currently no consensus about the best scheme for subdividing the southeastern Canadian Cordillera. For example, the westernmost high-grade rocks of the Omineca Belt located west of Revelstoke (Fig. 3.2) have been referred to as “Kootenay Terrane” (e.g. Johnston et al., 2000); “undivided high grade rocks” (Carr and Simony, 2006); “Middle Unit of the Shuswap Complex” (Teyssier et al., 2005); “Late Proterozoic-Mesozoic metamorphic assemblage of the Middle-crustal zone” (Kruse and Williams, 2007); “Neoproterozoic Windermere Supergroup of the Selkirk allochthon” (Gibson et al, 2005); “Monashee Cover sequence” (Thompson et al., 2006). The boundaries between units also differ, which adds to the confusion. In this section, I present a new subdivision scheme by describing the salient features of a regional map and of three cross-sections.

This section also serves as a review of data that will be used later to test tectonic models of exhumation.

It is possible to subdivide the southeastern Canadian Cordillera in three tectonometamorphic units based on structural, metamorphic and geochronologic characteristics. The map of Fig. 3.2 and three simplified cross-sections shown on Fig. 3.3 present the subdivision adopted in this paper along with thermochronologic data from which it is based. The northernmost section A-B that runs SW-NE is modified from Gibson (2003); the C-D cross-section that runs W-E is modified from Johnson and Brown (1996); whereas section E-F that runs W-E combines the interpretation of the Lithoprobe seismic reflection lines 7-9 by Glombick et al. (2006a) and the interpretation of the Lithoprobe seismic refraction line 7 by Burianyk and Kanesevich (1995). The main criteria that distinguish the three units are the timing of ductile deformation and thermal peak. On the map of Fig. 3.2, most cooling ages are for hornblende except when an adjacent letter indicates otherwise (*t* for U-Pb titanite, *b* for $^{40}\text{Ar}/^{39}\text{Ar}$ of biotite, and *ap* for apatite fission track of apatite). Numerical values of these ages are shown on the simplified cross-sections of Fig. 3.4, and references on Table 1.1 (p.40). For $^{40}\text{Ar}/^{39}\text{Ar}$ analyses, only plateau ages defined by spectra of more than 80% of ^{39}Ar are shown. Black symbols represent ages older than 120 Ma, gray symbols represent ages between 120 and 60 Ma, and white symbols are ages between 59 and 47 Ma. From all these data, four main *tectonometamorphic* units are singled-out. From upper to lower structural levels: the light gray unit is the Upper Selkirk allochthon (USa) that was deformed and metamorphosed before 120 Ma (black symbols); the medium gray unit is the Lower Selkirk allochthon (LSa) that was mainly deformed and metamorphosed between 120 and

60 Ma (gray symbols); and the dark gray units is the cover sequence of the Monashee Complex that was deformed and metamorphosed between ~59 and 47 Ma (white symbols). The lowermost white unit with crosses represents basement rocks with affinities with the North American craton (Armstrong et al., 1991; Crowley, 1999). These rocks were deformed and metamorphosed in the Paleoproterozoic (Crowley et al., 2001; Crowley et al., 2008; chapter 2 of this thesis) and were heterogeneously deformed and metamorphosed again between ~59 and 47 Ma. The three unit names were chosen rather than the upper-middle-lower crustal zones of Carr (1991) because they assure continuity with the traditional subdivision presented in Read and Brown (1981). In the following sections, the main features of LSa and its boundaries are described. The focus is first put on high-grade rocks cropping out west of the Monashee Complex. Then it is argued that the high-grade rocks cropping out in the northern Selkirk and Monashee mountains (Fig. 3.2) share similar characteristics as, and are physically linked to LSa rocks cropping out west of the Monashee Complex.

3.3.1 The Lower Selkirk allochthon west of the Monashee Complex

West of the Monashee Complex, the LSa consists mainly of semi-pelitic schist, calc-silicate, amphibolite and dioritic gneiss, with subordinate amount of pelitic schist, marble and quartzite (Carr, 1991; Johnson, 1994; Vanderhaeghe et al., 1999; Glombick et al., 2006a). Most rocks of the LSa along sections C-D and E-F (Fig. 3.3) are at the sillimanite-K-feldspar grade (Glombick, 2005, modifying Carr, 1991) except in the Pinnacles area (Pnc on Fig. 3.2) where the muscovite-out, and sillimanite-in isograds have been mapped toward the USA. Leucocratic material, in the form of leucosome and leucogranitic sheets, commonly comprises 35-60% of any outcrop by volume. Most of

the leucogranitic sheets have been interpreted to be part of the Ladybird suite emplaced between ~62 and 54 Ma (Carr, 1992). Ladybird intrusions postdate much of the regional deformation of L_{Sa} rocks and are variously strained depending on their temporal and spatial relationships (Carr, 1992).

Because of extensive retrograde exchange and net-transfer reactions, it is often difficult to determine peak P-T conditions in migmatites (e.g. Spear et al., 1999). Nonetheless, 20 out of 22 locations where geothermobarometry has been conducted in the southern part of the map of Fig. 3.2 fall in the range 650-850°C and 700-1000 MPa (Teyssier et al., 2005; Glombick, 2005). This range is compatible with P-T estimates based on petrogenetic grids of metapelites consisting of the peak mineral assemblage garnet-K-feldspar-sillimanite-biotite (e.g. Spear et al., 1999) that characterize most rocks of the L_{Sa}.

Rocks of the L_{Sa} are characterized by a well-developed L-S transposition fabric (Journeay, 1986; Carr, 1991; Scammell, 1993; Glombick et al., 2006a; Johnson, 2006). The planar fabric is formed by the original bedding, various generations of veins and flattened leucosome, and aligned planar minerals that have all been transposed into a transposition foliation. The linear fabric is formed by elongated mineral aggregates, elongated quartzo-feldspathic pressure shadows around garnet, aligned metamorphic minerals, and rod-shape leucosome that together define a stretching lineation (Journeay, 1986; Carr, 1991; Scammell, 1993; Johnston, 1998; Glombick et al., 2006a). On equal area lower hemisphere projections, lineations form well-defined clusters that plunge shallowly to the west (from WNW to SW; Fig. 3.2 e, g, h, i). Johnson (1994) reported that kinematic indicators and mylonites are rare in the structural centre of the L_{Sa},

whereas kinematic indicators abound within mylonites of the two oppositely verging shear zones bounding the LSa. Johnson (1994) interpreted these features as reflecting an increase of the kinematic vorticity number (Hanmer and Passchier, 1991) from the centre of the LSa to its margins. Similar observations were reported by Glombick et al. (2006a) for cross-section E-F. Furthermore, Teyssier et al. (2005) studied the anisotropy of magnetic susceptibility in leucogranite of the Ladybird suite along a transect passing between cross-sections C-D and E-F (Fig. 3.3) and concluded that there is a kinematic hinge approximately centred on Sugar Mountain, which is located east of Trinity Hills (TH on Fig. 3.2). According to Teyssier et al., top-to-the-west kinematic indicators prevail westward of the hinge, whereas top-to-the-east kinematic indicators prevail eastward of the hinge. Carr (1991) also reported a vergence reversal in unannealed shear structures within Ladybird leucogranite. However, the kinematic hinge described by Carr (1991) is located ~15 km further east than the hinge reported by Teyssier et al. (2005). Therefore, it appears that there is a vergence reversal across the LSa that is recorded in both the transposition fabric and in leucogranite of the Ladybird suite.

3.3.2 Western boundary between LSa and USa: the Okanagan Valley fault system

The westerly-dipping, top-to-the-west brittle-ductile Okanagan Valley fault system separates the USa from the underlying LSa (Fig. 3.3b). Notice that I follow Johnson (2006) and will refer to the Okanagan Valley fault system (OVfs) for the western boundary between the USa and LSa that includes the Adams-North Thompson fault, the Eagle River fault, the Kalamalka shear zone, and the Silver Star shear zone. The hanging wall of the OVfs consists of rocks at the greenschist facies or lower, whereas the immediate footwall of the fault consists of migmatitic rocks at the sillimanite-K-feldspar

grade that are mylonitic and contain numerous top-to-the-west shear sense indicators (Johnson, 1994 & 2006). Between Vernon and Nakusp (Fig. 3.2) along section E-F (Fig. 3.3c), two tectonic windows expose LSa rocks within USa exposures, whereas three outliers of USa are present within LSa exposures. Glombick et al. (2006a) showed that these observations are due to the subhorizontal attitude of the USa-LSa boundary at this latitude. Johnson (2006) and Glombick et al. (2006a) documented that top-to-the-west structures of the OVfs are formed by the peak metamorphic minerals garnet-sillimanite-K-feldspar, by the retrograde minerals muscovite and chlorite, as well as by brittle structures. These authors interpreted this sequence of mineral assemblages as indicating cooling and decompression during top-to-the-west shearing.

There are relatively good timing constraints for top-to-the-west shearing along the OVfs near Sicamous (Fig. 3.2 for location). Hangingwall rocks cooled through the apatite fission track temperature ($110 \pm 10^\circ\text{C}$, Green et al., 1989) by 67 Ma (Lorencak et al., 2001), whereas footwall rocks record complex and protracted zircon growth in the Cretaceous to Paleocene. An interesting study that has not been published is that of Gilley (1999) who conducted SHRIMP U-Pb dating of zircons from three leucosome specimens collected in the immediate footwall of the OVfs near the town of Sicamous (Fig. 3.2). Gilley (1999) documented an important episode of zircon growth occurring between ~160 and 155 Ma. Of the 25 youngest $^{206}\text{Pb}/^{238}\text{U}$ analyses, 11 are distributed between ~120 and 75 Ma, and 13 are distributed between ~75 and 60 Ma. The youngest $^{206}\text{Pb}/^{238}\text{U}$ date of 57.0 ± 1.3 Ma is a concordant analysis from an elongated and euhedral zircon rim having sharp crystal faces (Gilley, 1999 p. 51). This zircon is from a leucosome in a diorite gneiss that yielded concordant $^{206}\text{Pb}/^{238}\text{U}$ zircon dates of ~419 Ma,

~157 Ma, ~86 Ma, ~75 Ma, ~71 Ma and ~57 Ma. Although the emplacement age of this rock is not known, the spread of dates precludes its crystallization at ~57 Ma. Rather, the shape of the ~57 Ma grain indicates it grew in the presence of a melt phase. In addition, Johnson (1994) obtained a U-Pb age (IDTIMS) of ~56 Ma for a leucogranite emplaced during top-to-the-west shearing at Pukeashun Mountain (PM on Fig. 3.2). Johnson (1994) also obtained well-defined plateau $^{40}\text{Ar}/^{39}\text{Ar}$ ages for three hornblende samples located less than ~1 km structurally below the OVfs. If combined, these three dates yield a weighted mean average age of 56.1 ± 5.3 Ma (MSWD = 1.7; prob. = 0.18). Combined with structural and metamorphic observations described above, these geochronologic data are crucial for testing tectonic models because they demonstrate that the last increments of top-to-the-west shearing at high temperature along the OVfs took place during the Late Paleocene (61-55 Ma) along a cooling path. The progression of top-to-the-west shear sense indicators from high temperature ductile, to low temperature ductile, and then to brittle further suggest that decompression accompanied cooling. However, it is also possible that high temperature shearing occurred along an isobaric cooling path that was followed, or reactivated later, by low temperature and brittle shearing along a decompression path.

Glombick et al. (2006b) conducted detailed U-Pb geochronologic studies in the area surrounding the town of Vernon (SW corner of the map of Fig. 3.2). They demonstrated that the hanging wall of the OVfs was deformed between ~172 and ~162 Ma; whereas migmatitic rocks of the Aberdeen Complex in the footwall of the fault record a protracted history of zircon growth between ~155 and 150 Ma, at ~90 Ma, and between ~66 and ~51 Ma. Noteworthy is the 61.6 ± 4.2 Ma age obtained for two concordant titanite

fractions in a migmatitic diorite gneiss that suggests cooling below the titanite closure temperature at ~62 Ma (~600-700°C depending on grain size and cooling rate (Cherniak, 1993). The minimum age of deformation in the complex is constrained by the ~52 Ma emplacement age of a monzonite pluton that postdates the gneissosity in enclosing gneiss (Glombick et al., 2006b). The last increments of ductile top-to-the west shearing along the OVfs are constrained to have occurred between ~50 and ~47 Ma: the U-Pb age of a deformed felsic dyke and of an undeformed mafic dyke, respectively (Glombick et al., 2006b citing Heaman et al., 1999). These results are consistent with the $^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages obtained by Bardoux (1993) just outside of the southern limit of Fig. 3.2. Biotite in the hanging wall of the OVfs yielded a plateau $^{40}\text{Ar}/^{39}\text{Ar}$ age of ~160 Ma, whereas hornblende and biotite in the footwall rocks yielded $^{40}\text{Ar}/^{39}\text{Ar}$ ages between ~52 and ~49 Ma, respectively. Therefore, data from the Vernon area, especially the ~62 Ma titanite age, are compatible with that described above for Sicamous, which indicated high-temperature shearing along the OVfs in the Late Paleocene. However, it appears that rocks of the Vernon area remained above the hornblende closure temperature for a longer period of time until ~50 Ma when the last increments of shearing along the OVfs occurred.

3.3.3 Boundary between the LSa and the Monashee Complex: the Monashee décollement

The lower boundary of the LSa with the underlying Monashee Complex is a top-to-the east shear zone named Monashee décollement (cf. Brown et al., 1992). It is from studies conducted at the latitude of cross-section A-B (Fig. 3.3a) that the Monashee décollement was best documented (Read and Brown, 1981; Journeay, 1986; Scammell, 1986; Brown et al., 1992). There it marks: 1) a lithologic break with rocks characteristic

of the Monashee Complex stratigraphic assemblage (quartzite, marble, pelitic schist) in the footwall, and rocks characteristic of the semi-pelite amphibolite (SPA) division of the Neoproterozoic Horsethief Group in the hangingwall; 2) an abrupt increase in leucogranite intrusions by outcrop volume from <15 % in the footwall to more than ~20-30% in the hangingwall; 3) a metamorphic discontinuity with kyanite migmatite in the footwall and sillimanite migmatite in the hangingwall; 4) a geochronologic break. Indeed, rocks from the LSA in the hangingwall of the décollement record protracted zircon growth in sheared leucosome and leucogranite at ~118-115, ~97, ~93 and ~85 Ma (Scammell, 1993; Parrish, 1995) and top-to-the-east shearing ending prior to the ~57 Ma crystallization age of an undeformed leucogranite dyke cutting the foliation at right angles (Scammell, 1993). In contrast, extensive geochronologic studies conducted in rocks of the Monashee Complex indicated that high temperature metamorphism and deformation occurred from ~60-50 Ma (Crowley and Parrish, 1999; Gibson et al., 1999; Crowley et al., 2001; Foster et al., 2004).

On the map of Fig. 3.2, the Monashee décollement delimits the gray area (LSa) comprising mainly dark gray symbols from the dark gray area (Monashee Complex) comprising mainly white symbols. Timing for the last increments of shearing along the Monashee décollement at the latitude of northern Frenchman Cap dome is loosely bracketed between ~85 and ~57 Ma. Shearing took place at high temperature because shear fabrics are formed by the peak metamorphic mineral assemblage sillimanite-garnet-K-feldspar.

The presence of the Monashee décollement further south at the latitude of Thor-Odin dome has been questioned by Johnston (1998), Vanderhaeghe et al. (1999) and Williams

and Jiang (2005). However, these authors recognized that the trace of the Monashee décollement around Thor-Odin dome does coincide with a lithologic boundary that is less than ~1 km (horizontal distance)) from the kyanite-out isograd (Harrap, 1990; Carr, 1991; McNicoll and Brown, 1995; Johnston et al., 2000). In addition, this lithologic and metamorphic boundary occurs in an area where a geochronologic boundary must also be present to separate rocks deformed and metamorphosed between ~120 and 60 Ma (gray symbols on Fig. 3.2 and 3.3) from rocks deformed and metamorphosed between ~59 and 47 Ma (white symbols). Specifically, near Three Valley Gap (TVG on Fig. 3.2), zircon grew in leucosome between ~100 and 70 Ma with a cluster of dates between ~75 and 72 Ma (Gilley, 1999; Kuiper, 2003). A ~73 Ma leucogranite sheet cross-cutting the transposition foliation south of TVG indicates that the transposition foliation started to develop prior to ~73 Ma (Johnston et al., 2000); and a ~67 Ma mylonitic pegmatite containing top-to-the-east kinematic indicators north of TVG indicate that the last increments of shearing occurred after ~67 Ma (Johnson, 1994). In contrast, there is no evidence of deformation and high temperature metamorphism prior to ~65-60 Ma in the underlying Monashee Complex (Johnston et al., 2000; Kuiper, 2003).

In southern Thor-Odin dome, at Cariboo Alp (CA on Fig. 3.2), the last increments of top-to-the-east shearing along the Monashee décollement were well bracketed at ~59-58 Ma from the crystallization ages of deformed leucosome and undeformed leucogranite (Carr, 1992). It is likely that the last increments of shear strain occurred along a cooling path because hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages (>90% of ^{39}Ar) of ~57 and ~59 Ma were obtained in the immediate hangingwall of the décollement in southern Thor-Odin dome (Spark, 2001). In contrast, deformation and partial melting in the underlying Monashee

Complex took place from ~60 –to 52 Ma (Vanderhaeghe et al., 1999; Hinchey et al., 2006 & 2007).

Nyman et al. (1995) conducted a detailed study of strongly deformed sillimanite-bearing migmatites of the LSa within the Monashee décollement at Three Valley Gap (TVG on Fig. 3.2). They estimated that K-feldspar-garnet-sillimanite-biotite schists reached peak metamorphic conditions of 750-900 MPa/ 720-820°C that resulted in the production of ~54% melt, ~29% of which would have left the system. Importantly, Nyman et al. (1995) reported extensive garnet resorption and interpreted the resorption to be related to garnet breakdown in the presence of melt via the retrograde net-transfer reaction:

(ReNtr 1): garnet + melt + K-feldspar = sillimanite + biotite + plagioclase

that requires cooling in the presence of melt (Spear et al., 1999; Kohn and Spear, 2000). Similar observations were made by Scammell (1993) ~2 km structurally above the Monashee Complex near station A on Fig. 3.2.

Consequently, although a strain gradient was not observed (Johnston, 1998; Vanderhaeghe et al., 1999; Williams and Jiang, 2005), there is a lithologic, metamorphic and geochronologic boundary at the latitude of Thor-Odin dome. This boundary separates sillimanite-bearing migmatites of the LSa, which record protracted high temperature deformation and metamorphism between ~90-60 Ma, from kyanite-bearing migmatites of the underlying Monashee Complex, which record much shorter high temperature deformation between ~60-50 Ma. Because this boundary is defined by the same characteristics as the well-defined Monashee décollement at the latitude of Frenchman

Cap dome, one has to conclude that the décollement is the same structure along the ~150 km length of the Monashee Complex. The top-to-the-east shear fabric is formed by metamorphic textures suggesting that shearing started at high temperature, likely in the presence of melt, and continued along a cooling path. The best estimate of timing of the last increments of shearing comes from the southern end of Thor-Odin dome where it was bracketed between ~59-58 Ma (Carr, 1992). Although less precise, geochronologic data from the Three Valley Gap and from the northern Frenchman Cap dome are compatible with a Late Paleocene cessation of top-to-the-east shearing.

3.3.4 Columbia River Fault

The Columbia River Fault is a brittle-ductile, top-to-the-east normal fault that juxtaposes the staurolite-kyanite and lower grade rocks of the USA in its hanging wall to the sillimanite- and kyanite-bearing migmatites of the Monashee Complex and LSa in its footwall (Brown and Brown and Psutka, 1980; Read and Brown, 1981; Brown and Murphy, 1982; Lane et al., 1989). Because the Monashee Complex is at a lower structural level than the LSa, there is a reduction of relative dip-slip displacement from the core of the Monashee Complex toward the south and north. Johnson and Brown (1996) restored the unmodified version of cross-section C-D to its pre-extension state. They showed that 45 km of net extension accommodated by the OVfs and the Columbia River faults links together stratigraphic units in the hanging wall of both faults that are similar “...in terms of stratigraphy, metamorphic grade, structural style, and their unique association with widespread Devonian-Mississippian orthogneisses.” It is important for this thesis to note that this restoration puts rocks of the Monashee Complex and LSa at a depth of less than ~14 km. An exhumation mechanism other than extension would thus

be required to bring these rocks from mid-crustal levels at >25 km depth to less than ~14 km.

3.3.4 Cross-section A-B

The northern cross-section A-B is discussed last because its interpretation is less straightforward than the other two. The pitch of the average regional stretching lineation in the average foliation plane changes from almost down-dip to almost strike-parallel from south to north across the map (Fig. 3.2). Hence, by contrast with the other two sections, which are approximately parallel to the average regional stretching lineation and perpendicular to the average regional foliation, drawing this section perpendicular to the foliation strike results in an obliquity larger than 45° to the lineation. In addition, an assumption of the cross-section is that rocks of the LSa located north of the Monashee Complex and west of the Columbia River (between stations A and C on Fig. 3.2) project along the NW-SE strike *beneath* the USA that is exposed east of the river. This assumption finds its justification in the continuous exposure of the LSa from the Selkirk Mountains, to the northern Monashee Mountains, and then toward the SW around the Monashee Complex (Fig. 3.2). Hence, rocks of the LSa projected beneath the USA on cross-section A-B (Fig. 3.3a) were most likely adjacent to rocks exposed west of the Columbia River (between stations A-C; Fig. 3.2) before normal faulting along the Columbia River Fault.

Scammell (1993) integrated structural, metamorphic and geochronologic data from the LSa located between stations A and C on Fig. 3.2 into a model of Cretaceous synconvergent “dynamic spreading”, or ductile extrusion. It is important to elaborate on Scammell’s study because it carries important implications for the current work and

because some data can be interpreted differently. On equal-area lower hemisphere projections, the cluster of poles to minor fold axial planes lies within 5° of the cluster of poles to the transposition foliation. A west-plunging cluster of minor fold axes, for which a sheath geometry is common, also lie within 5° of the cluster of stretching lineations (Fig. 3.2a, b). A top-to-the-east rotational component of flow is indicated by the ubiquitous presence of kinematic indicators (see Fig. 3.4 for typical outcrop photographs). Scammell argued that a transposition of this magnitude by simple shear alone requires very high strain and that a pure shear component was probably required. From U-Pb dating of variously transposed leucogranite sheets, Scammell documented that high temperature top-to-the-east shear strain lasted several ten's of my in the Cretaceous. Furthermore, an important conclusion of Scammell was that large amounts of shear strain accumulated in the presence of melt via garnet breakdown by ReNtr 1. Microstructural evidence of this reaction is presented on Fig. 3.5a (taken from Scammell, 1993 p.93) in which garnet porphyroblasts are preferentially resorbed and replaced by products of ReNtr 1 in pressure shadows inclined to the lineation, a microstructure indicating a top-to-the-east shear sense. This microstructure implies that melt migrated toward the direction of minimal shortening (top left to bottom right quadrants) during top-to-the-east shearing while ReNtr 1 was proceeding. Furthermore, this microstructure implies cooling during shearing because ReNtr 1 is nearly isothermal and takes place continuously upon cooling in the melt field (cf. Spear et al., 1999). Scammell (1993) also documented that this cooling was accompanied by decompression in the order of 300-500 MPa. The timing of this synshearing exhumation episode was bracketed between ~ 100 Ma, the U-Pb ages of transposed leucogranite and U-Pb ages of monazite grains in schist;

and ~90 Ma, the oldest U-Pb titanite age. However, as discussed by Scammell (1993), there is a trend of cooling ages decreasing from northeast to southwest. This trend is illustrated on cross-section A-B of Fig. 3.3a where titanite U-Pb ages decrease from ~98-90 Ma at Scrip Creek to ~74 Ma at Ruddock Creek, while $^{40}\text{Ar}/^{39}\text{Ar}$ ages decrease from ~75-66 Ma to ~63-61 Ma across the same transect. It is thus possible that cooling and exhumation were diachronous across the area.

Along cross-section A-B, the Columbia River Fault separates footwall kyanite migmatites of the Monashee Complex, which were deformed and metamorphosed in the Paleocene-Eocene period, from hangingwall lower-amphibolite to greenschist facies rocks of the USA, which were deformed and metamorphosed in the Jurassic (Colpron et al., 1996; Gibson et al. 2005 & 2008). Sillimanite migmatitic schists that record penetrative east-verging deformation resurface in the footwall of the Selkirk detachment in the Selkirk Mountains (Brown and Gibson, 2006), and are in lithologic, structural and metamorphic continuity with rocks located west of the Columbia River, near A on Fig. 3.2. Numerous transposed leucosome pods and leucogranite sheets, as well as monazite in migmatitic sillimanite-bearing schists, yielded dates between ~105 and 90 Ma (Sevigny et al., 1989; Scammell, 1993; Crowley et al., 2003; Gibson et al., 2005). This suggests that peak metamorphic conditions were reached, and that large amounts of east verging shear strain accumulated during this time period.

A feature specific to the northeast exposure of the LSa is the prevalence of map-scale post-transposition and post-metamorphic folds, which commonly dip to the southwest (Brown et al., 1978; Ghent et al., 1980; Simony et al., 1980; Perkins, 1983; Gibson et al., 2005). Shortening associated with these post-metamorphic folds was estimated to be 30-

60% (Ghent et al., 1980; Simony et al., 1980). Restoring 50% of NE-SW shortening between A and Kinbasket Lake (KL on Fig. 3.2) would bring area A close to the NW-SE axis of the Monashee Complex. The minimal age of deformation is poorly constrained by the ~71-57 Ma crystallization ages of numerous undeformed leucogranites that crosscut the transposition foliation and post-transposition folds (Sevigny et al., 1989; Scammell, 1993; Crowley et al., 2003; Gibson et al., 2005). Mica $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages range from ~55 to 52 Ma in the northern part of this transect (Scammell, 1993; Ghent and Villeneuve, 2006), which is only 2-6 my older than mica $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages obtained in the LSa cropping out to the west of the Monashee Complex (Fig. 3.3).

It is important to stress that isograds and map units are continuous *along the NW-SE strike* across the Columbia River from the Selkirk to the Monashee Mountains and that no tectonic boundary has been identified *across strike* toward the southwest in the Monashee Mountains (Fig. 3.2). Therefore, high-grade rocks located in the Selkirk and northern Monashee Mountains are physically linked to rocks located west of the Monashee Complex and they all share similar metamorphic, structural and geochronologic characteristics. The continuity of the LSa around the Monashee Complex and into the Selkirk Mountains is emphasized because it has been ignored in most studies that were conducted in the southern part and further south of the map of Fig. 3.2 (e.g. Parrish et al., 1988; Bardoux and Mareschal, 1994; Vanderhaeghe and Teyssier, 1997; Vanderhaeghe et al., 1999; Teyssier and Whitney, 2002; Teyssier et al., 2005). A key implication of this observation is that the northeasternmost exposure of LSa is not bounded by brittle-ductile normal faults, as is the case along section E-F (Fig. 3.2 & 2.3c). Another key implication is that the Selkirk detachment occupies the same

structural position as the OVfs with USa in its hangingwall and LSa in its footwall. The two structures thus likely formed one that was eroded away above the Monashee Complex and down-faulted by the Columbia River Fault to the east of the complex, as suggested by Brown and Gibson (2006) and illustrated on cross-section A-B of Fig. 3.3a.

From studies conducted near the Purcell Thrust (Fig. 3.3), Crowley et al. (2003) and Gibson (2003) documented that migmatitic rocks of the LSa, which were deformed and metamorphosed from Mid-Cretaceous up to Paleocene, structurally overlie lower-amphibolite and greenschist facies rocks that record a main history of deformation and metamorphism between ~171-130 Ma. Boggs (2004) obtained $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende plateau ages of ~89 and ~83 Ma in the lower-plate of the same area suggesting that the main period of cooling below 550-500°C took place while overlying rocks of the LSa were at their thermal peak. There would thus be another USa-LSa boundary, with an opposite polarity, in the northeastern part of the map of Fig. 3.2. Determining the nature of this boundary will be a matter of future research because it passes within rocks inferred to have remained stitched together throughout their tectonic history that involved three generations of folding accompanied by the development of metamorphic isograds (Simony et al., 1980; Ghent et al., 1980; Raeside and Simony, 1983; Digel et al., 1998; Ghent and Simony, 2005). Finally, the western boundary of the LSa in the northern part of Fig. 3.2 is not well constrained. It is tentatively traced parallel to the North Thompson segment of the Okanagan Valley fault system.

3.3.5 Discussion and summary

In summary, from the three cross-sections described above, it is clear that rocks of the LSa share many similarities in terms of strain patterns, metamorphic grade; and

timing of deformation, metamorphism and cooling. In all three sections, the LSa consists mainly of sillimanite-bearing migmatites intruded by >30% of variously strained leucogranite sheets. Structurally, these rocks are L-S tectonites characterized by westerly plunging stretching lineation (Fig. 3.2). The paucity of kinematic indicators in the centre of the LSa along cross-sections C-D (Johnson, 1994) and E-F (Glombick et al., 2006a), and the perfect alignment of minor fold axes and stretching lineations documented by Scammell (1993) north of the Monashee Complex (Fig. 3.2a, b) suggest a coaxial component of flow. By contrast, the ubiquitous presence of kinematic indicators in the mylonites forming the oppositely verging shear zones that bound the LSa suggest that simple shear was the dominant flow type there. Along the three cross-sections, a top-to-the-west brittle-ductile fault—the Okanagan Valley fault system—separates the overlying lower-amphibolite and greenschist facies rocks of the USa, which were deformed and metamorphosed in the Jurassic, from sillimanite-migmatites of the LSa, which remained hot and ductile until the Late-Cretaceous to Paleocene. By contrast, a top-to-the-east ductile shear zone—the Monashee décollement—separates the LSa from underlying kyanite-bearing migmatites of the Monashee Complex that were deformed, metamorphosed and cooled between ~60-50 Ma. All rocks of the LSa, from the Selkirk Mountains to the Vernon area, record a complex and protracted history of high-grade metamorphism. At many localities, this history started in the Jurassic, and most localities record an important period of high temperature metamorphism and deformation from the Late Cretaceous to the Early Eocene (~120-56 Ma). Therefore, it is appropriate to group all these rocks into one unit—the LSa—as done herein.

An important criterion for testing the different tectonic models presented on Fig. 3.1

is the timing of motion along the OVfs and the Monashee décollement. Zircon growth in leucosome from the Monashee décollement and the OVfs overlap between ~90-59 Ma. Because hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages in the footwall of the OVfs are the same within uncertainty as the youngest dates of magmatic zircon at ~57 Ma, it is most likely that the last increments of high-temperature (close to solidus) top-to-the-west shear strain along the fault took place at, and a few millions years earlier, than ~57 Ma. Similarly, along the Monashee décollement at Cariboo Alp (CA on Fig. 3.2), the age of the youngest leucosome, the age of undeformed cross-cutting dykes, and hornblende cooling ages, all fall within the bracket ~59-57 Ma. In northern Frenchman Cap dome (near C on Fig. 3.2) and at Three Valley Gap (TVG on Fig. 3.2) the brackets are less precise but compatible with these numbers, being ~85-57 Ma and ~65-58 Ma, respectively. These data thus suggest synchronous motion along the two oppositely verging shear zones between ~65-57 Ma. Alternatively, it is possible that shearing in northern Frenchman Cap dome occurred closer to ~85 Ma than ~60 Ma. The next section presents results of U-Pb geochronology of a synkinematic leucosome pod from the Monashee décollement in northern Frenchman Cap dome that resolves this issue.

Another piece of crucial information for testing tectonic models of exhumation is the P-T-t path followed by rocks located at different structural levels, especially along the two bounding shear zones of the LSa. In pelitic schists of the two shear zones, the shear fabric is formed by reactant and products of ReNtr 1. Because ReNtr 1 requires cooling, shearing along the two shear zones occurred along a cooling path. Furthermore, cooling was most likely fast because hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ ages are within uncertainties of the youngest melting episode. However, it is not known whether this cooling path was

accompanied by decompression or was isobaric. The progression of shear fabrics from high temperature ductile to brittle observed along the OVfs (Johnson, 2006) does suggest that decompression accompanied cooling, but it is equally possible that brittle faulting occurred millions of years later. In the next section, a pressure-temperature-time-deformation path is constructed for rocks of the Monashee décollement at the latitude of northern Frenchman Cap dome. The information gained from this path fills the gap mentioned above and facilitates testing different tectonic models of exhumation illustrated on Fig. 3.1.

3.4 New pressure-temperature-time-deformation (P-T-t-d) paths from the base of the LSa

3.4.1 Methodology

See section 1.5

3.4.2 Metamorphic constraints

3.4.2.1 Field observations

Specimens from the Monashee décollement at the base of the LSa were collected at area C on Fig. 3.2 to reconstruct a P-T-t-d path. Rocks in this area consist of interlayered migmatitic sillimanite-bearing semi-pelite, metapelite, quartzo-feldspathic gneiss, and boudinaged amphibolite. They are mylonitic and record top-to-the-east sense of shear in a variety of kinematic indicators (Fig. 3.4). Fig. 3.4c illustrates leucosome pooling preferentially in an amphibolite boudin neck where amphibolites are segmented into angular bodies. These observations indicate that melt was present during deformation. Metapelites at the base of the LSa within the Monashee décollement, consist of mylonitic sillimanite-bearing migmatitic schists that contain between 15 and 30% of leucosome.

The outcrop volume of leucosome and leucogranitic intrusions is more than 40%.

Leucosome forms millimetric granitic veinlets that are part of a C-S fabric indicating a top-to-the-east shear sense; or occurs as millimetric pods that crosscut the foliation.

These small veinlets and pods are uniformly distributed throughout the rocks; whereas larger, centimetric to decametric, granitic pods are common, but impossible to distinguish from dismembered granitic veins because of the high degree of strain that has affected these rocks during and after melt crystallization. Kyanite was observed by Journeay (1986) in only a few specimens and was reported to be extensively resorbed or almost completely transformed into sillimanite. Garnet is rather rare, commonly surrounded by leucosome, and highly resorbed on faces normal to the stretching lineation, which suggests resorption in the presence of melt that migrated toward the direction of minimum shortening.

3.4.2.2 Specimen Fg462a: migmatitic sillimanite-bearing semi-pelitic schist

Specimen Fg462a is a migmatitic rutile-garnet-ilmenite-sillimanite-K-feldspar-biotite schist that was used to estimate peak metamorphic conditions. Sillimanite is present throughout the matrix and around resorbed garnet porphyroblasts and has a prismatic or fibrolitic shape. No relict kyanite was observed. Garnet comprises only 3-5% of the mineral assemblage and commonly show preferential resorption of their sigmoidal tails that indicate a top-to-the-east sense of shear (Fig. 3.5a). Sillimanite, biotite, ilmenite, plagioclase and quartz are the most common minerals observed in these sigmoidal tails (Fig. 3.5a). The largest garnet porphyroblast observed in this sample has a diameter of ~3 mm. Its core has a flat compositional zoning of $X_{\text{sps}} = 0.05$, $X_{\text{grs}} = 0.04$, $X_{\text{pyr}} = 0.23$ (Table 3.1), but its margins show an increase of X_{sps} (0.08), and an increase in the

Fe/Fe+Mg ratio (0.80). A smaller garnet porphyroblast (0.5 mm) has a similar core composition but its resorbed margins show a larger increase in X_{sps} (0.16) and in the Fe/Fe + Mg ratio (0.85; Table 3.1). Such compositional spikes at garnet margins are commonly interpreted as evidence for garnet breakdown in the presence of melt by ReNtr 1 (Spear et al., 1999; Kohn and Spear, 2000). The anorthite content of plagioclase increases from ~ 0.20 to ~ 0.23 in proximity to these margins.

More quantitative information on the P-T path followed by specimen Fg462a is obtained from phase equilibria modeling with the software *Perple_X'06* (Connolly and Petrini, 2002; Connolly, 2005). P-T diagrams calculated with this software are divided into a mosaic of polygons with distinct metamorphic assemblages. It is thus possible to simplify such a diagram by tracing lines marking the disappearance, or first appearance of certain minerals. Fig. 3.6 presents the resulting simplified pseudosection calculated with *Perple_X'06* for the bulk chemical composition of specimen Fg462a in the system TiMnCaNaKFMASH, including a melt model (White et al., 2001) and a Ti-Mn-biotite solid-solution model (Powell and Holland, 1999). This biotite solid-solution model was the only available with the version of *Perple_X* used herein. However, because this solution model underestimates the upper stability temperature of biotite (White et al., 2007; Tajcmanova et al., 2009), the biotite-out (Bt-out) curve on Fig. 3.6 probably represents a minimum. Also shown on Fig. 3.6 are important reactions and results of geothermobarometry for specimen FG-462d described in the next paragraph. P-T fields delineated by these lines of appearance/disappearance provide important constraints on the possible P-T path followed by specimen Fg462a. Hence, the presence of K-feldspar, leucosome and biotite, combined with the absence of kyanite, muscovite and cordierite

constrain peak metamorphic conditions and the retrograde path to a P-T space delineated by the muscovite-out, biotite-out, cordierite-out and sillimanite-in lines.

A more quantitative estimate of pressure and temperature conditions during homogenization of garnet is provided by garnet isopleth thermobarometry. *Perple_X'06* does not calculate uncertainty, but, because the garnet solid-solution model of Holland and Powell (1998, revised 2002) is used, it is possible to calculate uncertainty associated with garnet composition with the program *AX* (Holland and Powell, 1998). The black triangle on Fig. 3.6 represents the intersection of $X_{\text{pyr}}-X_{\text{grs}}-X_{\text{sps}}$ within calculated uncertainty represented by the three zones with different shades of gray. The calculated P-T condition of ~ 800 MPa/ 760°C is also in agreement with the anorthite isopleth ($An = 0.20$ for plagioclase away from resorbed garnet margin) that is parallel to the muscovite-out reaction (not shown). Garnet consumption by *ReNrt 1* requires cooling because garnet volume isopleths in the sillimanite-melt field are nearly isothermal (not shown). Similarly, production of more calcic plagioclase requires decompression because anorthite isopleths are nearly parallel to the muscovite-out reaction and increase toward high-temperature and low-pressure (not shown).

In summary, the results indicate that garnet of specimen FG-462a was homogenized at ~ 800 MPa/ 760°C . These P-T conditions represent a minimum estimate of the metamorphic peak, because it is possible that the composition of garnet reequilibrated with that of the matrix along the retrograde path. The retrograde path is constrained to a P-T field delineated by the muscovite-out, the biotite-out, the kyanite-sillimanite, and the cordierite-out curves. The presence of the product of *ReNtr 1* in pressure shadows of resorbed garnet further suggests that this retrograde path took place during top-to-the-east

shearing.

3.4.2.3 Specimen Fg462d: synkinematic clinopyroxene-bearing leucosome

The second specimen examined, Fg462d, is a leucosome containing a sigmoidal clinopyroxene-plagioclase-titanite aggregate (Fig. 3.4d) that is used to reconstruct the lower pressure evolution of the area and to constrain the timing of the retrograde path. The host of the leucosome is a clinopyroxene-hornblende-biotite mylonitic leucogneiss that contains numerous boudinaged amphibolite layers (Fig. 3.4a, c). As shown on Fig. 3.4c, coarse clinopyroxene-bearing leucosome pods are commonly associated with amphibolite boudins and interpreted as the product of amphibolite melting via a reaction such as:

This reaction is the low-pressure equivalent of amphibolite melting reaction with garnet as an additional product (e.g. Wyllie and Wolf, 1993). Products of Rx 2, without orthopyroxene nor garnet, have been observed during partial melting experiments of amphibolite conducted at pressures below ~1000 MPa (Wyllie and Wolf, 1993 and references therein; Kawamoto, 1996; Springer and Seck, 1997; Lopez and Castro, 2001; Nair and Chacko, 2008). These experiments were conducted at temperature above 800°C, but it is likely that the product of Rx 2 could also form by fluid-present melting in the temperature range ~700-750°C (Wyllie and Wolf, 1993; Liu et al., 1996).

Specimen Fg462d consists of a clinopyroxene aggregate surrounded by K-feldspar + plagioclase + quartz leucosome (Fig. 3.4d). The aggregate itself mainly consists of an intergrowth of clinopyroxene, titanite, plagioclase, K-feldspar, and quartz. Fig. 3.5b illustrates the stark contrast between the fine-grained, recrystallized and quartz-

ribbon microstructure of the host (left side); and the coarse-grained, euhedral and intergrowth texture of the leucosome (right side). Clinopyroxene is 0.5-1.5 mm wide, is euhedral to subhedral, and commonly presents crystal faces against quartzo-feldspathic material (Fig. 3.5c). Titanite is euhedral (Fig. 3.5c, d; 3.8d) and has an intergrowth texture with clinopyroxene (Fig. 3.5c). Feldspars are also euhedral to subhedral and commonly have crystal faces, as can be observed on Fig. 3.5d. The intergrowth texture, the euhedral shape of grains and the numerous crystal faces are interpreted as evidence for growth of clinopyroxene and titanite in the presence of melt (criteria from Vernon, 2004). Because similar clinopyroxene-bearing leucosome pods are common in the study area and interpreted as products of amphibolite partial melting via Rx 2, specimen Fg462d is also interpreted to be the product of Rx 2.

A striking feature of specimen Fg462d is the concentration of leucosome and hornblende in the top-to-the-east pressure shadows of the clinopyroxene aggregate (Fig. 3.4d). These observations are interpreted as evidence for migration of melt in the pressure shadows of the aggregate during top-to-the-east shearing and for formation of hornblende by the inverse of Rx 2 upon cooling.

Conventional thermobarometry was conducted on specimen Fg462d to examine its cooling P-T path. It was not possible to adopt a more powerful phase equilibria modeling approach because the metamorphic evolution of specimen Fg462d involved partial melting of an amphibolite, for which there is currently (now in 2009) no appropriate solid-solution model. Clinopyroxene, hornblende and plagioclase are unzoned and their compositions do not vary with textural setting. Clinopyroxene is diopsidic with Mg/Mg+Fe ratio of 0.68, and Na/Ca+Na ratio of 0.11-0.12. Hornblende is a magnesio-

hornblende (Table 3.2). Plagioclase has a uniform anorthite activity of 0.16-0.17 (as per Holland and Powell, 1992; Table 3.1). Mineral activities used for thermobarometric calculations with Thermocalc were calculated with the program A-X version 2005 (Holland and Powell, 1994).

The average PT method of Thermocalc version 3.31 (Holland and Powell, 1994) was used in combination with other experimentally-calibrated thermobarometers for specimen Fg462d. The average composition of hornblende, clinopyroxene and plagioclase were used for the calculations. Thermocalc identified a set of four independent reactions between end-members of the assemblage clinopyroxene-hornblende-plagioclase-quartz:

- 1) Ca-tschermak (cpx) + quartz = anorthite
- 2) 2 diopside + tschermakite + 2 quartz = tremolite + 2 anorthite
- 3) 5 hedenbergite + tschermakite + 2 quartz = 3 diopside + ferroactinolite + 2 anorthite
- 4) diopside + 3 Ca-tschermak (cpx) + tschermakite + albite = pargasite + 4 anorthite

The average P mode yields unacceptable results with $\sigma_{\text{fit}} = 4.0$, which is much higher than the cut-off value of 1.61. However, Ca-tschermak (cpx) can be identified as the outlier end-member because of an e_k^* value of -3.8, which is 1.4 from the cut-off value (see Holland and Powell, 1994 for a comprehensive explanation). Removing this end-member from the calculation results in three independent reactions that are the same as reactions 2-4 above, except for 4 in which Ca-tschermak (cpx) is removed from the reactant and tremolite added to the product. The average P now yields acceptable results with a $\sigma_{\text{fit}} = 1.0$ -1.1 and yields pressures ranging from 280 ± 128 MPa at 650°C to $260 \pm$

135 MPa at 750°C (uncertainties at the 1σ level; line 3 on Fig. 3.6). In addition, it is possible to use the previously discarded reaction 1 for barometric calculations because this reaction has been experimentally calibrated by McCarthy and Patiño-Douce (1998). This barometer, which is commonly called “SCAN”, yields pressures of 270 ± 100 MPa at 650°C and 300 ± 100 MPa at 750°C (uncertainties at the 1σ level; line 5 on Fig. 3.6). The experimentally-calibrated Al-in hornblende barometer of Anderson and Smith (1995) can be used because specimen Fg462d contains the necessary buffer assemblage consisting of K-feldspar-quartz-biotite (present in trace amount)- titanite-ilmenite, and a melt phase. The barometer yields pressure of 168 ± 60 MPa at 650°C, and 77 ± 60 MPa at 750°C (uncertainties at the 1σ level). Finally, an estimate of temperature is obtained from the two hornblende-plagioclase thermometers of Holland and Blundy (1994). The edenite + albite = richterite + anorthite, yields 617 ± 40 °C at 0 MPa and 627 ± 40 °C at 500 MPa; and the quartz-saturated edenite + 4 quartz = tremolite + albite thermometer yields 671 ± 40 °C at 0 MPa and 628 ± 40 °C at 500 MPa.

The uniform composition of clinopyroxene, plagioclase and hornblende in specimen Fg462d, combined with the excellent $\sigma_{\text{fit}} = 1.0\text{-}1.1$ for three independent reactions obtained by Thermocalc indicate that equilibrium between these three minerals was reached. Average pressures of ~ 270 MPa at 700°C calculated with Thermocalc (Fig. 3.6) line are consistent with the pressure estimate by the SCAN barometer at ~ 285 MPa and 700°C (Fig. 3.6 stippled lines 3 and 5). The Al-in hornblende barometer of Anderson and Smith (1995) yields lower pressure between 168-77 MPa. However, Bachmann and Dungan (2002) demonstrated that this barometer yields pressures lower by at least 200 MPa for low-Al hornblende ($< 6\%$ Al_2O_3) compared to high-Al hornblende in igneous

rocks of the Fish Canyon magmatic system. Adding 200 MPa to the obtained pressures for specimen Fg462d brings the pressure for the Al-in barometer to similar pressures as the other two barometers (line 4 on Fig. 3.6). Such a procedure is justified because the Al_2O_3 content is below 6% for hornblende of specimen Fg462d (5.5%, Table 3.1). For an estimate of temperature, the quartz-saturated hornblende-plagioclase thermometer (line 6 on Fig. 3.6; Holland and Blundy, 1994) is preferred because it yields temperature closer to the wet-granite solidus (Fig. 3.6; curve 2 from Ebadi and Johannes, 1991). Because hornblende is interpreted to have formed during crystallization of a melt phase, the preferred P-T condition of equilibrium during hornblende formation in specimen Fg462d is estimated at $\sim 700^\circ\text{C}/300\text{ MPa}$ (ellipse on Fig. 3.6), which is: 1) the intersection of the granite solidus curve; 2) the pressure curve derived from the average P method of Thermocalc; 3) the SCAn barometer; 4) the Al-in hornblende barometer with an addition of 200 MPa; and 5) within uncertainty of the quartz-saturated hornblende-plagioclase thermometer.

3.4.3 *U-Pb LA-MC-ICP-MS geochronology*

3.4.3.1 Zircon

Twenty-six U-Pb analyses from eighteen zircon grains extracted from specimen Fg462d were conducted by the Laser Ablation MC-ICP-MS method (following the procedure outlined in the methodology section 1.5 and Simonetti et al., 2005 & 2006). The $^{206}\text{Pb}/^{204}\text{Pb}$ ratio of all analyses was much higher than a 1000, such that no correction for common Pb was applied (Lugwig, 1998). I followed Ludwig (2003) for uncertainty corrections and multiplied the 1σ analytical uncertainty by the Student's-t factor and by the square root of the MSWD when the calculated probability of fit was less than 0.05, or

when the number of analyses examined was less than six. CL images of zircon grains from this specimen show rounded cores surrounded by dark rims that are usually faceted but rarely oscillatory zoned (Fig. 3.7a-c). U-Pb dates of zircon cores range in age from Archean to Devonian (Table 3.3) and are thus interpreted to reflect xenocrystic inheritance. U-Pb dates of zircon rims that fall on, or near, concordia yield dates of ~120, 107, 83, 62 Ma with two clusters at ~83 Ma and ~62 Ma. The older cluster of rim dates represents four analyses from three grains and yields a weighted mean average $^{206}\text{Pb}/^{238}\text{U}$ age of 83.2 ± 4.9 Ma (MSWD = 1.6; prob. fit = 0.19; Fig. 3.7d). The younger cluster of rim dates represents five analyses from four grains and yields a weighted mean average $^{206}\text{Pb}/^{238}\text{U}$ age of 62.2 ± 2.0 Ma (MSWD = 3.3; prob. fit = 0.01; Fig. 3.7d). An interesting grain is zircon L-2 (Fig. 3.7a) from which four analyses from the two clusters were measured. Three analyses from the core of the grain yields $^{206}\text{Pb}/^{238}\text{U}$ dates of ~83 Ma and one analysis of the rim of the grain yields a date of ~65 Ma. The core of this grain has faint concentric zoning (Fig. 3.7a) and is separated from its rim by a thin black layer in CL image that defines a euhedral faceted interior crystal. The rim overgrows the euhedral faceted core and is itself faceted parallel to the boundaries of the core. Euhedral faceted zircon is commonly interpreted as evidence for the presence of a melt phase (Corfu et al., 2003). Therefore, I interpret the euhedral faceted internal part of grain L-2 and the euhedral faceted morphology of every ~62 Ma rim as indicative of the presence of a melt phase during these two periods of zircon growth.

3.4.3.2 Titanite

Titanite in specimen Fg-462d is ~300x600 μm , is euhedral, and is in textural equilibrium with clinopyroxene and plagioclase (Fig. 3.5c,d). Titanite has dark cores

surrounded by light rims when examined on BSE images (Fig. 3.8a-d). Titanite naturally contains a significant amount of common Pb. Consequently, great care must be taken in the method used for the common Pb correction (see Storey et al., 2006). Three methods can be used to correct for initial common Pb in isotopic techniques that cannot precisely measure ^{204}Pb such as the LA-MC-ICP-MS technique used herein (Aleinikoff et al., 2004). The first is to regress the uncorrected data on a Terra-Wasserburg semi-total $^{207}\text{Pb}/^{206}\text{Pb}$ vs. $^{238}\text{U}/^{206}\text{Pb}$ plot and use the lower concordia intercept as the age of titanite growth. The second method is to use the y- $^{207}\text{Pb}/^{206}\text{Pb}$ intercept on the Terra-Wasserburg and calculate the weighted average of $^{206}\text{Pb}/^{238}\text{U}$ dates corrected by the ^{207}Pb correction method (c.f. Stern, 1997; Williams, 1998). The third method is to use the model terrestrial Pb curve of Stacey and Kramers (1975) as the initial $^{207}\text{Pb}/^{206}\text{Pb}$ for the ^{207}Pb correction method. I follow Aleinikoff et al. (2004) and Storey et al. (2006) and use the second method because it provides a quantitative estimate of common Pb and yields more precision. Regression of 21 of the 22 uncorrected analyses on a Terra-Wasserburg plot intercepts concordia at 54.3 ± 26.9 Ma (MSWD= 10.9; prob. fit = 0.0) and yields a y-axis $^{207}\text{Pb}/^{206}\text{Pb}$ intercept of 0.61 ± 0.11 (Fig. 3.8e). Using the y-axis for common Pb correction yields a corrected weighted mean average $^{206}\text{Pb}/^{238}\text{U}$ age of 54.4 ± 1.7 Ma (MSWD =1.08; prob. fit = 0.36; Fig.3.8f; Table 3.4). This age is the same within uncertainty as the age of magmatic titanite grains from the undeformed leucogranite RS-465 collected at the same outcrop as specimen Fg462d that yielded a U-Pb age of 54.1 ± 1.4 Ma (Scammell, 1993).

3.4.4 Titanite diffusion modeling

Diffusion modeling of titanite cooling was conducted in order to unravel the T-t path

followed by rocks of the Monashee décollement. The extended Dodson's closure temperature formulation (Dodson, 1973) derived by Ganguly and Tirone (1999) was used via the program DIFF-CL, and the diffusion parameters of titanite determined by Cherniak (1993) were incorporated into the calculations. The dataset derived from specimen Fg462d as well as the U-Pb age of zircon and titanite from a ~3m thick leucogranite dyke (RS465 from Scammell, 1993) crosscutting the foliation within the same outcrop as specimen Fg462d impose important constraints on possible cooling paths. IDTIMS U-Pb ages of zircon and titanite fractions from leucogranite RS465 yielded concordant ages of 57.0 ± 0.2 Ma and 54.1 ± 1.4 Ma, respectively (Scammell, 1993). Because it has a thickness of only ~3 m, the dyke would have homogenized to the temperature of its host in a very short period of time (e.g. Spear, 1993, p.45). Accordingly, the first assumption of the modeling is that the zircon U-Pb age is the emplacement age of the dyke. The temperature range of zircon crystallization in dyke RS465 is estimated at ~650-750°C from zircon saturation thermometry conducted on a suite of Late-Cretaceous leucogranite collected in the northern Monashee Mountains (Sevigny et al., 1990). Because of the euhedral shape of titanite grains and the inferred fast crystallization of the dyke, the second assumption is that titanite crystallized in the melt at the same time as zircon. Similarly, the third assumption is that titanite from specimen Fg462d — dated at 54.4 ± 1.7 Ma (Fig. 3.8) — crystallized in the melt at the same time as zircon grains/rims formed at ~62 Ma. The U-Pb age of titanite grains from the two specimens investigated are thus younger than their crystallization age, which implies crystallization of titanite grains from both specimens at a temperature higher than the closure temperature for diffusion of Pb in titanite. Furthermore, the postkinematic

dyke RS465, must have been emplaced after formation of the synkinematic hornblende in the pressure shadows of specimen Fg462d, for which results of geothermobarometry suggest a crystallization at a temperature of $\sim 700^{\circ}\text{C}$ (Fig. 3.6). Hence, titanite grains of specimen Fg462d must have crystallized at a higher temperature than the emplacement of the dyke. In addition, the cooling rate between the emplacement of dyke RS465 at ~ 57 Ma and the age at which titanite grains from both specimens closed to Pb diffusion at ~ 54 Ma must be the same for both specimens because both specimens are from the same outcrop. Finally, the radius of titanite grains from specimens RS465 and Fg462d average $\sim 75\text{ }\mu\text{m}$ (Scammell, 1993) and $\sim 125\text{ }\mu\text{m}$, respectively.

Fig. 3.9 presents several age profiles for titanite grains of specimens RS465 (Fig. 3.9a) and Fg462d (Fig. 3.9b) modeled for different temperatures of formation and cooling rates. Other permutations were attempted but are not shown because they were not realistic. Initial modeling conducted at cooling rates above $30^{\circ}\text{C}/\text{my}$ demonstrated the impossibility of obtaining the same U-Pb age in titanite grains crystallized ~ 5 my apart. Three cooling curves fall within the age bracket of 53-56 Ma given by the measured U-Pb age of titanite from leucogranite dyke specimen RS465 plus or minus uncertainty (Fig. 3.9a), whereas two curves fall within this age bracket for titanite from the clinopyroxene-bearing leucosome specimen Fg462d. For specimen RS465, two curves involve cooling from 700°C at cooling rates of 20 and $30^{\circ}\text{C}/\text{my}$, and the other involves cooling from a temperature of 660°C at a rate of $10^{\circ}\text{C}/\text{my}$. Initial modeling results indicated that cooling at rates higher than $\sim 10^{\circ}\text{C}/\text{my}$ starting at ~ 62 Ma for titanite grains of specimen Fg462d would yield U-Pb age older than the 53-56 Ma age bracket, but that a two-step cooling pattern would be possible for a cooling rate of $20^{\circ}\text{C}/\text{my}$. One of the

two possible cooling curves for specimen FG462d that falls within the 53-56 Ma age bracket involves cooling from a temperature of 760°C at a rate of 20°C/my. This curve was calculated assuming that cooling at a rate of 20°C/my started at a temperature of 760°C, which is the minimum estimate for the thermal peak (Fig. 3.6). This temperature would have been attained at 60 Ma in order to reach a temperature of 700°C at 57 Ma, as indicated by the emplacement age of the dyke RS465. At these temperatures, modeling indicates that titanite is completely open for Pb diffusion. This scenario thus implies no cooling and temperature equal to, or higher than 760°C, between 62 and 60 Ma. The second possible scenario involves cooling at a constant rate of 10°C/my from a temperature of 710°C at 62 Ma.

Consequently, two possible temperature-time models fit the imposed constraints described in the previous paragraph. The first involves formation of titanite in specimen Fg462d at ~62 Ma at a temperature higher than 760°C. Titanite would then have remained open for Pb diffusion until 60 Ma when cooling started and/or continued at a rate of 20°C/my from 59 to 54 Ma. In this model, titanite grains from leucogranite RS465 crystallized with zircon at 57 Ma at a temperature of 700°C and closed to Pb diffusion at 54 Ma at a temperature of 640°C. The second scenario involves formation of titanite in specimen Fg462d at ~62 Ma at a temperature 710°C followed by cooling at a rate of 10°C/my down to a temperature of 630°C, reached at 54 Ma. For leucogranite RS465, the difference with the first model is the crystallization of titanite at a temperature of 660°C instead of 700°C.

3.4.5 *P-T-t-d path*

The dataset presented herein provides good constraints to derive a pressure-

temperature-time-deformation path for the Monashee décollement at the base of the LSa. Phase equilibria modeling of migmatitic metapelitic schist Fg-462a indicates that garnet was homogenized at ~ 800 MPa/ 760°C in equilibrium with plagioclase. At these metamorphic conditions, partial melting of amphibolite via Rx 2 is possible depending on bulk composition and water activity. For example, the coexistence of clinopyroxene and melt was obtained at $\sim 725^{\circ}\text{C}/500$ MPa for one amphibolite composition in the partial melting experiments of Springer and Seck (1997) conducted with excess water. For amphibolite layers embedded in partially molten metasediment, water could be provided by water expelled from crystallizing leucogranite. It is impossible with the available data to constrain the timing of this thermal peak better than the bracket of ~ 120 and 62 Ma imposed by U-Pb dates of zircon rims from specimen Fg462d. The retrograde path is better constrained by conventional thermobarometry for specimen Fg462d, coupled to titanite and zircon U-Pb ages from this specimen as well as from leucogranite dyke RS465 (Scammell, 1993), and to diffusion modeling for titanites from these two specimens. Results of thermobarometry indicate equilibration for the assemblage clinopyroxene-hornblende-plagioclase-quartz of specimen Fg462d within a P-T field centered at $\sim 700^{\circ}\text{C}/300$ MPa (ellipse on Fig. 3.6). The retrograde P-T path is shown to pass through the upper pressure limit of this P-T field in order to avoid crossing the cordierite-in reaction above the granite solidus (Fig. 3.6). The retrograde path thus has a steep $\Delta P/\Delta T$ slope (Fig. 3.6).

Two possible cooling paths were derived from diffusion modeling. The first path involves cooling from 760°C to 700°C between 60 and 57 Ma. According to the P-T path presented on Fig. 3.6, this cooling would be associated with decompression of >500 MPa

at a rate of >167 MPa/my or >6 mm/y, assuming an average crustal density of ~ 2850 kg/m³. This path leaves a small time window of less than 1 my for the synkinematic crystallization of hornblende in specimen Fg-462d (estimated to have occurred at $\sim 700^{\circ}\text{C}$) before the emplacement of the postkinematic dyke RS465. Furthermore, a pressure of ~ 800 MPa at ~ 62 Ma would imply decoupling with overlying rocks of LSa at Ruddock Creek (Fig. 3.2 & 3.3a for location) studied by Scammell (1993) that were at less than ~ 450 MPa and had cooled below the hornblende closure temperature by ~ 62 Ma. By contrast, the second possible diffusion model indicates a temperature of 710°C at 62 Ma, which corresponds to a pressure of ~ 400 MPa along the P-T path of Fig. 3.6. In this model, dyke RS465 would have been emplaced at 660°C , which leaves time and temperature windows of ~ 4 my and $\sim 30^{\circ}\text{C}$, respectively, between the formation of the synkinematic hornblende in specimen Fg462d and the emplacement of the dyke. The second diffusion model is preferred because it leaves appropriate time and temperature windows between the formation of the synkinematic hornblende in specimen Fg462d and the emplacement of the dyke RS465, and it does not require a major structural decoupling with overlying rocks from Ruddock Creek at ~ 62 Ma.

Consequently, the P-T-t-d path proposed for rocks of the Monashee décollement at the base of LSa involves a thermal peak of $\sim 760^{\circ}\text{C}/800$ MPa loosely bracketed between ~ 120 and 62 Ma, which was most likely a period of protracted high temperature metamorphism accompanied by episodic zircon growth. From the thermal peak, the retrograde path was nearly isothermal and took place during top-to-the-east shearing that resulted in the formation of hornblende via Rx 2 in the pressure shadows of a clinopyroxene-titanite-feldspar aggregate between ~ 62 and 57 Ma at P-T conditions of

~700°C/300 MPa. The end of deformation is bracketed by the ~57 Ma U-Pb age of the leucogranitic dyke RS465 that cross-cuts the foliation at right angle. Titanite diffusion modeling indicates cooling at a rate of 10K/my from a temperature of ~710°C at ~62 Ma down to a temperature of ~630°C at ~54 Ma.

3.5 Timing of motion along the two oppositely verging shear zones bounding the LSa

The P-T-t-d path constructed in the previous section fills crucial gaps for testing tectonic models of exhumation. First, it is the first high spatial resolution geochronologic technique to confirm previous results obtained by the IDTIMS U-Pb geochronologic technique (Carr, 1992; Johnson, 1994) that indicated that the last increments of top-to-the-east shear strain along the Monashee décollement occurred between ~62 and 57 Ma. Second, these results indicate that this late motion was synchronous along the entire ~150 km length of the Monashee décollement, from northern Frenchman Cap dome to southern Thor-Odin domes. Most importantly, this ~62-57 Ma bracket coincides with the timing of the last increments of high temperature top-to-the-west shear strain along the OVfs (see section 3.3.2 and Table 3.4). The two oppositely verging shear zones bounding the LSa were thus shearing at the same time, at least in the Early to Late Paleocene (Table 3.4). Finally, this P-T-t-d path is the first clear evidence that shearing along the Monashee décollement led to ~500 MPa of exhumation at high temperature (>650°C). Given that top-to-the-east shearing at the base of the LSa was synchronous with top-to-the-west shearing at the roof of the LSa in the Early to Late Paleocene, the exhumation path recorded at the base of the LSa implies that the LSa was getting closer to the Earth's surface by coeval, oppositely verging shearing at its base and roof.

3.6 Depth-time paths

Fig. 3.10 presents seven depth-time paths, five of which are for rocks of the LSa and the remaining from the underlying Monashee Complex, respectively labelled A-E and F, G. From A to E, the P-t paths are located progressively southwestward along an azimuth perpendicular to the NW-SE axis of the Monashee Complex (Fig. 3.3). The axis of the complex passes a few kilometers NE of area C (Fig. 3.2). Pressure was converted into depth by assuming an average crustal density of 2850 kg/m^3 . Table 3.5 lists all the data used to construct the P-t paths. On Fig. 3.10, P-T-t points are illustrated by solid ellipses and the different paths have been adjusted in order to pass through these points. All paths of the LSa investigated are at the sillimanite-K-feldspar grade with muscovite and kyanite absent from aluminous schist, which imposes a maximum of $\sim 1000 \text{ MPa}$ for peak pressure. In addition, the large majority of peak pressure estimates for the LSa lie in the range $700\text{--}900 \text{ MPa}$ for temperature between the experimentally and thermodynamically constrained upper stability temperature of muscovite and biotite, at ~ 700 and 850°C , respectively (see Fig. 3.6; Vielzeuf and Montel, 1994; Patiño-Douce and Harris, 1998; Gardien et al., 2000; Patiño-Douce and Beard, 1996; White et al., 2007). The average geothermal gradient calculated from retrograde P-T points is 1.6°C/MPa , and this value was used to assign a pressure to the estimated closure temperature of a given thermometer for data that do not have a specific P-T-t data point. Closure temperatures were assumed to be: $650 \pm 50^\circ\text{C}$ for titanite (Cherniak, 1993), $500 \pm 50^\circ\text{C}$ for hornblende (Harrison, 1981), $350 \pm 50^\circ\text{C}$ for muscovite (Hames and Bowring, 1994), and $300 \pm 50^\circ\text{C}$ for biotite (Harrison et al., 1985). Data for paths F and G are derived from chapter 4 of this thesis.

The major conclusion from the pressure-time paths presented on Fig. 3.10 is that rocks of the LSa and rocks of the Monashee Complex have contrasting evolution. Areas A-E from the LSa all share similar paths with a protracted residence time in the middle crust and exhumation to less than ~14 km depth between ~80 and 50 Ma. Also noteworthy is the trend of progressively younger exhumation from A to E along an azimuth perpendicular to the axis of the Monashee Complex. Because the LSa is allochthonous relative to the autochthonous Monashee Complex (Read and Brown, 1981; Crowley, 1999; Price and Monger, 2003; Crowley et al., 2008; this thesis chapter 2), this trend implies that rocks A and B were emplaced northeastward above the autochthonous Monashee basement, whereas rocks D and E were never emplaced above the basement. Along a NE-SW trend, rocks A and B are thus located at the front of the LSa whereas rocks D and E are located at the rear. P-t paths from the underlying Monashee Complex are strikingly different with burial path to >35 km depth between ~75 and 55 Ma before being rapidly exhumed to less than ~14 km depth between ~59 and 50 Ma (Fig. 3.10; chapter 4). Two crucial conclusions emerge from these paths. First, exhumation of the LSa to upper crustal conditions (<14 km) preceded, or was synchronuous, with burial of the underlying Monashee Complex. Second, at ~60 Ma, rocks that are now juxtaposed (stations C and F on Fig. 3.10) were separated by a vertical distance of >25 km (>700 MPa at a density of 2850 kg/m³). Hence rocks of the LSa and Monashee Complex were not juxtaposed until the Eocene.

3.7 Discussion

3.7.1 Testing extensional models of exhumation

Exhumation of high-grade rocks exposed in the southern Omineca Belt has generally

been attributed to gravitational collapse of a crustal welt previously thickened by protracted convergence (Coney and Harms, 1984; Parrish et al., 1988; Bardoux and Mareschal, 1994; Vanderhaeghe and Teyssier, 1997; Vanderhaeghe et al., 1999; Teyssier and Whitney, 2002; Teyssier et al., 2005). In these models, exhumation is attributed to large displacement along crustal-scale brittle-ductile normal faults (Fig. 3.1a), like the OVfs and Columbia River Fault; to dissection of the upper crust by brittle normal faults (Fig. 3.1 a & b); and to ductile thinning (Fig. 3.1a; Bardoux and Mareschal, 1994) or channel flow and diapirism (Fig. 3.1b; Teyssier et al., 2005) in the middle crust.

However, these models find little support in the data presented in this contribution. First, the USA in the hangingwall of the Columbia River Fault is not dissected by listric normal faults rooting at a master detachment (Lane, 1984; Crowley and Brown, 1994; Colpron et al., 1996 and references therein). One area where the USA should be dissected by normal faults is in the southern part of the study area (Fig. 3.2) along the strike of the cross-section E-F (Fig. 3.3c) where the USA-LSa boundary (i.e. the OVfs) is subhorizontal. However, Glombick et al. (2006a) documented that the few observed normal faults commonly cut through the OVfs and do not root into it. Secondly, whereas the LSa is flanked by two major brittle-ductile normal faults verging outward along cross-section E-F, the northeastern end of LSa along cross-section A-B is bounded by the Purcell thrust that was reactivated into a normal fault with normal displacement estimate of less than 5 km (Boggs, 2004). Third, metamorphic core complexes in western United-States commonly present a clear pattern of mica cooling ages becoming younger toward the master detachment (e.g. Sullivan and Snoke, 2007). Rocks of the Selkirk and northern Monashee mountains do record slightly younger mica cooling ages at 57-53 Ma (Ghent

and Villeneuve, 2006; Fig. 3.3) than rocks cropping out west of the Monashee Complex, but along cross-sections C-D and E-F (Fig. 3.3b, c), such a pattern is absent, with $^{39}\text{Ar}/^{40}\text{Ar}$ mica clustering at ~50-48 Ma. Last and foremost, extensional models are irreconcilable with the synchronous exhumation of LSa and burial of the underlying rocks of the Monashee Complex in Late-Cretaceous/Paleocene indicated by the P-t paths of Fig. 3.10. These P-t paths rather imply a synconvergent model of exhumation.

3.7.2 *Testing synconvergent exhumation models*

3.7.2.1 Ductile thrusting

Three synconvergent models have been proposed for the exhumation of high-grade rocks of the southern Canadian Cordillera. The first model is ductile thrusting above a ramp (Fig. 3.1c) that has been proposed for the Valhalla Complex located just south of the map of Fig. 3.2 (Spear, 2004; Carr and Simony, 2006). Decompression during reverse-sense shearing, as documented herein, is possible by ductile thrusting above a ramp provided synchronic denudation at the surface. Exhumation would then be controlled by focussed erosion and/or extensional faulting in areas of excess topography above the ramp. On a map view this should result in a gradient of metamorphic grade and age of deformation, not in a sharp discontinuity as observed to the west of the LSa. Furthermore, exhumation by ductile thrusting above a ramp is a single-stage process that is not compatible with the protracted ~30 my of exhumation indicated by the P-t paths presented on Fig. 3.10. Another model is therefore required.

3.7.2.2 Progressive underthrusting

The second model is the progressive underplating model of Brown (2004) illustrated on Fig. 3.1d. Parrish (1995) initially proposed a similar model based on a pattern of

younger temperature-time paths with structural depth. The compilation of data presented herein clearly demonstrates the existence of this pattern between the USa, the LSa and the Monashee Complex. However, *within* the LSa, the pattern of younger exhumation documented on Fig. 3.3a and 3.10 is more lateral than vertical. Furthermore, the sharp lithologic, metamorphic and geochronologic breaks that characterize the western boundary between the USa and LSa (OVfs) are not compatible with the gradient expected for exhumation controlled by denudation. Kohn (2008) summarized features that should distinguish channel flow from progressive underplating in critical taper. Two distinctive characteristics of the latter model are a short duration at high temperature (~5 my) and a retrograde path involving isobaric cooling. The protracted zircon and monazite growth in leucosome pods, schists and in deformed intrusions recorded throughout the LSa (Fig. 3.3; 3.10), imply a protracted residence time at high temperature in the middle crust that is incompatible with the short duration at high temperature predicted by the progressive underplating model. Finally, the nearly isothermal decompression path documented for area C (Fig. 3.6) concludes the arguments against the progressive underplating model.

3.7.2.3 Ductile extrusion

The third synconvergent exhumation model is the ductile extrusion model initially proposed for exhumation of the LSa by Scammell (1993) under the name “dynamic spreading”. A similar model was proposed for the exhumation of the LSa together with rocks of the Monashee Complex by Johnston et al. (2000). Channel flow has recently emerged in tectonic models of the southeastern Canadian Cordillera (Williams and Jiang, 2005; Brown and Gibson, 2006; Glombick et al., 2006a; Kuiper et al., 2006), and two studies suggested that channel flow (and ductile extrusion) could have been related to the

exhumation of LSa rocks (Glombick et al., 2006a; Brown and Gibson, 2006). Godin et al. (2006b), Jones et al. (2006) and Kohn (2008) discussed distinctive testable criteria of ductile extrusion, including: 1) a panel of migmatite surrounded by lower-grade rocks that record an older age of deformation, metamorphism and cooling; 2) structures that cannot be traced from the lower-grade rocks into the channel; 3) coeval top-to-the-hinterland and top-to-the-foreland shear zones at the top and base of the panel, respectively; 4) shearing along the two bounding shear zones should be associated with a P-T-t path having a component of decompression; 5) protracted residence time at high temperature; 6) similar timing of deformation and peak metamorphism throughout the exhumed panel; 7) if the front and rear part of the panel are exposed, cooling ages should be older at the front than at the rear; 8) retrograde P-T path should record nearly isothermal decompression; 9) decrease in peak pressure and peak temperature toward higher structural levels of the panel.

The extensive structural, metamorphic and geochronologic database compiled on Fig. 3.2 and 3.3 clearly demonstrate that the USA was deformed, metamorphosed and exhumed above the ductile-brittle transition in the Jurassic to Early Cretaceous, whereas the LSa records protracted high temperature deformation in the Cretaceous. Criteria 1 and 2 are thus met. The P-T-t-d path constructed on Fig. 3.6 confirms previous studies conducted at the latitude of the Trans-Canada Highway (Parrish, 1995; Gilley, 1999; Kuiper, 2003) and at the southern end of the Monashee Complex (Carr, 1992) and demonstrate that the last increments of top-to-the-east shear strain along the Monashee décollement at the base of LSa took place in the Paleocene along the entire length of the Monashee Complex. This new dataset also demonstrates that this shearing was associated

with nearly isothermal decompression. As discussed above, geochronologic and structural data combined with metamorphic textures suggests that top-to-the-west shearing along the OVfs took place in the Paleocene along a decompression and cooling path. A more precise P-T-t-d path from the Okanagan Valley fault system zone is required, but available data suggest that criteria 3, 4 and 8 are also met. The P-t paths presented on Fig. 3.10 point to an overlap of tens of my for the LSa residence time in the middle crust in the Late-Cretaceous. Criteria 5 and 6 would thus be met. Finally, the pattern of younger exhumation from NE to SW along an azimuth perpendicular to the Monashee Complex illustrated on Fig. 3.10 is consistent with criterion 7. The relatively low level of precision for P-T conditions in the footwall of the OVfs precludes testing criterion 9.

3.7.2.4 Summary

Consequently, the extensive structural, metamorphic and geochronologic dataset of the LSa do not meet the distinctive criteria of any other model than the synconvergent ductile extrusion model, for which 7 out of 8 distinctive criteria are met. Evidence for ductile extrusion is strongest between ~65 and 57 Ma, for areas B, C and D (Fig. 3.2, 3.3) because data strongly suggest coeval motion along the two oppositely-verging shear zones bounding the LSa. Because of the lithologic, structural, metamorphic and geochronologic continuity from areas D to A, and then SE toward the Selkirk Mountains (Fig. 3.2), it is likely that ductile extrusion started as early as ~90 Ma, the estimate on the timing at which area A started to be exhumed. However, this hypothesis would require further data before being considered. The best test of this model would be to demonstrate that top-to-the-west shearing along the Selkirk detachment (in the Selkirk Mountains see

Fig. 3.2) was synchronous with top-to-the-northeast shearing in its footwall, and that shearing along the two shear zones occurred along a decompression path.

3.7.3 *Final exhumation by Eocene extension*

Another issue concerns the period between ~60 and 54 Ma when the Ladybird leucogranite suite was emplaced in the LSa west of the Monashee Complex (Carr, 1992). Dykes of the ~60-54 Ma Ladybird suite commonly crosscut part of the transposition foliation but are themselves variously deformed, from completely transposed and sheared to undeformed (Carr, 1992). Teyssier et al. (2005) and Carr (1992) both documented a vergence reversal for low-temperature unannealed shear structures in Ladybird leucogranite between Vernon and Nakusp (Fig. 3.3). This low-temperature vergence reversal is based on solid and convincing data, but the high-temperature vergence reversal inferred by Teyssier et al. (2005) is equivocal because no sense of shear was determined east of the proposed kinematic hinge, and because the confidence level for the determination of the sense of shear was judged to be good to excellent for only 5 out of the 14 specimens examined (Fig. 11 in Teyssier et al., 2005). Ductile high-temperature deformation of the older dykes documented by Carr (1992) was most likely due to the ductile extrusion episode documented herein. However, it is reasonable to assume that the low-temperature deformation of the dykes was linked to normal shearing along the outward-dipping OVfs and Columbia River Fault.

This brings us to the issue of whether extension and normal faulting played any role in the exhumation of the LSa. Four features strongly suggest that extension and normal shearing contributed to the final exhumation of the LSa after it had been exhumed to less than ~14 km depth by ductile extrusion. First, normal shearing along the OVfs and

Columbia River Fault outlasted synconvergent deformation of underlying rocks of the LSa and Monashee Complex, because the timing for the last increments of ductile normal shear strain along both faults is well constrained between ~50 and 47 Ma (Heaman et al., 1999; Mulch et al., 2006). Secondly, the low-temperature vergence reversal documented by Carr (1992) and Teyssier et al. (2005) is compatible with normal shearing along two outward-dipping shear zones. Third, $^{40}\text{Ar}/^{39}\text{Ar}$ mica cooling ages cluster at 53-47 Ma in high-grade rocks exposed in the southern Omineca Belt. This age range coincides with shearing along the two normal faults and is consistent with quenching induced by rapid extensional denudation. Finally, Johnson and Brown (1996) demonstrated that 45 km of net extension, accommodated by the OVfs and the Columbia River Fault, links together stratigraphic units in the hanging wall of both faults that are similar “...in terms of stratigraphy, metamorphic grade, structural style, and their unique association with widespread Devonian-Mississippian orthogneisses.” It is noteworthy that the burial depth of LSa prior to extension was estimated at 8-14 km, which is entirely consistent with the P-t paths presented on Fig. 3.10.

3.7.4 *Driving force for ductile extrusion and hypothetical channel flow*

It is interesting to examine the possible driving force(s) of the ductile extrusion episode documented herein by looking at the tectonic setting of the southern Canadian Cordillera. Thermomechanical models of channel flow (e.g. Beaumont et al., 2001, 2004, 2006) and field-based studies in the Himalayas (e.g. Burbank, 2002 and references therein) identified erosion as the main factor for the creation of a pressure gradient that drove ductile extrusion. In the southern Canadian Cordillera, erosion was probably not the driving force of exhumation. First, the tip of the channel exhumed above the brittle-

ductile transition (~14 km) at ~75 Ma, would likely have been eroded away by the end of extrusion at ~59 Ma if the driving force was erosion. Second, palinspastic restoration of the Foreland Belt and thermochronologic data from the Belt-Purcell Supergroup indicated minimal denudation during the main episode of contraction between ~74 and 59 Ma (Price and Sears, 2000; Sears, 2001). Third, no metamorphic clasts of the LSa have ever been found in the Foreland Basin (Ross et al., 2005 and references therein).

Alternatively, expulsion of ductile lower crust driven by tectonic forcing of an underthrusting indenter has been reproduced in thermomechanical models of large, hot orogens (e.g. Beaumont et al., 2001 & 2006; Jamieson et al., 2007). Results presented above indicate that ductile extrusion took place between ~65 and ~57 Ma, and could have started as early as ~90 Ma. If we zoom out to an orogenic scale, the ~74-59 Ma period coincides with (cf. section 1.1, p. 1): 1) a major depositional change in both basins (Sustut and Foreland) that were flanking the Omineca Belt at this time with voluminous high-energy sedimentation starting at ~72 Ma (Evenchick et al., 2007); 2) a northward migration, and narrowing of the locus of tectonic loading, as indicated by reciprocal stratigraphy in the Western Interior foreland basin (Catuneanu et al., 2000); 3) burial of the craton and overlying shelf sediment now forming the Monashee Complex (Fig. 3.10); 4) the fastest convergence rate between the northern Farallon Plate and North America recorded from past plate motion reconstruction (Stock and Molnar, 1988); and foremost 5) the main convergence phase in the Foreland Belt with ~200 km of shortening recorded between ~74 and 59 Ma at this latitude (Price and Sears, 2000; Sears, 2001). Exhumation of the LSa between ~75 and 57 Ma thus occurred during a major compressive episode. In this context it is likely that underthrusting of a stiff orthogneissic

basement was a driving force of ductile extrusion. If the LSa was not extruded at the surface, as would have been the case for erosion-driven ductile extrusion, the frontal part of the channel should have been shortened to accommodate the flow at the rear. The frontal part of the LSa, located northeast of area A and northeast of the Selkirk detachment (Fig. 3.2) is characterized by abundant map-scale, upright to steeply-inclined, northeast-verging folds referred to as F_3 (Ghent et al., 1980; Simony et al., 1980; Raeside and Simony, 1983). Although such folds have been interpreted as immature folds formed during east-directed shearing in the Monashee Complex (Williams and Jiang, 2005; Kruse, 2007), F_3 folds in the Selkirk and Monashee mountains clearly postdate the metamorphic climax because the foliation and lineation formed in part by the alignment of high temperature minerals (sillimanite, kyanite, oblate garnet), are folded by F_3 , and there is no neocrystallization of metamorphic minerals associated with these folds (Ghent et al., 1980; Simony et al., 1980; Raeside and Simony, 1983). Total shortening accommodated by F_3 folds was estimated at 30-60% (Ghent et al., 1980). Assuming 50% of NE-SW shortening would bring area A approximately to the latitude of the Monashee Complex axis (near area B on Fig. 3.2 & 3.3a); or at the inferred location of the autochthonous basement ramp (Price and Sears, 2000; Price and Monger, 2003; Crowley et al., 2001 & 2008; chapter 2 of this thesis). These data thus suggest that underthrusting of a stiff basement ramp expelled a channel of partially molten rocks (LSa) toward higher structural levels. This process would have continued for a protracted period of time in the Cretaceous-Paleocene by the combined effect of horizontal flow at the rear of the channel, within the orogenic core (channel flow); and buckling at the front, above the basement flat and past the ramp. Although appealing, this type of ductile extrusion

induced by tectonic forcing was critically questioned by Williams et al. (2006) who argued that channel flow driven by an internal pressure gradient is much more likely. I will argue in chapter 4 that underthrusting of a felsic basement having a lower density than the overlying wedge produces a significant pressure gradient that can also drive ductile extrusion.

3.8 Summaries and conclusions

1. The LSa is a 10-15 km thick panel consisting mainly of migmatitic rocks at the sillimanite-K-feldspar grade. The LSa is characterized by isoclinal folds that are completely transposed into parallelism with a well-developed, unimodal E-W stretching lineation (Fig. 3.2), and a transposition foliation.
2. The LSa is bounded by a top-to-the-west shear zone at its roof (the brittle-ductile Okanagan Valley fault system) and a top-to-the-east shear zone at its base (the Monashee décollement). The roof top-to-the-west shear zone separates migmatitic rocks of the LSa that were deformed at high temperature for a protracted period of time between ~120 to ~75-57 Ma, from overlying lower amphibolite/greenschist rocks of the USa that were deformed, metamorphosed and cooled above the brittle field before ~120 Ma. The basal top-to-the-east shear zone separates the sillimanite-bearing migmatites of the LSa from underlying, kyanite-bearing migmatites of the Monashee Complex that were deformed and metamorphosed between ~60 and 48 Ma.
3. This contribution presents a P-T-t-d path from the top-to the-east basal shear zone. The path is reconstructed by combining laser ablation U-Pb dating of zircon and titanite, phase equilibria modeling, conventional thermobarometry,

and titanite diffusion modeling. It is concluded that top-to-the-east shearing took place along a steep decompression path from thermal peak conditions of $\sim 760^{\circ}\text{C}/800\text{ MPa}$ to $\sim 700^{\circ}\text{C}/300\text{ MPa}$ between ~ 83 and $\sim 57\text{ Ma}$, with the last increments of shear strain taking place between ~ 62 and $\sim 57\text{ Ma}$ (Fig. 3.6). This P-T-t-d path is consistent with timing constraints as well as with microstructural and metamorphic textures documented along the entire length of the Monashee décollement.

4. The timing of the roof shear zone is not as well constrained, but compiled geochronologic, structural and metamorphic data from the zone strongly suggest that part of the top-to-the-west shearing was synchronous with top-to-the-east shearing at the Monashee décollement and was also occurring along a decompression and cooling path.
5. Five depth-time paths from the LSa and two from the underlying Monashee Complex indicate progressive exhumation of the LSa to $<14\text{ km}$ depth along a trend perpendicular to the axis of the Monashee Complex; from $\sim 75\text{ Ma}$ in the Selkirk Mountains to $\sim 54\text{ Ma}$ west of the Monashee Complex to $\sim 50\text{ Ma}$ near the town of Vernon (Fig. 3.2, 3.3, 3.10). The $\sim 75\text{-}54\text{ Ma}$ exhumation of the LSa was coeval with burial of the Monashee Complex to $>35\text{ km}$ depth (Fig. 3.10), as well as with most of the 200 km of shortening recorded in the Foreland Belt at this latitude (Price and Sears, 2000; Sears, 2001).
6. This pattern of coeval exhumation of the LSa and burial of the Monashee Complex in the middle to lower crust along with shortening in the upper crust repudiate the widely publicized models of exhumation by Eocene

gravitational collapse and large-magnitude extension (Coney and Harms, 1984; Parrish et al., 1988; Bardoux and Mareschal, 1994; Vanderhaeghe and Teyssier, 1997; Vanderhaeghe et al., 1999; Norlander et al., 2002; Teyssier and Whitney, 2002; Teyssier et al., 2005). The data are rather entirely consistent with synconvergent ductile extrusion above the underthrusting Paleoproterozoic basement of the Monashee Complex (Crowley et al., 2001 & 2008; chapter 2 of this thesis).

Table 3.1. Mineral compositions

Min	cpx	hbl	pl	grt. 2	grt. 2	grt. 1	grt. 1	pl
	avrg.(6)	avrg. (5)	avrg.(6)	core (5)	rim (6)	core (2)	rim (3)	avrg. (5)
SiO₂	53.03	48.66	65.89	37.92	37.63	37.41	37.13	64.18
TiO₂	0.12	0.47	0.00	0.01	0.01	0.01	0.01	0.00
Al₂O₃	1.51	5.48	21.67	21.35	21.14	21.16	21.02	23.51
Fe₂O₃	2.51	2.15	0.19	0.11	0.35	0.52	0.14	0.00
FeO	8.05	12.60	0.00	30.67	31.76	31.21	30.00	0.00
MnO	0.59	0.44	0.00	2.05	3.43	2.74	7.20	0.22
MgO	11.38	13.07	0.00	5.94	4.56	5.35	3.05	0.00
CaO	21.65	10.96	2.31	1.53	1.43	1.47	1.43	4.31
Na₂O	1.47	1.31	10.28					9.21
K₂O	0.01	0.64	0.26					0.21
Totals	100.07	95.56	100.60	99.58	100.31	99.86	99.99	101.64
Oxygens	6.00	23.00	8.00	12.00	12.00	12.00	12.00	8.00
Si	1.98	7.27	2.88	3.00	3.00	2.98	2.99	0.00
Ti	0.00	0.05	0.00	0.00	0.00	0.00	0.00	2.03
Al	0.07	0.97	1.12	1.99	1.98	1.99	2.00	0.49
Fe₃	0.07	0.24	0.01	0.00	0.00	0.00	0.00	0.37
Fe₂	0.25	1.57	0.00	2.04	2.14	2.11	2.03	0.00
Mn	0.02	0.06	0.00	0.14	0.23	0.18	0.49	0.01
Mg	0.63	2.91	0.00	0.70	0.54	0.64	0.37	0.00
Ca	0.87	1.75	0.11	0.13	0.12	0.13	0.12	0.20
Na	0.11	0.38	0.87					0.78
K	0.00	0.12	0.02					0.01
Sum	4.00	15.41	5.00	8.00	8.00	8.00	8.00	5.00
Na/Na+Ca	0.11		0.89					0.79
Fe/Fe+Mg	0.28	0.35		0.74	0.80	0.77	0.85	
Xpyr				0.23	0.18	0.21	0.12	
Xgrs				0.04	0.04	0.04	0.04	
Xsps				0.05	0.08	0.06	0.16	

Table 3.2. U-Pb laser ablation MC-ICP-MS spot analyses of zircon grains from leucosome sample 05-FG-462d.

Anal.#	$^{207}\text{Pb}/^{235}\text{U}$	$\pm 2\sigma$	$^{206}\text{Pb}/^{238}\text{U}$	$\pm 2\sigma$	rho	$^{206}\text{Pb}/^{238}\text{U}$ age (Ma)	$\pm 2\sigma$ (Ma)
<i>rims</i>							
02-2	0.1889	0.0060	0.0247	0.0017	0.45	157	11
03-1	0.1112	0.0035	0.0167	0.0005	0.86	107	3
05-2	0.1262	0.0040	0.0187	0.0007	0.96	119	4
L02-2	0.0814	0.0027	0.0130	0.0004	0.90	83	3
L02-3	0.0615	0.0022	0.0101	0.0003	0.85	65	2
L02-4	0.0806	0.0027	0.0132	0.0004	0.87	84	3
L02-5	0.0668	0.0024	0.0131	0.0004	0.71	84	3
S05-1	0.0853	0.0030	0.0164	0.0010	0.99	105	6
S-12	0.0927	0.0041	0.0136	0.0016	0.36	87	10
S-13	0.0599	0.0023	0.0093	0.0003	0.76	60	2
S-13-2	0.0603	0.0025	0.0095	0.0003	0.64	61	2
S-15	0.0634	0.0032	0.0097	0.0003	0.30	62	2
S15-2	0.0405	0.0018	0.0100	0.0005	0.79	64	3
S-16	0.0608	0.0029	0.0099	0.0003	0.36	63	2
S-19-1	0.0813	0.0058	0.0125	0.0005	0.78	80	4
<i>detrital</i>							
02-1	4.5261	0.1359	0.2858	0.0107	0.96	1620	61
04-1	11.6970	0.3510	0.4400	0.0146	0.94	2351	78
08-1	0.5103	0.0154	0.0650	0.0021	0.92	406	13
09-1	0.6671	0.0201	0.0822	0.0027	0.94	509	17
11-1	0.8458	0.0254	0.0987	0.0033	0.95	607	20
L01-1	2.0724	0.0622	0.1920	0.0064	0.94	1132	38
L01-2	2.0084	0.0603	0.1893	0.0064	0.93	1118	38
S03-1	0.5549	0.0175	0.0718	0.0057	0.40	447	36
S07-1	0.6482	0.0198	0.0845	0.0045	0.58	523	28
S23-1	0.7095	0.0216	0.0805	0.0043	0.56	499	27

Table 3.3 U-Pb laser ablation MC-ICP-MS spot analyses of titanite grains from leucosome sample 05-FG-462D.

	Uncorrected isotopic ratios and errors				Corrected age ^a		
	²⁰⁷ Pb/ ²⁰⁶ Pb	± 2σ	²⁰⁶ Pb/ ²³⁸ U	± 2σ	ρho	²⁰⁶ Pb/ ²³⁸ U ± 2σ	Age (Ma) Ma
<i>cores</i>							
M01_1	0.2522	0.0052	0.0149	0.0005	0.79	60.7	7.1
M02_1	0.2580	0.0052	0.0145	0.0005	0.80	58.3	7.1
M04_2	0.3062	0.0053	0.0165	0.0006	0.86	57.2	9.8
S 01	0.3276	0.0045	0.0186	0.0006	0.91	59.7	11.8
S 01-6	0.3216	0.0051	0.0182	0.0006	0.88	59.7	11.3
S 01-2*	0.3395	0.0043	0.0247	0.0008	0.92	76.2*	16.2
S 05-2	0.3151	0.0040	0.0170	0.0006	0.93	57.3	10.4
M04_1	0.3060	0.0063	0.0153	0.0005	0.80	53.2	9.1
<i>rims</i>							
S 12	0.3115	0.0065	0.0152	0.0006	0.85	51.7	9.2
M01_2	0.2585	0.0073	0.0135	0.0005	0.65	54.3	6.7
M02_2	0.2230	0.0042	0.0121	0.0005	0.86	53.5	5.2
M02_3	0.2368	0.0072	0.0121	0.0004	0.63	51.7	5.5
M02_5	0.3276	0.0056	0.0140	0.0005	0.86	44.9	8.9
M02_6	0.3222	0.0058	0.0145	0.0006	0.91	47.7	9.1
M03_1	0.3223	0.0070	0.0172	0.0006	0.76	56.4	10.8
S 01-3	0.2165	0.0028	0.0129	0.0005	0.95	57.7	5.3
S 01-4	0.2343	0.0032	0.0134	0.0005	0.92	57.3	5.9
S 01-5	0.2673	0.0030	0.0138	0.0005	0.98	53.9	7.1
S 05-1	0.3064	0.0051	0.0140	0.0006	0.96	48.6	8.4
S 05-3	0.3257	0.0056	0.0162	0.0005	0.86	52.6	10.2

* Discarded from calculations. ^a ²³⁸U/²⁰⁶Pb age corrected by the ²⁰⁷Pb method (Stern, 1997) by taking the y-intercept on the Terra-Wasserburg plot (Fig. 3.8) as the value of common Pb.

Table 3.4. Comparison Monashee décollement (MD) vs. OVfs

	MD	OVfs
shear-sense	top-east	top-west
metamorphic grade	sil-Kfs	sil-Kfs
P-T path	↓T, ↓P	↓T, ?↓P?
Timing HT	~118-59 Ma	~120-57 Ma
Timing last increments HT shear strain	~62-57 Ma	>57 Ma
Timing end of ductile defn.	~57 Ma	down to ~49 Ma
Timing cooling <500-600°C	~59-57Ma	~58-54 Ma

Table 3.5. Source of data for depth-time paths of Fig. 3.10

Area	P-T peak	P-T retrograde	Timing peak	Timing retrograde
A	~800 MPa at ~700°C ^{1,2}	~500 MPa at ~600°C ¹	~170, 110-90, 68 Ma Mnz schists ^{1, 2, 3, 4} ~122 Ma pretransposition dyke ⁴ ~100 U-Pb syntransposition. dyke ^{1, 2} ~104 Ma posttransposition dyke ³ ~63 Ma posttransposition dyke ^{3, 4}	~91 Ma Tnt ¹ ~75-66 Ma Hbl ¹ ~52 Ma Ms ¹
B	~800 MPa at ~760°C ¹	~400 MPa at ~600°C ¹	~100 Ma Mnz schists ¹ ~135 Ma pre to syntransposition dyke ¹ ~71 Ma posttransposition dyke ¹	~74 Ma Tnt ¹ ~70-64 Ma Hbl ¹ ~53 Ma Rt ¹ ~50 Ma Ap ¹
C	~800 MPa at 760°C ⁵	~200 MPa at ~700°C ⁵	~120-155; 87-83; ~62 Ma Zrc in leucosome ^{5, 6} ~132; 120-110; 86-81; ~65 Ma Mnz leucosome ⁶ ~97 syntransposition dyke ¹ ~57 Ma posttransposition dyke ¹	~54 Ma Tnt ^{1, 5}
D	700-1000 MPa at 700-850°C ^{7, 8}	n/a	~70 Ma Zrc in leucosome ⁹ ~61-54 Ma Mnz in leucosome ⁹	~58-54 Ma Hbl
E	700-1000 MPa at 700-850°C ^{7, 10}	~450 MPa at ~600°C ¹⁰	M1: 155-150 Ma Zrc in leucosome ¹¹ M2: ~90 Ma Zrc in leucosome ¹¹ M3: 79-51 Ma Zrc rims + Tnt ¹¹	~79-51 Ma Tnt ¹¹ ~52-49 Ma Hbl ¹⁰ ~49 Ma Bt ¹⁰
F	1000-12000 MPa at 800-850°C ¹²	~300 MPa at ~650°C ¹²	~59 Ma Zrc in ky-leucosome ¹³ ~75-59 Ma prograde Mnz in schist ¹⁴	~59-56 Ma Mnz in schist ¹⁴ ~55 Ma Zrc post-transposition pegmatite ~55 Ma Hbl ¹⁵ ~50 Ma Bt ¹⁵
G	~800-12000 MPa 675-775°C ¹²	~400 MPa at ~560°C	~60-53 Mnz in schist ^{13, 14}	n/a

Ref: 1: Scammell, 1993; 2: Sevigny et al., 1990; 3 : Gibson et al., 2005 & 2008; 4 : Crowley et al., 2003; 5 : This chapter; 6 : Parrish, 1995; 7: Glombick, 2005; 8: Teyssier et al., 2005; 9: Johnson, 1994; 10: Bardoux, 1993; 11: Glombick et al., 2006b; 12: This study chapter 3; 13: Crowley and Parrish, 1999; 14: Foster et al., 2004; 15: Sanborn, 1996.

Figure 3.1. Exhumation models investigated in this study. See text for explanation. In d) MD3 represents the active shear zone, whereas MD1 & 2 are shear zones that are being passively exhumed by accretion of material from below and denudation at the surface.

a) Large-magnitude extension (as in Parrish et al., 1988; Bardoux and Mareschal, 1994)

b) Flow toward a rolling-hinge (as in Teyssier et al., 2005)

c) Ductile thrusting above an indenter (as in Spear, 2004; Carr and Simony, 2006)

d) Progressive underthrusting in critical taper (as in Brown, 2004)

e) Channel flow and ductile extrusion (as in Brown and Gibson, 2006; Glombick et al., 2006a)

Legend

- Crustal contraction
- Crustal extension
- Sense of shear (top direction)
- Erosion
- Isotherms

- Low-grade rocks
- Medium-grade rock
- Migmatites cooled below brittle-ductile transition
- Flowing migmatites

Figure 3.1

LEGEND (Map & cross-sections)

- Devonian, Jurassic and Cretaceous Plutons
- Paleocene Ladybird leucogranite

Tectonometamorphic units

- Upper Selkirk allochthon (Usa)
- Lower Selkirk allochthon (Lsa)

Monashee Complex

- Cover sequence: metasediments infolded with basement slices
- Basement: polymetamorphosed para-orthogneiss

Geochronologic data**Type of dates**

- Deformed intrusions (U-Pb zrc, mnz)
- Thermal peak schist (U-Pb mnz)
- Crosscutting dykes (U-Pb zrc, mnz)
- Hbl cooling ages or
b:biotite, m: muscovite, t: titanite, ap: apatite
Ar/ Ar for hbl, b, m; U-Pb for t;
fission track for ap

Colour chart

- < 59-47 Ma
- 120-60 Ma
- > 120 Ma
- Locations for depth-time paths

Structural symbols

- Normal brittle-ductile fault
- Thrust ductile shear zone

Figure 3.2. Revised tectonometamorphic map of the southeastern Canadian Cordillera surrounding the Monashee Complex.

Inset (above) shows the five morphogeological belts of the Canadian Cordillera. Symbols on the map correspond to the collection site of specimens analyzed for geochronology, see legend for signification of symbols, Fig. 3.3 for value of ages and Table 1.1 (p. 40) for references. A-B, C-D and E-F are the traces of the cross-sections shown on Fig. 3.3. Lower hemisphere equal-area projections for stretching lineations (L_s) and mature folds (F_T). Structural data: a-d from Scammell (1993); e-f from Journeay (1986); g-h from Johnson (2006); and i) from Glombick et al. (2006a). Circles with capital letters A-E are areas discussed in the text for which depth-time paths were reconstructed and depicted on Fig. 3.10. It is of note that the Purcell thrust was reactivated as a normal fault (Boggs, 2004). See Table 1.1 p.38 for abbreviations.

Figure 3.2

Figure 3.3. Cross-sections on which all available ages have been projected. See text for explanation. Legend is the same as for Fig. 3.2 except for: 1) the black fill that represents a mafic lower crust inferred from seismic refraction profiles (Burianyk and Kanesevich, 1995); and 2) for the basement of the Monashee Complex (X pattern) that only includes rocks inferred to have escaped Cordilleran tectonism, whereas on the map of Fig. 3.2 basement refers to lithologic units irrespective of their timing of deformation.

Figure 3.3

Figure 3.4: Field photographs from the Monashee décollement at the base of the LSa on the north flank of the Monashee Complex (location C on Fig. 3.2). All photos show surfaces perpendicular to the foliation and parallel to the stretching lineation (XZ plane of the strain ellipsoid). a) Mylonitic straight gneiss with boudinaged amphibolite layers. Three types of boudins can be considered as reliable shear sense indicators as per Goscombe and Passchier (2003): 1) foliation-oblique shear band boudins of a leucogranite dyke; 2) foliation-parallel shear band boudins of an amphibolite; 3) foliation-parallel domino boudins of amphibolite. Shear sense is invariably top-to-the-east. Compass pointing north for scale. b) Foliation-oblique shear-band boudin in leucogranite indicating a top-to-the-east sense of shear. c) The centre of the photograph is an amphibolite boudin that contains clinopyroxene-bearing leucosome. Leucosome is preferentially located at the neck where amphibolite is broken into angular fragments. These relationships are interpreted as evidence for melt-present deformation and amphibolite melting via Rx2: $\text{Hbl} + \text{Pl} + \text{Qtz} \pm \text{H}_2\text{O} = \text{Cpx} + \text{Liq} + \text{Tnt}$. Stair-step structures (Hanmer and Passchier, 1991) in this mylonitic gneiss above the boudin (ellipse) indicate top-to-the-east sense of shear. Pen for scale. d) Sample FG-462d used for conventional thermobarometry and LA-MC-ICP-MS U-Pb geochronology of zircon and titanite. It consists of a sigma-type clinopyroxene aggregate. Notice preferential location of leucosome in the upper-right pressure shadow where hornblende crystallized, most likely by the inverse of Rx 2 upon cooling. See text for further explanation. Eraser tip for scale. Mineral abbreviation from Kretz (1983).

Figure 3.4:

Figure 3.5: Photomicrographs for specimens Fg462a & d. a) One of the rare garnet of specimen FG-462a, a semi-pelitic migmatitic schist from zone A. The long axis of this oblate garnet is inclined relative to the foliation (which is parallel to the picture) and is oriented from bottom-left to top-right of the image. Notice the preferential resorption of garnet in the bottom-left to top-right quadrants, with sillimanite-biotite-plagioclase concentrated in these pressure shadows. These microtextures are interpreted as structurally-controlled garnet resorption by ReNtr 1 during top-to-the-east shearing (dextral on the picture). Thin-section cut in the XZ plane of the strain ellipsoid. b) Contact between host and leucosome of specimen FG462d. Note the coarse grain size and euhedral shape of quartz and feldspar grains forming the leucosome. See text for detail. c) Core of the clinopyroxene aggregate of specimen FG-462d. Note the euhedral and faceted shape of clinopyroxene and titanite, as well as the intergrowth texture between these two minerals. d) Margin of the clinopyroxene aggregate of sample FG-462d where clinopyroxene is directly pseudomorphed into hornblende. Note the perfect crystalline shape of titanite and crystal faces of feldspar (Fld) and quartz (Qtz) crystals (circles). Observations in b-d are interpreted as evidence of partial melting. Scale bar is 1 mm long on all photomicrographs.

Figure 3.5

Figure 3.6: P-T-t path for the Monashee décollement north of the Monashee Complex (location C on Fig. 3.2). Background shows lines of disappearance/appearance of key metamorphic minerals of specimen Fg462a, a migmatitic rutile-garnet-ilmenite-sillimanite-K-feldspar-biotite schist. The lines of disappearance/appearance were calculated by phase equilibria modeling in the TiMnCaNaKFMASH system with the program Perple_X'06. Bulk composition of specimen Fg462a in weight% is TiO₂: 1.06; MnO: 0.09; Ca: 1.18; Na₂O: 2.80; K₂O: 4.49; FeO: 7.56; MgO: 3.72; Al₂O₃: 18.70; SiO₂: saturated; H₂O: 2.00. Also shown are P-T conditions estimated with conventional thermobarometry analyses conducted on specimen FG-462d, a clinopyroxene-titanite leucosome interpreted to result from amphibolite melting. Curve (1) is the water-saturated melting curve of metabasalt S6 in the experiments of Springer and Seck(1997). Curve (2) is the wet-granite solidus of Ebadi and Johannes (1991). Curve (3) is the result of Thermocalc's average P mode for specimen Fg462d. Curve (4) is the result of the Al-in barometry (Anderson and Smith, 1995) for specimen Fg462d with 200 MPa added (see Bachmann and Dungan, 2002 and text for justification). Curve (5) is the SCAN barometer of McCarthy and Patiño-Douce (1998). Curve (6) is the quartz-saturated hornblende-plagioclase thermometer of Holland and Blundy (1994). Stippled boxes are the uncertainties associated with thermobarometry. The ellipse shows the preferred P-T of equilibration for specimen Fg462d. Explanation of the P-T-t path in the text.

Figure 3.6

Figure 3.7: Results of laser ablation MC-ICP-MS U-Pb geochronology for zircon grains of specimen FG462d, a clinopyroxene-titanite leucosome. a-c) Cathodoluminescence images of zircon crystals with location of spot (40 μ m) analyses. Notice the sharp crystal faces on the three grains and the faint zoning in (a) and (c). In (a), the boundary between the ~83 Ma core and the ~65 Ma rim is a thin black line that delimits a euhedral core with sharp faces. These textures are interpreted as growth of the two generations of zircon in the presence of a melt phase. d) $^{206}\text{Pb}/^{238}\text{U}$ vs. $^{207}\text{Pb}/^{235}\text{U}$ concordia plot showing the distribution of metamorphic dates. Notice the protracted concordant dates from ~120-60 Ma. The three analyses lying above the concordia curve (gray ellipses) were discarded for the calculations of the weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages of the two clusters. See text for further details.

Figure 3.7

Figure 3.8: Results of laser ablation MC-ICP-MS U-Pb geochronology from titanite grains of specimen FG462d, a clinopyroxene-titanite leucosome. a-d) BSE images of titanite grains and location of spots analyzed for U-Pb geochronology. Note the euhedral shape the grains in (d), a photo taken from the thin-section. e) Terra-Wasserburg $^{207}\text{Pb}/^{206}\text{Pb}$ vs. $^{238}\text{U}/^{206}\text{Pb}$ plot. The upper y-intercept of 0.61 ± 0.11 is inferred to represent the initial common Pb value. f) Weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages. See text for details.

Figure 3.8

Figure 3.8. Results of laser ablation MC-ICP-MS U-Pb spot analyses from titanite grains in leucosome sample FG462d.

Figure 3.9. Diffusion modeling for Pb in titanite grains from (a) specimen RS465 (Scammell, 1993) and (b) specimen Fg462d (this study). Modeling was conducted with the program DIFF-CL of Ganguly and Tirone (1999), using titanite diffusion parameters of Cherniak (1993). Only the most representative results are shown. The first number is the initial temperature in °C and the second is the cooling rate. In order to be viable, a modeled curve has to lie within the age bracket provided by the U-Pb age of titanite with uncertainty shown as solid lines. See text for explanation.

a)

b)

Figure. 3.9.

Figure 3.10: Depth-time paths for the LSa and underlying Monashee Complex (MC). Paths A-E are for different areas of the LSa whereas paths F-G are for the Monashee Complex. Filled circles are well-constrained P-T-t date points. Stippled rectangles are uncertainties on pressures and timing. Source of data presented in Table 3.5 (p. 172). See text for explanation.

CHAPTER 4

FOLD NAPPE FORMATION BY DUCTILE EXTRUSION IN THE NORTHERN MONASHEE COMPLEX OF THE CANADIAN CORDILLERA

ABSTRACT

The purpose of this paper is to examine processes responsible for the formation and exhumation of migmatitic fold nappes in the northern Monashee Complex of the southeastern Canadian Cordillera. Key to the interpretation is the construction of pressure-temperature-time (P-T-t) paths at different structural levels of a ~10 km long transect that crosses two fold nappes. Stratigraphic, structural and metamorphic data from unpublished theses are first summarized and in some instances reinterpreted in light of new field investigations conducted on the west flank of the northern Monashee Complex (Frenchman Cap dome). Then, P-T-t paths are derived for five structural levels by combining results of phase equilibria modeling with previously published geochronologic data. The derived P-T-t paths are then combined with structural constraints to reconstruct the evolution of five rock particles in an isostatically-balanced crust. It is shown that fold nappes of the Monashee Complex formed by sequential extrusion of partially molten rocks above an underthrusting basement between ~59 and 50 Ma.

4.1 Introduction

Map-scale recumbent folds—defined as fold nappes by Twiss and Moore (2006) and other authors (Price and McClay, 1981; Ramsay and Huber, 1987; Merle, 1994)—are characteristic structures of the core zones of orogenic belts. Formation of fold nappes has been at the heart of the vertical-horizontal tectonics debate. Gravity sliding from

topographic highs has been a process commonly proposed for the formation of fold nappes in the external parts of fold belts (cf. Merle, 1994), but these models are obviously not applicable to fold nappes formed in the middle to lower crust. Based on field studies in the Scandinavian Caledonides and centrifuge analogue modeling, Ramberg (1980 & 1981) proposed that fold nappes could form by lower crustal flow induced by gravitational collapse. In Ramberg's model, a fold nappe geometry results from the combined effect of flow and shear stress at the base of the crust. By contrast, building on field studies from the same Scandinavian Caledonides and from the southern Monashee Complex, Williams and Jiang (2005) presented a kinematic model of fold nappe formation in high-grade terrains according to which fold nappes would form by horizontal crustal-scale shearing. Their studies supported previous interpretations of fold nappe formation by Hatcher and Hooper (1992) and Krabbendam et al. (1997). Alternatively, thermomechanical numerical modeling of orogens suggests that fold nappes could form during ductile extrusion of deep-seated rocks. Jamieson et al. (2007) and Stoeckhert and Gerya (2005) discussed the similarities between their model and fold nappes of the western Grenville Province and western Alps, respectively. However, these models remain to be entirely tested with field data.

This chapter explore the processes of fold nappe formation and exhumation of andalusite-sillimanite-kyanite migmatitic schists from the northern Monashee Complex of the southeastern Canadian Cordillera. This is accomplished by integrating previously published data to new Pressure-Temperature-time paths constructed for five metamorphic zones that are spatially linked to the structural geometry of the fold nappes.

4.2 Tectonic setting

See chapter 1.

4.3 Geology of Frenchman Cap dome

4.3.1 *Methods of field investigation*

See chapter 1.

4.3.2 *Lithostratigraphic units*

Lithostratigraphic studies conducted in Frenchman Cap dome led to the construction of a detailed stratigraphic column that allowed for a better understanding of its structural geometry (Weeler, 1965; Journeay, 1986; Scammell and Brown, 1990; Höy, 2001). The dome is cored by Proterozoic basement orthogneiss and subordinate paragneiss and is mantled by a metasedimentary cover sequence composed of siliciclastic and carbonate rocks interlayered with extrusive and intrusive alkaline igneous rocks. The cover sequence and the upper basement level were collectively metamorphosed to upper-amphibolite facies during the Cordilleran orogeny. A simplified lithostratigraphic column of the metasedimentary cover sequence consists, from bottom to top, of: 1) a thick layer of pure quartzite; 2) calc-silicates interlayered with minor pelitic schist; 3) a white calcite marble layer closely associated with a distinctive extrusive carbonatite layer and commonly with a mineralized Pb-Zn layer; 4) aluminous metapelites; 5) quartzite interlayered with amphibolites and hornblende gneisses (McMillan, 1970; Read and Brown, 1981; Höy, 1987, 2001; Journeay, 1986; Scammell and Brown, 1990). Only the first 4 units are exposed within the area covered by the map of Fig. 4.2 because unit 5 crops out only in the northern part of the dome (Scammell, 1986). The white calcite marble and the associated extrusive carbonatite constitute a crucial structural marker that

has been mapped almost continuously around the dome, as illustrated on Fig. 4.2 by the thick white line that encircles the dome. Scammell and Brown (1990) argued that the amagmatic basal part of the cover sequence formed on a subsiding cratonic platform; and that the upper part of the sequence—containing a distinctive extrusive carbonatite layer, numerous alkaline intrusions, and mafic sills – was deposited in a rift setting. The age of deposition of the basal part is constrained between ~1.85 Ga, the youngest detrital zircon in the basal quartzite, and ~724 Ma, the crystallization age of a syenite intrusion (Crowley, 1999). The upper, magmatic part of the sequence was deposited between ~1.2 Ga, the age of the youngest detrital zircon from a quartzite in the upper part (Crowley, 1999), and 541 Ma, the crystallization age of an amphibolite sill (Parrish, 1995). This possible Cambrian age would agree with the Cambrian Pb-Pb model age obtained by Höy and Godwin (1988) for the extrusive carbonatite and suggests that the upper part of the cover sequence was deposited during the second rifting event of the Windermere basin (Colpron et al. 2002). If this interpretation is correct, the absence of deep-water facies sediments that typifies Windermere sediments would imply that the cover sequence was deposited either inboard of the Windermere basin or above a rifted continental block as suggested by Thompson et al. (2006). The former interpretation is supported by palinspastic reconstruction of the Foreland Belt that restores back the edge of the miogeocline to the present location of the Monashee Complex (Price and Sears, 2000; Sears, 2001).

At the latitude of Frenchman Cap dome, there is a sharp lithologic break between the Monashee Complex and overlying Lower Selkirk allochthon (LSa). The LSa comprises migmatitic semi-pelitic schists, hornblende-bearing quartzo-feldspathic gneisses, and

amphibolite layers; quartzite, aluminous schist and marble, which are typical lithologic units of the Monashee Complex, are rare (Read and Brown, 1981; Journeay, 1986; Scammell, 1993; this study). Furthermore, the lithologic break coincides with a sharp gradient in the volume of leucogranite intrusions, which comprise over 30% of outcrop by volume in the upper plate, and only ~10% of the lower plate. This lithologic boundary is associated with a structural, metamorphic and geochronologic break that has been referred to as the Monashee décollement (e.g. Brown et al., 1992).

4.3.3 *Structural geology*

4.3.3.1 Fold nappes

Despite receiving a lot of attention in the last ten years (e.g. Vanderhaeghe and Teyssier, 1997; Vanderhaeghe et al., 1999; Vanderhaeghe and Teyssier, 2001; Teyssier and Whitney, 2002; Teyssier et al., 2005), gneiss dome formation constitutes only a small part of the complicated structural history of the Monashee Complex. The most important map-scale structures are fold nappes that are only a few kilometers thick but have been mapped continuously for tens of kilometers along strike. In Frenchman Cap dome, two overlying fold nappes — here named the Sybley and Mount-Grace — have axial surface traces more than 50 km long that encircle the northern part of the dome (Fig. 4.2). Each fold nappe consists of an anticline-syncline pair that folds the stratigraphic layering. In order to facilitate reading of this manuscript “anticline” will refer to rocks from the upper part of each fold nappe down to the axial surface of the syncline, whereas “syncline” will refer to rocks located below the axial surface of the underlying syncline. As can be seen from Fig. 4.2 and 4.3, the distinctive structural marker composed of white calcite marble-extrusive carbonatite is repeated along the limbs of the two fold nappes. Stratigraphic

layering is not entirely continuous from fold to fold, however, and stratigraphic breaks have been interpreted as thrust faults separating fold nappes by Journeay (1986). The same conclusion was reached by Duncan (1984) for fold nappes of Thor-Odin dome (Fig. 4.1 for location), but Williams and Jiang (2005) reinterpreted these breaks to result from transposition of discontinuities during high shear strain. Axial surfaces of the fold nappes are subparallel to a moderately dipping transposition foliation that wraps around the dome. The only exposed fold nappe closure is that of the Mount-Grace syncline exposed south of Fig. 4.2.

4.3.3.2 Top-to-the-east shear strain

A striking structural feature of the area is the almost perfect parallelism between minor fold axes and a pervasive stretching lineation that is formed by the alignment of elongate minerals (sillimanite, kyanite, hornblende), recrystallized mineral aggregates and rod-shaped leucosome pods. On the southwest-dipping flank of Frenchman Cap dome shown on Fig. 4.2, W-plunging stretching lineations and minor fold axes cluster on lower hemisphere equal area projections (Fig. 4.3). Asymmetric minor folds dominantly verge to the north, although south-verging folds are also present (Journeay, 1986). There is no systematic minor fold vergence reversal across the hinges of fold nappes. Kinematic indicators such as C-S fabric, rotated porphyroclasts, and oblique boudin trains (Hanmer and Passchier, 1991) are common and generally indicate a top-to-the-east sense of shear (Scammell, 1986; this study). However, top-to-the-west kinematic indicators abound within a localized zone at the top of the complex (see section 4.3.3.3, p. 198).

My interpretation of structural data agrees with kinematic analyses conducted in the Monashee Complex during the last 25 years (Van den Driesshe, 1983; Journeay, 1986,

p.114-119; Scammell, 1986, p.55-65; McNicoll and Brown, 1995, p.22-26; Gibson et al., 1999, p.759-760; Crowley et al., 2001; Spark, 2001; Brown, 2004; Williams and Jiang, 2005; Kruse, 2007). These analyses demonstrated that the total strain observed in rocks of the Monashee Complex was acquired through progressive simple shearing with a direction of maximum stretch approximately parallel to the penetrative W-plunging regional lineation. Drag folds developed throughout shearing and were progressively rotated toward the W-plunging direction of maximum stretch (cf. Williams et al., 2006). In such a zone of progressive shearing, different attitudes of folds with respect to the transposition foliation and stretching lineation, commonly labelled F_1 , F_2 etc, are not related to distinct deformation events, but rather reflect the degree of maturity of folds that are being rotated toward the direction of maximal stretch (see Williams and Jiang, 2005). According to the equations presented in Williams and Jiang (2005), the subparallelism between stretching lineations and fold axes observed in Frenchman Cap dome implies high shear strain of ~ 32 to >100 depending of the initial plunge of the fold. Strain would be lower if there was a pure shear component as suggested by Scammell (1993) for the LSa (cf. Chapter 3). Assuming an initial eastward vergence, the common northward vergence of transposed minor folds indicates a sinistral rotation on plan view; the significance of which would require further analysis. Consequently, an appropriate mechanism of fold nappe formation must account for the recumbent folding of the stratigraphic layering accompanied by large magnitude foreland-directed shear strain.

4.3.3.3 Top-to-the-west Anstey Range detachment

A zone where top-to-the-west kinematic indicators abound crops out at the upper structural level of the dome, immediately below the Monashee décollement. At its

southernmost exposure on Fig. 4.2, the Monashee décollement is not only characterized by a lithologic break, but also coincides with an abrupt shear sense reversal. Indeed, pervasive top-to-the-east kinematic indicators prevail within rocks of the LSa, whereas pervasive top-to-the-west kinematic indicators are dominant in rocks of the underlying Monashee Complex. In some places, top-to-the-east C-S fabrics are overprinted by top-to-the-west shear bands, a feature also noticed by Journeay (1986). This zone is confined to the Sybley anticline and has a maximum thickness of 500 m on the cross-section of Fig. 4.3. Similar observations were made on the north flank of the dome, within the upper limb of the Sybley syncline, in an area located along strike of a zone of top-to-the-west shearing reported by Scammell (1986). Unfortunately, these two zones were discovered at the very end of the field season and their exact lateral extent and structural thickness remain to be thoroughly documented. Nevertheless, the two zones are inferred to be continuous because similar structural and metamorphic features were observed at both localities.

There is evidence that top-to-the-west shearing started in the presence of melt in the kyanite field and proceeded along a decompression and cooling path. Fig. 4.4a & b present two examples of specimens interpreted to have been deformed mostly in the magmatic or in the solid-state, respectively. The fabric of specimen Fg487 shown on Fig. 4.4a, is interpreted to have developed in the presence of melt because feldspars are euhedral and commonly show crystal faces against quartz that do not form ribbons. The fabric is constrictional with weak mineral alignment on the YZ section that contrasts with the strong mineral alignment observed on the XZ section. On the XZ section of Fig. 4.4a, most kyanite grains are imbricated and have their long axes running from the bottom left

to top right of the picture. This type of structure is interpreted as tiling formed during top-to-the-west shearing in the presence of melt. Following Mulchrone et al. (2005) tiling in this case is considered a reliable shear-sense indicator because the large majority of tiling structures, which are in excess of 60, indicate the same sense of shear. In contrast, Fig. 4.4b presents an example of top-to-the-west shearing that may have started in the presence of partial melt, but that mostly developed under solid-state conditions. With a volume of ~40%, the amount of leucosome is similar to that in the previous specimen, but in contrast with the even leucosome distribution observed on Fig. 4.4a, the leucosome in this specimen has segregated into bands, pods, and garnet pressure shadows. Feldspar and quartz in leucosome are recrystallized and the rock is characterized by C-S structures as well as shear bands that imply subsolidus deformation and leucosome recrystallization (Fig. 4.4b). Other top-to-the-west kinematic indicators observed within this zone include sillimanite-bearing shear bands, delta-shaped leucosomes in garnet tails, delta leucosome porphyroclasts, and shear band boudins (see Goscombe and Passchier, 2003). The zone will be referred to as the “Anstey Range detachment” throughout this chapter.

4.3.4 Field observations of metamorphic relationships

The geological map presented on Fig. 4.2 is divided into five metamorphic zones that are closely related to the map-scale structural geometry. From upper to lower structural levels, the five zones are: A) the upper sillimanite migmatitic zone confined to the base of LSa; B) the sillimanite-kyanite migmatitic zone confined to the Sybley anticline; C) the kyanite-muscovite zone confined to the Sybley syncline; D) the (muscovite)-kyanite-sillimanite migmatitic zone confined to the Mount-Grace anticline; E) the (muscovite)-sillimanite migmatitic zone confined to the Mount-Grace syncline. Muscovite in

parentheses indicates it is a retrograde phase, minerals are listed in order of abundance, and the label migmatitic also implies the presence of K-feldspar. As can be deduced from the name of the five zones, criteria used to delineate their boundaries were the type of Al_2SiO_5 , the presence of abundant leucosome and the presence of prograde muscovite. Andalusite is not pervasive, but has been locally observed in zones B-E. Data came from field and petrographic observations collected during this study combined with a reinterpretation of metamorphic assemblages reported in McMillan (1970), Journeay (1986) and Höy (1987), as well as a re-examination of the entire set of rock chips collected by Journeay (1986), many of which were stained for feldspar distinction. This new subdivision differs from these previous studies because it takes into account the progressive nature of metamorphism, which comprises a prograde and a retrograde path, rather than ascribing metamorphic overprints to discrete metamorphic events. Part of the rationale for this interpretation was provided by the geochronologic studies of Crowley and Parrish (1999) and Gibson et al (1999) which demonstrated that there was no evidence for discrete metamorphic events in Frenchman Cap dome, but rather evidence for protracted, progressive, metamorphism with U-Pb dates of monazite in schist decreasing with structural depth.

4.3.4.1 Zone A: sillimanite migmatitic zone of the base of the LSa

Metapelites at the base of the LSa within the Monashee décollement zone, consist of mylonitic sillimanite-bearing migmatitic schists that contain between 15 and 30% of leucosome. The outcrop volume of leucosome and leucogranitic intrusions is more than 30%. Leucosome forms millimetric granitic veinlets that are part of a C-S fabric indicating a top-to-the-east shear sense; or occurs as millimetric pods that crosscut the

foliation. These small veinlets and pods are uniformly distributed throughout the rocks; whereas larger, centimetric to decametric, granitic pods are common, but impossible to distinguish from dismembered granitic veins because of the high degree of strain that has affected these rocks during and after melt crystallization. Kyanite was observed by Journeay (1986) in only a few specimens and was reported to be extensively resorbed or almost completely transformed into sillimanite. Garnet is rather rare, commonly surrounded by leucosome, and highly resorbed on faces normal to the stretching lineation, which suggests resorption in the presence of melt that migrated toward the direction of minimum shortening. FG-462a & FG-462d are the two specimens collected on the north flank of Frenchman Cap dome (see Fig. 4.1 for location) that were used in Chapter 3 to derive the detailed P-T-t-d path presented on Fig. 3.6 (p. 184).

4.3.4.2 Zone B: Sillimanite-kyanite migmatitic zone, Sybley anticline

Rocks from the sillimanite-kyanite migmatitic zone occur within the anticline of the Sybley fold nappe, immediately below the Monashee décollement. In addition to coinciding with a boundary of lithologic units and with a sharp vergence reversal (cf. section 4.3.2 and 4.3.3.2), the Monashee décollement also coincides with a metamorphic discontinuity separating kyanite-bearing schists in the footwall from sillimanite-bearing schists in the hangingwall. The top-to-the-west Anstey Range detachment occupies the upper part of zone B. The main lithologic units are garnet-kyanite-K-feldspar-biotite ($\pm$ sillimanite, $\pm$ andalusite) schists and garnet-clinopyroxene migmatitic calc-silicates. Leucosome forms between 15-30% of outcrop areas. It is generally evenly distributed and occurs in pressure shadows of porphyroblasts, or locally occurs in layers as in Fig. 4.4b. Fig. 4.4c presents one of the areas where leucosome constitutes the highest rock volume.

It is also the area where the largest garnet and kyanite porphyroblasts were observed. Although not quantified, field observations pointed to the existence of a correlation between porphyroblast size and volume of leucosome. If real, this correlation suggests that part of garnet and kyanite growth occurred in the presence of melt (see below for possible reaction). Besides, centimetric to metric leucosome pods or melt-filled shear zones were not observed. The absence of these structures, which are so common to other migmatitic terrains (e.g. Collins and Sawyer, 1996; Gervais et al., 2004), suggests that melt extraction through a 3D network of dilatant sites was limited (cf. Brown, 2007 and references therein). Spectacular pink andalusite pseudomorphs after kyanite, similar to Fig. 4.4e, are common and andalusite has also been observed in pulled-apart kyanite (Journeay, 1986; Scammell, 1986). Sillimanite is present, locally around kyanite, but mostly around resorbed garnet and commonly in the top-to-the-west shear bands of the Anstey Range detachment.

Specimen DG-136 and JC-223 will be used in the next section to derive a P-T-t path for zone B. Specimen Dg-136 was collected by Gibson (1997) on the north-dipping flank of the dome (location on Fig. 4.1), whereas specimen JC-223 was collected by Crowley (1997) on the west flank of the dome on Ratchford Creek road (location on Fig. 4.2). It is noteworthy that fieldwork around the location of specimen DG-136 revealed it was collected in one of these highly migmatitic areas illustrated on Fig. 4.4c where top-to-the-west kinematic indicators of the Anstey Range detachment are common.

4.3.4.3 Zone C: kyanite- muscovite zone of the Sybley syncline

Metamorphic grade decreases down section toward the lower limb of the Sybley fold nappe syncline in the kyanite-muscovite zone. The transition from zone B to zone C

occurs over a distance of a few hundred meters without noticeable structural break. The following features have been observed in two transects from the west flank of Frenchman Cap dome and similar observations can be inferred from the description of specimens collected on the north flank of the dome by Gibson (1997). The decrease in metamorphic grade is manifested in many ways: 1) muscovite becomes abundant and parallel to the transposition foliation (Fig. 4.5c); 2) leucosome volume decreases from 15-30% to <10% (compare Fig. 4.4a-c with Fig. 4.4d); 3) the size of the largest garnet and kyanite porphyroblast decreases to < 1 cm; 4) compositional layering (relict bedding?) is more apparent (Fig. 4.4d); and 5) biotite color changes from brown-red to green in thin-section (Fig. 4.5c). Indeed, biotite color is related to its TiO_2 content, which in turn has been shown to increase as function of temperature (Henry et al., 2005 and references therein). Taken together, these features imply that rocks from zone C formed at a temperature lower than the muscovite dehydration melting reaction. On the map of Fig. 4.2, zone C is located between two oppositely facing muscovite-out isograds because rocks structurally above (zone B) and below (zone D) have clearly been partially melted above that reaction. Andalusite was reported by Scammell (1986) and Gibson, (1997) at a few outcrops of zone C.

Specimen DG-167 will be used in the next section to construct the P-T-t path of zone C. This specimen was collected by Gibson (1997) on the north flank of the dome (location on Fig. 4.1) and was the focus of extensive geochronologic studies (Gibson et al., 1999; Foster et al. 2002, 2004; Foster and Parrish, 2003).

4.3.4.4 Zone D: (muscovite)-kyanite-sillimanite migmatitic zone of the Mount-Grace anticline

Leucosome is abundant down-section within Mount-Grace anticline. Basement orthogneiss forms the core of the anticline and commonly contains 30-50% leucosome and some meter-scale outcrops without structural coherence can be considered as diatexite (Sawyer, 1999). As demonstrated in chapter 2 and in Crowley et al. (2008), basement rocks were also partially melted in the Proterozoic. It was thus difficult to determine whether or not basement rocks contain Precambrian or Eocene leucosome. The overturned limb of the fold nappe is dominated by metapelites that share most characteristics described above for zone B except that sillimanite and muscovite are more abundant and garnet more resorbed, as illustrated on Fig. 4.5d. A P-T-t path for this zone is reconstructed by combining already published ages from specimen JC-191 — an andalusite-kyanite leucosome shown on Fig. 4.4e — with P-T information gained from its host schist that was collected by Crowley (1997) on the northwest-dipping flank of Frenchman Cap dome (location on Fig. 4.2).

4.3.4.5 Zone E: (muscovite)-sillimanite migmatitic zone of the Mount-Grace syncline

The (muscovite)-sillimanite migmatitic zone is located within the lower limb of the Mount-Grace syncline, upsection from basement rocks (Fig. 4.2, 4.3). Leucosome forms 15-30% of outcrops and is either uniformly distributed or segregated into veins that gives the rock a stromatitic structure. Kyanite has been observed in some leucosome of Zone E, but is conspicuously absent from the host schist. The few garnet grains observed were extensively resorbed and were in the core of sillimanite-biotite-plagioclase-quartz pods.

These pods give the rock a faserkiesel texture. Retrograde muscovite, recognized because it crosscuts the foliation, is also common. FG-459 is a key specimen collected within zone E on the northwest flank of Frenchman Cap dome (location on Fig. 4.2). It consists of a sillimanite-muscovite-biotite schist, typical from zone E, that hosts a spectacular garnet + kyanite leucosome pod. The leucosome, shown on Fig. 4.4f, is rich in euhedral K-feldspar and surrounded by a dark biotite restite that contains many euhedral garnet crystals, 3-10 mm in diameter, and kyanite up to 8 mm long. Kyanite is also present within the leucosome and one crystal is 30 mm long (not shown). On the bottom right of Fig. 4.4f, pink euhedral crystals are interpreted to be andalusite pseudomorphs after kyanite. This specimen is very important because it demonstrates that rocks from zone E crossed the muscovite dehydration melting reaction in the kyanite field. Hence, the typical assemblage of zone E, consisting of sillimanite-muscovite-biotite with rare garnet, must have a retrograde origin. The retrograde P-T-t path is constructed by combining these observations with timing constraints provided by specimen JC-258 collected by Crowley (1997) on the west flank of the dome on Ratchford Creek road (location on Fig. 4.2).

4.4 Pressure-Temperature-time paths

4.4.1 Methodology

In the following sections, metamorphic microtextures, mineral composition and textural relationships between minerals are first described in light of results of rock specific metamorphic phase diagram modeling conducted on specimens for which U-Pb dates are available. Specimen names beginning with “FG” were collected during this study. Specimen names beginning with “JC” were dated by the U-Pb ID-TIMS method

by Crowley and Parrish (1999). The JC specimens discussed herein were investigated for their petrography, analyzed for their mineral composition, and modeled to calculate pseudosections. Specimen names beginning with “DG” were collected by Gibson (1997) and have been extensively investigated to better understand the U-Th-Pb isotopic system in monazite by different geochronologic methods (Gibson et al., 1999, Foster et al. 2002, 2004; Foster and Parrish, 2003). Foster et al. (2004) further linked monazite composition of DG-136 and DG-167 with metamorphic reactions involving major silicates. I did not conduct any new work on DG specimens, but FG specimens collected nearby were investigated.

4.4.2 *Analytical methods*

See chapter 1.

4.4.3 *Zone A: sillimanite migmatitic zone at the base of the LSa:*

Chapter 3 of this thesis presented a detailed retrograde P-T-t path for zone A located within mylonitic rocks of the Monashee décollement at the base of the LSa (Fig. 3.6). This path suggested that peak metamorphic conditions of $\sim 760^{\circ}\text{C}$ at 800 MPa, within the sillimanite + melt field, were reached between ~ 120 and ~ 62 Ma, after which the rock cooled and decompressed to less than $\sim 700^{\circ}\text{C}$ and ~ 300 MPa between ~ 62 and ~ 57 Ma. Structurally controlled hornblende growth in both tails of a sigmoidal clinopyroxene aggregate further indicated that decompression and cooling occurred during top-to-the-east shear strain. A similar interpretation is implied by the texture observed on Fig. 4.5a, from specimen Fg-462a, that shows preferential garnet resorption in the bottom left and top right quadrants, which indicates a top-to-east sense of shear. Resorption of garnet grains occurred in the presence of melt in the sillimanite field by the retrograde net-

transfer reaction (ReNtr):

because the products of this reaction are present in garnet pressure shadows.

Consequently, rocks from zone A of the Monashee décollement were cooled and exhumed to upper crustal conditions by ~60 Ma, while footwall rocks of zone B were reaching partial melting condition at more than ~40 km depth, as described below.

4.4.4 Zone B: sillimanite-kyanite migmatitic zone of the Sybley anticline

4.4.4.1 Metamorphic textures and mineral composition

Specimen JC-223 is an andalusite-sillimanite-garnet-kyanite-K-feldspar-biotite schist hosting a kyanite-bearing leucosome that was dated by Crowley and Parrish (1999).

These authors reported that the leucosome contains kyanite grains much larger than its host schist and that these kyanite grains are pseudomorphed into andalusite. The host schist specimen JC-223 is medium grained and is not mylonitic. It has a foliation defined by biotite aggregates, quartzo-feldspathic layers, kyanite and K-feldspar crystals, and a lineation defined by sillimanite-biotite seams, kyanite, and elongated leucosome rods. Leucosome commonly consists of quartz and K-feldspar. Quartz forms either equant or elongated grains and shows undulose extinction, whereas K-feldspar is aligned with the lineation but is generally euhedral. Parts of leucosome show evidence of solid-state strain in the form of trains of subgrains at the margins of subhedral larger grains. This microtexture suggests that deformation accompanied melt crystallization or that a small amount of strain postdated crystallization. Kyanite is pulled-apart and filled with what appear to be tiny crystals of muscovite and altered plagioclase. Garnet has a diameter of 1-6 mm, has inclusion-rich cores and inclusion-free rims, and is partially resorbed in a

direction oblique to the lineation (Fig. 4.5b). Sillimanite, biotite, ilmenite and rutile are the main phases surrounding resorbed garnet which could suggest resorption during shearing similar to the interpretation presented above for specimen Fg-462a (Fig. 4.5a). Unfortunately, it is impossible to relate these textures to a specific kinematics because the specimen was not oriented at collection. However, It is likely that the structures observed in specimen JC-223 indicate top-to-the-east shearing because top-to-the-east C-S fabric were present at outcrops located near the collection site of specimen JC-223. Inclusions in garnet cores comprise quartz, plagioclase, biotite, muscovite, allanite and ilmenite, whereas inclusions in garnet rims comprise rutile, ilmenite, monazite, zircon and biotite.

Exceptional for such high-grade rocks is the preservation of growth zoning in garnet (Fig. 4.6). Only six microprobe analyses of plagioclase and biotite were conducted because I judged that conventional thermobarometry was not likely to provide reliable results. This judgement was based on the study of Spear and Parrish (1996), which examined rocks of similar metamorphic grade from the Valhalla Complex located immediately south of Fig. 4.1 and which demonstrated that temperatures obtained by garnet-biotite thermometry could vary by many hundreds of degrees. The focus of electron microprobe investigation was rather on garnet that can provide important information for interpreting results of phase equilibria modeling (e.g. Foster et al., 2004; Indares et al., 2008). The only plagioclase grains observed in the vicinity of garnet has a core composition 0.22 anorthite, whereas the rim has a composition of 0.26 anorthite. Biotite Fe/Fe+Mg ratio varies between 0.64 and 0.72. Fig. 4.6a shows compositional profiles across the garnet with the highest spessartine content of all garnets from the two thin-sections examined. Profiles across this garnet are not symmetric, but other garnets

had symmetric profiles. Only the left part of the profiles shown on Fig. 4.6a is thus described. Spessartine (X_{sps}) is bell-shaped with a core composition of ~9 mol % that decreases to ~2.5 mol % at the inner rim and then increases again to ~8.5 mol % at the outer rim. Pyrope (X_{pyr}) has a core composition of ~10 mol % that progressively increases to ~17 mol % at the inner rim and decreases to ~10.5 mol % at the outer rim. Grossular (X_{grs}) has a core composition of ~9 mol % that progressively decreases to ~4 mol % at the outer rim. Despite variation in composition at cores and outer rims, the inner rims of the three garnet grains for which a compositional profile was acquired did not show any variation.

4.4.4.2 Phase equilibria modeling

Phase equilibria modeling with the program PerpleX'06 (Connolly and Petrinì, 2002; Connolly, 2005) in the extended chemical system TiMnCaNaKFMASH tightly constrains the P-T path of specimen JC-223. The calculation included the melt model of White et al. (2001) and the Ti-Mn-biotite solution model of Powell and Holland (1999). Table 4.1 presents the bulk composition of specimen JC-223 and Figure 4.7a shows the calculated pseudosection. A pseudosection for specimen DG-136, which is used in the next section to derive timing constraints, was also calculated and shows similar topology to the pseudosection of specimen JC-223.

The prograde path is constrained by garnet core isopleths, the garnet inclusion assemblage, and the matrix paragenesis. Calculated spessartine, grossular and pyrope isopleths for the garnet core analysis with the highest spessartine intersect at ~550 MPa and ~575°C, which is very close to the 1% garnet volume isopleth (bottom left of Fig. 4.7a). The inclusion assemblage consisting of quartz, plagioclase, biotite, muscovite,

allanite and ilmenite, constrains the initial P-T path to lie within the staurolite-plagioclase-garnet-biotite-muscovite-ilmenite field. The absence of staurolite as an inclusion in garnet is interpreted to reflect its consumption during garnet growth. It is noteworthy that a good intersection is obtained only for the garnet core with the highest manganese content and that this value is near the value predicted for the first increment of garnet growth ($X_{\text{grs}} = 0.09$; $X_{\text{pyr}} = 0.10$; $X_{\text{sps}} = 0.09$; vs. $X_{\text{grs}} = 0.11$; $X_{\text{pyr}} = 0.11$; $X_{\text{sps}} = 0.11$). Diffusion commonly smoothens garnet zoning at high temperature and should split isopleths from the initial predicted composition. It would be highly fortuitous that garnet isopleths smoothed by diffusion still yield a good intersection within a P-T field consistent with the inclusion assemblage. It could be possible if isopleths spacing was narrow, but if this were the case, a good intersection would be obtained for other core composition as well. Consequently, it must be concluded that diffusion was limited for this sample.

The absence of prograde sillimanite and the abundance of kyanite indicate that the staurolite-out reaction was crossed in the kyanite field. The absence of prograde muscovite, as well as the predominance of kyanite as the main Al_2SiO_5 within both leucosome and host schist imply that the prograde net-transfer (PrNetr) muscovite-dehydration melting reaction (e.g. Spear et al., 1999 and references therein):

was crossed in the kyanite field. Also, the correlation between garnet size and leucosome volume illustrated on Fig. 4.4c suggests that some garnet grew in the presence of melt, most likely by the continuous biotite dehydration melting reaction (Le Breton and

Thompson, 1988; Patiño-Douce and Johnston, 1991; Gardien et al., 2000; Spear et al., 1999):

PrNetr 2: biotite+ aluminosilicate + plagioclase + quartz = garnet + K-feldspar + melt

On Fig. 4.7a, the products of PrNetr 1 and PrNetr 2 are predicted to be stable together in a restricted P-T field bounded by the Ms-out, Bt-out and Ky-in lines, which occurs between pressures of ~1000 and 1300 MPa and temperatures of ~800-835°C (Fig. 4.7a). A P-T trajectory passing at higher pressure would lead to subsolidus biotite breakdown via a reaction such as:

PrNetr 3: biotite+ kyanite = muscovite + garnet

The extent of these two reactions must have been limited or else the reactant kyanite would not be as abundant and would show evidence of resorption. Furthermore, these estimate of peak P-T conditions are most likely minima because the only biotite solution model available at the time of the calculation (Powell and Holland, 1999) as been shown to underestimate the upper thermal stability of biotite (White et al., 2007; Tajcmanova et al., 2009). Nevertheless, a conservative estimate of peak P-T condition of ~825°C at 1100 MPa is considered reasonable because a higher temperature would have consume kyanite via PrNetr 2 and a higher pressure would be difficult to reconcile with the ~32 km crustal thickness below the Monashee Complex (see below).

The proposed prograde path is thus almost parallel to the kyanite-sillimanite reaction, but at higher pressure in the kyanite field ending in the garnet + biotite + kyanite + rutile + liquid + K-feldspar field (Fig. 4.7a). The peak pressure estimate of ~1100 MPa is consistent with the peak pressure of ~1020 MPa calculated for specimen DG-136, which

is located in zone B on the north-dipping flank of the dome (see Fig. 4.1 for location), by the average P-T mode of Thermocalc (Foster et al., 2004). The peak temperature estimate of $\sim 728^{\circ}\text{C}$ calculated by Thermocalc is too low by at least 75°C , because the peak metamorphic assemblage is the same for specimen DG-136 as that of JC-223. This discrepancy most likely reflects modification of biotite composition during retrogression (see below and Spear and Parrish, 1996). A lower peak temperature also implies an underestimation of peak pressure

The retrograde path is constrained by mineral parageneses in garnet pressure shadows, by the presence or absence of key minerals, and by andalusite pseudomorphs after kyanite. Unpublished modeling assuming that melt loss had occurred suggested that it would not modify the topology of the pseudosection enough to invalidate its use to describe the retrograde P-T path. The presence of biotite, sillimanite, plagioclase, and quartz in garnet pressure shadows suggests that ReNtr1 was crossed on the retrograde path within the sillimanite field. Further evidence for this reaction includes the structurally controlled resorption texture around most garnet prophyroblasts (Fig. 4.5b), and the sharp increase in both Fe/Fe^{+} and X_{sps} ratios at the rims of garnet (Fig. 4.6; Spear et al., 1999; Kohn and Spear, 2000). Garnet volume isopleths (Fig. 4.7a) indicate that the continuous ReNtr 1 is nearly isothermal. Consumption of garnet by ReNtr 1 thus required cooling. The predominance of rutile instead of ilmenite in garnet pressure shadows further indicates that the initial stage of cooling and decompression took place in the rutile field. Furthermore, the absence of cordierite implies a retrograde path passing above the cordierite-in reaction that has a positive shallow slope passing through the lower right quadrant of Fig. 4.7a. Additionally, the rarity of retrograde muscovite

suggests that the reaction:

, which is the inverse of PrNtr 1, was not crossed on the retrograde path. Furthermore, the spectacular pink andalusite pseudomorphs after kyanite, similar to the texture depicted on Fig. 4.4e, suggest that water or a melt phase was still present to facilitate the reaction. A retrograde P-T path passing close to the intersection between the vapour present melting reaction and PrNtr 1 ($\sim 660^\circ\text{C}/300 \text{ MPa}$), which is labeled IP1 on Fig. 4.7a (cf. Spear et al., 1999), would not cross ReNtr2, but would lead to water release upon crystallization of the remaining melt phase.

Consequently, all these observations indicate a retrograde path from the thermal and baric peak of $\sim 825^\circ\text{C}/1100 \text{ MPa}$ involving an initial stage of cooling and decompression within the sillimanite + rutile + melt field, followed by further decompression and cooling in the sillimanite + ilmenite + melt field, and then passing through the andalusite field near IP1 at $\sim 660^\circ\text{C}/300 \text{ MPa}$. Because garnet is still present in the rock, it is likely that the proposed retrograde path was too rapid, or the reaction kinetic too sluggish, for garnet exhaustion by ReNtr 1. It is also possible that progressive melt crystallization upon cooling would have prevented much reaction between garnet and melt.

4.4.4.3 Timing constraints and P-T-t path

Specific dates can be assigned to several P-T points along the P-T path presented on Fig. 4.7a by integrating petrologic data with results of six detailed geochronologic studies conducted on the same rocks or other rocks of zone B (Crowley and Parrish, 1999;

Gibson et al., 1999; Crowley et al., 2001; Foster et al. 2002; Foster and Parrish, 2003; Foster et al., 2004). Specimen JC-223 shares similar field relationships, reaction textures, garnet zoning and IDTIMS monazite U-Pb dates as specimen DG-136 (see Fig. 4.1 and 4.2 for locations). Timing constraints from both specimens, as well as from the crosscutting pegmatite specimen JC-263 (Crowley et al., 2001), are combined with results from the pseudosection calculated for specimen JC-223 to construct a detailed P-T-t path for rocks of zone B.

Important relationships between accessory and major metamorphic phases have been documented by Foster et al. (2004) for specimen DG-136. Allanite is the common LREE-rich phase included within garnet and kyanite cores, whereas monazite is the LREE-rich phase within rims of these two phases as well as in the matrix. The same relationships were observed in specimen JC-223. Furthermore, Foster et al. (2004) documented that monazite in specimen DG-136 has distinct yttrium contents, depending on its textural position. High-yttrium zone-1 monazite is observed within rims of some kyanite crystals; within low-yttrium rims of garnet beyond a high-yttrium annulus; and in cores of matrix monazite grains. Medium-yttrium zone-2 monazite is observed within rims of other kyanite crystals and within the inner rims of matrix monazite grains. By contrast, high-yttrium zone-3 monazite is observed at the outer rims of two matrix monazite grains. These relationships — summarized as a sketch on Fig. 4.8— were interpreted by Foster et al. (2004) as evidence for: 1) growth of zone-1 monazite via allanite breakdown near the kyanite-in isograd while garnet was still growing and was sequestering yttrium from monazite; 2) growth of zone-3 monazite on the retrograde path during garnet breakdown that liberated yttrium; and 3) growth of zone-2 monazite during garnet growth on the

prograde path between growth of zone-1 and zone-3 monazite.

Monazite of specimen DG-136 was dated by Foster et al. (2004) by the in-situ laser ablation method (LA-ICP-MS). Six analyses of low-Y zone 1 monazite yielded concordant $^{206}\text{Pb}/^{238}\text{U}$ dates between ~ 77 and ~ 74 Ma that together gave a concordia age (cf. Ludwig, 1998) of 75 ± 1.4 Ma (MSWD=1.1; probability=0.39). Seven analyses of medium-Y zone-2 monazite yielded concordant $^{206}\text{Pb}/^{238}\text{U}$ dates between ~ 65 and ~ 60 Ma that together gave a concordia age of 63.2 ± 1.5 Ma (MSWD=1.3; probability=0.22). Two analyses of high-Y zone 3 monazite yielded concordant $^{206}\text{Pb}/^{238}\text{U}$ dates of 55.8 ± 2.6 and 59.9 ± 3.0 Ma. Three studies published since Foster et al. (2004) documented a spatial correlation between the allanite-monazite isograd and the Al_2SiO_5 -in isograd (Wing et al., 2003; Tomkins and Pattison, 2007; Janots et al., 2008). As shown on figures 4.7a and 4.8, the ~ 75 Ma age of zone-1 monazite is interpreted as the age at which the rock crossed the kyanite-in isograd. The proposed P-T path presented on Fig. 4.7a crosses the kyanite-in line within a trivariant field in which kyanite and staurolite are stable, and across which more than 8% vol. of garnet is predicted to grow. This trivariant field is located immediately below the 75 Ma ellipse on Fig. 4.7a. Such an important garnet growth episode, which reached 10% across the same field on the pseudosection calculated for specimen Dg-136 (not shown), would have sequestered yttrium. Zone-1 monazites would thus have grown in a low yttrium environment. As suggested by Foster et al. (2004) the high amount of yttrium that must have been available for the growth of zone-3 monazite was most likely provided by garnet that broke down on the retrograde path via ReNtr 1. The two monazite dates of ~ 59 and ~ 56 Ma are thus interpreted to reflect two increments of monazite growth during garnet breakdown by ReNtr 1. The

significance of the medium-yttrium zone-2 monazite is more equivocal, but can be deduced from the topology of the pseudosection and the time gap between zone-1 and zone-2 monazite. As can be seen on Fig. 4.7a, the P-T path is predicted to cross a wide tetravariant field between the staurolite-in and the melt-in reactions. It is not likely that monazite grew in this field because there would be monazite dates closer in age to ~75 Ma if this had been the case. It is rather likely that the ~63.2 Ma, medium-yttrium, zone-2 monazite growth was linked to dissolution of apatite and/or zone-1 monazite during the vapour-present melting (labelled melt-in on Fig. 4.7a; cf. Zeng et al., 2005). Because there are medium-yttrium monazites in specimen DG-167 (described in the next section) that did not reach temperature high enough for muscovite-dehydration melting, but preserve evidence of vapour-present melting, I suggest that medium-yttrium monazite zones/grains in specimen DG-136 and DG-167 grew in the vapour-present melting field between the melt-in and muscovite-out reactions.

Additional timing constraints are provided by the study of Crowley and Parrish (1999). JC-223 is the host of a kyanite-bearing leucosome dated by U-Pb IDTIMS of zircon and monazite fractions extracted from the leucosome by standard mineral separation techniques. Three zircon fractions lying slightly below concordia yielded $^{207}\text{Pb}/^{235}\text{U}$ dates of 56.4 ± 0.2 , 58.4 ± 0.2 and 59.0 ± 0.2 , whereas reversely discordant $^{207}\text{Pb}/^{235}\text{U}$ monazite dates yielded two groups, one between ~65 and 63 Ma and the other between ~57 and 55 Ma. JC-334 is an adjacent pelitic schist that yielded reversely discordant $^{207}\text{Pb}/^{235}\text{U}$ monazite dates ranging between ~61 and ~56 Ma. There are two ways to interpret these results. The first is to assume that all zircon grains of the leucosome crystallized synchronously and to average the age of the three zircon fractions

to get an age of 57.5 ± 1.5 Ma (as in Crowley and Parrish, 1999). However, the single-age crystallization hypothesis is unlikely because the P-T path of specimen JC-223 implies progressive melt crystallization during cooling and decompression. Alternatively, all these dates can be considered meaningful and harmonized with results of Foster et al. (2004). According to this new interpretation monazite older than ~ 60 Ma would have grown on the prograde path, whereas monazite and zircon younger than ~ 59 Ma would have grown on the retrograde path during progressive melt crystallization and garnet consumption by ReNtr 1. Finally, the 55.5 ± 1.0 Ma age of the undeformed pegmatite dyke JC-263 crosscutting the transposition foliation provides a minimum age for ductile deformation in zone B (Crowley et al., 2001).

In summary, rocks from the sillimanite-kyanite migmatitic zone B, located in the anticline and underlying overturned limb of the Sybley fold nappe, followed a hairpin P-T-t path (Fig. 4.7a) involving relatively slow burial and heating from $\sim 650^\circ\text{C}$ and 700 MPa at ~ 75 Ma to $\sim 800^\circ\text{C}$ and 1100 MPa at ~ 59 Ma. This prograde path, which reached partial melting conditions in the kyanite field, was then followed by a much faster retrograde trajectory down to the andalusite field by ~ 55 Ma. This P-T-t path contrasts with those derived for immediately overlying and underlying rocks of zone A and C, respectively.

4.4.5 Zone C: Kyanite-muscovite zone of the Sybley syncline

4.4.5.1 Petrography and mineral composition

The metamorphic evolution of the Sybley syncline is investigated by combining previously published data from specimen DG-167 (Foster et al. 2004) with results of new phase equilibria modeling from that specimen and petrographic observations made on

other pelitic schists from that zone. Specimen DG-167 is an andalusite-sillimanite-kyanite-garnet-muscovite-biotite schist that contains garnet up to 6 mm in diameter (bulk composition in Table 4.1). Three other pelitic schists from this zone were examined and revealed that: 1) there is less than 10% of trondhjemitic leucosome; 2) biotite has a green color (in contrast with the red color of biotite in zones B, D, and E); 3) muscovite is everywhere parallel to the foliation and appears in textural equilibrium with other matrix minerals; 4) ilmenite is the main Ti oxide inclusion in garnet; 5) sillimanite, where present, occurs preferentially at the margin of partially resorbed garnet.

Compositional profiles of a 6 mm wide garnet from specimen DG-167 is presented on Fig. 4.6b (from Fig. 2j in Foster et al. 2004). This garnet is characterized by an asymmetric zoning profile with the left portion typical of prograde zoning and a flat right portion typical of homogenization at high temperature. Furthermore, a X-Ray Mn map from Foster et al. (2004) shows that the prograde zoning is cut by a resorbed margin. The right portion of the profile is characterized by a bell-shaped X_{sps} that decreases from ~0.03 in the core, to ~0.01 at the inner rim, before rising again to ~0.05 at the outer rim. X_{grs} also has a bell shape that decreases from ~0.19 in the core to ~0.1 at the outer rim. The profile for the Fe/Fe + Mg ratio is similar to X_{sps} and decreases from ~0.88 in the core, to ~0.80 at the inner rim, before rising again to ~0.85 at the outer rim. Such spikes at the outer rim for X_{sps} and the Fe / Fe + Mg ratio are diagnostic of garnet breakdown by a net-transfer reaction similar to ReNtr 1 (see Kohn and Spear, 2000). The flat profiles on the left have compositions similar to the inner rim on the right side.

4.4.5.2 Phase equilibria modeling

The prograde path of specimen DG-167 is constructed by comparing predictions of

thermodynamic modeling to the measured garnet core composition and to the inclusion assemblages in garnet. Isopleths of X_{sps} , X_{pyr} , X_{grs} for garnet having the highest X_{sps} (Foster et al., 2004) intersect at ~ 590 MPa and 530°C on the bottom left corner of Fig. 4.7c. In contrast to the pseudosection of specimen JC-223, the trivariant field within which staurolite and kyanite are stable is rather narrow and can be treated as a single reaction. This reaction is inferred to have been crossed at pressure above the sillimanite-kyanite boundary, because of the absence of prograde sillimanite and the abundance of kyanite; and below the ilmenite-to-rutile reaction, because garnet from all specimens observed includes mainly ilmenite. The P-T path of Fig. 4.7c is thus shown to cross the kyanite-in reaction at ~ 750 MPa/ 630°C . The peak metamorphic assemblage consisting of biotite, muscovite, garnet, and kyanite is predicted to be stable over a large trivariant field and provides only a loose constraint on peak P-T conditions (Fig. 4.7c). Interestingly, calculated garnet isopleths ($X_{\text{grs}} = 0.1$; $X_{\text{alm}} = 0.7$; $X_{\text{pyr}} = 0.18$) for the inner rim composition to the right of the compositional profiles, which is the same as the composition of the flat profiles to the left (Fig. 4.6b), intersect at $\sim 660^\circ\text{C}/900$ MPa. These P-T conditions are the same P-T estimates calculated with the average P-T mode of THERMOCALC that yielded $665 \pm 23^\circ\text{C}$ at 880 ± 120 MPa (Foster et al., 2004) and were interpreted as an estimate of the metamorphic peak. However, because there is textural and chemical evidence of a net-transfer retrograde reaction (and maybe of Fe-Mg exchange reaction as well) operating at the rim of the garnet, it is likely that this estimate represents a minimum (Spear, 1993; Kohn and Spear, 2000). Furthermore, growth zoning in garnet should preclude isopleths intersection elsewhere than at the core of the grain with the highest Mn content because it modifies the bulk composition (e.g. Evans,

2004). However, given that 1/3 of this 6 mm wide garnet has been homogenized at high temperature, it is likely that most garnet zoning in this sample have also been smoothened by diffusion. This way, the bulk composition would not have been modified significantly by the rare garnet zoning. Accordingly, if the homogenized garnet equilibrated with the matrix phases during retrogression at high temperature, it would be possible to obtain isopleth intersection at lower metamorphic conditions than the peak. The intersection is thus interpreted as a minimum estimate. Finally, the presence of ~10% trondhjemitic leucosome suggests that the rock crossed the wet-solidus, as indicated by partial melting experiments of metapelites (Patiño-Douce and Harris, 1998; Garcia-Casco et al., 2003). The P-T path of Fig. 4.7 is thus shown to cross the wet-solidus along the decompression path.

Information on the retrograde P-T path is provided by the observation that muscovite, biotite, plagioclase, sillimanite and quartz occur within pressure shadows of resorbed garnet in most specimens from zone C. The presence of prograde muscovite constrains the retrograde path at pressure higher and temperature lower than the muscovite-out reaction. Furthermore, evidence for garnet resorption implies decompression because garnet volume isopleths have a shallow positive slope at temperature below the muscovite-out reaction (Fig. 4.7c). This suggests an initial nearly isothermal decompression followed by cooling and further decompression to the andalusite field occurring within the muscovite stability field (Fig. 4.7b).

4.4.5.3 Timing constraints and P-T-t path

Good timing constraints on the P-T path presented on Fig. 4.7c are provided by the study of Foster et al. (2004), which dated monazite of specimen DG-167 by the in-situ

LA-ICP-MS method. Specimen DG-167 shares the same textural relationships between allanite, garnet, kyanite and monazite as specimen DG-136, with allanite occurring in the core of garnet and kyanite, and monazite occurring at the rims of garnet and kyanite as well as in the matrix (Fig. 4.8). In contrast with specimen DG-136, most monazites of specimen DG-167 have a medium yttrium concentration similar to zone-2 monazites from specimen Dg-136, but it is still possible to distinguish two zones based on yttrium content (Foster and Parrish, 2003). Zone-1 monazite has an yttrium content of 0.4-0.8 wt% and occurs in the core of matrix monazite grains and in inclusions at the rims of garnet; whereas zone-2 monazite has an yttrium content of 0.9-1.1% and occurs at the rims of matrix grains. Zone-1 monazites were analyzed 24 times in the combined studies of Foster and Parrish (2003) and Foster et al. (2004) and yielded concordant dates between ~59 and ~54 Ma. I re-examined the combined dataset of those studies and found it is possible to isolate three different ages that do not overlap within uncertainty by using the concordia age approach of Ludwig (1998). The six oldest dates yield a concordia age of 59.0 ± 0.8 Ma (MSWD = 1.40; prob. = 0.15); the 13 medium dates yield a concordia age of 57.0 ± 0.4 Ma (MSWD = 1.10; prob. = 0.33); the 5 youngest dates yield a concordia age of 54.8 ± 0.7 Ma (MSWD = 1.10; prob. = 0.36); all MSWD and prob. are given for concordance and equivalence (cf. Ludwig, 1998 & 2003). Only one zone-2 monazite, which occurs as a thin rim over zone-1 monazite, was analyzed and yielded the youngest date of all analyses at ~53 Ma, which is within uncertainty of the youngest zone-2 ages. It was thus included in the calculation of the youngest concordia age. Following the same rationale presented for specimen DG136, the oldest monazite date of ~59 Ma is considered the best approximation for the age at which the rock entered the

kyanite field. By contrast with modeling results for specimen JC-223 (Fig. 4.7a), only 3-4 vol.% of garnet is predicted to grow at the staurolite-kyanite transition. In this case, the smaller amount of garnet would not have sequestered as much yttrium as if garnet was more abundant, and more yttrium would then have been available for the initial growth of a higher yttrium monazite relative to zone-1 monazite of specimen DG136. The ~57 Ma zone 1 monazite is interpreted to have grown within the muscovite + melt field, following the same reasoning as for the growth of zone-2 monazite in specimen DG-136. Finally, the ~55 Ma youngest monazite probably grew during garnet breakdown along the retrograde path, but the lower amount of melt present in the rocks prevented extensive garnet resorption and associated release of yttrium. Thermal and baric peak conditions would thus have been reached between ~57 and ~56 Ma.

In summary, rocks from the Sybley syncline followed a hairpin path from ~550°C/510 MPa, entered the kyanite field at a pressure of ~800 MPa at ~59 Ma, reached peak conditions of ~660°C/900 MPa at ~57 Ma, and were retrogressed to the andalusite field sometime after ~53 Ma (Fig. 4.7c). In contrast with rocks of all the other zones, rocks from zone C did not cross the muscovite dehydration melting reaction. Most importantly, zone C rocks from the Sybley syncline were at ~700-800 MPa at ~59 Ma while zone B rocks from the overlying Sybley anticline were reaching their baric peak at ~1100 MPa.

4.4.6 *Zone D: (muscovite)-kyanite-sillimanite migmatitic zone of the Mount-Grace anticline:*

4.4.6.1 Petrography

Metapelites of zone D located on the overturned limb of the Mount-Grace fold nappe

present similar metamorphic assemblages to rocks from the sillimanite-kyanite migmatitic zone B. Specimen JC-191 is a spectacular leucosome collected by Crowley (1997a) north of Blais Creek in the NW corner of Fig. 4.2. The leucosome contains large blue kyanite crystals 2-4 cm long that were variously pseudomorphed into pink andalusite (Fig. 4.4e). Fig. 4.5d is a photomicrograph of the migmatitic muscovite-sillimanite-garnet-kyanite-biotite schist hosting this leucosome. Muscovite is a retrograde phase because it occurs only at the edges of partially resorbed kyanite (Fig. 4.5d) or as porphyroblasts 1-3 mm long that contains garnet, biotite and kyanite inclusions. Sillimanite and biotite seams are common around resorbed garnet (Fig. 4.5d), which suggest garnet resorption in the presence of melt by ReNtr 1 (e.g. Kohn and Spear, 2000). As originally noted by Crowley and Parrish (1999), the larger size of kyanite in leucosome than in matrix suggests it grew in the presence of melt.

4.4.6.2 P-T-t path

The host schist of specimen JC-191 has a similar bulk composition as specimen JC-223 (Table 4.1), and the pseudosection of Fig. 4.7a can thus be used to describe its metamorphic evolution. The presence of andalusite pseudomorphs after kyanite in the leucosome and the microtextural evidence for garnet resorption by ReNtr 1 suggest that rocks of zone D followed a similar retrograde P-T path as that of zone B (Fig. 4.7a vs. 4.7c). The presence of retrograde muscovite in rocks of zone D could be explained by a retrograde P-T trajectory passing above IP1, and thus crossing ReNtr 2 (e.g. Spear et al., 1999). Furthermore, Crowley and Parrish (1999) reported a 57.8 ± 1.5 Ma age for crystallization of leucosome JC-191 based on a similar rationale described in section 4.2.3.3 for leucosome JC-223. This age is the same, within uncertainty, as the

crystallization age of leucosome specimen JC-223 of zone B. A minimum age estimate for cooling below $\sim 550 \pm 50^\circ\text{C}$ is provided by an hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age of 55.6 ± 1.7 Ma (95% of ^{39}Ar released) obtained from a garnet amphibolite collected within the Mount-Grace anticline on the west-dipping flank of the dome (Sanborn 1996). It is noteworthy that this age is the same as the minimum age for deformation provided by the crystallization age of 55.5 ± 1.0 Ma of the crosscutting pegmatite JC-263 from zone B. The similarity between the geochronologic dataset from zones B and D is interpreted as evidence that rocks from Sybley and Mount-Grace anticlines followed a similar retrograde P-T-t path presented on Fig. 4.7a for zone B and Fig. 4.7c for zone D. If this interpretation is correct, it suggests that the lower-grade, zone C rocks of the Sybley syncline were engulfed above and below during the flow of zones B and D (see section 4.6.2 below).

4.4.7 Zone E: (*muscovite*)- *sillimanite* migmatitic zone, Mount-Grace syncline:

4.4.7.1 P-T path

Pelitic schists of zone E present different textural relationships than pelitic schists from the overlying metamorphic zones, despite similar bulk chemical composition (Table 4.1). A typical specimen from that zone is FG-459 that was collected on the north-dipping flank of Frenchman Cap dome (cf. Fig. 4.2 for location). This specimen is described because it has a very similar bulk chemical composition as specimen JC-223 (Table 4.1), hence the pseudosection of Fig. 4.7a can also be used to decipher its metamorphic evolution. Also, this specimen is the host of a kyanite + garnet leucosome, described in section 4.3.4.6 and illustrated on Fig. 4.4f, that provides information on peak metamorphic conditions reached in zone E. Specimen FG-459 is a rutile-sillimanite-

muscovite-biotite schist with trace amount of resorbed garnet and ragged kyanite. The specimen is divided in microlithons of sillimanite (fibrolite) -rutile and plagioclase-quartz-muscovite. Apart from muscovite that is commonly slightly oblique, all these minerals are aligned parallel to the transposition foliation. Fibrolitic sillimanite forms faserkiesel nodules. Muscovite has a modal volume of ~15-25%, is concentrated in the quartzofeldspathic-rich areas of the thin-section, and is seen to engulf biotite and sillimanite. Highly ragged kyanite is present in trace amount and is surrounded by muscovite and sillimanite.

Combined with the pseudosection presented on Fig. 4.7a, the above textural relationships provide strong constraints on the possible P-T path followed by rocks of zone E. Foremost, the presence of a K-feldspar-bearing leucosome containing large crystals of kyanite and surrounded by a biotite-restite, which contains large euhedral garnet crystals, are clear evidence of partial melting above the muscovite-dehydration reaction accompanied by kyanite and garnet growth. Therefore, the metamorphic peak must have been in the kyanite + melt field that lies between the muscovite-out and biotite-out lines (800-840°C/1000-1300 MPa). This P-T estimate is the same as the estimated baric peak reached by rocks of the two other kyanite-sillimanite migmatitic zones B and D. The implication of this conclusion is that muscovite and sillimanite present in the host schist ought to be retrograde. Textural relationships supporting formation of retrograde muscovite via ReNtr 2 include: 1) its abundance within quartzofeldspathic areas; 2) its occurrence around resorbed garnet and kyanite; 3) the presence of sillimanite and biotite inclusions within muscovite; and 4) its obliquity to the foliation. Furthermore, the absence of garnet and kyanite in specimen FG-459 — a feature common

to all pelitic schists of this metamorphic zone — implies that retrogression in the sillimanite field by ReNtr 2 proceeded to near-completion because phase equilibria modeling predicts that more than 20% garnet should have been present at the metamorphic peak. Zone E is thus composed entirely of retrograde products, in contrast to pelitic schists of the structurally overlying metamorphic zones that largely preserve their peak mineral assemblage. Fig. 4.7c shows the retrograde P-T path suggested for zone E. Thermal and baric peak conditions were reached in the kyanite + melt field, followed by cooling and decompression in the rutile + sillimanite field above the muscovite-out reaction, and then by cooling and decompression to the andalusite field. This path is similar to the two other paths reconstructed for zone B and C; what differs is the timing.

4.4.7.2 Timing constraints

Excellent timing constraints for Zone E and the upper basement level are provided by monazite dates from 11 pelitic schists as well as crystallization ages from four pegmatites and one leucosome dated by Crowley and Parrish (1999). Monazite U-Pb dates range in age from ~60 to 49 Ma with a clear predominance of a ~52-49 Ma population. The last increment of northeast-vergent ductile strain is tightly bracketed between 50.5 ± 1.0 Ma — the crystallization age of deformed sillimanite-bearing leucosome specimen JC-258 collected on the west flank of the dome— and 49.5 ± 0.5 Ma — the age of the undeformed and discordant pegmatite specimens collected within the upper basement level (Fig. 4.2; Crowley and Parrish, 1999). These data suggest a similar timing of prograde metamorphism for zone E as for the three other zones of the Monashee Complex, but clearly indicate that deformation and high-temperature metamorphism in zone E

outlasted that in the other zones.

Hence, migmatites from the upper and lower kyanite-sillimanite zones B and D were rapidly cooled and exhumed after reaching their peak metamorphic conditions, whereas migmatites from the lower sillimanite zone E remained at high temperature and medium pressure until ~51 Ma, which allowed more time for ReNtr 1 and ReNtr 2 to proceed and almost totally remove traces of the peak assemblage. On Fig. 4.7d, the ~51 Ma age was tentatively assigned to the point at which the retrograde P-T path crossed the solidus, but crystallization of leucosome JC-258 could have occurred anywhere in the sillimanite + melt field. The conclusion that rocks of zone E were cooled and exhumed ~4-5 my after the rocks of the overlying zones B-D has important tectonic implications that are discussed below.

4.5 Rates of burial and exhumation

The detailed P-T-t path for specimen JC223 allows the determination of rates of burial/heating and exhumation/cooling. Pressure was converted into depth by assuming an average density of 2900 kg/m^3 , corresponding to the calculated density of K-feldspar-sillimanite-garnet-biotite schist (Table 4.2), in agreement with average upper crustal density modeled from the Bouguer anomaly by Lowe and Ranalli (1993). Uncertainties on pressures and temperatures were derived directly on the pseudosection of Fig. 4.7a from the size of the P-T fields along the P-T path. Uncertainties on ages were considered at the two-sigma level and uncertainties of pressure to depth conversion were taken for densities between 2700 kg/m^3 to 3000 kg/m^3 . Average burial rates of 1 km/my (min: 0.1 km/my; max: 1.4 km/my) and heating rates of 10°C/my (min: 6°C/my ; max: 12°C/my) characterize the prograde path. Geological rates changed drastically at ~59 Ma,

immediately following partial melting. Calculated exhumation (vertical) rates yield average of 7.7 km/my (min: 2.9 km/my; max: 16.5 km/my) and cooling rates of 57°C/my (min: 20°C/my; max: 124°C/my).

Exhumation rates documented in the current study are on the upper end of exhumation velocity range documented elsewhere. The rates are slower than those of some ultra-high-pressure (UHP) rocks from Papua New Guinea (~17 km/my; Baldwin et al., 2004), Pakistan (>30 km/my; Parrish et al., 2006) and the Alpine Dora Maira (16-34 km/my; Rubatto and Hermann, 2001); in the same range as exhumed UHP rocks of Norway (2-8 km/my; Labrousse et al., 2004) and as the Saxon granulites of Germany (~10 km/my; Romer and Roetzler, 2001); but faster than exhumed high-grade rocks from the Variscan Vosges Mountains granulites (3-4 km/my; Schulmann et al., 2002), and the classic Barrovian sequence of the Scottish and Irish Grampian Orogenic belt (>3 km/my; Oliver et al., 2000). It is interesting to note that extrusion models have been suggested for many of the above examples, often to reconcile fast exhumation rates coupled with synconvergent extension.

4.6 Mechanical tectonic model of fold nappe formation

4.6.1 Parameters

The P-T-t paths constructed in this chapter combined with the extensive database of geophysical, structural, metamorphic and geochronologic information available for the southeastern Canadian Cordillera allow the construction of the quantitative mechanical tectonic model presented in Fig. 4.9. The model is not based on automated thermomechanical forward modeling, but rather consists of a series of cross-sections that show the evolution of particles in time and space in an isostatically balanced lithosphere.

Calculations were done on an Excel spreadsheet, and the results were drawn on isostatically-balanced cross-sections. Only the areas surrounding the Monashee Complex are shown in Fig. 4.9 although the calculations include a much larger area.

The first step was to determine the appropriate geometry and density distribution within the crust of the southern Canadian Cordillera at a latitude of $N51^{\circ}$ today. Many features of the model are derived from the extensive database of geophysical data that was acquired through the Lithoprobe project in the 80's and 90's (e.g. Cook et al., 1992; Varsek and Cook, 1994; Buriannyk and Kanesewich, 1995; Buriannyk et al., 1997; Hyndman and Lewis, 1999). First, the Moho is relatively flat with a minor increase from a depth of ~ 32 km depth below the Intermontane Belt to the west, to ~ 35 km below the Monashee Complex ~ 150 km further east (Hyndman and Lewis, 1999). The eastern margin of the Monashee Complex marks the start of a rapid change of Moho depth that thickens from 35 to 50 km depth over a horizontal distance of ~ 50 km. Second, Hyndman and Lewis (1999) demonstrated that this Moho depth gradient coincides with a transition from a hot Cordilleran crust that has a thin mantle lithosphere (~ 20 km thick); to a cold craton that has a thick mantle lithosphere (in excess of 100 km). A mantle lithosphere thickness of 20 km beneath the Intermontane Belt and Monashee Complex was used for the initial density and thickness calculations. A mantle lithosphere thickness of 45 km was used for the mechanical model because it was likely thicker at the orogenic climax, but not too thick so that isostatic compensation would have precluded the high elevations that have been reported by various studies (e.g. Mulch et al., 2006). For the eastern part of the model, a mantle lithosphere thickness of 73.3 km is used to reflect the intermediate position of this area between the thin orogen and the thick craton. Third,

wide-angle seismic refraction profiles indicate that the bottom 12-15 km of the crust has a mafic composition because this layer is characterized by high P-wave velocities (Burianyk and Kanasewich, 1995; Zelt and White, 1995). A typical basalt density of 3000 kg/m^3 is used for this layer. Topographic elevations used to calculate density and thickness vary from an elevation of $\sim 1200 \text{ m}$ in the relatively flat Intermontane Belt (average elevation taken from Tribe, 2005 and Flück et al, 2003), to $\sim 1600 \text{ m}$ in the Monashee Complex (average elevation from Mulch et al., 2007), to $\sim 1800 \text{ m}$ in the Foreland Belt (Flück et al, 2003), to a base level of $\sim 1 \text{ km}$ in the Alberta Foreland basin. In the model, the thickness of the North-American cratonic basement thins from 20 km below the Foreland Fold Belt to less than 10 km below the Intermontane Belt, based on the geological cross-section of the southeastern Canadian Cordillera at a latitude of $N51^\circ$ (cf. Fig. 3.3, p. 178). The basement “bulge” at the Monashee Complex is not considered a primary geometry but was included for the initial density and thickness calculations because it is the geometry today. Densities for the Monashee orthogneissic basement, for the sillimanite-migmatites of the LSA, and for the kyanite-migmatites of the Monashee Complex were calculated with the program *Perple_X'06* for appropriate bulk chemical compositions and peak P-T conditions (Table 4.2). Mantle densities throughout the evolution of the model were adjusted to the temperature at half the thickness of a given mantle layer by using a coefficient of thermal expansion of $3.4 \times 10^{-5} \text{ K}^{-1}$ (see Afonso et al. 2005 for justification). The geotherm for the mantle lithosphere is assumed to be a linear variation between the temperature at the base of the crust and at the top of the asthenosphere. The geotherm for the crust today is based on the geotherm presented in Currie and Hyndman (2006), whereas the geotherm for the crust between 80-50 Ma is

derived from baric peak conditions derived in this chapter (Table 4.3). The geotherm for the asthenosphere is derived from the mantle adiabat presented in Currie and Hyndman (2006; Table 4.3). All the above data were used to calculate the average density of the USA, as well as the thickness of the crust and lithospheric mantle underneath the craton. It was found that a USA density of 2890 kg/m^3 , a cratonic crustal thickness of 50 km, and a mantle lithosphere thickness of 130 km beneath the craton, would maintain the crust in isostatic equilibrium while reproducing the observed average elevations of $\sim 1200 \text{ m}$ and $\sim 1600 \text{ m}$ for the Intermontane Belt and the Monashee Complex, respectively.

The next step was to determine the convergence and denudation rate. The P-T-t path for rocks of zone B indicates they were underthrust from a depth of $\sim 26 \text{ km}$ to a depth of $\sim 39 \text{ km}$ between ~ 75 and $\sim 59 \text{ Ma}$ (assuming an average density of 2900 kg/m^3). Assuming a constant dip of 30° , this is equivalent to a horizontal distance of 21.3 km , or to a convergence rate of 1.3 km/my . This convergence rate was kept constant until the last stage of the model at $\sim 50 \text{ Ma}$ based on the interpretation presented in chapter 2 that the $\sim 50 \text{ Ma}$ deformation recorded at the deepest structural level of Frenchman Cap dome was contractional. In the model, the upper plate is fixed relative to the underthrusting lower plate. Horizontal movements are depicted with the use of a vertical black bar that remains pinned to the underthrusting basement and moves relative to the frame of the model.

This model uses a similar, but simpler method of denudation by slope-dependant erosion as presented in Beaumont et al. (2004) and involves solving the following equation between each time step:

$$\text{denudation} = \int_{t_1}^{t_2} f * g * (\text{slope}_1 + \text{average slope rate change}) dt \quad (\text{km})$$

where f is the erosion factor (km/my) and g is a measure of the spatial variation of aridity. The eastward : westward ratio for the aridity factor g varied from 2:1 between 80 and 59 Ma, to 1:1 between 59 and 55 Ma, to 2:3 between 55 and 50 Ma. These ratios were chosen to reflect progressively drier winds coming from the east as the Western Interior seaway regressed in the Paleogene (Catuneanu et al., 2000). Continentward winds were suggested by the high-resolution climate modeling of the North American Laramide foreland in the Early-Paleogene (Sewall and Sloan, 2006). In contrast, the erosion factor f was increased from 15 to 40 km/my at 55 Ma to reflect the increase in mean annual temperature during the warmest period recorded in the Phanerozoic that should have brought more humidity in the air by ocean evaporation (Wilf, 2000; Sewall and Sloan, 2006). Denudation calculated for the western and eastern parts of the model were added to obtain the total amount of denudation. Three parts of the model illustrated on Fig. 4.9 were denuded: a western part, centred on the LSa; a middle part, centred on the Monashee Complex; and an eastern part, centred on the eastern USA. Westward slopes were calculated as the difference in elevation relative to the Intermontane Belt divided by the horizontal distance between the Intermontane Belt and the LSa for the western part of the model, estimated at 125 km; or the horizontal distance between the Intermontane Belt and the Monashee basement culmination for the middle part, estimated at 50 km. Denudation in between these areas was visually interpolated. Eastward slopes were calculated as the difference in elevation relative to the craton divided by the horizontal distance between the flat craton and the eastern USA, estimated at 125 km; or the horizontal distance between the flat craton and the Monashee Complex, estimated at 200 km. These parameters are not only compatible with what is known about

paleoclimate, but more importantly; best matched the P-T-t paths presented above.

Denudation rates varied from 0.1 to 1.4 km/my, which are reasonable if compared to the fastest denudation rates recorded on Earth today in the Himalayan Nanga Parbat that are in the order of 2 to 10 km/my (Burbank, 2002).

4.6.2 *Model description*

The mechanical model presented on Fig. 4.9 illustrates how fold nappes can form during sequential ductile extrusion of a partially molten channel. The model consists of five cross-sections that each represents snap-shots of the evolving system at 80, 59, 57, 55 and 50 Ma. At each time-step, the geometry that best matches the data was drawn, while the MOHO depth and thickness of the upper layer were adjusted to maintain most of the crust in isostatic equilibrium. The part of the crust that is going to flow as a channel by the next time-step— and thus leave its lid and base behind— is shown in pale red for clarity. Stations labelled A-E on Fig. 4.9 refer to the five metamorphic zones described in this chapter. It is important to stress that the term “ductile extrusion” is used herein as a process by which a slice of deformable rock material is detached from its base and lid and flows towards the surface of the Earth, without necessarily reaching it. The upward translation of the slice is accommodated by ductile shear zones of opposite shear sense and by internal ductile deformation. This definition has no genetic implications with regards to the driving force; which could be: 1) a lateral pressure gradient induced by a topographic gradient or density contrast between two rock columns; 2) active erosion at the orogenic front; 3) convergence of two blocks that expels the rock mass between them; or 4) any combination of the above.

The ~80 Ma time-step (Fig. 4.9a), is characterized by a ~12 km thick channel within

rocks of the L_{Sa} that had started to be extruded above the underthrusting craton. Station A was placed at ~30 km depth to reflect the baric peak of ~800-900 MPa derived from the P-T-t path presented in Fig. 3.6 (p. 184). As the channel tunnelled its way above the ramp, rocks at the top deformed by top-to-the-west shearing, whereas rocks at the base deformed by top-to-the-east shearing. Evidence for this kinematic reversal is observed along the Okanagan Valley fault system and the Monashee décollement at the top and bottom of the L_{Sa}, respectively. Stratigraphic studies summarized in section 4.3.2 indicate that stations B, C, D and E are part of the same stratigraphic layering that had not yet been folded at ~80 Ma. Hence they lay on one inclined plane above an underthrusting basement. Station B was located at a depth of ~24 km on its way to entering the kyanite field (at ~750 MPa), which was reached at ~75 Ma from the P-T-t path presented on Fig. 4.7a.

The next time-step presents the modeled crust at ~59 Ma (Fig. 4.9b), for which precise depth-time points exist for stations A, B, C and D. Based on the P-T-t path presented on Fig. 3.6 (p. 184), rocks from station A were exhumed to <14 km depth by ~59 Ma. In contrast, part of this basement and its cover sequence, which now form the Monashee Complex, had been buried to more than 35 km depth and were being partially melted. Based on the P-T-t paths reconstructed in this chapter, station B and D were at their baric peak of ~39 km (or ~1100 MPa) in the kyanite + melt field, while station C was just crossing the kyanite-in reaction at a depth of ~25 km further up the underthrust basement ramp. Fig. 4.9b shows the presence of a 5-8 km thick channel within which incipient fold nappes were being formed. Extensive partial melting recorded in rocks from zones B and D is inferred to have lowered their viscosity enough

to allow their flow down the pressure gradient produced by the combined effect of a higher surface elevation and the presence of denser rocks to the west of the model (see below).

There are fewer quantitative data for the ~57 Ma section, but this time-step of the model is important to illustrate the process that led to the final assembly of the two fold nappes. At ~57 Ma, station C was at its baric peak of ~900 MPa and most likely started to be exhumed (Fig. 4.7c). P-T-t paths for station B and D both indicate a rapid exhumation path between ~59 and ~55 Ma. Furthermore, the kyanite-migmatites of station B and D are part of two basement-cored fold nappe anticlines located above and below muscovite-kyanite schist of station C. A possible process that could lead to this geometry while complying with the P-T-t paths is shown on Fig. 4.9b&c. At ~59 Ma, the incipient Sybley fold nappe (station B), located at the front of the channel, started to be extruded above the ramp — and thus above station C — while the incipient Mount-Grace fold nappe (station D) was still flowing horizontally at the rear of the channel. It will be demonstrated in section 4.7 that the flow velocity at the front of the channel should have been reducing upon cooling and melt crystallization that accompanied the decompression path documented on Fig. 4.7a. In contrast, the rear of the channel was still hot and flowing at a high velocity until it got stuck against the decelerating front of the channel. At this point, rocks located at the rear of the channel could have thickened, or continued to flow above or below the front of the channel. The path of least resistance would have been to flow below station C because of a shallower slope. Hence, rocks from station C were engulfed within a flowing channel. Eventually, rocks from above (station B) and below (station C) reached the same velocity and all three stations flowed together toward the top of the

ramp.

The next time-step at ~55 Ma marks the end of ductile deformation and cooling for most of the two fold nappes of the Monashee Complex. P-T-t paths presented in this contribution indicate that stations B and D were exhumed to the andalusite field at less than ~14 km depth by ~55 Ma. Thus, between time-steps ~57 and ~55 Ma, station D was extruded beneath station C and then stations B, C and D were extruded together. At this point, the Sybley fold nappe was entirely formed, station B was overlying station C, and both were located below station A (Fig. 4.9d).

The last time-step shows the modeled crust at ~50 Ma during the waning stage of ductile strain recorded in the southeastern Canadian Cordillera. After ~55 Ma, flow and ductile extrusion was concentrated within the deepest and hottest structural level, below station D. The ~50.5 Ma age of crystallization for sillimanite-bearing leucosome JC-258; the numerous ~50 Ma monazite dates from schists collected within the lower part of the Mount-Grace fold nappe and within the upper basement level; and the extensive retrogression textures observed within the lower sillimanite-migmatitic zone E attest that rocks from this zone remained at high temperature longer than rocks from above and that newly underthrust basement rocks were heated above ~600°C. Formation of the Mount-Grace syncline is interpreted to have resulted from the drag induced by the extrusion of the underlying migmatitic felsic orthogneiss, which now forms the upper basement level that was less dense than the overlying garnet pelitic schists, and was thus flowing faster. The inferred lower temperature for this extrusion phase would have been associated with a higher viscosity, which in turn would have resulted in slower extrusion than previous phases. After the channel locked up because of the high viscosity of

crystallized coarse-grained migmatitic schists, the last increments of contraction were potentially accommodated by large amplitude folding that would have produced domes (cf. chapter 2). It is noteworthy that channel locking followed by large amplitude folding has also been document after ductile extrusion of the Great Himalayan Sequence (Godin et al., 2006a).

4.7 Driving forces for extrusion

In order to be viable, a mechanical model has to satisfy physical laws to the first-order. For upward motion by ductile extrusion to occur, the positive (updip) force exerted by the lateral pressure gradient (F_{IP}) must overcome the resistance to detachment that is approximately given by the integral of the critical shear stress along the crust slice (F_{τ} ; statement modified from Ranalli et al. 2005, p.169). In order to exhume a rock slice by extrusion to less than ~15 km depth, as suggested by the P-T-t paths presented herein, the up-dip resulting force must have remained positive throughout the entire extrusion process. That is:

$$(eq.1) F_{IP} > F_{\tau}$$

at any given depth during extrusion. In the model illustrated on Fig. 4.9a, the lateral pressure gradient at the base of the basement ramp (at location D) is caused in part by the topographic gradient of ~600m, and in part by the density difference of ~100 kg/m³ between an orogenic wedge consisting of garnet schists and metavolcanic arc rocks (~2900 kg/m³) and underlying felsic orthogneissic basement (~2800 kg/m³). The lateral pressure gradient can be integrated over the vertical thickness of the inclined slice to obtain the up-dip component of the lateral force F_{IP} by the equation:

$$(eq.2) F_{IP} = [(\Delta\rho g (z - z_{tb})) + \rho_w g h] \times D/2 \quad (N/m)$$

for which $\Delta\rho = 100 \text{ kg/m}^3$ is the difference in density between a felsic basement and the overlying wedge, $g = 9.8 \text{ m/s}^2$ the acceleration of gravity, z the depth in m, $z_{tb} = 11\,000 \text{ m}$ the depth at the top of the basement ramp, $\rho_w = 2900 \text{ kg/m}^3$ the average density of the wedge, h the difference in elevation, and D the thickness of the extruding slice = 2000 m derived from the thickness of the Mount-Grace fold nappe (Fig. 4.3). Integration of the shear stress acting at the base of the extruding slice yields an approximation of the down-dip force F_τ acting at the base of the slice. According to Medvedev (2002):

$$(eq.3) \quad F_\tau \sim VL\eta/2Z_{avw} \quad (\text{N/m})$$

for which V is the underthrusting velocity in m/s, L is the length of the extruding slice in m, η is the effective viscosity in Pa s, Z_{avw} is the average thickness of the wedge at a given depth in m. V can be estimated from the burial rate of specimen JC-223 derived from the P-T-t path presented on Fig. 4.7a, which is equivalent to $5.55 \text{ E} -11 \text{ m/s}$ assuming underthrusting along a 30° slope. There is no constraint on L and a value of 30 000 m is assumed from the length of each fold nappe on the reconstruction presented on Fig. 4.9d. The effective viscosity η is a complex function that depends on shear stress (σ_s), strain rate ($d\epsilon_s/dt$), rheological parameters (A , Q) and temperature (T). For a power-law material deforming in simple shear (cf. Ranalli, 1995):

$$(eq.4) \quad \eta = \sigma_s / (2 \, d\epsilon_s/dt) \quad (\text{Pa s})$$

Evidence presented in this contribution indicates that the rocks were partially molten during their extrusion. Rutter et al. (2006) derived a flow law for synthetic partially molten granite:

$$(eq.5) \quad d\epsilon_s/dt = \sigma^n A \exp(B \Phi^3) \exp(-Q/RT) \, (\text{s}^{-1})$$

for which $d\epsilon_s/dt$ is the strain rate in s^{-1} , $n = 1.8$ is the power-law exponent, $A = 6.46E-13$ $Pa^{-1.8} s^{-1}$, $B = 192$, Φ is the melt fraction, $Q = 2.2 E+05$ J/mol , R is the gas constant, T is the temperature (K). To get an estimate of the viscosity, it is necessary to fix the value of σ_s or $d\epsilon_s/dt$. It is difficult to estimate the stress, but geological rates are commonly estimated to be in the order of $1E-15 s^{-1}$, $1E-13 s^{-1}$ (Ranalli, 1995; Baxter and DePaolo, 2004). By fixing $d\epsilon_s/dt$, solving for σ in (eq.5), and replacing σ in eq. 4 it is possible to calculate η as a function of Φ and T . Visual estimates of ~30 volume % leucosome in specimen JC-223 are consistent with the predicted volume of melt calculated by phase equilibria modeling at thermal peak conditions. Assuming linear melt crystallization during cooling from thermal peak conditions at ~800°C/1100 MPa to crossing of the wet solidus at ~660°C/300 MPa (P-T-t path of Fig. 4.7a), it is possible to calculate Φ as a function of T , which in turn is function of the depth z (eq.2) by assuming an average wedge density of 2900 kg/m^3 .

Fig. 4.10 presents results of calculations conducted with the parameters described above for different strain rates. For comparison, calculations were also conducted with the quartzite rheological parameters of Hirth et al. (2001). On Fig. 4.10, the line with open circles depicts the variation with depth of the up-dip force resulting from the lateral pressure gradient (F_{IP}); lines with filled symbols and crosses depict the down-dip force resulting from shear stress (F_τ) at the base of the extruding slice composed of partially molten rocks or quartzite, respectively, for different strain rates. It is assumed that extrusion stops when the down-dip force equals the up-dip force ($F_{IP} = F_\tau$), which is represented on Fig. 4.10 by the points of intersection of the lines with filled symbols and crosses, with the line with open circles. These results indicate that, at the baric peak, the

lateral pressure gradient exerts an up-dip integrated force on the slice two orders of magnitude higher than the down-dip force exerted by shear stress integrated over the length of the partially molten slice. This significant up-dip resulting force should have driven extrusion up the ramp until the resulting force became negative. For the parameters used herein, migmatites deforming at a strain rate of $1\text{E}^{-14}\text{ s}^{-1}$ could have been exhumed to a depth of ~ 17 km by extrusion, whereas quartzite could only have been exhumed to a depth of ~ 25 km. At a faster strain rate of $1\text{E}^{-13}\text{ s}^{-1}$, which is probably more realistic for partially molten rocks (cf. Albritton et al., 2005), both migmatites and quartzite could be exhumed to a depth of ~ 12 km by extrusion. Notice that extrusion is favoured for a shorter extruding slice, and that a 50 km long slice of partially molten rocks would require a strain rate of $1\text{E}^{-13}\text{ s}^{-1}$ to be exhumed to less than 15 km depth by extrusion. Although the actual knowledge on migmatite rheology is still limited, the quartzite rheological parameters used herein are much more compelling because there is a good agreement between recent deformation experiments. The parameters are also compatible with microstructural studies of natural specimens (Tullis, 2002). Deformation experiments conducted on mica-rich rocks point to extremely low yield strength at mid-crustal conditions (Mariani et al., 2006). Therefore, I consider that the conclusions presented in this paragraph are robust because mica-rich partially molten rocks ought to be softer than quartzite, for which exhumation by extrusion is possible within uncertainty of the numerous parameters used herein. The model presented on Fig. 4.9 is thus physically viable to a first order.

4.8 Kinematics of fold nappe formation

Up to this section, I proposed a model of ductile extrusion during basement

underthrusting to explain how rocks from five different locations with contrasted P-T-t paths can be juxtaposed over each other. In this section I will argue that the fold nappes of the Monashee Complex are large-scale drag folds created by the internal density contrast within the channel. Williams et al. (2006) discussed the kinematics of drag fold formation during noncoaxial strain. Building on the experimental results of Bons and Cox (1994), they showed how initial open folds develop during simple shear and then amplify into asymmetric drag folds with high shear strain. Williams et al. (2006) further argued that shear zone thickening favours the formation of drag folds, whereas thinning hinders it. Evidence for extensive partial melting observed in rocks of the Monashee Complex implies that expansion took place to accommodate density reduction produced by partial melting (cf. Powell et al., 2005). This melting-induced positive volume change during underthrusting is equivalent to thickening during shearing; it should thus favour the formation of drag folds. Furthermore, the anticlines of the fold nappes are cored by felsic orthogneiss that have a maximum crustal density of $\sim 2720 \text{ kg/m}^3$, which is much lower than the density of the surrounding kyanite-garnet pelitic schist calculated at $\sim 3050\text{--}3125 \text{ kg/m}^3$ with the software *Perple_X* (cf. Table 4.2). The internal density contrast within the extruding slice creates a vertical stress that can be integrated over the thickness of the basement and multiplied by the sine of the slope to get the up-dip component of the buoyancy force F_g by:

$$\text{(eq.6)} \quad F_g = \frac{1}{2} * \rho g D_b^2 \sin 30^\circ$$

for which $\rho = 350 \text{ kg/m}^3$ is the density contrast between basement rocks comprising 70% of partially molten felsic orthogneiss and 30% of amphibolite; $D_b = 500 \text{ m}$ the approximate thickness of basement rocks within the Mount-Grace fold nappe (Fig. 4.3).

The resulting force of $2.4\text{E}+08$ N/m is significant (compare with Fig. 4.10) and should drive a faster flow of partially molten basement compared to surrounding partially molten garnet-kyanite schist. In such a gravitational instability, any perturbations should rapidly amplify into a basement-cored drag fold that would continue to amplify once formed.

Fig. 4.11 illustrates three stages expected during the formation of a fold nappe by buoyancy-driven differential flow. The faster flow of the partially molten orthogneiss would produce a kinematic reversal with rocks below deforming by reverse ductile shearing and rocks above deforming by normal ductile shearing. This is the strain pattern observed within the Sybley fold nappe that is dominated by reverse shearing except at its top part — the Anstey Range detachment — where normal shearing dominates.

4.9 Discussion

4.9.1 *Phase equilibria modeling*

In addition to providing important tectonic constraints on the formation of fold nappes, this study documents a previously unrecognized behaviour for partially molten rocks. To my knowledge, this study is the first to document intersection of garnet isopleths in migmatites. At peak temperature $>800^{\circ}\text{C}$, compositional profiles of minerals are usually homogenized and preservation of prograde zoning in garnet is usually interpreted as evidence of a short residence time at high temperature (Spear, 1993).

Furthermore, many studies from other migmatitic terrains documented that migmatites commonly evolve as open systems that progressively lose melt, which lead to the preservation of peak metamorphic assemblages (White and Powell, 2002; Brown, 2007; Indares et al., 2008). This chapter rather points to limited melt loss from metapelites of zones B and D. First, the isopleths intersection calculated by forward

modeling for the garnet core analysis with the highest X_{sps} in specimen JC-223 (Fig. 4.7a) would not be possible if the bulk composition had been significantly modified by melt loss. Second, peak metamorphic assemblages are preserved in migmatites of zones B and D, whereas peak metamorphic assemblages are entirely retrogressed in migmatites of zone E. Because the P-T-t paths presented herein indicate that former were cooled and decompressed in less than ~5 my, whereas the latter remained at high temperature ~5 my longer, it is much likely the extent of retrogression was related to different exhumation rates and not to different volume of melt removed from the system as suggested by White and Powell (2002). This proposition is supported by the macroscopic textures of the migmatites of the Monashee Complex that shows a relatively even leucosome distribution at the hand-specimen scale (Fig. 4.4a-c), rather than pooling of melt in dilatant sites as generally observed in migmatitic terrains (e.g. Gervais et al., 2004; Brown, 2007). Therefore, this study suggests that migmatites can, under certain circumstances, behave as closed systems.

4.9.2 *Isograds to locate cryptic structural boundaries*

Conclusions presented in this study have important implications for the understanding of isograds in migmatitic terrains. The definition of "isograd" given in Glossary of Geology (Jackson, 1997; p.337) is: "A line on a map at which metamorphism proceeded at similar values of pressure and temperature as indicated by rocks belonging to the same metamorphic facies. Such a line represents the intersection of an inclined surface with the Earth's surface corresponding to the boundary between two contiguous facies....". According to Carmichael (1978) such a common definition of isograd is unfortunate because it carries a genetic connotation. Indeed, this definition tacitly implies

several assumptions including: 1) that the entire body of rocks considered was metamorphosed as a coherent block; and 2) that quenching occurred at the same point of a P-T path, presumably at the thermal peak. It has been a common practice to use this definition in studies of metamorphic terrains and the study of Journeay (1986) in Frenchman Cap dome is one example. Hence, the sequence of kyanite-muscovite in zone C, to kyanite-sillimanite in zone D, to sillimanite in zone E was interpreted by Journeay (1986) as a typical Barrovian gradient with the kyanite-sillimanite reaction crossed at subsolidus conditions below the muscovite-dehydration melting reaction (cf. pseudosection of Fig. 4.7a as a template for a typical aluminous schist petrogenetic grid). In order to comply with the isograd definition presented above and their inherent implications, Journeay (1986) gave more weight to the presence of muscovite in zones D and E than to the presence of K-feldspar in these zones. Furthermore, Journeay (1986) assigned the abundance of K-feldspar and the common presence of sillimanite and andalusite in zone B, as evidence for a second, low-pressure-high-temperature, metamorphic event related to thrusting of the overlying Selkirk allochthon (redefined herein as the LSa). Gibson et al. (1999) followed by Crowley and Parrish (1999) and Foster et al. (2004), proved that this latter interpretation is wrong because there is no geochronologic evidence for a second high-temperature metamorphic event in zone B. On the contrary, the oldest U-Pb monazite ages were obtained within zone B that was supposed to have been overprinted by a younger high-temperature low-pressure event.

By considering the effect of retrogression and cooling rates, as done herein, the “isograds” traced on Fig. 4.2 do not violate any metamorphic nor geochronologic data. This is true as long as one considers muscovite from zone D and E as a retrograde phase.

This consideration is supported by textural relationships, by the presence of K-feldspar in many of these rocks, and is confirmed by the presence of the kyanite-garnet-K-feldspar leucosome (specimen FG-459) in a kyanite-free, sillimanite-muscovite migmatitic schist typical of zone E (Fig. 4.4e). The model presented in this contribution suggests that the three isograds from the study area (Fig. 4.2) are not isograds as defined above, but rather structural boundaries between fold nappes extruded at different time. Rather than the genetic definition discussed above, Carmichael (1978; footnote 2 on p.337) proposed a more descriptive definition of "isograd": "An isograd is the trace of a surface across which a specific change in metamorphic mineralogy takes place (in rocks of comparable bulk composition). Such an isograd is a mappable line whose physical significance, in terms of P, T and other externally or internally controlled parameters remains open to discussion." Isograds of Frenchman Cap dome correspond to this descriptive definition. This study thus highlights the importance of mapping isograds in the sense of Carmichael (1978) for they could be the only observable evidence for structural breaks.

4.9.3 *Comparison between Thor-Odin and Frenchman Cap domes*

The model of fold nappe formation by ductile extrusion above an underthrust basement is the first model that reconciles all data from the multidisciplinary database available at the latitude of Frenchman Cap dome, but is this model able to incorporate data at the latitude of Thor-Odin dome (see Fig. 4.1 for location) as well? Frenchman Cap dome (FCD) and Thor-Odin dome (TOD) have differences and similarities that must be reconciled for a model to be viable. Differences include: 1) basement of FCD preserves a Proterozoic fabric (Crowley et al, 2001; Crowley et al., 2008; chapter 2 of this thesis), whereas the Proterozoic fabric in TOD basement is apparently completely transposed

(see Williams and Jiang, 2005; Hinchey et al., 2006 and references therein); 2) there are more infolds of basement and cover in TOD than in FCD (compare Fig. 4.3 with Fig. 4 of Williams and Jiang, 2005); 3) the volume of Cordilleran migmatite in the core zone of Thor-Odin is higher, as evidenced by the presence of map-scale bodies of diatexite (Vanderhaeghe et al., 1999); 4) the transposition fabric and the enveloping surface of map-scale folds are steeper in TOD than FCD (compare Fig. 4.3 with Fig. 4 in Williams and Jiang, 2005); 5) part of TOD was isothermally decompressed whereas the P-T-t paths from FCD indicate cooling during decompression (Norlander et al., 2002 vs. this study). Similarities between both domes include: 6) a similar general lithostratigraphic sequence consisting of basement rocks having a North-American affinity overlain by a cover sequence characterized by a thick quartzite at its base (Reesor and Moore, 1971; Brown, 1980; Armstrong et al., 1991; Parkinson et al., 1991; Crowley, 1999); 7) a basal quartzite that has been mapped continuously between the two domes, implying that there is therefore no major structural discontinuity between them (Read, 1980), and both domes are at approximately the same structural level of exposure; 8) a similar structural evolution characterized by fold nappes of extensive continuity along strike and three generations of variably transposed, north to northeast-verging, asymmetric minor folds (e.g. Journeay, 1986 vs. Williams and Jiang, 2005); 9) a major normal shear zone — that started its motion at a high metamorphic grade — is present at the contact with the overlying LSA in both domes (Johnston et al., 2000; Kruse and Williams, 2007; this study); 10) last and foremost, according to the P-T-t-d paths constructed by Hinchey et al. (2006) for TOD and in this study for FCD, both domes reached their baric peak between ~60 and 56 Ma and were then rapidly exhumed to upper crustal conditions before ~55-52

Ma while they were deforming by foreland-directed ductile shearing. The last point partially conflicts with one of the conclusions presented by Hinchey (2005) who suggested, based on the P-T path of Journeay (1986) for FCD, that the domes have different tectonic histories because of their distinct P-T-t paths. In contrast, results of this study point to similar P-T conditions at baric peak as well as a similar timing and magnitude of synshearing decompression for both domes. The main difference is that rocks of FCD cooled as they decompressed, whereas rocks of TOD remained hot.

I propose that a model of ductile extrusion of a partially molten channel, which would have been thicker than the 2 km thick fold nappes of FCD, above an underthrusting basement could explain most of the differences between the domes. According to this model, differences between TOD and FCD would originate from a different basement composition. Indeed, TOD basement is formed by a high volume of paragneiss, whereas FCD is mainly cored by orthogneiss. Once underthrust to anatexis conditions, a thicker channel would have developed in TOD because of the higher fertility for partial melting of paragneiss. Accordingly, the lower viscosity of partially molten paragneiss compared to orthogneiss, would have resulted in a better deformation coupling with cover sequence rocks. In addition to providing an explanation for the more abundant infolds in TOD, this would have resulted in the development of a thicker channel. In turn, a thicker channel implies slower cooling by conduction for rocks located in the centre of the channel that would thus have followed a decompression path having a steeper P-T slope than a thinner channel such as in FCD. The discrepancy between the isothermal decompression path documented in TOD (Norlander et al., 2002) and the hairpin path documented herein could thus be related to a thickness difference between

two extruding channels. Furthermore, Norlander et al. (2002) documented that rocks of TOD crossed the two temperature-sensitive muscovite and biotite dehydration melting reactions (PrNtr 1) and (PrNtr 2), as well as the pressure-sensitive biotite dehydration melting reaction:

Hence, rocks of TOD crossed three melting reactions whereas rocks of FCD only crossed the first two. This additional melting reaction is the most probable explanation for the larger volume of Cordilleran leucosome documented in the core of TOD compared to FCD (Vanderhaeghe et al., 1999; Kruse, 2007). Accordingly, a larger volume of Cordilleran leucosome would have been produced in TOD. There would thus have been a feedback loop operating in TOD in which a thicker channel would have favoured isothermal decompression that would have, in turn, contributed to increasing the thickness of the channel. It is interesting to note that the diatexite body mapped by Vanderhaeghe et al. (1999) is located in the core of a northeast-verging fold nappe with a steep axial surface (cf. Fig. 5 in Hinchey et al., 2006). Williams and Jiang (2005) attributed the steep antiformal attitude of the dome to upright, north-south trending folding of the axial surfaces of the fold nappes and associated transposition fabrics. Chapter 2 of this thesis also reported the presence of upright, north-south trending folds associated with axial-planar hornblende-bearing leucosomes in the core of FCD. Blattner (1971) reported the presence of very similar hornblende-bearing leucosome associated with upright folds in the core zone of TOD. It may well be that the diatexite bodies of TOD are larger scale manifestations of the same pattern observed at the outcrop scale. Field investigation of a small area located near the core of a steep and upright map-scale

fold toward the hinge of which diatexite bodies would have flowed, could well be interpreted as a diapir as Vanderhaeghe, Teyssier, Whitney and co-workers have argued (e.g. Vanderhaeghe et al., 1999; Teyssier and Whitney, 2002; Teyssier et al., 2005).

Consequently, a model very similar to the model presented on Fig. 4.9, but involving a thicker section of basement paragneiss, could reconcile all the differences between the domes described in points 1-5 above. If this hypothesis were correct, the base of Cordilleran strain would lie beneath the exposure surface of TOD. Testing this hypothesis would require the construction of detailed P-T-t paths across fold nappes in TOD.

4.9.4 *Fold nappe formation and sequential ductile extrusion: a discussion*

The model of sequential ductile extrusion presented herein better reconciles available data than previously proposed models. Williams and Jiang (2005) argued that most regional high-grade rocks on Earth would belong to what they referred to as the “high-grade-nappe-association” (HGNA). They used structural characteristics of the Monashee Complex as a template to argue that HGNA generally forms by horizontal, crustal-scale shearing in the middle to lower crust. They further proposed that fold nappes, which are common map-scale structures in HGNA, would form by rotation and transposition of early upright folds, or by large-scale drag folding. Similar models of fold nappe formation have been proposed by Hatcher and Hooper (1992; their type-F thrust sheets) and by Krabbendam et al. (1997) for the generation of the Tay Nappe of Scotland. Although the horizontal crustal-scale shearing model elegantly explains the structural characteristics observed in the fold nappes of Frenchman Cap dome, it is not compatible with the P-T-t paths constructed herein. Notably, the HGNA model does not explain the

formation of the Sybley fold nappe by synchronous exhumation of its anticline and burial of its syncline, nor does it explain the pattern of baric peak and cooling ages decreasing with structural depth.

Alternatively, Gibson et al. (1999) combined geochronologic and structural data to derive a model of fold nappe formation that better reconciles the data. Their model involved eastward telescoping of the overturned limb of the Sybley fold nappe by eastward migration of an eastward-tapering orogenic wedge that would progressively bury and deform the cover sequence of the Monashee Complex located further toward the foreland. This model provided a mechanism of fold nappe development by foreland-directed noncoaxial flow while accounting for the diachroneity of metamorphism and deformation that becomes younger with structural depth. However, this model would not have resulted in synchronous burial of the anticline and exhumation of the syncline of a fold nappe (between ~59 and ~55 Ma Fig. 4.7a &b), nor would it provide an explanation for the ~25 km of vertical separation of currently juxtaposed rocks from zone A and B at 59 Ma (Fig. 4.9a). Apparently, the only way to reproduce these features is if rocks of zone B, which now form the Sybley anticline, left their lid and base behind and flowed above the underthrusting rocks of zone C, which now form the Sybley syncline (Fig. 4.9c). Similarly, the presence of zone-E rocks that were still flowing at high temperature ~5 my after exhumation and cooling of the immediately overlying rocks from zone B, C, and D requires that zone-E rocks were extruded beneath rocks of zone D (Fig. 4.9 d & e) in a similar way as zone B, C and D were extruded beneath rocks of zone A (Fig. 4.9 b-d). It is important to stress that these features would be produced by ductile extrusion along an inclined plane irrespective of the denudation rate at the surface. For the

construction of the model presented on Fig. 4.9, these rates were varied until a best fit was obtained, but the overall shape of the P-T-t paths can be reproduced even when no denudation is assumed.

Furthermore, this study indicates that density contrast could play a role in the formation of fold nappes. As discussed in section 4.8, the positive volume-change associated with partial melting combined with the inverted density gradient created during high-pressure metamorphism of a metasedimentary cover sequence underlain by felsic basement orthogneiss should lead to the formation of perturbations that would amplify into drag folds during higher shear strain. Once formed, such a basement-cored fold would flow faster than the surrounding heavier kyanite + garnet –bearing schists and create a kinematic reversal centred on the basement-cored fold (Fig. 4.11). The alternative of transposition of an early-formed upright fold (Rose and Harris, 2000; Williams and Jiang, 2005) is not applicable because the P-T-t paths of zone B and C, which are respectively located on the upright and overturned limb of the Sybley syncline, indicate that they were separated by a vertical distance of ~15 km at ~59 Ma and that the fold nappe formed between ~59 and ~55 Ma. Density contrast has rarely been considered in the kinematic analysis of folds (exception of Bucher, 1963; Ramberg, 1970; Blay et al., 1977; Vacas Peña et al., 2004), but the first order calculations presented in section 4.8 point to non-negligible driving forces that should influence the kinematics of flow as much as viscosity contrast. It is interesting to note that many fold nappes in the world are cored by basement gneiss. Whether this constitutes a cause and effect remains conjectural but would be worth considering in future work.

Other models of fold nappe formation during sequential ductile extrusion have been

presented for different tectonic settings including Andean-type subduction and continental collision. An interesting model of sequential burial and exhumation within a fold nappe formed in a subduction setting was proposed by Boyle et al. (1994) for the evolution of the Sulitjelma fold nappe of the northern Norwegian Caledonides. Similar to the model of Frenchman Cap dome, Boyle et al. suggested that the two parts of the fold nappe were sequentially exhumed when they came under the influence of the underthrusting Baltica basement ramp. In contrast with the southeastern Canadian Cordillera, P-T-t paths for two specimens of the Sulitjelma fold nappe suggest that the foreland-ward specimen was being exhumed while the hinterland-ward specimen was still being buried. The formation of the classical fold nappes of the Pennine Alps has also been ascribed to sequential ductile extrusion. For example, Babist et al. (2006) described a process of systematic sequential burial of foreland-ward nappes synchronous with rapid extrusion of the adjacent hinterland-ward nappes. According to Babist et al. (2006), this process would have been repeated up to eight times between ~70 and ~30 Ma.

Sequential ductile extrusion has also been documented in continent-continent collisional settings. In a comprehensive paper integrating stratigraphic, structural, metamorphic and geochronologic data from the eastern Nepalese Himalaya, Goscombe et al. (2006) presented evidence that the lower part of the Great Himalayan Sequence records high-pressure metamorphism millions of years after the main phase of extrusion of the overlying channel. As Goscombe et al. pointed out, a later phase of extrusion from beneath was required to bring these young, high-pressure, rocks in contact with the overlying and previously exhumed channel. Their model is thus very similar to the model of homogenous extrusion of the thick LSa followed by heterogeneous and sequential

ductile extrusion of the Monashee fold nappes. Another potentially good analogue to this model has been recently documented in the MesoProterozoic Grenville Province of northeastern Quebec by Jordan et al. (2006). Like the Monashee Complex, the Manicouagan area is located in a retro-wedge setting of a large, hot collisional orogen and involved extrusion of remobilized underthrust craton. The Manicouagan Imbricate Zone comprises HP-HT rocks that were extruded from the lower crust beneath rocks of the Hart Jaune Terrane that remained cold and rigid throughout Grenvillian orogeny (Indares and Dunning, 2004). Similar to the Monashee Complex, the footwall of this HP-HT zone is composed of a tonalitic basement overlain by a continental margin sequence including migmatitic kyanite-bearing metapelites that record HP metamorphism ~50 my after the main extrusion phase of overlying rocks. Although not explicitly suggested by Jordan et al. (2006), these data could suggest a later phase of extrusion beneath the main extruded panel, similar to the model documented for Frenchman Cap dome. Consequently, a pattern emerges from the cases described above and others as described, for example, by Hynes (2002): it appears that irrespective of the convergent tectonic settings (retro-wedge, pro-wedge, collisional, subduction), underthrusting of a cratonic basement promotes extrusion of deeply buried rocks, felsic or mafic, partially molten or not. Recent thermomechanical models of the Dalhousie Geodynamics Group suggested that underthrusting of stiff crustal indentors could squeeze-out rocks from the lower crust that would be extruded as “hot-fold nappes” (Culshaw et al., 2006; Jamieson et al., 2007). This process is akin to the “extrusive flow” referred to by Williams et al. (2006) that should, according to those authors, hinder the development of drag folds and result in tectonic overpressure. Alternatively, the force balance calculations presented in section

4.7 demonstrate that an important effect of basement underthrusting is to produce a lateral pressure gradient that can drive the flow of low-viscosity material. Ductile extrusion is probably the most efficient mechanism for redistributing mass and heat, and, therefore, for stabilizing an unstable crust. It is thus not surprising that a growing body of evidence suggests that ductile extrusion may have played an important role in bringing mid-crustal rocks within many orogenic belts to the surface of the Earth.

4.10 Conclusion

Results of this chapter can be summarized as follow:

1. The cover sequence of Frenchman Cap dome comprises four metamorphic zones that are each confined to a specific structural position within two fold nappes (Fig. 4.2 & 4.3). They occur in a ~5 km thick section below the base of the overlying LSa, which is considered as a fifth metamorphic zone (Zone A). From top to bottom these zones are: A) the sillimanite-migmatitic zone A within the Monashee décollement at the base of the LSa; 2) the sillimanite-kyanite migmatitic zone B, confined to the Sybley anticline; 3) the kyanite-muscovite zone C, confined to the Sybley syncline; 4) the (muscovite)-kyanite-sillimanite migmatitic zone D, confined to the Mount-Grace anticline; 5) the (muscovite)- sillimanite-migmatitic zone E, confined to the Mount-Grace syncline. Rocks from zones B, D and E reached the kyanite + melt field, but peak metamorphic assemblages of zone E have been almost completely retrogressed in the sillimanite field (e.g. Fig. 4.4 a, b, c, e, f).
2. Calculation of pseudosections together with previously published geochronologic data, provide the basis for the construction of P-T-t paths for the five metamorphic zones.

3. This study presents the first success of garnet isopleth thermobarometry in migmatitic rocks.
4. The P-T-t path for zone B, which now forms the Sybley anticline, indicates a slow burial and heating path in the kyanite field from $\sim 630^{\circ}\text{C}/750\text{ MPa}$ at $\sim 75\text{ Ma}$ to $\sim 800^{\circ}\text{C}/1100\text{ MPa}$ at $\sim 59\text{ Ma}$, and was followed by rapid cooling and decompression to $< 650^{\circ}\text{C}$, $< 400\text{ MPa}$ by $\sim 55\text{ Ma}$ (Fig. 4.7a). By contrast, the P-T-t path for zone C, which now forms the lower limb of the Sybley syncline, indicates a burial depth of only $\sim 700\text{--}800\text{ MPa}$ at $\sim 59\text{ Ma}$ (Fig. 4.7c).
5. These data imply formation of the Sybley fold nappe between ~ 59 and $\sim 55\text{ Ma}$ during synchronous exhumation of the anticline (zone B) and burial of the syncline (zone C). This could only have been achieved if rocks of the anticline flowed faster than their lid and base and were translated upward to upper crustal levels: i.e. they were exhumed by ductile extrusion (Fig. 4.9b, c).
6. The P-T-t path for the Mount-Grace anticline zone D indicates cooling and decompression from the kyanite + melt field to the andalusite field between ~ 59 and $\sim 55\text{ Ma}$ (Fig. 4.7d). The presence of kyanite-migmatites, which record similar retrograde P-T-t paths, above (zone B) and below (zone D) the kyanite-muscovite schists of zone-C, requires the latter to have been engulfed by flowing migmatites of the two other zones (Fig. 4.9c).
7. In contrast, a $50.5 \pm 1.0\text{ Ma}$ sillimanite-bearing leucosome in zone E (Crowley et al., 2001) implies that rocks from the Mount-Grace syncline were still flowing at high-temperature $\sim 5\text{ my}$ after the exhumation and cooling of the overlying zones that has a maximum age of $55.5 \pm 1.0\text{ Ma}$ (Crowley et al., 2001). Their juxtaposition also

implies extrusion of zone E beneath zone D (Fig. 4.9d, e).

8. A mechanical model is presented that reproduces the particle paths of the five zones and the formation of the two fold nappes while maintaining the crust in isostatic equilibrium. This model illustrates the formation of a channel at ~40 km depth that tunnels its way up an underthrusting basement ramp by ductile extrusion (Fig. 4.9).
9. Underthrusting of a basement composed of felsic orthogneisses (density of ~2800 kg/m³) underneath an orogenic wedge composed largely of garnet schists and metavolcanics (density ~2900 kg/m³) induces a lateral pressure gradient that adds to the lateral pressure gradient created by the topographic gradient. A force balance analysis demonstrates that the up-dip component of the force exerted by a lateral pressure gradient is much higher than the force exerted by shear stress at the base of the extruding slice and is sufficient to drive ductile extrusion throughout the model (Fig. 4.10).
10. The density contrast of ~300 kg/m³ between the kyanite + garnet-schists and the underlying felsic basement orthogneiss within the channel produces a buoyancy force that could drive a faster flow of basement rocks that would core fold nappe anticlines.
11. A comparison between Frenchman Cap and Thor-Odin domes of the Monashee Complex suggests that the model of ductile extrusion developed for Frenchman Cap dome is applicable to Thor-Odin, provided that the extruding channel was thicker in Thor-Odin

Table 4.1. Bulk chemical composition of metapelitic schists

Sample name	FG-462a	JC223	DG136	DG167	JC191	FG-459a
Zone	Zone A	Zone B	Zone B	Zone C	Zone D	Zone E
UTM/NAD83 ZONE 11	371817E 5724420N	376280E 5693390N	374670E 5723080N	377500E 5718025N	374260E 5711270N	376004E 5709362N
Major (wt%)						
SiO ₂	57.49	49.90	50.30	54.70	55.98	54.30
Al ₂ O ₃	18.75	26.54	28.00	24.10	22.40	23.90
CaO	1.22	0.54	1.58	2.31	0.70	0.59
K ₂ O	4.20	4.53	1.63	3.23	4.02	4.36
MgO	3.73	3.58	2.11	2.61	3.15	2.69
MnO	0.08	0.14	0.12	0.11	0.27	0.06
Na ₂ O	2.60	0.44	2.12	2.08	0.87	1.09
P ₂ O ₅	0.13	0.29	0.13	0.37	0.09	0.20
Fe ₂ O ₃ (T)	8.56	11.27	11.10	8.94	9.51	8.54
TiO ₂	1.07	1.04	0.92	1.00	0.84	0.89
L.O.I.	0.90	1.60	1.61	0.76	1.50	2.70
Sum	97.66	98.83	98.70	99.21	98.49	98.43
Trace (ppm)						
Ce	143.00	137.00	78.00	71.00	92.00	134.00
La	91.00	75.00	35.00	42.00	87.00	91.00
Nd	58.00	57.00	24.00	32.00	45.00	63.00
Th	29.00	21.00	16.00	15.00	27.00	30.00
U	11.00	8.00	4.00	5.00	10.00	13.00
Y	39.00	47.00	27.00	29.00	38.00	43.00
Zr	234.00	165.00	87.00	228.00	163.00	183.00
Ree Tot	292.00	269.00	137.00	145.00	224.00	288.00
A*	35.04	48.55	65.64	57.49	46.89	51.08
M	21.20	13.42	5.99	10.41	14.29	12.68
F	43.76	38.03	28.37	32.09	38.82	36.24
A	61.70	69.52	71.93	67.96	68.07	71.00
F	34.63	29.34	24.86	26.38	30.17	27.44
C	3.67	1.14	3.21	5.66	1.76	1.55
A	55.12	63.72	70.92	64.49	62.19	64.10
F	32.90	27.34	25.66	27.31	28.30	25.33
K	11.99	8.94	3.42	8.20	9.51	10.57
Al₂O₃/Shaw	1.10	1.56	1.65	1.42	1.32	1.41
CaO/Shaw	0.55	0.24	0.71	1.05	0.32	0.27

*AFM, ACF and AFK refer to the classification of Thompson (1957) and Eskola (1915).

Table 4.2. Density calculations.

Sample	Zone	Peak P (MPa)-T (°C)	Main minerals	Density kg/m ³
FG462*	A	800-750	Kfs -Sil-Grt-Bt	2950
JC223*	B	1100-800	Kfs-Ky-Grt-Bt	3050
DG167*	C	1000-750	Grt-Ky-Bt-Ms	3125
AH-02-07**	Basement	1000-800	Hbl-Bt-Kfs-Pl	2600
Basement + 30% amphibolite***		1000-800		2720
Core of Thor-Odin (FG204)****		500-750	Bt-Grt-Crd-Kfs	2710

*Calculated with Perple_X'06 directly from pseudosections.

**Calculated with Perple_X'06 from composition in Hinchey et al. (2006) assuming 20% melt was present.

***Calculated assuming an amphibolite density of 3000 kg/m³. The density is a maximum because less than 30% mafic rocks were observed, amphibolite density is likely lower, and more than 20% leucosome were observed at any given outcrop.

**** Calculated for a basement paragneiss of Frenchman Cap dome. Pseudosection for specimen Fg204 presented in Chapter 2 of this thesis better reproduces the assemblage described in Hinchey (2005), including the same amount of melt (30%), than a pseudosection for an aluminous metapelites such as specimen JC-223.

Table 4.3. Geotherms

Cordillera today (from Currie and Hyndman, 2006)	$T = 344 + 17.4 * Z$
Cordillera Eocene (from P-T path of JC223, this study)	$T = 331.5 + 11.8 * Z$
Craton (from Currie and Hyndman, 2006)	$T = 338 + 4.96 * Z$
Mantle lithosphere (mantle adiabat in Currie and Hyndman, 2006)	$T = 1289 + 0.4 * Z$

Where T is temperature in °C and Z depth in km

Table 4.4. Flow law parameters

Rock type	A (MPa⁻ⁿ s⁻¹)	n	Q (Kj/mol)	Source
Quartzite (1-2% melt)	1.8 x 10 ⁻⁸	4	137	Hirth et al, 2001
Migmatite	4.07 x 10 ⁻²	1.8	220	Rutter et al., 2006

Figure 4.1: Revised tectonometamorphic map of the southeastern Canadian Cordillera surrounding the Monashee Complex. Inset shows the five morphogeological belts of the Canadian Cordillera. BB is for the area including the Bowser Basin, Skeena Fold Belt and Sustut Basin. The gray area delineated by a solid line represents the current locations of these three entities, whereas the pale gray area delineated by a stippled line represents their approximate locations in the Late-Cretaceous prior to dextral strike-slip displacement (Gabrielse et al., 2006). Locations of specimens discussed in the text that are outside of Fig. 4.2 (location indicated by a square) are shown. The light gray unit is the Upper Selkirk allochthon (USa) that was deformed, metamorphosed and exhumed before ~120 Ma; the medium gray unit is the Lower Selkirk allochthon (LSa) that was mainly deformed, metamorphosed and exhumed between ~120 and 60 Ma; the dark gray unit is the high-grade cover sequence and infolded basement gneiss of the Monashee Complex that was deformed, metamorphosed and exhumed between ~59 and 47 Ma; the white unit with cross patterns represent basement rocks that record a polymetamorphic history in the Proterozoic and during the Cordilleran orogeny. Abbreviations: FCD: Frenchman Cap dome; TOD: Thor-Odin dome; MG: Malton gneiss.

Figure 4.1

Figure 4.2: Geological map of central Frenchman Cap dome, modified after Wheeler (1965), McMillan (1970), Höy and Brown (1980), Journeay (1986), Höy (1987), Crowley (1997), and Miller-Tait (2001); showing the map distribution of a simplified lithostratigraphic sequence and the location of specimens discussed in the text. The map also shows the trace of the four metamorphic isograds mapped in this study or modified from McMillan (1969), Journeay (1986), Höy (1997). These isograds delimit five metamorphic zones that coincide with specific structural levels, these are from west to east: zone A: the sillimanite migmatitic zone confined to the base of LSa; zone B: the sillimanite-kyanite migmatitic zone confined to the Sybley anticline; zone C: the kyanite-muscovite zone confined to the Sybley syncline; zone D: the (muscovite)-kyanite-sillimanite migmatitic zone confined to the Mount-Grace anticline; zone E: the (muscovite)-sillimanite migmatitic zone confined to the Mount-Grace syncline. The limit of Cordilleran strain is traced according to chapter 2 of this thesis. Notice the excellent control of the map distribution of the carbonatite/white marble/Pb-Zn marker horizon that was key to identifying the two fold nappes of the area (Höy and Brown, 1980; Journeay, 1986; Scammell, 1986; Höy, 1987).

Figure 4.2

Figure 4.3: Cross-section of central Frenchman Cap dome (modified after Journeay (1986) and Crowley (1997), which illustrates the geometry of the Sybley and Mount-Grace fold nappes, and the structural location of the five metamorphic zones discussed in the text. Also shown are lower hemisphere equal area projections of stretching and mineral lineation, labeled L, and minor fold axes, labeled F, collected within the two fold nappes on the west- to southwest-dipping flank of the dome. Data were collected by Journeay (1986) and during this study. Notice the parallelism between lineation and fold axes that both plunge gently to the west. ARD: top-to-the-west Anstey Range detachment; MD: top-to-the-east Monashee décollement; CRF: top-to-the-east Columbia River Fault.

Figure 4.3

Figure 4.4: Field and rock photographs of key metamorphic and structural features. a) Example of a specimen from the normal-sense Anstey Range detachment deformed largely in the presence of melt. Leucosome are not segregated into bands, pods or shear zones. Notice the constrictional-shear fabric with a weak and strong preferential mineral alignment on the YZ, and XZ sections, respectively. Kyanite crystals are oriented from bottom-left to top-right and are imbricated. This structure is interpreted as tiling in the presence of melt indicating a top-to-the-west sense of shear. b) Field photograph, looking north, of the normal-sense Anstey Range detachment deformed mostly in the solid-state. Leucosome are well segregated into bands, pods and in garnet pressure shadows. A C-S fabric indicating a top-to-the west sense of shear is formed by peak metamorphic assemblages composed of kyanite-garnet-biotite-leucosome. There are also shear bands going from top-right to bottom-left of the picture. Kyanite is partly replaced by sillimanite. c) Typical outcrop of a migmatitic aluminous metapelitic schists from zone B with garnet and kyanite megacrysts surrounded by leucosome. This feature suggests growth of garnet and kyanite in the presence of melt. d) Outcrop of a migmatitic aluminous metapelitic schists from zone C containing less than 10% leucosome, having a well-preserved compositional layering, no K-feldspar, and abundant muscovite. e) Leucosome JC-191 showing a spectacular pseudomorph texture of pink andalusite after blue kyanite. Similar textures were also observed in zone B. f) Leucosome collected beside the sillimanite-muscovite-biotite schist specimen FG-459. Clear evidence for the presence of melt includes: 1) crystal faces of feldspar; 2) the euhedral shape and large size of garnet in the leucosome; 3) the presence of biotite-rich restite surrounding the leucosome; 4) Garnet and kyanite crystals are larger near, or within the leucosome than in the matrix.

Figure 4.4

Figure 4.5: Photomicrographs of metamorphic textures. a) One of the rare garnet of specimen FG-462a, a semi-pelitic migmatitic schist from zone A. The long axis of this oblate garnet is inclined relative to the foliation (which is parallel to the picture) and is oriented from bottom-left to top-right of the image. Notice the preferential resorption of garnet in the bottom-left to top-right quadrants, with sillimanite-biotite-plagioclase concentrated in these pressure shadows. This microtexture is interpreted as structurally-controlled garnet resorption by ReNtr 1 during top-to-the-east shearing (dextral on the picture). b) Photomicrograph of the andalusite-sillimanite-garnet-kyanite-K-feldspar-biotite schist specimen JC-223 from zone B. Muscovite, plagioclase, ilmenite, biotite, quartz and allanite form the assemblage of the inclusion-rich core, whereas rutile, ilmenite, monazite, zircon and biotite form the assemblage of the inclusion-poor rim. Garnet is resorbed with sillimanite-biotite-plagioclase-rutile in its pressure shadows. Garnet is not as resorbed as garnet from zone D and E. c) Typical texture of an aluminous metapelitic schist from zone C. The lower metamorphic grade reached by rocks from this zone is indicated here by: 1) abundant muscovite; 2) the green color of biotite that contrasts with red biotite on the other photomicrographs; 3) smaller size of kyanite and garnet porphyroblasts, which suggest the absence of a melt phase that promoted growth in other zones. d) Migmatitic muscovite-sillimanite-garnet-kyanite-biotite schist hosting leucosome JC-191 from zone D. Garnet is much more resorbed than garnet from zone B (Fig. 4.5a). Notice that sillimanite is mainly located around resorbed garnet, whereas muscovite is only found at the margin of resorbed kyanite. e) A nodule cored by one of the rare garnet observed in a migmatitic metapelitic schist of zone E. The nodule is composite with an outer shell composed of sillimanite and biotite and an inner shell composed of plagioclase, quartz, muscovite and rutile. f) Typical sillimanite-biotite nodule of zone E interpreted as a replacement texture of garnet in the presence of melt by ReNtr 1. See text for additional interpretation. Scale bars are all 1mm long.

Figure 4.5

Figure 4.6. Compositional profiles for garnet of zones B and C.

a) Garnet from sample JC-223 of zone B. Notice the bell-shaped X_{sps}, which is typical of prograde zoning (e.g. Spear, 1993). The spike in X_{sps} and Fe / Fe + Mg ratio is typical of garnet breakdown in the presence of melt by ReNtr 1. b) Garnet from sample DG-167 of zone C (from Foster et al., 2004). Note the prograde zoning on the right side of the profile and the flat profile on the left side. See text for explanation.

Figure 4.7. P-T pseudosections calculated with *Perple_X'06* in the extended TiMnCaNaKFMASH system and P-T-t paths. Quartz is in excess in all fields except for fields containing the label (q). Pseudosections are composite, with the portion below 650°C calculated without a melt solution model and saturated in a C-O-H fluid as defined by Connolly and Cesare (1993). The amount of H₂O calculated at ~650°C and ~0.7 GPa was then used to calculate the higher temperature portion of the pseudosection, which includes a melt model. The composite nature of pseudosections required minor adjustments that were accounted for by visually adjusting the fields to agree with Schreinemaker's rules. a) Pseudosection calculated for the bulk composition of the sillimanite-andalusite-garnet-kyanite-K-feldspar-biotite schist specimen JC-223 from zone B (Table 4.1). Notice the perfect intersection of isopleths for the garnet core having the highest X_{sps} (profile illustrated on Fig. 4.6). Notice the detailed P-T-t path characterized by relatively slow burial and heating and very fast decompression and cooling after partial melting in the kyanite field at ~59 Ma. b) P-T pseudosection for the andalusite- sillimanite-kyanite-garnet-muscovite-biotite schist specimen DG-167. Notice the excellent isopleth intersection for garnet core and inner rim compositions. Wet-solidus line was calculated for the same bulk composition, but with excess water (6 wt%). c) P-T-t retrograde paths for zones D and E using the background of a). See text for details.

Figure 4.7a, b

Figure 4.7c

Figure 4.7c. P-T pseudosections and P-T-t paths for zones D and E.

Figure 4.8

Figure 4.8. Sketch of timing relationships between growth of accessory phases allanite and monazite with respect to growth of major phases garnet and kyanite in zone B and C. U-Pb ages of monazite. Modified from Foster and Parrish (2003). See text for detail.

Figure 4.9: Quantitative mechanical model showing the formation of fold nappes by ductile extrusion during basement underthrusting. Stations A-E refer to the respective five metamorphic zones mapped in this study. The upper part of the wedge is fixed during underthrusting. Black bar shows the horizontal displacement for each time-step. a) Homogeneous channel flow and ductile extrusion of the LSa accommodated by the top-to-the-west Okanagan Valley fault system (OVfs) at the roof, and top-to-the-east Monashee décollement (MD) at the base. b) Rocks of zones B and D have been buried and heated to the kyanite-melt field and start to flow. c) Rocks of zone B are exhumed by ductile extrusion, which led to the formation of the anticline of the Sybley fold nappe. The top-to-the-west Anstey Range detachment (ARD) accommodated the extrusion. d) Ductile extrusion of rocks from zone D underneath the Sybley fold nappe led to the formation of the syncline of the Sybley fold nappe and to the anticline of the Mount-Grace fold nappe. e) Ductile extrusion of rocks from zone E and from upper basement level finalized the exhumation of rocks of Frenchman Cap dome from middle to upper crustal level. Also shown is a dome formed by buckling, which is the preferred, but still speculative, doming mechanism. See text for details.

Figure 4.9

Figure 4.10. Results of force balance analysis.

The lateral force (open circles) results from the density contrast between two adjacent lithospheric columns and from the topographic gradient, integrated over the thickness of the channel. Shear force (filled symbols and crosses) is the integral of the shear stress acting at the base of the channel, which is function of rheological parameters and temperature. Extrusion is possible when the up-dip force produced by lateral pressure gradient is larger than the shear force. Notice that exhumation to less than 15 km depth by extrusion is expected for quartzite and migmatites at a geologically acceptable strain rate of $1E-13 \text{ s}^{-1}$, and that exhumation to a depth of $\sim 17 \text{ km}$ is expected for migmatites deforming at a strain rate of $1E-14 \text{ s}^{-1}$.

Figure 4.11. Sketch depicting the formation of a fold nappe within an extruding channel. The density contrast between felsic orthogneisses of the Monashee basement and garnet + kyanite schist of the cover sequence creates an up-dip pressure gradient that should drive a faster flow of basement rocks. This provides an explanation for the presence of felsic basement rocks in the cores of anticlines and for the kinematic reversal centered on basement rocks observed in the Sybley syncline.

CHAPTER 5

CONCLUSION AND GENERAL DISCUSSION

The aim of this thesis was to gain a better understanding of the processes that led to the exhumation of the high-grade rocks exposed in part of the southeastern Canadian Cordillera. In contrast to the generally accepted models of exhumation by gravitational collapse and extension of a thickened crustal welt (Coney and Harms, 1984; Parrish et al., 1988; Bardoux and Mareschal, 1994; Teyssier and Vanderhaeghe, 1997; Vanderhaeghe et al., 1999; Norlander et al., 2002; Teyssier and Whitney, 2002; Teyssier et al., 2005; Kruse and Williams, 2007), this thesis rather presents evidence in favour of sequential exhumation during convergence. The strongest evidence for synconvergent exhumation includes: 1) a maximum burial depth for the deepest structural level of Frenchman Cap dome that is at least 15 km less than the burial depth recorded at overlying structural levels (chapter 2); 2) Eocene strain patterns at the upper and lower basement levels of the dome are more compatible with contraction than extension (chapter 2); 3) the synchronous exhumation of the LSa and burial of the underlying cover sequence of the Monashee Complex in Late-Cretaceous-Paleocene time (chapter 3 & 4); 4) the synchronous exhumation of the anticline and burial of the underlying syncline of the Sybley fold nappe (chapter 4). Synconvergent exhumation between ~90 and ~50 Ma is also more compatible with plate motion reconstructions that indicate a peak convergence rate during this time period (Stock and Molnar, 1988), and with the timing of the main shortening phase in the Foreland fold belt that was bracketed between ~75-59 Ma (Price and Sears, 2000).

A mechanical, isostatically balanced model for the Mid-Cretaceous to Eocene tectonic evolution of the southeastern Canadian Cordillera at the latitude of the Monashee Complex is presented on Fig. 4.9. The geometry of the model is firmly constrained by an extensive database provided by field-based structural analyses (Van den Driessche, 1983; Journeay, 1986; Brown et al., 1992; Scammell, 1993; Gibson et al., 1999; Johnston et al., 2000; Brown, 2004; Williams and Jiang, 2005), complemented with subsurface geometry, provided by the Lithoprobe geophysical surveys (e.g. Cook et al., 1992; Burianyk et al. and Kanasewich, 1995). The feasibility of this model is demonstrated to a first order by the force balance analysis presented in chapter 4. As summarized in chapter 1 (from Evenchick et al., 2007), the Omineca Belt was thick enough in the Jurassic to induce midcrustal deformation. The orogenic wedge grew by underthrusting and accretion of pericratonic material that consisted mainly of sedimentary and volcanic rock. The thermal peak was probably reached at ~100 Ma as suggested by the clustering of U-Pb ages obtained on numerous leucogranites, leucosomes and monazites in schists from the LSa (chapter 3). After millions of years of shortening, it is most likely that the Omineca Belt was a wide plateau of high elevation by ~100 Ma, and that a significant portion of the middle crust was partially molten. The low viscosity of the partially molten mid-crustal layer facilitated its flow down the pressure gradient created by the difference in elevation between the high plateau and the Foreland basin. Lateral crustal flow is inferred to have started at ~100 Ma and exhumation by ductile extrusion to have started between ~98-90 Ma, as suggested by the depth-time path constructed for area A (Fig. 3.10). Because the depth-time paths from various locations of the LSa (Fig. 3.10) suggest a westward-younger trend of exhumation, it is likely that ductile extrusion in the east was

fed by easterly-directed lateral channel flow from underneath the orogenic core. This process could well have taken place continuously from ~98 Ma to ~60 Ma according to the exhumation paths presented on Fig. 3.10. The main driving force of flow between ~100 and ~75 Ma was most likely the topographic gradient. However, by ~75 Ma the P-T-t path from the Sybley anticline of the Monashee Complex (Fig. 4.7a) indicates that basement rocks of the Monashee Complex started to be underthrust. As indicated by the force balance analysis presented in chapter 4, underthrusting of felsic orthogneiss, which has a density $\sim 100 \text{ kg/m}^3$ lower than the average density of the overlying wedge, created a significant lateral pressure gradient that would have added up to the lateral pressure gradient induced by topography, thus favouring ductile extrusion. Forty years ago, Price and Mountjoy (1970) proposed a model according to which "upwelling and lateral spreading of tongues of hot gneissic rocks" was the main driving force for thrusting in the Foreland Belt. Conclusions presented in this thesis, which indicate that the main phase of shortening in the Foreland Fold Belt (~75-59 Ma) coincided with the main phase of ductile extrusion of the LSa, do suggest that ductile extrusion was a driving force for foreland thrusting. However, ductile extrusion was probably not the main driving force because: 1) rocks of the Monashee Complex were being underthrust during this time period; 2) ductile extrusion implies delamination with upper crustal rocks; 3) foreland folding and thrusting is recorded along the entire length of the orogen, not only where high-grade rocks of LSa are exposed.

The model presented on Fig. 4.9 suggests a change from homogenous ductile extrusion of the LSa between ~98-60 Ma to heterogeneous and sequential ductile extrusion that led to the formation of fold nappes in the Monashee Complex between

~59-50 Ma. It is possible that this change of flow type is only apparent and due to the scale of observation, and that the entire L_{Sa} is in fact formed by several slices of rocks juxtaposed together by sequential ductile extrusion as documented herein for the Monashee Complex. However, there is firm geochronologic evidence that rocks of the L_{Sa} remained at mid-crustal level for a much longer period of time than rocks of the Monashee Complex (>60 vs. <15 my, respectively). This suggests that it took several millions of years to develop conditions favourable for lateral ductile flow of a 10-15 km thick channel (i.e. the L_{Sa}; Fig. 4.9a), but that afterward, underthrust rocks were rapidly expelled from the middle crust once partially molten (Fig. 4.9b-e). I argue that the flow type change is real and is a natural consequence of orogenic evolution. It is indeed expected that several millions of year of shortening would be required to develop a high-elevation plateau and to raise the temperature high enough to induce partial melting of a significant portion of the crust, but that once reached, these conditions would be maintained until a major change in far-field stress. Accordingly, it would be expected that the first phase of lateral channel flow and ductile extrusion would involve a relatively thick channel that would flow relatively homogeneously, but that subsequently, underthrust material would sequentially be expelled from the middle crust once a sufficient volume of rocks would have reached a viscosity low enough to counteract the shear strength at their base (chapter 4). As discussed in chapter 4, similar processes have been documented in several other orogens.

Although not an initial objective, results of this thesis point to a new model of fold nappe formation. The P-T-t paths from the upper and lower limb of the Sybley syncline (Fig. 4.7a, b) imply formation of the fold nappe between ~59 and ~55 Ma by coeval

exhumation of rocks from the upper limb and burial of rocks from the lower limb. As discussed in chapter 4, these paths imply a process involving ductile extrusion coupled to underthrusting (Fig. 4.9 b, c). A first-order force balance analysis (chapter 4) suggests that the formation of fold nappes cored by felsic orthogneiss could be a direct consequence of the internal density contrast created during metamorphism by inducing a higher velocity of flow of the felsic orthogneiss.

Finally, unravelling the doming mechanism was a high priority when I started this project, but the exact process that led to the formation of the two gneiss domes of the Monashee Complex remains enigmatic. Nevertheless, an important step forward was accomplished by demonstrating that the widely publicized model of extension-assisted diapirism (Teyssier and Vanderhaeghe, 1997; Vanderhaeghe et al., 1999; Teyssier and Whitney, 2002; Teyssier et al., 2005) is not valid for Frenchman Cap dome because of the preservation of a Paleoproterozoic foliation, of the strain pattern and of the baric peak conditions recorded at the deepest exposed structural level of the dome (chapter 2). It is likely that the ductile extrusion process documented in chapter 4 contributed to the formation of the dome in a way analogous to doming of a channel above an underthrusting stiff indenter documented in the thermomechanical models of Beaumont et al. (2004). By doming the overlying USa, ductile extrusion would produce a lobate boundary between the low-grade rocks of the USa and underlying high-grade rocks. After erosion, this boundary would present many characteristics of a normal fault having a maximum net-slip displacement at the centre of the dome and minimum net-slip displacement at the margins. This could provide an explanation for the observed spatial relationship between the Columbia River Fault and the domes of the Monashee Complex.

However, the Beaumont et al. (2004) model alone would not predict the exposure of basement gneiss with a Paleoproterozoic foliation in the core of the dome. As discussed in chapter 2, a syncontractional doming model, such as buckling, is likely, but data presented herein are insufficient to conclude in favour of a specific mechanism.

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APPENDICES

Appendix 1: Field data from the cover sequence of Frenchman Cap dome

UTME	UTMN	OTCRP	ST_strk	ST_dip	L_plunge	L_trend	Shear sense	Tectonite	F2 plunge	F2 trend	F2 type	Unit#	AL2SIO5
Mount-Grace Camp													
370765	5701651	05-FG-301	170	28	28	290		L>S				2	
370713	5702235	05-FG-302	160	42	40	260						2	
370940	5702291	05-FG-303	152	24	22	276						2	
370988	5702009	05-FG-304										1	
371113	5701617	05-FG-305	135	38	36	262						1	
371163	5701448	05-FG-306	138	29	26	242						1	
371214	5701264	05-FG-307	135	36	36	270						1	
371213	5701409	05-FG-308	155	45	42	270			30	300	z	1	
371258	5701480	05-FG-309										1	
371263	5701938	05-FG-310	130	28	25	262						2	
371325	5702037	05-FG-311	185	32	30	290		L>S	30	260	z	2	
371369	5701898	05-FG-312	144	27	27	280		L=S				2	
371424	5701782	05-FG-313	140	30	25	270		S>L				2	
371479	5701569	05-FG-314										1	
371527	5701272	05-FG-315	140	10				S				1	
371559	5701149	05-FG-316										1	
371603	570963	05-FG-317										1	
371711	5701071	05-FG-318										1	
371922	5701370	05-FG-319	160	35	30	270						1	
371922	5701370	05-FG-320	145	35	35	280			10	290	s	1	
372065	5700763	05-FG-321	155	36	32	272		L>S				1	
372700	5700962	05-FG-322			30	281		L>S				1	
372600	5700963	05-FG-323						L>>S				1	

UTME	UTMN	OTCRP	ST_strk	ST_dip	L_plunge	L_trend	Shear sense	Tectonite	F2 plunge	F2 trend	F2 type	Unit#	AL2SIO5
Blais Creek Camp													
372156	5704667	05-FG-325	250	30								4	
372168	5704465	05-FG-326										4	
372265	5704659	05-FG-327	17	30	15	125						4	
371866	5704855	05-FG-328	195	51	47	300		L>S	47	300	?	4	
371929	5704669	05-FG-329										4	
372051	5704991	05-FG-330										4	
372063	5705046	05-FG-331	200	42	30	330		L>S				4	
371988	5705069	05-FG-332	190	40	32	315		L>S	35	280		4	
372473	5704911	05-FG-333	190	16	15	300	SE					4	
372459	5705190	05-FG-334	330	30	30	60			15	15	z	4	
372609	5705519	05-FG-335	220	35	30	275						4	S
373368	5704901	05-FG-336	145	35	35	245						4	S
372778	5705508	05-FG-337	295	15				S>L				4	S
Ratchford Creek Camp													
375213	5693312	05-FG-338										6	KSM
375629	5693125	05-FG-339										6	
375658	5693042	05-FG-340	145	40	38	260	E					6	SKM
376095	5692794	05-FG-341	140	30	30	245						6	M
376228	5692834	05-FG-342	190	10	10	270	W					6	M
376415	5692873	05-FG-343										6	KMs
376528	5692972	05-FG-344	140	50								6	KMs
376528	5692972	05-FG-344	155	42			E					6	
376635	5692973	05-FG-345	102	32	20	240	E					5	
376952	5693077	05-FG-346			32	260		L				dx	
377113	5693000	05-FG-347	175	45	44	275	W		30	340	S (F3)	4	Ms
377019	5693050	05-FG-348			20	270		L				4	Ms

UTME	UTMN	OTCRP	ST_strk	ST_dip	L_plunge	L_trend	Shear	Tectonite	F2 plunge	F2 trend	F2 type	Unit#	AL2SIO5
							sense						
377327	5692940	05-FG-349	150	35	32	272						1	
377511	5693088	05-FG-350	150	30	28	272						1	
377655	5693217	05-FG-351	150	50					47	235	Z	dx	
377882	5693711	05-FG-352	205	43	30	250						4	
378300	5694227	05-FG-353	150	45	43	205	W	L				6	KSM
378261	5694366	05-FG-354	130	50	50	220						6	M
378409	5694384	05-FG-355							25	284	M	6	SMsMK
377681	5692382	05-FG-356	130	35	32	255			16	275	S	1	
377274	5692397	05-FG-357	150	25	24	265						1	
377643	5693001	05-FG-358	105	25	15	260	N					1	
378549	5694406	05-FG-359					E					6	
376023	5693442	05-FG-360	125	35	25	262	NE					6	
375783	5693584	05-FG-361										6	KMA
375592	5693760	05-FG-362										gabbro	
376305	5693124	05-FG-363										6	KS
378582	5694342	05-FG-364	120	75	25	285						6	Ms
378661	5694319	05-FG-365					SW	L>S	30	280		6	S
378830	5694311	05-FG-366										6	
379011	5694384	05-FG-367	185	30					25	285	S	6	
379160	5694452	05-FG-368	155	40					28	260	M	6	SMK
379286	5694400	05-FG-369	165	42	35	286						5	
379442	5694460	05-FG-370	170	35	34	270	E	S=L				6	S
379579	5694486	05-FG-371										4	SKMMs
379882	5694504	05-FG-372										4	SMs
380083	5694600	05-FG-373	180	30	28	241						4d	SMs
380249	5694940	05-FG-374										4	M
380237	5695056	05-FG-375	150	40				S				4	
380274	5695135	05-FG-376										4	

UTME	UTMN	OTCRP	ST_strk	ST_dip	L_plunge	L_trend	Shear sense	Tectonite	F2 plunge	F2 trend	F2 type	Unit#	AL2SIO5
379892	5694868	05-FG-378										4	
380005	5694893	05-FG-379										6	
380195	5694819	05-FG-380										6dx	SM
375479	5693977	05-FG-381	150	28	26	250						6	KSM
375545	5694106	05-FG-382										6	M
375845	5693876	05-FG-383	146	30	29	272						6	KSM
Goat Lake Camp + Hop&Go													
374516	5711396	05-FG-458	230	30	25	300	NE		30	285		6	KM
376004	5709362	05-FG-459										6	SKM
373252	5723604	05-FG-460										6	SKM
373052	5724420	05-FG-461										6	MsKM
371817	5725255	05-FG-462	305	25	10	100	E					LSa	SM
372951	5724781	05-FG-463	215	15	12	290						LSa-6	SKMs
374792	5723318	05-FG-464	245	35									
375066	5722995	05-FG-465	255	60	20	270	W					6	SKMA
		05-FG-466					W					6	KM
373321	5715548	05-FG-467			45	265	E					6	ASKM
Anstey Range Camp													
378688	5686328	05-FG-484			43	263	W					6	KSM
379024	5686309	05-FG-485			35	263	W					6	SKAM
379123	5686252	05-FG-486			34	263	E					4	KMs
378983	5686239	05-FG-487										6-LSa	SKAM
378586	5686499	05-FG-488			35	270							SM
378333	5686504	05-FG-489			20	270						6	
380506	5684882	05-FG-490										6	KSMs

Legend

Lithologic units (as per Höy, 1987)

- 1: Basement orthogneiss
- 2: Basement paragneiss
- 3: Basal quartzite
- 4: Mainly calc-silicates
- 5: White marble
- 6: Mainly pelitic schist
- dx: diatexite

LSa: LSa

Metamorphic minerals

- K: Kyanite
- S: Sillimanite
- A: Andalusite
- M: Melt
- Ms: Muscovite

Appendix 2. Field data from the basement levels of Frenchman Cap dome.

OTCP	UTME	UTMN	Unit	ST strk	ST dp	ST L	ST L	F2 plng	F2 trnd	F2 plng	F2 trnd	F3 type	F3 plng	F3 trnd	F3 type	Shear	Shr	sns.	Tect.
<u>Upper basement level</u>																			
<u>Pettipiece Pass Camp</u>																			
04-FG-208	389044	5693593	2a	270	10	8	60	24	340	z							NE	L=S	
04-FG-209	389299	5693726	1	245	25	5	250	5	60	s							NE	L=S	
04-FG-210	389424	5693807	bg	320	15	15	65										NE	L=S	
04-FG-211	390162	5694513	3	145	50	45	245											L=S	
04-FG-212	390114	5694480	3	155	50	40	290	30	330	z								L=S	
04-FG-213	390125	5694374	3	180	15							10	350	z				L=S	
04-FG-214	390237	5694284	1	310	10	8	268											L=S	
04-FG-215	390237	5694284	1									25	190	z				L=S	
04-FG-216	389872	5694371	1,3															L=S	
04-FG-217	390051	5694318	1	270	77			30	300	z						E	L=S		
04-FG-218	390883	5694075	1,3									15	310	z			NE	L=S	
04-FG-219	390750	5693927	1,3	280	22	10	295					20	300	z				L=S	
04-FG-220	389916	5694283	1					15	60	s								L=S	
04-FG-221	389886	5694251	1	300	48			45	45	s						NW	L=S		
04-FG-222	389728	5694233	2a	290	10	4	70											L=S	
04-FG-223	389761	5694161	2b	272	8	5	80	5	80							NE	L=S		
04-FG-224	389744	5694138	2b									5	335	z				L=S	
04-FG-225	389759	5694096	2b															L=S	
04-FG-226	389659	5694150	2b	165	10			10	260	s								L=S	
04-FG-227	389514	5694145	2a			4	240					30	330	z		N	L=S		
04-FG-228	388141	5693834	2a									15	295	z				L=S	
04-FG-229	388165	5693887	2a	244	26	4	50	8	50	s								L=S	
04-FG-230	388114	5693751	2a													NE	L=S		
04-FG-231	388132	5693711	2a			8	50											L=S	

OTCP	UTME	UTMN	Unit	ST	ST	L	L	F2	F2	F3	F3	F3	Shear	Shr.sns.	Tect.	
				strk	dp	plng	trnd	plng	trnd	type	plng	trnd	type	strk.	dp	top-to-
04-FG-232	388292	5693731	2b	248	35	36	310							310	15	L=S
04-FG-233	388509	5693754	2b	250	35	6	60							130	45	L=S
04-FG-233	388509	5693754	2b											15	65	L=S
04-FG-233	388509	5693754	2b											328	45	L=S
04-FG-233	388509	5693754	2b													
Pettipiece Pass Camp																
04-FG-233	388509	5693754	2b											164	65	L=S
04-FG-233	388509	5693754	2b											298	52	L=S
04-FG-233	388509	5693754	2b											145	65	L=S
04-FG-233	388509	5693754	2b											15	65	L=S
04-FG-233	388509	5693754	2b											135	72	L=S
04-FG-233	388509	5693754	2b											285	65	L=S
04-FG-233	388509	5693754	2b											215	75	L=S
04-FG-233	388509	5693754	2b											190	75	L=S
04-FG-233	388509	5693754	2b											33	65	L=S
04-FG-233	388509	5693754	2b											340	55	L=S
04-FG-233	388509	5693754	2b													
04-FG-233	388509	5693754	2b											120	70	L=S
04-FG-234	388728	5693890	2a								15	320	z			L=S
04-FG-235	388842	5693949	2a	240	30						30	310	w			L=S
04-FG-236	389020	5694048	2a	295	20	15	65	15	65	s						L=S
04-FG-237	389091	5693887	2a	275	25	5	64	2	110	s						L=S
04-FG-238	389069	5693752	2a													L=S

OTCP	UTME	UTMN	Unit	ST	ST	L	L	F2	F2	F3	F3	F3	Shear	Shr.sns.	Tect.
				dp	dp	plng	trnd	plng	trnd	type	plng	trnd	type	strk.	dp top-to-
<u>Transition zone in basement</u>															
05-FG-410	389294	5688369	2b			15	185								S
05-FG-411	389264	5688412	2a												
05-FG-412	389140	5688313	2a												
05-FG-413	389114	5688309	2a												
05-FG-414	387485	5688112	2a	160	15	10	235								L
05-FG-415	387514	5687393	2a	150	12	10	245								
05-FG-416	388123	5687647	2b	185	20	18	230	18	230	s					
05-FG-417	388001	5687664	2b					25	245	s					
05-FG-418	387784	5687724	2a												
05-FG-419	388237	5687720	2a	180	25	20	265								
05-FG-420	388308	5687741	2b			45	255	22	230	s					
05-FG-421	388451	5687781	2a												
05-FG-422	388497	5687782	2a					20	180	s					
05-FG-423	388689	5687796	2a	195	45	30	240								L=S
05-FG-424	387200	5690773	2b												L>S
<u>Lower basement levels</u>															
<u>Bourne Glacier Camp</u>															
04-FG-121	389960	5688795	2a	290	2								205	18	
04-FG-122	389970	5688790	2a	190	10	0	350	15	355	s			255	45	
04-FG-122	389970	5688790	2a										285	32	
04-FG-122	389970	5688790	2a	10	150	10	150								
04-FG-123	?	?	2a	156	24	0	160								
04-FG-124	?	?	2a												
04-FG-125	?	?	2a	145	32	8	340								
04-FG-126	?	?	2a			17	130	330	54						
04-FG-128	389850	5689033	2a	300	4	3	130						180	16	
04-FG-129	389805	5689040	2a	320	50	20	80								
04-FG-130	389523	5688395	2b	160	35	30	235								S
								18	320	w					

OTCP	UTME	UTMN	Unit	ST	ST	L	L	F2	F2	F2	F2	F3	F3	F3	F3	Shear	Shr.	sns.	Tect.
				strk	dp	plng	trnd	plng	trnd	plng	trnd	type	trnd	type	strk.	dp	top-to-		
04-FG-131	389542	5688385	2b	155	30	22	246												
04-FG-132	389607	5688245	2b			24	260	30	260	s	25	313						L	
04-FG-132	389607	5688245	2b					0	175	z									
04-FG-133	389456	5688325	2b																
04-FG-134	389482	5688305	2b	160	30	30	216											L=S	
04-FG-135	389581	5688321	2b	168	30													S	
04-FG-136	389581	5688273	2b	170	34	5	190												
04-FG-137	589700	5688198	2b	200	20	20	307					18	310					L=S	
04-FG-138	?	?	2b	250	20	20	348												
04-FG-139	389760	5688124	2a	175	25	25	250												
04-FG-140	389769	5688095	2a	170	35	33	238											S>L	
04-FG-141	389787	5688090	2a	255	40	5	270								240	4		L=S	
04-FG-142	389731	5688030	2a	185	30	275	30								135	60		L=S	
04-FG-142	389731	5688030	2a																
04-FG-143	389739	5688000	2a																
04-FG-144	389758	5687941	2a	220	20														
04-FG-145	389789	5688058	2b	200	15										80	30			
04-FG-146	?	?	2b	230	45	24	280												
04-FG-147	389806	5688046	2b	240	65							0	240					L=S	
04-FG-148	?	?	2b	175	45	38	295												
04-FG-149	389824	5688102	2b	175	35	35	265											S>L	
04-FG-150	389787	5688157	2b	180	25	22	270											S	
04-FG-151	?	?	2b	180	20														
04-FG-152	?	?	2b	165	40	15	285												
04-FG-153	?	?	2b	150	40														
04-FG-154	?	?	2b	150	40	35	200												
04-FG-155	?	?	2a	160	35	35	220											S>L	
04-FG-157	390290	5686691	1	160	40	20	255								195	40			
04-FG-158	390242	5686700	1																

OTCP	UTME	UTMN	Unit	ST	ST	L	L	F2	F2	F2	F2	F3	F3	F3	Shear	Shr.	sns.	Tect.
				dp	dp	plng	trnd	plng	trnd	type	type	type	type	type	strk.	dp	top-to-	
04-FG-158	390242	5686700	1												210	35		
04-FG-158	390242	5686700	1												50	72		
04-FG-158	390242	5686700	1												130	55		
04-FG-159	390200	5686650	1															
04-FG-160	390341	5686672	amph	220	20	20	295											
04-FG-161	390415	5686640	1															
04-FG-162	390485	5686600	1	193	28	28	283										E	
04-FG-163	390515	5686620	1	190	25	4	245										E	
04-FG-164	389855	5687820	1	160	37	40	240											L=S
04-FG-164	389855	5687820	1	148	44	34	304											
04-FG-165	389882	5687960	2b	165	30													S
04-FG-166	389895	5688085	2b															
04-FG-167	389956	5688147	1	135	35	18	270											L=S
04-FG-169	391500	5687730	1	174	33	33	257	10	355	z								L=S
04-FG-170	391520	5687690	1															
04-FG-171	391474	5687690	1	110	55	5	110	55	145	w								
04-FG-172	391405	5687680	2b	310	85	15	330											
04-FG-173	391340	5687710	2b	175	40	35	300											
04-FG-174	391205	5687700	2b	345	25	15	290											
04-FG-174	391205	5687700	1	155	14	8	292											
04-FG-175	390260	5688460	1	285	35	15	312											
04-FG-176	390140	5687860	1	158	38	35	304											
04-FG-184	390240	5688440	1	260	25	4	320	0	340	z								
04-FG-185	390160	5688430	1	10	10	10	310	12	255	s								
04-FG-186	390180	5688355	1	135	20	10	275											
04-FG-187	390387	5686506	1	135	25	22	195											
04-FG-188	390487	5686607	1	125	35	25	210											
04-FG-188	390487	5686607	1	210	65													
04-FG-189	390771	5686295	2a	145	28													

OTCP	UTME	UTMN	Unit	ST	ST	L	L	F2	F2	F2	F2	F3	F3	F3	Shear	Shr	sns.	Tect	
				strk	dp	plng	trnd	plng	trnd	type	plng	trnd	type	plng	trnd	type	strk.	dp	top-to-
Frisby Creek Camp																			
05-FG-425	394005	5677681	2a	130	45							30	165	M	160	70			S
05-FG-426	394005	5677681	2a	150	75										150	75			S
05-FG-427	394005	5677681	2a												350	80			S
05-FG-426	394543	5677555	2a	135	45							30	185	M					S
05-FG-427	394217	5677698	2a	0	75							45	180	S					S
05-FG-428	394789	5677126	2a	120	55			15	125	S					160	70			S
05-FG-429	394693	5677160	2a	115	35														S
05-FG-430	394553	5677292	2a	72	25	25	145	35	80	Z									S
05-FG-431	394473	5677335	2a	155	60							15	190						S
05-FG-432	394801	5677174	2a	95	35							30	165		165	75			S
05-FG-433	394788	5677279	2a																S
05-FG-434	394753	5677361	2a									30	150						S
05-FG-435	394687	5677493	2a	80	35			15	140	S									S
05-FG-436	394565	5677541	2a	118	38							20	145	S	180	65			S
05-FG-437	394456	5677659	2a	155	49							25	170	S	170	75			S
05-FG-438	394351	5677645	2a	60	50			14	45	S		35	190		15	85			S
05-FG-439	392714	5677531	2a	120	45							30	180		90	60			S
05-FG-439	392714	5677531	2a	130	60			15	295	Z					180	60			S
05-FG-440	392932	5677488	2a			0	120												S
05-FG-441	393928	5677696	2b																S
05-FG-442	393908	5677604	2a					15	290	S		30	175	S					S
05-FG-443	393969	5677540	2a									30	180						S
05-FG-444	393897	5677540	2a																S
05-FG-445	393813	5677455	2b																S
05-FG-446	393702	5677332	2a												160	80			S
05-FG-447	393689	5677604	2b	115	60			40	145	Z									S
05-FG-448	393689	5677604	2a	125	60	30	170	42	115	S		5	160						S
05-FG-449	393441	5677437	2a	125	70			40	135	S									S

OTCP	UTME	UTMN	Unit	ST	ST	L	L	F2	F2	F3	F3	F3	Shear	Shr.	sns.	Tect.
				stri	dp	plng	trnd	plng	trnd	type	plng	trnd	type	strk.	dp	top-to-
<u>Perry River Camp</u>																
05-FG-469	389488	5675312	2a	155	75	10	270									L>S
05-FG-470	389443	5675348	2a													L
05-FG-471	389355	5675396	2a			20	90									L
05-FG-472	389304	5675358	2a	170	70	10	270							W		L>S
05-FG-473	389562	5675206	2a			15	285									L
05-FG-474	390535	5675409	2b-um	140	30											
05-FG-475	390584	5675488	um													
05-FG-476	390535	5675409	um,2b			20	275									L
05-FG-477	390648	5675595	um													
05-FG-478	390590	5675653	2b,amph													
05-FG-480	390612	5675321	um													
05-FG-481	389315	5675371	2a													
05-FG-482	389447	5675427	2a	236	50											
05-FG-483	389414	5675359	2a	240	20	15	90	z								

Legend

Lithologic units

2a: Granodioritic augen gneiss b: Bourne granite stock

2b: Leucocratic gneiss with amphibolite 1: Basement paragneiss

um: ultramafic rocks 3: Basal quartzite

Abbreviations for appendices 1 & 2: OTCP: outcrop; UTME/UTMN: easting and northing geographic coordinates; ST: transposition foliation; strk: strike; dp: dip; L: mineral or stretching lineation; plng; plunge; trnd: trend; F2: mature fold; F3: immature fold; Shear: macroscopic shear zone or shear band; top-to-...: sense of shear; Tect: type of tectonite.

Appendix 3. Rb-Sr and Sm-Nd isotopic analyses for felsic and mafic intrusions from basement level of Frenchman Cap dome.

Specimen	Rock type	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$ calc. ^a	$^{87}\text{Sr}/^{86}\text{Sr}$ m ^b	$2\sigma^c$ $\pm$	Sm ^d ppm	Nd ^d ppm	$^{147}\text{Sm}/^{144}\text{Nd}$ corr. ^e	$^{143}\text{Nd}/^{144}\text{Nd}$ corr. ^e	$2\sigma^c$ $\pm$	T Dm ^f Ma
Granite												
FG425b	Bourne granite	216	165	3.83	0.816901	18	8.25	58.73	0.084948	0.511012	14	2471
JC33	Leucogranite	33	354	0.27	0.770784	18	10.01	43.52	0.139092	0.511771	14	2718
JC50	Bourne granite	155	212	2.13	0.788665	18	14.20	74.88	0.114660	0.511409	14	2604
Mafic rocks												
Fg444	Frisby diabase	29	186	0.45	0.717292	18	2.91	11.23	0.156814	0.512377	13	1960
FG475	clinopyroxenite	13	154	0.24	0.7108377	18	2.31	8.78	0.159011	0.512059	14	2910
FG478d	Frisby diabase	59	405	0.42	0.714496	18	4.73	22.49	0.127070	0.511684	14	2497

^a Calculated using ppm concentration from XRF analysis. ^b Measured. ^c Uncertainties are at the fifth and sixth decimal place. ^dCalculated from a $^{148}\text{Nd}/^{149}\text{Sm}$ spike. ^e Corrected for the ^{143}Nd and ^{144}Nd contained in the spike. ^f Calculated assuming present-day values of $^{143}\text{Nd}/^{144}\text{Nd} = 0.513114$ and $^{147}\text{Sm}/^{144}\text{Nd} = 0.214000$.

Appendix 4. Major and trace elements composition of mafic intrusions in the core of Frenchman Cap dome				
Sample #	FG-187	FG-444	FG-475	FG-478d
Rock type	Bourne granite	hornblende diabase	Clino-pyroxenite	hornblende diabase
Locality	Bourne Glacier	Frisby Creek	Perry River	Perry River
UTM NAD83 Zone 11	390387E 586506N	393897E 5677540N	390584E 5675488N	390590E 5675653N
Major (wt%)				
SiO ₂	70.44	48.52	47.67	48.15
Al ₂ O ₃	13.77	14.84	6.31	15.57
CaO	1.73	10.83	14.39	9.04
Fe ₂ O ₃ (T)	3.68	13.35	11.28	13.19
K ₂ O	5.67	0.54	0.43	1.58
MgO	0.43	7.48	16.16	6.42
MnO	0.04	0.20	0.19	0.17
Na ₂ O	2.6	2.16	1.02	3.18
P ₂ O ₅	0.07	0.11	0.09	0.25
TiO ₂	0.33	1.31	0.62	1.32
L.O.I.	0.30	1.40	2.90	2.50
Sum	99.04	99.44	98.37	99.06
Trace (ppm)				
Ba	948	102	94	487
Ce	333	16	0	27
Co	6	47	58	49
Cr	12	95	950	53
Ga	20	14	8	12
La	188	10	18	15
Nb	25	9	1	8
Nd	120	16	11	16
Ni	17	149	250	142
Pb	47	10	15	15
Rb	225	29	13	59
Sr	153	186	154	405
Th	65	9	9	7
U	7	4	7	2
V	19	281	290	250
Y	25	27	19	25
Zn	57	90	71	106
Zr	355	83	43	119