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The Monashee décollement at Cariboo Alp, southern flank of the Monashee complex, southern British Columbia, Canada

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Abstract—The Monashee décollement is a crustal-scale compressional shear zone in the southern Canadian Cordillera; it is well exposed in alpine terrane and is imaged in the subsurface by LITHOPROBE seismic-reflection profiles. The décollement separates middle- to upper-crustal allochthonous assemblages from underlying middle- to lower-crustal rocks of the orogen. Development of the exposed part of the décollement at Cariboo Alp in the southern Thor–Odin culmination has included formation of major isoclinal folds, superposition of smaller scale folds during progressive deformation, and transposition of early formed fabrics and stratigraphic boundaries into the dominant SW-dipping mylonitic planar fabric. Hinge lines of early folds have been rotated toward the strong consistent SW-plunging mineral stretching lineation. As a result of NE-directed shearing of fold limbs and thrust-stacking of these folded sheets between a basal shear zone and overlying allochthonous assemblages, an imbricate zone developed within the middle crust during peak metamorphic conditions.

INTRODUCTION

The Monashee décollement is a crustal-scale compressional shear zone within the Canadian Cordillera; it links middle-crustal strain within the hinterland of the orogen to upper-crustal thrust faulting in the foreland (Brown *et al.* 1992, Cook *et al.* 1992 and references therein). This shear zone and the underlying Monashee complex are exposed in a tectonic window within the Omineca Belt of the hinterland (Fig. 1).

The Monashee complex consists of two assemblages: Early Proterozoic basement gneisses (Armstrong *et al.* 1991 and references therein, Parkinson 1991) and unconformably overlying Proterozoic to lower Paleozoic platformal metasedimentary rocks (Scammell & Brown 1990 and references therein). The complex, considered to be a remnant of the disrupted cratonic margin of North America, was deformed during Mesozoic to early Tertiary compression of the Cordilleran orogen and exhumed from middle- to upper-crustal levels by tectonic denudation during Eocene crustal extension (Brown & Journeay 1987, Parrish *et al.* 1988 and references therein).

The Monashee décollement is recognized regionally as a zone in which stratigraphy is completely transposed, fabrics are mylonitic, and overlying rocks record a different deformational and metamorphic history than that of the underlying Monashee complex (Brown 1980, Read & Brown 1981, Brown *et al.* 1986, Journeay 1986). The overlying rocks are at sillimanite grade, and rocks of the underlying complex locally preserve inverted isograds (Journeay 1986, Journeay & Brown 1986).

The overlying assemblages have been collectively referred to as the 'Selkirk allochthon' (Read & Brown 1981), a composite thrust sheet that includes slices of displaced North American continental margin deposits, accreted rocks, and plutonic and extrusive rocks which were emplaced during the evolution of the orogen. However, the interpretation that the décollement extends eastward beneath the Rocky Mountain Belt and westward beneath the Intermontane Belt (Brown *et al.* 1992), implies that the part of the orogen that lies above the décollement is allochthonous with respect to the underlying middle and lower crust. The regional validity of this notion of an 'orogenic float' (see Oldow *et al.* 1990, Brown *et al.* 1993) requires further testing, but in the southeastern Canadian Cordillera the data indicate that a middle-crustal shear zone or set of interrelated zones were generated in the Mesozoic, above which northeastward displacement, thickening, and shortening occurred.

Palinspastic restoration of the assemblages that currently overlie the Monashee complex indicates that the assemblages were transported ~300 km northeastward relative to the North American craton (Price & Mountjoy 1970, Brown *et al.* 1992). The magnitude of displacement of the allochthonous assemblages relative to the underlying Monashee complex is not well constrained but is considered to be at least 80 km (Read & Brown 1981, Brown *et al.* 1986). There is clear evidence of contractional strain within the complex, and it has been suggested that crustal-scale duplexing has shortened and thickened the complex during transport of the overlying allochthonous assemblages (Brown *et al.* 1986).

Our purpose in this paper is to present new data acquired from the southernmost exposed part of the Monashee complex at Cariboo Alp on the southern flank of the Thor–Odin culmination, where the Monashee décollement and the base of the overlying alloch-

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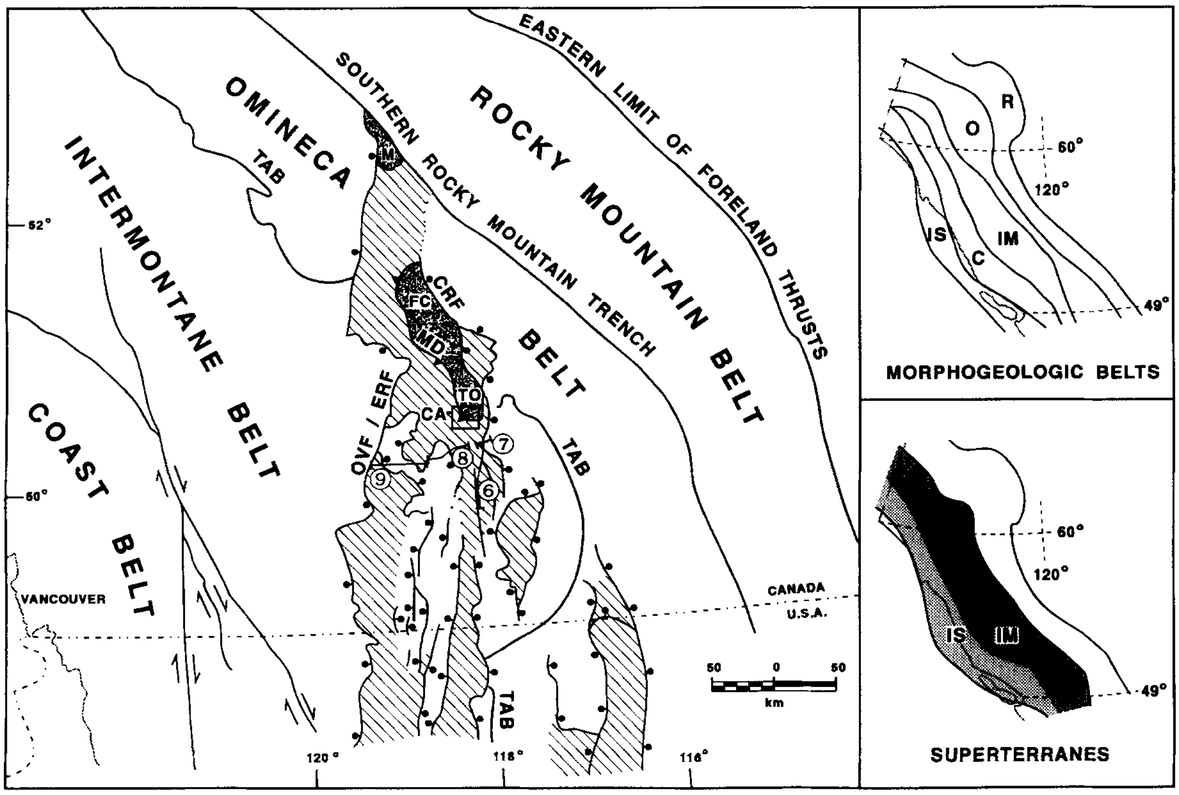

Fig. 1. Tectonic map of southern Omineca Belt modified from Parrish *et al.* (1988) and Brown *et al.* (1992). Insets locate the morphogeological belts of the Cordillera as well as the Intermontane (IM) and Insular (IS) superterranes. The Omineca Belt is bounded on the east by the southern Rocky Mountain Trench and on the west by the Okanagan Valley–Eagle River normal fault system (OVF/ERF) and the terrane accretion boundary (TAB). The Intermontane Superterrane (IM) lies west of the terrane accretion boundary. The diagonal stripe pattern marks the Shuswap complex. Basement rocks (mottled grey pattern) are exposed in Frenchman Cap (FC) and Thor–Odin (TO) culminations of the Monashee complex and in the Malton complex (M). MD marks the Monashee décollement, a Mesozoic–Paleocene thrust fault. Eocene normal faults include the Columbia River fault (CRF) and the Okanagan Valley–Eagle River fault system (OVF/ERF). Numbers 6–9 refer to LITHOPROBE lines across the Omineca Belt. Box labelled CA locates Cariboo Alp region (see Figs. 2 and 4).

thonous assemblages are beautifully displayed. We document the deformation and metamorphism associated with the history of the Monashee décollement, outline available timing constraints, and relate the structures within the shear zone to adjacent LITHOPROBE seismic-reflection data.

GEOLOGY OF CARIBOO ALP

The geology of the Cariboo Alp region is conveniently described in terms of three structural levels (Fig. 2). Penetratively strained but comparatively undisrupted Monashee complex basement gneisses with infolds of overlying cover gneisses form the deepest structural level, hereafter called the Monashee complex fold zone. In the middle level, imbricated and transposed Monashee complex basement and cover rocks occur in a highly sheared zone ~2 km thick; the top of this high-strain imbricate zone is defined by a marked change in stratigraphy, plutonism, and structural style and by a sharp metamorphic discontinuity. The overlying allochthonous assemblages of the highest level have been correlated with rocks of the Selkirk allochthon (Journeay & Brown 1986, Carr 1991).

Monashee complex fold zone

The Monashee complex fold zone is located in the Gladsheim Lake—Mt. Skade area and includes basement gneisses unconformably overlain by metasedimentary cover rocks. Detail of the geology in this area is shown in Fig. 3. In Thor–Odin culmination, basement rocks include both paragneisses and orthogneisses (see also Reesor & Moore 1971). The predominant basement lithology in southern Thor–Odin is migmatitic biotite–quartz–feldspar paragneiss (+/−garnet, +/−sillimanite, +/−kyanite +/− hornblende), which is characterized by strongly contrasting biotite-rich and quartzofeldspathic layers that alternate on the scale of millimetres to centimetres (Fig. 6a). The migmatitic paragneiss contains minor interlayers of semipelitic, pelitic, psammitic, calcsilicate gneisses and schists that are discontinuous along strike. Rare lenses of phlogopite–gedrite and gedrite–cordierite–kyanite–sillimanite–biotite schist are also present (see also Duncan 1984). Zones containing abundant amphibolite boudins, lenses and discontinous layers can be traced for up to 100 m. Very homogeneous, white-weathering felsic orthogneiss, ranging in composition from hornblende-biotite granodiorite to leucocratic quartz monzonite,

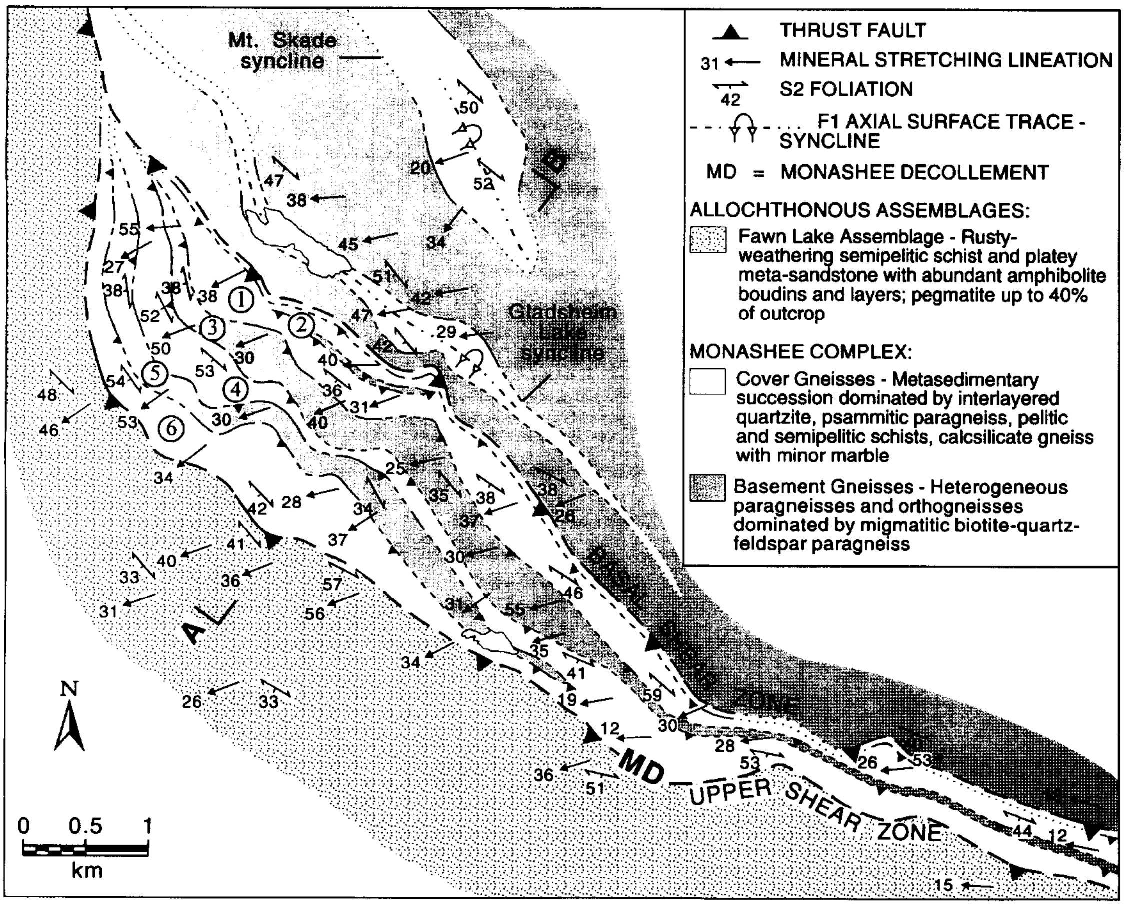

Fig. 2. Simplified geological map of Cariboo Alp region (location given by box in Fig. 1). Cross-section A–B is shown in Fig. 5.

forms monotonous sheets within the heterogeneous migmatitic paragneiss sequence.

The basement gneisses are unconformably overlain by a sequence of metasedimentary cover gneisses. The basement-cover unconformity is a regional feature; plutonic rocks within the basement gneisses are truncated at the contact of the basal quartzite of the cover gneisses and are not found within the overlying cover stratigraphy (Journeay 1986, Scammell & Brown 1990).

Within the basement rocks of southern Thor–Odin culmination, cover gneisses are preserved as infolds (see the Gladsheim Lake and Mt. Skade synclines on Fig. 3). Cover-gneiss stratigraphy in both of the synclines includes the following: a basal unit (a in Fig. 3) of quartzite with pelitic partings and local quartz pebble conglomerate; a heterogeneous paragneiss unit (b) composed of a succession of interlayered pelitic and semipelitic schists, calcsilicate gneiss with minor impure marble, psammitic gneiss, and minor quartzite; a second major quartzite unit (c) with interlayers of pelitic and semipelitic schists; and a paragneiss unit (d) dominated by psammitic gneisses with interlayers of diopsidic quartzite, rusty-weathering semipelitic and pelitic schists, and a diopside-bearing calcsilicate gneiss at the top of the sequence. Discontinuous layers and boudins of amphibolite gneiss are present throughout the entire cover-gneiss sequence.

A stratigraphic succession of the cover gneisses has been established, and correlations of laterally extensive layers have been made throughout the Monashee complex despite intense penetrative strain and recrystallization (see Scammell & Brown 1990 and references therein). Cover gneisses in southern Thor–Odin dome can be correlated with the lowest part of the cover-gneiss stratigraphy described in the northern Monashee complex. Preserved thickness of the cover gneisses in the southern Thor–Odin area is only about 200 m, as compared to 750 m between Frenchman Cap and Thor–Odin culminations and greater than 2000 m on the north flank of Frenchman Cap dome (Fig. 1).

The peak metamorphic assemblage in Monashee complex basement and cover gneisses is sillimanite+/−kyanite–potassium feldspar−melt. Kyanite and sillimanite coexist in these rocks, although kyanite is metastable and has been extensively replaced by synkinematic sillimanite. No isograd marking the appearance of stable kyanite has been found within the mapped area of Thor–Odin culmination.

A well developed SW-dipping foliation (designated S_2) and a penetrative SW-plunging mineral stretching lineation are the dominant structures observed in Monashee complex basement and cover rocks (Fig. 2). S_2 foliation is defined by schistosity and gneissic layering and is often mylonitic. It generally parallels transposed and recrystallized compositional layering in the metasedimentary rocks. S_2 contains minerals of the peak meta-

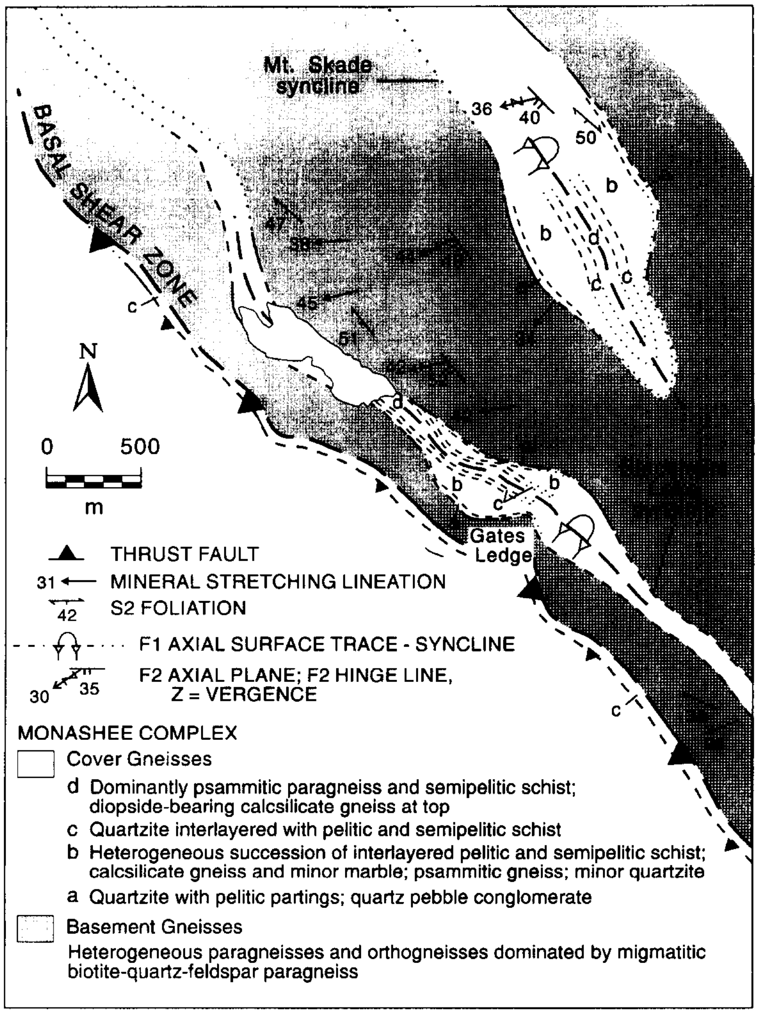

Fig. 3. Geology of the Monashee complex fold zone in the Gladsheim Lake–Mt. Skade area.

morphic assemblage and concordant lenses of leucosome. The mineral stretching lineation is defined by the orientation of elongate minerals including sillimanite, kyanite, hornblende and quartz rods. Fractured and pulled apart kyanite blades are commonly observed with sillimanite growth between segmented kyanite crystals. These planar and linear fabrics are well developed throughout the examined Monashee complex, but become less pronounced and more variable in orientation farther away from the overlying shear zone.

Three phases of folding that post-date the deposition of cover on basement rocks are preserved in Monashee complex rocks (Fig. 4). Deformation and metamorphism of the basement gneisses prior to the deposition of cover rocks probably occurred, but younger events have obscured any direct evidence.

The map-scale distribution of lithologic units is dominated by large first-phase folds (F_1), such as the Gladsheim Lake and Mt. Skade synclines, which are infolds of cover gneisses within the basement gneisses. These folds are well defined by stratigraphic units and have been mapped on the basis of stratigraphic symmetry (inversion and repetition of regionally continuous markers) (Fig. 3). The Gladsheim Lake and Mt. Skade synclines have long attenuated limbs; the overturned limbs are much more attenuated, resulting in structurally thinned or locally omitted stratigraphic units. Overall, cover-gneiss units in the Gladsheim Lake syncline are more structurally thinned than those in the Mt. Skade syncline; presumably this attenuation is the result of a closer proximity to the structurally overlying high-strain zone. Minor folds and fabrics associated with F_1 folding are rare as they have been overprinted and transposed by later folding, shear strain, and metamorphism.

The large F_1 folds are overprinted by syn-peak-metamorphic F_2 folds (Fig. 3). F_2 folds are N-verging, asymmetric, tight to isoclinal folds which plunge gently to moderately to the west and have axial planes oriented at varying low angles to the S_2 foliation. Annealed mineral assemblages and leucosome melt pods associated with the peak conditions of metamorphism are contained within the axial planes of F_2 folds.

F_3 folds are post-peak-metamorphic broad warps to tight crenulations that fold the mylonitic fabrics and the SW-plunging mineral stretching lineations. Open, upright F_3 folds plunge gently to the northwest and have steep SW-dipping axial planes. Randomly oriented retrograde minerals including chlorite, muscovite and epidote overprint these structures.

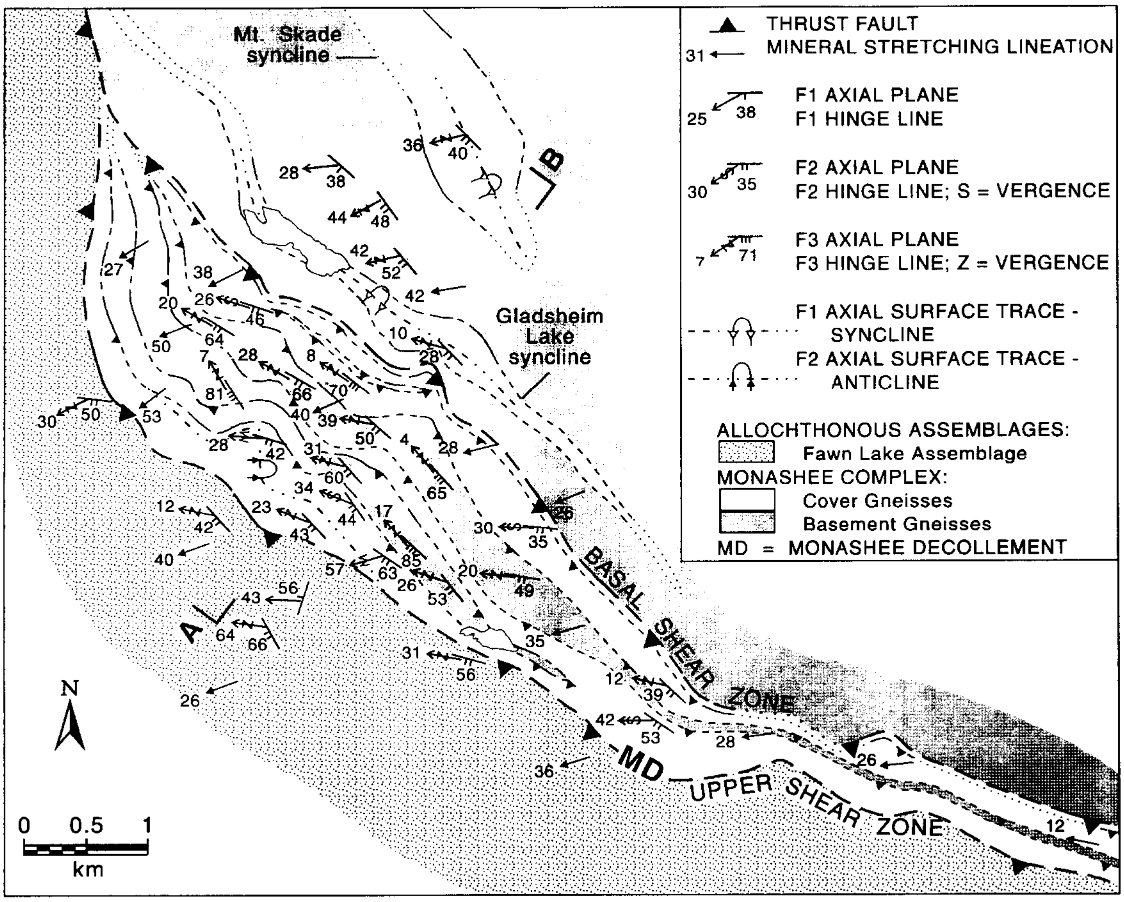

Fig. 4. Simplified structural map of Cariboo Alp region (location given by box in Fig. 1).

The amount of strain in Monashee complex fold zone rocks increases towards the overlying imbricate zone. The basal shear of the imbricate zone is marked by a large strain gradient between coherent stratigraphy of the Monashee complex fold zone and highly strained disrupted stratigraphic units above.

Monashee complex imbricate zone

This highly strained zone involving imbricated units of the Monashee complex is well displayed in Cariboo Alp (Fig. 2), where it is ~2 km thick. The lowest part of the imbricate zone at Gates Ledge is marked by isolated lozenges of foliated basement gneiss surrounded by more highly sheared gneiss overlain by several metres of strongly sheared quartzite (Fig. 3). This sheared quartzite is interpreted as correlating with unit c of the cover-gneiss stratigraphy, indicating the disruption and omission of stratigraphy at this lower boundary.

The imbricate zone is bound above and below by SW-dipping shear zones and contains SW-dipping fault-bounded slices of alternating Monashee complex basement and cover gneisses (labelled 1–6 on Fig. 2). Stratigraphic units have been completely transposed. A strong SW-dipping mylonitic foliation, which is parallel to the ductile thrust faults bounding the imbricate slices, and a penetrative down-dip, SW-plunging mineral stretching lineation are the most dominant features of this imbricate zone. NE-directed shear-sense indicators are ubiquitous and sheath folds are common.

The Cariboo Alp region contains the thickest section of the imbricate zone. Six map-scale slices vary between 50 and 500 m in thickness (Figs. 2 and 5). Figure 5 is a SW–NE cross-section through Cariboo Alp (location of cross-section is given in Fig. 2). East of Cariboo Alp, towards the trace of the Columbia River normal fault, the thrust slices are greatly attenuated, and some are omitted. On the western side of Thor–Odin culmination, thrust slices are attenuated into a narrower ductile fault zone that has no clear evidence of basement and cover-gneiss imbrication.

Since the imbricated lithologies within this strained zone are recognized as having been derived from the Monashee complex, we have placed them in the footwall of the Monashee décollement. Regionally, the décollement marks the zone of disruption of rocks of the Monashee complex and the sheared base of structurally overlying allochthonous assemblages.

Basement thrust slices (labelled 2, 3, 5 on Fig. 2) of the imbricate zone are composed mainly of migmatitic biotite–quartz–feldspar augen paragneiss with interlayers of sillimanite+/−kyanite–garnet–biotite pelitic schist and garnet–biotite-bearing quartzofeldspathic paragneiss (Fig. 6b). The basement paragneisses are characterized by distinct segregations of leucocratic and melanocratic minerals that define the foliation. Rare lenses of gedrite–cordierite–kyanite–sillimanite–biotite

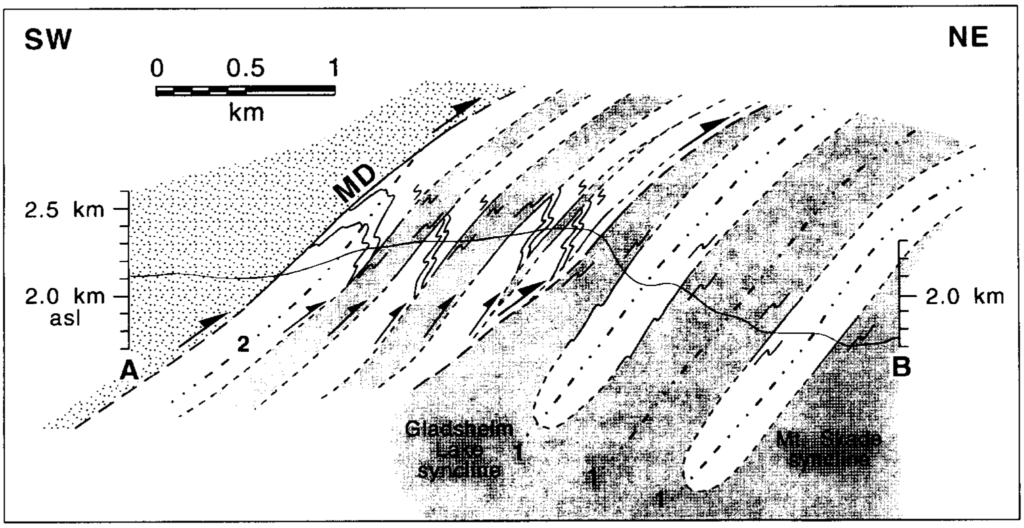

Fig. 5. SW–NE cross-section A–B through Cariboo Alp. Location of the cross-section and a description of the symbols is given in Fig. 2. Fold patterns represent the vergence of F_2 minor folds. The axial planes of F_1 and F_2 folds are marked by 1 and 2, respectively. Horizontal scale = vertical scale.

schist and phlogopite–gedrite schist lie parallel to the foliation. These rocks are correlated with Monashee complex basement rocks underlying the imbricate zone on the basis of lithological similarity. Correlation of basement thrust slices 2 and 5 with basement below the basal shear zone is unambiguous. Basement slice 3 contains siliceous paragneisses and layers, up to 20 m in thickness, with diopside–hornblende–quartz 'clasts' in a calcareous to siliceous matrix and cannot be easily correlated with basement rocks immediately below the imbricate zone, although these rocks do occur elsewhere in the basement of the Monashee complex.

Cover-gneiss thrust slices (labelled 1, 4, 6 on Fig. 2) are composed of interlayered paragneisses including: rusty-weathering migmatitic sillimanite–biotite–garnet +/−kyanite semipelitic schist with thin pelitic interlayers; thickly-bedded biotite–garnet-bearing psammitic gneiss; quartzite layers up to 5 m in thickness; and diopside-bearing calcsilicate gneiss with minor impure marble interlayers (Fig. 6c). Lithologies in the cover-gneiss slices can be correlated with lithologies in the infolds of cover gneiss below the imbricate zone. The stratigraphy is not coherent, however, as there are internal disruptions such as folding and imbrication of the units within the slices.

Boudins and discontinuous layers of amphibolite gneiss are common, both within the basement and cover-gneiss slices. A suite of undeformed sillimanite-bearing pegmatite dykes cut through all the thrust slices of the imbricate zone (Fig. 6d).

The annealed peak pelitic metamorphic assemblage within Monashee complex rocks of the imbricate zone and underlying fold zone includes sillimanite +/−kyanite–K-feldspar−leucosome melt pods. Peak assemblages containing both sillimanite and kyanite occur within the mylonitic foliation and the axial planes of F_2 folds. Sillimanite and kyanite are also oriented parallel to the mineral stretching lineation.

Although some pristine blue kyanite exists, most of the bladed kyanite grains have been almost completely replaced and overgrown by oriented syn-kinematic sillimanite and sillimanite–biotite–feldspar mats (Fig. 6e). In thin section, crystals of kyanite are kinked and ragged in appearance and have sweeping extinction. These textures suggest that kyanite is a metastable phase of the metamorphic assemblage. Mats of sillimanite occur on discrete shear surfaces that truncate F_2 folds and related fabrics, thereby post-dating both kyanite and older sillimanite. Based on these textural observations, a high-pressure assemblage containing kyanite is inferred to have nucleated either prior to or during the initial stages of deformation and subsequently has been replaced by syn-kinematic sillimanite as deformation proceeded.

There is a very strong SW-dipping mylonitic foliation (S_2) in rocks of the imbricate zone (Fig. 2). This foliation is parallel to the compositional layering and represents a composite fabric into which earlier formed fabrics have been transposed. It contains a strong, consistent, penetrative, SW-plunging mineral stretching lineation (Fig. 6f), which is defined by the preferred alignment of elongate minerals including sillimanite, fibrolite–biotite mats, kyanite, hornblende and quartz and by the long axes of pressure shadows composed of quartz, feldspar and biotite. Some of the prismatic blades and mats of sillimanite are pulled apart parallel to the mineral stretching lineation and have muscovite in the pull-aparts.

The transport direction of these strongly sheared rocks is interpreted as approximately parallel to the mineral stretching lineation (Hanmer & Passchier 1991 and references therein). The sense of shear was determined using an assemblage of mesoscopic and microscopic asymmetric deformation structures that developed during the shearing process. Shear-sense indicators, which were studied in outcrop, hand sample, and thin section, include the following: *S–C* fabrics (Berthé *et al.* 1979), asymmetrical extensional shear bands (Platt & Vissers 1980, White *et al.* 1980, Hanmer

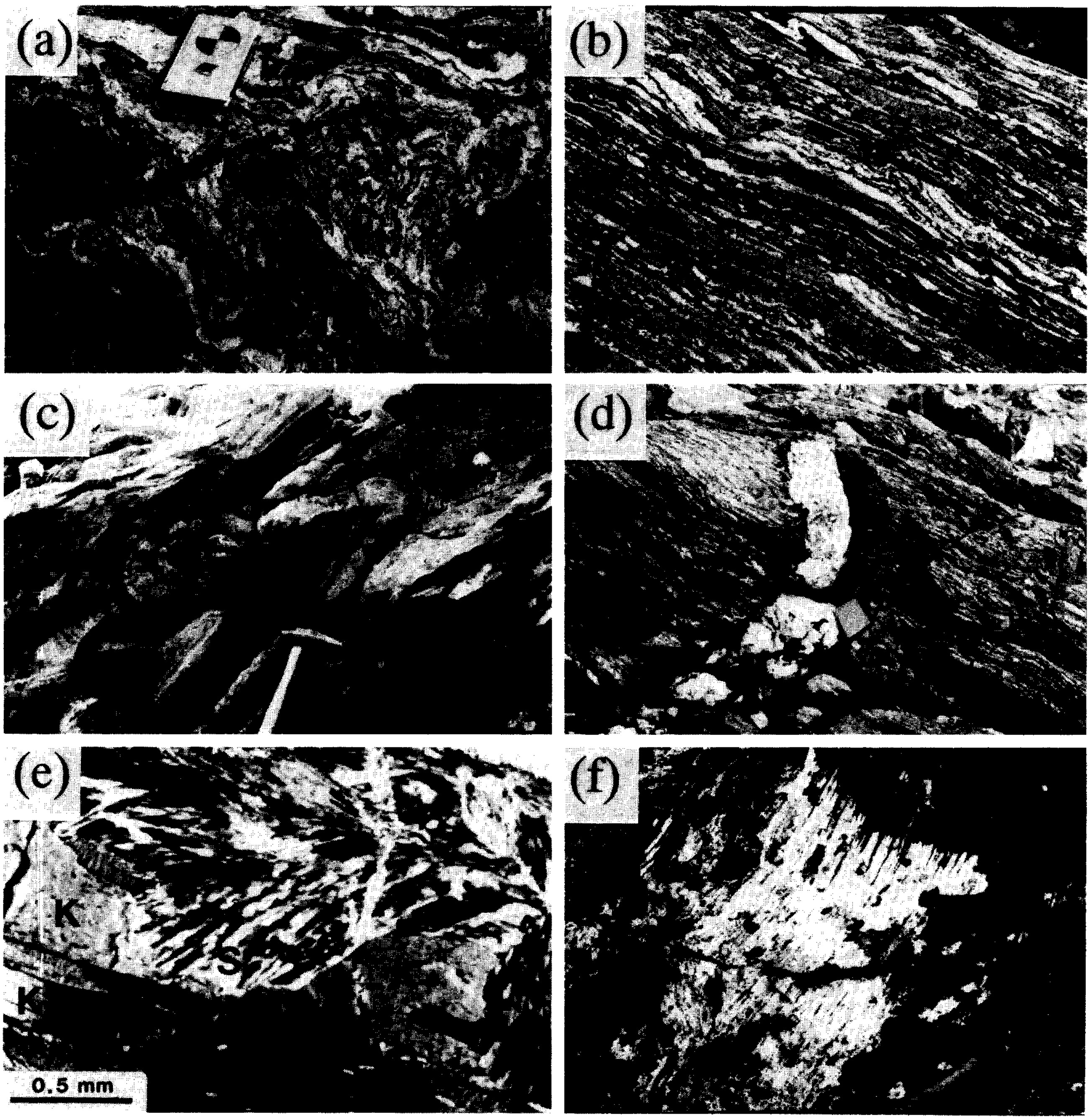

Fig. 6. (a) Banded migmatitic biotite–quartz–feldspar paragneiss, characterized by strongly contrasting biotite-rich and quartzofeldspathic layers, on Mt. Gunnarsen within the Monashee complex fold zone. Interference pattern is a result of the superposition of F_2 folds on F_1 folds. (b) Migmatitic biotite–quartz–feldspar paragneiss from a basement thrust slice of the imbricate zone. Photo is taken looking southeast, on a face parallel to the mineral stretching lineation. Asymmetric pinch and swells in ductile boudins and asymmetrical extensional shears give a NE-directed sense of shear. (c) F_2 fold deforming psammitic gneiss, semipelitic schist, calcsilicate gneiss, and impure marble of the cover gneiss slice 6 of the imbricate zone. (d) Discordant, undeformed pegmatite from a suite of pegmatite dykes that cut across all the thrust slices and structures of the imbricate zone in the Cariboo Alp region. This pegmatite dyke has been dated at 58 +/− 0.5 Ma (Carr 1992). (e) Kyanite (K) replaced by fibrous sillimanite (S), which is somewhat unoriented. Note bent kyanite grain with undulose extinction below the large kyanite grain. X-nicols. (f) Typical strong SW-plunging mineral stretching lineation, defined by the orientation of mats of sillimanite, within basement paragneisses of the imbricate zone.

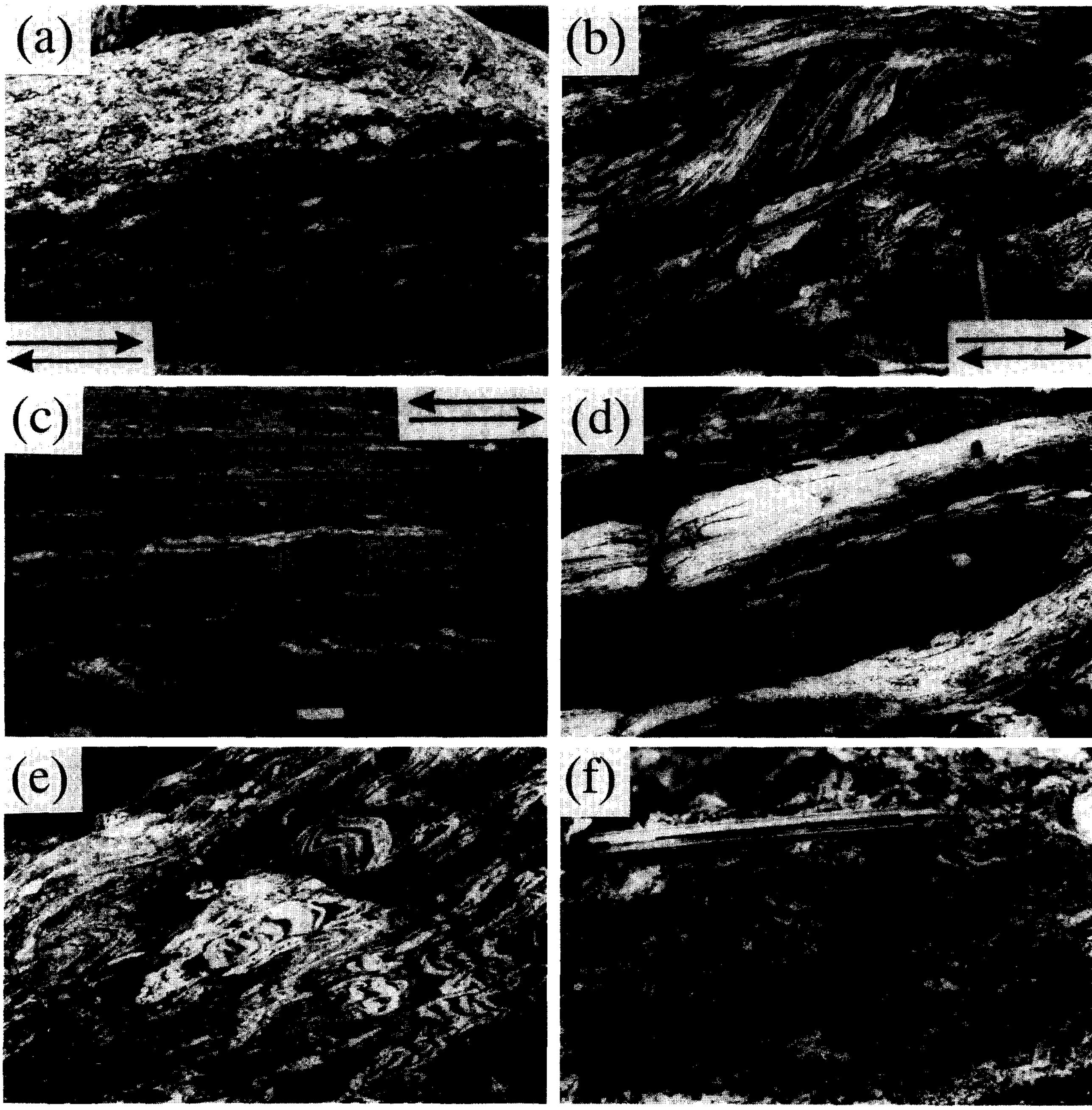

Fig. 7. (a) Sheared gneisses in the immediate hangingwall of the Monashee décollement. *C–S* fabrics and asymmetrical extensional shears indicate a NE-directed sense of shear. The light coloured layer is a concordant deformed pegmatite with a protomylonitic fabric. (b) Small back-rotated lozenge of mylonite within mylonite gives a NE-directed sense of shear in deformed basement gneisses of the imbricate zone. (c) Back-rotated asymmetrical boudins within sheared basement gneisses of the Monashee complex imbricate zone indicate a NE-directed sense of shear. Discrete zones of shear (analogous to asymmetrical extensional shears) are present at the ends of the boudins, oriented at angles less than 30° to the shear plane. The ends of the boudins and the internal layering are deflected into these discrete shears. (d) F_1 rootless isoclinal fold in basement rocks with fold hinge parallel to the mineral stretching lineation. F_2 *z*-folds overprint this F_1 fold. (e) Transposed F_2 folds with sheared-off limbs in basement slice 5 of the Monashee complex imbricate zone. (f) F_3 folds in a basement slice of the imbricate zone that crenulate the mylonitic fabrics and the sillimanite and biotite that define the mineral stretching lineation.

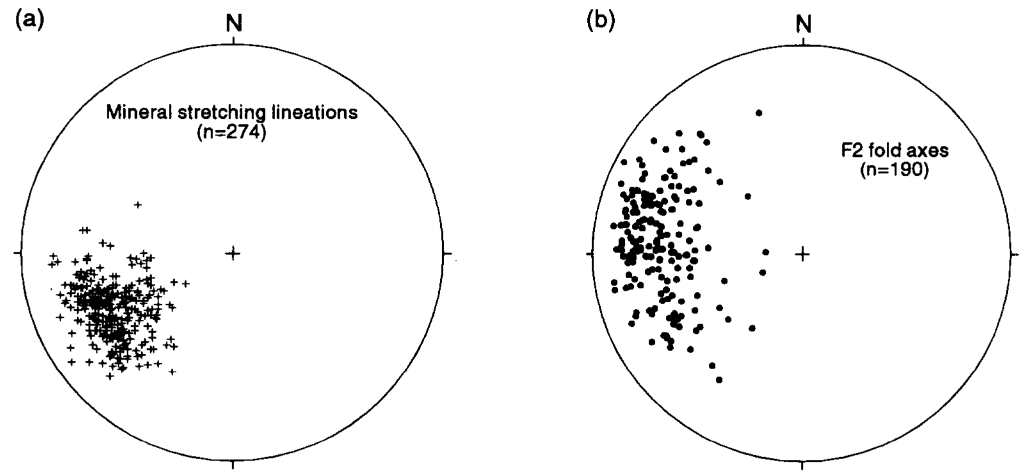

Fig. 8. Lower hemisphere stereonet plots of (a) mineral stretching lineations (crosses) and (b) fold axes of F_2 folds (filled circles) within the Monashee complex imbricate zone. The fold axes of tight to isoclinal F_2 folds with highly attenuated limbs are subparallel to the mineral stretching lineation, whereas the axes of more open F_2 folds are inclined at a higher angle to the lineation.

& Passchier 1991 and references therein), rotated porphyroclasts (Passchier & Simpson 1986, Van den Driessche & Brun 1987), back-rotated foliation segments (Hanmer 1984, 1986, Lister & Snoke 1984) and strain-insensitive foliations (Means 1881, Hanmer & Passchier 1991 and references therein) (Figs. 7a–c). Shear-sense indicators consistently record a NE-directed sense of shear; a diverse assemblage shows remarkable consistency within single outcrops, from outcrop to outcrop, and across kilometres of terrain. Mesoscopic asymmetric structures are generally better preserved than the microstructures; high syn- and post-tectonic temperatures apparently induced recrystallization and grain-boundary adjustments that tend to mask the strain-induced microstructures.

Three phases of folding are preserved in rocks of the imbricate zone (Fig. 4). F_1 minor folds are N- to NE-verging overturned structures that occur in outcrop as refolded folds (type 3 interference patterns with F_2 folds) or as rootless isoclinal folds with hinges parallel to the mineral stretching lineation (Fig. 7d). The limbs of minor F_1 folds display moderate to extreme attenuation. Minor structures and fabrics associated with this phase of folding are rare: they have largely been overprinted and transposed by later high-grade metamorphism, folding, and shear strain. The axial planar cleavage associated with F_1 folds is defined by the planar preferred alignment of biotite, elongate quartz and feldspar. A lack of sillimanite in the axial planes indicates that F_1 folds formed prior to the peak of metamorphism. A symmetrical stratigraphic repetition of distinctive units in basement slice 3 of the imbricate zone outlines a possible remnant of a map-scale F_1 fold (Coleman 1990).

Syn-peak-metamorphic N- to NE-verging F_2 folds plunge gently to moderately to the west to southwest and are the dominant fold type observed in outcrop (Figs. 6c and 7e). They are asymmetric, tight to isoclinal folds with axial planes at varying low angles to the mylonitic foliation. In this zone of high strain, F_2 folds with highly attenuated sheared-off limbs are common (Fig. 7e). F_1 and early F_2 folds have been rotated, resulting in sheath fold geometries, some up to 1 m in size.

F_2 folding was at least partly coeval with the development of fabrics related to shearing in the imbricate zone. At the outcrop scale, F_2 minor folds both deform the mylonitic foliation and are disrupted and truncated by superimposed discrete shears. Within the imbricate zone, the vergence of F_2 minor folds changes abruptly at each thrust contact of the slices (see Figs. 4 and 5). Each of the slices is interpreted as a preserved sheared-off limb of large map-scale F_2 folds (see discussion below). A large F_2 overturned antiform in the uppermost cover thrust slice (6) is truncated by the upper shear zone (Monashee décollement) of the imbricate zone.

There is a variation in the geometry and orientation of F_2 minor folds related to the amount of strain they have undergone. Fold axial planes have been rotated with progressive tightening of F_2 folds: the axial plane of a tight F_2 fold is at a lower angle to the mylonitic foliation compared to the axial plane of a more open F_2 fold. F_2 fold axes have been variably rotated towards parallelism with the mineral stretching lineation: the fold axes of tight to isoclinal F_2 folds with highly attenuated limbs are subparallel to the mineral stretching lineation, whereas the axes of more open F_2 folds are inclined at a higher angle to the lineation (Fig. 8). F_2 folds are interpreted as having formed during shearing deformation and then became progressively tightened, flattened, and rotated towards the shear plane during continuing flow in this highly strained zone (cf. Quinquis *et al.* 1978, Cobbold & Quinquis 1980).

Open to tight upright F_3 folds plunge gently to the northwest and have steep SW-dipping axial planes, which are defined by the preferred orientation of biotite+/−garnet. F_3 folds formed as peak metamorphic conditions were waning: they are broad warps to tight crenulations that fold the mylonitic fabrics and the SW-plunging mineral stretching lineations (Fig. 7f). Locally,

F_2 folds are refolded by F_3 folds to form basin-and-dome interference patterns at the outcrop scale. No major map-scale F_3 folds were mapped within the imbricate zone or the underlying Monashee complex fold zone. F_3 structures are overprinted by randomly oriented retrograde minerals including chlorite and muscovite.

Allochthonous assemblages

The top of the imbricate zone (the Monashee décollement, labelled MD on Fig. 2) is defined by a marked change in stratigraphy, plutonism, and structural style and by a sharp metamorphic discontinuity. Immediately overlying this zone in Cariboo Alp is a well exposed, 1.5–3 km thick package of metasedimentary rocks consisting mainly of semipelitic schists and quartzofeldspathic paragneisses intruded by extensive concordant layers, boudins, and cross-cutting dykes of pegmatite. This package of rocks has been called the Fawn Lake assemblage (Carr 1991). Rusty-weathering semipelitic schists, which are interlayered with very thin pelitic schists, are migmatitic with leucosome pods and lenses forming up to 40% of the rock by volume. Quartzofeldspathic paragneisses are thinly layered and consist mainly of roughly 5 cm thick mica-poor layers separated by thin schistose partings, giving the rocks their characteristic 'platy' appearance. Quartzite, marble, and diopside-bearing calcsilicate gneiss make up a minor portion of this sequence of rocks. Boudins and layers of amphibolite gneiss are very abundant, forming up to 30% of the outcrop in some places.

The metasedimentary package is correlated with rocks of the Late Proterozoic Windermere Supergroup based on lithological similarities (Journeay & Brown 1986, Carr 1991). Similar correlations have been made for migmatitic paragneisses of the Shuswap assemblage in the Hunters Range (50 km NW of Cariboo Alp) (Johnson 1988, 1990) and for rocks in the immediate hangingwall of the Monashee décollement on the western and northwestern flanks of the Monashee complex (Brown 1980, Brown & Read 1983, Journeay 1986, Scammell 1986, 1990).

This package of rocks is intruded by extensive Late Paleocene–Early Eocene leucocratic granite and pegmatite of the Ladybird suite (Carr 1991, 1992). Below the Monashee décollement, leucogranite is found primarily as a suite of discrete dykes (Fig. 6d). The amount of leucogranite substantially increases in the allochthonous rocks of the immediate hangingwall and includes synkinematic sheets, plutons, and pegmatite and post-kinematic batholiths, plutons, and pegmatite dykes (Carr 1992).

The peak pelitic metamorphic assemblage of rocks of the Fawn Lake assemblage is sillimanite–*K*-feldspar − leucosome melt (see also Carr 1990, 1991). Sillimanite is the only aluminosilicate and occurs as fibrous sprays and mats on foliation planes; no kyanite has been found above the Monashee décollement at the top of the Monashee complex imbricate zone.

Penetrative strain in the allochthon, related to its northeastward displacement, extends approximately 5 km upward from its base. The intensity of strain decreases southward, structurally higher above the Monashee décollement. A strong, consistent SW-dipping mylonitic foliation in the hangingwall rocks, designated S_2, contains the peak metamorphic assemblage, including concordant lenses of leucosome. The S_2 foliation contains a penetrative SW-plunging mineral stretching lineation defined by the orientation of sillimanite, hornblende, fibrolite–biotite mats and quartz rods.

At least three phases of folding have been recorded in allochthonous rocks of the Fawn Lake assemblage (Fig. 4; see also Carr 1991). Structures are similar in geometry and orientation to structures of the Monashee complex in the footwall; however, there is a marked decrease in the abundance of minor folds and a much more planar appearance to this belt of rocks. Rare pre-peak-metamorphic F_1 folds are documented as refolded folds and as isoclinal folds with hinges parallel to the mineral stretching lineation. Syn-peak-metamorphic F_2 folds are tight to isoclinal N-verging folds and are the dominant fold type in outcrop. F_2 fold hinges have been rotated towards the mineral stretching lineation; sheath folds are locally present. Post-peak-metamorphic F_3 folds deform the mylonitic fabrics and crenulate the penetrative mineral stretching lineation.

Columbia River normal fault

The E-dipping ductile–brittle Columbia River normal fault truncates the Monashee décollement on the eastern margin of the Monashee complex (Lane 1984 and references therein) (Fig. 1). Geochronology and fabric relationships of footwall mylonites in the South Fosthall pluton south of the Thor–Odin dome constrain movement on this fault to the Eocene (Parrish *et al.* 1988).

In the footwall of the Columbia River fault, upper-amphibolite-facies rocks of the Monashee complex and overlying allochthonous assemblages are overprinted by greenschist-facies mineral assemblages and unannealed mylonitic fabrics. Sillimanite+/−kyanite-bearing mineral stretching lineations have been transposed into an E–W orientation and are overprinted by chlorite–muscovite-bearing assemblages. These lower grade mineral assemblages and fabrics occur in a belt approximately 3 km in width that parallels the Columbia River fault and are interpreted to have been generated during extensional ductile deformation (see also Carr 1991).

Geochronologic constraints

Tight timing constraints have been placed on the final displacement of the Monashee décollement by U–Pb dating of zircon and monazite. Deformed sillimanite-bearing 62–59 Ma syn-tectonic leucogranites above the décollement provide evidence that compression was still ongoing in the Late Paleocene. Compressional deformation was finished by 58 ± 0.5 Ma, the U–Pb zircon age of a cross-cutting pegmatite dyke (Carr 1992). This dated dyke is from a suite of discordant, undeformed sillimanite-bearing pegmatite dykes that cut across all of the structures and thrust slices in the Cariboo Alp region (Fig. 6d).

Mineral dating suggests that the metamorphic peak occurred in the Late Cretaceous–Paleocene with the culmination of compressional deformation. In the immediate hangingwall of the Monashee décollement, *ca.* 62 Ma magmatic zircons within strained sillimanite-bearing migmatite leucosome indicate that the onset of melting occurred in the latest Paleocene (Carr 1992). A Late Paleocene thermal peak can also be documented in footwall rocks of the Monashee complex with the growth of magmatic zircons in leucosome melt pods of deformed basement paragneiss, metamorphic monazite in basement and cover paragneisses (Coleman 1990), and metamorphic zircons in amphibolite gneiss (Carr 1990). These data suggest that the syn-peak-metamorphic folding and transposition foliation that are preserved in the Cariboo Alp region developed in the Paleocene during the final stages of emplacement of the allochthonous assemblages.

SEISMIC DATA

LITHOPROBE seismic-reflection profiles to the south of the surface exposure of the Monashee décollement (Fig. 1) have recently been interpreted by several authors (Carr 1991, Brown *et al.* 1992, Cook *et al.* 1992). A comprehensive review of these interpretations is beyond the scope of this paper; in the following, we give consideration to those results that have a direct bearing on the subsurface nature of the shear zone. The composite transposition foliation (S_1–S_2) of the Monashee décollement zone dips to the west on the western flank of the Monashee complex and to the south on the southern flank. This change in strike of the foliation is reflected in its apparent dip as interpreted from the seismic-reflection data of profiles 6 and 7 (Fig. 9). The NS-trending profile 6 is approximately normal to the surface strike of the foliation at the southern end of the Monashee complex. The moderate southerly dip (35–50°) of the foliation at the surface is seen on the reflection profile of the subsurface extending down through the middle crust to an estimated depth of 20 km (6–7s two-way travel time), where the reflections flatten out to become sub-horizontal in the lower crust. Westerly apparent dips are recorded in the EW-trending profiles 7, 8 and 9 with a similar change to sub-horizontal attitudes toward the base of the middle crust. From the surface data we have concluded that the allochthonous rocks have been displaced northeastward. Scammell & Brown (1990) have suggested that the shear zone ramps upwards to the northeast in the footwall. To the extent that the orientation of the subsurface reflectors record an orientation that existed during thrusting, the data support the presence of a lateral ramp along the southern margin of the Monashee complex; however, it is recognized that the dips of foliation have been modified by post-thrusting extensional events.

In this paper, we describe the structural fabric of the shear zone related to the Monashee décollement with particular emphasis on the rocks of the footwall. Carr (1991) has presented a detailed description of the rocks in the hangingwall. Together these descriptions account for a crustal thickness of over 7 km of intensely sheared and well layered mylonitic rocks. The excellent exposure and detailed mapping of these rocks indicate that the shear zone has crustal dimensions and that the projection of these highly sheared rocks below the surface correlate to the reflectors recorded in LITHOPROBE profiles.

DISCUSSION

Minor structures developed within the Monashee complex imbricate zone correlate with those observed below the basal shear within the Monashee complex fold zone. There is no indication of a metamorphic break across the boundary, and F_2 folds above and below exhibit the same relationship to peak metamorphic assemblages. F_1 folds below the basal shear zone are preserved as coherent infolds of cover gneisses in the basement; the rootless F_1 folds within the imbricate zone are interpreted as highly sheared and transposed equivalents of these structures. F_3 folds both above and below the basal shear zone are late minor structures that most likely formed at the same time, after completion of major shearing on the Monashee décollement.

There are insufficient data to determine the amount of shortening that occurred during formation of the imbricate zone. A minimum estimate of 15 km of shortening is obtained by 'unstacking' the slices and arranging them in stratigraphic order immediately below the basal shear zone, but this conservative estimate does not account for the dissimilarities in the stratigraphy of the cover and basement rocks above and below the shear zone.

The thrust slices of the imbricate zone at Cariboo Alp are interpreted as a result of progressive tightening, limb attenuation, and eventual disruption of F_2 folds. With progressive tightening of folds and transposition of compositional layering, the limbs of the folds became attenuated and ultimately disrupted as shear zones propagated through the structures. At the present level of erosion, we see alternating slices of Monashee complex basement and cover gneisses with discontinuities in vergence of F_2 folds. The imbricated zone with upper and lower bounding shear zones resembles a duplex structure as proposed by Coleman (1989, 1990). This geometrical similarity, however, does not imply a kinematic origin analogous to duplexes in foreland thrust belts (Boyer & Elliott 1982).

Duncan (1984) suggested that displacement of fault-bounded slices in the Cariboo Alp region was N-directed, at right angles to local F_2 fold axes. He apparently did not recognize the significance of the uniform attitude of the pervasive NE–SW-oriented mineral stretching lineations and abundant NE-directed shear-sense indicators. As pointed out earlier in the paper, we

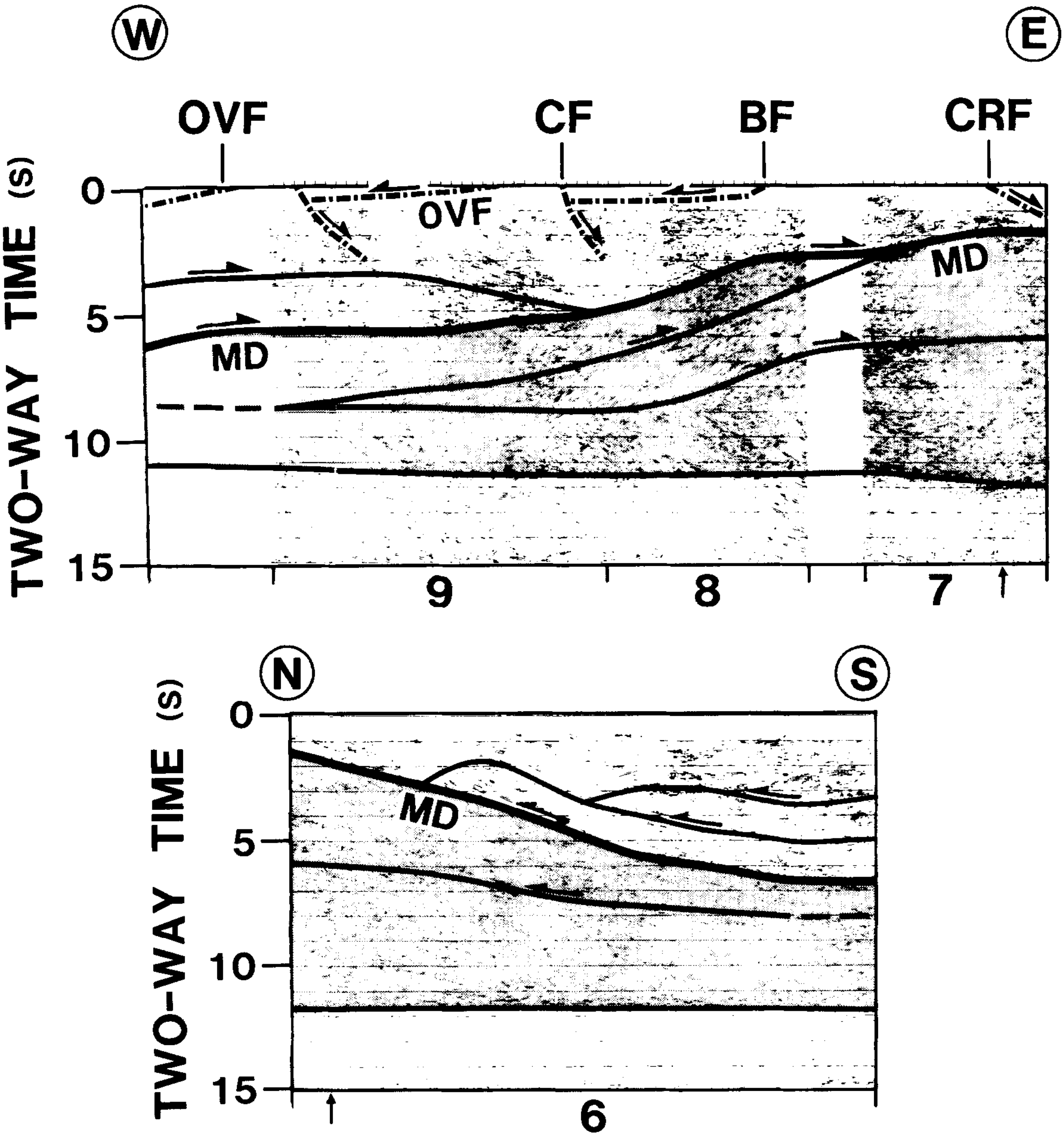

Fig. 9. LITHOPROBE seismic-reflection data with structural interpretation from lines 6–9 across the Omineca Belt. Shaded pattern represents crystalline basement. The Monashee décollement is shown with a bold line. Eocene normal faults are shown as alternating long and short dashes: OVF = Okanagan Valley fault, BF = Beaven fault, CF = Cherryville fault, CRF = Columbia River fault. See Fig. 1 for location of lines. Note that arrows representing movement to the northeast point to the right on W–E cross-sections and to the left on the N–S cross-section for line 6.

interpret the mineral stretching lineation to reflect the direction of displacement of the allochthonous assemblages, at least during the final stage of emplacement.

Tight timing constraints have been placed on the culmination of compressional deformation; however, there are no direct data to constrain the onset of displacement on the Monashee décollement. Shortening of the Rocky Mountain Belt is thought to have been underway by the latest Jurassic (Price & Mountjoy 1970), and it has been proposed that the Monashee décollement was kinematically linked to the sole thrust of the Rocky Mountain Belt throughout its history (see Brown *et al.* 1992).

Pre-peak-metamorphic structures such as major isoclinal folds within the Monashee complex fold zone are undated. Since these structures are NE-verging and appear to record the early stages of progressive shearing within the complex, determination of the age of these structures may clarify the time of earlier thrusting.

CONCLUSIONS

The Monashee décollement marks the western and southern boundary of the Thor–Odin culmination, separating allochthonous rocks from those of the underlying Monashee complex. Contrasting lithologies and intrusive layers, together with well developed mylonitic planar fabrics, are considered to be the cause of reflectors in LITHOPROBE seismic profiles that track the shear zone down to the lower crust.

In Cariboo Alp on the southern flank of the Thor–Odin culmination, NE-directed displacement on the Monashee décollement resulted in an imbrication of footwall rocks of the Monashee complex. Development of the shear zone outlasted early pre-peak-metamorphic F_1 folding, which included formation of major isoclinal folds, and culminated with syn-metamorphic F_2 folding. Progressive deformation at middle-crustal levels transposed early structures and stratigraphic boundaries into the dominant SW-dipping mylonitic planar fabric. Hinge lines of early folds were rotated toward the strong consistent SW-plunging mineral stretching lineation.

A significant part of the deformation recorded at Cariboo Alp occurred in the Late Cretaceous–Paleocene with the culmination of penetrative compressional deformation and with the accompanying peak metamorphism. NE-directed mylonitic fabrics are cut by an undeformed 58 ± 0.5 Ma dyke (Carr 1992), indicating that movement had ended by this time.

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