U-Pb geochronology in Frenchman Cap dome of the
Monashee complex, southern Canadian Cordillera:
early Tertiary tectonic overprint of a Proterozoic history

by

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of the requirements for the degree of
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ABSTRACT

The Monashee complex lies beneath a crustal-scale ductile thrust fault, the Monashee décollement, in a deep structural exposure in the southern Canadian Cordillera. Paleoproterozoic basement gneisses in the complex, the largest exposure of such rocks in the Cordillera, record a Paleoproterozoic geologic history that was partly overprinted by tectonism related to early Tertiary Cordilleran orogenesis. The gneisses were intruded into by granitoids and unconformably overlain by a metasedimentary-dominated cover sequence. Pelitic schists in the cover sequence, which was only affected by Cordilleran tectonism, display an inverted metamorphic sequence denoted by sillimanite - K-feldspar assemblages in the structurally highest rocks and staurolite-kyanite-muscovite assemblages in deeper rocks. U-Pb geochronologic data and field relationships from 39 samples from Frenchman Cap dome in the northern part of the complex are used to investigate the following: (i) the Paleoproterozoic geologic history, (ii) the deposition age of the cover sequence, (iii) the age and nature of Cordilleran tectonism.

The Paleoproterozoic history includes six intrusive episodes, two metamorphic events, two depositional sequences, and one deformation event. The most voluminous intrusive episode occurred at 2.10-2.08 Ga. These rocks and an older paragneiss sequence underwent the most significant Paleoproterozoic metamorphism and deformation, which are recorded by ~2.06 Ga monazite in pelitic schist and a migmatitic gneissosity that is interpreted as coeval based on the intrusive relationships of an amphibolite dyke. Other intrusive episodes occurred at ~2.27, ~1.97, 1.86, 1.85-1.83, and 1.76 Ga. Although interpretation of precise U-Pb ages is hindered by Pb loss and/or overgrowth that occurred in the dated minerals during the Cordilleran overprint, there is sufficient age resolution on the intrusive events to conclude that the Monashee complex has affinities with the western Canadian Shield and other Paleoproterozoic exposures in the Cordillera.

Rocks in the lower part of the cover sequence are dated at 1.86-1.85 Ga, making them the oldest known metasedimentary rocks in the Cordilleran miogeocline. The upper part of the cover sequence is dated at younger than ~1.21 Ga. Detrital zircons from throughout the sequence suggest a source in the western Canadian Shield rather than underlying basement. The thickness of Neoproterozoic metasedimentary rocks in the Monashee complex is considerably less than that outside the complex in the hanging wall of the Monashee décollement, suggesting that deposition above and below the décollement occurred in different parts of the Cordilleran miogeocline.

U-Pb isotopic data show that the age and intensity of Cordilleran metamorphism and deformation vary with structural level, over a thickness of 6 km. Monazite U-Pb dates that young downward from ~64 Ma in the highest level to ~49 Ma in the deepest level are interpreted as representing diachronous peak metamorphism. Modeling of Pb

loss data from Proterozoic and Cretaceous monazite and titanite indicates that deeper levels were at elevated temperatures ($\sim 700^{\circ}\text{C}$, as suggested by pelitic mineral assemblages and a petrogenetic grid) for less than 1 Ma and higher levels were at equivalent or moderately greater temperatures ($\sim 50^{\circ}\text{C}$ hotter) for up to a few million years longer. The thermal peak in higher levels was (i) synchronous with thrusting on the Monashee décollement, (ii) immediately followed by a period of rapid cooling that occurred during initial motion on overlying normal faults, which are associated with exhumation of the complex, and (iii) attained while deeper levels were at lower temperatures but presumably heating toward a thermal peak that occurred a few million years later during the main phase of extension on overlying faults. Dating of intrusive rocks with various relationships to the dominant Cordilleran deformation shows that deformation was synmetamorphic and lessens in intensity (a factor related to the amount of transposition) downward within basement rocks in deeper levels, locally having no effects.

The U-Pb data that suggest metamorphism was less intense with increasing structural depth agree with the mineral assemblage data by showing an inverted metamorphic sequence. Evidence for the associated thermal gradient being inverted was possibly revealed in deeper levels, where the pattern of downward decreasing metamorphic intensity exists in basement rocks that are known to have been only weakly strained since the Proterozoic. In higher levels, it is uncertain whether the thermal gradient was inverted because there are numerous east-directed structures capable of significant synmetamorphic displacements, such as kilometre-scale isoclinal folds with sheared limbs, that could have telescoped a once laterally extensive normal metamorphic sequence into an inverted sequence without ever placing warmer rocks over cooler ones. The resulting tectonic model contends that the inverted metamorphic sequence formed mainly by the juxtaposition of rocks with widely varying initial positions that reached peak temperatures at different times and in different positions within the orogen, and perhaps there was a minor component of local downward heat conduction.

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ORIGINAL CONTRIBUTION

During the six months of fieldwork in 1993-1995, I worked with a mapping partner in the following 1:20000 map sheets: 82M.028, 037, 038, 046, 047, 048, 056, 057. The work mostly consisted of collecting samples for U-Pb geochronology and making detailed observations of structural fabrics, metamorphic mineral assemblages, and intrusive relationships. Zircon, monazite, and titanite from 39 samples were dated by U-Pb geochronology. I completed nearly all of the geochronology, from sample collection through age interpretation, on the ~300 dated analyses. The parts that I did not complete are a minor amount of the mass spectrometry, which was performed by a technician at the Geological Survey of Canada in Ottawa, and a minor amount of the U-Pb chemistry and most of the general laboratory maintenance, which were performed by Dr. Sharon Carr at Carleton University.

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CHAPTER 1

U-PB GEOCHRONOLOGIC CONSTRAINTS ON THE COVER SEQUENCE OF THE MONASHEE COMPLEX, CANADIAN CORDILLERA: PALEOPROTEROZOIC DEPOSITION ON BASEMENT

ABSTRACT

The cover sequence in the Monashee complex is a platformal metasedimentary succession that occupies a nearly unique position in the Canadian Cordillera due to its unconformable contact with exposed crystalline basement. Zircon U-Pb data and field relationships show that the lower part of the sequence contains Paleoproterozoic rocks, the oldest known metasedimentary rocks in the Cordilleran miogeocline, and the upper part of the sequence is Mesoproterozoic or younger. Maximum age constraints on the lower part are provided by 1.99 Ga detrital zircons from the basal unit and a 1862 ± 1 Ma orthogneiss upon which it was presumably deposited. Minimum age constraints are provided by rocks that intruded into the lower part: 1852 ± 4 Ma pegmatite, 1762 ± 6 Ma leucogranite, and 724 ± 5 Ma syenitic gneiss. The upper part of the sequence must be considerably younger than the lower part because it contains a detrital zircon dated at ~ 1.21 Ga. Other detrital zircons, dated at Neoproterozoic (2.95-2.86 Ga) and Paleoproterozoic (1.85-1.81, 1.75 Ga), suggest a source in the western Canadian Shield. These ages constrain the thickness of Meso- and Neoproterozoic metasedimentary rocks

in the cover sequence to be <2 km. Combining these ages with previously interpreted Paleozoic deposition ages for the middle and upper parts of the sequence constrains the thickness to be <0.2 km, considerably less than coeval rocks above the Monashee complex in the hanging wall of the Monashee décollement. Such a contrast suggests that deposition above and below the décollement occurred in different parts of the Cordilleran miogeocline.

INTRODUCTION

The cover sequence in the Monashee complex is a 2-3 km thick, laterally extensive succession of platformal metasedimentary and subordinate metaigneous rocks that unconformably overlies a rare exposure of crystalline basement in the southern Canadian Cordillera (Figs. 1-1 to 1-5; Wheeler 1965; Reesor and Moore 1971; McMillan 1973; Psutka 1978; Höy and Brown 1980; Read 1980; Journeay 1986; Höy 1987; Scammell and Brown 1990). Knowledge of the deposition age and thickness of units in the cover sequence is necessary for correlation with rocks above the complex in the hanging wall of a major thrust fault, which could assist in restoration of the Cordillera to a pre-shortening configuration. This knowledge is also necessary for understanding the deposition history of the west margin of the Canadian Shield, a prerequisite for Proterozoic plate reconstruction. The only means of confidently constraining the deposition age of the middle and upper amphibolite facies metasedimentary rocks in the cover sequence is by U-Pb dating of granitoid rocks that intruded into them, detrital minerals, and the basement gneiss upon which they were deposited. Previous dating places broad constraints on the

age of the sequence; the lower part was deposited in the Proterozoic (Ross and Parrish 1991; Parrish and Scammell 1988) and the upper part was deposited in the Paleozoic (Höy and Godwin 1988), though a paucity of data makes this interpretation uncertain.

In this chapter, zircon U-Pb data and field relationships are presented from (i) rocks that intruded into the lower part of the Monashee cover sequence, (ii) detrital grains from the lower and upper parts of the sequence, and (iii) the youngest known basement gneiss upon which the sequence was deposited. These constraints on the deposition age of the cover sequence are used to make correlations with metasedimentary rocks above the Monashee complex. Implications for the restoration of the Cordillera to a pre-shortening configuration are also discussed.

GEOLOGIC SETTING

The Monashee complex lies in the southern Omineca Belt, the metamorphic and plutonic hinterland to the Canadian Cordillera that developed during collision between accreted terranes and the North American craton (Monger et al. 1982). The complex is composed of basement gneiss and overlying metasedimentary and metaigneous rocks (the cover sequence) that are exposed in the tectonic window beneath the Monashee décollement (Figs. 1-2, 1-3; Read and Brown 1981; Brown et al. 1986; Journeay 1986), a ductile crustal-scale thrust fault correlated with the sole thrust of the Rocky Mountain Foreland Belt (Brown et al. 1992). Palinspastic restoration of the Selkirk allochthon in the hanging wall indicates up to ~300 km of northeastward displacement relative to the North American craton (Brown et al. 1993), with the amount of displacement relative to the

Monashee complex being less well constrained, but estimated at more than 80 km (Read and Brown 1981).

Crystalline basement gneiss forms the core of the two structural culminations in the Monashee complex (Fig. 1-2), namely, Frenchman Cap dome in the north (Wheeler 1965; Höy and Brown 1980; Journeay 1986) and Thor-Odin dome in the south (Reesor and Moore 1971; Read 1980). Frenchman Cap basement is dominated by kilometre-wide granitic orthogneiss bodies dated at $\sim 2.27\text{--}2.07$ Ga (Fig. 1-4; Armstrong et al. 1991; Chapter 2). Orthogneiss dated at 1.86 Ga (this chapter) and orthogneiss in Thor-Odin dome dated at 1.93 and ~ 1.87 Ga (Parkinson 1991) are also considered to be basement because rocks of this age only exist beneath the cover sequence. A 0.5 km thick paragneiss layer overlies the basement orthogneiss, and thinner paragneiss horizons are interlayered at deeper levels (Figs. 1-4, 1-5). The paragneiss is considered to be basement because it lies beneath the cover sequence, in a relationship interpreted as unconformable (McMillan 1973; Journeay 1986), and because it is thought to have been intruded by the basement orthogneiss (Journeay 1986). There are three other Paleoproterozoic basement exposures in the Canadian Cordillera (Fig. 1-1): the ~ 1.85 Ga Sifton Ranges orthogneiss in north-central British Columbia (Evenchick et al. 1984), 1.99 and 1.87 Ga orthogneiss in the Malton complex (McDonough and Parrish 1991), located 80 km north of the Monashee complex, and the ~ 2.09 Ga Gold Creek orthogneiss (Murphy et al. 1991), located 10 km northwest of the Malton complex. The Gold Creek gneiss is the only other exposure in unequivocal depositional contact with a cover sequence, a quartzite- and

marble-dominated succession that Murphy et al. (1991) interpreted as being unconformably overlain by the Neoproterozoic Windermere Supergroup.

Zircon, monazite, and titanite U-Pb data indicate that high grade tectonism in Monashee basement occurred at ~2.06 Ga and during Cordilleran orogenesis in the early Tertiary (Chapters 2, 3, 4). The cover sequence was strongly affected by Cordilleran tectonism (Parrish 1995; Chapters 3, 4), during which the rocks attained middle and upper amphibolite facies (with a typical migmatitic assemblage of kyanite-sillimanite-muscovite in the pelitic rocks) and deformation resulted in kilometre-scale isoclinal folds, a transposition fabric, and penetrative mineral and stretching lineations (Journey 1986 and references therein).

MONASHEE COVER SEQUENCE

The Monashee cover sequence is a 2-3 km thick succession of laterally extensive siliciclastic and carbonate metasedimentary rocks and minor metaigneous rocks (Fig. 1-4 to 1-6; Wheeler 1965; Reesor and Moore 1971; McMillan 1973; Psutka 1978; Read 1980; Höy and Brown 1980; Journey 1986; Höy 1987; Scammell and Brown 1990). Although there has been intense deformation and metamorphism in the sequence, correlation of rock units at the local scale and across the complex was achieved with the use of distinctive and laterally continuous (up to 150 km along strike) horizons of quartzite and marble, stratified Pb-Zn layers, and stratiform carbonatite. The contact between the cover sequence and basement paragneiss was interpreted as unconformable, largely because of the continuity of the basal unit in the sequence and the presence of a thin (5-10 m)

paleoweathering horizon that is gradational between the sequence and basement (McMillan 1973; Journeay 1986). There are no exposed contacts between the cover sequence and basement orthogneiss due to the intervening paragneiss layer, and thus previous workers could only infer that the sequence is younger based on the absence of orthogneiss layers within it. This inference was confirmed by U-Pb dating which shows that detrital zircons from the basal unit (Ross and Parrish 1991; data presented below) are indeed younger than the largest basement orthogneiss bodies (Armstrong et al. 1991; Chapter 2).

Journeay (1986) defined three upward-fining depositional assemblages in the cover sequence of northern Frenchman Cap dome that are herein referred to as the lower, middle, and upper parts (Fig. 1-6). Each part consists of (i) a basal unit of quartzite interlayered with upward increasing amounts of semipelitic and calcareous schist; (ii) an intermediate unit of intercalated pelitic and semipelitic schist, and calc-silicate gneiss; and (iii) an upper unit of intercalated pelitic schist, quartzite, calc-silicate and amphibolitic gneiss, and regionally continuous thin marker horizons of pure and impure marble. The contacts between the units were interpreted as conformable because of their gradational nature. The nature of the contacts between the parts could not be determined, although it was noted that they are sharp but not demonstrably erosive. No unconformities have been mapped in the sequence by previous workers, largely because contact relationships do not require drastic changes in depositional setting, thick conglomerate or regolith units do not exist above the basal unit, and thin (metre-wide) marker units throughout the sequence

have great lateral continuity. Journeay concluded that the facies patterns fit a model of episodic subsidence, onlap, and stabilization of an open, shallow-marine shelf.

The middle and upper parts of the sequence in the northernmost part of the dome record synsedimentary rifting in a magmatically active environment (Scammell and Brown 1990). Extensional tectonism was inferred from the mature and immature siliciclastic metasedimentary rocks, with abrupt facies changes, that are intercalated with ultramafic and mafic sills and flows, as well as minor felsic pyroclastic deposits. The oldest evidence of syndepositional magmatism is a laterally extensive (~100 km along strike) stratiform pyroclastic carbonatite of unknown age that is correlated to the south with carbonatite in the middle part of the sequence (McMillan 1973; Psutka 1978; Journeay 1986; Höy 1987 and references therein).

The lower part of the sequence was interpreted to be Neoproterozoic or older based on the inferred intrusive relationships of the Mount Copeland syenitic orthogneiss, dated at 740 ± 36 Ma (Parrish and Scammell 1988). The position of this and other previously dated rocks is shown in the schematic stratigraphic column in Fig. 1-6. The middle part of the sequence was interpreted to be Cambrian based on a Pb-Pb model age on galena from a stratiform Pb-Zn horizon (Höy and Godwin 1988) that forms a laterally extensive (~100 km along strike) and narrow (metre-wide) marker layer (Höy 1987 and references therein). The base of the upper part of the sequence must be older than a $359 \pm 46/-43$ Ma discordant intrusive carbonatite (Parrish 1995). At a higher stratigraphic level in the upper part of the sequence, a 541 ± 11 Ma amphibolite (Parrish 1995) is separated by a few tens of metres from a pyroclastic metavolcanic rock dated at 388 ± 1 Ma or

younger (Scammell and Parrish 1993). Uncertainties concerning these samples include whether (i) the metavolcanic rock overlies the amphibolite, or vice-versa (R.R. Parrish personal communication 1995); (ii) the protolith of the amphibolite is intrusive or volcanic, and (iii) the one zircon grain dated at 388 Ma in the metavolcanic rock is an inherited grain or one that crystallized during extrusion; nine other zircons from this rock range in age from 2.9 to 1.7 Ga and were interpreted as xenocrysts (Parrish and Scammell 1993), and seven single-grain zircon analyses from a nearby pyroclastic metavolcanic rock yield similar ages (D.H. Gibson personal communication 1996).

GEOCHRONOLOGY

U-Pb geochronology was performed on seven samples from the Pettipiece Pass area of Frenchman Cap dome that constrain the deposition age of the Monashee cover sequence. Sample localities are shown on the map in Fig. 1-4 and on the schematic stratigraphic column in Fig. 1-6. Analyses are shown on concordia diagrams in Fig. 1-7 and in Table 1-1. Sample descriptions and interpretations of the U-Pb data are presented in this section, which is followed by discussion of possible correlations between metasedimentary rocks in the cover sequence and those outside the complex.

Analytical methods

The U-Pb geochronology followed procedures outlined by Parrish et al. (1987): microcapsules for mineral dissolution (Parrish 1987), a mixed ^{233}U - ^{235}U - ^{205}Pb tracer

(Parrish and Krogh 1987), multicollector mass spectrometry (Roddick et al. 1987), and estimation of errors using numerical error propagation (Roddick 1987). Decay constants used are those recommended by Steiger and Jäger (1977). Discordia lines through analyses were calculated with the use of a modified York (1969) regression (Parrish et al. 1987), which takes scatter into account. Mineral selection and most of the U-Pb chemistry were performed in the geochronology laboratory at Carleton University. All zircons were abraded (Krogh 1982). For zircon, monazite, and xenotime chemistry, U blanks were less than 5 pg and Pb blanks were usually 5-20 pg. For titanite chemistry, U blanks were less than 5 pg and Pb blanks were ~50 pg. Isotopic ratios were determined on a MAT 261 mass spectrometer at the Geological Survey of Canada in Ottawa.

Intrusive rocks

Sample 187 is from the Kirbyville gneiss, a hornblende biotite granodioritic orthogneiss north of Blais Creek (Fig. 1-4). It is one of many orthogneiss bodies that form the core of the Kirbyville anticline, an overturned, kilometre-scale isocline lying west of the main exposure of Frenchman Cap basement (Höy and Brown 1980; Journeay, 1986). The strongly deformed nature of the contacts between the bodies does not permit determination of intrusive relationships. Höy and Brown (1980) mapped the orthogneiss bodies as basement because they are widespread beneath the cover sequence and do not exist within it. A thin layer of paragneiss separates the gneiss from the basal quartzite unit of the cover sequence. The Kirbyville gneiss was dated because of the interpretation that it is basement, requiring that the cover sequence be younger. Four zircon analyses (one to

three grains each) and three multigrain titanite analyses (Table 1-1) define a discordia line with upper and lower intercepts of 1862 ± 1 and 63 ± 14 Ma, respectively (MSWD = 2.3) (Fig. 1-7a). Zircon discordance is 7-14% and titanite discordance is 94-96%, as measured from the upper intercept along the discordia. The upper intercept is the interpreted igneous crystallization age, and the lower intercept coincides with the age of Cordilleran metamorphism that was determined from monazite and zircon U-Pb dates from nearby samples (Parrish 1995; Chapter 3). The discordance is attributable to Pb loss during this thermal overprint.

Sample 341 is from a 2-3 m thick pegmatite layer northeast of Pettipiece Pass, on the east side of the dome (Fig. 1-4). It is boudinaged and is concordant with the main deformation fabric and lithologic layering in the host amphibolitic gneiss (Fig. 1-8). It presumably intruded into the host, which was mapped by Psutka (1978) as stratigraphically high in the lower part of the cover sequence (Fig. 1-6), and appears to have been subjected to the entire deformation history. The host is confidently considered to be part of the cover sequence because it is separated from basement by the basal quartzite unit of the cover sequence. Also, mapping of marker horizons by Psutka shows that the cover sequence in this part of the dome is a homoclinal section, which rules out the possibility of isoclinal folding of basement into the sequence. Sample 341 was dated because of the interpretation that the host rocks and stratigraphically underlying rocks are older. Four single-grain zircon analyses (A, D, E, F) and an analysis composed of two grains (C; Table 1-1) define a discordia line with upper and lower intercepts of 1852 ± 4 and 149 ± 20 Ma, respectively (MSWD = 9.8, discordance is 19-46%; Fig. 1-7b).

Another single-grain zircon analysis (B) lies slightly above the discordia for unknown reasons and is not used in the regression. Three single-grain xenotime analyses (Table 1-1) lie slightly below concordia between 58 and 55 Ma (Fig. 1-7b) and one single-grain monazite analysis (Table 1-1) is concordant at ~51 Ma. The upper intercept of the discordia is interpreted as the pegmatite crystallization age, and the lower intercept is interpreted as mainly reflecting Pb loss and/or metamorphic overgrowth during Cordilleran metamorphism. However, the lower intercept is significantly older than the early Tertiary monazite and xenotime U-Pb analyses from this sample and nearby samples that are interpreted to represent the age of Cordilleran metamorphism (Parrish 1995; Chapter 3). This may suggest that the oldest analyses on the discordia contained minor amounts of inherited material prior to the overprint, resulting in upper and lower intercepts that are slightly older than reality (the accuracy of the lower intercept would be affected to a greater degree because the analyses plot much closer to the upper intercept). The zircon discordance is mostly attributed to Pb loss rather than metamorphic overgrowth because abrasion (Krogh 1982) removed the outer part of the grains, the grains had igneous-looking crystal faces, and a rim-core relationship was not visible with a microscope. Ross and Parrish (1991) arrived at this conclusion for detrital zircons from the southern Omineca Belt that attained a similar metamorphic grade, largely on the basis of a lack of visible metamorphic rims.

The upper intercept is interpreted as a crystallization age rather than the age of inherited material (with crystallization occurring at the lower intercept age) because that would require all zircons to contain a significant amount of inherited material (composing

55-80% of the Pb in the grains) of a single ~ 1.85 Ga age. This is considered unlikely based on comparison with nearby pegmatite and leucogranite layers of early Tertiary age (Chapters 3, 4). Three of eight early Tertiary samples contain zircons with obvious inherited material, which varies widely in age (from ~ 2.9 to 1.7 Ga) and constitutes $<50\%$ (usually $<5\%$) of the Pb in the analyses; such characteristics contrast with those from sample 341. The concordant and nearly concordant analyses of early Tertiary xenotime and monazite are interpreted as having grown during Cordilleran metamorphism.

Sample 180 is from a 0.2 m thick quartz-rich leucocratic layer south of Ratchford Creek and west of Myoff Creek, in the core of the Kirbyville anticline (Fig. 1-4). Biotite is distributed throughout the rock and it also exists in thin (1-2 mm wide) biotite-rich seams within which kyanite of presumed metamorphic origin exists. The layer was interpreted in the field to be an intrusive rock on the basis of its leucocratic appearance and the existence of feldspar within it. The sample was collected in the hinge of an outcrop-scale isoclinal fold defined by the layer, in a locality ~ 300 m structurally below the Monashee décollement, in the zone of maximum metamorphism (upper amphibolite facies) and deformation in the study area. It presumably intruded into the host, which was mapped by Journeay (1986) as pelitic schist in the lower part of the cover sequence (Fig. 1-6), and appears to have been subjected to all deformation that affected the host. Sample 180 was dated because of the interpretation that the host rocks and stratigraphically underlying rocks are older. Zircon has two basic morphologies, elongate (aspect ratio of $\sim 3.5:1$) and subequant (aspect ratio of $\sim 1.5:1$), with most grains belonging to these endmembers but some of intermediate shape. Five analyses (F, G, H, K, L) of elongate morphology zircon

(one to nine grains each; Table 1-1) define a discordia line with upper and lower intercepts of 1762 ± 6 and 76 ± 20 Ma, respectively (MSWD = 2.8, discordance is 24-38%; Fig. 1-7c). Two other analyses (B, E) of elongate morphology zircons (five grains each) lie slightly below the discordia and are interpreted as containing minor amounts of inherited material. Five analyses (A, C, D, I, J) of subequant morphology zircons (one to five grains each) lie further below the discordia and are interpreted as containing significant inheritance. Four monazite analyses (one to three grains each; Table 1-1) have reverse discordance between 64 and 60 Ma ($^{207}\text{Pb}/^{235}\text{U}$ ages) (Fig. 1-7c). The upper intercept of the discordia defined by the elongate morphology zircons is interpreted as the leucogranite crystallization age, and the lower intercept is interpreted as approximately the time of Pb loss and/or metamorphic overgrowth during Cordilleran metamorphism, for the reasons given for sample 341. The monazites are interpreted as having grown during Cordilleran metamorphism.

Sample 207 is from the Mount Grace gneiss, a syenitic orthogneiss that forms a concordant, ~12 km long and up to 1 km thick body in the overturned limb of the Mount Grace syncline (Journeay 1986; Höy 1987). The syncline is a kilometre-scale isocline on the west side of Frenchman Cap dome (Fig. 1-4; Höy and Brown 1980; Journeay 1986; Höy 1987). The sample is mostly composed of quartz, K-feldspar, biotite, and hornblende, and it was collected atop Mount Grace, ~100 m structurally above the lower contact of the body. This contact is sharp and concordant with gneissosity in the body and with west-dipping lithologic layering and gneissosity in the host amphibolitic gneiss. The body is thought to have been subjected to the entire deformation history because the

gneissosity and stretching lineation within it are parallel with the dominant regional fabrics and the lack of intrusive contacts suggests that it was highly strained. The protolith is interpreted as intrusive into the host amphibolitic gneiss, which was mapped by Journeay (1986) and Höy (1987) as stratigraphically high in the lower part of the sequence (Fig. 1-6), despite the absence of intrusive contacts. The host is confidently considered to part of the cover sequence because it lies adjacent to marker horizons in the middle part of the cover sequence and it is separated from basement by the basal quartzite unit of the cover sequence. Journeay (1986) and Höy (1987) placed thrust faults below and above the body, respectively, but no evidence for faulting was found during this study (similar host gneiss exists above and below the orthogneiss body and there are no strain gradients near its margins). The body is located 0.2 km below (in a stratigraphic sense) a stratiform pyroclastic carbonatite horizon and a stratiform Pb-Zn horizon with a Cambrian Pb-Pb model age (Höy and Godwin 1988), both of which are markers in the middle part of the sequence. The Mount Grace gneiss was dated because of the interpretation that the host gneiss and stratigraphically underlying rocks are older, and to establish a more precise age for alkaline magmatism in the complex.

Zircons are slightly elongate, euhedral, typically cloudy in the core and clear in the rim. Five analyses (one to seven grains each; Table 1-1) define a discordia line with upper and lower intercepts of 724 ± 5 Ma and 53 ± 3 Ma, respectively (MSWD = 1.3, discordance is 55-85%; Fig. 1-7d). The upper intercept is interpreted as the crystallization age of the syenitic protolith for the reasons given for sample 341 and for the reasons given for a similar rock at a similar stratigraphic level in the southern part of Frenchman Cap

dome, the Mount Copeland syenitic orthogneiss (Parrish and Scammell 1988). The Mount Copeland gneiss has zircon analyses that define a discordia with upper and lower intercept ages of 740 ± 36 and 62 ± 1 Ma, respectively, and two other zircon analyses are concordant at 59 ± 1 Ma. Like the analyses in the Mount Grace gneiss, the discordant analyses from the Mount Copeland gneiss plot much closer to the lower intercept. Parrish and Scammell interpreted the upper intercept as the igneous crystallization age because the amount of discordance in each analysis was correlated with the amount of visible core material (corresponds with cloudiness in the core), suggesting that ~ 59 Ma metamorphic zircon overgrowth was responsible for the discordance. Overgrowth is also interpreted to be largely responsible for the discordance in the analyses from the Mount Grace gneiss because the same relationship is observed.

Metasedimentary rocks

Detrital zircons were dated to ascertain source ages, the youngest of which is a maximum deposition age. The analyses are 2-23% discordant. In order to obtain upper intercepts that are interpreted as approximate crystallization ages, the analyses are individually regressed through forced lower intercepts of 60-50 Ma, which is generally the time of peak Cordilleran metamorphism (Parrish 1995; Chapter 3). The forced regressions yield reliable source ages if the zircons lacked inherited material and they suffered Pb loss and/or overgrowth only in the early Tertiary. Zircon, monazite, and titanite data from 1.86-1.83 Ga rocks in Frenchman Cap dome (Armstrong et al. 1991; Parrish 1995; Chapter 2) are pertinent to this problem because they are coeval with most of the detrital

zircons. These data show that there was a single episode of Pb loss and/or overgrowth and it occurred at ~60-50 Ma (discordias generally have lower intercepts within uncertainty of this age), and inheritance is a minor problem (3 of 26 zircon analyses have substantial inheritance, excluding one suite of granite dykes). It is therefore likely that the forced regressions yield reliable source ages if the Paleoproterozoic zircons were derived from the underlying rocks or the western Canadian Shield (as suggested by age populations, see below), where no post-Paleoproterozoic high temperature events have been documented (Villeneuve et al. 1993). Forced regressions through Archean analyses are probably less accurate because it is likely that Pb loss and/or overgrowth occurred in the Paleoproterozoic in addition to the early Tertiary.

Sample 335 is from quartzite that forms the basal unit of the cover sequence, collected a few tens of metres above the unconformity with basement paragneiss in Pettipiece Pass (Fig. 1-4; Wheeler 1965; Journeay 1986). It lies above the metaconglomerate unit containing detrital zircons dated by Ross and Parrish (1991) (Fig. 1-4). Eight single-grain zircon analyses (Table 1-1) yield upper intercept age groups of 2.05-2.04 (n = 2), 2.02 (n = 1), and 2.00-1.99 Ga (n = 5) (Fig. 1-7e). Discordance is 2-11%.

Sample 300 is from pelitic schist on the east side of the dome, northeast of Pettipiece Pass (Fig. 1-4). Psutka's (1978) mapping places it in the base of the upper part of the sequence (Fig. 1-6), above a stratiform Pb-Zn horizon with a Cambrian Pb-Pb model age (Höy and Godwin 1988) and beneath Cambrian amphibolite (Parrish 1995). It lies in the least metamorphosed (staurolite-bearing assemblage in the pelitic rocks) and

least deformed part of the complex. Five single-grain zircon analyses (Table 1-1) yield upper intercept age groups of 2.86 ($n = 1$) and 1.84-1.81 Ga ($n = 4$) (Fig. 1-7e).

Discordance is 2-19%.

Sample 257 is from biotite feldspar semipelitic schist in the core of the Mount Grace syncline, south of Mount Grace (Fig. 1-4). Journeay's (1986) mapping places it in the base of the upper part of the sequence (Fig. 1-6), at about the same stratigraphic level as sample 300. Ten single-grain zircon analyses (Table 1-1) yield upper intercept age groups of 2.95-2.94 ($n = 2$), 2.88 ($n = 2$), 1.85-1.81 ($n = 4$), 1.75 ($n = 1$), and 1.21 Ga ($n = 1$) (Fig. 1-7f). Discordance is 5-23%. Significant discordance (23%) in the 1.21 Ga analysis leaves open the possibility that early Tertiary Pb loss and/or overgrowth was superposed on zircon that is a mixture between a grain with a crystallization age that is younger than 1.21 Ga and an older inherited core. Assuming that most discordance results from an early Tertiary overprint and that the age of any inheritance is 2.9-1.8 Ga (the dominant age of possible sources in the western Canadian Shield, Fig. 1-1) yields a minimum crystallization age of ~ 0.94 Ga. From these assumptions, it is surmised that the zircon is Grenville-age, which is a sufficient age resolution for the purposes of this detrital study. Monazites from this sample yield U-Pb analyses that are concordant at early Tertiary (Chapter 3).

DISCUSSION

Comparison of intrusion ages

The intrusive rocks dated in this chapter are the same age as intrusive rocks dated elsewhere in the Monashee complex and in the other Paleoproterozoic exposures in the Cordillera. The Kirbyville gneiss (sample 187) is coeval with ~ 1.87 Ga biotite granite gneiss in Thor-Odin dome (Parkinson 1991) and 1.87 Ga orthogneiss in the Malton complex (McDonough and Parrish 1991). Pegmatite sample 341 is coeval with 1848 ± 3 Ma granite (Armstrong et al. 1991; Chapter 2) and it is slightly older than three 1845-1834 Ma pegmatites that intruded into the underlying basement in Frenchman Cap dome (Chapter 2). Leucogranitic sample 180 is slightly younger than the Rimbey granites of the Alberta basement (Fig. 1-1; Villeneuve et al. 1993), and coeval zircons are found as (i) a xenocrystic component in pyroclastic metavolcanic rocks from northern Frenchman Cap dome (Scammell and Parrish 1993; D.H. Gibson personal communication 1996), (ii) a detrital component in sample 257 (see above), and (iii) an inherited component in three samples of pegmatite and leucosome generated during Cordilleran tectonism in the dome (Chapters 3, 4). The Mount Grace gneiss (sample 207) is coeval (within error) with the 740 ± 36 Ma Mount Copeland syenitic gneiss in the southern part of the dome (Parrish and Scammell 1988), the $736 +23/-17$ Ma Hugh Allen orthogneiss in the Malton complex (McDonough and Parrish 1991), and the $728 +9/-7$ Ma Deserters orthogneiss in north-central British Columbia (Evenchick et al. 1984).

Paleoproterozoic deposition

A conservative age estimate for deposition of the lower part of the Monashee cover sequence is derived from the youngest detrital zircons in the basal quartzite unit (sample 335) and the oldest pegmatite that intruded into the sequence (sample 341); deposition occurred between 1.99 Ga and 1852 ± 4 Ma. The presence of 1762 ± 6 Ma leucogranite (sample 180) in this part of the sequence is confirmation of Paleoproterozoic deposition. This deposition age is older than the Belt-Purcell Supergroup (Anderson and Davis 1995 and references therein), and thus the lower part of the Monashee cover sequence contains the oldest known metasedimentary rocks in the Canadian Cordilleran miogeocline (i.e., those deposited above crystalline basement). A more restricted age estimate for the lower part of the cover sequence is derived from the youngest rock upon which the sequence was presumably deposited (Kirbyville gneiss, sample 187) and the oldest pegmatite (sample 341); deposition occurred between 1862 ± 1 and 1852 ± 4 Ma. Supporting evidence for deposition during this more restricted interval comes from the lack of data in the cover sequence for a 1854 ± 13 Ma metamorphic event that was dated in the basement (Parrish 1995; Chapter 2). However, it is cautioned that U-Pb data from Paleoproterozoic rocks suggest this was not a high grade event, and therefore the lack of evidence for it in the sequence may be a result of a strong Cordilleran overprint on a weak to moderate Paleoproterozoic event. Other evidence for deposition during this more restricted interval is that 1874 ± 21 Ma biotite granite gneiss in Thor-Odin dome (Parkinson 1991) only exists beneath the sequence.

Detrital zircons from sample 335 (basal unit) were dated at 2.05-1.99 Ga, coeval with seven of the eight zircons from the underlying metaconglomerate (Ross and Parrish 1991) and younger than those dated in ~2.27-2.07 Ga orthogneiss bodies that dominate the underlying basement. The age population suggests that a source is one of the numerous domains in the Alberta basement that contain ~2.0 Ga rocks, such as the Buffalo Head and Thorsby terranes (Fig. 1-1; Villeneuve et al. 1993).

Mesoproterozoic to Paleozoic (?) deposition

The presence of a detrital zircon from sample 257 dated at ~1.21 Ga requires that the upper part of the cover sequence must have been deposited at a considerably younger time than the lower part. The maximum present thickness of rocks deposited between 1.85 (host of sample 341) and ~1.21 Ga (sample 257) is 1.0 km and the minimum thickness is unconstrained. Deformation certainly thinned the cover sequence, but major omissions due to deformation are unlikely, given the great lateral continuity of marker horizons. Rocks that are coeval with the ~1.5-1.24 Ga Belt-Purcell Supergroup (Anderson and Davis 1995 and references therein) could exist in this interval. No unconformities have been recognized here, but one probably exists, given that intrusion of 1.85 Ga pegmatite (sample 341) and 1.76 Ga leucogranite (sample 180) into the lower part of the sequence implies that it was overlain by at least a few kilometres of pre-1.85 Ga rocks that now do not exist.

The 2.95-2.86 and 1.85-1.75 Ga detrital zircon ages from the base of the upper part of the sequence (samples 257 and 300) fit in the bimodal distribution of Archean (3.3-

2.5 Ga) and Paleoproterozoic (1.86-1.74 Ga) ages that Ross and Parrish (1991) and Ross (1995) suggested is a "fingerprint" of sources in the Alberta basement (Fig. 1-1; Villeneuve et al. 1993). A possible source for the Neoarchean zircons in the sequence is the Medicine Hat Block of the Alberta basement. The Paleoproterozoic zircons are coeval with granitoids that intruded into the underlying basement (Armstrong et al. 1991; Chapter 2) and the lower part of the sequence (sample 341). However, it is unlikely that rocks in the Monashee complex are a source because the detrital population lacks grains that are coeval with those in ~2.27-2.07 Ga basement orthogneiss, the overwhelmingly dominant underlying rock. A possible source for these zircons is the Rimbey granites in the Alberta basement. A source for the ~1.21 Ga detrital zircon from sample 257 is unknown. Rocks with ages between 1.2 and 1.0 Ga, broadly referred to as "Grenville", are rare in western North America, yet the presence in the Cordillera of Grenville-age detrital grains (Ross et al. 1992 and references therein), xenoliths and/or xenocrysts in diatremes (Parrish and Reichenbach, 1991), and metamorphism (Anderson and Davis 1995) suggests orogenic activity of this age.

Thickness of Windermere Supergroup in the cover sequence

The Windermere Supergroup (Gabrielse and Campbell 1991 and references therein) is a succession of Neoproterozoic strata of great thickness (up to 7 km) and lateral extent (at latitude 52-53°N it stretches across the width of the Omineca Belt into the western main ranges of the Rocky Mountains; Figs. 1-2, 1-3). Knowledge of the thickness of the Windermere Supergroup in the Monashee cover sequence is necessary for

estimating the paleogeographic position of the sequence in the Cordilleran miogeocline relative to rocks above the complex in the Selkirk allochthon. Such estimates could potentially broadly constrain the amount of displacement on the crustal-scale thrust fault between the allochthon and the complex, Monashee décollement. Present geochronologic data from the upper part of the sequence only allow for an approximate thickness of Windermere rocks, as there are three scenarios, depending on the validity of interpretations that infer Paleozoic ages for the middle and upper parts of the sequence.

If it is assumed that the entire Windermere Supergroup is younger than the $728 \pm 9/-7$ Ma Deserters gneiss (Evenchick et al. 1984), upon which at least part of the Supergroup was deposited, then the setting of the 724 ± 5 Ma Mount Grace gneiss has implications for the thickness of Windermere rocks in the cover sequence. The protolith of the Mount Grace gneiss intruded high into the lower part of the sequence, in a position that is now 0.2 km from a stratiform Pb-Zn horizon in the middle part of the sequence with a Cambrian Pb-Pb model age. This Pb-Pb age and the stratigraphic relationships between the horizon and the host rocks were used by Höy and Godwin (1988) to contend that the middle part of the cover sequence was deposited in the Cambrian. In this scenario, the thickness of the Windermere Supergroup in the sequence must be <0.2 km. The 1852 ± 4 Ma pegmatite (sample 341) intruded into the east side of the dome at the same stratigraphic level as the Mount Grace gneiss, and therefore rocks in this interval must represent ~ 1.3 Ga of deposition. If an unrecognized unconformity exists here, then it is most likely beneath the quartzite-semipelitic schist unit at the base of the middle part of the sequence. The validity of this scenario depends on the Cambrian Pb-Zn horizon

having been formed during deposition of the host rocks. Höy (1987) gave compelling evidence for such an interpretation, but it is not conclusive due to the tectonized nature of these rocks.

A second scenario is based on the Devonian or younger pyroclastic metavolcanic rock in the upper part of the cover sequence (Scammell and Parrish 1993). In this scenario, rocks above the metavolcanic rock are Devonian or younger, which requires that the thickness of the Windermere Supergroup in the cover sequence is <1.5 km and deposition of the entire cover sequence must have occurred on a platform that was starved for sediment compared with the now up to ~ 10 km thick section of Mesoproterozoic to Devonian rocks in the Selkirk allochthon, assuming the sequence lacks major unconformities. The validity of this scenario depends on the pyroclastic metavolcanic rock being Devonian or younger, which is considered tenuous because the one single-grain zircon U-Pb analysis upon which the age is based has not been replicated.

A third scenario rejects the interpretations that require the middle and upper parts of the cover sequence to be Paleozoic; the age of these rocks remains unknown, except that the upper part is younger than ~ 1.21 Ga (age of youngest detrital zircon in sample 257). In this scenario, the Windermere Supergroup could be composed of the <2 km thick middle and upper parts of the sequence. Additional U-Pb dating of detrital grains and igneous rocks in the middle and upper parts is needed to resolve this issue.

Potential for Cordilleran restoration

The significant contrast between the thickness of the Windermere Supergroup in the Monashee cover sequence (certainly <2 km and possibly <0.2 km) and that in the Selkirk allochthon (up to 7 km) leads to the conclusion that Windermere rocks above and below the Monashee décollement were deposited in different parts of the Cordilleran miogeocline. This conclusion obviously assumes that thinning of the cover sequence during deformation related to Cordilleran orogenesis was similar to thinning of Windermere rocks in the allochthon. Determination of the difference in depositional settings could potentially be used to broadly restore the allochthon to its pre-shortening position relative to the complex, which is inferred to have always been part of the western Canadian Shield (Chapter 2) but not necessarily autochthonous. Such a restoration constrains the amount of displacement on the Monashee décollement.

Possible explanations for the thickness contrast include (i) the Monashee complex was a basement high within the Windermere basin, (ii) Windermere rocks in the allochthon were deposited outboard (west) or inboard (east) of the complex, or (iii) Windermere rocks were deposited north or south of the complex. The explanation invoking outboard deposition is preferred because Price and Mountjoy's (1970) restoration of the Rocky Mountain Foreland Belt places most of the Windermere basin to the west of the complex and the deep water nature of the Windermere rocks (Gabrielse and Campbell 1991 and references therein) suggests that deposition occurred outboard of the platformal metasedimentary rocks that compose the cover sequence. Such a setting requires that hundreds of kilometres of east-directed motion of the Selkirk allochthon displaced the east

edge of the Windermere basin to its present position in the western main ranges of the Rocky Mountains (Fig. 1-1). The possible explanations that invoke Windermere deposition to the east, north, or south of the complex are incompatible with deposition facies and known displacements.

SUMMARY

Zircon U-Pb data and field relationships show that the lower part of the Monashee cover sequence contains the oldest known metasedimentary rocks in the Canadian Cordilleran miogeocline; it was certainly deposited on crystalline basement between 1.99 Ga and 1852 ± 4 Ma, and less firm relationships indicate that it was deposited between 1862 ± 1 and 1852 ± 4 Ma. Detrital zircon ages from the basal unit suggest a source in the Alberta basement of the western Canadian Shield rather than underlying basement. A 1.21 Ga detrital zircon dated in the upper part of the sequence yields a maximum deposition age, and Neoarchean and Paleoproterozoic detrital zircons from this part suggest a source in the Alberta basement. If previous interpretations that require the middle and upper parts of the sequence to be Paleozoic are valid, then the thickness of Meso- and Neoproterozoic metasedimentary rocks in the sequence is <0.2 km. If these interpretations are invalid, then rocks of this age are only constrained to be <2 km thick. These thicknesses are considerably less than those of coeval Belt-Purcell and Windermere supergroup rocks above the Monashee complex in the hanging wall of the Monashee décollement, implying that deposition of rocks above and below the décollement did not occur in the same part of the Cordilleran miogeocline.

Table 1-1. U-Pb analytical data.

Analysis ^a	Wt. ^b U (μg)	U (ppm)	Pb ^{c*} (ppm)	²⁰⁶ Pb ^d ²⁰⁴ Pb (pg)	Pb ^e (pg)	²⁰⁸ Pb ^f (%)	²⁰⁶ Pb ^g ²³⁸ U	²⁰⁷ Pb ^g ²³⁵ U	Corr. coeff.	²⁰⁷ Pb ^g ²⁰⁶ Pb	²⁰⁷ Pb ^h ²⁰⁶ Pb age (Ma)	
180 (Fig. 1-7c) - Leucogranitic layer, kyanite-bearing and quartz-rich, 0.2 m thick, in hinge of isoclinal fold, intruded into lower part of cover sequence; located in Anstey Range, south side of pass between Anstey River and Myoff Creek at 2160 m, 378590E-5684970N (UTM zone 11, BCGS map sheet 82M.037)												
A	4,149,se	50	256.3	49.39	12187	11	13.3	0.1740 ± 0.09%	2.768 ± 0.10%	0.96	0.11538 ± 0.03%	1885.8 ± 1.1
B	5,105,el	34	367.4	79.89	7936	21	5.5	0.2158 ± 0.11%	3.184 ± 0.12%	0.95	0.10702 ± 0.03%	1749.2 ± 1.3
C	1,149,se	29	409.2	121.3	6596	29	12.8	0.2705 ± 0.16%	4.126 ± 0.17%	0.96	0.11064 ± 0.05%	1809.9 ± 1.7
D	2,105,se	19	376.9	83.25	13694	7	5.4	0.2168 ± 0.09%	3.599 ± 0.10%	0.94	0.12039 ± 0.03%	1962.0 ± 1.2
E	5,105,el	14	421.7	97.31	11206	8	6.6	0.2264 ± 0.09%	3.346 ± 0.10%	0.93	0.10718 ± 0.04%	1752.0 ± 1.4
F	6,105,el	15	355	72.93	7684	9	7.7	0.1992 ± 0.09%	2.925 ± 0.10%	0.91	0.10652 ± 0.04%	1740.6 ± 1.6
G	6,105,el	29	81.82	16.78	2559	11	12.3	0.1889 ± 0.16%	2.764 ± 0.17%	0.94	0.10613 ± 0.06%	1734.0 ± 2.1
H	9,105,el	26	277.6	57.07	11236	8	7.0	0.2009 ± 0.13%	2.947 ± 0.14%	0.98	0.10641 ± 0.03%	1738.7 ± 1.1
I	1,149,se	25	555	112.4	16224	11	1.8	0.2085 ± 0.09%	3.168 ± 0.10%	0.96	0.11021 ± 0.03%	1802.9 ± 1.1
J	1,149,se	20	407.1	88.19	3126	34	4.3	0.2173 ± 0.09%	3.274 ± 0.10%	0.91	0.10927 ± 0.04%	1787.3 ± 1.6
K	1,105,el	10	266.3	64.7	5814	7	8.6	0.2332 ± 0.09%	3.439 ± 0.10%	0.92	0.10694 ± 0.04%	1747.9 ± 1.5
L	1,105,se	8	523.7	107.6	6171	9	4.4	0.2065 ± 0.09%	3.032 ± 0.10%	0.94	0.10652 ± 0.04%	1740.8 ± 1.3
1	3,149	54	5303	118.7	3646	50	59.6	0.009983 ± 0.15%	0.06437 ± 0.17%	0.94	0.04676 ± 0.06%	37.2 ± 2.7
2	3,105	17	7945	148.9	3354	25	52.8	0.009782 ± 0.13%	0.06311 ± 0.15%	0.89	0.04680 ± 0.07%	38.8 ± 3.2
3	1,149	24	5201	125.8	3426	22	64.0	0.009626 ± 0.10%	0.06176 ± 0.13%	0.87	0.04654 ± 0.06%	25.5 ± 3.1
4	2,105	16	4688	105.2	3253	14	62.0	0.009434 ± 0.10%	0.06067 ± 0.16%	0.72	0.04664 ± 0.11%	30.8 ± 5.4
187 Kirbyville gneiss (Fig. 1-7a) - Granodioritic orthogneiss in core of Kirbyville anticline; located ~4 km SW of Kirbyville Lake at 2130 m, 372720E-5710740N (UTM zone 11, BCGS map sheet 82M.056)												
A	4,149	47	233.9	78.68	4325	47	11.5	0.3108 ± 0.12%	4.878 ± 0.13%	0.96	0.11381 ± 0.04%	1861.1 ± 1.3
B	4,149	68	330.4	102.7	22286	19	7.3	0.3010 ± 0.11%	4.715 ± 0.12%	0.97	0.11362 ± 0.03%	1858.1 ± 1.1
C	1,149	40	183.7	58.84	7766	18	10.1	0.3004 ± 0.11%	4.702 ± 0.12%	0.96	0.11354 ± 0.03%	1856.9 ± 1.2
D	1,149	30	275	82.96	11917	13	8.2	0.2890 ± 0.12%	4.525 ± 0.12%	0.96	0.11355 ± 0.04%	1857.0 ± 1.3
T1	20,149	369	31.51	0.9581	50	593	7.1	0.0301 ± 0.86%	0.386 ± 4.0%	0.69	0.09316 ± 3.5%	1491.1 ± 133
T2	41,149	384	34.97	0.8808	49	589	5.6	0.0254 ± 0.87%	0.316 ± 4.3%	0.69	0.08998 ± 3.7%	1425.1 ± 142
T3	42,149	409	36	0.7979	44	660	6.2	0.0223 ± 1.0%	0.266 ± 5.2%	0.69	0.08666 ± 4.5%	1352.9 ± 175
207 Mount Grace gneiss (Fig. 1-7d) - Syenitic orthogneiss, body is ~12 km long and ~1 km thick, intruded into lower part of cover sequence; located 1.2 km W of Mount Grace at 1830 m, 373000E-5700500N (UTM grid zone 11, BCGS map sheet 82M.046)												
A	1,149,clr	56	395.7	9.16	1792	18	9.0	0.0231 ± 0.21%	0.186 ± 0.26%	0.60	0.05827 ± 0.21%	540.0 ± 9.4
B	2,149,clr	55	281.4	8.854	522	59	10.3	0.0309 ± 0.14%	0.255 ± 0.36%	0.59	0.05977 ± 0.30%	595.1 ± 13.0
C	3,149,cld	40	867.8	25.72	4698	14	9.5	0.0294 ± 0.24%	0.243 ± 0.30%	0.82	0.05987 ± 0.17%	598.7 ± 7.4
D	7,149,cld	60	385.6	12.05	362	129	9.2	0.0311 ± 0.28%	0.257 ± 0.59%	0.62	0.05983 ± 0.47%	597.3 ± 20.5
E	3,105,cld	27	751.6	42.49	5493	13	8.1	0.0569 ± 0.09%	0.487 ± 0.10%	0.87	0.06211 ± 0.05%	677.9 ± 2.2
257 (Fig. 1-7f) - Feldspathic semipelitic schist, base of upper part of Monashee cover sequence; located on logging road at 1190 m, 7.8 km SE of Mount Grace, 378730E-5694490N (UTM grid zone 11, BCGS map sheet 82M.037)												
A	1,149	16	86.41	15.27	1544	9	11.1	0.1691 ± 0.18%	1.872 ± 0.23%	0.83	0.08029 ± 0.13%	1204.1 ± 5.1
B	1,149	23	313.7	96.12	19911	6	14.4	0.2755 ± 0.10%	4.059 ± 0.11%	0.96	0.10685 ± 0.03%	1746.3 ± 1.1
C	1,149	15	114.9	67	8392	6	11.4	0.4978 ± 0.29%	14.197 ± 0.30%	0.99	0.20683 ± 0.03%	2880.8 ± 0.9
D	1,149	12	63.64	39.38	4832	5	15.2	0.5014 ± 0.30%	14.913 ± 0.30%	0.99	0.21573 ± 0.03%	2949.1 ± 1.1
E	1,149	11	105.5	70.46	5310	7	16.3	0.5379 ± 0.32%	15.364 ± 0.32%	0.99	0.20714 ± 0.03%	2883.3 ± 1.0
F	1,149	8	57.33	19.92	1418	5	23.0	0.2792 ± 0.39%	4.348 ± 0.41%	0.96	0.11295 ± 0.12%	1847.4 ± 4.3
G	1,149	10	59.02	37.73	3138	6	16.2	0.5121 ± 0.29%	15.154 ± 0.29%	0.98	0.21461 ± 0.06%	2940.7 ± 2.1
H	1,149	13	94.56	34.84	2457	9	23.8	0.2932 ± 0.16%	4.538 ± 0.17%	0.93	0.11227 ± 0.06%	1836.5 ± 2.3
I	1,149	12	37.6	11.31	703	12	8.1	0.2897 ± 0.38%	4.406 ± 0.43%	0.88	0.11029 ± 0.21%	1804.1 ± 7.6
J	1,149	7	69.53	25.85	488	19	22.2	0.3024 ± 0.38%	4.624 ± 0.50%	0.78	0.11090 ± 0.31%	1814.1 ± 11.4
300 (Fig. 1-7e) - Pelitic schist, base of upper part of cover sequence; located 10.0 km NE of Pettipiece Pass at 1890 m, 391980E-5705050N (UTM grid zone 11, BCGS map sheet 82M.048)												
A	1,149	10	51.39	17.54	1886	5	10.9	0.3183 ± 0.43%	4.845 ± 0.43%	0.98	0.11040 ± 0.10%	1805.9 ± 3.5
B	1,149	9	348.9	189.1	22134	4	3.1	0.5072 ± 0.11%	14.279 ± 0.12%	0.98	0.20418 ± 0.03%	2859.9 ± 0.9
C	1,149	10	197.9	62.51	5846	6	9.2	0.3001 ± 0.14%	4.595 ± 0.15%	0.97	0.11105 ± 0.04%	1816.6 ± 1.4
D	1,149	9	214.4	73.72	6773	5	17.4	0.2964 ± 0.14%	4.592 ± 0.15%	0.96	0.11235 ± 0.04%	1837.8 ± 1.4
E	1,149	10	56.54	17.65	2407	4	17.0	0.2709 ± 0.38%	4.174 ± 0.40%	0.97	0.11175 ± 0.09%	1828.1 ± 3.2

Table 1-1. (concluded).

Analysis ^a	Wt. ^b (μg)	U (ppm)	Pb* ^c (ppm)	²⁰⁶ Pb/ ²⁰⁴ Pb ^d (pg)	Pb _c ^e (pg)	²⁰⁸ Pb ^f (%)	²⁰⁶ Pb/ ²³⁸ U ^g	²⁰⁷ Pb/ ²³⁵ U ^g	Corr. coef.	²⁰⁷ Pb/ ²⁰⁶ Pb ^g	²⁰⁷ Pb/ ²⁰⁶ Pb ^h age (Ma)	
335 (Fig. 1-7e) - Quartzite, basal unit of cover sequence, a few tens of metres above unconformable contact with basement; located S of hydroelectric line in Pettipiece Pass at 1760 m, 386930E-5695690N (UTM grid zone 11, BCGS map sheet 82M.047)												
A	1,149	27	185.8	68.48	2790	36	10.8	0.3392 ± 0.08%	5.889 ± 0.10%	0.92	0.12591 ± 0.04%	2041.6 ± 1.5
B	1,149	15	72.95	28.3	606	40	11.6	0.3548 ± 0.13%	6.010 ± 0.21%	0.75	0.12285 ± 0.14%	1998.0 ± 5.1
C	1,149	25	127.6	54.87	1453	46	20.7	0.3516 ± 0.10%	6.111 ± 0.13%	0.86	0.12604 ± 0.07%	2043.5 ± 2.3
D	1,149	22	88.53	32.17	6590	6	9.8	0.3392 ± 0.14%	5.736 ± 0.15%	0.97	0.12263 ± 0.03%	1994.9 ± 1.2
E	1,149	25	74.27	29.07	4359	9	15.0	0.3446 ± 0.15%	5.811 ± 0.16%	0.96	0.12230 ± 0.05%	1990.1 ± 1.6
F	1,149	12	140	55.49	362	99	14.3	0.3516 ± 0.30%	5.920 ± 0.47%	0.74	0.12209 ± 0.32%	1987.1 ± 11.3
G	1,149	12	35.44	12.28	1718	5	9.2	0.3255 ± 0.52%	5.571 ± 0.52%	0.99	0.12416 ± 0.09%	2016.8 ± 3.1
H	1,149	30	101.7	37.17	7161	9	11.1	0.3365 ± 0.11%	5.689 ± 0.12%	0.96	0.12263 ± 0.04%	1994.9 ± 1.3
341 (Fig. 1-7b) - Pegmatite layer, up to 3 m thick, boudinaged and concordant with schistosity in host, intruded into lower part of cover sequence; located 6.8 km NNE of Pettipiece Pass at 2250 m, 389530E-5702660N (UTM zone 11, BCGS map sheet 82M.048)												
A	1,105	20	2826	517.9	2534	249	1.1	0.1900 ± 0.09%	2.874 ± 0.11%	0.90	0.10972 ± 0.05%	1794.7 ± 1.8
B	1,105	23	2649	577	13902	58	1.1	0.2255 ± 0.09%	3.437 ± 0.10%	0.96	0.11052 ± 0.03%	1807.9 ± 1.0
C	2,105	20	2637	582.1	5738	124	1.3	0.2283 ± 0.09%	3.490 ± 0.10%	0.95	0.11088 ± 0.03%	1813.9 ± 1.2
D	1,105	9	2182	523.2	34441	9	0.8	0.2489 ± 0.10%	3.831 ± 0.11%	0.97	0.11163 ± 0.03%	1826.1 ± 1.0
E	1,105	7	1720	457.3	20082	11	0.8	0.2759 ± 0.09%	4.271 ± 0.10%	0.96	0.11227 ± 0.03%	1836.4 ± 1.1
F	1,105	7	1374	329	9493	16	0.8	0.2486 ± 0.10%	3.828 ± 0.11%	0.96	0.11170 ± 0.03%	1827.3 ± 1.2
I	1,149	26	4154	145.1	981	58	79.5	0.007910 ± 0.12%	0.05121 ± 0.28%	0.57	0.04696 ± 0.23%	46.9 ± 11.2
X1	1,149	20	10354	85.9	5873	20	2.2	0.009020 ± 0.13%	0.05880 ± 0.15%	0.90	0.04728 ± 0.07%	63.1 ± 3.2
X2	1,149	20	12930	102.9	19858	7	2.3	0.008641 ± 0.12%	0.05627 ± 0.13%	0.96	0.04723 ± 0.04%	61.0 ± 1.8
X3	1,149	15	10038	80.73	6897	12	4.2	0.008505 ± 0.10%	0.05574 ± 0.12%	0.88	0.04720 ± 0.06%	59.4 ± 2.9

Isotopic composition of Carleton University lab blank (uncertainty is one standard deviation) - 206:207:208:204 = 19.01 ± 0.36 : 15.64 ± 0.20 : 38.23 ± 0.74 : 1

^aIn first column, A-L = fraction codes for zircon analyses, 1-4 = fraction codes for monazite analyses, X1-X3 = fraction codes for xenotime analyses, T1-T3 = fraction codes for titanite analyses; in second column, 1-9 = number of grains analyzed, 105-149 = average size in μm prior to abrasion with pyrite (Krogh 1982), se = subequant shape (length : width = ~1.5 : 1), el = elongate shape (length : width = ~3.5 : 1), cl = clear in core, cld = cloudy in core.

^bWeighing uncertainty is 1 μg.

^cRadiogenic Pb.

^dMeasured ratio, corrected for spike and Pb fractionation of 0.09% ± 0.03%/AMU.

^eTotal common Pb in analysis, corrected for spike and fractionation.

^fRadiogenic ^{208}Pb , expressed as % of total radiogenic Pb.

^gCorrected for blank Pb and U and common Pb (Stacey-Kramers model Pb composition equivalent to the interpreted age of the analyses); errors are one standard deviation in %.

^hErrors are two standard deviations in Ma.

Figure 1-1. Age domain map of the crystalline basement of the western Canadian Shield, modified after Ross et al. (1991), based on geophysical properties and U-Pb data presented in Villeneuve et al. (1993). The position of the four crystalline basement exposures in the Cordillera are also shown. The approximate restored positions of the Malton complex and Sifton Ranges gneiss (shown in white) are taken from McDonough and Parrish (1991). The approximate position of the present east limit of the Windermere Supergroup is taken from Gabrielse and Campbell (1991). GSL = the Great Slave Lake shear zone.

Figure 1-2. Tectonic assemblage map in the vicinity of the Monashee complex, modified after Wheeler and McFeely (1991), showing the location of Fig. 1-4. The dark grey square in inset locates the map with respect to the morphogeologic belts of the Canadian Cordillera.

Figure 1-4. Geologic map of northern Frenchman Cap dome, modified after Wheeler (1965), McMillan (1973), Psutka (1978), Höy and Brown (1980), Journeay (1986), and Höy (1987), showing localities of the samples with U-Pb isotopic data. A-A' locates the cross section shown in Fig. 1-5.

Figure 1-5. Cross section of northern Frenchman Cap dome, modified after Journeay (1986). Location of section and patterns for rock units are shown in Fig. 1-4.

Figure 1-6. Schematic stratigraphic column from the Ratchford Creek area of northern Frenchman Cap dome, modified after Journeay (1986), showing the position of geochronologic samples from previous work and this study.

Figure 1-7. U-Pb concordia plots. Ellipses for the analyses represent the two sigma uncertainty (95% confidence level) and analyses that are plotted with + and x symbols have ellipses that are too small to be seen at this scale. Zircon analyses are denoted with letters, monazite analyses are denoted with numbers, titanite analyses are denoted with T1-T3, and xenotime analyses are denoted with X1-X3. See Table 1-1 for analytical data.

Figure 1-8. Sample 341 is from the 1852 ± 4 Ma pegmatite layer that lies a few metres to the left of the geologist (thickest white concordant layer). The view is to the north. The collection site was ~ 100 m along strike to the south. The host rock is amphibolitic gneiss that was mapped by Psutka (1978) as high in the lower part of the cover sequence.

CHAPTER 2

U-PB GEOCHRONOLOGY OF PALEOPROTEROZOIC ROCKS IN THE MONASHEE COMPLEX, SOUTHERN CANADIAN CORDILLERA: ELUCIDATING AN OVERPRINTED GEOLOGIC HISTORY

ABSTRACT

The Monashee complex contains the largest of four exposures of Paleoproterozoic rocks in the Canadian Cordillera. These rocks are composed of basement para- and orthogneisses, metasedimentary rocks in the base of an unconformably overlying cover sequence, and younger intrusive rocks. U-Pb geochronologic data and field relationships from the northern part of the complex are used to outline a Paleoproterozoic geologic history that includes six intrusive episodes, two metamorphic events, two depositional sequences, and one deformation event. Evidence for this history is best preserved in rocks that lie in deep structural levels of the complex, where overprinting related to middle and upper amphibolite facies Cordilleran (early Tertiary) orogenesis was incomplete. Although interpretation of precise U-Pb ages is complicated by Pb loss and/or overgrowth that occurred in the dated minerals during one or more thermal overprints, there is sufficient age resolution on the intrusive events to conclude that the complex has affinities with the western Canadian Shield and the other Paleoproterozoic exposures in the Cordillera. The dominant Paleoproterozoic rocks in the complex are large orthogneiss

bodies dated at 2272 ± 8 Ma, 2099 ± 17 , and 2077 ± 2 Ma. These rocks and an older paragneiss sequence underwent the most significant recorded Paleoproterozoic metamorphism and deformation, which are recorded by ~ 2.06 Ga monazite in pelitic schist and a migmatitic gneissosity that is interpreted as coeval based on the intrusive relationships of an amphibolite dyke. Other intrusive rocks include a pegmatite layer, which may be similar in age to a previously dated ~ 1.93 Ga orthogneiss in the southern part of the complex, and a 1862 ± 1 Ma orthogneiss that is coeval with a metamorphism recorded by titanite in amphibolitic gneiss. The rocks that predate 1.86 Ga formed the basement upon which the lower part of an extensive cover sequence was deposited. Deposition occurred prior to the intrusion of a 1852 ± 4 Ma pegmatite. Basement was intruded by four granitoids dated at 1848 ± 3 to 1834 ± 2 Ma, and the Paleoproterozoic history in the Monashee complex concluded with intrusion of a 1762 ± 6 Ma leucogranite layer into the cover sequence.

INTRODUCTION

The four known exposures of Paleoproterozoic rocks in the Canadian Cordillera exist in deep structural levels of the orogen and are isolated from the western Canadian Shield by the deformed Cordilleran miogeocline and cratonic cover (Fig. 2-1). Although previous U-Pb geochronology shows that the protoliths of some orthogneisses in the Cordillera intruded at 2.09-1.85 Ga (Wanless and Reesor 1975; Evenchick et al. 1984; Armstrong et al. 1991; McDonough and Parrish 1991; Murphy et al. 1991; Parkinson 1991) and were metamorphosed at ~ 1.85 Ga (Parrish 1995), there was little knowledge of

the Paleoproterozoic geologic history of these exposures. Ages of deformation and metamorphism were largely unknown and the resulting fabrics were not identified, mainly because a high grade metamorphic and deformation overprint related to the early Tertiary Cordilleran orogenesis affected the Paleoproterozoic rocks to such a degree that their geologic history can only be differentiated from the Cordilleran history with a significant amount of U-Pb dating and mapping.

Uranium-lead geochronology and field relationships are used in this chapter to outline a basic Paleoproterozoic geologic history in the Monashee complex (Figs. 2-1, 2-2), which contains the largest exposure of Paleoproterozoic rocks in the Canadian Cordillera. Six intrusive episodes, two metamorphic events, one deformation event, and two depositional sequences were identified and bracketed in age. A major problem encountered during this study was differentiating between the effects of the Paleoproterozoic tectonic events and the effects of an early Tertiary Cordilleran overprint, and thus this chapter outlines how this obstacle was partly overcome by working in deep structural levels of the complex where the overprint was shown to be less pervasive. Although the details of the Paleoproterozoic events remain uncertain due to the Cordilleran overprint, there is sufficient age resolution to permit correlation between intrusive episodes in the complex with plutonism in the other Paleoproterozoic exposures in the Cordillera and the western Canadian Shield.

GEOLOGIC SETTING

The Monashee complex lies in the southern Omineca Belt, the metamorphic and plutonic hinterland of the Canadian Cordillera (Monger et al. 1982), in the antiformal tectonic window beneath the Monashee décollement (Fig. 2-2). The décollement is a ductile crustal-scale thrust fault (Read and Brown 1981; Brown et al. 1986; Journeay 1986; McNicoll and Brown 1995) that was correlated with the sole thrust of the Rocky Mountain Foreland Belt (Brown et al. 1992) and known to have been active in the Paleocene (Carr 1992). The Selkirk allochthon in the hanging wall was displaced up to 300 km eastward relative to the North American craton (Brown et al. 1993), with more than 80 km of motion occurring on the décollement (Read and Brown 1981).

Paleoproterozoic rocks in the Monashee complex represent the southernmost of four Paleoproterozoic exposures in the Canadian Cordillera (Fig. 2-1). The other exposures are (i) the Sifton Ranges orthogneiss in north-central British Columbia, dated at 1851 ± 13 Ma (Evenchick et al. 1984); (ii) the Malton complex orthogneisses located 80 km north of the Monashee complex, dated at 1887 ± 7 , 1870 ± 3 , and 1866 ± 3 Ma (McDonough and Parrish 1991); and (iii) the Gold Creek orthogneiss located 10 km northwest of the Malton complex, dated at $2086 +23/-15$ Ma (Murphy et al. 1991).

Paleoproterozoic rocks in the Monashee complex consist of basement para- and orthogneisses (Wanless and Reesor 1975; Armstrong et al. 1991; Parkinson 1991; Chapter 1; this chapter), metasedimentary rocks in the lower part of an unconformably overlying cover sequence (Journeay 1986; Scammell and Brown 1990; Chapter 1), and granitoids that intruded into basement and cover (Armstrong et al. 1991; Chapter 1; this chapter).

Basement forms the core of two elongate structural culminations in the complex (Fig. 2-2) that developed during Cordilleran orogenesis, namely, Frenchman Cap dome in the north (Wheeler 1965; Journeay 1986) and Thor-Odin dome in the south (Reesor and Moore 1971; Read 1980). This chapter presents data from the Pettipiece Pass area of northern Frenchman Cap dome (Fig. 2-3).

Basement

Basement is composed of rocks that are known or are assumed to be older than the lower part of the unconformably overlying cover sequence. Frenchman Cap dome is dominated by large (kilometre-wide) homogeneous bodies of K-feldspar augen gneiss and granitic to tonalitic gneiss (Fig. 2-3; Wheeler 1965; Journeay 1986), which have only been mapped in detail by Journeay (1986) in the north-central part of the dome. Intrusive contacts of these bodies do not exist, as the contacts at the outcrop scale are concordant with the gneissosity. Armstrong et al. (1991) reported that (i) orthogneiss contains highly radiogenic Sr that scatters about a 2206 ± 117 Ma (1 sigma) Rb-Sr errorchron; (ii) Sm-Nd whole-rock and crustal-residence (T_{DM}) ages are 2.3 ± 0.2 Ga; and (iii) U-Pb isotopic data show that augen gneiss intruded at 2066 ± 8 Ma and granite at 1951 ± 8 or 1851 ± 8 Ma (sample PCA-305B-83, discussed extensively below). They proposed that migmatization and gneissosity development occurred in the Paleoproterozoic; Sr and Nd data from closely related felsic-mafic sample pairs were interpreted as indicating that the layering and chemical heterogeneity of the orthogneisses are the same age as the rocks themselves. These data and the minor Pb loss in monazite from the granite led to the

conclusion that the effects of post-Paleoproterozoic metamorphism on basement were "relatively mild" (p. 1188). Parrish (1995) used titanite U-Pb data to deduce a metamorphism at $\sim 1.90\text{--}1.85$ Ga and he inferred from the moderate amount of Pb loss that a substantial part of the metamorphic characteristics of the basement are Paleoproterozoic. In Thor-Odin dome, where orthogneiss forms a minor part of the basement (Reesor and Moore 1971), granodioritic augen gneiss was dated at $1960 \pm 35\text{--}45$ Ma (Wanless and Reesor 1975) and 1934 ± 6 Ma (Parkinson 1991), and homogeneous granitic gneiss was dated at 1874 ± 21 Ma (Parkinson 1991).

Frenchman Cap orthogneiss is overlain by and intercalated with paragneiss layers (up to 0.5 km thick) that consist mostly of hornblende biotite paragneiss, with local pelitic and semipelitic schist (Fig. 2-3; Wheeler 1965; Psutka 1978; Journeay 1986). Previous workers mapped the paragneiss as basement because of its unconformable relationship with the cover sequence (McMillan 1973; Psutka 1978; Höy and Brown 1980; Journeay 1986). It was interpreted as having been intruded by basement orthogneiss based on the inference that metre-scale interlayering of the para- and orthogneisses and map-scale discordance between the units resulted from deformation of intrusive contacts (Journeay 1986).

Basement was affected by Paleoproterozoic metamorphism and deformation, whereas the cover sequence and rocks that intruded into the basement were only subjected to events related to Cordilleran orogenesis. Cordilleran events in northern Frenchman Cap dome are well documented; upper amphibolite facies (with a typical assemblage in pelitic schist of kyanite-sillimanite-garnet-biotite-muscovite-melt) metamorphism was dated at

early Tertiary (U-Pb data from Parrish (1995); Chapters 3, 4) and deformation resulted in kilometre-scale isoclinal folds, a transposition fabric, and penetrative mineral and stretching lineations (Journey 1986 and references therein). Most outcrops of basement orthogneiss in the Pettipiece Pass area contain a single gneissosity that U-Pb dating suggests developed in the early Tertiary (Chapter 4). However, the dating and field relationships presented in this chapter show that a Paleoproterozoic gneissosity is preserved in centimetre- to metre-scale areas that escaped significant early Tertiary deformation. Because the two gneissosities are similar in appearance and orientation, they are only distinguishable in outcrops with unambiguous dating and fabric relationships.

Cover sequence

The cover sequence, interpreted as having been deposited as a platformal succession, is composed of upper amphibolite facies metasedimentary and subordinate metaigneous rocks (Wheeler 1965; McMillan 1973; Psutka 1978; Höy and Brown 1980; Journey 1986; Höy 1987; Scammell and Brown 1990). The sequence is 2-3 km thick and laterally extensive, with some units continuous for 150 km along strike. It occupies a nearly unique position in the Canadian Cordillera due to its depositional contact with crystalline basement (McMillan 1973; Höy and Brown 1980; Journey 1986). Although there the cover sequence is not in exposed contact with the basement orthogneiss (due to an intervening ubiquitous paragneiss horizon), previous workers inferred an unconformable relationship because layers of orthogneiss only exist beneath the cover sequence (McMillan 1973; Höy and Brown 1980; Journey 1986). This inference was

confirmed by U-Pb dating that shows that detrital zircons in the cover sequence (Ross and Parrish 1991; Chapter 1) were deposited after intrusion of four basement orthogneiss bodies (Armstrong et al. 1991; this chapter). Rocks in the lower part of the cover sequence were deposited in the Paleoproterozoic (Chapter 1), and those in the middle and upper parts may be as young as Paleozoic (Höy and Godwin 1988; Scammell and Parrish 1993).

Younger intrusive rocks

Basement and the lower part of the cover sequence were intruded by Paleoproterozoic granite and pegmatite (Armstrong et al. 1991; this chapter). Basement was also intruded by syenitic orthogneiss (Höy and Brown 1980) that is correlated with nearby 724 ± 5 Ma syenitic orthogneiss Chapter 1) and by early Tertiary pegmatite (Chapter 4).

GEOCHRONOLOGY

Zircon, monazite, and titanite from 15 samples in the Pettipiece Pass area of northern Frenchman Cap dome were dated in order to better constrain the Paleoproterozoic geologic history. U-Pb data, sample lithologies, and sample locations from most of the samples are given in part in Table 2-1 and in part in Chapter 1. Sample locations and ages are shown on the map in Fig. 2-3, concordia diagrams are given in Fig. 2-4, and ages are summarized in Table 2-2 and in the timeline in Fig. 2-5. Ages and

interpretations for intrusive rocks are presented first and are followed by those for pelitic schist and amphibolitic gneiss. Intrusive rocks that are roughly coeval are grouped into six intrusive episodes, but this grouping does not imply magmatic affinity.

Analytical methods

The U-Pb geochronology followed procedures outlined by Parrish et al. (1987): microcapsules for mineral dissolution (Parrish 1987), a mixed ^{233}U - ^{235}U - ^{205}Pb tracer (Parrish and Krogh 1987), multicollector mass spectrometry (Roddick et al. 1987), and estimation of errors using numerical error propagation (Roddick 1987). Decay constants used are those recommended by Steiger and Jäger (1977). Discordia lines through analyses were calculated with the use of a modified York (1969) regression (Parrish et al. 1987), which takes scatter into account. Mineral selection and most of the U-Pb chemistry were performed in the geochronology laboratory at Carleton University. All zircons were abraded (Krogh 1982). For zircon, monazite, and xenotime chemistry, U blanks were less than 5 pg and Pb blanks were usually 5-20 pg. For titanite chemistry, U blanks were less than 5 pg and Pb blanks were ~50 pg for initial analyses and ~20 pg for the final analyses. Other U-Pb chemistry was performed at the geochronology laboratory at the Geological Survey of Canada in Ottawa, where blanks were similar to those at Carleton University. Isotopic ratios were determined on a MAT 261 mass spectrometer at the Geological Survey of Canada in Ottawa.

Intrusive rocks

All U-Pb isotopic analyses from intrusive rocks are discordant, a result attributed to any combination of the following: inheritance of U- and Pb-bearing xenocrystic material, episodic Pb loss, and mineral growth at a later time. The latter two conditions are probable because the rocks were substantially reheated during early Tertiary Cordilleran metamorphism and the oldest rocks were also subjected to one or two Paleoproterozoic reheating events. Inheritance is also probable because igneous minerals in crustal melts are susceptible to incorporation of xenocrysts. The interpreted igneous crystallization ages are the upper intercept ages of collinear arrays through discordant analyses (here termed discordia lines) that are thought to lack inherited material. Because inheritance increases the ages of analyses to variable amounts, analyses that are thought to lack inheritance are those that form discordias, composed of three or more points, that lie above the analyses on the concordia diagram that are thought to contain inherited material. Xenocrystic cores in grains with interpreted inheritance were not observed with the use of a microscope. Interpretation in this manner yields reliable crystallization ages if the chosen analyses indeed lack inherited material. The only evidence for this is that they yield the youngest possible upper intercept ages amongst the dated analyses and they were the same age prior to the event that produced the discordance. If the chosen analyses did contain inherited material, then the crystallization ages must be younger than the upper intercepts.

Lower intercepts of the discordias from samples that postdate 1.86 Ga are interpreted as approximately the time of Pb loss and/or overgrowth during early Tertiary

metamorphism. The same does not hold for samples that predate 1.86 Ga because they were affected by multiple reheating events. Upper intercepts are interpreted as crystallization ages rather than the ages of inherited material (with crystallization occurring at the lower intercept age) because that would require all grains to contain a significant amount of inherited material (composing an average of 70-85% of the Pb in zircon and 80-90% of the Pb in monazite) of a single age, which is considered unlikely based on comparison of the analyses with those in nearby early Tertiary pegmatites and leucogranites (Chapter 4). Three of eight early Tertiary samples contain zircons with obvious inherited material, which varies widely in age (from ~2.9 to 1.7 Ga) and constitutes <50% (typically <5%) of the Pb in the analyses; such characteristics contrast with those in the samples that are interpreted as Paleoproterozoic.

~2.27 and 2.10-2.08 Ga orthogneisses: dominant basement rocks

Orthogneiss bodies dated at ~2.27 (sample 311) and 2.10-2.08 Ga (samples 60C and 65) are the dominant basement rocks in Frenchman Cap dome (Fig. 2-3). The strongly deformed nature of the contacts with adjacent rocks does not permit determination of intrusive relationships. Zircon U-Pb analyses are 10-30% discordant, as measured from the upper intercepts along the discordia lines (Fig. 2-4a,b). It is inferred that a substantial part of the discordance resulted from early Tertiary metamorphism because younger zircons that are known to have only been subjected to this metamorphism (those from post-1.86 Ga samples, see below) have an equal or greater amount of discordance. However, reheating during early Tertiary metamorphism cannot

be the only explanation for the discordance in zircon from the ~2.27 and 2.10-2.08 Ga samples because the lower intercept ages are considerably older (by >0.2 Ga) than early Tertiary and do not correspond with any known reheating events. Therefore, it is thought that the zircons contain inherited material and/or were affected by multistage Pb loss and/or overgrowth, resulting in upper intercepts that are not accurate igneous crystallization ages. Nevertheless, upper intercept ages are thought to be a close estimate of the igneous crystallization ages because each sample has at least one analysis with only moderate discordance and the fits to the discordias are good to excellent.

Sample 311 is from the Feline gneiss, a hornblende biotite quartz dioritic orthogneiss west of Feline Peak and east of the toe of Bourne Glacier (Fig. 2-3). This kilometre-wide body contains a gneissosity of unknown age and it is generally more mafic than adjacent orthogneisses. Three single-grain zircon analyses (C, D, G; Table 2-1) define a discordia line with upper and lower intercepts of 2272 ± 8 and 804 ± 66 Ma, respectively (MSWD = 5.3, discordance is 7-25%; Fig. 2-4a). Four other single-grain zircon analyses (A, B, E, F) lie below this discordia and are thought to contain inherited material. Although its age is imprecisely known, the array of zircon analyses is interpreted to indicate that the Feline gneiss is considerably older than the other dated rocks.

Sample 60C is from the Pettipiece gneiss, a garnet biotite granitic orthogneiss south of Pettipiece Pass (Fig. 2-3), thought to be the largest orthogneiss body in the study area. It locally contains two gneissosities. The older gneissosity is interpreted as Paleoproterozoic based on relationships presented below and in Figs. 2-6 and 2-7, and the younger gneissosity is interpreted as early Tertiary based on relationships presented in

Chapter 4. Three single-grain zircon analyses (A, B, C; Table 2-1) define a discordia line with upper and lower intercepts of 2099 ± 17 and 435 ± 73 Ma, respectively (MSWD = 3.6, discordance is 17-32%; Fig. 2-4b). Two other single-grain zircon analyses (D, E) lie slightly below this line and are thought to contain inherited material.

Sample 65 is from the Frenchman Cap gneiss, a garnet biotite K-feldspar augen granitic orthogneiss south of Pettipiece Pass (Fig. 2-3), thought to be the second largest orthogneiss body in the study area and considered by Journeay (1986) to be the youngest of the large bodies. It locally contains the same two gneissosities that exist in the Pettipiece gneiss. Four single-grain zircon analyses (Table 2-1) define a discordia line with upper and lower intercepts of 2077 ± 2 and 291 ± 19 Ma, respectively (MSWD = 0.1, discordance is 10-20%; Fig. 2-4b). More confidence is placed in this upper intercept age as being the igneous crystallization age, compared with the Feline and Pettipiece gneisses, because the lower intercept is closer to the known age of Pb loss and/or overgrowth, the discordia line has a considerably better fit, and all zircon analyses apparently lack inherited material.

~2.05 Ga amphibolite dyke: postdates a gneissosity

Sample 337 is from a plagioclase hornblende amphibolite dyke ~0.2 m thick that intruded into basement orthogneiss south of Pettipiece Pass, near samples 65 and 79 (Fig. 2-3). Crystallization is interpreted as having postdated the steeply dipping gneissosity and tectonically produced compositional layering in the host gneiss because intrusive contacts are straight, subhorizontal, and highly discordant to these elements (Fig. 2-6). The layer

and a foliation within it (defined by aligned hornblende and biotite) parallel the axial surfaces of folds that deformed the gneissosity and compositional layering in the host. In nearby outcrops, similar-looking folds that are thought to be the same age based on style and orientation formed, at least in part, in the early Tertiary (Chapter 4). Amphibolite dykes with this structural setting are uncommon in the Pettipiece Pass area. Zircons are clear, colourless, prismatic, and elongate; their morphology indicates an igneous origin. Three zircon analyses composed of four grains each (A, B, C) and one single-grain zircon analysis (D; Table 2-1) define a discordia line with upper and lower intercepts of 2051 ± 12 and 140 ± 95 Ma, respectively (MSWD = 32.6, discordance is 11-29%; Fig. 2-4c). The upper intercept is the interpreted crystallization age of this sample. Two other single-grain zircon analyses (E, F) lie slightly below this line and are thought to contain inherited material. Three multigrain titanite analyses (Table 2-1) define a discordia line with upper and lower intercepts of $1780 +247/-216$ and $29 +33/-39$ Ma, respectively (discordance is 89-98%; Fig. 2-4c). Although the discordance is large, it is certain that these titanites grew after igneous crystallization of the layer. It is hypothesized that they grew during the ~1.85 Ga event during which titanite growth occurred in nearby (0.2 km) amphibolitic gneiss (sample 79, see below). Forcing an upper intercept through this age yields a lower intercept of 38 ± 20 Ma, which is the interpreted age of the Pb loss and/or overgrowth that caused the discordance.

~1.97 Ga (?) pegmatite

Sample 291 is from a 1 m thick garnet biotite pegmatite layer that intruded into basement south of Pettipiece Pass (Fig. 2-3). Intrusive contacts are slightly discordant to a gneissosity in the host gneiss, and flattened quartz and feldspar in the layer define a fabric that parallels the gneissosity. The relationship between the gneissosity and intrusion of the pegmatite is uncertain. It was dated because of the interpretation that deformation occurred, at least in part, after intrusion. Seven zircon analyses (one to two grains each; Table 2-1) are discordant and do not form a reasonably fitting discordia line, but an approximate upper age limit on crystallization can be deduced if it is assumed that discordance results from (i) variable amounts of Pb loss and/or overgrowth that occurred only during early Tertiary metamorphism (dated in the host gneiss at 50 Ma; Chapter 3) and (ii) inheritance of multiple ages. In this scenario the analysis on the left side of the elongate array on the concordia diagram (F) shows the least inheritance, which leads to the inference that the protolith crystallization age must be younger than or equal to 1.97 Ga, the upper intercept of a chord drawn through the analysis from a forced 50 Ma lower intercept (discordance is 19%; Fig. 2-4d). How much younger the rock could be depends on the amount of inheritance in this analysis.

1.86 Ga orthogneiss: youngest basement rock

Sample 187 is from the Kirbyville gneiss, a hornblende biotite granodioritic orthogneiss north of Blais Creek (Fig. 2-3). Its setting and U-Pb data are presented in Chapter 1. It is one of many orthogneiss bodies that lie in the core of the Kirbyville

anticline, an overturned, kilometre-scale isocline lying west of the main exposure of Frenchman Cap basement (Höy and Brown 1980; Journeay 1986). The strongly deformed nature of the contacts between the bodies does not permit determination of intrusive relationships. Höy and Brown mapped the bodies as basement because of their widespread existence beneath the cover sequence and their absence within it. The crystallization age is interpreted as 1862 ± 1 Ma based on the upper intercept of a discordia line defined by four zircon analyses (one to three grains each) and three multigrain titanite analyses (MSWD = 2.3, lower intercept is 63 ± 14 Ma, zircon discordance is 7-14% and titanite discordance is 94-96%).

1.85-1.83 and 1.76 Ga granitoids: intrusions into basement and cover

Granitoids dated at 1.85-1.83 and 1.76 Ga intruded into the lower part of the cover sequence and basement. The discordance in the analyses that apparently do not contain inherited material is interpreted as the result of a single episode of Pb loss and/or growth during the early Tertiary metamorphism (Chapter 3) because the lower intercepts of the discordia lines are approximately coeval with this age. Pegmatites of this age are grouped into the Pettipiece pegmatite suite.

Sample 341 is from a 2-3 m thick pegmatite layer that intruded into the cover sequence on the east flank of the dome, northeast of Pettipiece Pass (Fig. 2-3). Its setting and U-Pb data are presented in Chapter 1. The crystallization age is interpreted as 1852 ± 4 Ma based on the upper intercept of a discordia line defined by five single-grain zircon analyses (MSWD = 9.8, lower intercept is 149 ± 20 Ma, discordance is 19-46%).

Bourne granite is grey medium-grained biotite granite that intruded into basement orthogneiss between Pettipiece Pass and Bourne Glacier. It is mostly composed of quartz, plagioclase feldspar, K-feldspar, biotite, and epidote, and it exists in ubiquitous 0.2-1.0 m thick layers that emanate from a small number of stocks that are up to ~0.2 km in diameter. The layers and stocks were correlated in the field on the basis of colour, structural state, and mineralogy, and in thin section on the basis of abundant coarse epidote (of presumed igneous origin) and zircon with clear, rod-shaped inclusions. Bourne granite appears nearly undeformed in outcrop due to its equigranular and homogeneous nature, but it does contain a weak to moderate lineation, which developed, at least in part, during the early Tertiary deformation (Chapter 4). The effects of this deformation are best seen in coarser grained pegmatite that intruded into the granite. Granite intrusion is interpreted as having postdated a migmatitic gneissosity in the host orthogneiss based on the local high discordancy between the layers and the gneissosity, as well as on the pristine appearance of the granite where it is in contact with the highly tectonized host (Fig. 2-7a-d) (Crowley and Schaub 1994). This is thought to be the same gneissosity that was postdated by a 2051 ± 12 Ma amphibolite dyke (sample 337). Bourne granite was dated because its ubiquity, characteristic appearance, and intrusive relationships make it the most valuable structural and time marker in the study area. U-Pb data have been obtained from three rocks that are considered to be part of the Bourne granite suite. Sample 60B is from a 0.5 m thick layer that emanates ~300 m from the stock from which sample 324A was collected. Sample PCA-305B-83, which was collected 6 km to the south of samples 60B and 324A for the Armstrong et al. (1991)

study, is considered to be Bourne granite on the basis of (i) descriptions given by Armstrong et al.; (ii) observations made during this study of its field relationships; and (iii) similarity of hand specimens, thin sections, and zircon U-Pb data to those from samples 60B and 324A.

Data from sample PCA-305B-83 led Armstrong et al. (1991) to suggest two interpretations of its crystallization age. In one interpretation the age is 1851 ± 8 Ma, which is the upper intercept of a discordia line defined by four multigrain monazite analyses and two multigrain zircon analyses (lower intercept is 122 ± 74 Ma, discordance is 9-18%). The five multigrain zircon analyses with visible cores that plot below this discordia are thought to contain inherited material. In another interpretation the age is 1951 ± 8 Ma, which is the upper intercept of a discordia defined by four multigrain zircon analyses (lower intercept is 319 ± 51 Ma, discordance is 21-39%). In this case ~ 1851 Ma is the age of metamorphism. A crystallization age of ~ 1851 Ma is preferred here because (i) data from pelitic schist (samples 70 and 85) do not indicate growth of monazite at ~ 1851 Ma, (ii) monazite data are preferred because most zircon in Bourne granite obviously contains inherited material (see samples 60B and 324A for additional evidence), (iii) titanite from sample 324A is the same age (see below), and (iv) the lower intercept age of the monazite-dominated discordia overlaps with the known age of early Tertiary metamorphism.

Monazite from sample PCA-305B-83 and zircon and titanite from samples 60B and 324A were dated during this study to better define the discordia line given by Armstrong et al. (1991). Two single-grain monazite analyses from PCA-305B-83 (1, 3;

Table 2-1) lie on the discordia (Fig. 2-4e,f) and the resulting upper and lower intercepts are 1850 ± 4 and 120 ± 48 Ma, respectively (MSWD = 4.5, monazite discordance is 9-18% and zircon discordance is 15-16%). A third single-grain monazite analysis (2) lies slightly above the discordia, the reason for which is unknown. Eight zircon analyses (one to eight grains each; Table 2-1) from samples 60B and 324A lie below the discordia from sample PCA-305B-83 and plot in the same part of the concordia diagram as five of the seven zircon analyses from sample PCA-305B-83 (Fig. 2-4e-f). The analyses are not collinear and therefore are thought to contain inherited material that could be related to the clear, rod-shaped inclusions that typify these zircons. Samples 60B and 324A contain titanite and lack monazite. Three multigrain titanite analyses from sample 324A (Table 2-1) lie on the monazite-dominated discordia and the resulting upper and lower intercepts are 1848 ± 3 and 95 ± 35 Ma, respectively. The discordia defined by titanite only has upper and lower intercepts of $1845 +117/-100$ and $70 +101/-109$ Ma, respectively. The amount of discordance is 52-80%. The interpreted crystallization age of the Bourne granite is 1848 ± 3 Ma, the upper intercept of the discordia defined by monazite, titanite, and zircon. Titanite from sample 60B was not dated because of its low abundance and poor quality.

Sample 308 is from a 2-4 m thick muscovite biotite pegmatite dyke that intruded into basement east of the toe of Bourne Glacier (Fig. 2-3). Intrusive contacts are straight and subvertical, highly discordant to the gently dipping gneissosity in the host gneiss (Fig. 2-8). The dykes contains a weak foliation defined by aligned coarse biotite that parallels the gneissosity in the host. It is considered unlikely that the fabric in the dyke is related to

the deformation that developed the gneissosity in the host because of the strong discordancy between the intrusive contacts and the gneissosity. Therefore, crystallization is interpreted as having postdated the older gneissosity, and postcrystallization strain in its vicinity is thought to have been relatively minor. Four zircon analyses (B, C, D, E, composed of two to three grains each; Table 2-1) define a discordia line with upper and lower intercepts of 1845 ± 3 and 84 ± 19 Ma, respectively (MSWD = 4.2, discordance is 13-31%; Fig. 2-4g). A single-grain zircon analysis (A) lies below this line and is thought to contain inherited material.

Sample 84 is from a 2 m thick garnet muscovite biotite pegmatite dyke that intruded into basement augen gneiss east of the toe of Bourne Glacier (Fig. 2-3). Structural relationships are similar to those in sample 308, except the intrusive contacts are gently folded and the foliation within the dyke is better defined by aligned coarse biotite (Fig. 2-9). Five monazite analyses (four are single-grain and another is composed of two grains) and three zircon analyses (D and F, composed single-grain tips of grains, and B, composed of six tips; Table 2-1) define a discordia line with upper and lower intercepts of 1836 ± 2 and 76 ± 15 Ma, respectively (MSWD = 8.3, monazite discordance is 6-17% and zircon discordance is 19-34%; Fig. 2-4h). One zircon analysis (E, composed of two tips) lies slightly below the discordia and two other zircon analyses (C, composed of one single-grain, and A, composed of three whole grains) lie further below. These analyses are thought to contain inherited material.

Sample 91 is from a 0.5-1.0 m thick garnet muscovite biotite pegmatite dyke that intruded into basement and sample 308 pegmatite east of the toe of Bourne Glacier (Fig.

2-3). Intrusive contacts are straight and subvertical, highly discordant to the gently dipping gneissosity in the host gneiss (Fig. 2-8). The dyke has random igneous textures and does not contain a deformation fabric (Fig 8c). Therefore, crystallization is interpreted as having postdated the older gneissosity, and postcrystallization strain is thought to have been nonexistent in its vicinity. Three monazite analyses (1, composed of two grains, and 2 and 4, composed of single-grains; Table 2-1) define a discordia line with upper and lower intercepts of 1834 ± 2 and 85 ± 17 Ma, respectively (MSWD = 0.2, discordance is 12-22%; Fig. 2-4i). Another single-grain monazite analysis (3) lies slightly below the discordia. Three zircon analyses (A, B, E, composed of one to five grains each) and a single-grain monazite analysis (5; Table 2-1) define a discordia with upper and lower intercepts of 1844 ± 2 and 30 ± 20 Ma, respectively (MSWD = 4.3, discordance is 5-22%; Fig. 2-4i). Two other single-grain zircon analyses (C, D) lie slightly below the discordia. The interpreted age is 1834 ± 2 Ma, the upper intercept of the discordia line defined by the monazite analyses. All zircon analyses and two of the monazite analyses that lie below the discordia are thought to contain inherited material. These interpretations are made because zircon is more susceptible to inheritance than the monazite in granitoids of similar age (samples 84 and PCA-305B-83) and data from pelitic schist (samples 70 and 85) do not indicate disturbance of the monazite U-Pb system at ~ 1834 Ma.

Sample 180 is from a 0.2 m thick quartz-rich leucogranitic layer that intruded into the cover sequence south of Ratchford Creek and west of Myoff Creek (Fig. 2-3). Its setting and U-Pb data are presented in Chapter 1. It was thoroughly deformed during early Tertiary tectonism, and biotite-rich seams within it contain kyanite of presumed

metamorphic origin. The crystallization age is interpreted as 1762 ± 6 Ma based on the upper intercept of a discordia line defined by five zircon analyses (one to nine grains each) (MSWD = 2.8, lower intercept is 76 ± 20 Ma, discordance is 24-38%).

Pelitic schist

Sample 85 is migmatitic garnet muscovite sillimanite biotite schist and sample 70 is similar but contains kyanite and is not migmatitic. These samples are from basement paragneiss horizons that are interlayered with basement orthogneiss south of Pettipiece Pass (Fig. 2-3). It is uncertain when the rock-forming minerals grew, as they could have grown during Paleoproterozoic or early Tertiary metamorphism. Three monazite analyses from sample 85 (1, 3, 4) and one from sample 70 (2), each composed of one to three grains (Table 2-1), are collinear, and therefore the data are discussed together. The discordia line defined by these analyses has upper and lower intercepts of 2057 ± 13 and 48 ± 7 Ma, respectively (MSWD = 262; Fig. 2-4j). Discordance is 10-88%, with one single-grain analysis having the least discordance and another having 74% discordance. Five other monazite analyses (three from sample 70 (1, 3, 4) and two from sample 85 (2, 5), each composed of one to five grains; Table 2-1) are concordant or have slight (<0.5 Ma) reverse discordance with $^{207}\text{Pb}/^{235}\text{U}$ ages of 51.5-49.5 Ma (Fig. 2-4j). It is uncertain how much of the discordance in the older single-grain analyses resulted from Pb loss and how much resulted from overgrowth at ~ 50 Ma. The discordant multigrain analyses probably reflect a combination of ~ 50 Ma grains and older grains. The reverse discordance in some of the ~ 50 Ma analyses suggests that they reflect monazite growth at

this time rather than complete Pb loss, assuming that the reverse discordance is due to unsupported Pb that resulted from incorporation of excess ^{230}Th upon crystallization (Schärer 1984; Parrish 1990). These data are interpreted as indicating that monazite first grew or cooled below its Pb closure temperature at ~ 2.06 Ga and underwent Pb loss and growth during ~ 50 Ma metamorphism.

Four single-grain zircon analyses (Table 2-1) from sample 85 yield upper intercept ages, when forced through 50 Ma lower intercepts, of 2.30, 2.19, 2.00, and 1.85 Ga. Discordance is 8-15%. These grains are small (74-105 μm prior to abrasion) and have no distinct morphological features suggestive of either detrital or metamorphic origin. A detrital origin is suggested by the similarity of the two youngest zircon ages with detrital zircon ages from schist and quartzite in the cover sequence (Chapter 1). However, this scenario is considered to be unlikely because it constrains the basement paragneiss to be coeval or younger than the lower part of the cover sequence; this requires substantial post-1.85 Ga folding and/or faulting to have caused interlayering of ortho- and paragneiss in an area that is known to have been weakly deformed to undeformed after intrusion of 1848 ± 3 Ma Bourne granite. This scenario is also ruled out by the presence of 2057 ± 13 Ma monazite in this schist; monazite has an assumed metamorphic rather than detrital origin because four dated grains were the same age prior to the overprint, which is unlikely if they are detrital. Given the improbability of a detrital origin, the zircon ages are interpreted as resulting from Pb loss and/or growth associated with Paleoproterozoic and early Tertiary metamorphic events. Evidence for metamorphism that is coeval with the youngest zircon is provided by the following sample.

Amphibolitic gneiss

Sample 79 is from a hornblende plagioclase amphibolitic gneiss layer ~0.3 m thick that lies within the 2077 ± 2 Ma Frenchman Cap gneiss south of Pettipiece Pass (Fig. 2-3). The layer is moderately folded and concordant with (i) a gneissosity that exists in the layer, (ii) a gneissosity in the Frenchman Cap orthogneiss, and (iii) numerous other amphibolitic gneiss layers. The protoliths of the amphibolitic gneiss layers are presumed to be mafic dykes or sills that intruded into the protolith of the Frenchman Cap gneiss. The layer and the gneissosity must predate 2051 ± 12 Ma because 0.2 km to the north there are layers with identical composition and structural setting that were intruded by an amphibolite dyke of that age (sample 337; Fig. 2-6). The folds that deformed the layer and gneissosity are thought to be early Tertiary based on dating in nearby samples (Chapter 4). Sample OP-90-12 is amphibolitic gneiss collected for the Parrish (1995) study from a locality 0.4 km north of sample 79. The two samples are considered to be the same generation because amphibolitic gneiss of this type is the overwhelmingly most common mafic rock in their vicinity and because of the similarity in their U-Pb data. Parrish (1995) used U-Pb data from sample OP-90-12 to deduce a metamorphism at ~1.90-1.85 Ga.

Three multigrain titanite analyses from sample 79 (Table 2-1) are collinear with three analyses from sample OP-90-12 (Parrish 1995), and therefore the data are discussed together. The discordia line defined by these analyses has upper and lower intercepts of $1859 +54/-44$ and $71 +174/-180$ Ma, respectively (MSWD = 0.02, discordance is 26-

43%). Forcing the discordia through a lower intercept through 50 Ma, which is the age of early Tertiary peak metamorphism (Chapter 3) and the interpreted age of major Pb loss in Proterozoic rocks, yields an upper intercept of 1854 ± 13 Ma (MSWD = 0.03; Fig. 2-4k). Three analyses of zircon of inferred metamorphic origin from sample OP-90-12 define a discordia with upper and lower intercepts of $1917 +75/-59$ and $154 +166/-173$ Ma, respectively (Parrish 1995). These data lead to the same conclusion as reached by Parrish, which is that the minerals in the amphibolitic gneiss grew during metamorphism at ~ 1.85 Ga and suffered moderate Pb loss during early Tertiary metamorphism.

TIMING OF THE PALEOPROTEROZOIC GEOLOGIC HISTORY

U-Pb isotopic data and field relationships are used to place constraints on the age of Paleoproterozoic geologic events in the Monashee complex. These events are compiled on a timeline in Fig. 2-5, where they are compared against igneous ages in other Paleoproterozoic exposures in the Cordillera and the Alberta basement of the western Canadian Shield. Ages in the Shield are known from U-Pb dating of rocks that were drilled through cratonic cover sediments and interpretation of magnetic signatures of basement domains (Villeneuve et al. 1993).

Intrusive episodes

The 2272 ± 8 Ma Feline gneiss (quartz dioritic orthogneiss, sample 311) is the oldest dated rock in the Canadian Cordillera. The dominant orthogneiss bodies in

Frenchman Cap dome are those of the 2.10-2.08 Ga intrusive episode, including the 2099 ± 17 Ma Pettipiece gneiss (granitic orthogneiss, sample 60C), 2077 ± 2 Ma Frenchman Cap gneiss (augen granitic orthogneiss, sample 65), and 2066 ± 8 Ma augen gneiss in southern part of the dome (Armstrong et al. 1991). The ~ 1.98 -1.93 Ga intrusive episode is represented by 1934 ± 6 Ma augen gneiss in Thor-Odin dome (Parkinson 1991), and it is possible that sample 291 pegmatite is part of this episode, but the presence of inherited material and/or multiple episodes of Pb loss and/or overgrowth make such a correlation uncertain. The 1.86 Ga episode is represented by the 1862 ± 1 Ma Kirbyville gneiss (granodioritic orthogneiss, sample 187) and 1874 ± 21 Ma biotite granite gneiss in Thor-Odin dome (Parkinson 1991). The 1.85-1.83 Ga episode is represented by 1848 ± 3 Ma Bourne granite (samples PCA-305B-83, 60B, 324A) and the 1852-1834 Ma Pettipiece pegmatite suite (samples 84, 91, 308, and 341). The 1.76 Ga episode is represented by 1762 ± 6 Ma leucogranite (sample 180), the youngest dated Paleoproterozoic rock in the Canadian Cordillera.

Deformation and metamorphism

~ 2.06 Ga events

The most significant Paleoproterozoic deformation in Frenchman Cap dome is constrained to be younger than the 2077 ± 2 Ma Frenchman Cap gneiss, the youngest orthogneiss that contains the gneissosity associated with this deformation, and older than a 2051 ± 12 Ma amphibolite dyke (sample 337; Fig. 2-6), the oldest rock that postdates the

gneissosity. This minimum age holds whether the zircon in the amphibolite has an igneous or metamorphic origin. Further evidence for this gneissosity developing in the Paleoproterozoic is seen in the many outcrops in which 1848 ± 3 Ma Bourne granite layers are highly discordant to the gneissosity (Fig. 2-7), and in the cliff containing the 1834 ± 2 Ma postdeformational pegmatite dyke (sample 91; Fig. 2-8). Metamorphism is thought to have been synchronous with this deformation because the gneissosity is migmatitic (Figs. 2-6 to 2-8) and the 2057 ± 13 Ma monazite in pelitic schist (samples 70 and 85) is the same age.

The early Tertiary overprint in the basement is considered to have been generally severe because Paleoproterozoic accessory minerals suffered moderate amounts of Pb loss and Paleoproterozoic fabrics and folds were largely transposed. Hence, there is not much preserved evidence pertaining to the structural history and metamorphic conditions of the Paleoproterozoic events. For example, there is only one known outcrop containing an isoclinal fold that was presumably related to the development of the ~ 2.06 Ga gneissosity. Also, it cannot be determined whether the rock-forming metamorphic minerals in pelitic schist (sillimanite, kyanite, garnet, etc.) grew at ~ 2.06 Ga because the schists lack Paleoproterozoic fabrics against which to compare growth; these minerals probably grew, at least in part, during the early Tertiary overprint because they are present in proximal schists in the cover sequence, which are known to only have been affected by the overprint. Nevertheless, it can be inferred from the migmatitic nature of the Paleoproterozoic gneissosity (Figs. 2-6 to 2-8) and concordancy of amphibolitic gneiss layers (Fig. 2-6) that high grade metamorphic conditions were attained and deformation

was severe enough to cause transposition. Postmetamorphic folds of Paleoproterozoic age exist in the cliff containing the 1.83 Ga postdeformational pegmatite dyke (sample 91; Fig. 2-8). Gneissosity in the host rocks is gently west-dipping and folded by upright, open, northwest-verging folds with northeast-trending hinge lines. It is unknown whether these folds developed shortly after development of the ~ 2.06 Ga gneissosity or at a later time, such as during the ~ 1.85 Ga events (see below). Additional study of rocks with Paleoproterozoic fabrics and intrusive relationships is clearly needed before the details of these events can be deduced.

~ 1.85 Ga events

The presence of 1854 ± 13 Ma titanite in amphibolitic gneiss (samples OP-90-12 (Parrish 1995) and 79) suggests that metamorphism occurred at this time, possibly related to intrusion of the 1862 ± 1 Ma protolith of the Kirbyville gneiss (sample 187). Titanite in sample 337 probably also grew at this time. These titanites are interpreted as having a metamorphic rather than igneous origin on the basis that their age postdates igneous crystallization by ~ 200 Ma. The metamorphism is inferred to have been a low to medium grade event, probably at temperatures below those necessary for Pb loss in monazite because monazite in pelitic schist (samples 70 and 85) did not suffer Pb loss at this time. The presence of a 1.85 Ga zircon in the pelitic schist (sample 85) suggests that metamorphic zircon growth occurred at this time. Deformation at this time is known to have been insignificant in one locality where a 2051 ± 12 Ma amphibolite dyke (sample 337) appears to have only been affected by early Tertiary deformation.

Deposition

Although intrusive contacts between the basement paragneiss and the 2.10-2.08 Ga orthogneiss bodies are not preserved, Journeay (1986) inferred that metre-scale interlayering of the rock types and map-scale discordance between them resulted from deformation of intrusive contacts. This inference requires that the protolith of the paragneiss was deposited prior to 2.10 Ga.

The lower part of the cover sequence is certainly younger than 1.99 Ga, the age of the youngest detrital zircons dated in the basal unit, and it is probably younger than the 1862 ± 1 Ma Kirbyville gneiss, the youngest dated orthogneiss upon which it was presumably deposited. It is older than a 1852 ± 4 Ma pegmatite (sample 341), the oldest dated rock intrusive rock. These ages require that the lower part of the cover sequence contains the oldest dated metasedimentary rocks in the Canadian Cordilleran miogeocline (i.e., those deposited on crystalline basement), and that there was a <15 Ma period during which basement gneisses were intruded by the protolith of the Kirbyville gneiss, slightly to moderately metamorphosed, exhumed to the surface, overlain by platformal sediments in the lower part of the sequence, and then buried again and intruded by pegmatite. Also, because the mostly ~1.99 Ga detrital zircons in the basal unit are younger than those dated in orthogneiss bodies that dominate the underlying basement it is suggested that a source is one of the domains in the Alberta basement that contain ~2.0 Ga rocks, such as the Buffalo Head and Thorsby terranes (Fig. 2-1; Villeneuve et al. 1993).

CORRELATION WITH THE WESTERN CANADIAN SHIELD

Figure 2-5 shows that intrusive episodes in the Monashee complex are the same age as those in other Paleoproterozoic exposures in the Cordillera (Evenchick et al. 1984; McDonough and Parrish 1991; Murphy et al. 1991) and they correspond with all major Paleoproterozoic plutonic episodes in the Alberta basement of the western Canadian Shield (Fig. 2-1; Villeneuve et al. 1993). Specifically, the ~2.27 and 2.10-2.08 Ga intrusive episodes in the complex have correlatives in the Chinchaga Low and Buffalo Head terrane of the Alberta basement and in the Gold Creek gneiss of the Cordillera. The ~1.98-1.93 Ga intrusive episode has correlatives in the Taltson, Ksituan, Thorsby, and Hottah magmatic belts of the Alberta basement and in the Malton complex (augen gneiss) of the Cordillera. The 1.86 Ga and 1.85-1.83 Ga intrusive episodes have correlatives in the Great Bear, Fort Simpson, and Rimbey magmatic belts of the Alberta basement and in the Malton complex (Bulldog and Yellowjacket gneisses) and Sifton Ranges of the Cordillera. Finally, the 1.76 Ga intrusive episode is slightly younger than the Rimbey granites of the Alberta basement.

Previous workers (Armstrong et al. 1991; Parkinson 1991) contended that the Monashee complex has an affinity with the western Canadian Shield on the basis of correlations involving the volumetrically minor ~1.93 and 1.86-1.85 Ga intrusive rocks. However, this correlation is not firm because plutons of this age, although prevalent in the Shield, form a significant part of other cratons in the northern hemisphere. The correlation proposed here involving the ~2.27 and ~2.10-2.08 Ga basement orthogneisses, the dominant Paleoproterozoic rocks in the complex, is considered to be more significant

because plutons of this age are characteristic of the western Canadian Shield (they are unusual in other northern hemisphere cratons). A final conclusion is that the widespread extent of plutonic episodes in the Alberta basement means that these correlations should not be used to infer whether the complex remained autochthonous during early Tertiary orogenesis.

Table 2-1. U-Pb analytical data.

Analysis ^a	Wt. ^b (μg)	U (ppm)	Pb ^{*c} (ppm)	²⁰⁶ Pb/ ²⁰⁴ Pb ^d	Pb _c ^e (pg)	²⁰⁸ Pb/ ²⁰⁶ Pb ^f (%)	²⁰⁶ Pb/ ²³⁸ U ^g	²⁰⁷ Pb/ ²³⁵ U ^g	Corr. coeff.	²⁰⁷ Pb/ ²⁰⁶ Pb ^g	²⁰⁷ Pb/ ²⁰⁶ Pb ^h age (Ma)
60B Bourne granite (Fig. 2-4e,f) - Grey granite layer, 0.5-1.0 m thick, moderately deformed, postdates older gneissosity, intruded into basement orthogneiss; located 2.8 km SE of Pettipiece Pass at 2230 m, 388920E-5694190N (UTM zone 11, BCGS map sheet 82M.038)											
A* 3,105	33	499.4	136	37223	7	9.4	0.2568 ± 0.08%	4.128 ± 0.10%	0.96	0.11657 ± 0.03%	1904.3 ± 1.0
B* 5,105	23	426.2	100.5	21396	6	9.4	0.2226 ± 0.09%	3.556 ± 0.10%	0.96	0.11589 ± 0.03%	1893.7 ± 1.0
C* 8,74	24	514.3	136.7	823	219	8.3	0.2526 ± 0.09%	4.211 ± 0.17%	0.73	0.12090 ± 0.12%	1969.5 ± 4.3
D 4,74	23	213.2	54.86	5072	14	13.3	0.2315 ± 0.09%	3.826 ± 0.11%	0.95	0.11987 ± 0.03%	1954.4 ± 1.2
60C Pettipiece gneiss (Fig. 2-4b) - Granitic orthogneiss, contains older gneissosity; located 2.8 km SE of Pettipiece Pass at 2230 m, 388920E-5694190N (UTM zone 11, BCGS map sheet 82M.038)											
A 1,149	22	264.9	79.06	3531	30	7.2	0.2864 ± 0.10%	4.910 ± 0.11%	0.91	0.12434 ± 0.05%	2019.4 ± 1.6
B 1,149	14	310.3	111.8	866	102	10.4	0.3326 ± 0.15%	5.843 ± 0.20%	0.80	0.12744 ± 0.12%	2062.9 ± 4.3
C 1,149	12	466.2	142.1	13917	7	7.9	0.2904 ± 0.11%	4.996 ± 0.12%	0.95	0.12479 ± 0.04%	2025.8 ± 1.3
D 1,105	7	337.5	103.2	3974	10	8.1	0.2901 ± 0.12%	5.026 ± 0.13%	0.96	0.12566 ± 0.04%	2038.1 ± 1.4
E 1,149	17	171.9	48.91	3800	12	8.5	0.2692 ± 0.15%	4.639 ± 0.16%	0.97	0.12498 ± 0.04%	2028.6 ± 1.4
65 Frenchman Cap gneiss (Fig. 2-4b) - Augen orthogneiss, contains older gneissosity; located at inlet to lake 4 km SSE of Pettipiece Pass at 2100 m, 388830E-5692810N (UTM zone 11, BCGS map sheet 82M.038)											
A* 2,149	19	344.1	130.1	20079	7	11.0	0.3468 ± 0.09%	6.094 ± 0.10%	0.96	0.12745 ± 0.03%	2063.1 ± 1.0
B* 3,149	30	337.8	116.8	25247	8	10.2	0.3204 ± 0.09%	5.588 ± 0.10%	0.96	0.12648 ± 0.03%	2049.7 ± 1.0
C 4,149	40	389.1	131	7223	40	10.1	0.3121 ± 0.10%	5.430 ± 0.11%	0.96	0.12617 ± 0.03%	2045.2 ± 1.1
D 2,149	28	355.7	122.9	12898	15	10.0	0.3209 ± 0.17%	5.597 ± 0.18%	0.97	0.12650 ± 0.04%	2049.8 ± 1.4
70 (Fig. 2-4j) - Sillimanite schist from basement paragneiss; located 4.8 km SE of Pettipiece Pass at 2130 m, 390010E-5692570N (UTM zone 11, BCGS map sheet 82M.038)											
1* 1,202	29	5167	113.7	1224	61	68.1	0.007751 ± 0.11%	0.05008 ± 0.19%	0.75	0.04686 ± 0.13%	41.9 ± 6.1
2* 3,105	16	3775	1395	6083	32	86.2	0.0528 ± 0.09%	0.853 ± 0.10%	0.95	0.11720 ± 0.03%	1914.0 ± 1.2
3 4,105	12	5068	108.3	1717	18	67.0	0.007789 ± 0.15%	0.05041 ± 0.22%	0.72	0.04693 ± 0.16%	45.9 ± 7.4
4 5,74	18	4312	103.5	1161	33	70.9	0.007724 ± 0.11%	0.04998 ± 0.19%	0.69	0.04693 ± 0.14%	45.8 ± 6.8
79 (Fig. 2-4k) - Amphibolitic gneiss layer, 0.5 m thick, folded, contains older gneissosity, intruded into basement orthogneiss; located in cliffs SW of lake 4 km SSE of Pettipiece Pass at 2250 m, 389020E-5692490N (UTM zone 11, BCGS map sheet 82M.038)											
T1 11,149	202	10.41	3.749	100	335	38.9	0.2292 ± 0.45%	3.554 ± 1.4%	0.65	0.11245 ± 1.1%	1839.3 ± 40.6
T2 20,149	229	11.26	4.157	107	381	39.7	0.2320 ± 0.41%	3.601 ± 1.3%	0.66	0.11257 ± 1.0%	1841.3 ± 37.7
T3 40,149	264	10.13	3.391	87	448	40.9	0.2064 ± 0.46%	3.193 ± 1.6%	0.67	0.11221 ± 1.3%	1835.5 ± 48.4
84 Pettipiece pegmatite (Fig. 2-4h) - Pegmatite dyke, 1.5 m thick, slightly deformed, postdates older gneissosity, intruded into basement orthogneiss; located 0.4 km E of toe of Bourne Glacier at 1860 m, 389940E-5687940N (UTM zone 11, BCGS map sheet 82M.038)											
A* 3,105,cl,crs	18	409	139.5	24629	6	6.3	0.3309 ± 0.08%	5.632 ± 0.10%	0.96	0.12344 ± 0.03%	2006.5 ± 1.0
B* 6,105,tan,tp	24	2208	579.2	73141	12	2.1	0.2689 ± 0.09%	4.135 ± 0.10%	0.96	0.11155 ± 0.03%	1824.8 ± 1.1
C 1,149,cl,crs	14	709.1	232.5	9825	19	9.5	0.3089 ± 0.24%	4.952 ± 0.24%	0.97	0.11627 ± 0.06%	1899.6 ± 2.1
D 1,149,tan,tp	21	2308	519	15156	43	1.4	0.2321 ± 0.14%	3.561 ± 0.15%	0.98	0.11129 ± 0.03%	1820.6 ± 1.0
E 2,tan,tp	9	2301	517.7	23976	12	0.8	0.2331 ± 0.08%	3.661 ± 0.10%	0.96	0.11393 ± 0.03%	1863.0 ± 1.0
F 1,tan,tp	7	606.8	129.6	4126	14	1.0	0.2214 ± 0.11%	3.388 ± 0.12%	0.95	0.11098 ± 0.04%	1815.5 ± 1.4
1 1,149	40	1655	6000	31979	37	92.1	0.2979 ± 0.09%	4.599 ± 0.11%	0.95	0.11198 ± 0.03%	1831.8 ± 1.2
2 2,149	30	1480	4284	15973	48	90.2	0.2935 ± 0.12%	4.523 ± 0.13%	0.96	0.11176 ± 0.03%	1828.2 ± 1.3
3 1,149	34	2175	6529	31865	43	90.2	0.3039 ± 0.11%	4.697 ± 0.12%	0.97	0.11209 ± 0.03%	1833.6 ± 1.0
4 1,149	53	1894	6111	3608	443	91.8	0.2750 ± 0.22%	4.236 ± 0.23%	0.98	0.11174 ± 0.04%	1827.9 ± 1.5
5 1,149	29	1875	1865	35198	28	70.3	0.3074 ± 0.10%	4.751 ± 0.11%	0.96	0.11209 ± 0.03%	1833.6 ± 1.1

Table 2-1. (continued).

Analysis ^a	Wt. ^b (μg)	U (ppm)	Pb ^{c*} (ppm)	²⁰⁶ Pb/ ²⁰⁴ Pb ^d	Pb _c ^e ²⁰⁸ Pb/ (pg) (%)	²⁰⁶ Pb/ ²³⁸ U ^f	²⁰⁷ Pb/ ²³⁵ U ^g	Corr. coef.	²⁰⁷ Pb/ ²⁰⁶ Pb ^g	²⁰⁷ Pb/ ²⁰⁶ Pb ^h age (Ma)
85 (Fig. 2-4j) - Sillimanite schist from basement paragneiss; located 0.7 km E of toe of Bourne Glacier at 1880 m, 390130E-5687830N (UTM zone 11, BCGS map sheet 82M.038)										
A 1,105	15	276.1	86.79	2482	30	7.3 0.3044 ± 0.09%	4.752 ± 0.11%	0.89	0.11324 ± 0.05%	1852.1 ± 1.8
B 1,105	15	280.4	93.16	3289	26	4.1 0.3301 ± 0.09%	5.590 ± 0.10%	0.92	0.12282 ± 0.04%	1997.6 ± 1.5
C 1,105	5	245.3	92.46	3473	8	4.0 0.3699 ± 0.20%	6.996 ± 0.21%	0.95	0.13717 ± 0.07%	2191.8 ± 2.4
D 1,105	5	470.3	176.6	3774	14	4.3 0.3650 ± 0.13%	7.316 ± 0.14%	0.97	0.14538 ± 0.04%	2292.3 ± 1.2
1* 2,105	14	5077	1803	7090	75	63.7 0.1328 ± 0.09%	2.275 ± 0.11%	0.96	0.12424 ± 0.03%	2018.0 ± 1.1
2* 2,105	15	6042	125.5	691	68	65.4 0.007946 ± 0.13%	0.05101 ± 0.28%	0.68	0.04656 ± 0.21%	26.7 ± 10.3
3 1,149	10	5355	1669	10636	31	67.7 0.1037 ± 0.13%	1.741 ± 0.14%	0.96	0.12170 ± 0.04%	1981.3 ± 1.4
4 1,105	17	2977	3378	28821	36	71.1 0.3369 ± 0.10%	5.900 ± 0.11%	0.97	0.12702 ± 0.03%	2057.1 ± 1.0
5 4,74	9	6633	136.3	697	42	64.7 0.008014 ± 0.13%	0.05172 ± 0.41%	0.51	0.04681 ± 0.37%	39.3 ± 17.5
91 Pettipiece pegmatite (Fig. 2-4i) - Pegmatite dyke, 0.5 m thick, postdeformational, intruded into basement orthogneiss; located 1.4 km SSE of toe of Bourne Glacier, in cliffs along glacier edge at 1990 m, 390220E-5686730N (UTM zone 11, BCGS map sheet 82M.038)										
A 4,149	97	1186	375.4	83614	26	4.7 0.3151 ± 0.11%	4.899 ± 0.12%	0.97	0.11274 ± 0.03%	1844.1 ± 1.0
B 5,149	74	2165	634.4	120008	24	2.1 0.3000 ± 0.28%	4.659 ± 0.28%	0.99	0.11265 ± 0.03%	1842.7 ± 1.0
C 1,149	59	1234	408.2	45894	30	8.1 0.3176 ± 0.11%	4.949 ± 0.12%	0.97	0.11301 ± 0.03%	1848.3 ± 1.0
D 1,149	30	1026	325.2	35379	17	7.7 0.3057 ± 0.26%	4.766 ± 0.26%	0.99	0.11310 ± 0.03%	1849.8 ± 1.2
E 1,149	35	1391	421.1	48575	19	4.3 0.3029 ± 0.10%	4.702 ± 0.11%	0.97	0.11258 ± 0.03%	1841.4 ± 1.0
1 2,149	43	4646	6199	50688	66	79.3 0.2870 ± 0.17%	4.420 ± 0.17%	0.99	0.11171 ± 0.03%	1827.5 ± 1.0
2 1,149	31	5734	7492	17256	158	80.9 0.2588 ± 0.17%	3.975 ± 0.18%	0.99	0.11137 ± 0.03%	1821.8 ± 1.1
3 1,149	13	3612	5938	34832	25	81.9 0.3087 ± 0.09%	4.776 ± 0.10%	0.96	0.11223 ± 0.03%	1835.8 ± 1.1
4 1,149	20	4030	4892	40421	34	76.8 0.2933 ± 0.10%	4.519 ± 0.11%	0.97	0.11176 ± 0.03%	1828.3 ± 1.0
5 1,149	11	1238	3108	10225	20	90.3 0.2517 ± 0.04%	3.902 ± 0.07%	0.93	0.11245 ± 0.03%	1839.4 ± 1.1
291 (Fig. 2-4d) - Pegmatite layer 1 m thick, deformed, postdates older gneissosity, intruded into basement orthogneiss; located in cliffs S of lake 4 km S of Pettipiece Pass at 2195 m, 388290E-5692520N (UTM zone 11, BCGS map sheet 82M.037)										
A 2,149	40	474	147.3	24623	14	8.0 0.2956 ± 0.16%	5.068 ± 0.17%	0.97	0.12435 ± 0.04%	2019.6 ± 1.3
B 2,149	32	583.5	195.7	1093	307	10.4 0.3108 ± 0.14%	5.306 ± 0.18%	0.87	0.12379 ± 0.09%	2011.5 ± 3.3
C 2,149	19	603.7	204.8	23740	10	8.9 0.3191 ± 0.11%	5.532 ± 0.12%	0.97	0.12573 ± 0.03%	2039.0 ± 1.0
D 2,149	27	495.3	137.1	3024	67	8.2 0.2625 ± 0.09%	4.512 ± 0.11%	0.92	0.12464 ± 0.04%	2023.7 ± 1.5
E 1,149	14	487.9	143.5	11948	10	6.4 0.2851 ± 0.08%	4.823 ± 0.10%	0.96	0.12268 ± 0.03%	1995.6 ± 1.1
F 1,149	12	609.1	179.3	13441	10	4.9 0.2906 ± 0.09%	4.826 ± 0.10%	0.96	0.12043 ± 0.03%	1962.6 ± 1.1
G 1,149	13	343.5	113.4	11025	8	8.0 0.3143 ± 0.11%	5.346 ± 0.12%	0.97	0.12337 ± 0.03%	2005.5 ± 1.1
308 Pettipiece pegmatite (Fig. 2-4g) - Pegmatite dyke, 3 m thick, slightly deformed, postdates older gneissosity, intruded into basement orthogneiss; located 1.4 km SSE of toe of Bourne Glacier, in cliffs along glacier edge at 2030 m, 390300E-5686680N (UTM zone 11, BCGS map sheet 82M.038)										
A 1,149,crs	12	538.1	186.7	14277	9	8.1 0.3301 ± 0.10%	5.608 ± 0.11%	0.96	0.12320 ± 0.03%	2003.1 ± 1.1
B 2,149,tan,tp	34	2467	564.3	6472	173	3.6 0.2308 ± 0.12%	3.552 ± 0.14%	0.97	0.11163 ± 0.03%	1826.1 ± 1.2
C 3,105,crs,tp	22	1619	406.3	6696	80	3.0 0.2545 ± 0.09%	3.933 ± 0.10%	0.95	0.11207 ± 0.03%	1833.2 ± 1.2
D 2,105,crs,tp	18	1482	426.2	6422	72	4.0 0.2887 ± 0.10%	4.478 ± 0.11%	0.95	0.11248 ± 0.03%	1839.9 ± 1.2
E 3,105,crs,tp	10	1926	549.6	21570	16	4.2 0.2860 ± 0.09%	4.430 ± 0.11%	0.96	0.11234 ± 0.03%	1837.5 ± 1.1
311 Feline gneiss (Fig. 2-4a) - Quartz dioritic orthogneiss; located 2.0 km SE of toe of Bourne Glacier at 2355 m, 391230E-5686880N (UTM zone 11, BCGS map sheet 82M.038)										
A 1,149	10	233.7	100	4643	11	12.7 0.3804 ± 0.12%	7.407 ± 0.13%	0.94	0.14122 ± 0.05%	2242.2 ± 1.6
B 1,149	10	536.4	265	9220	14	18.3 0.4100 ± 0.11%	8.153 ± 0.12%	0.95	0.14423 ± 0.04%	2278.6 ± 1.4
C 2,149	22	219	96.88	9432	13	12.5 0.3941 ± 0.15%	7.671 ± 0.16%	0.99	0.14119 ± 0.03%	2241.8 ± 1.1
D 1,149	26	178.4	71.4	9207	11	11.3 0.3627 ± 0.10%	6.891 ± 0.11%	0.96	0.13780 ± 0.03%	2199.7 ± 1.1
E 1,149	23	199	82.86	7680	13	12.2 0.3727 ± 0.09%	7.174 ± 0.10%	0.96	0.13962 ± 0.03%	2222.5 ± 1.1
F 1,149	6	130.9	60.55	2888	7	13.4 0.4070 ± 0.26%	8.049 ± 0.26%	0.98	0.14344 ± 0.05%	2269.2 ± 1.7
G 1,105	7	131.1	58.44	2027	11	12.6 0.3965 ± 0.25%	7.715 ± 0.26%	0.98	0.14111 ± 0.05%	2240.9 ± 1.7

Table 2-1. (concluded).

Analysis ^a	Wt. ^b (μg)	U (ppm)	Pb* ^c (ppm)	²⁰⁶ Pb/ ²⁰⁴ Pb ^d	Pb _c ^e (pg)	²⁰⁸ Pb/ (%) ^f	²⁰⁶ Pb/ ²³⁸ U ^g	²⁰⁷ Pb/ ²³⁵ U ^g	Corr. coef.	²⁰⁷ Pb/ ²⁰⁶ Pb ^g	²⁰⁷ Pb/ ²⁰⁶ Pb ^h age (Ma)
324A Bourne granite (Fig. 2-4e,f) - 0.5 km wide stock, moderately deformed, postdates older gneissosity, intruded into basement orthogneiss; located 3.0 km SE of Pettipiece Pass at 2320 m, 389380E-5693960N (UTM zone 11, BCGS map sheet 82M.038)											
A 1,149	36	287.6	89.2	7481	24	10.4	0.2892 ± 0.27%	4.639 ± 0.27%	0.98	0.11632 ± 0.05%	1900.4 ± 1.7
B 1,149	33	356.8	111.5	5491	39	6.5	0.3034 ± 0.15%	4.993 ± 0.16%	0.96	0.11934 ± 0.04%	1946.4 ± 1.6
C 1,149	13	1180	291.6	19720	12	2.5	0.2497 ± 0.10%	4.208 ± 0.11%	0.97	0.12223 ± 0.03%	1989.0 ± 1.1
D 3,105	15	476.9	150.2	12827	10	6.5	0.3048 ± 0.09%	5.149 ± 0.10%	0.96	0.12251 ± 0.03%	1993.0 ± 1.1
T1 22,149	364	14.93	2.934	127	476	18.4	0.1675 ± 0.35%	2.590 ± 1.1%	0.66	0.11219 ± 0.92%	1835.1 ± 33.4
T2 33,149	350	20.07	3.099	128	489	17.0	0.1345 ± 0.27%	1.997 ± 1.1%	0.69	0.10773 ± 0.92%	1761.4 ± 33.6
T3 54,149	298	25.41	2.218	89	454	17.5	0.0756 ± 0.40%	1.106 ± 1.7%	0.70	0.10607 ± 1.5%	1732.9 ± 53.3
337 (Fig. 2-4c) - Amphibolite dyke 0.3 m thick, postdates older gneissosity, intruded into basement orthogneiss; located in cliffs SW of lake 4 km SSE of Pettipiece Pass at 2260 m, 389070E-5692230N (UTM zone 11, BCGS map sheet 82M.038)											
A 4,149	67	362.4	115.2	20826	21	9.1	0.2986 ± 0.09%	5.151 ± 0.10%	0.96	0.12510 ± 0.03%	2030.3 ± 1.0
B 4,149	66	379.4	131.1	22027	23	9.2	0.3239 ± 0.09%	5.619 ± 0.10%	0.96	0.12583 ± 0.03%	2040.6 ± 1.0
C 4,149	57	590.1	199.9	40683	16	9.2	0.3127 ± 0.10%	5.420 ± 0.11%	0.96	0.12571 ± 0.03%	2038.8 ± 1.1
D 1,105	10	299.1	87.66	3173	16	8.3	0.2777 ± 0.11%	4.788 ± 0.13%	0.94	0.12507 ± 0.04%	2029.8 ± 1.5
E 1,105	12	439.9	157	10058	11	8.2	0.3378 ± 0.09%	5.898 ± 0.11%	0.96	0.12663 ± 0.03%	2051.6 ± 1.1
F 1,105	8	418.5	143.9	8824	8	10.0	0.3190 ± 0.12%	5.580 ± 0.13%	0.97	0.12687 ± 0.03%	2055.1 ± 1.1
T1 17,149	342	46.49	2.674	41	1567	28.1	0.0435 ± 1.1%	0.618 ± 5.0%	0.70	0.10306 ± 4.3%	1680.0 ± 168
T2 22,149	355	40.1	1.502	29	1544	35.0	0.0257 ± 2.1%	0.351 ± 9.7%	0.70	0.09906 ± 8.4%	1606.5 ± 351
T3 27,149	317	42.81	0.705	24	1040	35.5	0.0114 ± 3.4%	0.132 ± 18.2%	0.70	0.08426 ± 16.1%	1298.5 ± 799
PCA-305B-83 Bourne granite (Fig. 2-4e,f) - Grey granite dyke, slightly deformed, postdates older gneissosity, intruded into basement orthogneiss, collected by R.R. Parrish, previous data in Armstrong et al. (1991); located 0.5 km NNW of toe of Bourne Glacier at 1860 m, 389330E-5688340N (UTM zone 11, BCGS map sheet 82M.038)											
1 1,105	11	1159	3301	10103	23	89.7	0.3038 ± 0.08%	4.722 ± 0.10%	0.96	0.11274 ± 0.03%	1844.1 ± 1.1
2 1,105	10	1081	3265	5357	34	90.8	0.2894 ± 0.09%	4.463 ± 0.10%	0.95	0.11187 ± 0.03%	1830.0 ± 1.2
3 1,105	15	624	1490	9530	17	88.5	0.2848 ± 0.09%	4.412 ± 0.10%	0.95	0.11236 ± 0.03%	1837.9 ± 1.2

Isotopic composition of Carleton University lab blank (uncertainty is one standard deviation) - 206:207:208:204 = 19.01 ± 0.36 : 15.64 ± 0.20 : 38.23 ± 0.74 : 1

^aIn first column, A-G = fraction codes for zircon analyses, 1-5 = fraction codes for monazite analyses, T1-T3 = fraction codes for titanite analyses, * = chemical analysis done at the Geological Survey of Canada in Ottawa (all other analysis was done at Carleton University); in second column, 1-54 = number of grains analyzed, 105-149 = average size in μm prior to abrasion with pyrite (Krogh 1982), cl = clear, crls = colourless, tan = tan colour, tp = tips of grains.

^bWeighing uncertainty is 1 μg.

^cRadiogenic Pb.

^dMeasured ratio, corrected for spike and Pb fractionation of 0.09% ± 0.03%/AMU.

^eTotal common Pb in analysis, corrected for spike and fractionation.

^fRadiogenic ²⁰⁸Pb, expressed as % of total radiogenic Pb.

^gCorrected for blank Pb and U and common Pb (Stacey-Kramers model Pb composition equivalent to the interpreted age of the analyses); errors are one standard deviation in %.

^hErrors are two standard deviations in Ma.

Table 2-2. Summary of U-Pb ages.

Sample	Rock type	Key relationship	Age (Ma)	Mineral
311 - Feline gneiss	quartz dioritic orthogneiss	oldest basement orthogneiss	2272 ± 8	zircon
60C - Pettipiece gneiss	granitic orthogneiss	predates older gneissosity	2099 ± 17	zircon
65 - Frenchman Cap gneiss	augen orthogneiss	predates older gneissosity	2077 ± 2	zircon
337	amphibolite dyke	postdates older gneissosity	2051 ± 12	zircon
291	pegmatite dyke		~ 1.97 Ga ?	zircon
187 - Kirbyville gneiss	granodioritic orthogneiss	basement in Kirbyville anticline	1862 ± 1	zircon & titanite
341 - Pettipiece pegmatite	pegmatite layer	intruded into cover sequence	1852 ± 4 58-55 51	zircon xenotime monazite
PCA-305B-83, 60B, 324A - Bourne granite	granite dyke	postdates older gneissosity	1848 ± 3	zircon & monazite
308 - Pettipiece pegmatite	pegmatite dyke	postdates older gneissosity	1845 ± 3	zircon
84 - Pettipiece pegmatite	pegmatite dyke	postdates older gneissosity	1836 ± 2	zircon & monazite
91 - Pettipiece pegmatite	pegmatite dyke	postdates older gneissosity	1834 ± 2	monazite
180	leucogranitic layer	intruded into cover sequence	1762 ± 6 64-60	zircon monazite
70 & 85	pelitic schist	metamorphic monazite	2057 ± 13	monazite
79	amphibolitic gneiss	metamorphic titanite	1854 ± 13	titanite

Figure 2-1. Crystalline basement map of the western Canadian Shield, based on geophysical properties and U-Pb data presented in Villeneuve et al. (1993), showing the position of the four Paleoproterozoic exposures in the Cordillera. The approximate restored positions of the Malton complex and Sifton Ranges gneiss (shown in white) are taken from McDonough and Parrish (1991). GSL = the Great Slave Lake shear zone.

Figure 2-2. Tectonic assemblage map in the vicinity of the Monashee complex, modified after Wheeler and McFeely (1991), showing the location of Fig. 2-3. The dark grey square in inset locates the map with respect to the morphogeologic belts of the Canadian Cordillera.

Figure 2-3. Geologic map of northern Frenchman Cap dome, modified after Wheeler (1965), McMillan (1973), Psutka (1978), Höy and Brown (1980), Journeay (1986), and Höy (1987), showing localities of U-Pb samples (sample numbers are shown in parentheses) discussed in text. D = deformation during the ~2.06 Ga event.

Figure 2-4. U-Pb concordia plots. Ellipses for the analyses represent the two sigma uncertainty (95% confidence level) and analyses that are plotted with +, x, and * symbols have ellipses that are too small to be seen at this scale. Zircon analyses are denoted with letters, monazite analyses are denoted with numbers, and titanite analyses are denoted with T1-T3. See Table 2-1 for analytical data.

 

Figure 2-5. Timeline showing the age of Paleoproterozoic geologic events in the Monashee complex.

Figure 2-6a. Sample 337 is from the 2051 ± 12 Ma amphibolite dyke (the horizontal black layer) that is highly discordant to the nearly vertical folded gneissosity in the host orthogneiss, dated nearby at 2077 ± 2 Ma (Frenchman Cap augen gneiss (sample 65)), and amphibolitic gneiss layers (similar to sample 79). The view is to the southwest. These relationships suggest that the gneissosity and tectonically produced compositional layering developed between 2077 ± 2 and 2051 ± 12 Ma.

Figure 2-6b. Line drawing of (a), labeling details.

Figure 2-7a. Bourne granite, dated at 1850 ± 4 Ma from samples PCA-305B-83, 60B, and 324A, exists in layers and small stocks that are highly discordant to the gneissosity and migmatitic layering in the host gneiss. The relationships seen in this photograph and in (b) and (c) suggest that the gneissosity and migmatitic layering developed prior to granite intrusion.

Figure 2-7b. The granite contains a planar fabric that parallels the axial surfaces of folds that deformed the gneissosity and migmatitic layering in the host. These folds and the axial planar fabric are thought to have a Cordilleran origin based on nearby dating.

Figure 2-7c. This layer of Bourne granite is highly discordant to the gneissosity and migmatitic layering in the host gneiss. The layer contains a planar fabric that parallels the axial surfaces of folds that deformed the gneissosity and migmatitic layering in the host. The view is to the southwest. These folds and the axial planar fabric are thought to have a Cordilleran origin based on dating in nearby rocks (Chapter 4).

Figure 2-7d. Line drawing of (c), labeling details.

Figure 2-8a. Samples 308 and 91, dated at 1845 ± 3 Ma and 1834 ± 2 Ma, respectively, are from nearly vertical pegmatite dykes in a cliff at the toe of Bourne Glacier. The view is to the east. The dykes are highly discordant to the gently west-dipping gneissosity in the host gneisses, suggesting that the gneissosity developed prior to dyke intrusion.

Figure 2-8b. Line drawing of (a), labeling details.

Figure 2-8c. A detailed view of sample 91 pegmatite (see (b) for location) shows that the margins of the dyke sharply truncate the gneissosity in the host and that the dyke lacks deformation fabrics. These relationships are used to propose that rocks in this locality were little strained since 1834 ± 2 Ma.

Figure 2-9. Sample 84 is from the 1836 ± 2 Ma pegmatite that is gently folded and discordant to the gneissosity in the host augen orthogneiss, which is similar to nearby 2077 ± 2 Ma orthogneiss (sample 65). The dyke contains biotite that is aligned parallel to the axial surface of the folded dyke and to the gneissosity in the host. These relationships are used to propose that the gneissosity mostly developed prior to 1836 ± 2 Ma.

CHAPTER 3

U-PB GEOCHRONOLOGIC CONSTRAINTS ON THE AGE AND INTENSITY OF CORDILLERAN METAMORPHISM IN THE MONASHEE COMPLEX, BRITISH COLUMBIA

ABSTRACT

The Monashee complex lies beneath a crustal-scale thrust fault, the Monashee décollement, in a deep structural exposure in the southern Canadian Cordillera. An inverted metamorphic sequence is considered to exist across the décollement into the complex because sillimanite - K-feldspar rocks overlie kyanite-sillimanite-muscovite $\pm$ staurolite rocks. U-Pb isotopic data from the northern part of the complex show that the age and intensity of this metamorphism, which is a result of Cordilleran orogenesis, vary with structural level. The amount of Pb loss that occurred during this event, as determined from linear arrays of discordant U-Pb isotopic analyses, is used to interpret the thermal significance of concordant and reversely discordant monazite and xenotime U-Pb dates that young downward from ~64 Ma in the highest level to ~49 Ma in the deepest level. Upper intercepts of the arrays are Paleoproterozoic and Cretaceous, and lower intercepts suggest that the discordance in the analyses, some of which are composed of single grains, resulted from an early Tertiary overprint. The discordance is mostly attributed to Pb loss rather than mineral overgrowth because the samples generally lack U-

Pb evidence for growth of the U-Pb accessory minerals at the lower intercept ages. The amount of Pb loss in the deepest structural level (6-22% in monazite and 25-31% in titanite) is less than that in moderately higher rocks (52-65% in titanite) and significantly less than that in much higher rocks (~30-56% in monazite and 94-96% in titanite).

Quantitative estimates of the metamorphic intensity are made by modeling Pb diffusion in episodes of heating of fixed duration and temperature, specific combinations of which were calculated by solving equations for measured fractional Pb loss and assumed diffusion radius and geometry. Diffusion coefficients are taken from natural and experimental data. If peak temperatures were ~700-750°C and similar throughout all structural levels, consistent with mineralogy and a petrogenetic grid, then Pb loss occurred within 1 Ma in the deeper levels and within 2-5 Ma in higher levels. However, the differences in Pb loss amounts could also be due to the attainment of greater temperatures (up to 50°C) in higher levels. The knowledge that Pb loss in monazite that existed throughout this metamorphism was incomplete leads to the interpretation that the downward younging dates of 64-49 Ma from monazite and xenotime closely reflect the ages of mineral growth. These growth ages are taken to closely represent the ages of peak metamorphism based on (i) the calculated short durations of elevated temperatures, (ii) previously obtained $^{40}\text{Ar}/^{39}\text{Ar}$ dates that indicate rapid cooling occurred immediately after the peak, (iii) the widespread and consistent nature of the U-Pb dates, and (iv) U-Pb ages that indicate kyanite- and sillimanite-bearing leucosomes were generated at this time. Major conclusions made from these interpretations are that peak metamorphism in the

Monashee complex was diachronous according to structural level and metamorphic intensity increased upward.

INTRODUCTION

The Monashee complex in the core zone of the southern Canadian Cordillera is exposed in the tectonic window through the Monashee décollement (Figs. 3-1, 3-2; Read and Brown 1981), a crustal-scale ductile thrust fault. Middle and upper amphibolite facies metamorphism during Cordilleran orogenesis resulted in an inverted metamorphic sequence that exists across the décollement downward into the northern part of the complex (McMillan 1973; Psutka 1978; Brown 1980; Journeay 1986; Scammell 1986; Höy 1987). U-Pb isotopic data are used in this chapter to quantitatively investigate the age and intensity of the Cordilleran metamorphic history in greater detail than was previously done. U-Pb data from minerals that existed prior to this metamorphism are used to interpret the thermal significance of U-Pb dates from minerals that grew during Cordilleran metamorphism. For example, if it could be shown that minerals grown early in the metamorphism lost substantial amounts of Pb during the thermal peak, then U-Pb dates would be interpreted as underestimating the growth ages. On the other hand, if it could be shown that minerals grown early in the metamorphism lost minimal amounts of Pb during the thermal peak, then U-Pb dates would be interpreted as closely representing the growth ages. In addition, the amount of Pb loss in minerals that existed prior to this metamorphism is used to quantitatively estimate the intensity of the overprint. This is done by modeling Pb diffusion in episodes of heating of fixed duration and temperature,

specific combinations of which are calculated by solving equations for measured fractional Pb loss given the diffusion radius and geometry.

The study area in the northern part of the Monashee complex is the preferred locality to investigate the thermal history of the complex because (i) it contains the lowest grade rocks in the complex, which were subjected to conditions that are below the nominal Pb closure temperatures of monazite and titanite (evidence for this postulation is presented in this chapter), and (ii) U-Pb accessory minerals that grew prior to Cordilleran metamorphism exist in all structural levels. The purpose of this chapter is to show that the early Tertiary U-Pb dates, which young structurally downward, closely represent the ages at which the U-Pb accessory minerals grew and the ages of peak metamorphism. Another purpose is to present the evidence for the conclusion that metamorphic intensity (a factor dependent on temperature and duration) decreased downward. This degree of insight into the age and intensity of metamorphism, as determined by U-Pb methods, has not yet been attained in other orogenic belts. The tectonic implications of these constraints on metamorphism, such as those concerning the inverted metamorphic sequence in the complex, are presented in Chapter 4.

TECTONIC SETTING

The Monashee complex lies in the southern Omineca Belt, the metamorphic and plutonic hinterland to the Rocky Mountain Foreland Belt of the Canadian Cordillera that developed subsequent to collision between accreted terranes and the North American craton (Monger et al. 1982). The complex contains one of four known exposures of

Paleoproterozoic (~2.2-1.8 Ga) rocks in the Canadian Cordillera (Parrish 1991 and references therein). Most of the Paleoproterozoic rocks in the complex compose the basement that forms the core of two structural culminations (Fig. 3-1), namely, Frenchman Cap dome in the north and Thor-Odin dome in the south. Basement is unconformably overlain by the Monashee cover sequence, a platformal succession of metasedimentary and subordinate metaigneous rocks that is 2-3 km thick and laterally extensive (Wheeler 1965; Journeay 1986; Scammell and Brown 1990). The lower part of the cover sequence is Paleoproterozoic and the upper part is probably Paleozoic (Chapter 1 and references therein). Basement was affected by Paleoproterozoic metamorphism and deformation (Chapter 2), whereas the cover sequence was affected only by Cordilleran events, which generated middle and upper amphibolite facies metamorphic assemblages, kilometre-scale isoclinal folds, and penetrative planar and linear deformation fabrics (Wheeler 1965; Reesor and Moore 1971; McMillan 1973; Psutka 1978; Brown 1980; Read 1980; Höy and Brown 1980; Journeay 1986; Scammell 1986; Höy 1987; McNicoll and Brown 1995).

The Selkirk allochthon is interpreted as having been thrust onto the Monashee complex on the Monashee décollement (Figs. 3-1, 3-2; Brown 1980; Read and Brown 1981; Brown et al. 1986; Journeay, 1986; McNicoll and Brown 1995). Surface geology and Lithoprobe seismic reflection profiles are used to correlate the décollement with the Late Cretaceous to Paleocene basal thrust beneath the Rocky Mountain Foreland Belt, and as such it is considered to be a structure that links middle crustal strain in the hinterland of the orogen with upper crustal strain in the foreland (Brown et al. 1992; Cook et al. 1992 and references therein). Carr (1992) used U-Pb dating of granitic rocks to show that the

décollement was active in the Paleocene at the southern flank of Thor-Odin dome.

Compressional tectonism in the southern Omineca Belt was followed shortly thereafter by extension of the orogen (Parrish et al. 1988 and references therein). The east margin of the Monashee complex is defined by the Columbia River fault (Figs. 3-1, 3-2; Lane 1984 and references therein), an Eocene brittle-ductile normal fault with displacement estimated at >15 km (Parrish et al. 1988). This fault and a coeval normal fault to the west, the Okanagan - Eagle River system, bound the Shuswap metamorphic complex. Motion on these faults resulted in rapid exhumation of the footwall (Brown and Journeay 1987; Parrish et al. 1988).

INTERPRETATIONS OF THE METAMORPHIC SEQUENCE

The Cordilleran metamorphic sequence in Frenchman Cap dome is interpreted as being inverted (the highest grade rocks at the structurally highest position and grade decreasing systematically downward) based on mapping of pelitic mineral assemblages by previous workers (McMillan 1973; Psutka 1978; Journeay 1986; Scammell 1986; Höy 1987) and during this study (Figs. 3-3, 3-4). Rocks at the lowest metamorphic grade in the Monashee complex exist on the east side of Frenchman Cap dome, at the greatest distance from the exposed trace of the Monashee décollement (Fig. 3-4). The décollement is interpreted as being arched over the dome and presumably is many kilometres above the rocks on the east side. Here the staurolite-out isograd exists within nonmigmatized kyanite- and muscovite-bearing schists. This isograd lies on a P-T petrogenetic grid in the vicinity of ~700°C at >~6 kbar (Fig. 3-5; Spear and Cheney 1989). On the west side of

the dome, a broad kyanite-muscovite $\pm$ staurolite $\pm$ melt zone is overlain by a sillimanite - K-feldspar $\pm$ melt zone that is ~ 0.2 km thick beneath the décollement (Fig. 3-4) and extends upward for at least 10 km above it (Scammell 1993; Johnson 1994). In contrast to kyanite on the east side of the dome, kyanite on the west side is typically partly to wholly replaced by andalusite and sillimanite. Journeay (1986) interpreted the andalusite as being overprinted by a sillimanite - K-feldspar assemblage, but no conclusive evidence for such a petrogenesis was found during this study. It therefore remains uncertain to me whether the andalusite overprint records a distinct heating event or simply pseudomorphing of kyanite during the decompression part of the P-T-t path. Because the second sillimanite isograd has a steep slope on a P-T petrogenetic grid (Fig. 3-5), its presence is interpreted as indicating that higher structural levels attained greater temperatures. The basement and proximal rocks in the overlying cover sequence in the central part of the dome are in a sillimanite-muscovite $\pm$ melt zone. The isograds are essentially parallel and concordant with the Monashee décollement. They do not appear to have been affected by the kilometre-scale isoclinal folds that dominate the structure on the west side of the dome (Fig. 3-4), and hence have been interpreted by Journeay (1986) as forming after the significant development of these folds.

Based on the knowledge that most schists in the dome contain muscovite that appears to have been stable during peak metamorphism and that most schists were generally at or above the staurolite-out isograd, it is concluded that there is a fairly narrow temperature range recorded by the inverted metamorphic sequence. Although it is impossible to quantify this difference without the use of geothermometry, it is suggested

that temperatures throughout the dome were within 50°C of the ~700°C estimate for schists proximal to the staurolite-out isograd, except for the second sillimanite zone schists in which temperatures were probably significantly higher. If the inverted peak temperature range was indeed this narrow, it is probable that geothermometry would be unable to detect the inversion given that the geothermometers typically have errors of $\pm 50^\circ\text{C}$. Estimates of peak temperatures in basement are hindered by the uncertainty of whether a given assemblage reflects Proterozoic or Cordilleran metamorphism, but temperatures are thought to have been similar to those in nearby cover sequence rocks because assemblages are identical.

There is only one constraint on the peak pressure variation in the dome. Because the kyanite-sillimanite isograd has a moderate positive slope on a P-T grid (Fig. 3-5), the presence of kyanite in structurally higher levels and the absence of kyanite in deeper levels indicates increasing temperature and/or decreasing pressure with greater depth. Given that deeper levels are thought to have attained lower temperatures than higher levels, it is suggested that deeper levels attained lower pressures than higher levels. Such a situation requires an inversion of the pressure gradient, mostly likely being accomplished through structural inversion of the metamorphic sequence (i.e., emplacement of higher pressures rocks onto lower pressure rocks).

Journeay (1986) contended that the metamorphic sequence in Frenchman Cap dome represents two events based on pelitic mineral assemblages and other petrologic evidence: (i) a normal thermal gradient (cooler rocks were overlying warmer rocks at a given time) was proposed for an intermediate pressure event during which kyanite,

staurolite, and sillimanite grew, and (ii) an inverted thermal gradient (warmer rocks were overlying cooler rocks at a given time) was proposed for a low pressure overprint during which andalusite pseudomorphed kyanite and a sillimanite - K-feldspar assemblage formed in the highest temperature rocks. A Middle Jurassic to Late Cretaceous timing for these events was proposed based on numerous arguments, none of which involved geochronologic data from the dome. Such a multiple event petrogenesis relies on evidence that the sillimanite - K-feldspar overprint occurred coeval with the andalusite replacement of kyanite. As stated above, I found no such conclusive petrologic evidence for a high temperature overprint occurring at low pressures.

Parrish (1995) used a moderate amount of Pb loss in titanite from a Proterozoic amphibolitic basement gneiss to postulate that the Cordilleran overprint was short-lived (<5 Ma in duration), and it was inferred that metamorphism throughout the dome was of similar duration because there is no structural discontinuity between the basement and the cover sequence. It was contended that this metamorphism was associated with an inverted thermal gradient, with the heat source being the base of the Selkirk allochthon as it was thrust onto the complex. This conclusion, which contrasts with Journeay's (1986) interpretation that only a low pressure overprint was thermally inverted, is based on the short duration of the event and the fact that the base of the allochthon was at higher temperatures than the complex during the event.

PREVIOUS GEOCHRONOLOGY

Uranium-lead dating in the complex was recently reviewed by Parrish (1995) and Carr (1995), so only a brief summary will be given here. Zircon (Coleman 1990), monazite (Coleman 1990, Parkinson 1992), and titanite (Carr 1995) dates of 66-49 Ma from quartzite, pegmatite, paragneiss, and amphibolite in a high structural level of Thor-Odin dome were used as evidence that the cover sequence and basement were significantly metamorphosed in that period. Metamorphic zircon dates of ~60-52 Ma from syenitic orthogneiss (Parrish and Scammell 1988), amphibolite, carbonatite, and pelitic schist (Parrish 1995) in Frenchman Cap dome were used to contend that these cover sequence rocks at a high structural level underwent significant metamorphism at that time. Although Parrish (1995) admitted it is difficult to demonstrate that peak metamorphism was synchronous with zircon growth, he concluded that the "consistency in the pattern of ages argues for a lack of significant amphibolite facies metamorphism in the cover sequence prior to 60 Ma" (p. 1633). Armstrong et al. (1991) postulated that basement was little affected by Cordilleran metamorphism because there are no isotopic data indicating post-Proterozoic magmatism, metamorphism, or migmatization. Sanborn (1996) used hornblende and biotite $^{40}\text{Ar}/^{39}\text{Ar}$ dating to show that the cover sequence underwent rapid cooling in the early Tertiary.

GEOCHRONOLOGY

Zircon, monazite, xenotime, and titanite U-Pb data from 30 samples collected in the Pettipiece Pass area of Frenchman Cap dome (Fig. 3-3) have bearing on the age and

nature of Cordilleran metamorphism. In order to simplify data presentation, the field area is divided into three levels according to structural distance from the Monashee décollement, which defines the west margin of the dome and is interpreted as being arched over it (Journeay 1986). The Anstey Range fault (Figs. 3-3, 3-4; Journeay 1986) is the boundary between upper and middle levels, and the basement-cover unconformity is the boundary between middle and lower levels. However, basement that is exposed in the upper level is considered to be part of the upper level based on its structural position relative to the décollement (i.e., there is no relationship between stratigraphy and structural level). It is emphasized that boundaries between the levels do not have metamorphic age significance because there is an age overlap across them. The structural thickness from the décollement to the deepest basement exposure is 6 km (Fig. 3-4).

Data from the rocks discussed below are presented in part in this chapter and in part in Chapters 1 and 2. U-Pb data and sample locations are given in Table 3-1, ages are shown on the map in Fig. 3-3 and on the cross section in Fig. 3-6, concordia diagrams are depicted in Fig. 3-7, and Fig. 3-8 plots the ages according to structural level. U-Pb data from metamorphic minerals are summarized in Table 3-2 and U-Pb data from accessory minerals in Proterozoic rocks are summarized in Table 3-3.

Analytical methods

Uranium-lead geochronology followed procedures outlined in Chapter 2.

Reporting of U-Pb ages

Because many monazite U-Pb analyses have slight reverse discordance that is attributable to unsupported ^{206}Pb resulting from incorporation of excess ^{230}Th upon crystallization (Schärer 1984; Parrish 1990), the age reported for a sample is the range in $^{207}\text{Pb}/^{235}\text{U}$ ages of the analyses, including two sigma error. Corrections to the $^{206}\text{Pb}/^{238}\text{U}$ ages from intrusive rocks were not made using the method outlined by Schärer because Th/U whole rock ratios were not measured in this study. However, if typical Th/U whole rock ratios from intrusive rocks are taken from rocks with presumed similar heat production, the $^{206}\text{Pb}/^{238}\text{U}$ ages from intrusive rocks in this study can be corrected to equal the $^{207}\text{Pb}/^{235}\text{U}$ ages (i.e., the analyses become concordant). Ratios for U-Pb analyses with relatively low U concentrations were taken from nearby rocks studied by Parrish (1990) and Carr (1992) (ratios of 0.6-1.4), and ratios for U-Pb analyses with higher U concentrations were taken from Schärer (1984) (ratios of 0.1-0.2). As pointed out by Parrish (1990), there is little justification for the use of this correction on metamorphic monazite because it is difficult to determine the Th/U ratios of the fluids in which the mineral grew. Corrections for a deficit in ^{230}Th upon crystallization in zircon and xenotime (Schärer 1984; Parrish 1990) result in a maximum of 0.1 Ma increase in the $^{206}\text{Pb}/^{238}\text{U}$ ages. In some cases, especially for xenotime, this increase was enough to make the analyses concordant.

Most samples have an age spread of at least a few million years that typically is related to structural level, number of analyses in a sample, and number of grains in each analysis. The reported age for a sample (Figs. 3-3, 3-6, 3-8, Table 3-2) is thus not meant

to be an estimate of the age of all monazite in a sample, but rather is simply meant to reflect the ages of the analyses. Reasons why these ages are mostly interpreted as being growth ages are discussed in the section below that considers all the data.

Pelitic and semipelitic schist

Pelitic schist contains aluminosilicates, biotite, muscovite, plagioclase feldspar, quartz, accessory minerals (monazite, zircon, etc.), plus or minus garnet and is typically migmatitic, unless otherwise noted. Semipelitic schist lacks aluminosilicates.

Upper level

Sample 334 is from kyanite schist in the lower part of the cover sequence in the Kirbyville anticline south of Mount Grace (Fig. 3-3). Kyanite is partly pseudomorphed by andalusite. The Kirbyville anticline is a kilometre-scale isocline on the west flank of the dome that is underlain by the Mount Grace syncline (Fig. 3-3; Höy and Brown 1980; Journeay 1986; Höy 1987). Five single-grain monazite analyses (Table 3-1) have $^{207}\text{Pb}/^{235}\text{U}$ ages of 60.8-56.2 Ma (Fig. 3-7c).

Middle level

Samples 25 and 101 are from semipelitic schists in the base of the cover sequence south and southwest of Pettipiece Pass, respectively (Fig. 3-3). Three monazite analyses (one to three grains each) from sample 25 (Table 3-1) have $^{207}\text{Pb}/^{235}\text{U}$ ages of 51.0-50.1

Ma (Fig. 3-7b). Two monazite analyses (two and five grains each) from sample 101 (Table 3-1) have $^{207}\text{Pb}/^{235}\text{U}$ ages of 53.1-52.2 Ma (Fig. 3-7b).

Sample 126 is from nonmigmatitic kyanite schist in the middle part of the cover sequence northeast of Pettipiece Pass (Fig. 3-3). Nine monazite analyses (one to three grains each; Table 3-1) define a discordia line with upper and lower intercepts of $118 \pm 16/-8$ Ma and $37 \pm 23/-30$ Ma, respectively (MSWD = 0.9) (Fig. 3-7d). Discordance (as measured from the upper intercept along the discordia) is 22-44%. The upper intercept is interpreted as the age of monazite growth in this rock rather than the age of detritus because deposition is thought to have occurred at a considerably earlier time. The lower intercept is interpreted as the age of Pb loss rather than the age of growth because there is no visible indication that growth occurred in this rock at that time. The lower intercept overlaps with the known age of Pb loss and growth in nearby samples, which is ~ 56 Ma. Forcing a lower intercept through this age yields an upper intercept of 128 ± 5 Ma (MSWD = 0.9) and discordance is 37-62%.

Sample 165 is from sillimanite schist in the lower part of the cover sequence south of Blais Creek (Fig. 3-3). Six monazite analyses (two to four grains each; Table 3-1) have $^{207}\text{Pb}/^{235}\text{U}$ ages of 62.2-56.4 Ma (Fig. 3-7c). Disregarding the analysis (2) with the largest error ellipse yields a range of 60.9-56.4 Ma. Sample 214 is from sillimanite schist and sample 257 is from semipelitic schist in the upper part of the cover sequence in the core of the Mount Grace syncline south of Mount Grace (Fig. 3-3). Five monazite analyses (one to two grains each) from sample 214 (Table 3-1) have $^{207}\text{Pb}/^{235}\text{U}$ ages of 54.3-51.2 Ma (Fig. 3-7b), and one single-grain xenotime analysis (Table 3-1) has $^{207}\text{Pb}/^{235}\text{U}$ ages of

51.7-51.5 Ma (Fig. 3-7b). Three monazite analyses (three grains each) from sample 257 (Table 3-1) have $^{207}\text{Pb}/^{235}\text{U}$ ages of 57.4-55.5 Ma (Fig. 3-7c).

Lower level

Samples 24 and 112 are from basement sillimanite schists that lie just beneath the unconformity south and north of Pettipiece Pass, respectively (Fig. 3-3). Four monazite analyses (two to six grains each) from sample 24 (Table 3-1) have $^{207}\text{Pb}/^{235}\text{U}$ ages of 52.4-51.6 Ma (Fig. 3-7a). Four monazite analyses (one to two grains each) from sample 112 (Table 3-1) have $^{207}\text{Pb}/^{235}\text{U}$ ages of 51.9-50.3 Ma (Fig. 3-7a). Two single-grain zircon analyses have Proterozoic upper intercepts (J.L. Crowley, unpublished data), but significant discordance (29% and 56%) precludes derivation of meaningful ages. Sample 166 is from nonmigmatitic basement sillimanite schist that lies just beneath the unconformity at the head of Blais Creek (Fig. 3-3). Three monazite analyses (three to five grains each; Table 3-1) have $^{207}\text{Pb}/^{235}\text{U}$ ages of 51.1-50.5 Ma (Fig. 3-7a).

Samples 70 and 85 are from nonmigmatitic kyanite schist and migmatitic sillimanite schist, respectively, that form basement paragneiss horizons that are interlayered with orthogneiss south of Pettipiece Pass (Fig. 3-3). Data from these samples are presented in Chapter 2, but are also discussed and partly shown here (Fig. 3-7a). Four monazite analyses (three from sample 85 and one from sample 70, one to three grains each) define a discordia line with upper and lower intercepts of 2057 ± 13 and 48 ± 7 Ma, respectively (MSWD = 262). Discordance is 10-88%, with one single-grain analysis having the least discordance and another analysis having 74% discordance. Three other

monazite analyses from sample 70 (one to five grains each) have $^{207}\text{Pb}/^{235}\text{U}$ ages of 50.1-49.3 Ma (Fig. 3-7a) and two other monazite analyses from sample 85 (two and four grains each) have $^{207}\text{Pb}/^{235}\text{U}$ ages of 51.6-50.2 Ma (Fig. 3-7a). It cannot be determined for certain whether the single-grain analyses with normal discordance result from ~50 Ma Pb loss or overgrowth that was superposed on ~2.06 Ga grains. However, it is thought for reasons given below that the 10% discordant analysis reflects Pb loss and the 74% discordant analysis reflects Pb loss and overgrowth. The young ages are thought to result from new growth at ~50 Ma because most of the grains exhibit reverse discordance, which indicates that they incorporated excess ^{206}Pb upon crystallization.

Intrusive rocks

Pegmatite, leucogranite, and leucosome samples contain biotite, K-feldspar, quartz, accessory minerals (monazite, zircon, etc.), plus or minus muscovite, plagioclase feldspar, garnet, and aluminosilicates. Interpretation of whether monazite and xenotime ages represent the time of igneous crystallization, metamorphic growth, or cooling below the Pb closure temperature is based on the zircon U-Pb data. U-Pb data from many of the samples are presented in Chapters 1 and 2, but they are also discussed and partly shown in concordia diagrams here (Fig. 3-7) because of their relevance to the metamorphic history. Data from five other dated intrusive rocks are presented in Chapter 4 and are not discussed here because of the lack of direct relevance to the metamorphic history.

Upper level

Sample 180 is from a quartz-rich leucogranitic layer that intruded into the lower part of the cover sequence, ~300 m structurally below the Monashee décollement south of Ratchford Creek and west of Myoff Creek (Fig. 3-3). Its setting and U-Pb data are presented in Chapter 1. It was deformed by an isoclinal synmetamorphic fold, and biotite-rich seams within it contain kyanite of presumed metamorphic origin. Five zircon analyses (one to nine grains each) define a discordia line with upper and lower intercepts of 1762 ± 6 Ma and 76 ± 20 Ma, respectively (MSWD = 2.8, discordance is 24-38%). The upper intercept is interpreted as the igneous crystallization age, and the lower intercept is interpreted as the approximate time of Pb loss and/or overgrowth during Cordilleran metamorphism. Four monazite analyses (one to three grains each) have $^{207}\text{Pb}/^{235}\text{U}$ ages of 63.5-59.6 Ma (Fig. 3-7g).

Sample 187 is from the Kirbyville gneiss, a hornblende biotite granodioritic orthogneiss north of Blais Creek (Fig. 3-3). Its setting and U-Pb data are presented in Chapter 1. Four zircon analyses (one to three grains each) and three multigrain titanite analyses define a discordia line with upper and lower intercepts of 1862 ± 1 and 63 ± 14 Ma, respectively (MSWD = 2.3, zircon discordance is 7-14% and titanite discordance is 94-96%). The upper intercept is the interpreted igneous crystallization age, and the lower intercept is interpreted as the age of Pb loss. The latter interpretation is discussed further below.

Sample 191 is from an aluminous quartz-rich leucosome layer ~0.5 m thick that intruded into kyanite schist in the lower part of the cover sequence, in the lower limb of

the Kirbyville anticline north of Blais Creek (Fig. 3-3). Aluminosilicate minerals constitute ~30% of the sample. It appears as though kyanite grew first, and then ~50% of the grains were pseudomorphed by andalusite and minor sillimanite. A xenocrystic origin for the kyanite is ruled out because of its abundance and significantly greater size than kyanite in the host schist. Kyanite is interpreted as having crystallized from an aluminous igneous melt or an aluminous fluid, along with the zircon and monazite. Therefore, U-Pb dates from these latter two minerals presumably record the time at which the rocks were at pressures suitable for kyanite stability. The layer appears to have been affected by at least some of the synmetamorphic ductile deformation that affected the host because (i) kyanite is strongly lineated in an orientation that is coaxial with the dominant mineral and stretching lineations and (ii) the layer is concordant with the dominant foliation, which defines the domal shape of the Frenchman Cap culmination. These structural relationships are taken to indicate that deformation responsible for the dominant foliation occurred, at least in part, after intrusion. Seven zircon analyses (two to nine grains each; Table 3-1) define two discordia lines that contain the analysis lying closest to concordia (B; Fig. 3-7h). One discordia through three analyses (B, C, F) has upper and lower intercepts of 1723 ± 9 and 57.1 ± 0.4 Ma, respectively (MSWD = 1.5), and another discordia through four analyses (A, B, E, G) has upper and lower intercepts of 1860 ± 12 and 58.4 ± 0.5 Ma, respectively (MSWD = 4.9). The analyses are 95-99% discordant along the lines from the upper intercept (i.e., they lie much closer to the lower intercept). Another zircon analysis (D) lies below the discordias and is thought to have an older average age of inheritance. The upper intercepts of the discordias are interpreted as average ages of

inherited material and the lower intercepts are interpreted as crystallization ages of the aluminous melt or fluid, rather than metamorphic growth ages, because the grains have an euhedral morphology (elongate and prismatic). The lower intercepts are coeval with five monazite analyses (one to three grains each; Table 3-1) that have $^{207}\text{Pb}/^{235}\text{U}$ ages of 59.1-54.9 Ma (Fig. 3-7h). The monazites are interpreted as having crystallized from the aluminous melt or fluid because they share attributes with monazites that presumably crystallized from two other early Tertiary aluminous melts or fluids; they have significant reverse discordance (indicating incorporation of substantial ^{230}Th upon crystallization) and high U contents (0.6-1.0% U). Based on the monazite and zircon ages, the interpreted crystallization age of the melt or fluid is 57.8 ± 1.5 Ma.

Sample 207 is from the Mount Grace gneiss, a syenitic orthogneiss from a concordant, ~12 km long and up to 1 km thick body lying above the Anstey Range fault (Fig. 3-3). Its setting and U-Pb data are presented in Chapter 1. Five zircon analyses (one to seven grains each) define a discordia line with upper and lower intercepts of 724 ± 5 Ma and 53.4 ± 2.9 Ma, respectively (MSWD = 1.3, discordance is 55-85%). The upper intercept is interpreted as the igneous crystallization age, and the lower intercept is interpreted as the age of Pb loss and metamorphic overgrowth, for reasons given below.

Sample 223 is from an aluminous quartz-rich leucosome pod ~1 m wide that intruded into kyanite schist (sample 334) in the lower part of the cover sequence, in the Kirbyville anticline south of Mount Grace (Fig. 3-3). It differs from sample 191 in that (i) aluminosilicate minerals compose a greater proportion of this sample, ~50%, (ii) it is less certain that kyanite grew prior to all of the andalusite, and (iii) it is thought to have

intruded after the synmetamorphic deformation that affected the host schist on the basis that it does not contain ductile deformation fabrics. It is similar to sample 191 in that (i) a xenocrystic origin for the kyanite is ruled out because of its abundance and significantly greater size than kyanite in the host schist, (ii) kyanite and possibly andalusite are interpreted as having crystallized from an aluminous melt or fluid, and (iii) zircon and monazite U-Pb dates are thought to record the time at which the rocks were at depths suitable for kyanite and/or andalusite stability. Three zircon analyses (two to six grains each; Table 3-1) lie slightly below concordia between 59 and 56 Ma (Fig. 3-7f) and overlap in age with the younger of two groups of monazites. The younger monazite group is composed of two analyses (1, 2, composed of two grains each; Table 3-1) with $^{207}\text{Pb}/^{235}\text{U}$ ages of 56.8-55.1 Ma (Fig. 3-7f). These monazites have significant reverse discordance and high U contents ($\sim 1.7\%$ U). The older monazite group is composed of five single-grain analyses (3 to 7; Table 3-1) with $^{207}\text{Pb}/^{235}\text{U}$ ages of 65.4-63.8 Ma (Fig. 3-7f). These monazites have less reverse discordance and lower U contents ($\sim 0.5\%$ U, which is typical for monazite from schist) than the younger grains. The zircon discordance is attributable to any combination of the following: (i) minor inheritance; (ii) inherent slight deficiency of ^{206}Pb due to ^{230}Th fractionation during crystallization (Schärer 1984; Parrish 1990), which would decrease the $^{206}\text{Pb}/^{238}\text{U}$ date by 0.1 Ma; or (iii) inherent slight excess of ^{231}Pa (Schärer 1984), thereby increasing the $^{207}\text{Pb}/^{235}\text{U}$ date. Given that it is difficult to differentiate between these possibilities, a precise age for this rock is not known.

However, because the zircons overlap in age with the younger group of monazites, which have high U and Th concentrations that suggest crystallization from a melt or fluid (similar

to sample 191), the crystallization age is interpreted as corresponding with the part of concordia directly adjacent to the analyses, which is 57.5 ± 1.5 Ma. The older group of monazites are interpreted as grains inherited from the host schist, the assumed source for the melt or fluid. A problem with this explanation is that monazites from the host schist (sample 334), collected ~200 m from sample 223, are not the same age but rather a few million years younger at 60.8-56.2 Ma. A possible reason for this is that sample 223 was derived from rocks that were distal to the host.

Middle level

Sample 258 is from a 0.5 m thick leucosome layer that intruded into pelitic schist in the upper part of the cover sequence, in the core of the Mount Grace syncline south of Mount Grace (Fig. 3-3). It is one of numerous similar-looking layers that contain sillimanite (constituting a minor part of the rock), lie concordant with the dominant foliation, and contain quartz and feldspar that were flattened into parallelism with the margins of the layer. These structural relationships are taken to indicate that deformation responsible for the dominant foliation occurred, at least in part, after intrusion. The amount of deformation that took place prior to crystallization is unknown. Four zircon analyses (one to four grains each; Table 3-1) lie along concordia between 50.5 and 49.5 Ma, and are slightly younger than three single-grain monazite analyses (Table 3-1) with $^{207}\text{Pb}/^{235}\text{U}$ ages of 51.4-50.4 Ma (Fig. 3-7e). The monazites have significant reverse discordance and high U contents (~0.6-1.0% U), similar to monazite that presumably

crystallized from a melt or fluid in two other dates samples (191 and 223). Based on the monazite and zircon data, the interpreted igneous crystallization age is 50.5 ± 1.0 Ma.

Samples 303 and 341 are from similar-looking 1-3 m thick pegmatite layers that intruded into the lower part of the cover sequence, northeast of Pettipiece Pass (Fig. 3-3). U-Pb data from sample 341 are presented in Chapter 2. The layers are strongly boudinaged and concordant with the dominant foliation. These samples were dated to determine the age of pegmatite crystallization in this area and to compare monazite ages with those in nearby schist sample 126. The crystallization age of sample 341 is interpreted as 1852 ± 4 Ma based on the upper intercept of a discordia line defined by five single-grain zircon analyses (MSWD = 9.8, discordance is 19-46%). The lower intercept is 149 ± 20 Ma. Three single-grain xenotime analyses from sample 341 are concordant at 58.2-55.0 Ma and one single-grain monazite analysis is concordant at 51.0-50.4 Ma (Fig. 3-7g). Two nearly concordant single-grain monazite analyses from sample 303 (Table 3-1) have $^{207}\text{Pb}/^{235}\text{U}$ ages of 57.9-57.3 Ma (Fig. 3-7g).

Lower level

The settings and U-Pb data from six Proterozoic rocks that intruded into basement south of Pettipiece Pass are presented in Chapter 2. Most localities are shown in Fig. 3-3 and the data are summarized in Table 3-3 and Fig. 3-9. All U-Pb analyses are discordant. Upper intercept ages from discordia lines are interpreted as the igneous crystallization ages and the lower intercepts are interpreted as the ages of Pb loss and/or growth, for reasons given below. Bourne granite (some data from Armstrong et al. 1991) has two

multigrain zircon analyses, six monazite analyses (including two single-grain analyses), and three multigrain titanite analyses that define a discordia line with upper and lower intercepts of 1848 ± 3 and 95 ± 35 Ma, respectively (MSWD = 4.7, zircon discordance is 15-16%, monazite discordance is 9-18%, and titanite discordance is 52-80%). The titanite analyses are from a different sample than the zircon and monazite analyses, but the data are considered together because both samples were interpreted in the field as being part of the same suite of granites and the collinearity of the data suggest that they are the same age. Sample 337 is from an amphibolite dyke with three multigrain titanite analyses that define a discordia with upper and lower intercepts of $1780 +247/-216$ and $29 +33/-39$ Ma, respectively (discordance is 89-98%). Forcing an upper intercept through 1.85 Ga, the likely initial growth age of the titanite, yields a lower intercept of 38 ± 20 Ma. Sample 308 is from a pegmatite dyke with four zircon analyses (two to three grains each) that define a discordia line with upper and lower intercepts of 1845 ± 3 and 84 ± 19 Ma, respectively (MSWD = 4.2, discordance is 13-31%). Sample 84 is from a pegmatite dyke with five monazite analyses (four are single-grain and another is composed of two grains) and three zircon analyses (two are single-grains and another is multigrain) that define a discordia line with upper and lower intercepts of 1836 ± 2 and 76 ± 15 Ma, respectively (MSWD = 8.3, monazite discordance is 6-17% and zircon discordance is 19-34%). Sample 91 is from a pegmatite dyke with analyses that form two discordias. Three monazite analyses (two single-grain and one composed of two grains) have upper and lower intercepts of 1834 ± 2 and 85 ± 17 Ma, respectively (MSWD = 0.2, discordance is 12-22%). Three zircon analyses (one to five grains each) and a single-grain monazite

analysis have upper and lower intercepts of 1844 ± 2 and 30 ± 20 Ma, respectively (MSWD = 4.3, discordance is 5-22%). Samples OP-90-12 (Parrish 1995) and 79 are from amphibolitic gneiss layers that have multigrain titanite analyses with upper and lower intercepts of $1859 +54/-44$ and $71 +174/-180$ Ma, respectively (MSWD = 0.02, discordance is 26-43%).

METAMORPHIC OVERPRINT OF PREEXISTING MINERALS

Zircon, monazite, and titanite U-Pb analyses from Proterozoic rocks in Frenchman Cap dome form discordia lines (Table 3-3, Fig. 3-9), and monazite U-Pb analyses from Early Cretaceous monazites also form a discordia (Fig. 3-7d). Interpretation of whether the discordance results from Pb loss or growth during the metamorphic overprint is the first step toward determining its thermal significance. This interpretation is made next and is followed by discussion of the thermal significance of the discordance.

Pb loss or growth?

The degree to which the discordance reflects Pb loss or metamorphic growth can only be determined for certain using dating techniques capable of small-scale resolution, such as electron microprobe or ion probe analysis. Because such techniques were not employed in this study, the following inferences are based largely on the available U-Pb data and microscopic observations of the grains.

Discordance in most zircon analyses is thought to reflect Pb loss because the grains typically have igneous-looking morphologies and abrasion (Krogh 1982) of the grains

would have removed any thin overgrowths. Monazite morphology provides little insight into whether overgrowths exist. The discordance in the monazite analyses from samples that do not contain any dated early Tertiary monazites (as is the case for all of the intrusive rock samples and one of the pelitic schist samples) is assumed to result from Pb loss because apparently there was no growth during the overprint. However, there are two cases, one each for zircon and monazite, in which the discordance is thought to be partly due to mineral growth during the overprint. It is assumed that Pb loss and growth occurred simultaneously in these cases because the analyses lie along well-fitting discordias and each process is thought to have been responsible for a nontrivial part of the discordance; Pb loss is considered to be responsible for 10-40% of the discordance based on data from nearby samples. In the first case, significant discordance (55-85%) in zircon analyses (some of which are composed of single grains) from the Proterozoic sample 207 (Mount Grace syenitic orthogneiss) is attributable mostly to overgrowth because the amount of discordance is related to the amount of visible core material. The discordia has a lower intercept of 53.4 ± 2.9 Ma. Parrish and Scammell (1988) made a similar interpretation for the coeval Mount Copeland syenitic orthogneiss in the southern part of the dome based on analyses that plot in the same part of the concordia diagram and on two analyses that are concordant at the lower intercept. In the second case, discordant monazite analyses from schist (samples 70 and 85) are interpreted as follows: (i) a single grain analysis with 10% discordance is thought to mostly reflect Pb loss because it falls within the range of 6-22% discordance measured in monazite from nearby intrusive rocks, which is attributable to Pb loss; (ii) a single grain analysis with 74% discordance is thought

to mostly reflect overgrowth because this amount of discordance is significantly greater than that measured in nearby intrusive rocks and there is ample early Tertiary growth dated in samples 70 and 85 (Fig. 3-7a); and (iii) two multigrain analyses are thought to be combinations of early Tertiary grains and older grains. The resulting discordia has a lower intercept of 48 ± 7 Ma.

Discordance in titanite analyses is more difficult to evaluate because the analyses are composed of a dozen or more grains and it is known that ample titanite growth occurred during Cordilleran metamorphism in similar rock types (J.L. Crowley, unpublished data). Hence, it cannot be ruled out that the analyses reflect a combination of grains that suffered Pb loss and grains that grew during the overprint, as was documented by Verts et al. (1996) in a contact aureole in Wyoming. However, in contrast to the morphological and colour differences amongst the grains found by Verts et al., the grains dated in this study had a uniform appearance. Another piece of evidence that suggests the discordance is largely due to Pb loss is that the amount of discordance within each sample varies consistently with grain size. Smaller grains have a greater amount of discordance than larger grains, which is a pattern that is expected for Pb loss because a shorter diffusion path results in more efficient Pb loss. The grain size was rather uniform within each analysis (all grains were between 202 and 149 μm in diameter, as determined by sieve mesh), so the analyses were picked based on visually estimated grain sizes. The visual size estimates agree with the average grain weights that were calculated for each analysis. The interpretations that are presented below are based on the premise that Pb loss was the reason for nearly all the discordance in titanite from samples 79, 187, and 324A. There is

an isotopic characteristic of the titanite analyses from another sample, 337, that suggests its discordance reflects a combination of grains that suffered Pb loss and grains, or parts of thereof, that grew during the Cordilleran overprint. This characteristic is the concentration of common Pb in the analyses, which is considerably higher in sample 337 (3.2-4.5 ppm) than in samples 79 (1.51-1.58 ppm), 187 (1.45-1.54 ppm), and 324A (1.25-1.45 ppm). A high concentration of common Pb is thought to reflect the presence of titanite that grew during the overprint because common Pb concentrations are high in metamorphic titanites that are known to have grown in the Late Cretaceous - early Tertiary (3.9 ppm in one sample and 7.5-7.9 ppm in another; J.L. Crowley, unpublished data).

Estimated age of the overprint

Because diffusion is a thermally activated process that occurs most rapidly at elevated temperatures, the lower intercept of a discordia that resulted from a substantial amount of Pb loss is interpreted as the age of the Cordilleran thermal overprint. The lower intercepts of such discordias from Proterozoic analyses that presumably only underwent Cordilleran metamorphism are as follows: 149 ± 20 , 84 ± 19 , and 76 ± 20 Ma based on zircon discordias, 76 ± 15 and 30 ± 20 Ma based on discordias formed by monazite and zircon, 85 ± 17 Ma based on a monazite discordia, $71 +174/-180$, 63 ± 14 , 38 ± 20 Ma based on discordias controlled entirely or mostly by titanite, and 95 ± 35 Ma based on a discordia formed by titanite, monazite, and zircon (Table 3-3). A lower intercept of $37 +23/-30$ Ma is given by Pb loss in Early Cretaceous monazite from sample

126. The dispersion in the ages is interpreted to reflect Pb loss that was superimposed on grains that were not coeval prior to the overprint (presumably mostly due to inheritance), resulting in skewed discordia lines. The lower intercept ages that are considered to be the most accurate are discussed below.

Estimated temperature and duration of the overprint

Because diffusion is a thermally activated process, the amount of Pb loss suffered by a grain reflects the intensity (a factor dependent on temperature and duration) of the Cordilleran thermal overprint. The amounts of Pb loss that were measured in titanite and monazite from Proterozoic and Cretaceous samples are used to quantitatively estimate the thermal intensity by modeling diffusion in episodes of heating of fixed duration and temperature, specific combinations of which can be calculated by solving equations for fractional Pb loss, f , and assumed diffusion radius and geometry (Heaman and Parrish 1991 and references therein). For a cylinder (the geometry that is thought to best approximate the shape of the grains) of radius a (the grain size, determined by sieving the grains through meshes of various sizes) and $0 < f < 0.6$,

$$f = 2.257 \times (Dt/a^2)^{1/2} - (Dt/a^2) \quad [1]$$

and for $0.6 < f < 1.0$,

$$f = 1 - 0.692 \times \exp(-5.78Dt/a^2) \quad [2]$$

where t is time, and D is the diffusion coefficient. Values for D were found by solving the Arrhenius relationship

$$D = D_0 \times \exp(-E/RT) \quad [3]$$

where T is temperature in Kelvins, R is the gas constant, D_0 is the diffusion coefficient at infinite temperature, and E is activation energy.

Diffusion parameters for titanite were taken from (i) natural data from rocks that were metamorphosed at temperatures estimated from petrology and for durations estimated from the geologic history (Heaman and Parrish 1991; $D_0 = 1.95 \times 10^{-13} \text{ cm}^2/\text{s}$ and $E = 1.11 \times 10^5 \text{ J/mol}$) and (ii) experimental data (Cherniak 1993; $D_0 = 1.11 \text{ cm}^2/\text{s}$, $E = 3.29 \times 10^5 \text{ J/mol}$). Three fractions of Proterozoic titanite from Frenchman Cap basement amphibolitic gneiss (sample OP-90-12) that were dated by Parrish (1995) have 25-43% discordance. Parrish interpreted the discordance as a result of Pb loss during Cordilleran metamorphism, and upon applying the diffusion parameters taken from natural data to equation [1], assuming a diffusion radius of 75 μm (the measured size of the grains), it was shown that 43% Pb loss occurs in 2 Ma at 600°C and in <1 Ma at >650°C. As discussed above, on the basis of comparison of a petrogenetic grid (Fig. 3-5) with the pelitic mineral assemblages, peak temperatures in the dome are estimated at 700-750°C. Three titanite analyses from a nearby rock of similar lithology (sample 79) were dated in this study. Two analyses with the largest grain size (75 μm radius) indicate 31% Pb loss, which is slightly more than that recorded by two of the three analyses dated by Parrish. The third analysis was composed of grains with a radius of <75 μm , which is considerably smaller than those in the first two analyses and those in the analyses from other dated samples. Predictably, this analysis indicates a greater amount of Pb loss, 39%. Because of the differences in grain size, 25-31% is considered to closely approximate the amount of Pb loss suffered by 75 μm radius grains at this structural level. Application of the

diffusion parameters of Heaman and Parrish (1991) and Cherniak (1993) to 31% Pb loss yields durations at temperatures above 600°C that are slightly shorter than those estimated by Parrish (1995). The temperature-duration relationships that come from solving equations [1] and [2] are plotted in Fig. 3-10.

Two titanite analyses (with 75 µm radius grains) from the Proterozoic Bourne granite (sample 324A) in a structurally high position in the lower level indicate 52 and 62% Pb loss in the early Tertiary. Data from three other analyses (J.L. Crowley, unpublished data), from which only $^{207}\text{Pb}/^{206}\text{Pb}$ dates were recorded, indicate Pb loss amounts of 58-64% when combined with the U-Pb data. An analysis composed of 50 µm radius grains indicates 80% Pb loss. Pb loss of 52-65% is considered to closely approximate the amount of loss suffered by 75 µm radius grains at this structural level. This amount yields maximum durations at 650 and 700°C of 2 and 1 Ma, respectively, using the parameters of Heaman and Parrish (1991), and durations of 1 and ~0 Ma, respectively, using the parameters of Cherniak (1993) (Fig. 3-10). Three titanite analyses with grains of 75 µm radius from the Proterozoic Kirbyville orthogneiss (sample 187) in the upper level indicate 94-96% Pb loss at 63 ± 14 Ma. This amount yields durations at 650 and 700°C of 8 and 4 Ma, respectively, using the parameters of Heaman and Parrish (1991), and durations of 3 Ma and 0.5 Ma, respectively, using the parameters of Cherniak (1993) (Fig. 3-10).

In a similar manner, Pb loss in monazite can be used to calculate specific combinations of temperature and duration. Diffusion parameters were taken from (i) natural Pb loss that was measured in elemental Pb profiles in monazite that was

metamorphosed at temperatures estimated from petrology and for durations estimated from the geologic history (Suzuki et al. 1994; $D_0 = 3.40 \times 10^{-7} \text{ cm}^2/\text{s}$, $E = 2.44 \times 10^5 \text{ J/mol}$) and (ii) experimental data (Smith and Giletti 1997; $D_0 = 6.6 \times 10^{-11} \text{ cm}^2/\text{s}$, $E = 1.8 \times 10^5 \text{ J/mol}$). Single-grain monazite analyses from three Proterozoic granitic rocks (Bourne granite, samples 84 and 91) in the lower level indicate 6-22% Pb loss in the early Tertiary. Using the diffusion parameters of Suzuki et al. (1994) and diffusion radii of 50 and 75 μm (from the measured sizes of the grains) shows that 22% Pb loss occurs in 2 and 4 Ma, respectively, at 650°C and in <1 Ma at 700°C (Fig. 3-10). A comparison of the temperature-time plots in Fig. 3-10 shows that the monazite data yield durations at 650°C that are longer than those calculated from the titanite data by a few million years, but durations at 700°C are similar. It is cautioned that estimates based on the Suzuki et al. (1994) data are strongly dependent on the assumed temperature and duration of the metamorphism that affected the monazite in that study. For example, if the duration is half the assumed amount, which cannot be ruled out by the available geochronologic data, then the above durations are reduced by 50%. Similar results are found upon accounting for uncertainties in the assumed temperature. Application of the experimentally derived diffusion parameters of Smith and Giletti (1997) to the 22% Pb loss yields durations that are similar to those deduced from the Suzuki et al. (1994) parameters.

The 10% discordance in a single-grain monazite analysis from basement schist sample 85 is thought to be due to Pb loss, suggesting shorter durations than those given for 22% Pb loss. The discordance in five Early Cretaceous single-grain monazite analyses from schist sample 126 is thought to be due to Pb loss. Pb loss amounts of 22-44% are

found using the discordia line with an unforced lower intercept. However, the amounts are considered to be higher than 22-44% because using a discordia with a forced lower intercept of 56 Ma, the age of monazite and xenotime growth in nearby rocks and presumably the time of Pb loss, gives 37-62% loss. For the purposes of this discussion, a Pb loss amount of 56% is chosen to represent the maximum amount of Pb loss suffered by the monazite in sample 126. The diffusion radius is generally $\sim 50 \mu\text{m}$ and the amount of Pb loss does not correspond to grain size. Using the diffusion parameters of Suzuki et al. shows that 56% Pb loss occurs in 12 Ma at 650°C and in 2.5 Ma at 700°C (Fig. 3-10). Application of the Smith and Giletti (1997) parameters to the 56% yields durations that are ~ 2 Ma longer than those deduced from the Suzuki et al. (1994) parameters.

In summary, the calculations made using the monazite data agree with the titanite data in suggesting that metamorphism in deeper levels was up to a few million years shorter in duration and/or attained temperatures up to $\sim 50^\circ\text{C}$ less than metamorphism in higher levels (Fig. 3-10).

INTERPRETATION OF THE EARLY TERTIARY U-PB DATES

Growth ages

The first step toward attaching thermal significance to the early Tertiary monazite and xenotime U-Pb dates from Frenchman Cap dome (Figs. 3-3, 3-6 to 3-8, Table 3-3) is determining whether they represent growth or cooling ages, or spurious ages resulting from partial Pb loss. Smith and Barreiro (1990) found conditions sufficient for monazite

growth to be as low as $\sim 525^{\circ}\text{C}$ at 310 MPa (at the first appearance of staurolite), meaning that monazite in the dome could have grown during a postpeak event. The dates probably do not represent cooling ages because there are no known examples in which normal monazite has shown entirely open system behavior, even at 800°C (e.g., Spear and Parrish 1996), but partial Pb loss does occur at elevated temperatures (Parrish 1990). The closure temperature for xenotime, although poorly constrained, is thought to be similar to monazite (Heaman and Parrish 1991). The discussion below presents the interpretation that the U-Pb dates closely represent mineral growth ages, which when combined with other evidence, leads to the conclusion that the dates essentially record the time of peak metamorphism.

The limited amount of Pb loss that occurred during the entire Cordilleran metamorphism, as measured in minerals that existed prior to the event, is considered to be important evidence in the determining the significance of the U-Pb dates. It is clear that this metamorphism failed to cause complete Pb loss: in the lower level, Proterozoic monazite had 6-22% Pb loss and Proterozoic titanite had 25-31% Pb loss in a relatively deep locality and 52-65% Pb loss in higher locality within the lower level; in the middle level, Early Cretaceous monazite had ~ 30 -56% Pb loss; in the upper level, Proterozoic titanite had 94-96% Pb loss. Minerals throughout the dome probably suffered similar amounts of Pb loss because pelitic assemblages (Journeay 1986) suggest that rocks across the dome reached similar temperatures. The relatively minor Pb loss in the lower level supports the interpretation that early Tertiary monazite $^{207}\text{Pb}/^{235}\text{U}$ dates from there are close to the ages of mineral growth. The moderate amount of Pb loss in monazite from

the middle level and the substantial amount of Pb loss in titanite from the upper level suggest that early Tertiary monazite $^{207}\text{Pb}/^{235}\text{U}$ dates from there underestimate growth ages. However, the following two pieces of evidence are used to contend that the actual growth ages in the middle and upper levels are probably not much older, perhaps only a few million years, than the early Tertiary $^{207}\text{Pb}/^{235}\text{U}$ dates. First, many analyses have reverse discordance, which is regarded as being due to unsupported ^{206}Pb that resulted from incorporation of excess ^{230}Th upon crystallization (Schärer 1984; Parrish 1990), so Pb loss cannot have reduced their age significantly (Parrish 1990). Second, the interpreted short duration of high temperatures (only a few million years according to calculations made above) and rapid cooling from peak temperatures (inferred from $^{40}\text{Ar}/^{39}\text{Ar}$ data of Sanborn 1996) require that the maximum of ~50 % Pb loss cannot have reduced the ages significantly. For example, because rocks in the middle level cooled through the Ar closure temperature in hornblende (taken as $530 \pm 40^\circ\text{C}$), which is considered to be a temperature at which monazite suffers negligible Pb loss (Parrish 1990)), at ~53 Ma, then a monazite with a date of 56 Ma cannot have grown prior to ~59 Ma, assuming that the estimate of a maximum of 50% Pb loss is valid.

Peak metamorphic ages

Determining the relationship between peak metamorphism in Frenchman Cap dome and the growth of monazite and xenotime is the next step toward attaching thermal significance to the U-Pb dates. Peak metamorphism cannot have been significantly younger than the mineral growth ages because hornblende and biotite $^{40}\text{Ar}/^{39}\text{Ar}$ dates

(Sanborn 1996) are just a few million years younger than nearby mineral growth ages. Proving that peak metamorphism cannot be older than the mineral ages is a more difficult task, but the following three pieces of evidence are considered to be convincing. First, discordant U-Pb analyses from the Proterozoic minerals yield Pb loss ages that generally overlap the mineral growth ages (Table 3-3, Fig. 3-9). Pb loss ages are considered to be significant, although typically imprecise due to the generally relatively minor amounts of Pb loss, because of the thermally activated nature of diffusion. The most important Pb loss ages are those from analyses that plot closest to the lower intercepts because they yield more precise and accurate ages; the nearly complete Pb loss in titanite from the upper level indicates a thermal event at 63 ± 14 Ma, and the moderate amount of Pb loss in monazite from the middle level indicates a thermal event that is younger than 60 Ma (lower intercept age is $37 +23/-30$ Ma). The linear array of titanite and monazite along discordia lines indicates that no other events caused discordance between the Proterozoic and latest Cretaceous - early Tertiary. This does not rule out lower temperature metamorphism in the interval between these ages, such as in the Early Cretaceous (the time of monazite growth in sample 126), but it does preclude high grade heating. The second piece of evidence for the growth ages reflecting the times of peak metamorphism is that kyanite-bearing leucosome samples 191 and 223 in the upper level crystallized at 57.8 ± 1.5 and 57.5 ± 1.5 Ma, respectively, and sillimanite-bearing leucosome sample 258 in the middle level crystallized at 50.5 ± 1.0 Ma. Although approximate P-T conditions represented by these rocks cannot be estimated because it is unknown (i) whether the aluminosilicates grew during or after crystallization of these rocks and (ii) whether the

rocks crystallized from igneous melts or aluminous fluids, the existence of these minerals supports the conclusion that the host rocks were at considerable depths and temperatures at or after the time of crystallization. Third, all zircon, monazite, xenotime, and titanite U-Pb analyses from Frenchman Cap and Thor-Odin domes record dates in the latest Cretaceous - early Tertiary (Parrish and Scammell 1988; Coleman 1990; Parkinson 1992; Carr 1995; Parrish 1995; this chapter). The occurrence of regional high temperature metamorphism at this time, rather than a period of growth after peak metamorphism, is supported by the consistency of dates throughout the complex, especially with the knowledge that the dearth of older minerals cannot be due to isotopic resetting.

Diachronous metamorphism

The above discussion outlines the interpretation that the monazite and xenotime growth U-Pb dates represent an early Tertiary age of peak metamorphism in Frenchman Cap dome. The step beyond this general interpretation is a more precise interpretation that mineral growth ages from each sample closely represent when it and the surrounding rocks were at their thermal peak. A corollary of this interpretation is that the times in which monazite and xenotime did not grow in a sample are times when lower temperatures existed. Unlike the general interpretation, the more precise interpretation cannot be supported by the ages of Pb loss in preexisting minerals because of the large uncertainties associated with the lower intercept ages. However, the more precise interpretation is supported by the evidence for short durations of elevated temperatures and the consistency of ages with structural level. A conclusion that leads from the more

precise interpretation is that the variation in U-Pb dates with structural level (Figs. 3-3, 3-6 to 3-8) records the diachronous existence of peak metamorphic conditions in the dome. The tectonic implications of diachronous metamorphism and a model to explain it are presented in Chapter 4.

SUMMARY

Uranium-lead geochronologic data from Frenchman Cap dome of the Monashee complex show that the age and intensity of Cordilleran metamorphism vary with structural level. Discordant linear arrays of U-Pb isotopic analyses from preexisting monazite and titanite were used to interpret the thermal significance of concordant and reversely discordant monazite and xenotime U-Pb dates that young downward from ~64 Ma in the highest level to ~49 Ma in the deepest level. The lower intercepts of the arrays suggest that the discordance resulted from an early Tertiary overprint, and the discordance was mostly attributed to Pb loss rather than overgrowth because the samples generally lack U-Pb evidence for mineral growth at the lower intercept ages. The amount of Pb loss increases significantly upward from the deepest structural level to the highest level. Quantitative estimates of the metamorphic intensity were made by modeling Pb diffusion in episodes of heating of fixed duration and temperature, specific combinations of which were calculated by solving equations for measured fractional Pb loss and assumed diffusion radius and geometry. With the use of diffusion coefficients taken from natural and experimental data, it was estimated that if peak temperatures were ~700-750°C and similar throughout all structural levels, consistent with mineralogy and a petrogenetic grid,

then Pb loss occurred within 1 Ma in the deeper levels and within 2-5 Ma in higher levels. However, the differences in Pb loss amounts could also be due to the attainment of greater temperatures (up to 50°C) in higher levels. The knowledge that Pb loss in monazite that existed throughout the event was incomplete led to the interpretation that the downward younging dates of 64-49 Ma from monazite and xenotime closely reflect the ages of mineral growth. These ages were taken to closely represent the thermal peak based on (i) the calculated short durations of elevated temperatures, (ii) previously obtained $^{40}\text{Ar}/^{39}\text{Ar}$ dates that indicate rapid cooling occurred immediately after the peak, (iii) the widespread and consistent nature of the U-Pb dates, and (iv) U-Pb ages that indicate kyanite- and sillimanite-bearing leucosomes were generated at this time. Major conclusions made from these interpretations are that peak metamorphism in the Monashee complex was diachronous according to structural level and metamorphic intensity increased upward.

Table 3-1. U-Pb analytical data.

Analysis ^a	Wt. ^b U (μg)	U (ppm)	Pb* ^c (ppm)	²⁰⁶ Pb ^d / ²⁰⁴ Pb	Pb ^e (pg)	²⁰⁸ Pb ^f (%)	²⁰⁶ Pb ^g / ²³⁸ U	²⁰⁷ Pb ^g / ²³⁵ U	Corr. coeff.	²⁰⁷ Pb ^g / ²⁰⁶ Pb	²⁰⁷ Pb ^h / ²⁰⁶ Pb age (Ma)	
24 (Fig. 3-7a) - Sillimanite schist, basement paragneiss; located 3.2 km SE of Pettipiece Pass at 2400 m, 389720E-5694510N (UTM zone 11, BCGS map sheet 82M.038)												
1*	3,105	19	3491	89.32	1637	22	71.2	0.008138 ± 0.11%	0.05287 ± 0.18%	0.69	0.04711 ± 0.13%	55.0 ± 6.3
2*	2,105	14	3248	106.1	1583	15	77.4	0.008156 ± 0.10%	0.05316 ± 0.19%	0.67	0.04727 ± 0.15%	63.1 ± 6.9
3	3,105	10	4035	118	649	34	74.9	0.008091 ± 0.18%	0.05266 ± 0.49%	0.49	0.04720 ± 0.43%	59.4 ± 20.4
4	6,105	16	3490	94.18	1826	16	72.9	0.008080 ± 0.12%	0.05242 ± 0.18%	0.62	0.04705 ± 0.14%	51.8 ± 6.9
25 (Fig. 3-7b) - Sillimanite schist, lower part of cover sequence; located 3.5 km SE of Pettipiece Pass at 2430 m, 390070E-5694490N (UTM zone 11, BCGS map sheet 82M.038)												
1*	3,149	36	4235	115.4	1506	52	73.7	0.007902 ± 0.12%	0.05117 ± 0.17%	0.79	0.04696 ± 0.10%	47.4 ± 4.9
2	1,149	21	4165	114.4	1493	30	74.0	0.007876 ± 0.44%	0.05098 ± 0.41%	0.68	0.04694 ± 0.34%	46.2 ± 16.4
3	2,105	33	4499	124.5	1413	53	74.1	0.007920 ± 0.11%	0.05133 ± 0.17%	0.77	0.04701 ± 0.11%	49.8 ± 5.3
101 (Fig. 3-7b) - Sillimanite schist, lower part of cover sequence; located along edge of Bourne Glacier at head of east branch of Myoff Creek at 2150 m, 386110E-5686660N (UTM zone 11, BCGS map sheet 82M.037)												
1*	2,105	7	3361	47.73	2480	5	47.7	0.008221 ± 0.09%	0.05347 ± 0.19%	0.61	0.04717 ± 0.15%	57.7 ± 7.4
2	5,74	9	5406	126.5	1536	17	68.5	0.008148 ± 0.11%	0.05293 ± 0.24%	0.58	0.04711 ± 0.19%	54.9 ± 9.3
112 (Fig. 3-7a) - Sillimanite schist, basement paragneiss; located 6.1 km NNW of Pettipiece Pass at 2320 m, 385280E-5702190N (UTM zone 11, BCGS map sheet 82M.047)												
1*	1,149	39	8941	204	994	182	68.1	0.008050 ± 0.17%	0.05198 ± 0.25%	0.80	0.04684 ± 0.16%	41.0 ± 7.5
2*	2,149	22	9607	196.8	1365	81	64.5	0.008045 ± 0.12%	0.05195 ± 0.18%	0.80	0.04683 ± 0.11%	40.7 ± 5.3
3	1,149	18	6057	126.7	900	64	65.6	0.008149 ± 0.21%	0.05215 ± 0.35%	0.49	0.04642 ± 0.31%	19.2 ± 14.7
4	1,149	18	5787	140.7	773	72	70.2	0.008004 ± 0.20%	0.05119 ± 0.35%	0.51	0.04639 ± 0.31%	17.9 ± 14.7
126 (Fig. 3-7d) - Sillimanite kyanite schist, lower part of cover sequence; located 7.0 km NNE of Pettipiece Pass at 2030 m, 390080E-5702730N (UTM zone 11, BCGS map sheet 82M.048)												
1*	3,149	43	5314	256.8	1706	124	73.0	0.014412 ± 0.12%	0.09584 ± 0.16%	0.85	0.04823 ± 0.09%	110.5 ± 4.2
2*	3,149	65	4552	292.1	1338	225	77.9	0.015665 ± 0.14%	0.10431 ± 0.19%	0.83	0.04829 ± 0.11%	113.6 ± 5.2
3	2,105	10	4803	293.5	1319	38	76.6	0.015728 ± 0.10%	0.10461 ± 0.17%	0.72	0.04824 ± 0.12%	111.0 ± 5.8
4	2,105	14	4368	221.6	1193	47	74.0	0.014534 ± 0.10%	0.09659 ± 0.18%	0.73	0.04820 ± 0.13%	109.1 ± 6.1
5	1,105	6	4632	258.4	1233	23	74.5	0.015670 ± 0.12%	0.10411 ± 0.22%	0.63	0.04818 ± 0.17%	108.3 ± 8.1
6	1,105	7	4311	221.1	1216	23	73.7	0.014959 ± 0.13%	0.09925 ± 0.21%	0.66	0.04812 ± 0.16%	105.2 ± 7.4
7	1,105	9	2912	173.9	1159	21	78.8	0.013950 ± 0.14%	0.09253 ± 0.24%	0.59	0.04811 ± 0.19%	104.5 ± 9.1
8	1,105	8	2385	158.9	411	49	79.3	0.015170 ± 0.17%	0.10058 ± 0.47%	0.63	0.04809 ± 0.39%	103.6 ± 18.4
9	1,105	6	5627	232.3	1171	24	71.6	0.012921 ± 0.12%	0.08555 ± 0.22%	0.57	0.04802 ± 0.18%	100.4 ± 8.4
165 (Fig. 3-7c) - Sillimanite schist, lower part of cover sequence; located 3.8 km NE of Mount Grace at 2200 m, 376140E-5704250N (UTM zone 11, BCGS map sheet 82M.047)												
1	4,74	10	4745	110.2	533	50	65.4	0.008865 ± 0.17%	0.05761 ± 0.42%	0.60	0.04714 ± 0.35%	56.2 ± 16.5
2	2,74	5	6621	153.7	411	48	63.1	0.009455 ± 0.20%	0.06231 ± 0.64%	0.51	0.04780 ± 0.57%	89.3 ± 26.9
3	4,74	10	4904	135.1	1558	20	68.8	0.009495 ± 0.21%	0.06142 ± 0.33%	0.68	0.04692 ± 0.24%	45.0 ± 11.7
4	4,74	8	5575	137.5	1403	19	65.2	0.009477 ± 0.10%	0.06161 ± 0.21%	0.59	0.04715 ± 0.17%	56.8 ± 8.1
5	3,74	7	5733	136.9	2755	8	65.0	0.009241 ± 0.10%	0.06006 ± 0.20%	0.46	0.04714 ± 0.18%	56.4 ± 8.7
6	4,74	15	4263	111.4	2695	14	68.3	0.009159 ± 0.12%	0.05934 ± 0.18%	0.69	0.04699 ± 0.13%	48.8 ± 6.3
166 (Fig. 3-7a) - Sillimanite schist, basement paragneiss; located 7.3 km NE of Mount Grace at 2015 m, 377560E-5707380N (UTM zone 11, BCGS map sheet 82M.057)												
1	5,105	43	6006	120.4	1996	66	64.3	0.007900 ± 0.13%	0.05142 ± 0.17%	0.78	0.04721 ± 0.10%	59.7 ± 4.9
2	3,74	20	4219	75.17	1184	37	59.7	0.007940 ± 0.11%	0.05137 ± 0.19%	0.71	0.04692 ± 0.14%	45.3 ± 6.6
3	5,74	24	6472	133	1541	52	65.1	0.007918 ± 0.12%	0.05124 ± 0.17%	0.80	0.04693 ± 0.11%	45.7 ± 5.0

Table 3-1. (continued).

Analysis ^a	Wt. ^b (μg)	U (ppm)	Pb* ^c (ppm)	²⁰⁶ Pb/ ²⁰⁴ Pb ^d	Pb _c ^e (pg)	²⁰⁸ Pb/ ²⁰⁸ Pb ^f (%)	²⁰⁶ Pb/ ²³⁸ U ^g	²⁰⁷ Pb/ ²³⁵ U ^g	Corr. coef.	²⁰⁷ Pb/ ²⁰⁶ Pb ^g	²⁰⁷ Pb/ ²⁰⁶ Pb ^h age (Ma)	
191 (Fig. 3-7h) - Leucosome layer, ~0.5 m thick, kyanite-bearing, concordant with dominant foliation, intruded into lower part of cover sequence; located ~3 km SSW of Kirbyville Lake at 2010 m, 374260E-5711270N (UTM zone 11, BCGS map sheet 82M.056)												
A	2,105	7	1273	21.5	444	21	7.5	0.016831 ± 0.24%	0.18234 ± 0.64%	0.56	0.07857 ± 0.54%	1161.4 ± 21.6
B	4,105	23	3237	31.34	2345	21	2.7	0.010384 ± 0.12%	0.07982 ± 0.24%	0.64	0.05575 ± 0.18%	442.6 ± 8.2
C	3,74	10	1889	29.05	1576	12	5.6	0.015717 ± 0.75%	0.15962 ± 0.81%	0.92	0.07366 ± 0.32%	1032.1 ± 12.9
D	5,74	20	1624	35.84	5515	8	12.1	0.020718 ± 0.25%	0.25171 ± 0.26%	0.96	0.08811 ± 0.08%	1385.0 ± 2.9
E	6,74	14	1804	25.65	2400	10	5.1	0.014637 ± 0.14%	0.14673 ± 0.21%	0.70	0.07270 ± 0.15%	1005.7 ± 6.1
F	8,74	20	1569	27.86	4031	9	5.8	0.018043 ± 0.10%	0.19307 ± 0.13%	0.78	0.07761 ± 0.08%	1136.9 ± 3.2
G	9,74	26	586.8	12.93	2743	8	7.7	0.021749 ± 0.10%	0.26096 ± 0.13%	0.79	0.08702 ± 0.08%	1361.0 ± 3.1
1	3,105	30	10332	168.5	2207	81	50.5	0.008950 ± 0.15%	0.05588 ± 0.21%	0.76	0.04528 ± 0.13%	-40.4 ± 6.5
2	2,105	12	10212	183.8	1451	50	53.3	0.009304 ± 0.22%	0.05805 ± 0.27%	0.85	0.04525 ± 0.14%	-42.3 ± 7.0
3	3,105	16	9932	177.5	1038	91	53.8	0.009145 ± 0.23%	0.05612 ± 0.49%	0.41	0.04450 ± 0.44%	-82.7 ± 21.8
4	1,105	15	7998	176.4	1710	44	61.4	0.009435 ± 0.12%	0.05904 ± 0.17%	0.79	0.04539 ± 0.10%	-34.8 ± 5.0
5	1,105	8	6207	121.1	1426	22	56.8	0.009332 ± 0.24%	0.05927 ± 0.49%	0.56	0.04606 ± 0.41%	0.9 ± 19.5
214 (Fig. 3-7b) - Kyanite sillimanite schist, upper part of cover sequence; located 7.6 km SE of Mount Grace, on logging road at 1145 m, 378430E-5694540N (UTM zone 11, BCGS map sheet 82M.037)												
1	2,149	40	4590	104.1	1766	54	67.7	0.008089 ± 0.17%	0.05190 ± 0.20%	0.75	0.04653 ± 0.14%	25.2 ± 6.5
2	2,149	30	8603	147.4	1591	84	57.5	0.008048 ± 0.21%	0.05199 ± 0.25%	0.89	0.04685 ± 0.11%	41.7 ± 5.3
3	1,149	16	7407	144	1980	31	61.7	0.008229 ± 0.11%	0.05301 ± 0.16%	0.79	0.04672 ± 0.10%	34.9 ± 4.9
4	1,149	18	7270	137.3	2344	29	59.9	0.008372 ± 0.10%	0.05420 ± 0.14%	0.78	0.04696 ± 0.09%	47.0 ± 4.3
5	1,105	10	6947	128.9	1899	19	58.9	0.008436 ± 0.10%	0.05476 ± 0.16%	0.73	0.04708 ± 0.11%	53.4 ± 5.4
X	1,149	20	9756	72.53	11786	9	2.8	0.008030 ± 0.12%	0.05212 ± 0.13%	0.95	0.04707 ± 0.04%	52.9 ± 2.0
223 (Fig. 3-7f) - Leucosome pod, ~1 m wide, kyanite-bearing, lacks deformation fabrics, intruded into lower part of cover sequence; located 7.8 km SE of Mount Grace, on logging road at 1130 m, 376280E-5693390N (UTM zone 11, BCGS map sheet 82M.037)												
A	5,105	29	2107	16.62	3614	9	0.3	0.008743 ± 0.10%	0.05708 ± 0.14%	0.75	0.04735 ± 0.09%	66.9 ± 4.3
B	2,105	38	1913	15.84	3301	13	0.3	0.009161 ± 0.15%	0.05982 ± 0.18%	0.88	0.04736 ± 0.09%	67.2 ± 4.1
C	6,105	46	1276	10.4	3325	10	0.3	0.009034 ± 0.10%	0.05919 ± 0.14%	0.72	0.04752 ± 0.10%	75.5 ± 4.6
1	2,149	50	17332	322.9	2865	175	56.4	0.009005 ± 0.46%	0.05607 ± 0.47%	0.99	0.04516 ± 0.07%	-47.3 ± 3.2
2	2,149	40	16797	282	2622	151	50.7	0.009168 ± 0.32%	0.05714 ± 0.35%	0.96	0.04521 ± 0.10%	-44.4 ± 5.0
3	1,149	63	6017	183.8	1648	153	69.6	0.010258 ± 0.17%	0.06601 ± 0.38%	0.56	0.04667 ± 0.32%	32.6 ± 15.3
4	1,149	44	5718	174.2	1294	129	69.6	0.010231 ± 0.21%	0.06618 ± 0.26%	0.86	0.04691 ± 0.13%	44.8 ± 6.3
5	1,149	49	5218	184	1091	154	73.8	0.010211 ± 0.12%	0.06556 ± 0.20%	0.77	0.04657 ± 0.14%	27.1 ± 6.5
6	1,149	49	4595	158.2	1289	114	73.3	0.010139 ± 0.13%	0.06504 ± 0.20%	0.78	0.04652 ± 0.12%	24.9 ± 5.9
7	1,149	32	5239	146.4	1137	97	67.4	0.010049 ± 0.11%	0.06506 ± 0.20%	0.75	0.04696 ± 0.13%	47.1 ± 6.4
257 (Fig. 3-7c) - Feldspathic semipelitic schist, upper part of cover sequence; located on logging road at 1190 m, 7.8 km SE of Mount Grace, 378730E-5694490N (UTM grid zone 11, BCGS map sheet 82M.037)												
1	3,149	15	6044	190.6	1275	39	74.7	0.008793 ± 0.11%	0.05712 ± 0.19%	0.73	0.04712 ± 0.13%	55.3 ± 6.4
2	3,149	9	4795	195.6	1543	16	80.2	0.008911 ± 0.11%	0.05790 ± 0.20%	0.59	0.04713 ± 0.16%	55.6 ± 7.9
3	3,149	12	1994	68.02	830	16	76.8	0.008727 ± 0.11%	0.05669 ± 0.25%	0.60	0.04711 ± 0.21%	54.9 ± 9.8
258 (Fig. 3-7e) - Leucosome layer, 0.5 m thick, sillimanite-bearing, concordant with dominant foliation, intruded into lower part of cover sequence; located 8.5 km SE of Mount Grace, on logging road at 1220 m, 379920E-5694710N (UTM zone 11, BCGS map sheet 82M.037)												
A	1,149	30	1564	11.05	720	34	0.0	0.007857 ± 0.15%	0.05075 ± 0.30%	0.65	0.04684 ± 0.23%	41.3 ± 11.0
B	3,149	25	3295	22.88	3386	12	0.3	0.007696 ± 0.15%	0.05003 ± 0.18%	0.79	0.04715 ± 0.11%	56.6 ± 5.3
C	2,149	17	3317	23.17	6923	4	0.0	0.007761 ± 0.16%	0.05028 ± 0.17%	0.82	0.04698 ± 0.10%	48.4 ± 4.7
D	4,149	37	2070	14.52	4169	9	0.1	0.007789 ± 0.13%	0.05053 ± 0.18%	0.77	0.04705 ± 0.11%	51.9 ± 5.3
1	1,149	19	10758	164.3	1152	95	52.0	0.008120 ± 0.13%	0.05137 ± 0.21%	0.79	0.04588 ± 0.13%	-8.5 ± 6.3
2	1,149	48	6885	132.5	1574	112	61.2	0.008265 ± 0.13%	0.05111 ± 0.18%	0.83	0.04485 ± 0.10%	-64.0 ± 5.0
3	1,149	22	6474	118.5	1130	65	59.8	0.008152 ± 0.11%	0.05174 ± 0.20%	0.75	0.04603 ± 0.13%	-0.7 ± 6.4
303 (Fig. 3-7g) - Pegmatite layer, 1 m thick, boudinaged, concordant with gneissosity (schistosity) in host, intruded into lower part of cover sequence; located 7.0 km NNE of Pettipiece Pass at 2180 m, 389460E-5703130N (UTM zone 11, BCGS map sheet 82M.048)												
1	1,149	29	3892	53.8	3770	17	41.2	0.009001 ± 0.17%	0.05841 ± 0.21%	0.87	0.04707 ± 0.10%	52.6 ± 4.9
2	1,149	25	3028	54.55	1051	42	54.7	0.009016 ± 0.24%	0.05837 ± 0.31%	0.81	0.04696 ± 0.18%	46.9 ± 8.6

Table 3-1. (concluded).

	Wt. ^b	U	Pb* ^c	²⁰⁶ Pb ^d	Pb _c ^e	²⁰⁸ Pb ^f	²⁰⁶ Pb ^g	²⁰⁷ Pb ^g	Corr.	²⁰⁷ Pb ^g	²⁰⁷ Pb ^h	
Analysis ^a (μg)	(ppm)	(ppm)	²⁰⁴ Pb	(pg)	(%)		²³⁸ U	²³⁵ U	coef.	²⁰⁶ Pb	²⁰⁶ Pb age	
											(Ma)	
334 (Fig. 3-7c) - Sillimanite kyanite schist, lower part of cover sequence; located 7.8 km SE of Mount Grace, just N of logging road at 1140 m, 376350E-5693420N (UTM zone 11, BCGS map sheet 82M.037)												
1	1,149	50	7280	160.7	2383	94	61.0	0.009527 ± 0.39%	0.06120 ± 0.40%	0.98	0.04659 ± 0.09%	28.2 ± 4.1
2	1,149	36	6640	147.4	2656	52	62.6	0.009168 ± 0.25%	0.05912 ± 0.26%	0.96	0.04677 ± 0.07%	37.4 ± 3.4
3	1,149	24	6563	131.7	3885	23	60.2	0.008833 ± 0.10%	0.05699 ± 0.14%	0.81	0.04679 ± 0.08%	38.4 ± 3.9
4	1,149	25	7060	147.6	3423	29	61.6	0.008875 ± 0.11%	0.05718 ± 0.13%	0.89	0.04673 ± 0.06%	35.3 ± 2.9
5	1,149	25	5339	129.9	2219	36	65.9	0.009172 ± 0.10%	0.05901 ± 0.14%	0.81	0.04666 ± 0.08%	32.0 ± 4.0

Isotopic composition of Carleton University lab blank (uncertainty is one standard deviation) - 206:207:208:204 = 19.01 ± 0.36 : 15.64 ± 0.20 : 38.23 ± 0.74 : 1

^aIn first column, A-G = fraction codes for zircon analyses, 1-9 = fraction codes for monazite analyses, X = fraction code for xenotime analysis, * = chemical analysis done at the Geological Survey of Canada in Ottawa (all other analysis was done at Carleton University); in second column, 1-10 = number of grains analyzed, 74-149 = average size in μm prior to abrasion with pyrite (Krogh 1982).

^bWeighing uncertainty is 1 μg.

^cRadiogenic Pb.

^dMeasured ratio, corrected for spike and Pb fractionation of 0.09% ± 0.03%/AMU.

^eTotal common Pb in analysis, corrected for spike and fractionation.

^fRadiogenic ²⁰⁸Pb, expressed as % of total radiogenic Pb.

^gCorrected for blank Pb and U and common Pb (Stacey-Kramers model Pb composition equivalent to the interpreted age of the analyses); errors are one standard deviation in %.

^hErrors are two standard deviations in Ma.

Table 3-2. Summary of metamorphic monazite and xenotime U-Pb ages, arranged according to structural level.

Sample (number of analyses)	Rock type	Mineral	$^{207}\text{Pb}/^{235}\text{U}$ Ages (Ma)
<i>Upper Level</i>			
180 (n=4)	leucogranite	monazite	63.5-59.6
334 (n=5)	schist	monazite	60.8-56.2
<i>Middle Level</i>			
126 (n=9)	schist	monazite	118 +16/-8 ¹ 37 +23/-30 ²
303 (n=2)	pegmatite	monazite	57.9-57.3
341 (n=3)	pegmatite	xenotime	58.2-55.0
(n=1)		monazite	51.0-50.4
214 (n=5)	schist	monazite	54.3-51.2
(n=1)		xenotime	51.7-51.5
257 (n=3)	schist	monazite	57.4-55.5
165 (n=5)	schist	monazite	60.9-56.4
101 (n=2)	schist	monazite	53.1-52.2
25 (n=3)	schist	monazite	51.0-50.1
<i>Lower Level</i>			
24 (n=4)	schist	monazite	52.4-51.6
166 (n=3)	schist	monazite	51.1-50.5
112 (n=4)	schist	monazite	51.9-50.3
70 (n=3)	schist	monazite	50.1-49.3
85 (n=2)	schist	monazite	51.6-50.2
70 & 85 (n=4)	schist	monazite	48 ± 7 ²

¹Upper and lower intercept age

²Lower intercept age

Table 3-3. Summary of upper and lower intercept U-Pb ages and amounts of Pb loss from accessory minerals in Proterozoic rocks, arranged according to structural level.*

Sample	Rock type	Analyses (grains in each, m=multigrain)	Upper & lower intercept ages (Ma) of discordia lines	Amount of discordance
<i>Upper Level</i>				
180	leucogranite	5 zircon (1-9)	1762 ± 6 & 76 ± 20	24-38%
187 Kirbyville gneiss	granodioritic orthogneiss	4 zircon (1-3) & 3 titanite (m)	1862 ± 1 & 63 ± 14	7-14% 94-96%
207 Mount Grace gneiss	syenitic orthogneiss	5 zircon (1-7)	724 ± 5 & 53.4 ± 2.9	55-85%
<i>Middle Level</i>				
341	pegmatite	5 zircon (1)	1852 ± 4 & 149 ± 20	19-46%
<i>Lower Level</i>				
Bourne granite PCA-305B-83 324A	granite	6 monazite (1-m), 2 zircon (m) & 3 titanite (m)	1848 ± 3 & 95 ± 35	9-18% 15-16% 52-80%
308	pegmatite	4 zircon (2-3)	1845 ± 3 & 84 ± 19	13-31%
84	pegmatite	5 monazite (1-2) & 3 zircon (1-6)	1836 ± 2 & 76 ± 15	6-17% 19-34%
91	pegmatite	3 monazite (1-2) 3 zircon (1-5) & 1 monazite (1)	1834 ± 2 & 85 ± 17 1844 ± 2 & 30 ± 20	12-22% 5-22%
70 & 85	pelitic schist	4 monazite (1-3)	2057 ± 13 & 48 ± 7	10-88%
337	amphibolite	3 titanite (m)	$1780 +247/-216$ & $29 +33/-39$ (unforced) 38 ± 20 (with forced 1.85 Ga upper intercept)	89-98%
79 & OP-90-12	amphibolitic gneiss	6 titanite (m)	$1859 +54/-44$ & $71 +174/-180$	25-43%

*Data from Chapters 1 and 2, except for OP-90-12 (Parrish, 1995) and most of the data from PCA-305B-83 (Armstrong et al. 1991).

Figure 3-1. Tectonic assemblage map in the vicinity of the Monashee complex, modified after Wheeler and McFeely (1991), showing the location of Fig. 3-3. The dark grey square in inset locates the map with respect to the morphogeologic belts of the Canadian Cordillera.

Fig. 3-2. Crustal cross section through the southern Omineca Belt, taken from Johnson and Brown (1996). The section crosses the Monashee complex through the northern part of Thor-Odin dome (see Fig. 2 in Johnson and Brown (1996) for exact location).

Figure 3-3. Geologic map of northern Frenchman Cap dome, modified after Wheeler (1965), McMillan (1973), Psutka (1978), Höy and Brown (1980), Journeay (1986), and Höy (1987), showing localities of U-Pb samples (sample numbers are shown in parentheses) discussed in text. The number placed before the mineral name represents the number of U-Pb analyses upon which the age is based. ky = kyanite, si = sillimanite. A-A' locates the cross sections shown in Figs. 3-4 and 3-6.

Fig. 3-5. P-T petrogenetic grid (system $K_2O-FeO-MgO-Al_2O_3-SiO_2-H_2O$) from Spear and Cheney (1989). Mineral abbreviations are almandine (alm), aluminosilicate (as), biotite (bt), cordierite (cd), chlorite (ch), garnet (gt), K-feldspar (Kf), quartz (qz), and staurolite (st).

Figure 3-6. Cross section of northern Frenchman Cap dome, modified after Journeay (1986), showing approximate structural position of the samples with U-Pb ages (in Ma) and Pb loss data. Location of section and patterns for rock units are shown in Fig. 3-3.

Figure 3-7. U-Pb concordia plots. Ellipses for the analyses represent the two sigma uncertainty (95% confidence level) and analyses that are plotted with + symbols have ellipses that are too small to be seen at this scale. Zircon analyses are denoted with letters, monazite analyses are denoted with numbers, and xenotime analyses are denoted with X1-X3. See Table 3-1 and Chapters 1 and 2 for analytical data.

Figure 3-8. Monazite, xenotime, and zircon U-Pb ages from pelitic schist and intrusive rocks plotted against relative structural depth.

Figure 3-9. U-Pb concordia plots showing titanite and monazite analyses from Proterozoic rocks. Ellipses for the analyses represent the two sigma uncertainty (95% confidence level) and analyses that are plotted with + symbols have ellipses that are too small to be seen at this scale. See Chapters 1 and 2 for analytical data.

Figure 3-10. Plots showing metamorphic temperature versus duration, as deduced by application of diffusion parameters (references are given) to diffusion equations (see text) given the maximum amount of Pb loss recorded by titanite and monazite from different structural levels. For titanite, the 31%, 65%, and 96% Pb loss are from samples 79, 324A, and 187, respectively. For monazite, the 22% Pb loss is from rocks that intruded into basement and a basement pelitic schist and the 56% Pb loss is from sample 126. Peak temperatures are thought to have been ~ 700 - 750 degrees C based on comparison of the pelitic mineral assemblages with a petrogenetic P-T grid.

CHAPTER 4

U-PB GEOCHRONOLOGIC CONSTRAINTS ON THE DOWNWARD YOUNGING AND LESSENING IN INTENSITY OF CORDILLERAN TECTONISM IN THE MONASHEE COMPLEX, BRITISH COLUMBIA, WITH IMPLICATIONS FOR INVERTED METAMORPHISM

ABSTRACT

The Monashee complex lies beneath a crustal-scale ductile thrust fault, the Monashee décollement, in a deep structural exposure in the southern Canadian Cordillera. The presence of an inverted metamorphic sequence is denoted by sillimanite - K-feldspar assemblages in the structurally highest rocks and staurolite-kyanite-muscovite assemblages in deeper rocks, and structural data indicate that the dominant deformation fabric formed during this metamorphism. U-Pb isotopic data and field relationships from the northern part of the complex show that the age and intensity of Cordilleran metamorphism and deformation vary with structural level. Monazite U-Pb dates that young downward from ~64 Ma in the highest level to ~49 Ma in the deepest level, over a structural thickness of 6 km, are interpreted as representing diachronous peak metamorphism. Pb loss data from Proterozoic and Cretaceous monazite and titanite indicate that the metamorphic intensity (a factor dependent on temperature and duration) decreases downward. Modeling of Pb loss indicates that deeper levels were at elevated temperatures (~700°C, as suggested by

pelitic mineral assemblages and a petrogenetic grid) for less than 1 Ma and that higher levels were at equivalent or moderately greater temperatures ($\sim 50^{\circ}\text{C}$ hotter) for up to a few million years longer. The thermal peak in higher levels was (i) synchronous with thrusting on the Monashee décollement, (ii) immediately followed by a period of rapid cooling that occurred during initial motion on overlying normal faults, which are associated with exhumation of the complex, and (iii) attained while deeper levels were at lower temperatures but presumably heating toward a thermal peak that occurred a few million years later during the main period of extension on overlying faults. Dating of intrusive rocks with various relationships to the dominant Cordilleran deformation shows that deformation was synmetamorphic and lessens in intensity (a factor related to the amount of transposition) downward within basement rocks in deeper levels, locally having no effects. The U-Pb data that suggest metamorphism was less intense with increasing structural depth agree with the mineral assemblage data by showing an inverted metamorphic sequence. Evidence for the associated thermal gradient being inverted was possibly revealed in deeper levels, where the pattern of downward decreasing metamorphic intensity exists in basement rocks that are known to have been only weakly strained since the Proterozoic. In higher levels, it is uncertain whether the thermal gradient was inverted because there are numerous east-directed structures capable of significant synmetamorphic displacements, such as kilometre-scale isoclinal folds with sheared limbs, that could have telescoped a once laterally extensive normal metamorphic sequence into an inverted sequence without ever placing warmer rocks over cooler ones. The resulting tectonic model contends that the inverted metamorphic sequence formed

mainly by the juxtaposition of rocks with widely varying initial positions that reached peak temperatures at different times and in different positions within the orogen, and perhaps there was a minor component of local downward heat conduction.

INTRODUCTION

The Monashee complex in the core zone of the southern Canadian Cordillera is composed of rocks exposed in the tectonic window through the Monashee décollement (Figs. 4-1, 4-2; Read and Brown 1981), a crustal-scale ductile thrust fault that has been correlated with the sole thrust of the Rocky Mountain Foreland Belt (Brown et al. 1992; Cook et al. 1992). Tectonism in the complex during Cordilleran orogenesis generated middle and upper amphibolite facies metamorphic assemblages, kilometre-scale isoclinal folds, and penetrative planar and linear deformation fabrics (Wheeler 1965; Reesor and Moore 1971; McMillan 1973; Psutka 1978; Read 1980; Höy and Brown 1980; Journeay 1986; Scammell 1986; Höy 1987; McNicoll and Brown 1995). Knowledge of the age and nature of these features at this structural level, perhaps the deepest exposed in the southern Canadian Cordillera, has significant bearing on tectonic models of Cordillera orogenesis. Parrish (1995) used U-Pb isotopic data to propose a general age of peak metamorphism in the Monashee complex and to elucidate the thermal history of the Selkirk allochthon above the complex. The model based on these data contends that in the Mesozoic the complex was at low temperatures and was located east (on the foreland side) of the metamorphic front, not attaining high temperatures until the hotter allochthon

was emplaced onto it in the Late Cretaceous - early Tertiary. Parrish considered the thermal gradient in the complex to have been inverted because (i) the base of the allochthon was at higher temperatures than the complex during the metamorphism (based on mapping of metamorphic minerals in pelitic schist (Brown 1980; Journeay 1986; Scammell 1986) and U-Pb dating (Parrish 1995 and references therein)), (ii) the event was short-lived (<5 Ma in duration based on Parrish's modeling of Pb loss in Proterozoic titanite), and (iii) the metamorphic sequence in the complex is inverted (mapping of metamorphic minerals in pelitic schist (Psutka 1978; Journeay 1986) shows that the highest grade assemblage exists in the structurally highest rocks (see Chapter 3 for summary of the assemblage data)).

There are three main purposes of this chapter. The first is to show that the age and intensity of deformation associated with Cordilleran orogenesis vary with structural level in the northern part of the Monashee complex. This is done by presenting U-Pb isotopic data and field relationships from samples that were collected from intrusive rocks. The second purpose is to discuss the tectonic implications of the U-Pb data and interpretations concerning Cordilleran metamorphism that were presented in Chapter 3. The third purpose is to integrate the deformation and metamorphic interpretations into a new tectonic model for the northern Monashee complex. This model invokes Paleocene - Middle Eocene, short-lived metamorphism and deformation that young and lessen in intensity with increasing structural depth. A central part of the model is an inverted metamorphic sequence that is indicated by U-Pb isotopic and mineral assemblage data. Consideration of the deformation history leads to the conclusion that a normal

metamorphic sequence could have been telescoped during deformation without the thermal gradient in the complex ever being inverted. However, there is minor evidence for inversion of the thermal gradient in the structurally deepest rocks.

Inverted metamorphic sequences have been documented in many mountain belts (e.g., LeFort 1975; Spear et al. 1995). However, examples of inverted thermal gradients, where warmer rocks were above cooler rocks at some point in the metamorphic history, are rare because (i) proving their existence requires showing that the rocks attained peak conditions in approximately their present relative positions and (ii) inverted gradients are transient features, leaving little evidence for their existence in the absence of rapid cooling of the entire sequence. Proving an inverted gradient is a relatively easy task in rocks that were weakly deformed since the onset of metamorphism and cooled rapidly (e.g., contact metamorphism in the aureole of a high level postdeformational pluton). However, in deeply buried rocks that were strongly deformed since the onset of metamorphism, such as most rocks in the Monashee complex, it is difficult to prove that higher level rocks were warmer than underlying rocks during the juxtaposition. Two recent papers depict how an inverted metamorphic sequence can form in the absence of an inverted thermal gradient if there were substantial syn- to postmetamorphic displacements within the sequence.

Jamieson et al. (1996) showed that an inverted sequence can form by tectonic juxtaposition of rocks that had widely varying initial positions and reached peak temperatures at different times and in different positions within the orogen. Hubbard (1996) showed that an inverted sequence can form by ductile shearing of a normal thermal gradient. These models strongly suggest that displacements can telescope a normal

metamorphic sequence that was once laterally extensive into a stacked inverted sequence without ever placing warmer rocks over cooler rocks. The applicability of such models to the Monashee complex are considered at the end of the chapter.

TECTONIC SETTING

The Monashee complex lies in the southern Omineca Belt, the metamorphic and plutonic hinterland to the Rocky Mountain Foreland Belt of the Canadian Cordillera that developed subsequent to collision between accreted terranes and the North American craton (Monger et al. 1982). The complex contains one of four known exposures of Paleoproterozoic (~2.2-1.8 Ga) rocks in the Canadian Cordillera (Parrish 1991 and references therein; Chapter 2). The older of these rocks compose basement that forms the core of two structural culminations in the complex (Fig. 4-1), namely, Frenchman Cap dome in the north and Thor-Odin dome in the south. Basement is unconformably overlain by a Paleoproterozoic-Paleozoic cover sequence, a platformal succession of metasedimentary and subordinate metaigneous rocks that is 2-3 km thick and laterally extensive (Wheeler 1965; Journeay 1986; Scammell and Brown 1990; Chapter 1).

Previous workers (Read and Brown 1981; Journeay 1986; Scammell 1986; McNicoll and Brown 1995) mapped a ductile fault zone, the Monashee décollement, between the Monashee complex and the overlying Selkirk allochthon (Figs. 4-1, 4-2). A faulted contact was interpreted because kilometre-scale folds in the complex are truncated at this up to 1 km thick zone and because of the absence of rocks that typify the complex above the zone (i.e., there is a lithologic break at the fault; Figs. 4-1 to 4-3). The fault is

considered to be a thrust because kinematic indicators suggest top-to-the-east motion on its west dipping surface and because the allochthon appears to have originated at deeper levels than the complex (the allochthon contains a higher grade pelitic mineral assemblage and a higher proportion of leucogranitic rocks). Carr (1992) used U-Pb dating of leucogranitic rocks to show that the décollement was active in the Paleocene on the south side of Thor-Odin dome. Surface geology and Lithoprobe seismic reflection profiles were used to correlate the décollement with the Late Cretaceous - Paleocene basal thrust beneath the Rocky Mountain Foreland Belt, and thus it is considered to be a structure that links middle crustal strain in the hinterland of the orogen with upper crustal strain in the foreland (Brown et al. 1992; Cook et al. 1992 and references therein). Palinspastic restoration of the Selkirk allochthon indicates up to ~300 km of northeastward displacement relative to the North American craton (Brown et al. 1993), with more than 80 km of that occurring on the décollement (Read and Brown 1981; Brown et al. 1986).

The Selkirk allochthon has a protracted, multiple-event tectonic history spanning >100 Ma, from Early Jurassic in the highest structural level to early Tertiary in the lowest (Parrish 1995 and references therein). This pattern was considered by Parrish to be consistent with the thermal evolution of rocks presently in the hinterland of a fold and thrust belt that were progressively buried under the belt as it propagated toward the foreland. The long history of the allochthon contrasts with the relatively simple Cordilleran history of the Monashee complex, where only one thermal event (early Tertiary) is recorded and Paleozoic-Cretaceous plutons are unknown. Parrish used these characteristics to contend that the complex was part of the craton and overlying foreland

basin in the Mesozoic at the time the allochthon was tectonized to the west, not attaining high temperature and deep burial until the allochthon's hot base was thrust eastward onto the complex during early Tertiary motion on the décollement.

Compressional tectonism in the southern Omineca Belt was followed shortly thereafter by extension of the orogen (Parrish et al. 1988 and references therein). The east margin of the Monashee complex is delineated by the Columbia River fault (Figs. 4-1, 4-2; Lane 1984 and references therein), an Eocene brittle-ductile normal fault with an estimated >15 km of displacement (Parrish et al. 1988). This fault and a coeval normal fault to the west, the Okanagan - Eagle River system, bound the Shuswap metamorphic complex. Motion on these faults resulted in rapid exhumation of the footwall (Brown and Journeay 1987; Parrish et al. 1988).

METAMORPHISM

Basement in Frenchman Cap dome was affected by a ~2.06 Ga metamorphic event that predated deposition of the cover sequence (Chapter 2). Basement and the cover sequence underwent a younger metamorphism during Cordilleran orogenesis (McMillan 1973; Psutka 1978; Brown 1980; Journeay 1986; Scammell 1986; Höy 1987). Mapping of metamorphic minerals in pelitic schist in the cover sequence shows that the metamorphic sequence is inverted; Fig. 4-4 shows that the highest grade rocks (containing a sillimanite - K-feldspar $\pm$ melt assemblage) exist at the highest structural level, above and immediately beneath the Monashee décollement, and the lowest grade rocks (containing a kyanite-staurolite-muscovite assemblage) exist at the farthest distance from the exposed

trace of the décollement. The assemblage in the basement is sillimanite-muscovite $\pm$ melt, which is identical to the assemblage in the proximal rocks in the cover sequence. The similarity in the assemblages and the lack of evidence for postmetamorphic displacement on the unconformity suggest that the upper part of the basement reached the same temperatures as the base of the cover sequence. Because it is impossible to determine whether sillimanite in deeper parts of the basement grew during Proterozoic or Cordilleran events, estimates of peak metamorphic conditions cannot be made.

Journey (1986) used petrologic evidence to interpret the metamorphic sequence in Frenchman Cap dome as representing two events. A normal thermal gradient was proposed for an intermediate pressure event during which kyanite, staurolite, and sillimanite grew, and an inverted gradient was proposed for a low pressure overprint during which andalusite grew and the sillimanite - K-feldspar assemblages formed. A Middle Jurassic - Late Cretaceous timing for these events was proposed based on numerous arguments, none of which involved geochronologic data from the dome. Parrish (1995) used a moderate amount of Pb loss in titanite from a Proterozoic amphibolitic basement gneiss to postulate that the Cordilleran overprint was short-lived (<5 Ma in duration). He inferred that metamorphism throughout the dome (i) was of similar duration because there is no structural discontinuity at the unconformity and (ii) was associated with an inverted thermal gradient, with the heat source being the base of the Selkirk allochthon as it was thrust onto the complex. This latter conclusion, which contrasts with Journey's (1986) interpretation that only a low pressure overprint was thermally inverted, is based on the short duration of the event and the U-Pb data and

mineral assemblage data that indicate the base of the allochthon was at higher temperatures than the complex during the event.

Uranium-lead dating in the complex has recently been reviewed (Parrish 1995; Carr 1995), so only a brief summary will be given here. Zircon, monazite, and titanite dates of 66-49 Ma from various rock types in high structural levels of Frenchman Cap and Thor-Odin domes were used as evidence that significant metamorphism occurred at that time (Coleman 1990; Parkinson 1992; Carr 1995; Parrish 1995). Basement was considered to have been little affected by Cordilleran metamorphism because there are no isotopic data indicating post-Proterozoic magmatism, metamorphism, or migmatization (Armstrong et al. 1991). Hornblende and biotite $^{40}\text{Ar}/^{39}\text{Ar}$ dates indicate that the cover sequence cooled rapidly from elevated temperatures in the early Tertiary (Sanborn 1996).

In Chapter 3, U-Pb data were interpreted as indicating that the age and intensity of metamorphism vary with structural level. Knowledge of the intensity of the event was considered crucial to the interpretation of the thermal significance of concordant and reversely discordant metamorphic monazite and xenotime U-Pb isotopic dates that young downward from ~64 Ma in the highest level to ~49 Ma in the deepest level (Figs. 4-5, 4-7a). Such knowledge was gained by considering discordant linear arrays formed by monazite and titanite U-Pb analyses. Upper intercept ages are Proterozoic and Cretaceous, and lower intercept ages suggest that the discordance resulted from an early Tertiary overprint. The discordance was mostly attributed to Pb loss rather than overgrowth because the samples with the discordant linear arrays generally lack U-Pb evidence for mineral growth at the lower intercept ages. Quantitative estimates of the

metamorphic intensity were made by modeling Pb diffusion in episodes of heating of fixed duration and temperature, specific combinations of which were calculated by solving equations for measured fractional Pb loss and assumed diffusion radius and geometry. The results of this modeling are discussed below. The knowledge that Pb loss in monazite that existed throughout the event was incomplete led to the interpretation that the downward younging dates of 64-49 Ma from monazite and xenotime closely reflect the ages of U-Pb accessory mineral growth. These ages were taken to closely represent the thermal peak based on (i) the calculated short durations of elevated temperatures, (ii) previously obtained $^{40}\text{Ar}/^{39}\text{Ar}$ dates that indicate rapid cooling occurred immediately after the peak, (iii) the widespread and consistent nature of the U-Pb dates, and (iv) U-Pb ages that indicate kyanite- and sillimanite-bearing leucosomes were generated at this time. Major conclusions made from these interpretations are that peak metamorphism in the Monashee complex was diachronous according to structural level (Figs. 4-5, 4-7a) and metamorphic intensity increased upward.

DEFORMATION

Frenchman Cap basement was affected by a ~2.06 Ga deformation event that predated deposition of the cover sequence (Chapter 2). The dominant gneissosity in some basement outcrops is known, based on data presented in Chapter 2, to have developed during this event, but in most outcrops this gneissosity was overprinted by younger deformation during Cordilleran orogenesis. The following section is a summary of

mapping by Wheeler (1965), McMillan (1973), Psutka (1978), Höy and Brown (1980), Journeay (1986), Scammell (1986), and Höy (1987).

The largest folds associated with the Cordilleran deformation are east-verging, kilometre-scale, recumbent isoclines on the west side of the dome that were interpreted as having begun to form prior to peak metamorphism. These structures, herein termed the Frenchman Cap isoclines, include (from lowest to highest structural level) the Mount Grace syncline, the Kirbyville anticline (Figs. 4-3, 4-4), the Sibley Creek syncline, and the overlying anticlinal mate to the Sibley Creek syncline. Only the deepest two isoclines are exposed in the study area, and the upper two are exposed on the north side of the dome. Intensely sheared zones along the limbs of the isoclines were mapped as the Ratchford Creek and Anstey Range faults (Figs. 4-3, 4-4). Basement lies in the core of the Kirbyville anticline, but basement in the centre of the dome is interpreted as having been unaffected by kilometre-scale isoclinal folds on the basis that the basement-cover unconformity is a simple domed surface that was only affected by upright postmetamorphic folds. The Frenchman Cap isoclines were previously interpreted as having developed prior to the synmetamorphic motion on the Monashee décollement because the isoclines are apparently truncated at the fault surface. Synmetamorphic folds that were considered as postdating the Frenchman Cap isoclines are the most prevalent outcrop-scale structures in the dome and are associated with development of the dominant deformation fabrics, here termed "Cordilleran fabrics". Cordilleran fabrics include (i) a foliation (gneissosity in metaigneous rocks and schistosity in pelitic and semipelitic schists) that is axial planar to the synmetamorphic folds and defines the domal shape of the Frenchman Cap culmination

and (ii) mineral and stretching lineations that have a down-dip orientation within the foliation, coaxial with the hinge lines of the synmetamorphic folds. Synmetamorphic folds had negligible effects on the map-scale distribution of lithologic contacts in the study area. Folds that deformed the Cordilleran fabrics were mapped as postmetamorphic structures. These generally northeast-verging folds with subhorizontal hinge lines and upright to moderately inclined axial surfaces (Fig. 4-4) are best developed on the east side of the dome and are found locally in basement in the central part of the dome.

There were few previous U-Pb geochronologic age constraints on deformation in the Monashee complex. Thrusting on the Monashee décollement in southern Thor-Odin dome affected an intrusive rock dated at ~59 Ma and was postdated by a pegmatite dyke dated at ~58 Ma (Carr 1992). Four postdeformational pegmatite dykes from immediately above and below the décollement in the cover sequence of Frenchman Cap dome were dated at ~58-55 Ma (Scammell 1993; Parrish 1995). Deformation that resulted in variably strained 53-50 Ma pegmatite layers in Thor-Odin dome is inferred to be due to extension on the Columbia River fault (Parkinson 1992).

U-Pb geochronology

In the study of strongly tectonized rocks, it is difficult to determine when significant ductile deformation began and how the intensity of the strain varies spatially and temporally. The timing of early strain is typically difficult to ascertain because precise igneous crystallization ages must be obtained from pre- and syndeformational rocks; such structural interpretations are usually complex and interpretations based on U-Pb

geochronology are tenuous because the accessory minerals are subject to Pb loss during peak metamorphism. These problems can potentially be solved for Cordilleran deformation in Frenchman Cap dome for two reasons. First, the dome contains the lowest grade rocks in the Monashee complex, which are known to have been subjected to conditions that were below the nominal Pb closure temperatures in monazite and titanite because such mineral that grew in the Proterozoic suffered incomplete Pb loss during Cordilleran metamorphism (Chapter 3). This means that igneous accessory minerals grown before and during peak metamorphism yield U-Pb dates that closely reflect crystallization ages. Second, the dome contains numerous dated structural markers that were undeformed prior to the early Tertiary, such as Proterozoic gneissosity and granitic rocks (Chapters 1, 2), and it contains pegmatite and leucosome that crystallized throughout the Cordilleran deformation (Chapter 3; this chapter). These data allow for insights into the age and intensity of deformation that have not been obtained elsewhere in the Omineca Belt.

The tectonic significance of fabrics and correlation of structures across the study area were determined through U-Pb dating of 17 intrusive rocks (Fig. 4-3) and documentation of their relationships to the structures and fabrics. Nine samples are from Proterozoic rocks that were undeformed prior to the early Tertiary, and their U-Pb data were presented in Chapters 1 and 2. U-Pb data from five of the other samples, which intruded in the early Tertiary, are presented in this chapter, and data from three others were given in Chapter 3. In order to simplify data presentation, the study area is divided into three levels according to structural distance from the Monashee décollement, which

defines the west margin of the dome and is interpreted as being arched over it (Journey 1986). The Anstey Range fault (Figs. 4-3, 4-4; Journey 1986) is the boundary between upper and middle levels, and the basement-cover unconformity is the boundary between middle and lower levels. It is emphasized that boundaries between the levels do not have tectonic age significance because there is an age overlap across them. The structural thickness from the décollement to the deepest basement exposure is 6 km. U-Pb data and sample locations are given in Table 4-1, concordia diagrams are depicted in Fig. 4-6, ages are shown on the map in Fig. 4-3 and on the cross section in Fig. 4-5, and ages are summarized in Fig. 4-7c and Table 4-2.

Analytical methods

Uranium-lead geochronology followed procedures outlined in Chapter 2.

Early Tertiary intrusive rocks

Upper level

Sample 191 is from an aluminous quartz-rich leucosome layer ~0.5 m thick that intruded into the lower part of the cover sequence, in the lower limb of the Kirbyville anticline north of Blais Creek (Fig. 4-3). Its setting and U-Pb data are presented in Chapter 3. The layer appears to have been affected by at least some of the synmetamorphic ductile deformation that affected the host because (i) kyanite is strongly lineated in an orientation that is coaxial with the Cordilleran mineral and stretching

lineations and (ii) the layer is concordant with the Cordilleran foliation. Seven zircon analyses and five monazite analyses are interpreted as indicating an igneous melt or aluminous fluid crystallization age of 57.8 ± 1.5 Ma.

Sample 223 is from an aluminous quartz-rich leucosome pod ~1 m wide that intruded into the lower part of the cover sequence, in the Kirbyville anticline south of Mount Grace (Fig. 4-3). Its setting and U-Pb data are presented in Chapter 3. It differs from sample 191 in that it does not contain ductile deformation fabrics, and thus it is thought to have intruded after the synmetamorphic deformation that affected the host schist. Three zircon analyses and two monazite analyses are interpreted as indicating an igneous melt or aluminous fluid crystallization age of 57.5 ± 1.5 Ma.

Sample 263 is from a pegmatite stock ~0.5 km wide that intruded into amphibolitic gneiss and schist in the lower part of the cover sequence, in the lower limb of the Kirbyville anticline south of Mount Grace (Fig. 4-3). It is the largest postdeformational intrusion seen in the study area. Intrusion is interpreted as having postdated the synmetamorphic deformation that affected the host because the stock lacks ductile deformation fabrics and its contacts sharply truncate the Cordilleran foliation in the host (Fig. 4-8). Four zircon analyses (three to seven grains each; Table 4-1) lie slightly below concordia between 57.0 and 54.5 Ma (Fig. 4-6a), with the three youngest analyses just overlapping concordia. The minor discordance is attributable to any combination of the following: (i) minor inheritance; (ii) inherent slight deficiency of ^{206}Pb due to ^{230}Th fractionation during crystallization (Schärer 1984; Parrish 1990), which would decrease the $^{206}\text{Pb}/^{238}\text{U}$ date by a maximum of 0.1 Ma; or (iii) inherent slight excess of ^{231}Pa

(Schärer 1984), thereby increasing the $^{207}\text{Pb}/^{235}\text{U}$ date. Given that it is difficult to differentiate between these possibilities, a precise age for this rock is not known.

However, the cluster of slightly discordant analyses between 56.5 and 54.5 Ma suggests an igneous crystallization age of 55.5 ± 1.0 Ma.

Middle level

Sample 12 is from 0.5-1.0 m thick pegmatite layers that intruded into the lowest calcareous schist in the cover sequence, south of Pettipiece Pass (Fig. 4-3). Most layers are concordant with the Cordilleran foliation in the host and others are slightly discordant (Fig. 4-9). All layers are boudinaged, contain quartz and feldspar that were flattened into parallelism with the contacts of the layer, and were deformed by postmetamorphic folds. Sample 12 was dated because these relationships indicate that deformation responsible for the regional foliation and postmetamorphic folds occurred, at least in part, after intrusion. The amount of deformation that took place prior to intrusion is unknown. Zircon was dated from two pegmatite layers lying ~1 m apart. Three zircon analyses from each layer (three to ten grains each; Table 4-1) lie slightly below concordia between 52 and 50 Ma (Fig. 4-6b). On the concordia diagram, analyses from each layer plot in groups that span ~0.5 Ma (in $^{207}\text{Pb}/^{235}\text{U}$ ages) and there is a gap of ~0.5 Ma between the groups. The minor discordance is attributable to the same possible reasons given for sample 263. The igneous crystallization age is interpreted as corresponding with the part of concordia directly adjacent to the analyses, which is 51.0 ± 1.0 Ma.

Sample 258 is from a 0.5 m thick leucosome layer that intruded into the upper part of the cover sequence, in the core of the Mount Grace syncline south of Mount Grace (Fig. 4-3). Its setting and U-Pb data are presented in Chapter 3. It is one of numerous layers that lie concordant with the Cordilleran foliation in the host (Fig. 4-10) and contain quartz and feldspar that were flattened into parallelism with the margins of the layer. These structural relationships suggest that deformation responsible for the Cordilleran foliation occurred, at least in part, after intrusion. The amount of deformation that took place prior to crystallization is unknown. Four zircon analyses and three monazite analyses are interpreted as indicating an igneous crystallization age of 50.5 ± 1.0 Ma.

Lower level

Sample 57 is from a 0.7 m thick pegmatite dyke that intruded into basement orthogneiss, south of Pettipiece Pass (Fig. 4-3). Dyke intrusion is interpreted as having postdated the synmetamorphic deformation that affected the host because the contacts are straight and highly discordant to the Cordilleran gneissosity in the host, and the dyke lacks deformation fabrics. Seven zircon analyses (one to nine grains each; Table 4-1) lie slightly below concordia between 52 and 49 Ma (Fig. 4-6b). The analyses are disposed such that those with the youngest ages lie closest to concordia and the oldest analyses lie farthest away. Because the analyses probably contain slight inheritance and apparently define a discordia with a lower intercept ~ 49.5 Ma, the interpreted igneous crystallization age is 49.5 ± 0.5 Ma.

Sample 60A is from a 0.5 m thick pegmatite layer that intruded into basement orthogneiss and Proterozoic granite (Bourne granite, see below), south of Pettipiece Pass (Fig. 4-3). The pegmatite is deformed by folds that are interpreted as being responsible for the partial to complete transposition of a Proterozoic gneissosity (age based on cross-cutting relationships of nearby Bourne granite and amphibolite (Chapter 2)) into the Cordilleran foliation (Fig. 4-11d). These folds are correlated with the synmetamorphic Cordilleran folds mapped by Journeay (1986) because both sets have axial surfaces that parallel the Cordilleran foliation and hinge lines that are coaxial with the Cordilleran stretching and mineral lineations. The collection site was on the limb of an isoclinal fold, where strongly flattened quartz and feldspar in the pegmatite layer define a foliation that is parallel and continuous with the gneissosity in the host and slightly discordant to the contacts of the layer (Figs. 4-11a,b,c). A stretching lineation in the layer is parallel with the Cordilleran mineral and stretching lineations (Fig. 4-11e). Sample 60A was dated because these relationships indicate that deformation responsible for the Cordilleran foliation and lineation occurred, at least in part, after intrusion. Where the pegmatite layer truncates a layer of Bourne granite, the contacts of the granite are more discordant to the gneissosity in the host than those of the pegmatite (Figs. 4-11a,b), and more importantly, the granite appears to only have undergone the same deformation events as the pegmatite (i.e., the only fabrics contained in the two rocks are the Cordilleran planar and linear fabrics, and the rocks were apparently only affected by the same folds). These relationships imply that an undetectable amount of ductile strain occurred between the intrusion of Bourne granite in the Proterozoic and intrusion of sample 60A pegmatite.

Six zircon analyses from sample 60A (three to six grains each; Table 4-1) define two discordia lines (through three analyses each) that contain the analysis lying closest to concordia (Fig. 4-6c). One discordia has upper and lower intercepts of 1767 ± 16 and 51.6 ± 0.2 Ma, respectively (MSWD = 0.1), and another discordia has upper and lower intercepts of 2158 ± 116 and 52.0 ± 0.3 Ma, respectively (MSWD = 2.5). The analyses are 98-99% discordant from the upper intercepts. The upper intercepts are interpreted as average ages of inherited material and the lower intercepts as igneous crystallization ages. Another zircon analysis lies between the discordias and is thought to contain an inherited component of intermediate average age. Based on the lower intercepts, the interpreted igneous crystallization age is 51.8 ± 1.0 Ma.

Sample 292B is from a 0.3 m thick pegmatite layer that intruded into basement amphibolitic gneiss, south of Pettipiece Pass (Fig. 4-3). Intrusion is interpreted as having postdated a Proterozoic gneissosity in the host (age based on cross-cutting relationships of nearby rocks (Chapter 2)) because of the high discordance between the layer and the gneissosity (Figs. 4-12a,b). Similar to sample 60A pegmatite, sample 292B is deformed by folds (Figs. 4-12a,b) that are considered as having partly to completely transposed the Proterozoic gneissosity into the Cordilleran foliation. Quartz and feldspar in the layer were strongly flattened into a foliation that parallels the axial surface of these folds (Fig. 4-12c), which are correlated with the synmetamorphic Cordilleran folds because axial surfaces of each set are parallel and hinge lines are coaxial. Sample 292B was dated because these relationships indicate that deformation responsible for the Cordilleran foliation occurred, at least in part, after intrusion. The amount of deformation that took

place prior to intrusion is unknown, but it is inferred to be minimal based on the state of the Proterozoic gneissosity. Five zircon analyses (one to five grains each; Table 4-1) define two discordia lines that contain the analysis lying closest to concordia, which is composed of a single tip of a grain (Fig. 4-6d). One discordia through three analyses has upper and lower intercepts of 1782 ± 13 and 54.3 ± 2.0 Ma, respectively (MSWD = 85), and another discordia through three analyses has upper and lower intercepts of 1826 ± 13 and 56.2 ± 2.3 Ma, respectively (MSWD = 115). The analyses are 53-97% discordant from the upper intercepts. Upper and lower intercepts are interpreted as the average ages of inherited material and igneous crystallization ages, respectively. Based on the lower intercepts, the interpreted igneous crystallization age is 55.4 ± 3.0 Ma.

Proterozoic intrusive rocks

Upper level

Sample 180 is from a 0.2 m thick leucogranitic layer south of Ratchford Creek and west of Myoff Creek, in the core of the Kirbyville anticline (Fig. 4-3; Chapter 1). The sample was collected in the hinge of an outcrop-scale isoclinal fold, ~300 m structurally below the Monashee décollement, in a setting that suggests there was substantial strain after the interpreted crystallization age of 1762 ± 6 Ma. Sample 187 is from the 1862 ± 1 Ma Kirbyville gneiss, a hornblende biotite granodioritic orthogneiss north of Blais Creek (Fig. 4-3; Chapter 1). It contains a strong gneissosity that is interpreted as having a Cordilleran origin (Fig. 4-13a). Sample 207 is from the 724 ± 5 Ma Mount Grace gneiss,

a syenitic orthogneiss from a concordant, ~12 km long and up to 1 km thick body lying in the overturned limb of the Mount Grace syncline (Fig. 4-3; Chapter 1). It contains a strong gneissosity and lineation that are interpreted as having Cordilleran origins (Fig. 4-13b). The strong deformation fabrics and lack of intrusive contacts associated with these orthogneisses suggest that there was substantial post-Proterozoic strain in their vicinity.

Middle level

Sample 341 is from a 2-3 m thick pegmatite layer northeast of Pettipiece Pass (Fig. 4-3; Chapter 1). It is boudinaged and is concordant with the main deformation fabric and lithologic layering in the host amphibolitic gneiss, suggesting that there was substantial strain after the interpreted crystallization age of 1852 ± 4 Ma.

Lower level

Bourne granite is grey medium-grained biotite granite that intruded into basement orthogneiss between Pettipiece Pass and Bourne Glacier (Fig. 4-3; Chapter 2). It exists in ubiquitous 0.2-1.0 m thick layers that emanate from a small number of 100 m wide stocks (Fig. 4-11). The layers were correlated in the field on the basis of colour, structural state, and mineralogy, and in thin section on the basis of abundant coarse epidote and zircon with clear, rod-shaped inclusions. Bourne granite appears nearly undeformed in outcrop due to its equigranular and homogeneous nature, but it does contain a weak to moderate lineation that is shared with pegmatite layers that are thought to be coeval with sample

60A (Fig. 4-11e). Granite intrusion is interpreted as having postdated a migmatitic gneissosity based on the local high discordancy between the layers and the gneissosity, as well as on the pristine appearance of the granite where it is in contact with the highly tectonized host (Chapter 2). This is thought to be the same gneissosity that was postdated by a 2051 ± 12 Ma amphibolite dyke (sample 337, see below). An igneous crystallization age of 1848 ± 3 Ma was derived from samples dated in Armstrong et al. (1991) and Chapter 2.

Samples 84 and 308 are 2-4 m thick garnet muscovite biotite pegmatite dykes that intruded into basement augen gneiss east of the toe of Bourne Glacier (Fig. 4-3; Chapter 2). Intrusive contacts from sample 84 are gently folded and those in sample 308 are straight. In both rocks, the contacts are highly discordant to the gneissosity in the host gneiss. The dykes contain a weak foliation defined by aligned coarse biotite that parallels the gneissosity in the host. It is considered unlikely that the fabrics in the dykes are related to the deformation that developed the gneissosity in the host because of the strong discordancy between the intrusive contacts and the gneissosity. Therefore, postcrystallization strain in the vicinity of the dykes is thought to have been relatively minor after the interpreted crystallization ages of 1836 ± 2 Ma (sample 84) and 1845 ± 3 (sample 308).

Sample 91 is from a 0.5-1.0 m thick garnet muscovite biotite pegmatite dyke that intruded into basement and sample 308 pegmatite east of the toe of Bourne Glacier (Fig. 4-3; Chapter 2). Intrusive contacts are straight and highly discordant to the gneissosity in the host gneiss, and the dyke does not contain a deformation fabric, as it only has random

igneous textures. These relationships suggest that there was no postcrystallization strain in its vicinity after the interpreted crystallization age of 1834 ± 2 Ma.

Sample 337 is from a plagioclase hornblende amphibolite dyke ~0.2 m thick that intruded into basement orthogneiss south of Pettipiece Pass (Fig. 4-3; Chapter 2). Its intrusive contacts are highly discordant to the gneissosity in the host gneiss, suggesting that the gneissosity formed prior to the interpreted crystallization age of the protolith of the sample, which is 2051 ± 12 Ma. The relationships of this sample are important because they show that the younger Proterozoic samples described above were intruded after the dominant Proterozoic deformation event, suggesting that these samples were undeformed prior to the Tertiary. The layer and a foliation within it parallel the axial surfaces of folds that deformed the gneissosity in the host. On the basis of style and orientation, these folds are thought to be Cordilleran structures, similar to those that deformed samples 60A and 292B.

TECTONIC IMPLICATIONS

Early Cretaceous metamorphism

The discordant array of monazite U-Pb analyses from a schist on the east side of Frenchman Cap dome (sample 126, Chapter 3) suggests that conditions necessary for monazite growth were attained in the Early Cretaceous. The interpretation of partial Pb loss during early Tertiary metamorphism and the lack of analyses with evidence for Early Cretaceous growth outside of this schist indicate that this growth was localized, the

reasons for which are unknown. Growth was probably not associated with elevated temperatures because U-Pb data from Proterozoic monazite and titanite do not indicate Pb loss at this time. These data are simply taken to show that perhaps the complex was subjected to upper greenschist to lower amphibolite facies metamorphism in the Early Cretaceous, which would suggest that initial loading of thrust sheets onto the complex and/or deep sedimentary burial occurred at that time. More data are obviously needed before interpretations can be made with greater confidence.

Variation in the age and intensity of metamorphism

The interpreted relationship between U-Pb dates and peak metamorphism in Frenchman Cap dome (Chapter 3) requires a progressive downward younging in the presence of elevated temperatures (Figs. 4-3, 4-5, 4-7a). For example, higher structural levels were at the thermal peak when deeper levels were at lower temperatures, presumably below those necessary for monazite growth. Combining the U-Pb dates with hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ dates (Sanborn 1996) indicates that the thermal peak and the immediately following rapid cooling in higher levels occurred during the interval in which deeper levels were presumably heating toward a thermal peak that was attained a few million years later (Fig. 4-7b). Biotite $^{40}\text{Ar}/^{39}\text{Ar}$ dates indicate that higher levels had cooled significantly by the time the deepest schists attained the thermal peak. Peak metamorphism in the higher levels was synchronous with the ~59 Ma thrust motion on the Monashee décollement documented by Carr (1992) in southern Thor-Odin dome. The only age constraint on thrusting at the latitude of Frenchman Cap dome is that it must

predate 58-55 Ma (Parrish 1995). Peak metamorphism in lower levels was synchronous with motion on normal faults that are considered to have been largely responsible for exhumation of the complex, such as the Columbia River fault at the east margin of the complex (active between ~53 Ma and ~50 Ma (Parkinson 1992)) and the Okanagan - Eagle River fault system to the west (active after ~51 Ma (Parrish et al. 1988)).

Parrish (1995) showed that metamorphism in the Selkirk allochthon was diachronous, occurring from 170 Ma at the highest level to 60 Ma at the base, over a thickness of >10 km. Parrish considered the resulting “nested thermal evolution curves” (p. 1631) to be consistent with the evolution of rocks progressively buried under a tectonic load in the hinterland of a foreland fold and thrust belt as it propagated toward the foreland. In this model, warmer rocks progressively overrode cooler rocks as displacement migrated forward, synchronous with cooling and exhumation of previously tectonized rocks at higher levels. The fact that metamorphism in Frenchman Cap dome is also diachronous and is continuous in age and grade with that in the allochthon suggests a common tectonic process. The applicability of such a model to the dome is considered below.

The modeling of partial Pb loss in Proterozoic and Cretaceous monazite and titanite as diffusion in episodes of heating of fixed duration and temperature (Chapter 3) indicates that higher levels of the dome were metamorphosed to a greater intensity than deeper levels. Determining whether variations in duration or temperature were mostly responsible for the intensity differences is a difficult task, but some first order possibilities are suggested given the available data. Comparing pelitic mineral assemblages in the

cover sequence with a petrogenetic grid (Fig. 3-5; Spear and Cheney 1989) shows that peak temperatures were generally in the vicinity of $\sim 700\text{--}750^\circ\text{C}$. If this inference of similar peak temperatures is indeed true, then most of the measured intensity differences must be due to variable durations of the elevated temperatures. The modeling of Pb diffusion (Chapter 3) can give an indication of the duration differences. The temperature-duration curves for Pb loss in titanite (Fig. 3-10) show that durations within the lower level increase upward from a deep position to a high position by 1.5 and 1 Ma if temperatures throughout the level were 650°C and 700°C , respectively, and diffusion parameters are taken from Heaman and Parrish (1991). The curves also show that durations in the upper level are 4 Ma greater than those in the deepest level if temperatures in all levels were 700°C . The duration difference reduces to 2 Ma if the upper level was at 750°C . In all cases, the duration differences are negligible if diffusion parameters are taken from Cherniak (1993). Modeling of Pb loss in monazite (Fig. 3-10) indicates that durations in the middle level are 2 Ma greater than those in the lower level if all levels were at 700°C , and diffusion parameters are taken from Suzuki et al. (1994). The duration difference reduces to 1 Ma if the middle level was at 725°C and durations are the same at greater temperature differences. Duration differences are ~ 1 Ma longer if diffusion parameters are taken from Smith and Giletti (1997).

Because monazite and xenotime $^{207}\text{Pb}/^{235}\text{U}$ dates from schist are interpreted as representing the age of peak metamorphism, then independent evidence for variable durations of elevated temperatures are thought to be recorded by these dates. Figure 4-7a shows that seven deeper samples (24, 25, 70, 85, 101, 112, 166) have a 0.6-1.4 Ma spread

in dates within each sample and a 3.8 Ma spread between them, whereas four higher samples (165, 214, 257, 334) have a 1.9-4.6 Ma spread in dates within each sample and a 9.7 Ma spread between them.

Journeay (1986) interpreted the pelitic mineral assemblages as showing that an inverted metamorphic sequence exists beneath the Monashee décollement. The Pb loss data are compatible with an inverted sequence because they show that metamorphism in higher levels was more intense (longer-lived and/or hotter) than that in deeper levels. More importantly, U-Pb dates show that the thermal peak in higher levels occurred when deeper levels were below but presumably heating toward the thermal peak (Fig. 4-7a,b). This inverted sequence was interpreted by Journeay as having developed during a low pressure metamorphic overprint that was associated with an inverted thermal gradient. Parrish (1995) interpreted all metamorphism in the dome as having occurred during a single event that was associated with an inverted thermal gradient. This conclusion is based on the short duration of the event and the fact that the base of the Selkirk allochthon was at higher temperatures than the complex during the event. The possibility of an inverted thermal gradient is discussed in the tectonic model that is presented below, after consideration of the age and intensity of deformation.

Variation in age and intensity of deformation

The age and intensity of deformation in Frenchman Cap dome vary with structural level (Figs. 4-5, 4-7c; Table 4-2). In the upper level, the deformed nature of a 57.8 ± 1.5 Ma leucosome layer (sample 191) requires that synmetamorphic Cordilleran deformation

occurred, at least in part, at a later time but probably not much later because postdeformational leucosome (sample 223) and pegmatite (sample 263) to the south intruded at 57.5 ± 1.5 and 55.5 ± 1.0 Ma, respectively. There are no constraints on how much strain took place prior to ~ 58 Ma (e.g., whether the Frenchman Cap isoclinal folds developed by that time). The age brackets on deformation are similar to those in southern Thor-Odin dome, where motion on the Monashee décollement occurred at ~ 59 Ma and ceased by ~ 58 Ma (Carr 1992), and in the upper level of Frenchman Cap dome, where synmetamorphic deformation ceased by ~ 58 -55 Ma (Scammell 1993; Parrish 1995). In structurally low parts of the middle level, the deformed nature of 50.5 ± 1.0 Ma leucosome (sample 258) and 51.0 ± 1.0 Ma pegmatite (sample 12) layers requires that the syn- to postmetamorphic Cordilleran deformation occurred, at least in part, at a later time. There are no constraints on how much deformation took place prior to ~ 51 Ma, so it is unknown whether any of it was synchronous with deformation at higher levels. Variably deformed 53-50 Ma pegmatite layers dated by Parkinson (1992) show that high temperature strain occurred at that time in Thor-Odin dome. However, this strain was inferred to have resulted from extension on the Columbia River fault rather than regional compressional deformation. In the lower level, the deformed nature of 55.4 ± 3.0 (sample 292B) and 51.8 ± 1.0 Ma (sample 60A) pegmatite layers requires that the synmetamorphic Cordilleran deformation occurred, at least in part, at a later time but probably not much later because a nearby postdeformational pegmatite dyke (sample 57) intruded at 49.5 ± 0.5 Ma.

The highly strained state of three Proterozoic rocks (samples 180, 187, 207; Figs. 4-13a,b) that intruded into the upper level and of one rock (sample 341) that intruded into the middle level is evidence that Cordilleran deformation in the host rocks was substantial. In high structural positions in the lower level, the relationships between pegmatite layers (samples 60A and 292B), layers of the Proterozoic Bourne granite, and a Proterozoic migmatitic gneissosity show that deformation was insignificant between the Proterozoic and ~52 Ma (Figs. 4-11, 4-12). The presence of weakly strained and unstrained Proterozoic pegmatite dykes (samples 84, 91, 308) in deep exposures in the lower level is evidence that Cordilleran deformation was locally weak to nonexistent. The nature of the strain in the lower level is obviously best known in the limited areas with U-Pb data, but for the purpose of making the conclusions that are presented next, it was necessary to extrapolate from these areas. This was done largely by using layers and dykes of Bourne granite, which are the most important structural and time markers in the dome because they are distinctive-looking, ubiquitous, and known to have crystallized after the Proterozoic deformation.

The strain gradient that developed in the early Tertiary separates highly deformed rocks in the cover sequence from locally weakly deformed to undeformed rocks in the basement. The gradient is thought to exist almost entirely at the contact between the uppermost basement paragneiss horizon and the underlying orthogneiss bodies (Fig. 4-3) because (i) paragneiss contains a straight gneissosity and a lineation (Fig. 4-13c) that are interpreted as having Cordilleran origins on the basis of a similarity in orientations between these fabrics and those measured in nearby fabrics that are dated as early Tertiary (i.e.,

fabric in sample 60A pegmatite (Fig. 4-11)) and (ii) orthogneiss a few hundred metres below the paragneiss preserves Proterozoic fabrics and intrusive contacts. However, Cordilleran deformation must have been more complex than a simple gradient because strain was heterogeneous in deep levels, as it was strongly partitioned at various scales into high and low strain domains that are typically dependent on rock types (i.e., rheology of the rocks). Rock types are simplified in following discussion as paragneiss, granitic orthogneiss, and amphibolitic gneiss. Paragneiss in all localities is interpreted as having been highly strained by Cordilleran deformation because it contains a single gneissosity (Fig. 4-13c) that is presumed to be early Tertiary based on the orientation of its fabric elements. Paragneiss was also locally deformed by postmetamorphic folds that are rare in surrounding rocks, but well developed in the cover sequence on the east side of the dome. In contrast to paragneiss, granitic orthogneiss varies on a metre scale from containing a gneissosity that was dated as Proterozoic to containing a foliation that is interpreted as having a Cordilleran origin (Fig. 4-14). Again, a foliation in any given outcrop was mapped as Cordilleran on the basis of (i) similarities in orientation of fabric elements with nearby fabrics that are known to be early Tertiary (e.g., fabric in sample 60A pegmatite (Fig. 4-11)) and (ii) the existence of the foliation in the axial planes of folds that are correlated on the basis of style and orientation with those dated nearby as early Tertiary (e.g., folds that deformed sample 60A pegmatite (Fig. 4-11d)). Figure 4-14 shows that the Cordilleran foliation is a composite transposition fabric that resulted from reorientation and shearing of the Proterozoic gneissosity during Cordilleran folding and from fabric development and leucosome generation in the axial surfaces of related folds. This

composite fabric is best developed on the limbs of tight to isoclinal folds and where the Proterozoic gneissosity had a preexisting orientation that was coincident with the axial surfaces of Cordilleran folds. It is least developed in fold hinge zones. The Proterozoic gneissosity is generally well preserved in basement amphibolitic gneiss (e.g., host to sample 292B pegmatite (Fig. 4-12)), in which it was typically only drastically reoriented in the limbs of rare tight Cordilleran folds.

The downward weakening in the intensity of deformation to the deepest mapped part of the study area, where post-Proterozoic strain was heterogeneous according to rheology, leads to the question of whether the base of significant Cordilleran deformation is exposed in Frenchman Cap basement. The presence of thrust faults or ductile shear zones beneath the dome is an integral part of some previous tectonic models (Brown et al. 1986; Journeay 1986). Large-scale planar features in the basement are present on Lithoprobe seismic reflection profiles (Brown et al. 1992; Cook et al. 1992 and references therein), but their age and nature cannot be determined. Reconnaissance mapping outside the study area in the deepest exposure in the dome (at Frisby Glacier in the southern part of the dome) show that Cordilleran deformation weakly affected this level. Given these data, it is possible, but still unproven, that the base of significant Cordilleran deformation exists in the dome, which would mean that the rocks below are an intact part of the North American craton.

TECTONIC MODEL

An explanation for the inverted metamorphic sequence in Frenchman Cap dome, which was delineated by pelitic metamorphic mineral assemblages (Fig. 4-4; Journeay 1986 and references therein) and U-Pb isotopic data (Chapter 3), is a central part of tectonic models of the dome. Determining whether the sequence was associated with an inverted thermal gradient is a difficult endeavour in the dome because (i) it requires knowledge of which rocks were overlying other rocks at any given time during metamorphism and (ii) the dome was significantly deformed by structures that were active during deformation. Models presented by Jamieson et al. (1996) and Hubbard (1996) depict how an inverted metamorphic sequence can form in the absence of an inverted thermal gradient if there were syn- to postmetamorphic displacements within it and strain gradients were present. These models show that displacements can telescope a normal metamorphic sequence that was once laterally extensive into a stacked inverted sequence without ever placing warmer rocks over cooler rocks.

The existence of an inverted thermal gradient in the basement rocks of the dome (lower structural level) is suggested by (i) Pb loss data from titanite in two localities that show that the metamorphic intensity decreases downward and (ii) U-Pb dating and mapping of Proterozoic intrusive rocks that show that most of this level was only weakly strained since the Proterozoic. However, given that metamorphic intensity is related to temperature and duration, it cannot be ruled out that differences in duration rather than temperature were responsible for the variations in the intensity. The inverted metamorphic sequence in the cover sequence (middle and upper structural levels) cannot be attributed

to an inverted thermal gradient because it is probable that displacements associated with synmetamorphic structures resulted in the juxtaposition of rocks with widely varying initial positions that reached peak temperatures at different times and in different positions within the orogen. These structures include the Frenchman Cap isoclines (Journeay 1986 and references therein), which are kilometre-scale east-verging folds with intensely sheared zones along their limbs (such as the Ratchford Creek and Anstey Range faults). Previous workers mapped the isoclines as premetamorphic structures because they are overprinted by synmetamorphic folds that dominate at the outcrop scale. Yet, because there are no constraints on the temperatures that existed during development of the isoclines, it is possible that they formed just prior to the thermal peak. Significant deformation throughout the dome is known to have outlasted the thermal peak because metamorphic minerals, such as kyanite and sillimanite, are typically oriented within the Cordilleran lineation. Also, the U-Pb dating performed during this study shows that Cordilleran folds and fabrics developed during peak metamorphism.

The proposed tectonic model, which is schematically illustrated in Fig. 4-15, contends that the inverted metamorphic sequence in Frenchman Cap dome resulted mostly from the synmetamorphic tectonic assembly (e.g., Jamieson et al. 1996) of an originally normal metamorphic sequence, with a possible minor input from localized downward heat conduction. In this model, a normal metamorphic sequence that previously extended a significant distance (a few tens of kilometres or more) to the west of the presently exposed basement in the centre of the dome was telescoped via east-directed folding and thrusting into the present 6 km thick inverted sequence. It is possible that during juxtaposition

warmer rocks locally overrode cooler rocks, such as in areas adjacent to the Monashee décollement and in areas that were affected by penetrative ductile shearing associated with the east-verging folding. Stratigraphic marker horizons in the cover sequence have been mapped as nearly continuous through the dome (Figs. 4-3, 4-4), perhaps suggesting that the higher levels, which originated to the west, were not displaced more than a few tens of kilometres relative to the deeper levels. However, because the horizons are sheared and extensively thinned along the limbs of the isoclines, the apparent continuity does not constrain the amount of displacement associated with the structures, other than requiring that higher levels were probably displaced less than many tens of kilometres relative to deeper levels.

Thermotectonic history of the structural levels

An integral part of this model is that rapid cooling in each level began immediately after the attainment of the thermal peak, thereby preserving the inverted metamorphic sequence from further temperature increases that would have otherwise overprinted the mineral assemblage and obliterated the U-Pb isotopic record of the inversion.

The ~64-60 Ma peak metamorphism in the highest exposed rocks in Frenchman Cap dome is older and more intense than that in deeper rocks. These characteristics are attributed to the fact that these rocks were metamorphosed farthest to the west, where the crust reached a substantial thickness before it was thickened in the east. The burial depth was sufficient for kyanite stability, and temperatures adjacent to the allochthon were sufficient for the rocks to surpass the second sillimanite isograd. Cooling of the upper

level to temperatures below those necessary for monazite growth is thought to have taken place shortly after growth of the youngest dated monazite, generally at ~56 Ma. The retrograde part of the P-T path followed by these rocks must have passed through the sillimanite and andalusite fields, whereupon much of the kyanite was pseudomorphed. This cooling may have occurred during thrusting on underlying faults, such as the Anstey Range fault, that placed upper level rocks over cooler rocks below, rapidly changing the temperatures of rocks on both sides of the fault. Displacements associated with development of the Frenchman Cap isoclines is another process that may have placed warmer rocks in the cores of anticlines over cooler rocks in the cores of a synclines. Exhumation due to motion on overlying extensional faults, such as the Columbia River and Okanagan - Eagle River faults, is thought to have been too young to have caused much of this cooling because these faults were active at ~53 Ma and probably were active for just a few million years prior to that time. However, this extensional faulting coincides with rapid cooling of rocks at structurally high parts of the middle level, where cooling is known, on the basis of hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ data, to have begun by ~55 Ma.

The structurally low parts of the middle level and the lower level attained peak metamorphism from ~54 to 49 Ma, at lower grade conditions and after peak metamorphism in overlying rocks (Fig 4-7). The structural distance between the cooler rocks above and warmer rocks below is typically quite narrow. For example, rocks on the west side of the dome that cooled below the hornblende Ar closure temperature at ~55 Ma are separated by <1 km from underlying rocks (sample 214) that reached their thermal peak as early as ~54 Ma (Fig. 4-7a). Also, the typical metamorphic age relationships are

inverted in one locality on the west side of the dome; sample 257 was at its thermal peak slightly prior to the peak in overlying rocks, such as sample 214 (Fig. 4-7a). These relationships suggest that modification of the metamorphic sequence, including thinning and local duplication, occurred after the thermal peak.

It is tentatively inferred that the thermal gradient in the lower level at ~54-49 Ma may have been inverted on the basis that Pb loss data suggest that rocks in a structurally high position in this level were metamorphosed more intensely than rocks below. This possibility needs to be tested further, most directly by determination of the amounts of Pb loss that occurred in monazite and titanite throughout all structural positions within the lower level. A convincing argument against the existence of an inverted thermal gradient is that when deeper levels were at elevated temperatures higher level rocks that would have been the heat source had already undergone initial cooling. Based on Ar-Ar data, these higher level rocks were between ~500° and 300°C at ~54-49 Ma, which means that there was no significant overlying heat source to cause an inverted thermal gradient in the lower level.

The depth of the ~54-49 Ma metamorphism in the deeper levels is thought to have been less than the depth of the kyanite grade metamorphism in higher levels because there is no evidence suggesting that these rocks ever contained kyanite and because temperatures are inferred, based on the Pb loss data, to have been lower. The temporal overlap of peak metamorphism in the deeper rocks and extension-related exhumation of the dome is used to explain why metamorphism was less intense and at lower pressures

than that in higher levels; the rocks were exhumed as they began to attain elevated temperatures.

In conclusion, the model proposes that the variation in the age and intensity of tectonism with structural level in Frenchman Cap dome resulted from east-directed displacements within the dome, such as those associated with kilometre-scale isoclinal folds, possibly from local minor downward heat conduction during the displacements, and finally, exhumation that occurred shortly after or during the attainment of high temperatures (Fig. 4-15). Consideration of the physics of heat transfer, which is beyond the scope of the present study, would undoubtedly test and place constraints on the model.

Table 4-1. U-Pb analytical data.

Analysis ^a	Wt. ^b U (μg)	U (ppm)	Pb* ^c (ppm)	²⁰⁶ Pb/ ²⁰⁴ Pb ^d	Pb _c ^e (pg)	²⁰⁸ Pb/ ²⁰⁴ Pb ^f (%)	²⁰⁶ Pb/ ²³⁸ U ^g	²⁰⁷ Pb/ ²³⁵ U ^g	Corr. coeff.	²⁰⁷ Pb/ ²⁰⁶ Pb ^g	²⁰⁷ Pb/ ²⁰⁶ Pb ^h age (Ma)
12 (Fig. 4-6b) - Pegmatite layer, 0.5 km thick, boudinaged, concordant or nearly so with dominant Cordilleran foliation, intruded into lower part of cover sequence; located 1.4 km S of Pettipiece Pass at 2200 m, 387480E-5695060N (UTM zone 11, BCGS map sheet 82M.037)											
A* 4,105,L1	15	9915	70.2	2249	34	0.2	0.007853 ± 0.13%	0.05139 ± 0.16%	0.88	0.04746 ± 0.08%	72.4 ± 3.6
B* 9,74,t,L1	10	8967	63.67	2787	16	0.2	0.007875 ± 0.10%	0.05154 ± 0.14%	0.79	0.04747 ± 0.09%	73.1 ± 4.2
C 3,105,L1	23	17833	126	2105	97	0.3	0.007828 ± 0.19%	0.05121 ± 0.20%	0.80	0.04744 ± 0.12%	71.6 ± 5.9
D 3,105,L2	18	9260	66.66	1777	47	0.3	0.007978 ± 0.16%	0.05189 ± 0.27%	0.54	0.04718 ± 0.23%	58.2 ± 10.9
E 3,105,L2	16	6768	48.78	2165	26	0.2	0.007989 ± 0.10%	0.05216 ± 0.14%	0.79	0.04735 ± 0.09%	67.0 ± 4.2
F 3,105,L2	22	8128	58.26	3374	27	0.2	0.007952 ± 0.10%	0.05188 ± 0.12%	0.88	0.04732 ± 0.06%	65.2 ± 2.7
57 (Fig. 4-6b) - Pegmatite dyke, 0.7 m thick, postdeformational, intruded into basement orthogneiss; located 2.8 km S of Pettipiece Pass at 2080 m, 387830E-5693680N (UTM zone 11, BCGS map sheet 82M.037)											
A* 1,105,t,cs	14	20630	145.6	7585	18	0.5	0.007805 ± 0.15%	0.05097 ± 0.16%	0.97	0.04736 ± 0.04%	67.4 ± 1.9
B* 6,74,cs	3	11957	84.69	1826	11	0.4	0.007836 ± 0.12%	0.05140 ± 0.22%	0.61	0.04757 ± 0.17%	78.1 ± 8.3
C 1,149,tn	27	16614	116.7	1583	140	0.5	0.007763 ± 0.28%	0.05092 ± 0.30%	0.73	0.04757 ± 0.21%	78.1 ± 10.1
D 2,105,tn	10	12333	87.57	599	101	0.2	0.007868 ± 0.18%	0.05199 ± 0.74%	0.40	0.04792 ± 0.68%	95.3 ± 32.4
E 6,74,cs	6	10862	75.43	270	134	0.1	0.007709 ± 0.29%	0.04986 ± 0.88%	0.63	0.04691 ± 0.74%	44.7 ± 35.3
F 3,74,t,tn	4	21408	150.1	1158	40	0.3	0.007765 ± 0.11%	0.05061 ± 0.19%	0.71	0.04727 ± 0.14%	62.9 ± 6.5
G 9,74,cs	17	2079	14.67	1010	18	0.4	0.007807 ± 0.10%	0.05124 ± 0.22%	0.61	0.04760 ± 0.18%	79.5 ± 8.4
60A (Fig. 4-6c) - Pegmatite layer, 0.5 m thick, folded and contains dominant Cordilleran foliation, intruded into basement orthogneiss; located 2.7 km SSE of Pettipiece Pass at 2130 m, 387480E-5695064N (UTM zone 11, BCGS map sheet 82M.038)											
A* 3,149	39	7419	57.01	12372	13	0.1	0.008501 ± 0.13%	0.05993 ± 0.14%	0.97	0.05113 ± 0.04%	246.7 ± 1.6
B* 3,105	26	3842	35.56	5230	12	0.3	0.010127 ± 0.11%	0.08372 ± 0.12%	0.91	0.05996 ± 0.05%	602.1 ± 2.3
C* 5,105,t	24	5097	40.19	2832	24	0.2	0.008697 ± 0.10%	0.06383 ± 0.13%	0.88	0.05323 ± 0.06%	338.8 ± 2.7
D 4,105,t	14	5767	44.63	1070	42	0.1	0.008565 ± 0.12%	0.06006 ± 0.31%	0.47	0.05086 ± 0.28%	234.3 ± 12.8
E 6,74	12	5733	46.75	2149	19	0.1	0.008991 ± 0.11%	0.06708 ± 0.15%	0.81	0.05411 ± 0.09%	375.6 ± 4.1
F 6,74,t	22	10034	74.22	4937	24	0.0	0.008210 ± 0.17%	0.05475 ± 0.18%	0.96	0.04836 ± 0.05%	117.0 ± 2.2
263 (Fig. 4-6a) - Pegmatite stock, 0.5 km wide, postdeformational, intruded into lower part of cover sequence; located 7.5 km SE of Mount Grace at 1290 m, 376750E-5693840N (UTM zone 11, BCGS map sheet 82M.037)											
A 3,149	52	2437	19.36	9139	8	1.6	0.008686 ± 0.15%	0.05657 ± 0.15%	0.78	0.04724 ± 0.10%	61.1 ± 4.7
B 3,149	39	1514	12.9	1989	16	10.1	0.008501 ± 0.11%	0.05541 ± 0.17%	0.68	0.04727 ± 0.12%	63.0 ± 5.9
C 3,149	50	1439	11.52	3023	13	1.1	0.008817 ± 0.11%	0.05758 ± 0.17%	0.62	0.04737 ± 0.14%	67.7 ± 6.5
D 7,105	48	1445	11.6	4352	9	2.1	0.008731 ± 0.10%	0.05693 ± 0.14%	0.73	0.04729 ± 0.10%	63.9 ± 4.6
292B (Fig. 4-6d) - Pegmatite layer, 0.3 m thick, folded and contains dominant Cordilleran foliation, intruded into basement orthogneiss; located 3.7 km SE of Pettipiece Pass at 2300 m, 389350E-5693440N (UTM zone 11, BCGS map sheet 82M.038)											
A 2,105,cs,t	24	1413	105.9	3425	46	1.0	0.078096 ± 0.29%	1.13568 ± 0.30%	0.99	0.10547 ± 0.04%	1722.5 ± 1.6
B 4,74,cs,t	15	1968	144.3	11657	12	0.9	0.076595 ± 0.18%	1.09484 ± 0.19%	0.99	0.10367 ± 0.03%	1690.8 ± 1.2
C 1,149,tn,t	13	5805	97.39	1752	50	1.3	0.017835 ± 0.11%	0.19845 ± 0.14%	0.90	0.08070 ± 0.06%	1214.1 ± 2.3
D 1,105,tn	6	1578	242.8	4106	23	2.3	0.157603 ± 0.12%	2.38890 ± 0.13%	0.96	0.10993 ± 0.04%	1798.3 ± 1.4
E 5,74,cs,t	17	1450	30.1	1877	18	0.9	0.022061 ± 0.10%	0.26026 ± 0.15%	0.75	0.08556 ± 0.10%	1328.3 ± 3.9

Isotopic composition of Carleton University lab blank (uncertainty is one standard deviation) - 206:207:208:204 = 19.01 ± 0.36 : 15.64 ± 0.20 : 38.23 ± 0.74 : 1

^aIn first column, A-G = fraction codes for zircon analyses, * = chemical analysis done at the Geological Survey of Canada in Ottawa (all other analysis was done at Carleton University); in second column, 1-9 = number of grains analyzed, 74-149 = average size in μm prior to abrasion with pyrite (Krogh 1982), cs = colourless, tn = tan colour, t = tips of grains, L1 = layer one, L2 = layer two.

^bWeighing uncertainty is 1 μg.

^cRadiogenic Pb.

^dMeasured ratio, corrected for spike and Pb fractionation of 0.09% ± 0.03%/AMU.

^eTotal common Pb in analysis, corrected for spike and fractionation.

^fRadiogenic ²⁰⁸Pb, expressed as % of total radiogenic Pb.

^gCorrected for blank Pb and U and common Pb (Stacey-Kramers model Pb composition equivalent to the interpreted age of the analyses); errors are one standard deviation in %.

^hErrors are two standard deviations in Ma.

Table 4-2. Summary of U-Pb ages from early Tertiary intrusive rocks, arranged according to structural level.

Sample	Rock type	Key relationships ¹	Age (Ma)	Mineral
<i>Upper Level</i>				
191 ²	leucosome	contains kyanite, sillimanite, and andalusite; kyanite is lineated; concordant with S _C	57.8 ± 1.5	zircon, monazite
223 ²	leucosome	contains kyanite, sillimanite, and andalusite; igneous fabrics only	57.5 ± 1.5	zircon, monazite
263	pegmatite	igneous fabrics only; truncates S _C	55.5 ± 1.0	zircon
<i>Middle Level</i>				
12	pegmatite	boudinaged by D _C ; deformed by postmetamorphic folds	51.0 ± 1.0	zircon
258 ²	leucosome	contains sillimanite; concordant with S _C	50.5 ± 1.0	zircon, monazite
<i>Lower Level</i>				
292B	pegmatite	deformed by F _C folds; contains S _C	55.4 ± 3.0	zircon
60A	pegmatite	deformed by F _C folds; contains S _C	51.8 ± 1.0	zircon
57	pegmatite	igneous fabrics only; truncates S _C	49.5 ± 0.5	zircon

¹S_C = dominant Cordilleran foliation; D_C = dominant synmetamorphic Cordilleran deformation; F_C = folds associated with dominant synmetamorphic Cordilleran deformation.

²U-Pb data presented in Chapter 3.

Figure 4-1. Tectonic assemblage map in the vicinity of the Monashee complex, modified after Wheeler and McFeely (1991), showing the location of Fig. 4-3. The dark grey square in inset locates the map with respect to the morphogeologic belts of the Canadian Cordillera.

Figure 4-3. Geologic map of northern Frenchman Cap dome, modified after Wheeler (1965), McMillan (1973), Psutka (1978), Höy and Brown (1980), Journeay (1986), and Höy (1987), showing localities of U-Pb samples (sample numbers and names are shown in parentheses) discussed in text. Descriptions of intensity of deformation are made with respect to the dominant Cordilleran deformation. A-A' locates the cross section shown in Figs. 4-4 and 4-5.

Figure 4-4. Cross section of northern Frenchman Cap dome, modified after Journeay (1986), showing approximate structural position of the metamorphic mineral assemblages in pelitic schists of the cover sequence (data from Journeay (1986) and references therein). Assemblages in the basement are not shown because it is uncertain whether the minerals grew during Proterozoic or early Tertiary metamorphism. Location of section and patterns for rock units are shown in Fig. 4-3.

Figure 4-5. Cross section of northern Frenchman Cap dome, modified after Journeay (1986), showing approximate structural position of the samples with U-Pb ages (in Ma). Location of section and patterns for rock units are shown in Fig. 4-3. D_C = synmetamorphic Cordilleran deformation.

Figure 4-6. U-Pb concordia plots. Ellipses for the analyses represent the two sigma uncertainty (95% confidence level) and analyses that are plotted with + symbols have ellipses that are too small to be seen at this scale. All analyses are from zircons. See Table 4-1 for analytical data.

Figure 4-7. (a) Summary of monazite, xenotime, and zircon U-Pb ages from pelitic schist and intrusive rocks plotted against relative structural depth, from Chapter 3. Hornblende and biotite $^{40}\text{Ar}/^{39}\text{Ar}$ ages from Sanborn (1996) are also shown. (b) Inferred temperatures plotted against time. (c) Ages of intrusive rocks plotted against relative structural depth. D_C = synmetamorphic Cordilleran deformation.

Figure 4-8a. Sample 263 is from a 55.5 ± 1.0 Ma pegmatite stock that intruded into amphibolitic gneiss in the upper level. The intrusive contacts are highly discordant to the Cordilleran foliation in the host rock, and the pegmatite does not contain any ductile deformation fabrics. These relationships suggest that pegmatite intrusion postdated development of the Cordilleran foliation in the host rock.

Figure 4-8b. Another example of the high discordance between (i) the intrusive contacts of the pegmatite stock from which sample 263 was collected and (ii) the Cordilleran foliation in the host rock. The host rock is pelitic schist. The scale is a geologic hammer that lies just above the contact, on the left side of the photograph.

Figure 4-9a. Sample 12 is from the 51.0 ± 1.0 Ma pegmatite layers that are boudinaged and concordant with the Cordilleran foliation in the host rock. Two layers were collected for dating. Layer 1 lies in the upper left corner of the photo and layer 2 lies to the immediate left of the geologist. The host rock is calc-silicate schist in the middle level. The view is to the east-northeast.

Figure 4-9b. The pegmatite layers shown in (a) are projected about 10 m into this cliff face, which is viewed looking west-southwest. The scale is a geologic hammer that lies just beneath the boudinaged pegmatite layer on the left.

Figure 4-9c. Similar pegmatite layers as shown in (a) and (b) exist ~100 m along strike to the north. The layers are boudinaged and concordant with the Cordilleran foliation in the host rock over several tens of metres. The view is to the southeast. These relationships suggest that the synmetamorphic Cordilleran deformation occurred, at least in part, after intrusion.

Figure 4-10. Sample 258 is from the 50.5 ± 1.0 Ma leucosome layer lying below the geologist that is boudinaged and concordant with the Cordilleran foliation in the host rock. The view is to the north. This and similar-looking layers intruded into amphibolitic gneiss in the middle level. These relationships suggest that synmetamorphic Cordilleran deformation occurred, at least in part, after intrusion.

Figure 4-11a. Sample 60A is from the 51.8 ± 1.0 Ma pegmatite layer (thickest white layer in photograph) that intruded into Proterozoic granite (Bourne granite) and granitic gneiss, in the lower level. The layer is nearly concordant with the gneissosity in the host granitic gneiss and the foliation within the layer. The view is to the northeast. These relationships suggest that the synmetamorphic Cordilleran deformation occurred, at least in part, after intrusion. The interpretation that the layer of granite appears to have only been subjected to the same amount of strain as the pegmatite is used as evidence that deformation between the Proterozoic and 51.8 ± 1.0 Ma was minimal.

Figure 4-11b. Line drawing of (a), labeling details.

Figure 4-11c. Detailed view of part of (a), showing the fabric relationships.

Figure 4-11d. In a locality a few hundred metres from (a), a pegmatite layer that is thought to be the same age as sample 60A was deformed by folds that are interpreted on the basis of orientation and style to be related to those known to be associated with the Cordilleran deformation. The host is Bourne granite. The view is to the northeast.

Figure 4-11e. The foliation in a pegmatite layer that is thought to be the same age as sample 60A contains a stretching lineation that is coaxial with the dominant Cordilleran lineation. The view is perpendicular to the foliation in the pegmatite layer, which lies a few hundred metres from (a). The host is Bourne granite.

Figure 4-12a. Sample 292B is from the 55.4 ± 3.0 Ma pegmatite layer (thickest white layer in photograph) that intruded into amphibolitic gneiss in the lower level. The layer is discordant to a Proterozoic gneissosity in the host and contains a foliation that lies parallel with axial surfaces of folds that deformed that gneissosity. The view is to the northeast. These relationships suggest that the synmetamorphic Cordilleran deformation occurred, at least in part, after 55.4 ± 3.0 Ma. The interpretation that the Proterozoic gneissosity is apparently only slightly folded in the upper part of outcrop is evidence that Cordilleran deformation was locally mild.

Figure 4-12b. Line drawing of (a), labeling details.

Figure 4-12c. Detailed view of the pegmatite, showing the elements that define the foliation, including flattened quartz and feldspar and aligned biotite.

Figure 4-13a. Sample 187 is from the 1862 ± 1 Ma Kirbyville orthogneiss, which forms basement in the core of the Kirbyville anticline, in the upper level. The view is to the north. This rock contains a gneissosity that is interpreted as having a Cordilleran origin, and its presence is evidence that Cordilleran deformation was intense in this level.

Figure 4-13b. Sample 207 is from the 724 ± 5 Ma Mount Grace orthogneiss, which lies in the upper level. The view is perpendicular to the gneissosity, which contains a strong lineation that is interpreted as having a Cordilleran origin. The presence of the gneissosity is evidence that Cordilleran deformation was intense in this level.

Figure 4-13c. Basement paragneiss immediately beneath the unconformity, in the upper part of the lower level, contains a straight gneissosity that is interpreted having a Cordilleran origin. Many of the lighter coloured layers probably have igneous protoliths, which are of unknown age. The view is to the northeast. The presence of the gneissosity is evidence that Cordilleran deformation was intense in this part of the lower level.

Figure 4-14a. Two gneissosities typically exist in basement granitic orthogneiss in the lower level. The older one, which is thought to be Proterozoic based on nearby U-Pb dating (Chapter 2), is deformed by synmetamorphic Cordilleran folds (viewed down plunge). The younger one is a composite transposition fabric that resulted from reorientation of the older one during Cordilleran folding and from fabric development and leucosome generation in the axial surfaces. The preservation of an older gneissosity is evidence that Cordilleran deformation locally had minor effects in this level.

Figure 4-14b. Line drawing of (a), labeling details.

Figure 4-15. The tectonic model proposed for Frenchman Cap dome is depicted by cross sections drawn to schematically represent the inferred crustal structure and metamorphic conditions at four different times. The model shows how rocks that were initially greatly separated from east to west prior to 64 Ma, denoted by locations A-C, were structurally telescoped by ~52 Ma, largely during development of kilometre-scale isoclinal folds. These folds, referred to as the Frenchman Cap isoclines, include the Sibley Creek syncline (SCS), Kirbyville anticline (KA), and Mount Grace syncline (MGS). Stretching of the limbs of the folds has been included in the model, but faults along the limbs of the folds, such as on the Ratchford Creek and Anstey Range faults, have not been considered because the amounts of displacement are unknown. Adding such displacement to the model would obviously increase the initial distance between the locations A-C. The recorded diachronous attainment of peak metamorphic conditions and cooling from peak conditions (see text) is largely explained by this model, which shows that the rocks attained these conditions prior to and during the structural telescoping.

CHAPTER 5

SUMMARY OF CONCLUSIONS

The Monashee complex occupies a nearly unique position in the Canadian Cordillera because it contains Paleoproterozoic basement gneisses that lie in a deep structural position in the orogen. Knowledge of the early geologic history of the gneisses, which are the westernmost exposures of Paleoproterozoic crystalline basement in North America, is necessary for establishing tectonic links between the complex and the western Canadian Shield. Knowledge of the age of the metasedimentary rocks that were deposited unconformably on the gneisses has importance for determining the depositional history of the Cordilleran miogeocline, which is necessary for reconstruction of the orogen to a pre-shortening configuration. Tectonism that occurred during Cordilleran orogenesis overprinted the Paleoproterozoic rocks, thermally as well as structurally, to a degree that the specifics of their earlier geologic history may never be revealed; only general events have been elucidated thus far. Determining the age and nature of Cordilleran tectonism is important because the complex is possibly the deepest structural exposure in the orogen, lying beneath a ductile thrust fault, the Monashee décollement, that is correlated with the sole thrust of the Rocky Mountain Foreland Belt, and it was affected by the youngest known high grade regional metamorphism and deformation events. Obviously, knowledge gained from these events has a bearing on Cordilleran tectonic models.

U-Pb geochronologic data and field relationships from 39 samples from Frenchman Cap dome in the northern part of the complex were used to investigate the following: (i) the deposition age of the cover sequence (Chapter 1), (ii) the Paleoproterozoic geologic history (Chapter 2), and (iii) the age and nature of Cordilleran tectonism (Chapters 3, 4). The existence of Paleoproterozoic crystalline rocks in the complex greatly assists the study of Cordilleran tectonism because these rocks contain minerals that suffered Pb loss during the overprint, which serve as indicators of the intensity of the thermal events, and they contain intrusive contacts and gneissosities that were deformed during the overprint, which serve as structural markers of the intensity of the deformation. U-Pb data and field relationships from minerals that grew during the overprint and rocks that intruded during the overprint are also useful for revealing the Cordilleran tectonic history.

PROTEROZOIC HISTORY

Chapter 1

The Monashee cover sequence is a platformal metasedimentary succession that occupies a nearly unique position in the Canadian Cordillera due to its unconformable contact with exposed crystalline basement. Zircon U-Pb data and field relationships from Frenchman Cap dome were used to show that the lower part of the sequence contains Paleoproterozoic rocks, the oldest known metasedimentary rocks in the Cordilleran miogeocline, and the upper part of the sequence is Mesoproterozoic or younger. Maximum age constraints on the lower part are provided by 1.99 Ga detrital zircons from

the basal unit and a 1862 ± 1 Ma orthogneiss upon which it was presumably deposited. Minimum age constraints are provided by rocks that intruded into the lower part: 1852 ± 4 Ma pegmatite, 1762 ± 6 Ma leucogranite, and 724 ± 5 Ma syenitic gneiss.

The upper part of the sequence must be considerably younger than the lower part because it contains a detrital zircon dated at ~ 1.21 Ga. Other detrital zircons, dated at Neoproterozoic (2.95-2.86 Ga) and Paleoproterozoic (1.85-1.81, 1.75 Ga), suggest a source in the western Canadian Shield. These ages constrain the thickness of Meso- and Neoproterozoic metasedimentary rocks in the cover sequence to be < 2 km. Combining these ages with previously interpreted Paleozoic deposition ages for the middle and upper parts of the sequence constrains the thickness to be < 0.2 km, considerably less than coeval rocks above the Monashee complex in the hanging wall of the Monashee décollement. Such a contrast suggests that deposition above and below the décollement occurred in different parts of the Cordilleran miogeocline.

Chapter 2

The Monashee complex contains the largest of four exposures of Paleoproterozoic rocks in the Canadian Cordillera. These rocks are composed of basement para- and orthogneisses, metasedimentary rocks in the base of an unconformably overlying cover sequence, and younger intrusive rocks. U-Pb geochronologic data and field relationships from Frenchman Cap dome were used to outline a Paleoproterozoic geologic history that includes six intrusive episodes, two metamorphic events, two depositional sequences, and one deformation event. Evidence for this history is best preserved in rocks that lie in deep

structural levels of the complex, where overprinting related to middle and upper amphibolite facies Cordilleran (early Tertiary) orogenesis was incomplete. Although interpretation of precise U-Pb ages is complicated by Pb loss and/or overgrowth that occurred in the dated minerals during at least one thermal overprint, there is sufficient age resolution on the intrusive events to conclude that the complex has affinities with the western Canadian Shield and the other Paleoproterozoic exposures in the Cordillera.

The dominant Paleoproterozoic rocks in the complex are large orthogneiss bodies dated at 2272 ± 8 Ma, 2099 ± 17 , and 2077 ± 2 Ma. These rocks and an older paragneiss sequence underwent the most significant Paleoproterozoic metamorphism and deformation, which are recorded by ~ 2.06 Ga monazite in pelitic schist and a migmatitic gneissosity that is known to be coeval based on the intrusive relationships of a dated amphibolite dyke. Other intrusive rocks include a pegmatite layer, which may be similar in age to a previously dated ~ 1.93 Ga orthogneiss in the southern part of the complex, and a 1862 ± 1 Ma orthogneiss that is coeval with a metamorphism recorded by titanite in amphibolitic gneiss. Rocks that predate 1.86 Ga formed the crystalline basement upon which the lower part of a cover sequence was deposited prior to 1.85 Ga. Basement was intruded by four granitoids dated at 1848 ± 3 to 1834 ± 2 Ma.

CORDILLERAN HISTORY

Chapter 3

An inverted metamorphic sequence is considered to exist across the Monashee décollement into the complex because sillimanite - K-feldspar rocks overlie kyanite-sillimanite-muscovite $\pm$ staurolite rocks. U-Pb isotopic data from Frenchman Cap dome were used to show that the age and intensity of this metamorphism, which is a result of Cordilleran orogenesis, vary with structural level. The amount of Pb loss that occurred during this event, as determined from linear arrays of discordant U-Pb isotopic analyses, was used to interpret the thermal significance of concordant and reversely discordant monazite and xenotime U-Pb dates that young downward from ~64 Ma in the highest level to ~49 Ma in the deepest level. Upper intercepts of the arrays are Paleoproterozoic and Cretaceous, and lower intercepts suggest that the discordance in the analyses, some of which are composed of single grains, resulted from an early Tertiary overprint. The discordance was mostly attributed to Pb loss rather than mineral overgrowth because the samples generally lack U-Pb evidence for growth at the lower intercept ages. The amount of Pb loss in the deepest structural level (6-22% in monazite and 25-31% in titanite) is less than that in moderately higher rocks (52-65% in titanite) and significantly less than that in much higher rocks (~30-56% in monazite and 94-96% in titanite).

Quantitative estimates of the metamorphic intensity were made by modeling Pb diffusion in episodes of heating of fixed duration and temperature, specific combinations of which were calculated by solving equations for measured fractional Pb loss and assumed diffusion radius and geometry. Diffusion coefficients were taken from natural

and experimental data. If peak temperatures were $\sim 700\text{--}750^\circ\text{C}$ and similar throughout all structural levels, consistent with mineralogy and a petrogenetic grid, then Pb loss occurred within 1 Ma in the deeper levels and within 2-5 Ma in higher levels. However, the differences in Pb loss amounts could also be due to the attainment of greater temperatures (up to 50°C) in higher levels.

The knowledge that Pb loss in monazite that existed throughout this metamorphism was partial led to the interpretation that the downward younging dates of 64-49 Ma from monazite and xenotime closely reflect the ages of mineral growth. These growth ages were taken to closely represent the ages of peak metamorphism based on (i) the calculated short durations of elevated temperatures, (ii) previously obtained $^{40}\text{Ar}/^{39}\text{Ar}$ dates that indicate rapid cooling occurred immediately after the peak, (iii) the widespread and consistent nature of the U-Pb dates, and (iv) U-Pb ages that indicate kyanite- and sillimanite-bearing leucosomes were generated at this time.

Chapter 4

The interpretations presented in Chapter 3 were used to conclude that peak metamorphism in the Monashee complex was diachronous according to structural level and metamorphic intensity increased upward. The thermal peak in higher levels was (i) synchronous with thrusting on the Monashee décollement, (ii) immediately followed by a period of rapid cooling that occurred during initial motion on overlying normal faults, which are associated with exhumation of the complex, and (iii) attained while deeper levels were at lower temperatures but presumably heating toward a thermal peak that

occurred a few million years later during the main period of extension on overlying faults. Dating of intrusive rocks with various relationships to the dominant Cordilleran deformation showed that deformation was synmetamorphic and lessens in intensity (a factor related to the amount of transposition) downward within basement rocks in deeper levels, locally having no effects.

The U-Pb data that suggest metamorphism was less intense with increasing structural depth agree with the mineral assemblage data by showing an inverted metamorphic sequence. Evidence for the associated thermal gradient being inverted was possibly revealed in deeper levels, where the pattern of downward decreasing metamorphic intensity exists in basement rocks that are known to have been only weakly strained since the Proterozoic. In higher levels, it is uncertain whether the thermal gradient was inverted because there are numerous east-directed structures capable of significant synmetamorphic displacements, such as kilometre-scale isoclinal folds with sheared limbs, that could have telescoped a once laterally extensive normal metamorphic sequence into an inverted sequence without ever placing warmer rocks over cooler ones. The resulting tectonic model contends that the inverted metamorphic sequence formed mainly by the juxtaposition of rocks with widely varying initial positions that reached peak temperatures at different times and in different positions within the orogen, and perhaps there was a minor component of local downward heat conduction.

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